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(Article begins on next page)

1 **Evidence of subduction-related components in sapphirine-bearing gabbroic dykes (Finero**  
2 **Phlogopite–Peridotite): Insights into the source of the Triassic–Jurassic magmatism at the**  
3 **Europe–Africa boundary**

4

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26

27 **Abstract**

28 A gabbroic dyke swarm containing magmatic sapphirine occurs in the Finero phlogopite–peridotite  
29 (FPP), one of the major mantle massifs in the Ivrea–Verbano Zone (IVZ; western Southern Alps).  
30 Sapphirine is part of a particular mineral assemblage, including plagioclase, titanian pargasite,  
31 titanian phlogopite, and Cl-rich apatite; the latter mineral hosts calcite inclusions. The dykes cut the  
32 mantle foliation at a high angle, are bounded by orthopyroxenite layers, and show symmetric  
33 internal banding, represented by two outer hornblendite selvages and an inner leucogabbro band.  
34 The sapphirine occurs in up to 3 cm-thick irregular patches in both hornblendite salvages, along  
35 with Al-rich amphibole and green spinel.

36 We present major and trace elements of minerals and bulk rock, as well as mineral O, Sr, and Nd  
37 isotopic compositions of dykes and the host peridotite from two different outcrops in the FPP area.

38 Our data show that early melt migration developed through porous flow within cm-thick channels  
39 and was characterised by orthopyroxene dissolution. Following progressive percolation and  
40 reaction, the melt became silica saturated with segregation of orthopyroxenite in the centres of the  
41 channels. The banded internal structure of the dykes was caused by three different evolutionary  
42 stages, involving opening and enlargement of the conduits. The sapphirine and green spinel  
43 segregation took place at  $T > 1,000$  °C, in the presence of melt with transient composition, which  
44 interstitially migrated and reacted with the cumulus minerals to form the hornblendite layers. The  
45 mineral chemistry of the newly-formed amphiboles indicates that the sapphirine parental melt was  
46 Al-rich, depleted to strongly depleted in Hf, Zr, Nb, Ta, Ti, Sc, V, and middle and heavy rare earth  
47 elements, and characterised by a positive Eu anomaly and  $(Zr/Hf)_N < 1$ . These data suggest a  
48 parental melt with a significant amount of normative plagioclase. However, the studied veins do not  
49 show evidence of plagioclase assimilation, and we argue that this process could have occurred in  
50 magmatic bodies that are not outcropping today to the surface or in the melt source.

51 The  $\delta^{18}O$  values of vein amphiboles and plagioclases vary from 6.9 to 8.6‰ SMOW, which is well  
52 above the mantle range, even when considering fractionation upon cooling. Given that

53 orthopyroxene from the wall has “normal” mantle  $\delta^{18}\text{O}$  values (5.8‰), reaction with the host  
54 metasomatised peridotite cannot be responsible for the heavy  $\delta^{18}\text{O}$  signature, and the latter must  
55 have been imparted by crustal components deeper in the mantle.

56 Our petrographic and geochemical evidence demonstrates that the northern IVZ records an  
57 extremely prolonged release, from the Variscan orogenic cycle to the Mesozoic exhumation, of K-  
58  $\text{H}_2\text{O}$ -rich mantle-derived melts, mixed with subduction-related components. This finding provides  
59 valuable insights into the Triassic–Jurassic magmatism and the geodynamic environment at the  
60 Europe–Africa boundary.

61

## 62 **1. Introduction**

63 It is well known that the subcontinental lithospheric mantle (SCLM) may record significant  
64 heterogeneities in terms of both lithology and geochemical composition, resulting from the  
65 development of different tectono-magmatic events over a long time span (Mukasa and Shervais,  
66 1999; Rivalenti et al., 2007a, b; Mazzucchelli et al., 2009; 2010; 2016; Borghini et al., 2017;  
67 Princivalle et al., 2014; Ponce et al., 2015; Rocco et al., 2017; Consuma et al., 2019). The  
68 petrochemical records of the lithospheric mantle and uprising melts can be very different based on  
69 the geochemical affinity of the melt, the composition of the peridotite mantle column, the  
70 modalities of melt migration, and the  $P$ ,  $T$ ,  $f\text{O}_2$ , and  $f\text{H}_2$  conditions of the system. An exceptional  
71 locality where this can be studied in great detail is the phlogopite–peridotite mantle unit of Finero  
72 (Cawthorn, 1975; Siena & Coltorti, 1989). The Finero phlogopite–peridotite (FPP) is one of the  
73 most studied mantle massifs on Earth. It crops out in the northernmost part of the Ivrea–Verbano  
74 Zone (IVZ, western Southern Alps, Italy; Fig. 1) and consists of an association of phlogopite-  
75 bearing amphibole harzburgites and dunites, both locally associated with phlogopite-bearing  
76 amphibole pyroxenites. This is the result of an important episode of pervasive to channelled porous  
77 flow migration of melts containing a large volume of crustal components that induced virtually  
78 complete metasomatic recrystallisation of an older sequence of harzburgite and pyroxenite

79 associations (Lensch, 1968; Cawthorn, 1975; Griffin & O'Reilly, 1986; Siena & Coltorti, 1989;  
80 Shervais and Musaka, 1991; Hartmann & Wedepohl, 1993; Zanetti et al., 1999, 2016; Grieco et al.,  
81 2001, 2004; Morishita et al., 2003, 2008; Raffone et al., 2006; Selverstone & Sharp, 2011;  
82 Mazzucchelli et al., 2014; Tommasi et al., 2017). The crustal affinity of these metasomatic melts is  
83 implied by high  $\delta^{18}\text{O}$  values and by the Nd, Sr, Pb, H, S, Cl, and noble gas isotopic compositions of  
84 the FPP rocks (Hunziker and Zingg, 1982; Voshage et al., 1987, 1988; Cumming et al., 1987;  
85 Hartmann and Wedepohl, 1993; Obermiller, 1994; Seitz and Woodland, 2000; Downes, 2001;  
86 Matsumoto et al., 2005; Selverstone and Sharp, 2011). The melt migration processes have been  
87 mainly attributed to supra-subduction environments (e.g. Zanetti et al., 1999; Grieco et al., 2001,  
88 2004; Morishita et al., 2003, 2008; Matsumoto et al., 2005), but also to extensional settings (Garuti  
89 et al., 2001; Zaccarini et al., 2004).

90 The available geochronological data indicate that this main event of melt migration took place in  
91 the Palaeozoic (Zanetti et al., 2016; Malitch et al., 2017), but many field, petrochemical, and  
92 geochronological data point to the development of several other tectono-magmatic events in the  
93 Mesozoic as well (Stähle et al., 1990, 2001; Grieco et al., 2001; Matsumoto et al., 2005; Morishita  
94 et al., 2003, 2008; Zanetti et al., 1999; 2016; Malitch et al., 2017).

95 Giovanardi et al. (2013) studied the occurrence of a late sapphirine-bearing gabbroic dykelet swarm  
96 previously reported by Siena & Coltorti (1989) and discovered by M. Mazzucchelli (Giovanardi et  
97 al., 2013). Based on a detailed petrographic and geochemical survey, they provided evidence for a  
98 multi-stage dyke formation, which involved fractional crystallisation associated with different  
99 patterns of melt–rock interactions.

100 With the aim of better constraining the nature of the primitive melt and petrological processes and  
101 inducing the segregation of sapphirine in the dykes, in this paper, new and detailed petrochemical  
102 data are reported for dykes and host peridotites close to the area studied by Giovanardi et al. (2013)  
103 and on a second finding in an adjacent area. Particular care has been dedicated to characterising the  
104 compositional variability in terms of major and trace element mineral chemistry, as well as the

105 mineral O, Nd, and Sr isotopic composition, to document the geochemical fractionation of the  
106 flowing melt following reaction with the host peridotite and early cumulates. The genetic and  
107 temporal relationships between the dyke emplacement and other events of melt migration recorded  
108 by the FPP unit, as well as the related geodynamic scenarios, are also addressed.

109

## 110 **2. Geological setting**

111 In the Finero area (Fig. 1), the IVZ crops out in a pseudo-antiform structure with the FPP mantle  
112 unit at the core, flanked by what has been called the Finero Mafic Complex (Cawthorn, 1975; Siena  
113 & Coltorti, 1989; Lu et al., 1997a, b; Zanetti et al., 2013, 2014; Giovanardi et al., 2014;  
114 Mazzucchelli et al., 2014; Langone et al., 2017). The Mafic Complex is divided into three different  
115 units: the stratigraphically lower Layered Internal Zone unit (LIZ), in contact with the FPP; the  
116 Amphibole-Peridotite (Amphibole-Pd); and the External Gabbro (EG). The EG is placed in tectonic  
117 contact with the Amphibole-Peridotite by a Mesozoic high-T shear-zone (Langone et al., 2018). A  
118 tectonic contact also characterises the transition toward the amphibolite-facies metasediments and  
119 metabasites of the Kinzigite Formation (KF), which represent the metamorphic basement of the  
120 Adria Plate. Septa of KF rocks are embedded in the EG.

121 The Finero mafic-ultramafic sequence presents several differences with respect to the southern and  
122 central sectors of the IVZ (i.e. the Baldissero and Balmuccia peridotites and the Val Sesia Complex;  
123 Quick et al., 1995; Correia et al., 2012; Mazzucchelli et al., 2014 and references therein), providing  
124 evidence of a different geological evolution (Zanetti et al., 2013, 2014, 2016; Langone et al., 2017,  
125 2018). In particular, the FPP has been completely recrystallised by several events of migrating melts  
126 with crustal affinity (Zanetti et al., 1999; 2016; Mazzucchelli et al., 2014 and references therein),  
127 and the relationships with the parental melts of the surrounding Mafic Complex are far from being  
128 understood (Giovanardi et al., 2014). Conversely, the peridotites cropping out in the central and  
129 southern part of IVZ did not undergo similar melt-induced recrystallisation and were emplaced in

130 the Kinzigite Formation as tectonic slivers before the intrusion of the central and southern Mafic  
131 Complex (Quick et al., 1995; Mazzucchelli et al., 2014 and references therein).

132 The FPP is mainly represented by phlogopite-bearing amphibole harzburgites and associated  
133 phlogopite-bearing pyroxenites (Cawthorn, 1975; Siena & Coltorti, 1989; Hartmann & Wedephol,  
134 1993; Zanetti et al., 1999; Raffone et al., 2006; Selverstone & Sharp, 2011; Mazzucchelli et al.,  
135 2014; Giovanardi et al., 2018). These lithologies derived from pervasive metasomatism of a  
136 depleted peridotite, which formed secondary orthopyroxene, amphibole, and phlogopite (Zanetti et  
137 al., 1999; Tommasi et al., 2017). Channelled migration events formed dunite bodies containing  
138 stratiform to podiform chromitites and, rarely, pyroxenite and hornblendite layers (Grieco et al.,  
139 2001, 2004; Zanetti et al., 2016).

140 Late stages of porous-flow melt migrations crystallised apatite- and dolomite-bearing wehrlites and  
141 apatite-rich, orthopyroxene-bearing peridotites, which sometimes contain carbonate-bearing  
142 domains showing marked modal and geochemical gradients with the host rocks (Zanetti et al., 1999;  
143 Morishita et al., 2003, 2008; Matsumoto et al., 2005; Raffone et al., 2006). U–Pb analyses on  
144 apatite and isotopic noble gas data have yielded Triassic ages (Morishita et al., 2008).

145 Apatite and calcite also occur in dykes of nepheline-bearing syenites of the Triassic age, associated  
146 with hornblendites (Stähle et al., 1990, 2001). An Early Jurassic U–Pb zircon age was determined  
147 for an alkali pegmatite (Grieco et al., 2001).

148 The rocks of the FPP related to the main event and the dunite bodies show similar geochemical  
149 features and absence of geochemical gradients. The harzburgite–pyroxenite association and the  
150 chromitite and pyroxenite layers in dunite bodies are depleted in Nb and high field strength  
151 elements and significantly enriched in K, Rb, Ba, Sr, and light rare earth elements (LREEs)  
152 (Hartmann & Wedephol, 1993; Zanetti et al., 1999, 2016; Mazzucchelli et al., 2014). The  
153 mineralogical and compositional features have been considered by several authors as evidence of  
154 the role of a slab-derived crustal component in the percolating melts (Hartmann & Wedephol, 1993;  
155 Mazzucchelli et al., 1995, 2016; Rivalenti et al., 1995, 2007a; Zanetti et al., 1999, 2016; Grieco et

156 al., 2001, 2004; Morishita et al., 2003, 2008; Ponce et al., 2015; and others). Thus far, a variety of  
157 different geochemical components has been identified based on isotopic data. In particular, the Hf  
158 (in zircon) and O (in zircon and pyroxene) isotopic compositions of chromitites suggest the  
159 presence of large volumes of continental crust in the migrating melts (Zanetti et al., 2016; Malitch  
160 et al., 2017), whereas the hydrogen, oxygen, and chlorine isotopic compositions of amphibole and  
161 phlogopite from the harzburgite-pyroxenite association show variability ( $\delta D$  from  $-29$  to  $-86\%$ ,  
162  $\delta^{18}O$  from  $4.9$  to  $6.1\%$ , and  $\delta^{37}Cl$  from  $-2.0$  to  $+2.1\%$ ) consistent with mixtures of magmatic fluids  
163 contaminated by seawater (Hartmann & Wedepohl, 1993; Selverstone & Sharp, 2011).  
164 Giovanardi et al. (2013) reported the occurrence in the FPP of late sapphirine-bearing gabbroic  
165 dykes. These dykes crosscut at high angles the pervasive mantle foliation and other lithologies,  
166 showing different mineralogical and major element mineral chemistry features with respect to other  
167 FPP rocks.

168

### 169 **3. Samples and petrography**

170 Two different dykes and their host peridotites from the FPP unit were investigated. The first dike  
171 (sample FI09C06, Fig. 2) was collected along the road that connects National Road 631 to a  
172 peridotite quarry located on the right flank of the Rio Creves valley (less than 100 m from the  
173 outcrops studied by Zanetti et al., 1999). The second dike (sample FI9664) is from a boulder along  
174 the Rio Creves, about 30 m upstream of its intersection with the Rio Cannobino. The host peridotite  
175 was collected 8 cm from the FI09C06 dyke, and close to the contact with the dyke represented by  
176 sample FI9664.

177 The gabbroic dykes are centimetric in thickness (mostly 2–5 cm; Fig. 2). They show variable  
178 strikes, usually crosscutting the harzburgite-pyroxenite association at high angles (Fig. 2).

179 A Mesozoic age for these sapphirine-bearing dykes is constrained by the observation that they also  
180 crosscut the foliation of protomylonites in the external domains of Mesozoic shear zones (Matysiak  
181 & Trepmann, 2015 and references therein), but are themselves deformed in a few cm-wide



182 mylonitic to ultramylonitic bands parallel to the protomylonite foliation (Tommasi et al., 2017).  
183 Such shear zones were active at different crustal levels over a very long, Triassic–Jurassic time  
184 interval (235–180 Ma; Langone et al., 2018 and references therein). This finding is also consistent  
185 with the observation that all the late intrusive bodies, which are discordantly intruding the pervasive  
186 mantle foliation by hydraulic fracturing, do not show ages older than Triassic (Stähle et al., 1990,  
187 2001; Grieco et al., 2001; Matsumoto et al., 2005).

188 The gabbroic dykes contain sapphirine (Giovanardi et al., 2013) and show a banded symmetric  
189 structure consisting of melanocratic zones at the peridotite contacts and a leucocratic zone  
190 representing the dyke core (Fig. 2). The melanocratic zones, in turn, can be divided in three  
191 different bands.

192 From the host peridotite to the dyke core, the following layers can be recognised:

193 1) An orthopyroxenite zone (hereafter, the Orthopyroxene Zone), established within the ambient  
194 peridotite. Orthopyroxene locally shows recrystallised rims with the growth of fine neoblasts of  
195 orthopyroxene and rarely olivine. The orthopyroxene sometimes presents exsolution lamellae.  
196 Locally, black spinel and phlogopite occur as accessory phases. Phlogopite is concentrated in  
197 interstitial positions but rarely fills fractures within orthopyroxene crystals (Fig. 2). Giovanardi et  
198 al. (2013) reported the occurrence of a sapphirine- and amphibole-bearing recrystallisation front in  
199 the Orthopyroxene Zone.

200 2) A first melanocratic zone (hereafter, the Early Amphibole Zone) inside veins formed by dark-  
201 brown amphibole (up to 1 cm long, named 'early amphibole') and associated small plagioclase  
202 grains, spinel, phlogopite, and apatite; in these zones, the magmatic texture is preserved, as  
203 evidenced by amphibole twinning. Apatite and spinel mainly occur as rounded inclusions in  
204 amphibole, whereas phlogopite is in interstitial position.

205 3) A second melanocratic zone (hereafter the Late Amphibole Zone), consisting of light-brown to  
206 green amphibole (named 'late amphibole'), green spinel, sapphirine, and phlogopite. Sapphirine  
207 occurs in three textural positions: inclusions within late amphibole, coronas rimming spinel, and

208 isolated/aggregated crystals in interstitial positions (Fig. 2). Phlogopite is an accessory phase in  
209 interstitial positions. Late amphibole is smaller than early amphibole and is euhedral to anhedral in  
210 shape. Intermediate green–brown amphibole, often associated with spinel and sapphirine, is  
211 recognised near and through the Early Amphibole Zone–Late Amphibole Zone contact.  
212 Recrystallisation zones with a fine-grained texture occur. The Late Amphibole Zone is not  
213 continuous through the dykes. It forms patches, which can also occur only on one side of the  
214 leucocratic core of the dyke, or that can extend up to the Orthopyroxene Zone. The Late Amphibole  
215 Zone is more developed in sample FI9664 (up to about 1.5 cm in thickness) than in sample  
216 FI09C06 (up to 1 cm in thickness). In addition, sapphirine crystals in sample FI9664 can reach up to  
217 1.5 mm in size, whereas in sample FI09C06, they are commonly < 0.2 mm.

218 4) A leucocratic zone (Leucocratic Zone, hereafter) formed by plagioclase and subordinate  
219 amphibole (both magmatic and relict from melanocratic zones). Apatite occurs as an accessory  
220 phase, sometimes included in plagioclase or in relicts of brown early amphibole. The apatite rarely  
221 contains calcite inclusions. Plagioclase shows twinning (mainly pericline and subordinately crossed  
222 albite and pericline), which is often partially or totally erased by recrystallisation induced by late  
223 deformation. Magmatic amphibole has greenish pleochroism, whereas relict amphibole, ripped from  
224 the melanocratic zones, is brown. Amphibole often forms single-crystal alignments parallel to the  
225 dyke strike, like in a flow texture. Phlogopite is rare and is associated with amphibole.  
226 Recrystallisation zones show fine-grained equigranular texture. Rarely, Fe-oxides, Fe–Ni sulphides,  
227 and pyrite occur.

228 The host peridotite away from the contact is a hornblende-harzburgite in modal composition with a  
229 porphyroclastic texture (Giovanardi et al., 2018). It is characterised by the presence of olivine and  
230 orthopyroxene porphyroclasts, with a secondary, undeformed mineral assemblage dominated by  
231 amphibole (15% by volume), in association with orthopyroxene, spinel, phlogopite, and  
232 clinopyroxene, strictly similar to the dominant peridotite type described by Zanetti et al. (1999) and  
233 Tommasi et al. (2017). A detailed petrographic inspection highlights the occurrence of a reacted

234 peridotite zone approaching the orthopyroxene zone. It is characterised by the presence of a  
235 secondary mineral assemblage, modally dominated by long (up to 5 mm) phlogopite lamellae, with  
236 associated subordinate amounts of undeformed orthopyroxene, spinel, amphibole, and  
237 clinopyroxene. In the reacted zone, olivine was stable; conversely, the modal orthopyroxene content  
238 is slightly lower than that in the peridotite far from the vein. Primary (e.g. olivine) and secondary  
239 minerals in the reacted zones display elongation sub-parallel to the present-day vein strike.

240

#### 241 **4. Analytical methods**

242 Sample FI9664, representing the gabbroic dyke and the contact host harzburgite, was analysed for  
243 whole-rock major and trace elements. Whole-rock major elements and Sc, V, Cr, Co, Ni, Cu, and  
244 Zn were analysed by X-ray fluorescence spectrometry (XRF), whereas Li, Rb, Sr, Y, Zr, Nb, Ba,  
245 rare earth element (REE), Hf, Ta, and Pb contents were analysed by inductively coupled plasma-  
246 mass spectrometry (ICP-MS). Analyses were performed following the methods described by  
247 Mazzucchelli et al. (2010) (data reported in Supplementary Material A).

248 Samples FI9664 and FI09C06 were analysed for mineral major elements with the electron  
249 microprobe JEOL 8200 Super Probe housed at the University of Milano. Analyses were performed  
250 following methods described by Ponce et al. (2015). Data are reported in Supplementary Material  
251 A.

252 Mineral trace elemental contents were determined (as in Rivalenti et al., 2007b) with a LA-ICP-MS  
253 housed at Istituto di Geoscienze e Georisorse, Consiglio Nazionale delle Ricerche, Sede Secondaria  
254 di Pavia (data are reported in Supplementary Material A) using a Perkin Elmer SCIEX DCR-e  
255 coupled with a solid-state laser source (Q-switched Nd:YAG, Quantel Brilliant). Data reduction was  
256 performed using the GLITTER software (Griffin et al., 2008). NIST SRM 610 was used as an  
257 external standard. Ca was used as an internal standard for clinopyroxene, amphibole, and  
258 plagioclase, Si for olivine, orthopyroxene, and phlogopite, and Mg for spinel. Precision and  
259 accuracy were assessed by repeated analysis of the standard BCR-2g and were found to be better

260 than 10% at the ppm concentration level. Further information is reported by Giovanardi et al.  
261 (2017).

262 Sample FI09C06 was selected for O, Nd, and Sr isotopic analyses in mineral separates, after the  
263 combination of preliminary electron microprobe and LA-ICP-MS analyses suggested a more  
264 primitive nature for its parental melt.

265 O isotopes of pure mineral separates from sample FI09C06 were analysed at the I.G.G.-C.N.R.,  
266 Pisa, by conventional laser fluorination with a Finnigan Delta Plus mass spectrometer. For each  
267 phase, 1–1.5 mg aliquots were necessary to measure the oxygen isotope composition. Analyses  
268 were performed following the methods described by Perinelli et al. (2011). Data are reported in  
269 Table 1.

270 Sr and Nd isotopic ratios of amphibole and plagioclase separates from sample FI09C06 and the host  
271 harzburgite were analysed at the Marine Environmental Sciences Laboratory (LEMAR) of the  
272 Institut Universitaire Européen de la Mer (IUEM), Université de Bretagne Occidentale. Analyses  
273 were carried out after dissolution and chromatographic separation using a TRITON Thermo-  
274 Ionisation Mass Spectrometer (TIMS), following the procedure described by Janin et al. (2012).  
275 Analyses were corrected for the reference material NBS987 ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.710241 \pm 0.000019$ ,  $n =$   
276  $6$ ) for Sr and the La Jolla standard ( $^{143}\text{Nd}/^{144}\text{Nd} = 0.511847 \pm 0.000009$ ,  $n = 3$ ) for Nd. Data are  
277 reported in Table 2.

278

## 279 **5. Bulk rock chemistry**

280 The gabbroic dyke FI9664 presents higher  $\text{TiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{CaO}$ ,  $\text{Na}_2\text{O}$ , and  $\text{K}_2\text{O}$  contents and lower  
281  $\text{FeO}$  and  $\text{MgO}$  contents than the host harzburgite collected at the contact.

282 The gabbroic dyke has trace element abundances that are one order of magnitude higher than those  
283 of the host harzburgite; only Rb and Li contents are similar in both rocks (Fig. 3; Supplementary  
284 Material A). The two rock types show some similarities in element fractionation: marked  
285 enrichments in LREEs with respect to middle-heavy rare earth elements (M-HREE;  $(\text{La}/\text{Yb})_N$

286 values are 5.5 and 16–30 for the dyke and host harzburgite, respectively; primitive mantle data from  
287 McDonough and Sun, 1995);  $(\text{Th}/\text{U})_{\text{N}}$  and  $(\text{Zr}/\text{Hf})_{\text{N}}$  always  $< 1$ ; and positive Ba and Pb anomalies.  
288 Conversely, the  $(\text{Nb}/\text{Ta})_{\text{N}}$  values are 1.6 for the gabbroic dyke and 0.7 for the host harzburgite.  
289 Linearly fractionated LREE-enriched patterns were also shown by the nepheline-bearing alkaline  
290 dykes described by Stähle et al. (2001). However, the latter had significantly higher trace element  
291 contents than those of gabbroic dyke FI9664. The dykes reported by Stähle et al. (2001) also  
292 displayed significant peculiarities in terms of fractionation of highly incompatible trace elements,  
293 such as U, Th, Nb, Ta, Ba, and Pb, with respect to REEs.  
294 The spidergram of the peridotite close to the contact with dike FI9664 shows some relevant  
295 differences with respect to those reported by Hartmann & Wedephol (1993): large positive Pb and  
296 Hf anomalies, greater Rb, U, Ba, Ta, and Nb contents, and slightly lower LREE contents (Fig. 3).

297

## 298 **6. Major element mineral chemistry**

### 299 *6.1. Host harzburgites*

300 The mineral composition of the host harzburgite far from the veins is similar to that of the  
301 harzburgite-pyroxenite association reported by Zanetti et al. (1999). The amphibole is pargasitic in  
302 composition, with only one analysis of Mg-hornblende composition. The mica is phlogopite, and  
303 the clinopyroxene is diopside.

304 Minor differences include a slightly Fe-enriched composition shown by olivine ( $\text{Fo} = 100 \times$   
305  $\text{Mg}/(\text{Mg} + \text{Fe}^{2+}_{\text{tot}})$ , molar ratio of 90.4–91.1), orthopyroxene, spinel, and amphibole (Fig. 4).  
306 Orthopyroxene and spinel (hercynite to spinel in composition) are also richer in Al. Because of their  
307 low Cr#, the FI09C06 oxides straddle the hercynite–spinel boundary, whereas oxides in the FPP are  
308 chromite. Lower Cr content is shown by amphibole, which is also characterised by lower Na and  
309 higher K contents.

310 At the contact with the vein, the Mg# does not change significantly in the minerals. Aluminium  
311 decreases in orthopyroxene, spinel, amphibole, and phlogopite, whereas Ti content is higher in both

312 amphibole and phlogopite. In amphibole, Na decreases, balanced by higher K content (Fig. 5).  
313 Conversely, the Na/K ratio is very variable in phlogopite.  
314 Orthopyroxene from the Orthopyroxene Zone has lower Mg# and CaO content and higher Al<sub>2</sub>O<sub>3</sub>  
315 content (Fig. 4). Similarly, its composition is different from orthopyroxene from the sapphirine-  
316 bearing rock found in the LIZ northern unit of the Finero Mafic Complex by Sills et al. (1983). The  
317 latter has lower Mg# and higher Al<sub>2</sub>O<sub>3</sub> content (Fig. 4).

318

## 319 6.2. Gabbroic dykes

320 Notwithstanding the similar internal banding, the two studied dykes show marked differences in  
321 terms of major elemental mineral chemistry. Significant compositional changes are also shown for  
322 the different types of amphibole (i.e. the Early and Late Amphibole Zones and the Leucocratic  
323 Zone; Fig. 6). The unique common feature of the amphiboles of both gabbroic dykes is the higher  
324 Al content exhibited by the late amphibole (Giovanardi et al., 2013). Amphibole from gabbroic  
325 dykes is mostly pargasitic in composition (unit formula calculated according to Ridolfi et al., 2018),  
326 but sometimes Al substitution for Si in late amphibole is higher than 2 a.p.f.u., thereby entering the  
327 sadaganaite compositional field.

328 In particular, amphibole from sample FI09C06 has distinctly lower Mg# (0.73–0.82) than sample  
329 FI9664 (0.85–0.87). In sample FI09C06, Mg# is lower in early amphibole than in late and  
330 leucocratic amphibole, whereas it is exactly the reverse in sample FI9664 (Fig. 6).

331 FI09C06 amphibole also displays the lowest CaO and Na<sub>2</sub>O and the highest TiO<sub>2</sub> and K<sub>2</sub>O contents  
332 (TiO<sub>2</sub> 0.27–2.19 wt.% and K<sub>2</sub>O 0.31–1.28 wt.% respectively) (Fig. 6).

333 In sample FI09C06, TiO<sub>2</sub> linearly increases with the decrease in Mg#. Instead, TiO<sub>2</sub> is higher in the  
334 FI9664 early amphibole than in the leucocratic amphibole, but distinctly higher TiO<sub>2</sub> contents are  
335 shown by late amphibole. Moreover, in sample FI9664 early amphibole, the increase in Mg# is  
336 associated with increasing Cr and Ca, and with decrease in Na and Al contents.

337 As a whole, amphibole from the gabbroic dykes has higher  $\text{Al}_2\text{O}_3$  content and lower Mg# and  $\text{Cr}_2\text{O}_3$   
338 content with respect to others FPP lithologies, resulting in similarity to the amphibole from the  
339 sapphirine-bearing rock in the LIZ (Sills et al., 1983) (Fig. 6). Amphibole from the two gabbroic-  
340 dykes samples shows lower CaO and  $\text{K}_2\text{O}$  and higher  $\text{Na}_2\text{O}$  contents than the respective host  
341 amphibole (Fig. 6).

342 Phlogopite from sample FI9664 presents a narrow range of major elemental contents with respect to  
343 sample FI09C06. In sample FI9664, phlogopite has higher Mg# values than phlogopite from sample  
344 FI09C06 (0.92–0.94 and 0.70–0.90, respectively) and commonly lower  $\text{TiO}_2$  contents (Fig. 5).  
345 Phlogopite from the gabbroic dykes shows higher  $\text{Al}_2\text{O}_3$  content and lower Mg# than the phlogopite  
346 from the host peridotite and the harzburgite-pyroxenite association (Fig. 5).

347 Spinel is mainly found in the Late Amphibole Zone. Unlike the spinel from the host harzburgite, the  
348 harzburgite-pyroxenite association reported by Zanetti et al. (1999), and the chromitite layers in  
349 dunite bodies (Grieco et al., 2001; 2004; Zanetti et al., 2016), the spinel from the gabbroic dykes  
350 does not contain  $\text{Cr}_2\text{O}_3$  and can be classified as spinel (Mg# 0.46–0.75).

351 Plagioclase from sample FI9664 is commonly more anorthitic than the plagioclase from sample  
352 FI09C06 (An contents of 82–93 and 36–87, respectively). In both samples, some reversely zoned  
353 plagioclases are recognised.

354 The sapphirine composition falls near the 7:9:3 composition on the  $\text{SiO}_2$ –( $\text{FeO} + \text{MgO}$ )–( $\text{Al}_2\text{O}_3 +$   
355  $\text{Cr}_2\text{O}_3 + \text{Fe}_2\text{O}_3$ ) diagram (Fig. 7). Sapphirine from sample FI9664 is higher in  $\text{Al}_2\text{O}_3$  content with  
356 respect to sample FI09C06, whereas it is lower in  $\text{SiO}_2$  content.

357 Scanning electron microscope (SEM) investigations at the Centro Interdipartimentale Grandi  
358 Strumenti laboratories of the Università di Modena e Reggio Emilia, performed with an  
359 Environmental SEM Quanta-200 (Fei Company-Oxford Instruments), suggest that apatites from  
360 different zones of the dykes are Cl-apatites.

361

## 362 **7. Trace elemental compositions**

363 *7.1. Host harzburgite*

364 Amphibole and clinopyroxene from harzburgite FI09C06, 8 cm away from the contact with the  
365 gabbroic dyke, are characterised by LREE-enriched linearly fractionated patterns (Figs. 8, 9 and  
366 10). Their compositions are similar to those of the FPP harzburgite-pyroxenite association (Zanetti  
367 et al., 1999). Proceeding toward the contact (i.e. sample FI9664), amphibole and clinopyroxene  
368 show upward-convex REE patterns characterised by high variability in absolute contents (Figs. 8, 9  
369 and 10). These patterns are similar to those from apatite-rich domains reported by Zanetti et al.  
370 (1999) (Figs. 9 and 10).

371

372 *7.2. Gabbroic dykes*

373 The two gabbroic dykes show different trace element compositions. In sample FI09C06,  
374 amphiboles from the Early Amphibole Zone and Leucocratic Zone have L-MREE-enriched  
375 upward-convex patterns, similar to those of harzburgite near the contact (Fig. 8). In the Late  
376 Amphibole Zone, some crystals show more fractionated patterns characterised by lower M-HREE  
377 contents and positive Eu anomalies (Fig. 8). These variations are associated with marked depletion  
378 in Ta, Zr, Hf, Y, Sc and V (Fig. 10, V is not shown). As a whole, amphiboles from different zones  
379 of sample FI9664 show more fractionated REE patterns characterised by enrichment in LREEs and  
380 depletion in HREEs (Fig. 8) with a nearly flat pattern in the LREE region. Amphibole from the  
381 Early Amphibole Zone of sample FI9664 presents a small positive Eu anomaly ( $(Eu/Eu^*)_N$  within  
382 1.04–1.65), which becomes more evident in the amphibole from the Late Amphibole Zone  
383 ( $(Eu/Eu^*)_N$  within 1.29–1.62). Amphibole from sample FI9664 is also more enriched in Th, U, and  
384 Pb (and to lesser extent in Sr) compared with amphibole from sample FI09C06 (Fig. 10).

385 Similar to the amphibole, REE patterns of plagioclase from sample FI9664 are more fractionated  
386 than those of plagioclase from sample FI09C06, (Fig. 8). In both samples, the trace elemental  
387 abundances of the plagioclase cores are greater than those of the plagioclase rims.



388 Apatite displays the typical LREE enrichment and high Th, U and Pb contents (12.31–17.35 ppm,  
389 4.35–5.92 ppm, and 2.18–4.19 ppm, respectively). Nb, Ta, Zr, Hf, Ti, and Sc show negative  
390 anomalies with values that are often below the detection limit.

391 No systematic trace elemental variations were found in phlogopite.

392

### 393 **8. O isotopes**

394 The  $\delta^{18}\text{O}$  in silicates of sample FI09C06 shows a steady increase from the contact (Orthopyroxene  
395 Zone) to the leucocratic gabbro in the vein core, through the Early and Late Amphibole Zones (Fig.  
396 8, 10, and 11 and Table 1). In particular, it varies from 5.81‰ (standard deviation (std. dev.) 0.11)  
397 in orthopyroxene from the Orthopyroxene Zone to 6.9‰ (std. dev. 0.05) in amphibole from the  
398 Early and Late Amphibole Zones, and to 8.60‰ (std. dev. 0.01) in the plagioclase of the gabbroic  
399 core.

400 The  $\delta^{18}\text{O}$  values of orthopyroxene from the Orthopyroxene Zone lie within the mantle range. They  
401 are higher than the  $\delta^{18}\text{O}$  value reported by Hartmann & Wedephol (1993) for clinopyroxene from  
402 the phlogopite-bearing amphibole harzburgite (Fig. 11), and for orthopyroxene from olivine-  
403 chromitites of the FPP, but significantly lower than the orthopyroxene from orthopyroxene-  
404 chromitites (Zanetti et al., 2016).

405 Vein amphibole and plagioclase have  $\delta^{18}\text{O}$  values significantly higher than the mantle range (Fig.  
406 11). They are close to the highest values found in amphibole, phlogopite, orthopyroxene,  
407 clinopyroxene, and zircon from the phlogopite-bearing amphibole harzburgite-pyroxenites  
408 (Hartmann & Wedephol, 1993; Selverstone and Sharp, 2011) and chromitite layers of the FPP  
409 (Zanetti et al., 2016).

410 The  $\delta^{18}\text{O}$  values obtained for the green spinel associated with sapphirine are markedly lower than  
411 those of the associated late amphibole ( $\delta^{18}\text{O} = 4.38\text{‰}$ , std. dev. 0.10), as expected based on crystal-  
412 chemical constraints (see Bindeman, 2008 and references therein).

413

## 414 **9. Sr and Nd isotopes**

415 The trace elemental concentrations of the mineral separates of early and late amphibole match the  
416 differences highlighted by LA-ICP-MS analysis of thin sections. The Sr and Nd isotopic  
417 compositions of early amphibole and late amphibole from the FI09C06 dyke are coincident within  
418 analytical uncertainty (Table 2). Plagioclase from the FI09C06 Leucocratic Zone shows the same Sr  
419 isotopic composition of the amphiboles, and only a slightly more radiogenic  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio.  
420 Similar Sr isotopic compositions were documented in amphibole and apatite of discordant veins  
421 from the FPP, which were characterised by higher  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios (Morishita et al., 2008). The  
422 isotopic composition of the dyke minerals is enriched with respect to the depleted mantle reference  
423 value and the mid-oceanic ridge basalt (MORB) field; these minerals fall within the ocean island  
424 basalt (OIB) field (Fig. 12). In particular, they lie between the isotopic compositions of FPP  
425 hornblende-syenite dykes (bulk rock from Stähle et al., 1990, 2001) and peridotites (amphibole and  
426 clinopyroxene from Obermiller, 1994). Amphiboles separated from the host harzburgite (collected  
427 far from the contact) show the most radiogenic Sr and unradiogenic Nd values, but have never  
428 before been documented in the literature for FPP rocks (Fig. 12).

429

## 430 **10. Discussion**

431

### 432 *10.1. Petrogenesis of the sapphirine-bearing gabbroic dykes*

433 The two veins studied here share exactly the same internal banding as the sample described by  
434 Giovanardi et al. (2013). This suggests that all three veins were part of an interconnected swarm,  
435 which allowed them to record the same sequence of petrologic events. The symmetric structure of  
436 the dykes suggests that the intrusion occurred through several events of melt percolation. In the  
437 following section, we discuss, in detail, the different stages and processes that produced the  
438 gabbroic dyke swarm.

439

440 *10.1.1. Melt percolation through the peridotite*

441 According to the process governing the emplacement of other dyke swarms in mantle sectors (e.g.  
442 Mazzucchelli et al., 2010 and references therein), the melts flowing in the metasomatic haloes and  
443 in the conduit were originally similar, undergoing severe fractionation owing to assimilation and  
444 fractional crystallisation.

445 The characterisation of the wall peridotite has revealed some peculiarities. It is a common  
446 observation that metasomatic haloes of cm-to-m-scale surround late dykes and veins formed by  
447 melt segregation under mantle conditions. In most cases, the metasomatic haloes show marked  
448 geochemical and mineralogical gradients as a function of the distance from the dike contact (Zanetti  
449 et al., 1999; Ionov et al., 2002, Mazzucchelli et al., 2010; Borghini et al., 2016, 2017), which are  
450 interpreted as the result of porous flow percolation of melt escaping from the open conduits.  
451 Conversely, no progressive mineralogical or compositional variation is documented in the wall  
452 peridotite of sample FI9664 at a variable distance from the vein (see Supplementary Material B).  
453 Newly-formed clinopyroxene actually shows trace element heterogeneities, but these are randomly  
454 distributed. Conversely, the trace element composition of amphibole is very homogeneous,  
455 indicating late crystallisation from a unique melt.

456 The particular L/MREE-enriched convex-upward patterns shown by clinopyroxene and amphibole  
457 from the wall peridotite of sample FI9664 also demonstrate that their parental melt could not have  
458 been the same as that from which the adjacent early amphibole of dyke FI9664 crystallised (which  
459 show linearly-fractionated REE patterns). However, based on the amphibole trace element  
460 composition, the melt recorded by the wall peridotite of sample FI9664 must have been quite  
461 similar to the melt that produced the Early Amphibole Zone of dyke FI09C06.

462 The textural evidence that primary and secondary minerals from the metasomatic haloes display  
463 elongation sub-parallel to the present-day vein strike confirms that the development of melt  
464 migration channels was associated with some local deformation.

465 These observations can be reconciled by assuming that the metasomatic haloes were not produced  
466 by melt escaping normally from the conduit. They were likely established during an early stage of  
467 melt migration via focused porous flow along channels developed in correspondence with structural  
468 weakness, with a direction roughly parallel to the strike of the present-day veins.

469 The absence of resorption textures in olivine of the reacted wall peridotite indicates that this mineral  
470 was stable. Conversely, the modal content of orthopyroxene is slightly lower than in the peridotite  
471 far from the vein. This evidence leads us to consider that the orthopyroxene saturation of the melts,  
472 indicated by the segregation of the reactive Orthopyroxene Zone and the presence of few  
473 orthopyroxenes in the veins (Giovanardi et al., 2013), was not a clear, primary characteristic, but  
474 possibly a consequence of an early stage of reactive melt migration characterised by olivine  
475 precipitation and orthopyroxene dissolution (see Piccardo et al., 2017).

476 The mineralogical mismatch between the nearly monomineralic, hydrous-mineral-free  
477 Orthopyroxene Zone and the Early Amphibole Zone suggests that they were segregated by  
478 compositionally different melts. This hypothesis is confirmed by large differences in terms of O  
479 isotopic compositions recorded by the orthopyroxene (5.8‰ standard mean ocean water (SMOW))  
480 from the Orthopyroxene Zone and the amphibole from the Early Amphibole Zone (6.9‰ SMOW)  
481 within the vein. The lighter O isotopic composition of the Orthopyroxene Zone would support a  
482 hybrid composition of its parental melt caused by reactive porous flow through the FPP, with  
483 buffering of O isotopic composition at mantle values.

484 This conclusion led us to consider the possibility that the Orthopyroxene Zone formed before the  
485 segregation of the Amphibole Zones, possibly at the centre of the migration channels, as observed  
486 elsewhere in the FPP. It was successively split into two parts by the opening of the fracture.

487

#### 488 *10.1.2. Melt segregation in the open conduit*

489 The Early Amphibole Zone was segregated when melts started flowing in open fractures. It is likely  
490 that amphibole crystallisation was triggered by the presence of an ultramafic wall, suggesting that

491 plagioclase was unstable in contact with the orthopyroxenite layers. The development of  
492 amphibole-rich selvages where hydrous gabbroic rocks come in contact with ultramafic layers (e.g.  
493 in the LIZ of the Finero Complex; Mazzucchelli et al., 2014) has previously been documented in the  
494 literature.

495 The two gabbroic dykes of this study show significant differences in terms of major and trace  
496 element mineral compositions. In particular, amphibole from the FI9664 vein shows an unusual  
497 relative enrichment in highly incompatible trace elements (U, Th, LREEs, and Na) and compatible  
498 elements (Mg, Ca, and Cr) with respect to those from the FI09C06 vein. Consistently, phlogopite is  
499 more enriched in Mg and Na, and plagioclase is more anorthitic. Assuming that the dyke swarm  
500 was produced by the injection of a unique melt, this relationship cannot be reconciled based on a  
501 melt evolution governed only by fractional crystallisation. Mazzucchelli et al. (2010) documented  
502 the same correlation between compatible and incompatible elements in minerals from cm-thick  
503 diorite dykes, demonstrating that such geochemical features can be modelled considering an  
504 assimilation of host minerals concomitant with fractional crystallisation (i.e. assimilation and  
505 fractional crystallisation (AFC) processes). Accordingly, we argue that the parental melt of vein  
506 minerals in sample FI9664 can be considered more evolved through AFC processes than that of  
507 dyke FI09C06. The concomitant enrichment of Na and Ca in the vein minerals of sample FI9664  
508 evidences the role of amphibole in the assimilated component. This is also confirmed by the trace  
509 element patterns of FI9664 vein amphiboles, in which highly incompatible elements (such as  
510 LREEs, Th, and U) increase and moderately incompatible element (such as HREEs, Ti, and Y)  
511 decrease, moving toward the fractionation shown by those of host harzburgite away from the  
512 contact. The decrease of moderately incompatible elements with the progression of the melt  
513 evolution may be indicative of  $D^{\text{Amphibole/L}}$  higher than 1 (Tiepolo et al., 2007). It is noteworthy that  
514 the Al content in sample FI9664 early amphibole is comparable with to slightly lower than in those  
515 from the FI09C06 vein, suggesting that the assimilation of host minerals was not effective in  
516 boosting the Al concentration of the evolving melt.

517

518 *10.1.3. Formation of sapphirine-bearing patches*

519 The mm-to-cm-thick sapphirine-bearing patches and stripes are randomly distributed from the  
520 internal end of the hornblendite seam to the Orthopyroxene Zone, although they are generally  
521 concentrated towards the centres of the veins. They are apparently the results of interstitial  
522 migration of a melt in chemical disequilibrium with the early cumulus minerals, as indicated by the  
523 recrystallisation fronts inside large early amphiboles. Both interstitial and recrystallised patches  
524 show that the injected melt was saturated in late amphibole, sapphirine, and spinel, whereas the  
525 saturation in phlogopite is uncertain, because it occurs only interstitially and the textural  
526 relationships are equivocal.

527 The following series of petrographic and geochemical features suggest that the sapphirine-bearing  
528 mineral assemblages were not simply related to interaction with the parental melt of the Leucocratic  
529 Zone, but were segregated because of the injection in the dyke swarm of an additional melt  
530 component coming from outside the system:

- 531 1) Early amphibole at the contact with or embedded by the plagioclase of the Leucocratic Zone  
532 does not show any evidence of comparable reaction.
- 533 2) Notwithstanding the significantly different major elemental composition of early amphibole  
534 documented in the two veins of this study, the major elemental chemistry of sapphirine-  
535 associated late amphibole converges toward specific Al and Mg# values. Phlogopite shows  
536 consistent variation, even though its attribution to early or late amphibole assemblages is  
537 doubtful because of its interstitial position. The data reported by Giovanardi et al. (2013)  
538 confirm this trend. This finding indicates that a unique component with a specific  
539 composition was responsible for the sapphirine segregation in the three sectors of the dyke  
540 swarm.
- 541 3) Normalised patterns of late amphibole from sapphirine-bearing areas show particular  
542 features: stronger LREE/HREE fractionation, depletion in M-HREEs, Zr, Hf, Nb, Ta, Ti, Sc,

543 and V, and the appearance of a positive Eu anomaly. The results also show inversion of the  
544  $(Zr/Hf)_N$  values, which are  $> 1$  in the early amphibole, but  $< 1$  in the late amphibole. The  
545 overall fractionation mimics that normally exhibited by plagioclase.

546 The latter evidence, consistent with the apparent increase of the Al content in the system, suggests  
547 that the injected melt was characterised by a high normative plagioclase content.

548 The O, Nd, and Sr isotopic compositions of early and late Amphibole are very similar, suggesting  
549 that the plagioclase component belonged to the same magmatic cycle that produced the dyke  
550 swarm.

551 However, petrographic surveying has not provided evidence of plagioclase assimilation and/or  
552 replacement by amphibole or another mineral, excluding plagioclase components derived by local  
553 assimilation. Thus, we suggest that the plagioclase signature was acquired by the injected melt  
554 through assimilation of cumulus plagioclase in hidden magmatic bodies and/or addition in the melt  
555 source, where it was likely present as a high-P metamorphic equivalent of pristine plagioclase.

556

#### 557 *10.1.4. Segregation of the Leucocratic Zone*

558 The petrographic evidence confirms that the formation of the Leucocratic Zone was a high energy  
559 event, which determined the partial disaggregation of the Early Amphibole Zone, with evident  
560 breaking of large early amphibole crystals. It can be locally recognised that two parts of a formerly  
561 unique crystal apparently lie on different sides of a vein. Giovanardi et al. (2013) suggested that this  
562 stage was accompanied by enlargement of the conduits.

563 The Nd and Sr isotopic compositions of the plagioclase indicate that the parental melt had a cognate  
564 origin with those of the hornblendite selvages. This inference is also confirmed by the trace element  
565 composition of the leucocratic amphibole, which is strictly similar to that of the associated early  
566 amphibole.

567 Nevertheless, the observation that the major elemental chemistry of the leucocratic amphibole is  
568 intermediate between that of early and late amphibole leads us to conclude that the parental melt of  
569 the sapphirine-bearing assemblages could be still present in diluted proportions.

570

#### 571 *10.2. Nature of the parental melts*

572 Evaluating the nature of the parental melt of magmatic segregates belonging to narrow, cm-thick  
573 veins/dykes intruding mantle peridotites at high-T-P mantle conditions must be done with particular  
574 caution, with an effort first to characterise the modifications imparted by fractional crystallisation  
575 and the reaction/assimilation processes with the host rock (e.g. Mazzucchelli et al., 2010 and  
576 references therein). As discussed in the previous section, sample FI09C06 shows that its parental  
577 melt underwent the lowest degree of interaction with the host peridotite among the known dykes,  
578 and thus is the best candidate to obtain constraints on the nature of the parental melts.

579 An important contribution in defining the geochemical components in the melts is provided by the  
580 marked zoning in terms of O isotopic composition documented in sample FI09C06 between the  
581 Orthopyroxene Zone, the hornblendite selvage, and the Leucocratic Zone.

582 The  $\delta^{18}\text{O}$  values of early and late amphibole (6.9‰ SMOW) well exceed the mantle range  
583 (Bindeman, 2008; Polat et al., 2018), suggesting a large contribution by crustal components in the  
584 parental melt of sample FI09C06 hornblendite. The proportion of the crustal component is even  
585 larger in the leucocratic band, in which a  $\delta^{18}\text{O}$  value of 7.8‰ SMOW can be calculated for the  
586 amphibole in hypothetical equilibrium with the plagioclase composition (8.6‰ SMOW). This is  
587 done by conservatively considering a low closure T for the FPP of ~850 °C (provided by pyroxene-  
588 solvus geothermometry), an average value of An70 for plagioclase, and the  $\delta^{18}\text{O}$  mineral  
589 fractionation values reported by Bindeman (2008).

590 Published data on FPP minerals show a pronounced heterogeneity of the O isotopic composition,  
591 with  $\delta^{18}\text{O}$  values from below to well above the mantle range (Hartmann and Wedephol, 1993;



592 Selverstone and Sharp, 2011; Zanetti et al., 2016). This heterogeneity may be due to multiple melt-  
593 migration events.

594 Values of  $\delta^{18}\text{O}$  approaching those of early and late amphibole have sometimes been documented in  
595 harzburgites and chromitites (Hartmann & Wedephol, 1993; Zanetti et al., 2016), but the values of  
596 the leucocratic layers are markedly out of range.

597 In addition, the  $\delta^{18}\text{O}$  value shown by orthopyroxene (5.81‰ SMOW) from the Orthopyroxene Zone  
598 provides a very important constraint, indicating how migrating melts, after prolonged interaction  
599 with the FPP, may have oxygen isotopic compositions buffered to mantle values. The decrease of  
600  $\delta^{18}\text{O}$  values in the minerals from the hornblendite selvages with respect to those from the dyke core  
601 (i.e. the Leucocratic Zone) confirms such a buffering effect.

602 We conclude that the high  $\delta^{18}\text{O}$  shown by dyke minerals cannot be the result of interaction between  
603 melts and the FPP but is a primary feature of the melts, which indicates the occurrence of large  
604 amounts of crustal components, either added to the melt in the source regions or at greater mantle  
605 depths.

606 A subduction-related component was identified based on noble gas isotopic composition in Late  
607 Triassic apatite-rich layers (Matsumoto et al., 2005, Morishita et al., 2003, 2008), overprinting the  
608 crustal affinity of the host FPP. Amphiboles from such apatite layers show the same radiogenic Sr  
609 isotopic composition as the minerals of sample FI09C06, coupled to a slighter more radiogenic Nd  
610 isotopic composition.

611 The isotopic compositions of early amphibole, late amphibole, and plagioclase from the FI09C06  
612 dyke are significantly richer in radiogenic Sr and unradiogenic Nd isotopes compared with those  
613 reported from nepheline-bearing Triassic intrusions cutting the FPP (Stähle et al., 1990, 2001),  
614 which were interpreted as derived from alkaline melts of OIB affinity.

615 Combining data from nepheline-bearing alkaline dykes, sapphirine-bearing dykes, and the FPP, we  
616 can define a trend at low radiogenic Nd and high radiogenic Sr isotopic ratios, suggesting mixing  
617 between asthenospheric components (OIB, according to Stähle et al., 1990; 2001 and Schaltegger et

618 al., 2015) with materials derived from the continental crust (see compositional fields of Casetta et  
619 al., 2018a). We speculate that the crustal components were located at lithospheric deep levels after  
620 the Variscan orogenic cycle (e.g. Bonadiman et al., 1994) and mobilised by asthenospheric  
621 magmatism. This scenario is supported by the evidence of continental crustal metasomatism  
622 exhibited by mantle bodies involved in the Variscan collisional orogeny, such as the FPP, the Ulten  
623 Zone (Sapienza et al., 2009), and some of the Bohemian Massif (Becker et al., 1999; Schulmann et  
624 al., 2014).

625 In this scenario, it can be noted that the Nd and Sr isotopic compositions of sample FI09C06 lie on  
626 the mixing line between the composition of the Triassic alkaline dykes and the host harzburgite  
627 away from the contact. According to AFC simulation, assimilation of significant degrees of the  
628 metasomatised peridotite material (15%) into OIB must be considered to account for the  
629 composition of the parental melt of the FI09C06 veins.

630 Comparison of the trace elemental patterns of the amphibole from the vein apparently less  
631 contaminated by the host peridotite (FI09C06) with those of amphiboles segregated by primary  
632 hydrous alkaline basalts (e.g. Demeny et al., 2005) shows similarity in REE fractionation, but the  
633 early amphibole has an absolute content that is nearly one order of magnitude higher. In addition, a  
634 correlation with primary alkaline mantle melts is not directly supported by the negative Nb, Ta, and  
635 Ti anomalies shown by the early amphibole.

636 Equilibrium liquids calculated from the trace elemental composition of early amphibole from the  
637 FI09C06 dyke and amphibole-melt partition coefficients experimentally determined at 1,015 °C in  
638 the presence of moderately polymerised melts (dataset T2 1015; Tiepolo et al., 2007) match the  
639 REE contents and fractionation exhibited by shoshonitic rocks of the Triassic magmatism of the  
640 Dolomites area (Casetta et al., 2018a; 2018b; see Supplementary Material C). The calculated melts  
641 also show consistent negative Ti anomalies,  $(\text{Nb/La})_N$  values close to 1, and slight positive U, Th,  
642 and Pb anomalies; the latter are more pronounced in the shoshonitic melts. However, a best match  
643 in terms of U, Th, and Pb concentrations is shown by equilibrium liquids calculated in equilibrium

644 with amphiboles from 9,664 dykes. These observations suggest that the sapphirine-bearing gabbroic  
645 dykes may record the deep mantle input of Triassic K-rich calc-alkaline to shoshonitic melts  
646 erupted into the eastern Southern Alps, also documenting the fractionation trends responsible for the  
647 enrichment in Th, U, and Pb. This hypothesis is also supported by the Sr and Nd isotopic  
648 compositions of Triassic K-rich calc-alkaline to shoshonitic intrusive rocks and lavas of the Eastern  
649 Alps (Casetta et al., 2018a; Figure 12), which approach the isotopic trend defined by our model.

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### 651 *10.3. Constraints on the geodynamic evolution of the Europe–Africa boundary*

652

#### 653 *10.3.1. P–T constraints on the FPP*

654 Geothermobarometric estimates constrain the intrusion of sapphirine-bearing dykes at very high-P  
655 and T conditions. Ab-initio calculations indicate that the primary crystallisation field of sapphirine  
656 in the MAS diagram ( $\text{SiO}_2\text{–MgO–Al}_2\text{O}_3$ ) becomes considerably larger over 1.0 GPa and shrinks at  
657 2.0 GPa (Belmonte et al., 2014). This evidence confirms the experimental results of Liu & Presnall  
658 (1990, 2000) and Milholland & Presnall (1998), indicating that magmatic sapphirine in FPP veins  
659 likely crystallised at  $P \geq 1$  GPa. Equilibrium T estimates based on the sapphirine–spinel  $\text{Mg–Fe}^{2+}$   
660 exchange thermometer of Sato et al. (2006) and the amphibole–plagioclase thermometer of Holland  
661 & Blundy (1994) are mostly higher than 1,000 °C (up to 1,085 °C), confirming the T estimates of  
662 Giovanardi et al. (2013). The melt T was thus significantly higher than that in the host harzburgite,  
663 which shows solidus T typically corresponding to the water-oversaturated peridotite (965 °C at  $P =$   
664 1.1 GPa; Giovanardi et al., 2013). The absence of partial melting evidence in the host FPP confirms  
665 that the source of the uprising melts was at greater mantle depths. The high T, in combination with  
666 the high water and volatile contents, may have allowed the melt to migrate via porous flow along  
667 the direction of structural weakness (see Tommasi et al., 2017) before the opening of the conduits.  
668 The high-P emplacement conditions are consistent with the scenario where the FPP, after  
669 Palaeozoic metasomatism, remained at depths greater than the large lherzolitic mantle bodies south

670 of the Anzola-Val Grande high-T shear zone (namely, from north to south, Premosello, Balmuccia,  
671 and Baldissero; Quick et al., 1995), until its exhumation at shallower levels at ~180 Ma (Zanetti et  
672 al., 2013, 2016; Langone et al., 2017, 2018; Decarlis et al., 2017, Malitch et al., 2017; Petri et al.,  
673 2019).

674

### 675 *10.3.2. Constraints on the Mesozoic mantle sources at the Africa–Europe Boundary*

676 The outcomes of this study indicate that in the Mesozoic, melts extremely rich in volatiles (H<sub>2</sub>O, P,  
677 CO<sub>2</sub>, and Cl), K, Na, and highly incompatible elements rose from mantle depths toward the surface.  
678 The large crustal component present in the melts, as indicated by the O isotopic composition, leads  
679 us to consider that the high Al content, enrichment in large-ion lithophile elements and LREEs, and  
680 enriched Nd and Sr isotopic compositions are a primary feature inherited from the source.

681 This finding confirms the extreme complexity of the tectono-magmatic scenario recorded by the  
682 FPP. In particular, the northern IVZ records an extremely prolonged release of K-H<sub>2</sub>O-rich mantle-  
683 derived melts mixed with subduction-related components, which lasted from the Variscan orogenic  
684 cycle to the Mesozoic exhumation of lithospheric mantle at shallower levels. This explains why the  
685 FPP records many generations of phlogopite-bearing mineral assemblages, showing variable field  
686 relationships, geochemical signatures, and ages (Hartmann & Wedephol, 1993; Zanetti et al., 1999;  
687 2013; 2016; Stähle et al., 1990; 2001; Greco et al., 2001; 2004; Morishita et al., 2003, 2008;  
688 Malitch et al., 2017). It also provides a new interpretative frame for previous data indicating the  
689 emplacement of melts with subduction-related components in the Triassic (Mastumoto et al., 2005;  
690 Morishita et al., 2008, Malitch et al., 2017).

691 This magmatism-bearing subduction component appears to have roughly overlapped with the ascent  
692 of silica-undersaturated alkaline melts of OIB affinity (Stähle et al., 1990, 2001, Schaltegger et al.,  
693 2015), which likely also have some counterparts in the central IVZ (Fiorentini et al., 2018; Galli et  
694 al., 2019).

695 The FPP developed in a concomitant extensional–transtensional tectonic regime, the origin and  
696 geodynamic scenario of which are still strongly debated (Cassinis et al., 2008; Zanetti et al., 2013;  
697 Casetta et al., 2018a, b). Further investigations are needed to address whether the crustal  
698 components are remnants of old subduction events (Bonadiman et al., 1994), possibly located at  
699 lithospheric levels, reactivated by asthenospheric magmatism with OIB or depleted mantle affinity;  
700 represent crustal material recycled into asthenospheric mantle sources (Locmelis et al., 2016); or  
701 were added by active Mesozoic subduction (Cassini et al., 2008; Schmid et al., 2008; Morishita et  
702 al., 2008; Selverstone and Sharp, 2011; Zanetti et al., 2013).

703 It is a matter of fact that the major elements (in particular the enrichment in K and Al), trace  
704 elements, and Sr and Nd isotopic compositions of the sapphirine-bearing gabbroic rocks approach  
705 the geochemical affinity shown by Triassic K-rich calc-alkaline to shoshonitic magmatism of the  
706 Dolomites area (Eastern Alps) (Casetta et al., 2018a, b). Thus, the study of magmatic events at the  
707 roots of the continental crust of the Adria Plate can provide a unique opportunity to constrain the  
708 tectono-magmatic evolution of the Europe–Africa boundary.

709

## 710 **11. Concluding remarks**

711 A detailed survey of Mesozoic sapphirine-bearing gabbroic dykes within the FPP unit led us to  
712 recognise different stages of melt migration (from porous-flow migration in peridotite channels to  
713 flow in open conduits) and constrain the presence in the parental melts of a large continental crustal  
714 component acquired in the source region or at deeper lithospheric mantle levels.

715 The reaction between dyke melts and the strongly metasomatised FPP enhanced the crustal  
716 signature.

717 The high Al content of these melts allowed the segregation of magmatic sapphirine, which is a  
718 marker of high-P-T intrusive conditions.

719 This dyke swarm possibly represents a record of the input to the mantle of the K-rich calc-alkaline  
720 to shoshonitic melts that characterised the magmatism widespread during the Triassic in the  
721 Southern Alps area.

722

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1020 **Figure captions**

1021

1022 Figure 1: Geological map of the Finero area, modified after Mazzucchelli et al. (2014).

1023

1024 Figure 2: A) Sample FI09C06 crosscutting the host harzburgite foliation. The centre of the dyke is  
1025 formed by the Leucocratic Zone, whereas the melanocratic zones (i.e. the Orthopyroxene Zone,  
1026 Early Amphibole Zone, and Late Amphibole Zone) are indistinguishable; B) phlogopite vein  
1027 cutting the Orthopyroxene Zone; C–F) occurrences of sapphirine in sample FI09C06 (C and D) and  
1028 sample FI9664 (E and F). The figures show the increase of sapphirine size, from  $\mu\text{m}$  (C) to mm (D  
1029 and E) and cm scale (F).

1030

1031 Figure 3: Primitive mantle (PM)-normalised bulk rock trace elemental patterns of a gabbroic dyke  
1032 and host harzburgite. PM values are from McDonough and Sun (1995). Literature values from the  
1033 harzburgite-pyroxenite association of (1) Hartmann and Wedephol (1993) and nepheline-bearing  
1034 hornblende syenitic dykes of (2) Stähle et al. (2001) are reported for comparison.

1035

1036 Figure 4: Orthopyroxene major elemental contents from host harzburgites, dykes, and literature  
1037 data. Plotted literature data are the harzburgite-pyroxenite association orthopyroxene from (1)  
1038 Zanetti et al. (1999) and orthopyroxene from sapphirine-bearing rock from the Mafic Complex from  
1039 (2) Sills et al. (1983).

1040

1041 Figure 5: Phlogopite major elemental contents from host harzburgites, dykes, and literature data.  
1042 Plotted literature data are: harzburgite-pyroxenite association phlogopite from (1) Zanetti et al.  
1043 (1999).

1044

1045 Figure 6: Amphibole major elemental contents from host harzburgites, dykes, and literature data.  
1046 Plotted literature data are: harzburgite-pyroxenite association amphibole compositions from (1)  
1047 Zanetti et al. (1999) and (2) Morishita et al. (2008), amphibole from apatite-rich veins in the FPP

1048 from (1) Zanetti et al. (1999) and (2) Morishita et al. (2008), and amphibole from sapphirine-  
1049 bearing rock from the Mafic Complex from (3) Sills et al. (1983).

1050

1051 Figure 7: Sapphirine compositions plotted in the (MgO + FeO)–(Cr<sub>2</sub>O<sub>3</sub> + Fe<sub>2</sub>O<sub>3</sub> + Al<sub>2</sub>O<sub>3</sub>)–SiO<sub>2</sub>  
1052 diagram (mol. %). Literature data are from sapphirine of the Finero Mafic Complex (Sills et al.,  
1053 1983) and sapphirine 1 of Higgins et al. (1979).

1054

1055 Figure 8: REE patterns of clinopyroxene and amphibole from the gabbroic dykes and the respective  
1056 host rocks divided for samples and position. Values are normalised to Chondrite-I (CI, values from  
1057 Lyubetskaya and Korenaga (2007);  $\delta^{18}\text{O}$  values (normalised to SMOW) from phases of sample  
1058 FI09C06 are reported near the REE patterns according to their positions in the sample.

1059

1060 Figure 9: PM-normalised trace elemental patterns of clinopyroxene from host harzburgite. PM  
1061 values are from McDonough and Sun (1995). Plotted literature data are: clinopyroxene and  
1062 amphibole average compositions from the harzburgite-pyroxenite association from (1) Zanetti et al.  
1063 (1999) and (2) Morishita et al. (2008), clinopyroxene and amphibole average from apatite-rich veins  
1064 in the FPP from (1) Zanetti et al. (1999) and (2) Morishita et al. (2008), reported as apatite veins.

1065

1066 Figure 10: PM-normalised trace elemental patterns of amphiboles from gabbroic dyke and host  
1067 harzburgite. PM values are from McDonough and Sun (1995).

1068

1069 Figure 11:  $\delta^{18}\text{O}$  normalised to SMOW. Values from literature are the field of mantle and variation  
1070 of MORB from Rollinson (1993), and mantle ultramafics and mantle-derived melts (\*) from  
1071 Bindeman (2008). Finero literature data are reported from the harzburgite-pyroxenite association of  
1072 Hartmann and Wedephol (1993) and Selverstone and Sharp (2011) and from chromitite in dunite  
1073 bodies from Zanetti et al. (2016).

1074

1075 Figure 12:  $^{143}\text{Nd}/^{144}\text{Nd}$  vs  $^{87}\text{Sr}/^{86}\text{Sr}$  recalculated at 225 Ma for amphibole and plagioclase from the  
1076 various zones of sample FI09C06 (host and dyke): (1) amphibole data of the FPP from Obermiller  
1077 (1994); (2) bulk rock data of the Finero Mafic Complex from Lu et al. (1997b); (3) data from an  
1078 alkaline dyke in the FPP from Stähle et al. (1990); (4) data from an alkaline dyke in the FPP from  
1079 Stähle et al. (2001); (5) amphibole and apatite data from discordant apatite-bearing veins in the FPP  
1080 from Morishita et al. (2008); (6) Mesozoic shoshonitic magmatism (SS: silica saturated; US:  
1081 undersaturated in silica) in the Predazzo area from Casetta et al. (2018a, b); Depleted MORB  
1082 Mantel (DMM) from Workman and Hart (2005). The mixing model was calculated between the two  
1083 end-members: the alkaline dyke S9 of Stähle et al. (1990) and the hosting peridotite FI09C06. Data  
1084 for the melt in equilibrium with the alkaline dyke S9 are:  $^{143}\text{Nd}/^{144}\text{Nd} = 0.512607$ ,  $^{87}\text{Sr}/^{86}\text{Sr} =$   
1085  $0.703720$  (from Stähle et al., 1990; recalculated at 225 Ma), Nd = 8.333 ppm, and Sr = 830 ppm  
1086 (calculated from LA-ICP-MS plagioclase analysis of albitite dykes similar to the dyke of Stähle et  
1087 al., 1990, using the  $K_d$  of Dohmen and Blundy, 2014; average Nd = 0.25 ppm, Sr = 5810 ppm).  
1088 Data for the melt in equilibrium with the host FPP peridotite are  $^{143}\text{Nd}/^{144}\text{Nd} = 0.512130$ ,  $^{87}\text{Sr}/^{86}\text{Sr} =$   
1089  $0.708501$  (recalculated at 225 Ma), Nd = 67.5 ppm, and Sr = 909 ppm (calculated using the  $K_d$  of  
1090 Ionov et al., 2002).

Figure 1

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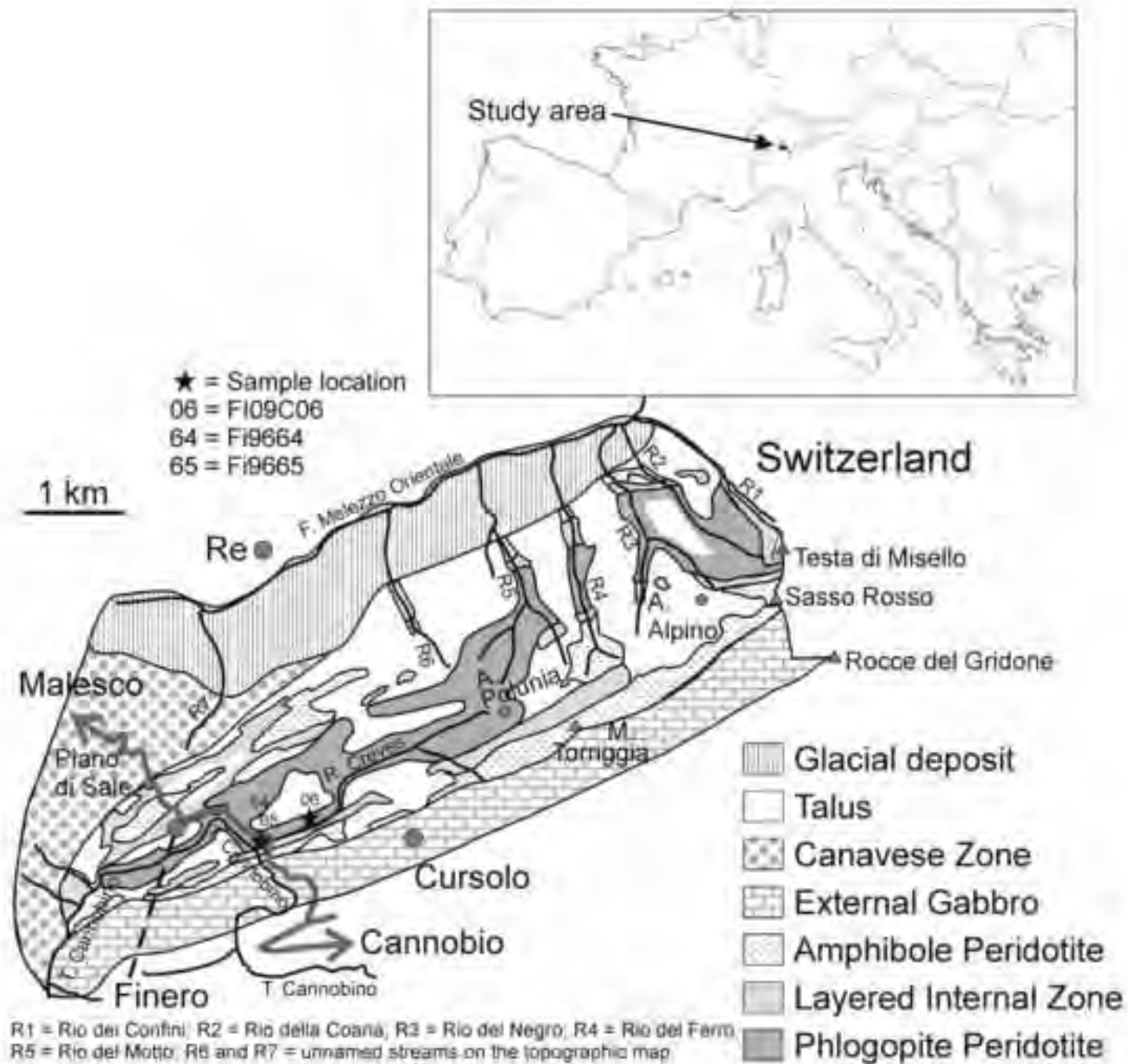


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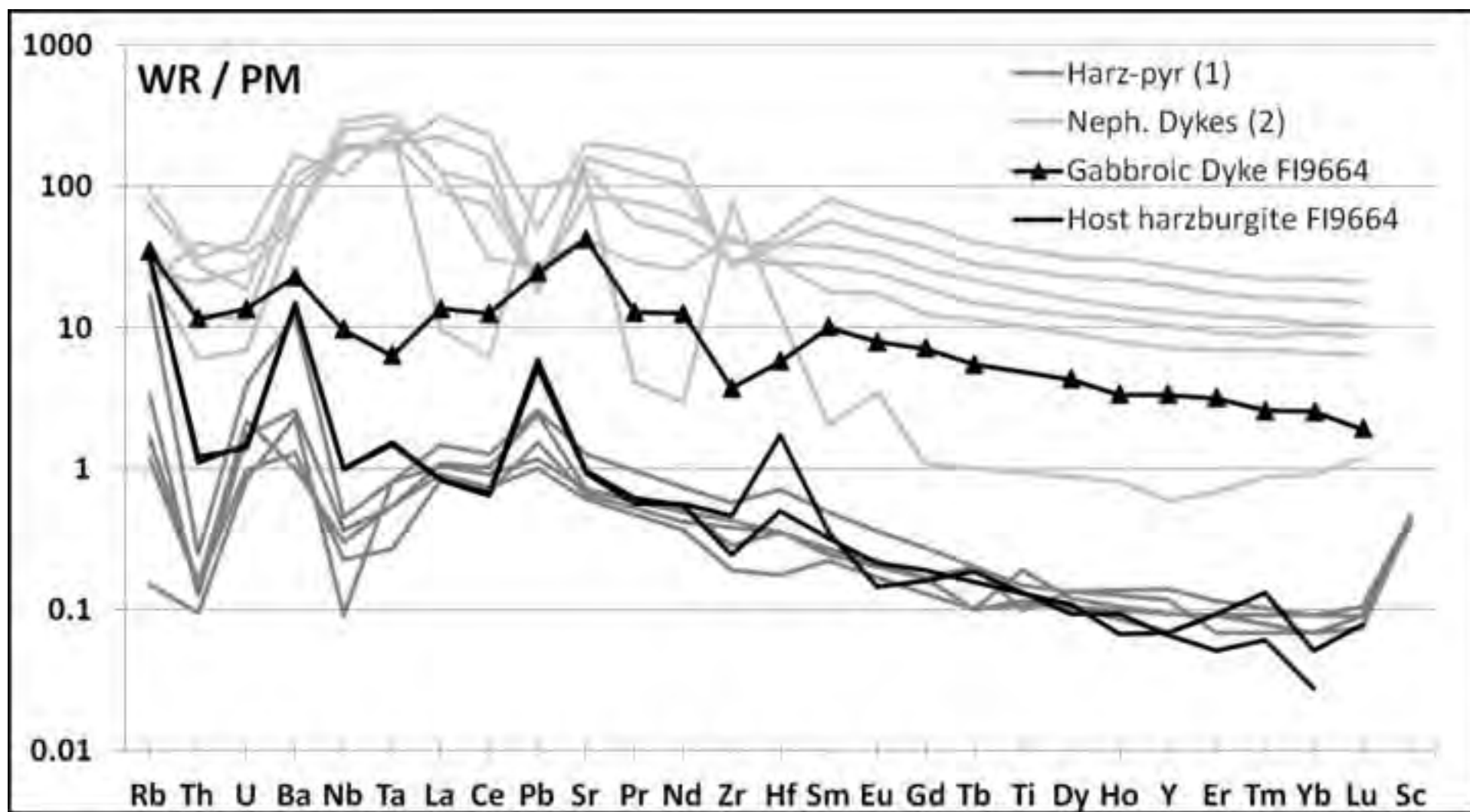


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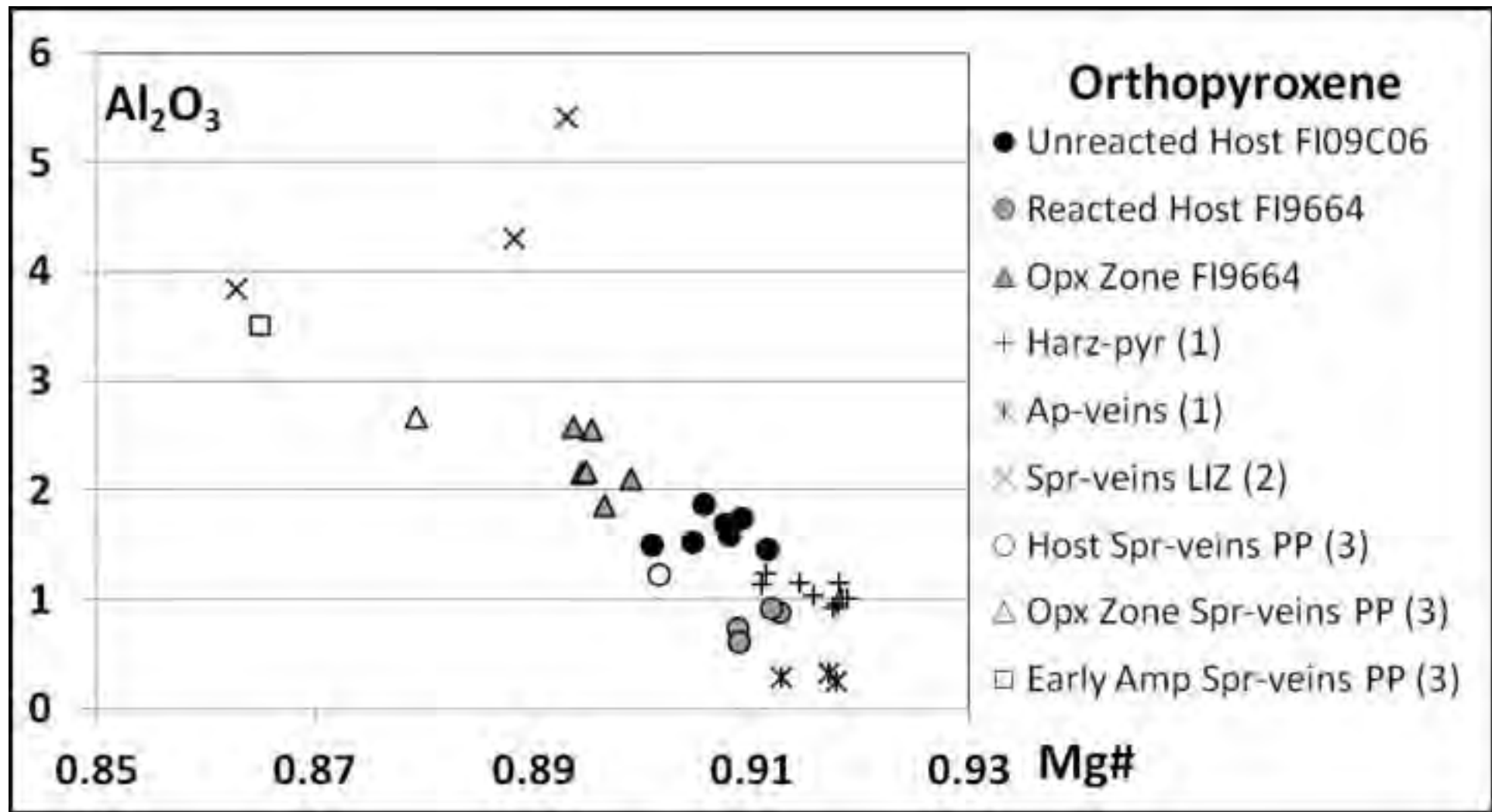




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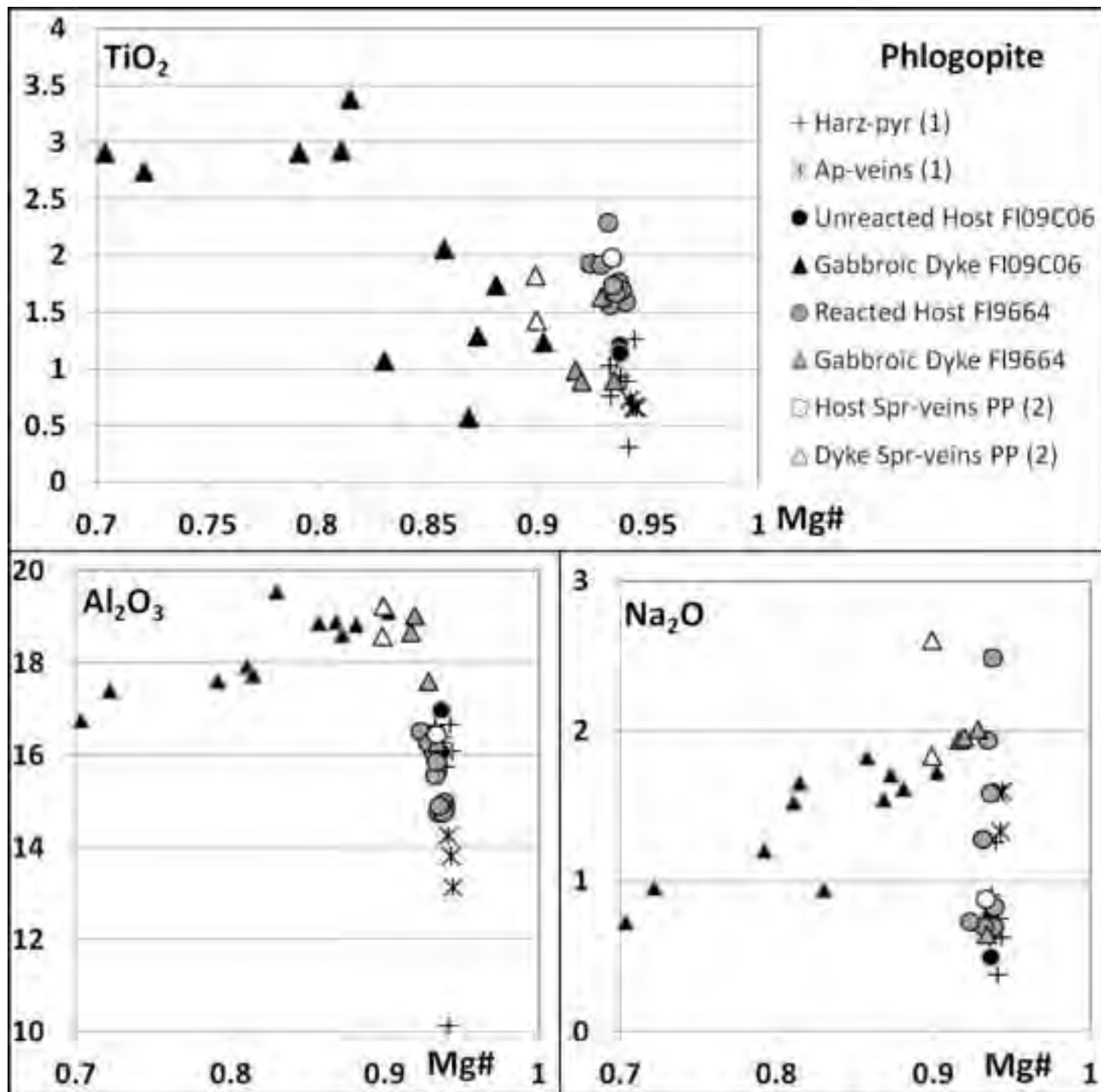


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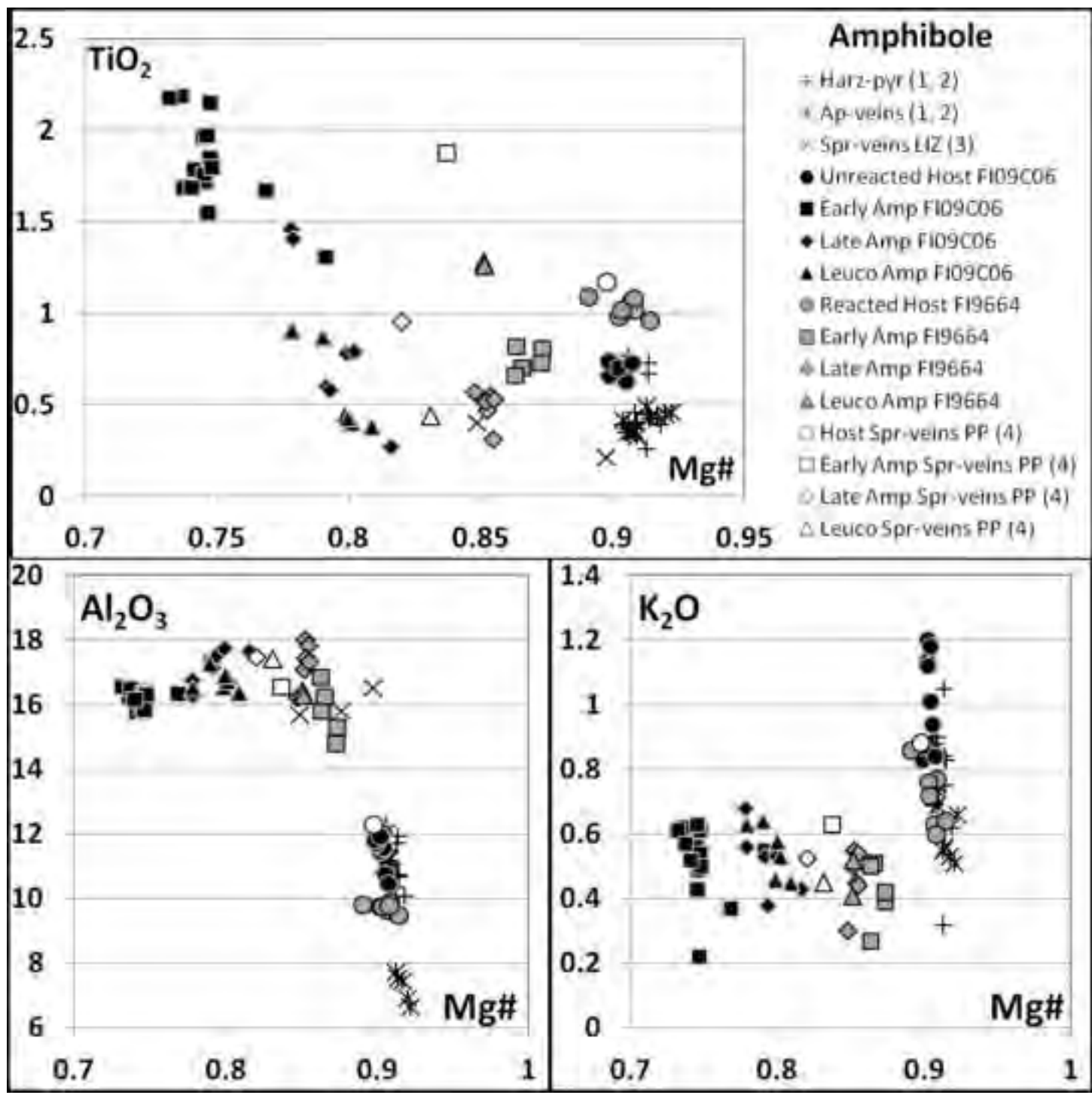


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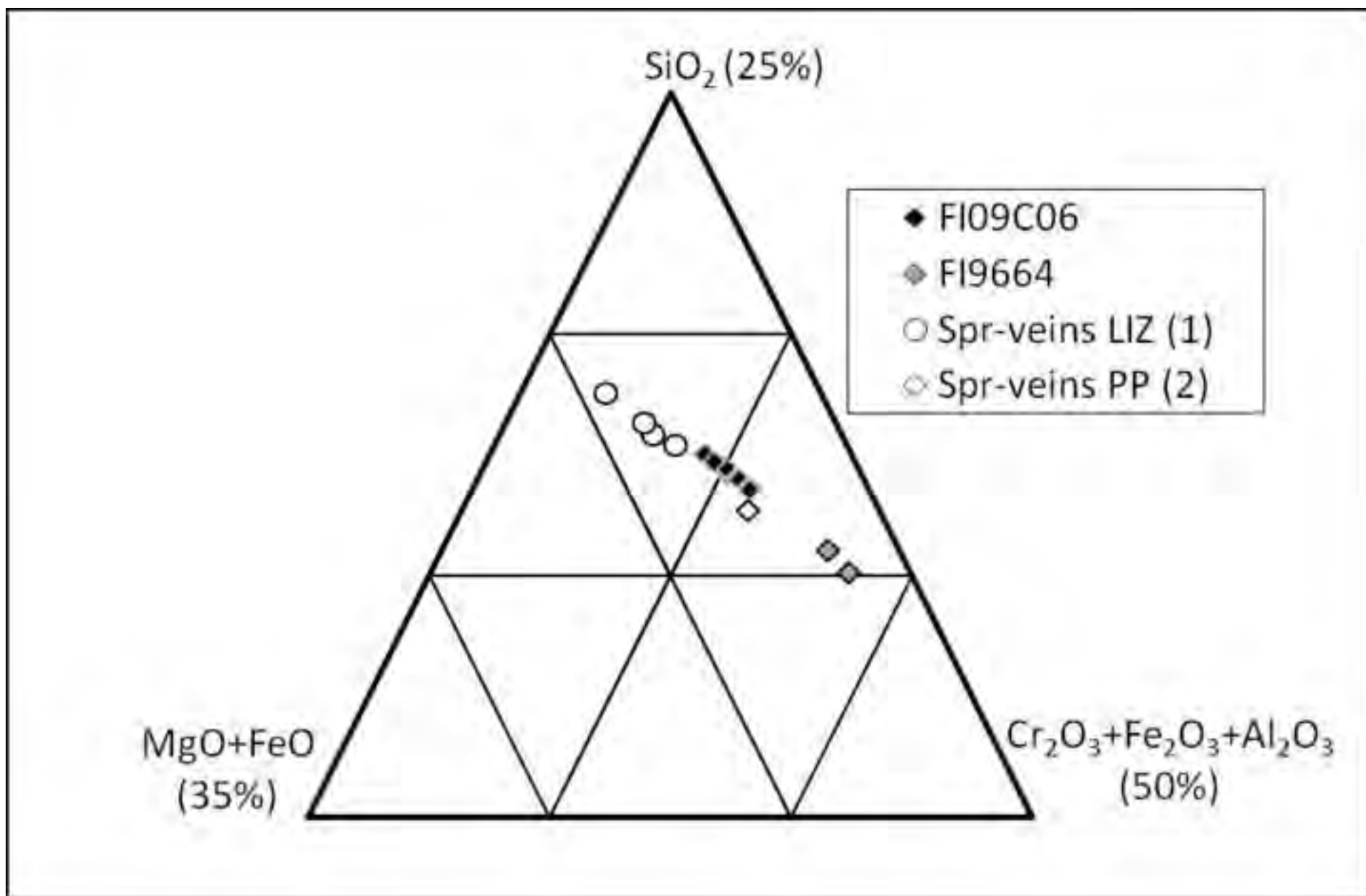


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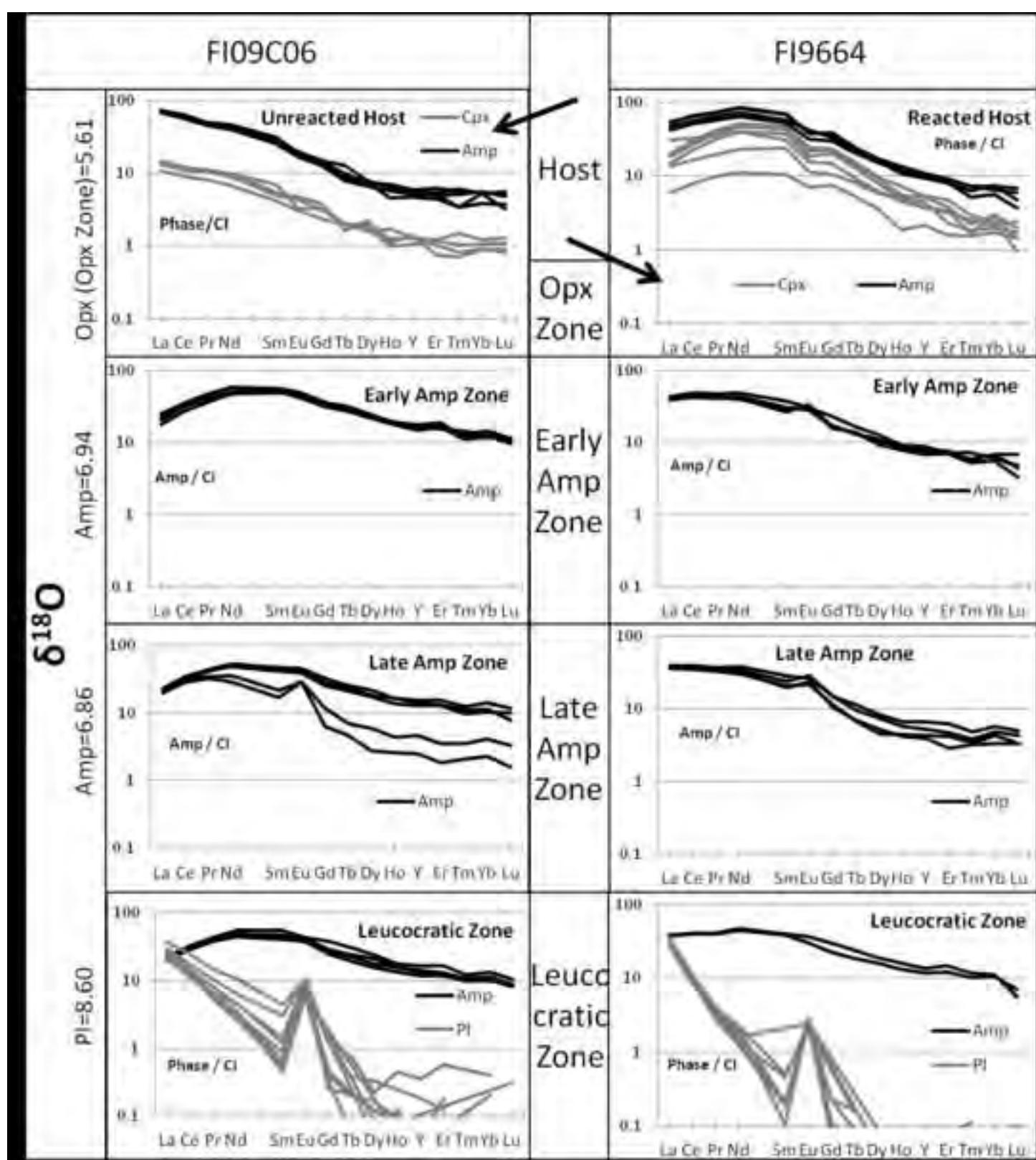


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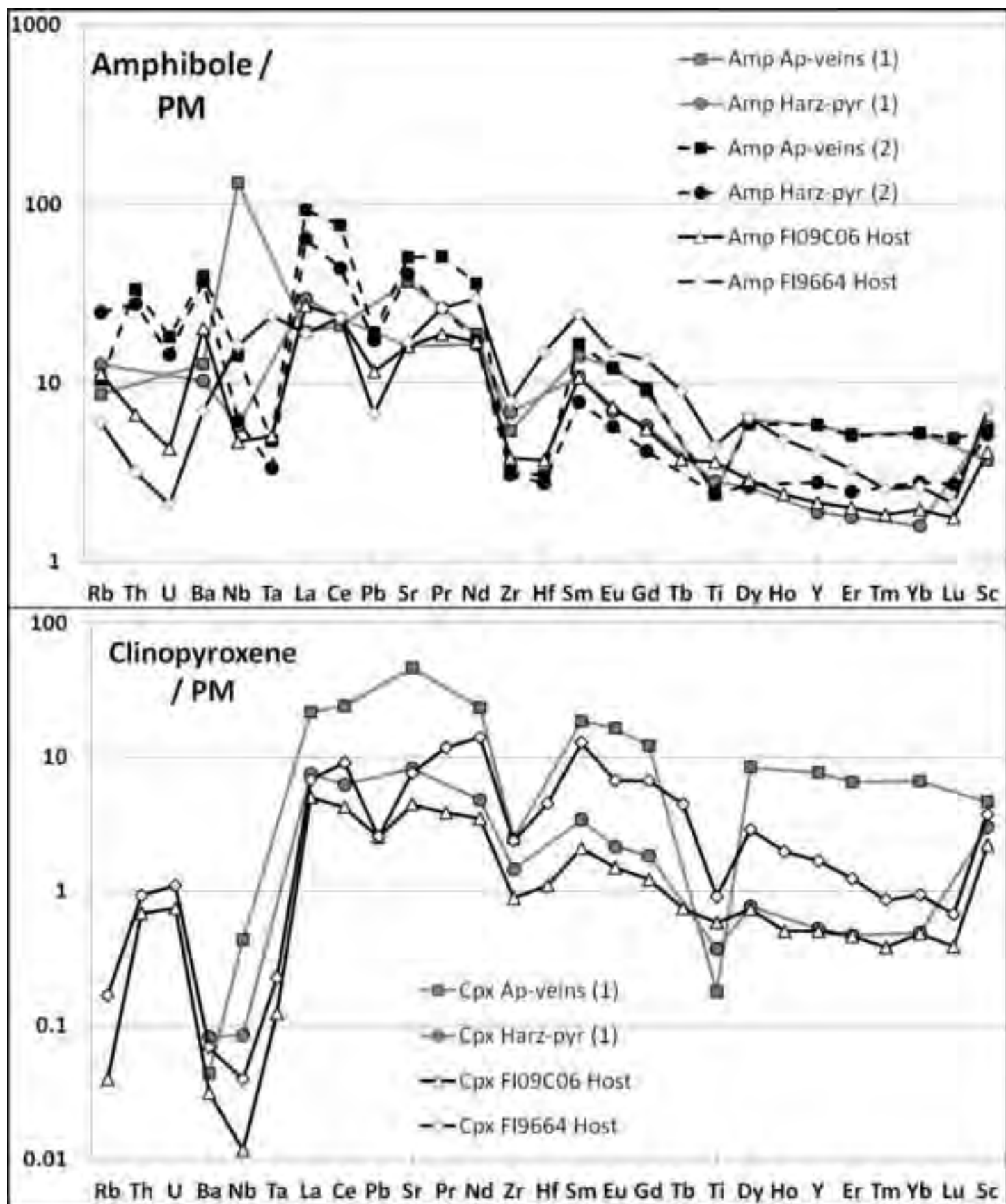


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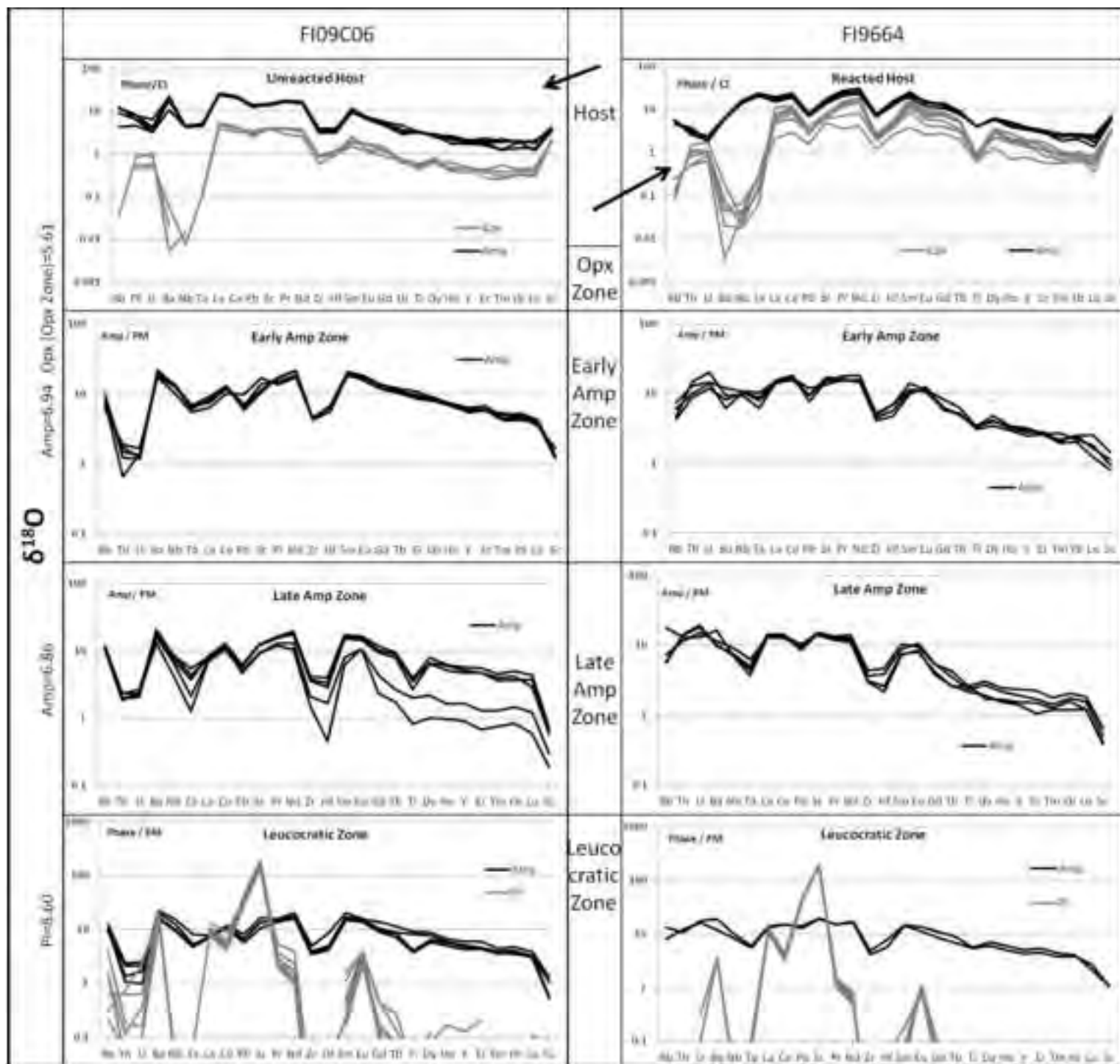


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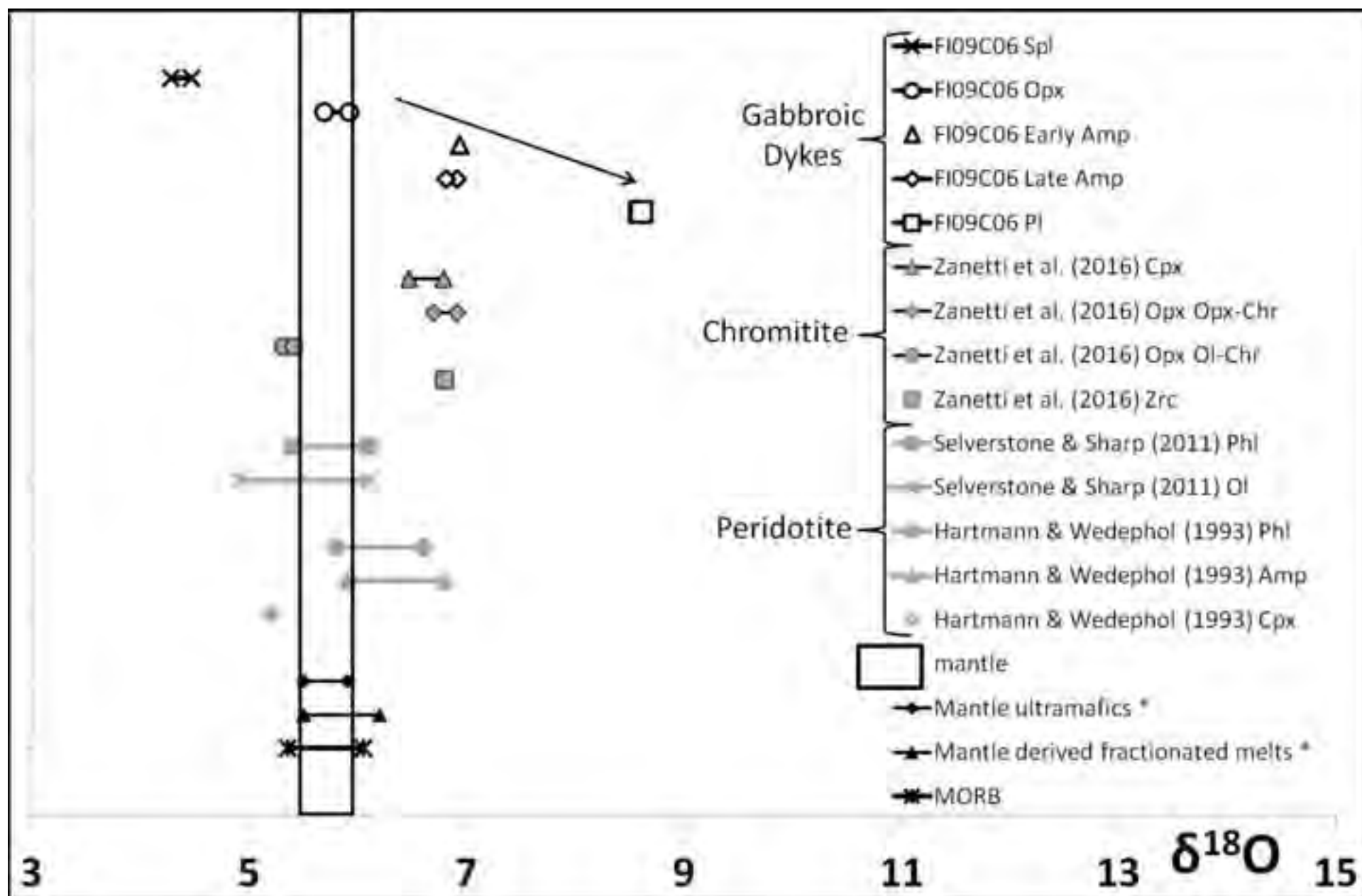
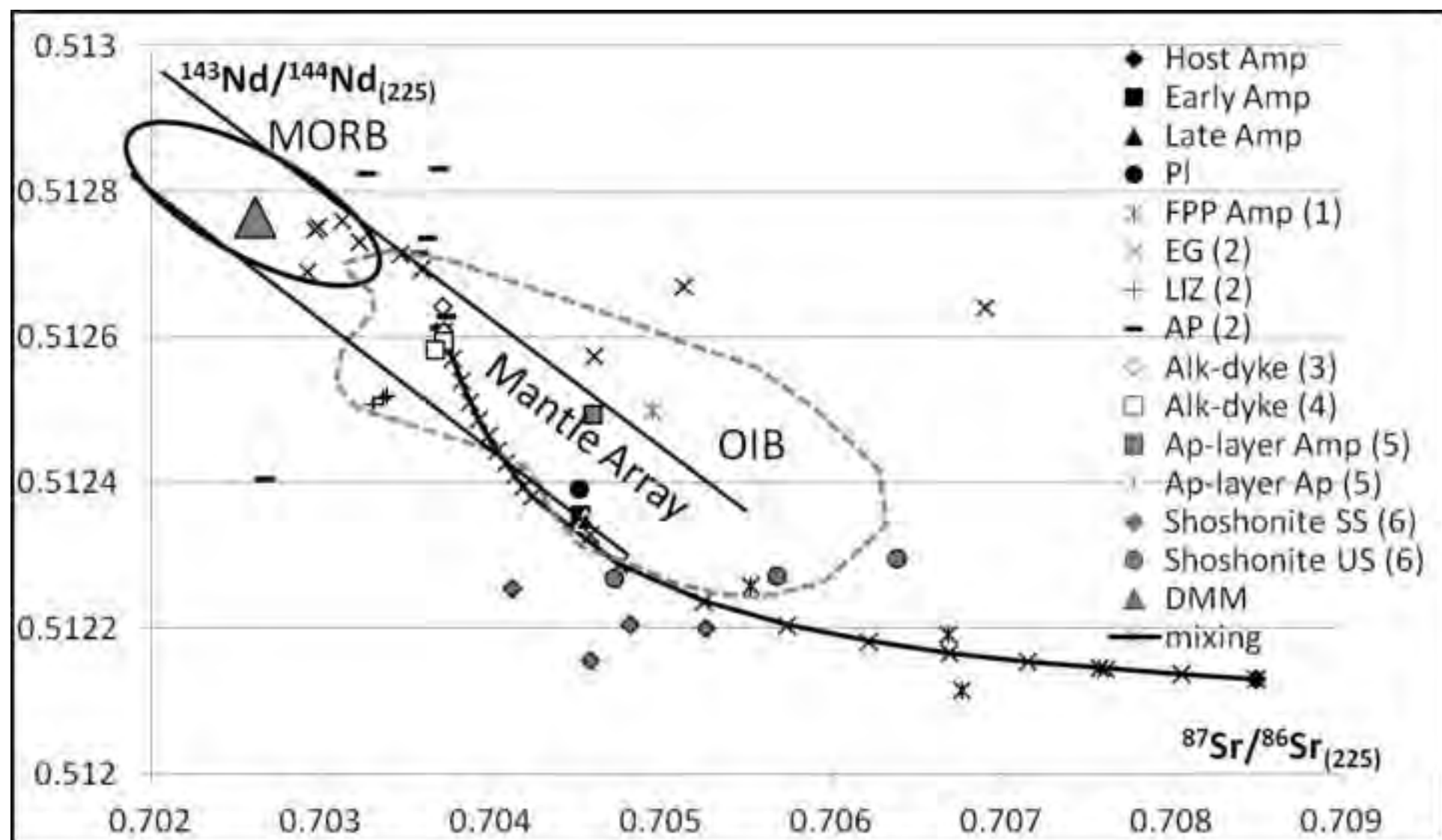


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**Table 1**[Click here to download Table: Table 1\\_rev.docx](#)

Table 1: O isotopic composition of minerals from the Spr-bearing gabbroic dykes.

<b>Sample</b>	<b>Phase</b>	<b><math>\delta^{18}\text{O}</math></b>	<b>std. dev.</b>
FI09C06	Early Amp	6.94	0.00
FI09C06	Late Amp	6.86	0.05
FI09C06	Pl	8.60	0.01
FI09C06	Opx	5.81	0.11

**Table 2**[Click here to download Table: Table 2\\_rev.docx](#)

Table 2: Sr and Nd isotopic composition of minerals from sample FI09C06 from the Spr-bearing gabbroic dykes and the host peridotite.

Rock	Phase	Rb	Sr	<sup>87</sup> Sr/ <sup>86</sup> Sr	2SE	<sup>87</sup> Rb/ <sup>86</sup> Sr	Nd	Sm	<sup>143</sup> Nd/ <sup>144</sup> Nd	2SE	<sup>147</sup> Sm/ <sup>144</sup> Nd
Host	Amp	6.5	281.2	0.708713	0.000008	0.066372	19.7	4.2	0.512317	0.000008	0.126796
Dyke	Early Amp	6.8	324.4	0.704722	0.000008	0.060834	24.8	8.1	0.512646	0.000008	0.197559
Dyke	Late Amp	8.6	189.5	0.704971	0.000008	0.131167	16.8	4.0	0.512559	0.000006	0.142593
Dyke	Pl	0.6	3109.7	0.704519	0.000008	0.000568	2.2	0.2	0.512482	0.000006	0.061238