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Cryptic crustal growth identified through Variscan post-collisional lamprophyre-granite composite dykes, French Massif Central.

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Abstract:	<p>The processes that control crustal growth by addition of new mantle-derived material to the crust and the recycling of crustal components that were introduced into the mantle during subduction remain poorly understood in collisional orogens. This is largely due to the fact that the sub-orogenic mantle has been fertilized by the introduction of crustal components during previous subduction events, and lacks the typical depleted mantle signature. Here, we present a study of lamprophyric-granitic composite dykes from the northern border of the post-collisional Variscan Aigoual pluton, French Massif Central (FMC). At Aigoual, lamprophyres and granites have gradational contacts and evidences in the field and similar geochemical features that they coexisted indicate that they are comagmatic. U-Pb dating of zircon yields crystallisation ages between 311 ± 2.9 Ma and 313 ± 3.0 Ma for dykes, synchronous within uncertainty with the ages of 312 ± 3.1 Ma and 313 ± 3.2 Ma obtained for the emplacement of the Aigoual pluton. The lamprophyres are metaluminous to slightly peraluminous and display enrichment in both compatible (Fe, Mg, Ni, Cr) and incompatible elements (K₂O, LILE, LREE) and have crustal isotopic signatures in both radiogenic (Sr, Nd and Hf) and stable (O) isotope systems. This dual crust/mantle signature of the lamprophyres indicates partial melting, within the spinel-lherzolite field, of an enriched mantle that was metasomatised by sediment-derived melts or fluids related to subduction preceding continental collision. The granites are peraluminous, display high K₂O and have shoshonitic affinities. Their trace elements patterns are similar to the lamprophyres. The isotopic compositions of the granites overlap those of the lamprophyres. The high Cr, Ni, FeO, and MgO contents of the granites are consistent with a mantle component in the magmas. Based on our results, we suggest that between 65 and 80% of material in the lamprophyres and granites was derived from the depleted mantle, and the rest from recycled crustal material. Although blurred from an isotopic point of view, this magmatism contributes with crustal growth in the post-collisional magmatism of the FMC. Coupled with their high preservation potential, post-collisional sites might represent significant sites of crust growth.</p>
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19 normalized trace-element patterns are similar to the lamprophyres and the isotopic compositions
20 of the two rock types overlap. The high Cr, Ni, FeO, and MgO contents of the granites are
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22 wt% of material in the lamprophyres and granites was derived from the depleted mantle, and the
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26 significant sites of crustal growth.

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Highlights

- Lamprophyre-granite composite dykes from late-Variscan Aigoual pluton, French Massif Central;
- Mantle-crust interaction in a post-collisional setting;
- Sr-Nd-Hf and O isotopes display a cryptic crustal growth.

[Click here to view linked References](#)

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46

47 **1. Introduction**

48 The process that allow continental crust to be generated and evolve involve the addition of
49 new, mantle-derived magma to the crust; the recycling and melting of oceanic crust and
50 sediments at subduction sites and the addition of these magmas to the continental crust; as well as
51 the anatectic reworking of the continental crust in collisional environments (Dhuime et al., 2011;
52 Jacob et al., 2021; Moyen et al., 2021). The continental crust grows when new, purely mantle-
53 derived material is added to the continents, not when continental crustal components introduced
54 into the mantle are returned to the crust; this is simply crust coming home. Growth of the
55 continental crustal occurred extensively in arc settings throughout geological time, where large
56 amounts of mantle-derived mafic and intermediate magmas (and their differentiates) were added
57 to the crust. Nevertheless, this does not necessarily imply significant long-term crustal-growth,
58 because arc settings display poor preservation potential. A high proportion of the crust generated
59 in this setting may be recycled back to the mantle shortly after formation (Kay and Kay, 1993;
60 Condie, 2014; Scholl and von Huene, 2009). Although less efficient than arc settings in
61 producing new continental crust, post-collisional sites display the greatest potential for
62 preservation in the geological record (Hawkesworth et al., 2009; Spencer et al., 2015). Post-
63 collisional settings are characterized by orogenic collapse, potential delamination of the
64 lithosphere and hence asthenospheric upwelling. This may produce a range of magmas, including
65 granites, from both mantle and crustal sources (Bonin, 2004; Clemens et al., 2009). Thus, post-
66 collisional settings represent a significant opportunity for crustal growth in an environment with
67 potential for long-term preservation (Couzinié et al., 2016; Moyen et al., 2017; Gómez-Frutos et
68 al., 2023).

69 In this context, high-K–Mg mantle-derived magmatic suites occur systematically in post-
70 collisional settings. Although relatively scarce, the suites are considered a significant contributor
71 to crustal growth throughout geological time (Prelević et al., 2012; Scarrow et al., 2008; Murphy,
72 2020). Post-collisional high-K–Mg mafic rocks have been described in the Variscan belt of
73 Western Europe (e.g. Turpin et al., 1988; Janoušek & Holub, 2007; Couzinié et al., 2016; von
74 Raumer et al., 2014). These rocks are characterized by a significant range of SiO₂ contents (45–
75 65 wt%), high-K, through shoshonitic to ultrapotassic affinity (K₂O = 1.5–6.0 wt%), and high
76 maficity (FeO_t + MgO up to 25 wt%). They are enriched in volatiles, mantle-compatible (Mg, Ni,
77 Cr) and incompatible elements (e.g., K, Rb, Ba, Sr).

78 The French Massif Central is a segment of the Variscan Belt, in which high-K–Mg mafic
79 rocks are spatially and temporally associated with the peak of granite magmatism (335 to 300
80 Ma; Couzinié et al., 2014; Laurent et al., 2017). These mafic rocks occur as either enclaves to
81 decametric small bodies in felsic igneous rocks and migmatites, or as lamprophyric dykes
82 (Couzinié et al., 2014; 2016; Sabatier, 1991). Despite being well defined in space and time, the
83 role of this mafic magmatism in the post-orogenic phase of the Variscan orogeny is still debated
84 (e.g. Tabaud et al., 2015; Couzinié et al., 2016). More specifically, it is unclear whether the
85 volumetrically more significant granites represent differentiated magmas from the mafic rocks
86 (Clemens et al., 2021; Moyen et al., 2017) or the products of hybridization between the mafic and
87 crustal melts (Laurent et al., 2017, Solgadi et al., 2007, Yamasaki, 2022). In turn, the balance
88 between crustal growth by addition of new mantle-derived material and crustal recycling into the
89 mantle in post-collisional sites is poorly understood (Jacob et al., 2021; Storck et al., 2021, and
90 references therein). To contribute to this debate, we selected co-magmatic lamprophyres and
91 granites in a unique system of parallel dykes associated with the northern boundary of the late-

92 Variscan Aigoual granite, French Massif Central, to investigate the petrogenesis of both rock
93 types, explore the potential genetic links between them and to constrain the contribution of
94 mantle-derived material to these post-collisional magmas.

95 **2. Geological setting**

96 **2.1 The French Massif Central**

97 The orogenic Variscan Belt in central and Western Europe formed between c. 400 and 280
98 Ma in a complex history of collision between two main continents, Gondwana and Laurussia, as
99 well as intervening micro-continents and oceanic basins, resulting in the formation of the Pangea
100 supercontinent at the end of the Paleozoic (Kröner and Romer, 2013; Stampfli et al., 2013; Matte,
101 2001). One of the largest exposures of the Variscan Belt in Western Europe is preserved in the
102 French Massif Central (FMC) (Vanderhaeghe et al., 2020). Igneous and metamorphic rocks here
103 represent the exhumed root zone of the orogen (Lardeaux et al., 2014), constructed over the
104 northern margin of Gondwana (Fig. 1b).

105 From the Ordovician to the late Carboniferous, the FMC experienced several
106 tectonometamorphic–magmatic events, with opening of ocean basins (ca. 500–450 Ma) and
107 subsequently closure and subduction between 420 and 360 Ma, to continental collision between
108 ca. 350 and 320 Ma. Orogenic collapse occurred in the late Carboniferous (ca. 320–300 Ma)
109 (Faure et al., 2009; Vanderhaeghe et al., 2020). The eastern FMC (E-FMC) consists of a south-
110 verging stack of nappes that are composed of four main metamorphic units (Faure et al., 2009;
111 Vanderhaeghe et al., 2020).

112 In the southeastern part of the FMC, a low-grade metamorphic unit called the
113 Parautochthonous Unit represents the lowermost unit of the nappe pile. It exposes predominantly
114 metasediments (micaschists, quartzites and graphitic metapelites) (Faure et al., 1999) deposited

115 between the Neoproterozoic and early Cambrian (Melleton et al., 2010), minor orthogneisses
116 formed in the late Cambrian to Early Ordovician (490–485 Ma), and metarhyolites. These rocks
117 underwent polyphase deformation and metamorphism between ca. 340 and 310 Ma under
118 greenschist- to amphibolite-facies conditions (Faure et al., 1999) (Fig. 1b). This unit was intruded
119 by the Aigoual–Saint Guiral–Liron Pluton (ASGL) at shallow depth (1–2 kbar; Najoui et al.,
120 2000) developing a contact metamorphic aureole. Based on U–Pb dating in zircon and monazite,
121 and $^{40}\text{Ar}/^{39}\text{Ar}$ dating in biotite, the main intrusion occurred between 305 and 310 Ma. Thus it was
122 synchronous with the late stage of migmatization at lower structural levels and the late Variscan
123 extensional event in this area (Brichau et al., 2008).

124 **2.2 Magmatic relations in the E-FMC and Aigoual granite**

125 The main phase of igneous activity in the French Massif Central occurred during the
126 transition from mid-Carboniferous collision to the syn- and post-orogenic collapse of the orogen
127 in late-Carboniferous to early-Permian times (Faure et al., 2009; Laurent et al., 2017). The
128 plutonic rocks of E-FMC can be subdivided in four main groups (Chantraine et al., 1996;
129 Barbarin, 1999). (1) Two-mica or muscovite-bearing peraluminous granitoids (“MPG”) cropping
130 out as small plutons and dykes. These rocks are granites *sensu stricto* with SiO_2 (>70 wt%); their
131 chemistry and isotopic characters indicate mature crustal source (Pin and Duthou, 1990); (2)
132 Cordierite-bearing peraluminous granitoids (“CPG”), which are granodiorites to granites with
133 biotite and cordierite ranging from abundant to rare. These rocks are characterized by K_2O
134 contents in the range of 3–7 wt%, with $\text{SiO}_2 > 65\text{wt}\%$. Isotopic characteristics suggest melting of
135 local crust as the source of the magmas that formed these rocks (Pin and Duthou, 1990;
136 Williamson et al., 1992). The compositions of the CPG are consistent with higher melting
137 temperatures (≥ 850 °C, biotite breakdown) than the MPG (Laurent et al., 2017; Villaros et al.,

138 2018) and/or greater proportions of orthogneisses in the source (Downes et al., 1997; Moyen et
139 al., 2017), although with limits imposed by their isotopic compositions. (3) K-feldspar-phyrlic,
140 biotite ± amphibole-bearing granites, granodiorites and quartz diorites (“KCG”) are high-K calc-
141 alkaline, metaluminous to weakly peraluminous and form large plutons (e.g., Aigoual–Saint
142 Guiral–Liron) or enclaves in the Velay Complex. In general, they are more mafic than the CPG,
143 having $\text{FeO}^{\text{t}} + \text{MgO} \sim 4$ wt%, with SiO_2 contents averaging around 69 wt%. (4) A suite of
144 biotite–amphibole ± clinopyroxene-bearing mafic to intermediate rocks that occur throughout the
145 Variscan Belt (von Raumer et al., 2014) and are locally known as vaugnerites (Couzinié et al.,
146 2016; Sabatier, 1991). The vaugnerites are metaluminous with low silica ($44 < \text{SiO}_2 < 63$ wt%),
147 high $\text{FeO}^{\text{t}} + \text{MgO}$ (10–24 wt%), Cr (200–600 ppm) and Ni (100–220 ppm) contents, Mg#
148 (molecular $\text{Mg}/(\text{Mg} + \text{Fe}) > 0.65$). They also have high K_2O contents (1.6–6.7 wt%) and high Ba
149 and Sr contents (1000–2000 ppm; Couzinié et al., 2016). In this contribution, we refer to these
150 rocks as lamprophyres.

151 The Aigoual granite forms the northern end of the L-shaped Aigoual-Saint Guiral-Liron
152 (AGSL) Pluton in the southeastern FMC. The ASGL rocks are classified as K-feldspar-phyrlic
153 calc-alkaline granitoids (KCG) (Chantraine et al., 1996). The northern termination of the Aigoual
154 granite consists of a swarm of NE–SW to NNE–SSW-trending lamprophyre–granite composite
155 dykes, which are the subject of this study (Fig. 1c).

156 **3. Analytical methods**

157 This study has integrated information on field relations, petrography, mineral chemistry,
158 whole-rock major and trace element compositions, whole-rock Nd, Sr and Hf isotopic ratios, U–
159 Pb dating and Lu–Hf analyses in zircon and oxygen isotopes in quartz and K-feldspar, from the
160 Aigoual lamprophyre-granite composite dykes and three samples from the Aigoual granite. The

161 whole-rock major- and trace-element data information of mineral major element compositions, as
162 well as U–Pb dating of zircon, were undertaken at the Central Analytical Facilities at
163 Stellenbosch University, South Africa. The whole-rock Nd, Sr and Hf isotope analyses and quartz
164 and K-feldspar oxygen isotope ratio analyses were undertaken at the University of Cape Town.
165 Lu–Hf analyses in zircon were carried out at Universidade Federal de Ouro Preto, Brazil. The
166 analytical details for each method are described in the Supplementary Material 1.

167 **4. Results**

168 **4.1 Field relations**

169 At the northern end of the Aigoual granite, several 100–1000 m long, NE-trending dykes
170 intrude the Cévennes micaschists, part of the Parautochthonous Unit described above. Most
171 dykes are composite and display gradation between granitic and lamprophyric components. The
172 wider dykes (described by Sabourdy, 1975) display lamprophyric dark borders with aphanitic
173 textures, progressively grading into porphyritic microgranites with some mafic enclaves in the
174 cores. In general, bulk compositional gradients that are interpreted to reflect different mixing
175 ratios, are aligned parallel to the dyke walls. The dyke illustrated in Fig. 2a varies from
176 lamprophyre to granite (SiO_2 57.20 to 67.81 wt%) from left to right in the image. An increase in
177 number and size of K-feldspar phenocrysts, from lamprophyric to granitic end-members, was
178 observed in the field (Fig. 2a). In other outcrops, lamprophyric and granitic components form
179 independent dykes that were observed to intrude one another (Fig. 2b).

180 The Aigoual granite proper contains many typically rounded, 1 to 3dm in diameter (Fig. 2c),
181 lamprophyric enclaves with partly resorbed phenocrysts of K-feldspar. In combination, the
182 compositionally gradational dykes and the existence of lamprophyric enclaves in the granites that
183 include crystals likely to have been captured from the granite magma. In combination, this

184 mineral textural and field evidence demonstrates the coeval nature of the granitic and
185 lamprophyric magmas.

186 The sample list and the coordinates of the sample points are presented in Supplementary
187 Data 2.

188

189 **4.2 Petrography and mineral chemistry**

190 **4.2.1 The composite dykes**

191 The lamprophyres have panidiomorphic–porphyritic textures with phenocrysts of euhedral
192 biotite (~ 25–30 vol%), plagioclase (~ 20%), quartz (~ 5%) and K-feldspar (~ 5%), set in a fine
193 matrix composed of quartz, alkali-feldspar, plagioclase, biotite, ± amphibole, ± carbonate (Fig.
194 3). Olivine and pyroxene were not observed but pseudomorphs indicate the former presence of
195 one or more ferromagnesian phases in some lamprophyric samples, as also reported by Sabourdy
196 (1975) from the same outcrops. Accessory minerals are apatite, zircon, ilmenite, ± allanite, ±
197 rutile.

198 A characteristic of the lamprophyres is the presence of carbonate in three different textural
199 forms (Fig. 3): 1) globular, ocellar; 2) interstitial to the silicate minerals, and 3), intergrown with
200 biotite/phlogopite. The carbonate-mica intergrowths appear to pseudomorph a prior blocky
201 mineral.

202 Carbonate (dolomite) was observed mantling corroded quartz that varies from 0.4 to 0.6 mm
203 in size (Supplementary Figure 1a). Ocellar carbonate aggregates vary from 0.2–1.6 mm in
204 diameter; the ocelli also contain very fine-grained alkali-feldspar (Fig. 3). The matrix
205 surrounding the ocelli is aphanitic and composed of quartz, plagioclase, biotite (partially altered
206 to chlorite in some places). Matrix plagioclase laths display an orientation tangential to the

207 margins of the carbonate globules indicating that the ocelli structures formed part of the magma
208 (Supplementary Figure 1b-c). Interstitial carbonate is present among euhedral quartz, plagioclase
209 and biotite (Supplementary Figure 1d). The interstitial character of the carbonate relative to
210 euhedral quartz, biotite and plagioclase crystals is shown in the Si X-ray map of Supplementary
211 Figure 1e. Plagioclase crystals associated with the carbonate display normal magmatic zoning
212 with Na content increasing towards the rim (Ab₅₆ to Ab₇₅) (Supplementary Figure 1f). The
213 carbonate associated with biotite/phlogopite (interpreted as pseudomorphing a high-temperature
214 ferromagnesian phase) was observed in both granitic and lamprophyric samples (Supplementary
215 Figure 1g, h). In these textures, calcite or dolomite cores are rimmed by blocky phlogopite and
216 both replace a former phenocryst with rectangular habit, possibly pyroxene.

217 Granitic samples from composite dykes display seriate porphyritic textures with extreme
218 average crystal size variations between the phenocrysts and the matrix. In general, the
219 phenocrysts (up to 5 cm) of plagioclase (~ 30%), K-feldspar (~ 30%), quartz (~ 20%) biotite (~
220 5–10%) and ± clinopyroxene (~ 3%) are set in a fine- to medium-grained matrix composed of
221 quartz, K-feldspar ± plagioclase. Accessories are apatite, zircon, ± allanite, ± titanite ± ilmenite.
222 In some samples, phenocrysts of plagioclase display oscillatory zoning (Fig. 3j) and some K-
223 feldspar phenocrysts have corroded margins (Fig. 3l).

224 **4.2.2 Aigoual granite**

225 The Aigoual granite proper consists predominantly of strongly porphyritic, coarse-grained
226 granite with phenocrysts of K-feldspar reaching 5–8 cm. The granite contains phenocrysts of K-
227 feldspar (~ 30%), plagioclase (~ 20%) and quartz (~ 15%) in a coarse matrix of quartz,
228 plagioclase and K-feldspar (~20%), mica (~7%) and clinopyroxene (~5%).

229 **4.2.4 Mineral Chemistry**

230 Mineral chemistry data is presented in Supplementary Data 4, Table S1. Euhedral crystals of
231 mica from the lamprophyres have Mg# varying from 0.47 to 0.80, corresponding to a transition
232 between biotite (Mg# < 0.65) and phlogopite (Mg# > 0.65) (Fig. 4a). Biotite/phlogopite has low
233 to moderate Cr₂O₃ content (0–0.05 apfu), high Ti (0.37–0.97 apfu) and in some samples displays
234 zoning in BSE images with dark grey cores and light gray discordant rims (e.g. Fig. 3g). The
235 cores have lower Ti (0.42–0.59 apfu) than the rims (0.78–0.97 apfu) and biotite in the matrix
236 shows similar compositions to the rims with high Ti (0.69–0.94 apfu). In the FeO^t vs. Al₂O₃ and
237 TiO₂ vs. Al₂O₃ mica discrimination diagrams, the biotite/phlogopite follows a minette trend,
238 typical of calc-alkaline lamprophyres (Fig. 4b–c; Rock, 1991). Dark micas in the granites from
239 the composite dykes are generally less magnesian than those in the lamprophyres, with an
240 average of Mg# ~ 0.52 and are thus classified as biotite. A few analyses with Mg# > 0.65 fall in
241 the phlogopite field (Fig. 4a). Biotite in the Aigoual granite proper has Mg# ~ 0.52.

242 Feldspar compositions are illustrated in Supplementary Figure 2a. Feldspar in the
243 lamprophyres is predominantly plagioclase, which ranges in composition from An₁₀ to An₆₀.
244 Alkali feldspar is a relatively minor component in these rocks and ranges in composition from
245 Or₅ to Or₄₀. Feldspar in the Aigoual granite consist of plagioclase with a narrow compositional
246 range clustered around An₃₀ and K- feldspar which varies in composition from Or₈₀ and 92. In
247 the granite dykes the plagioclase compositional range is similar to that exhibited by the
248 lamprophyric dykes. However, at low An contents significantly more K₂O (typically up to Or₃₀) is
249 present in some feldspar crystals with significant Ca, which plot in the fields of oligoclase and
250 anorthoclase. The more common alkali-feldspar compositions in these dykes vary from Or₆₀ to
251 Or₁₀₀.

252 All pyroxene in the Aigoual granite and in the granitic dykes is augite and clusters in a
253 narrow range of compositions characterized by Ca contents in the range of Di50 to 60 and with
254 Mg#s between 60 and 75 (Supplementary Fig. 2b). All amphibole from the lamprophyres is
255 actinolite with high Si contents (7.53–7.90 apfu) and Mg# (0.62–0.88), as is typical of secondary
256 amphiboles in lamprophyres (Supplementary Fig. 2c and d).

257 **4.3 Zircon U–Pb ages and Lu–Hf isotopes**

258 In this work, the geochronological data are from individual composite dykes, where
259 lamprophyre and granite occur in the same outcrop with gradational transition from one to the
260 other. The zircons are euhedral to subhedral, 50–450 μm across. The crystals have prismatic,
261 subrounded prismatic, pyramidal to bipyramidal tips and aspect ratios from 4:1 to 1:1. Under CL,
262 they exhibit strong to weak oscillatory zoning, diffuse zoning or no luminescence at all
263 (Supplementary Figure. 3). Many grains show a core–rim texture with diffusely or oscillatory
264 zoned core surrounded by a dark rim. Inherited cores display bright, diffuse to oscillatory,
265 truncated zoning with dark rims.

266 The $^{206}\text{Pb}/^{238}\text{U}$ dates show a large range of concordant ages, scattering from 340 to 300 Ma
267 (Fig. 5). However, this large dispersion displays no correlation with position of the analysis spot
268 in different zircon crystals. Older ages, between 320 and 340 Ma, were obtained from both dark
269 rims and CL-dark grain interiors with unusually high U contents (> 1000 ppm). Due to the high U
270 content, these zircons might no longer be crystalline (Ewing et al., 2003). The absence of a
271 correlation between age and Hf isotopic composition indicates that the age dispersion has no
272 geological significance. The older ages, between 320 and 340 Ma, are likely to be artefacts
273 potentially caused by inaccurate correction for downhole fractionation in these high U crystals
274 and strongly radiation damaged zircon. Accordingly, the Concordia ages were calculated using

275 data from zircons that were more likely to retain coherent crystallinity ($U < 1000$ ppm) and KDE
276 variation, which displays a peak at 315 - 310–Ma for most of the samples (Fig. 6).

277 Eleven concordant analyses from lamprophyres and granites of dyke 5 yielded a Concordia
278 date of $^{206}\text{Pb}/^{238}\text{U} = 313 \pm 3$ Ma (MSWD conc. + equiv. = 1.7) (Fig. 5a). Inherited grains were
279 largely found in the granite component and yielded $^{206}\text{Pb}/^{238}\text{U}$ ages of = 537 ± 10 , 541 ± 14 , 545
280 ± 9 , 649 ± 10 and 673 ± 7 Ma. Younger ages of 296 ± 4 and 291 ± 5 Ma are interpreted as
281 reflecting Pb loss. Magmatic zircons from dyke 5 have $\varepsilon_{311}^{\text{Hf}} = -1.6$ to -8.1 (± 1.2), $T_{\text{DM}} = 1.7$ to
282 1.3 Ga] (Fig. 6). The inherited zircons yielded $\varepsilon_{311}^{\text{Hf}}$ of $+6.8$ to $+0.9$.

283 Thirty-three concordant analyses from lamprophyres and granites of dyke 10 yielded a
284 Concordia date of $^{206}\text{Pb}/^{238}\text{U} = 311 \pm 3$ Ma (MSWD conc. + equiv. = 1.5) (Fig. 5b). A large
285 population of inherited grains found in the granite component yielded several date groups from
286 Paleoproterozoic (2490-2040 to 2480 Ma), through early Neoproterozoic (920–1000 Ma) to late
287 Neoproterozoic and early Paleozoic (450–730 Ma). Zircons from dyke 10 have $\varepsilon_{311}^{\text{Hf}} = +3.6$ to –
288 7.7 (± 1.5) (Fig. 6), with $T_{\text{DM}} = 1.7$ to 1.1 Ga. Inherited grains with $^{206}\text{Pb}/^{238}\text{U}$ ages of 450–600
289 Ma gave $\varepsilon_{311}^{\text{Hf}}$ of $+2.5$ to 0.0, whereas inherited grains (> 600 Ma) yielded $\varepsilon_{311}^{\text{Hf}}$ of $+2.9$ –13.5.

290 Samples 12A and 12B represent the eastern margin of Aigoual granite with microgranite and
291 monzonite occurring as part of the body of the granite, in contrast to all the other granite samples,
292 which form part of the composite dykes. Isotopic data from 10 concordant analyses display a
293 Concordia date of 312 ± 3 Ma (MSWD conc. + equiv. = 1.7) (Fig. 8c). No inherited grains were
294 observed. Zircon grains from margin of Aigoual granite (locality 12) have $\varepsilon_{311}^{\text{Hf}} = 0.4$ to -6.8 (\pm
295 1.4), $T_{\text{DM}} = 1.6$ to 1.2–Ga] (Fig. 6).

296 Zircons from the Aigoual granite proper yield a Concordia date of 313 ± 3 Ma (MSWD conc.
297 + equiv. = 1.3; n = 7) (Fig. 8d) with $\varepsilon_{311}^{Hf} = -5.6$ to -3.3 (± 1.1), $T_{DM} = 1.4$ – 1.5 Ga] (Fig. 6).

298 **4.4 Whole-rock major- and trace-element chemistry**

299 In the TAS system of Le Maitre et al. (2002), the rocks of this study would be classified as
300 granites, quartz monzonites and monzonites, but of course such alkali rich compositions are not
301 amenable to classification according composition discrimination diagrams. The compositions of
302 the three Aigoual granite samples are very similar and fall consistently between the compositions
303 of the granitic and lamprophyric portions of the composite dykes on plots of all elements vs. SiO_2
304 (Fig. 7). The dykes show a wide range in SiO_2 , varying from 53.16 to 72.57 wt%, recalculated on
305 a volatile-free basis. Lamprophyric and granitic samples are high-K (4.08–6.55 wt%) with
306 shoshonitic affinities (Fig. 8a) and high Al_2O_3 (13.63–17.18 wt%). In the A/NK vs. A/CNK
307 [molar $Al_2O_3/(Na_2O + K_2O)$ vs. $Al_2O_3/(CaO + Na_2O + K_2O)$] diagram of Shand (1943),
308 lamprophyric samples classify as metaluminous to slightly peraluminous with higher A/NK
309 values, whereas granitic samples are weakly to moderately peraluminous with some exceptional
310 cases having $A/CNK > 1.3$ (Fig. 8b).

311 A striking feature of the bulk rock chemistry of the entire sample set is that Cr, Ni and Ba are
312 high (30–573 ppm, 15–149 ppm and 404–2416 ppm, respectively), but variable and with only a
313 vague inverse correlation with SiO_2 . Mg# varies from 41 to 71 and along with TiO_2 and P_2O_5 is
314 one of the components which shows the best developed inverse correlation with SiO_2
315 concentration. For the incompatible elements, which commonly are positively correlated with
316 SiO_2 content, Na_2O demonstrates this behavior, despite a significant amount of scatter, for the
317 lamprophyres. However, for the granites, if there is any trend its flat. Similarly, over the whole

318 range of compositions K_2O shows no correlation with SiO_2 , with the majority of the rocks falling
319 between 5 and 6 wt% K_2O .

320 The lamprophyric–granitic rocks from composite dykes and the Aigoual granite are enriched
321 in large ion lithophile elements (LILE, e.g. Cs, Rb, Ba, Th, U; Fig. 9a) and Pb relative to high
322 field strength elements (HFSE, e.g. Nb, Ta, P, Zr and Ti). The lamprophyres show negative Nb,
323 Ta and Ti anomalies, but Ta anomalies are not observed in the granites. In general, the
324 lamprophyres are richer in REE and mantle-incompatible elements than the granites, as is
325 commonly observed in these types of rocks (e.g. Couzinié et al., 2016; Fowler et al., 1996).

326 In the primitive mantle -normalized (Sun and McDonough 1989) REE multielement plot
327 (Fig. 9b), lamprophyres and granites from composite dykes as well as the Aigoual granite
328 samples show similar patterns. They are enriched in LREE ($La_N = 24\text{--}147$) and depleted in
329 HREE ($La_N/Yb_N = 3.34\text{--}22.81$). In general, lamprophyres show smaller negative Eu anomalies
330 ($Eu/Eu^* = 0.60\text{--}0.85$) than granites ($Eu/Eu^* = 0.38\text{--}0.86$).

331

332 4.5 Whole-rock Sr–Nd–Hf isotopes

333 Lamprophyres and granites from the composite dykes show little variation in Sr, Nd and Hf
334 isotope ratios. Initial $^{87}Sr/^{86}Sr$ ratios of lamprophyres and granites overlap and vary from 0.7080
335 to 0.7106. Similarly, the Nd isotope ratios of the lamprophyres and granites overlap ($\epsilon_{311}^{Nd} = -3.60$
336 to -5.39 and $\epsilon_{311}^{Nd} = -4.62$ to -5.39 respectively). The Hf isotope ratios for both rock types also
337 overlap, with lamprophyres varying from $\epsilon_{311}^{Hf} = -0.26$ to -4.92 and the granites varying from
338 $\epsilon_{311}^{Hf} = -2.31$ to -3.47 (Fig. 14) (Table S2 from Supplementary Data 3). On average,
339 lamprophyres have slightly more radiogenic Nd and Hf than granites ($\epsilon_{311}^{Nd} = -4.40$ and -5.13 , and

340 $\epsilon_{311}^{Hf} = -2.81$ and -3.20 , respectively). The Aigoual granite has an $(^{87}\text{Sr}/^{86}\text{Sr})_i$ of 0.70941 , $\epsilon_{311}^{Nd} = -$
341 5.43 and $\epsilon_{311}^{Hf} = -3.71$.

342 4.6 Quartz, and K-feldspar O isotopes

343 The $\delta^{18}\text{O}$ values of quartz and K-feldspar ($\delta^{18}\text{O}_{\text{Qtz}}$ and $\delta^{18}\text{O}_{\text{Fsp}}$ respectively) from composite
344 dykes are listed in Table 3 from Supplementary Data 4 and illustrated in Fig. 10. The $\delta^{18}\text{O}_{\text{Qtz}}$
345 values vary from 10.76 to 11.21‰ ($n = 9$; average = 10.97‰) and, despite their narrow range,
346 show a positive correlation with whole-rock SiO_2 ($r^2 = 0.72$) (Fig. 10a) and a negative one with
347 $\text{Cr} + \text{Ni}$ ($r^2 = -0.66$) (Fig. 10b). The $\delta^{18}\text{O}_{\text{Fsp}}$ vary from 6.4 to 13‰ ($n = 9$; average = 9.4‰) and
348 are thus much more variable than for quartz. There is no correlation between $\delta^{18}\text{O}_{\text{Fsp}}$ and $\delta^{18}\text{O}_{\text{Qtz}}$
349 (Fig. 10c). The $\delta^{18}\text{O}_{\text{Qtz}}$ values display no correlation with whole-rock $(^{87}\text{Sr}/^{86}\text{Sr})_i$ ($r^2 = 0.13$) (Fig.
350 10d), ϵ^{Nd} ($r^2 = 0.17$; not shown) and ϵ^{Hf} ($r^2 = 0.23$; not shown). The three lamprophyre samples
351 plot between the diffusion closure and crystallization isotherms on a δ - δ plot (Fig. 10c). The
352 remaining samples show a wide range in feldspar $\delta^{18}\text{O}$ values that are both higher and lower than
353 $\delta^{18}\text{O}_{\text{Qtz}}$. The wide range in $\delta^{18}\text{O}_{\text{Fsp}}$ values is consistent with interaction with fluids at a range of
354 temperatures.

355 5. Discussion

356 5.1 Timing of magma emplacement

357 U-Pb analyses in zircon from lamprophyres and granites from dykes 5 and 10 yielded
358 $^{206}\text{Pb}/^{238}\text{U}$ ages of 313 ± 3 Ma and 311 ± 3 Ma, respectively. Locality 12, in the Aigoual granite
359 at its eastern border, gave a $^{206}\text{Pb}/^{238}\text{U}$ age of 312 ± 3 Ma and the granite from the center of the
360 body has a $^{206}\text{Pb}/^{238}\text{U}$ age of 313 ± 3 Ma. Therefore, all samples of dykes and the Aigoual granite
361 show the same crystallization ages, within uncertainties. These are statistically indistinguishable

362 from the U–Pb age of 304 ± 12 Ma, reported by Brichau et al. (2008), from the southern portion
363 of the ASGL Pluton.

364 The alignment of the high-K–Mg rocks in FMC in along NE–SW trends (Fig. 1b), that are
365 also parallel to major lithological boundaries, suggests that emplacement followed zones of
366 weakness in the crust. Emplacement was possibly also facilitated by NW–SE directed
367 extensional associated with crustal thinning of the Variscan Belt, during the syn- to post-
368 collisional period (Faure et al., 2009; Talbot et al., 2005). Dykes from northern Aigoual granite
369 display an AMS fabric indicative of NE–SW stretching. However, this fabric was controlled by
370 magmatic flow rather than regional extension, suggesting a southwestward source feeding the
371 magmatic system (Talbot et al., 2005).

372 The lack of inherited or xenocrystic zircon grains in lamprophyres can be explained by
373 zircon under-saturation due to the high temperature and mafic composition of these magmas
374 (Laurent et al., 2017). Inherited grains in granitic samples display a wide spread of
375 Paleoproterozoic to late-Ordovician ages (Fig. 8b) and \mathcal{E}^{Hf} signatures from -13 to +6.8, consistent
376 with age distribution and isotopic signatures from Variscan ortho- and paragneisses in the FMC
377 (Chelle-Michou et al., 2017).

378 Dating of granites that were proposed to be crustally derived and mantle-derived mafic
379 magmas in the E-FMC (Couzinié et al., 2014; Laurent et al., 2017) showed that the two were
380 related in space and time. They marked a protracted (40 Myr) magmatic period, in which granites
381 (337 ± 1 to 299 ± 2 Ma), emplaced coevally with vaugnerites (336 ± 2 and 299 ± 1 Ma – Laurent
382 et al., 2017). The age distribution displays migration of magmatism toward the south, where the
383 younger ages of 307 and 298 Ma were reported from the KCG plutons in the Cévennes area
384 (Couzinié et al., 2014; Laurent et al., 2017). These are substantially younger than the ages

385 reported in the current work. Late Carboniferous southward delamination of the lithospheric
386 mantle and asthenosphere upwelling may explain the long-lived lithospheric-scale thermal
387 anomaly that could have produced melts of both crustal and mantle sources (Laurent et al., 2017).

388 **5.2 Origin of the lamprophyres and granites**

389 **5.2.1 Origin of lamprophyres**

390 The lamprophyres have moderate to high compatible element contents, e.g. Cr (74–572 ppm)
391 and Ni (23–148 ppm), coupled with relatively high Mg# (up to ~ 70), consistent with a mantle
392 source for at least this component of the composite system. However, the lamprophyres also
393 display elevated concentrations in LILE (K₂O, Sr, Rb, Ba.), LREE and depletion in HFSE such
394 Nb, Ta and Ti, in relation to the primitive mantle. Whole-rock Sr–Nd–Hf and in-zircon Hf
395 isotopes display crust-like signatures with subchondritic ϵ^{Nd} and ϵ^{Hf} values. Also the $\delta^{18}\text{O}$ values
396 for the lamprophyres are on average 5‰ higher than the expected value for depleted mantle
397 (5.2‰, Eiler, 2001). The crust-like signatures of the lamprophyres can result from crustal
398 assimilation accompanying fractional crystallization of a primitive basaltic liquid (AFC process)
399 or partial melting of an enriched mantle source, that was metasomatised by interaction with
400 crustal material during the subduction stage that preceded the continental collision (e.g., Becker
401 et al., 1999; Janoušek and Holub, 2007; Lexa et al., 2011; Schulmann et al., 2014; Couzinié et al.,
402 2016).

403 The absence of correlations between Mg#, Cr, Ba, Ni and $^{87}\text{Sr}/^{86}\text{Sr}_i$, ϵ^{Nd} , ϵ^{Hf} suggests that the
404 isotopic enrichment did not result from assimilation of continental crust by a primitive mafic end-
405 member magma. Moreover, the most mafic members display crust-like isotopic signatures,
406 similar to the intermediate members and even to the granites from composite dykes. An
407 alternative interpretation that the lamprophyres are the result of assimilation of Fe-Mg rich crustal

408 rocks of suitable composition, e.g. the graphitic metapelites of the Parautochthonous Unit (Faure
409 et al., 2001), does not seem viable, as such sediments are unlikely to be sufficiently rich in Cr and
410 Ni. Furthermore, lamprophyres and other high-K–Mg rocks referred to as vaugnerites are
411 ubiquitous in the FMC, where they have intruded several different rock types of the nappe pile
412 (Montel & Weisbrod, 1986; Sabatier, 1991; Couzinié et al., 2016). These other high-K–Mg rocks
413 have chemical and isotopic signatures similar to the lamprophyres reported in this work, pointing
414 to the enriched mantle as the most likely source (Couzinié et al., 2016).

415 Magmas derived from mantle sources modified by subducted components usually differ from
416 MORBs and OIBs by enrichment in Th and Nb compared to Yb (Pearce, 2008) (Fig. 11a). As is
417 the case for modern sediments (GLObal Subducted Sediments – GLOSS; Plank and Langmuir,
418 1998), the studied calc-alkaline lamprophyres show jagged spidergrams marked by low Nb–Ta–
419 Ti concentrations and elevated Th, U and Pb. All these geochemical features coupled with low
420 Nb/La ratios, enrichment in LILE, LREE and depletion in HFSE in primitive-mantle-normalized
421 multi-element diagrams suggest that the enriched lithospheric mantle was the most likely source
422 for the lamprophyres. Likewise, the relatively low $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios (0.7080– 0.7106) are
423 consistent with recent mantle enrichment, probably during the Silurian–Devonian subduction.

424 The presence of phlogopite identified in the lamprophyres and also the elevated Rb/Sr and
425 low Ba/Rb ratios strongly indicate that phlogopite was the dominant hydrous phase in the mantle
426 source (Fig. 11b) (e.g., Butvina et al., 2021; Condamine et al., 2022). Moreover, high-K₂O
427 magmas with shoshonitic affinities are a typical product of partial melting of phlogopite-bearing
428 lithospheric mantle that has been metasomatised by fluids or melts derived from the continental
429 crust or marine sediments during subduction (Conceição and Green, 2004; Förster et al., 2019;
430 Prelević et al., 2012). (Fig. 11c).

431 The $\delta^{18}\text{O}_{\text{Qtz}}$ values can be used as a proxy for the original magma $\delta^{18}\text{O}$ values (e.g., Harris et
432 al., 2005). The parental magma would be about 1‰ lower than $\delta^{18}\text{O}_{\text{Qtz}}$ values (e.g. Fourie and
433 Harris, 2011). The calc-alkaline lamprophyres display high $\delta^{18}\text{O}_{\text{Qtz}}$ values ($\delta^{18}\text{O} = 10.76$ to
434 10.80), suggesting an average $\delta^{18}\text{O}$ value of $\sim 10\text{‰}$ for the original magma. This is close to the
435 boundary between I- and S-type granites suggested by Harris et al. (1997) (Fig. 12). The $\delta^{18}\text{O}$
436 values expected for an uncontaminated mantle-derived magma is about 5.7‰ (Eiler, 2001) and
437 fractional crystallization from a primitive magma can at most change $\delta^{18}\text{O}$ values by $\sim 1.0\text{‰}$ for a
438 20% increase in SiO_2 content (Bindeman, 2008; Bucholz et al., 2017). Either a very high $\delta^{18}\text{O}$
439 contaminant or a very high proportion of crustal assimilation would have been required to raise
440 the $\delta^{18}\text{O}$ value from 5.7 to 10‰. Simple mixing calculations between mantle and average of
441 upper continental crust ($\delta^{18}\text{O} = 15\text{‰}$) show that $\sim 50\%$ of crustal assimilation would be necessary
442 to raise $\delta^{18}\text{O}$ values above 10‰ (Fig. 13). This high degree of crustal assimilation is inconsistent
443 with the high compatible element contents and the relatively low $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios of the
444 lamprophyres which point to little or no crustal assimilation being involved in the origin of
445 lamprophyres. Therefore, we suggest that the high $\delta^{18}\text{O}$ values in lamprophyres result from
446 contamination with high $\delta^{18}\text{O}$ material already in the mantle source.

447 High $\delta^{18}\text{O}$ values are common in sedimentary rocks. Oceanic carbonate-bearing sediments
448 have $\delta^{18}\text{O}$ values of 25 to 32‰ and pelagic clays from 15 to 25‰ (Eiler, 2001). Values from the
449 silicate upper continental crust are around 10 and 20‰ (Taylor and Shepard, 1986). Extremely
450 high $\delta^{18}\text{O}$ values have been reported in peraluminous granite in ophiolite as a product of the
451 melting of subducted pelagic sediments, leading to magma $\delta^{18}\text{O}$ values above 20‰ (Spencer et
452 al., 2017) (Fig. 12). In this work, the high $\delta^{18}\text{O}$ values from lamprophyres ($> 10\text{‰}$) indicate that

453 the crustal component that metasomatised the mantle source was dominated by sedimentary
454 material.

455 To quantify the nature of the mantle source, mixing calculations using Sr, Nd and O isotopic
456 systems were applied. Simple mixing calculations between mantle and potential contaminants
457 indicate that ~ 20% of oceanic carbonate-bearing sediment ($\delta^{18}\text{O} = 28\text{‰}$) or ~ 35% of pelagic
458 clay ($\delta^{18}\text{O} = 20\text{‰}$) would be necessary to increase $\delta^{18}\text{O}$ values to $> 10\text{‰}$ (Fig. 13).

459 Following the same approach, simple magma mixing calculations using $^{87}\text{Sr}/^{86}\text{Sr}_i$ and ϵ^{Nd}
460 from the Variscan mantle and sedimentary material were applied. A proxy for the original,
461 subduction-modified depleted Variscan mantle can be derived from Devonian arc-related
462 magmatism and ophiolites of the Variscan Belt. Arc magmatism, resulting from fluid-fluxed
463 melting of the mantle wedge should have ϵ^{Nd} close to the canonical depleted mantle values but
464 $^{87}\text{Sr}/^{86}\text{Sr}$ offset to higher values (e.g., Bouilhol et al., 2013). The Devonian rocks of Massif
465 Central have ϵ^{Nd} up to +7 (Pin and Paquette, 2002). They also have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that are
466 invariably higher than the expected value of 0.703 on the “mantle array” for rocks with $\epsilon^{\text{Nd}} = +7$,
467 as expected for arc magmas and likely reflecting some continental crustal component.
468 Consequently, we estimate the Variscan mantle composition at that time to be $\epsilon^{\text{Nd}} = +7$, $^{87}\text{Sr}/^{86}\text{Sr}$
469 = 0.703.

470 In the case of FMC, the pre-Variscan metasediments are compositionally close to sediments
471 from the former northern Gondwana margin (Chelle-Michou et al., 2017; Couzinié et al., 2017).
472 We modelled binary mixing with several end-members, covering the range of Variscan
473 metasediments (Solgadi et al., 2007; Williamson et al., 1992) and GLOSS (Plank and Langmuir,
474 1998). Mixing calculations indicate that the amount of sedimentary material mixed with mantle

475 required to model the crustal isotopic signatures in Sr–Nd isotopic systems is 25–40% (Fig. 14a),
476 depending on the composition of the sediment end-member.

477 Lu–Hf and Sm–Nd isotopic systems display similar behavior during most magmatic
478 processes and the composition of mantle and crust are correlated, evolving in a “mantle array”
479 trend (Chauvel et al., 2008). Subchondritic whole-rock ϵ^{Nd} and ϵ^{Hf} imply mantle enrichment with
480 crustal materials.

481 The textural evidence from the lamprophyres and granites argues for the involvement of a
482 carbonate component in the magmas and the existence of an immiscible carbonate liquid in the
483 lamprophyres during crystallization. Ocellar carbonates are commonly described as the products of
484 immiscibility between silicate and carbonate liquids in potassic, alkaline, mantle-derived igneous
485 rocks (e.g., Huang et al., 2002 and references therein). The presence of carbonates in
486 lamprophyre might indicate that the crustal component was a carbonate-bearing sediment. This
487 composition corroborates the high oxygen fugacity identified in the mantle beneath FMC, which
488 is attributed to the addition of volatiles, including CO₂, in the mantle during Variscan subduction,
489 triggering CO₂, H₂O and K metasomatism (Martin et al., 2017).

490 After inferring the probable nature of the contaminant, the partial melting mechanism that
491 generated the lamprophyres was investigated. Non-modal partial melting models of a calculated
492 source with 65–80% of DM and 20–35% of a contaminant (GLOSS composition) show that
493 melts produced by 10 to 20% partial melting match the trace-element composition of
494 lamprophyres and associated granites (Fig. 15a, b). This model resembles results of experimental
495 melting of metasomatised mantle by Condomine et al. (2022).

496 The high HREE contents are consistent with shallow melting of a spinel peridotite, i.e., in the
497 absence of garnet. Lamprophyres and granites display higher LREE contents than the modelling

498 predicts, suggesting a pelagic clay richer in Rb and Ba than the average global sediments used in
499 the calculation (Plank and Langmuir, 1998), or melts derived from sediments, more enriched in
500 LREE than GLOSS.

501 In summary, the evidence from this study indicates that the magmas that produced the
502 lamprophyric component of the composite dykes was produced by 10–20% melting of a
503 metasomatised mantle source in the spinel-lherzolite field. This CO₂-, H₂O-, K-enriched
504 metasomatised mantle was phlogopite-bearing and resulted from reaction with sediment-derived
505 melts or fluids that were probably introduced into the mantle during Silurian–Devonian
506 subduction.

507 **5.2.2 Origin of granites**

508 The studied dykes are composite and include lamprophyric and granitic components, which
509 display field evidence indicating that they coexisted. They are also characterized by very similar
510 geochemical features. Thus, the lamprophyres and granites of the composite dykes were
511 comagmatic and show a mixture of “crust-like” and “mantle-like” geochemical features. Three
512 main hypotheses or combinations thereof are plausible for the origin of the granites:

- 513 1. The lamprophyres and the granites are related by differentiation processes, such as
514 fractional crystallization, although remelting of underplated lamprophyres to produce the
515 granites would result in broadly the same patterns;
- 516 2. The lamprophyres and the granites represent distinct magmas, an enriched-mantle derived
517 magma for the lamprophyres and a crust-derived magma for the granites, that mingled
518 and mixed;
- 519 3. The granite and the lamprophyres are both direct products of melting of the enriched
520 mantle source but reflect different degrees of partial melting. This could be achieved by

521 different source compositions, different temperatures of partial melting, different volatile
522 contents (more CO₂ in the lamprophyre source), or some combination.

523 We briefly discuss each hypothesis below.

524 The granites in the dykes share many geochemical characteristics with the lamprophyres,
525 including nearly identical isotopic ratios and trace-element concentrations. However, the granites
526 and the lamprophyres are not connected by differentiation trends, as illustrated by the fact that the
527 lamprophyres are richer in most incompatible elements than the granites. This enrichment is
528 inconsistent with a simple model of fractional crystallization of mafic into felsic magmas and is
529 also inconsistent with the granites being the product of partial melting of the lamprophyres.

530 Major elements such as K₂O, Na₂O and Al₂O₃ are relatively constant with increasing SiO₂
531 content, implying no significant fractionation of K-feldspar, plagioclase and biotite in high-K
532 magmas (e.g., Wang et al., 2021) (Fig. 6a, b, c). The increase in volume and size of phenocrysts
533 of K-feldspar in the granitic component demonstrate that a considerable decrease in K₂O is
534 expected during fractional crystallisation. Furthermore, a fractionation process involving
535 plagioclase would display similar behavior for CaO and Sr on the binary plots against SiO₂ in
536 Figure 6 display distinct behavior from CaO and Sr against SiO₂ contents, demonstrating that
537 fractional crystallization from the lamprophyres is unlikely to generate the granites.

538 In general, fractional crystallization involving clinopyroxene would remove compatible
539 elements such as Cr and Ni. The high Cr (30–140 ppm; 80.9 on average and in cases over 100
540 ppm in granite with SiO₂ = 70 wt%) and Ni (14–40 ppm; 24.4 on average) content in the granites
541 are inconsistent with fractional crystallisation of clinopyroxene playing an important role in the
542 formation of granites. Moreover, no cumulate textures are observed in lamprophyres and no
543 pyroxenites are described in the area. Therefore, this fractionation model is unlikely.

544 Cogenetic granite may result from mixing between lamprophyric and crust-derived magmas
545 at crustal conditions. Petrological evidence favoring this model consists of the presence in ca.
546 10% of zircon xenocrysts in granites from composite dykes and the existence of corroded quartz
547 xenocrysts mantled by carbonate and amphibole in lamprophyres. Quartz xenocrysts rimmed by
548 amphibole and carbonate are interpreted as typical magma-mixing texture (e.g., Janoušek et al.,
549 2004). In this way, we consider that assimilation of a crustal component is involved in the
550 petrogenesis of the granites from the composite dykes.

551 Linear trends observed in many (but not all) binary chemical plots between lamprophyre and
552 granite might indicate magma mixing between these two end members. However, similar trends
553 are observed in fractional crystallization or partial melting processes (Janoušek et al., 2016).
554 Radiogenic isotopes are generally not fractionated during partial melting or fractional
555 crystallization/accumulation and, therefore, are the best tools for deciphering open-system
556 processes such as magma mixing (e.g., Sun et al., 2010).

557 Large variations in zircon ϵ^{Hf} in granitic rocks have been attributed to magma mixing
558 between crustal and mantle-derived magmas (Griffin et al., 2002; Storck et al., 2021; Yang et al.,
559 2007) or inheritance of heterogeneities in detrital zircon record from the sedimentary source of
560 anatectic granites (e.g., Villaros et al., 2012).

561 In the present study, zircon ϵ^{Hf} values from lamprophyres and granites vary from -7.8 to
562 $+0.4$ with more than 90% of the data falling between -5.0 and -1.6 , i.e. within less than 4 ϵ^{Hf}
563 units. This is a limited variation with respect the analytical uncertainty ($c. \pm 1.5$ on average). In
564 the case of mixing of magmas with typical mantle and crustal Hf isotopic compositions, a
565 substantially wider range in ϵ^{Hf} values is expected (up to 10 ϵ^{Hf} units). Also, crustal and mantle
566 signatures would be more prominent, i.e., mantle component migrating towards positive and

567 crustal component towards negative ϵ^{Hf} values (e.g., Storck et al., 2021). Likewise, magmatic
568 rims mantling an old inherited core display different Hf isotopic signatures than cores, which
569 indicates that the inheritance does not control the isotopic composition of the magmatic rim. The
570 ϵ^{Hf} values from lamprophyres and granites are similar and overlap, suggesting that mixing within
571 different and contrasting magmas did not play an important role.

572 In terms of oxygen isotopes, the lack of correlation between $\delta^{18}\text{O}$ values and $^{87}\text{Sr}/^{86}\text{Sr}_i$, ϵ^{Nd}
573 and ϵ^{Hf} renders the slight correlation between SiO_2 and Cr or Ni meaningless. Therefore, the
574 small difference in $\delta^{18}\text{O}$ values between lamprophyres and granites (less than 0.5‰) is consistent
575 with source-related variation or homogenization of the magma than mixing between two different
576 and contrasting magmas.

577 Although crustal melts are regionally widespread in FMC, the ASGL Pluton lacks obviously
578 crustally-derived granites, such as strongly peraluminous cordierite- or muscovite-bearing
579 granites that are ubiquitous elsewhere in the FMC. In addition, the lack of well-defined mixing
580 trends for major elements, trace elements and isotopes suggests that magma mixing was not a
581 dominant process. Therefore, we favor only local and limited assimilation.

582 The comagmatic nature of the lamprophyres and granites, as well as their very similar
583 isotopic signatures, indicates that a third hypothesis needs to be evaluated: That the granites and
584 lamprophyres were both produced by directly by partial melting of metasomatised mantle. Mantle
585 peridotite or lherzolite metasomatised by either subducted metasediment or melts derived from
586 such metasediments is widely accepted to be the source of post-collisional K-rich magmatism
587 (e.g. Förster et al., 2019, 2020, 2021; Mallik et al., 2015). More specifically, recent studies
588 reported that mixing between peridotite and sediment may produce rocks that are likely to be

589 quartz normative (Förster et al., 2020, 2021) with major- and trace-element patterns similar to
590 those in the Aigoual dykes.

591 The available data clearly rule out the possibility that the granites formed by fractional
592 crystallization of a lamprophyric parent magma. It is also very unlikely that the granites formed
593 predominantly from a crustally derived magma because the granites have typical SiO₂ contents,
594 yet are strongly enriched in mantelitic elements. They also have isotopic compositions that are
595 very similar to the lamprophyres. The ages of the inherited zircon crystals in the granites do
596 indicate some assimilation of Variscan basement, but the range of ϵ^{Hf} values (approximately -8 to
597 +4) for these zircons relative to ϵ^{Hf} in the magmatic zircons (90% between approximately -5 and
598 -2) clearly indicates that this Variscan basement was not a major source of magma to the granites.
599 The data are consistent with the granite magma forming via the partial melting of a mantle source
600 metasomatised in the manner defined, with a minor amount of assimilation of Variscan crust to
601 explain the inherited zircons. As the granite and lamprophyre magmas are isotopically very
602 similar at source, the major and trace element arrays defined by the granitic and lamprophyric
603 components may be shaped by mixing between these magmas that is invisible from an isotopic
604 perspective.

605 **5.2.3 Texture of the microgranites**

606 The preservation of the studied composite dykes and the fine-grained, quench textures in the
607 granites may be a direct consequence of the presence of carbonate in the system. The absence of
608 clinopyroxene in the lamprophyres and its presence in the granites may be explained by the
609 textural association of biotite/phlogopite with carbonate pseudomorphing a prior mafic mineral in
610 the lamprophyres. This reaction replacing clinopyroxene and/or olivine can be explained as a
611 response to increasing $a\text{CO}_2$ and $a\text{H}_2\text{O}$ in the lamprophyric magma during crystallization (Blank

612 and Brooker, 1994; Solomatova et al., 2020). As the presence of carbonate facilitates partial
613 melting of the mantle (e.g., Dasgupta et al., 2013), carbonated portions of the metasomatised
614 mantle source are inferred to have melted preferentially to produce the lamprophyres.

615 The CO₂ plays a fundamentally different role in the melting of mantle and crustal rocks and
616 its solubility is strongly dependent of pressure, melt composition, particularly SiO₂ content
617 (Blank and Brooker, 1994; Ni and Keppler, 2013). During melting of mafic and ultramafic rocks
618 at mantle pressures, the solubility of CO₂ increases with increasing pressure and decreasing SiO₂
619 contents (Brooker et al., 2001). CO₂ has a very low solubility in felsic melts at crustal pressures,
620 whilst H₂O solubility is high, consistent with the powerful effect of H₂O in depressing the granite
621 solidus (Blank and Brooker, 1994; Brooker et al., 2001). The consequence of CO₂ being present
622 in these systems is to raise the solidus temperature by lowering the water activity (Scaillet et al.,
623 1997). Therefore, CO₂-fluids derived from cooling mantle-derived magmas that intruded the crust
624 may limit the possibilities for complete mixing between the mafic and felsic magma at shallow
625 levels. Our study potentially records an example of this behavior, in which the mantle-derived
626 lamprophyre magmas quenched the granitic magmas with which they mingled, once carbonate
627 saturation of the mafic magma occurred.

628 **5.3 Implications for crustal growth**

629 In any magmatic episode, the ratio of new mantle-derived magma relative to that derived
630 from the partial melting of older crustal material, determines the amount of crustal growth
631 achieved. In the case of the FMC, a large proportion of the magmatic rocks (MPG, CPG and
632 Velay migmatites) were derived from melting of local crust. Therefore, these rocks did not
633 contribute to crustal growth and corresponded to pure crustal recycling (e.g. Moyen et al., 2017).
634 On the other hand, vaugnerite and the KCG granites contain a contribution of mantle-derived

635 material and thus, to some degree, represent the formation of new continental crust (Couzinié et
636 al., 2016; Moyen et al., 2017). Couzinié et al. (2016) has demonstrated that Hf–O isotopes in
637 zircon in this case do not identify additions of new mantle-derived material to the crust.
638 Therefore, the process of crustal growth may be underestimated throughout geological time, since
639 many studies were carried out relying on the zircon record (e.g., Belousova et al., 2010;
640 Hawkesworth and Kemp, 2006).

641 This study has illuminated the reasons for this, clearly mantle derived magmas such as the
642 vaugnerites in this study, are derived from mantle source rocks with sufficient degrees of
643 metasomatism by a subducted crustal component that conventional radiogenic isotopic systems,
644 such as Rb–Sr, Sm–Nd and Lu–Hf, as well as O isotope ratios, attain values consistent with their
645 derivation from the continental crust. The high $\delta^{18}\text{O}$ of clay-rich sediments and the low
646 abundance the relevant radiogenic precursor isotopes in the mantle, result in a relatively small
647 proportion of sediment-rich crustal material (< 30%) recycled into the mantle being sufficient to
648 control all incompatible elements contents and form the “crust-like” isotopic signature observed
649 in mafic post-collisional magmatism (Couzinié et al., 2016; Jacob et al., 2021). However, in
650 detail, the calculated mantle:crust ratio may be dependent on the element, or isotopic system
651 used, as well as the exact composition of deep crustal contaminant and the mechanism of
652 metasomatism (Couzinié et al., 2016; Janoušek et al., 2020; 2021).

653 Our results demonstrate that between 20 and 35% of the budget of Sr, Nd, Hf and other
654 incompatible elements comes from the crust and therefore does not constitute additions of new
655 material. However, at the same time, the budget of compatible elements such as Mg, Fe, Ni, Cr is
656 strongly controlled by the mantle component and represents addition of new material to the crust.
657 In the lamprophyres, the mafic character and dual geochemical signature displays mantle and

658 crustal contributions. Even though the mantle contribution is hidden from an isotopic point of
659 view, between 65 to 80% of material comes from the mantle. In the case of the Aigoual
660 composite dykes, this broad finding applies to modelled mantle crust ratios constrained on the
661 basis of radiogenic isotope ratios, trace element compositions and oxygen isotope ratios (Fig.
662 13, 14 and 15).

663 Although less evident, the same process occurs in the granites of composite dykes.
664 Assimilation of a minor amount of crustal material is likely be involved in their petrogenesis, but
665 the relatively high Cr, Ni, also FeO, and MgO contents in relation to purely crustally-derived
666 granites in the FMC reflects the mantle component involved in their genesis. The intimate
667 relation (physically and geochemically) between lamprophyre and granite described in this work
668 from the composite dykes at northern border of the Aigoual granite provides constraints on the
669 mantle contribution in post-collisional granitoids from the FMC. The fact that the chemistry of
670 the Aigoual granite lies at the mafic end of the trend defined by the granites from the composite
671 dykes, as well as the common occurrence of rounded lamprophyric enclaves in the granite,
672 suggests that it formed by mixing with approximately the same amount of lamprophyric magma as
673 the most mafic granites in the composite dykes.

674 High-K–Mg mafic rocks are low in volume in the FMC, forming ~ 10% outcrops, and
675 therefore their contribution to crustal growth may be relatively minor. However, the findings of
676 this study demonstrate a significant fraction of mantle material in the granites of the composite
677 dykes, in the Aigoual granite and, thus, likely also in other KCG granites in the FMC (ca. 25% of
678 outcrop area). Considering that between 65% and 80% of the material from lamprophyres and
679 KGC granites came from the mantle and both represent < 35% of granitoids outcrops in the
680 FMC, approximately 22 to 28% of the magmatic material represented by the KGC likely came

681 from the mantle and, despite the crustal-like incompatible element contents and isotopic
682 signatures, represents a substantial new addition to the crust.

683 **6. Conclusions**

684 In summary, lamprophyric-granitic composite dykes from the northern border of the Aigoual
685 granite (southeastern Massif Central, France) record interaction between the mantle and crust in
686 the source of the magmas. The intrusion of these magmas into the late/post-orogenic continental
687 crust represents significant long-term additions of new crust with excellent potential for
688 preservation. Their intrusion was coeval with the emplacement of Aigoual granite between 311
689 and 313 Ma, which shares their source but underwent more complete homogenization because
690 the magmas that formed the granite did not freeze in their conduits. Crystallization of the granitic
691 component of the composite dykes in the conduits was likely a consequence CO₂-saturation of
692 the lamprophyric component and consequent quenching of the granite portion. Lamprophyres
693 display crust-like signatures in their Sr, Nd, Hf, O isotopes and other incompatible elements, and
694 were the product of 10–20% of partial melting from metasomatised mantle. This mantle can be
695 modelled by mixing between 20 and 35% of sediments from northern Gondwana margin with 65
696 to 80% of depleted mantle. The melting processes occurred at shallow depths in the spinel
697 stability field. Granites have a mantle-derived component, evident from their high Cr, Ni, FeO
698 and MgO contents compared with typical anatectic granites in the FMC. In terms of crustal
699 growth, the results imply that between 22 and 28% of magmatic material in the FMC was mantle-
700 derived. This study demonstrates that, although hidden by crustal signatures in terms of
701 radiogenic isotopes and incompatible elements, the mantle contribution in post-collisional
702 magmatism can be significant. Considering their greater preservation potential throughout

703 geological time, post-collisional sites represent a meaningful contributor to long-term crustal
704 growth.

705

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714

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1042

1043 **Figure Captions**

1044 Figure 1. (a) General location of Variscan belt in Western–Central Europe. (b) Geological map of
1045 the eastern French Massif Central modified from Moyen et al. (2017). Post-collisional mafic
1046 magmatic rocks (vaugnerites) occurrences from Couzinié et al. (2016). (c) Geological map of
1047 Aigoual granite and composite dykes modified from Talbot et al. (2005). At the scale illustrated,
1048 it is not possible to differentiate the different lamprophyric components of the dykes from the
1049 granites.

1050

1051 Figure 2. Selected field photos illustrating key field relations (a) A section of a composite dyke
1052 with the field of view orientated perpendicular to the strike of the dyke. A typical gradational
1053 contact from lamprophyre (left) to granite (right) is illustrated, also showing the increasing
1054 proportion and size of phenocrysts of K-feldspars; (b) Field relationships between lamprophyre
1055 and granite showing that two different generations of lamprophyre dykes intrude a granite dyke.
1056 The contacts in this case are sharp with no evidence of mingling; (c) Aigoual granite with a

1057 typical rounded enclave of lamprophyre cut by a later thin aplitic dyke. The larger pale coloured
1058 inclusions in the enclave represent partially resorbed K-feldspar phenocrysts.

1059

1060 Figure 3. Photomicrographs and BSE images of lamprophyres and granites from composite
1061 dykes. (a–b) Euhedral–subhedral biotite in lamprophyre; (c) Interstitial dolomite in lamprophyre
1062 associated with euhedral quartz, plagioclase and biotite; (d) Calcite associated with biotite in
1063 granite; (e) Ocellar carbonate and microcrysts of K-feldspar in lamprophyre; (f) Plagioclase and
1064 biotite phenocryst in lamprophyre; (g–h) BSE images with compositional variation in
1065 biotite/phlogopite from lamprophyre; (i) Fine-grained granite from composite dykes; (j–l)
1066 Textural variation in granite from composite dykes with phenocrysts of zoned plagioclase, K-
1067 feldspar (j), quartz (k) and corroded megacryst of K-feldspar (l). Mineral abbreviations from
1068 Warr (2021).

1069

1070 Figure 4. (a) Al – Mg – Fe²⁺ ternary diagram showing compositional variations of biotite to
1071 phlogopite. Minettes and lamproites fields after Rock (1991). (b) Al₂O₃ vs. FeO^T and (c) Al₂O₃
1072 vs. TiO₂ discrimination diagram for micas after Mitchell and Bergman (1991). Abbreviations: Es
1073 = Eastonite, Ph = Phlogopite, Sd = Siderophyllite, and An = Annite.

1074

1075 Figure 5. Zircon U–Pb Tera–Wasserburg concordia diagrams for composite dykes and Aigoual
1076 granite. Data-point error ellipses are 2σ.

1077

1078 Figure 6. Diagram of ²⁰⁶Pb/ ²³⁸U ages (Ma) versus ε^{Hf}_(t) from zircons of composite dykes and
1079 Aigoual granite. ε^{Hf}_(t) lower and upper limit for Depleted Mantle (DM) are from Næraa et al.

1080 (2012) and Griffin et al. (2004). The KDE curve (light pink) portrays the $^{206}\text{Pb}/^{238}\text{U}$ age
1081 distribution used to calculate the Concordia ages.

1082

1083 Figure 7. (a) SiO_2 vs. K_2O classification diagram of Peccerillo and Taylor (1976); (b) A/CNK vs.
1084 A/NK diagram.

1085

1086 Figure 8. Whole-rock major- and trace-element compositions of lamprophyre and granite from
1087 composite dykes and Aigoual granite.

1088

1089 Figure 9. (a) Primitive mantle-normalized rare earth element (REE) patterns from Sun and
1090 McDonough (1989); (b) Multi-element normalized to Primitive mantle from Sun and
1091 McDonough (1989). GLOSS composition from Plank and Langmuir (1998); PCMM – post-
1092 collisional mafic magmatism from FMC of Couzinié et al. (2016); CAL – China – Calc-alkaline
1093 lamprophyre derived by partial melting of subducted sediments from Su et al. (2017).

1094

1095 Figure 10. (a) SiO_2 versus $\delta^{18}\text{O}$ quartz; (b) $\log(\text{Ni} + \text{Cr})$ versus $\delta^{18}\text{O}$ quartz; (c) $\delta^{18}\text{O}$ quartz
1096 versus $\delta^{18}\text{O}$ feldspar. Quartz–feldspar isotherms for 550 °C and 950 °C calculated using the
1097 calibrations of Chiba et al. (1989).

1098

1099 Figure 11. (a) Nb/Yb vs. Th/Yb diagram after Pearce (2008) suggesting the involvement of
1100 continental crust in the formation of the magmas that produced the composite dykes. (b) Ba/Rb
1101 vs. Rb/Sr diagram (after Furman and Graham, 1999) suggesting the presence of phlogopite in the
1102 mantle source region as a consequence of the metasomatism by a crustal component; (c) La/Yb

1103 vs. Nb/La diagram (Abdel and Nassar, 2004). Granites are also plotted (in pale colors), as they
1104 are comagmatic and may be derived from similar source rocks. Primitive mantle (PM) values
1105 from Hofmann (1988).

1106

1107 Figure 12. Histogram of $\delta^{18}\text{O}$ quartz analyses from this study. Mantle whole-rock O isotopic
1108 value ($\sim 5.7\text{‰}$) from Eiler (2001); I- and S-type granites from Harris et al. (1997) and
1109 peraluminous granite from Oman-United Arab Emirates (UAE) ophiolite from Spencer et al.
1110 (2017).

1111

1112 Figure 13. (a) Results of mixing models between mantle ($\delta^{18}\text{O} = 5.7\text{‰}$ and $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.703$)
1113 and three different contaminants all characterised by $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}} = 0.715$ (GLOSS, Plank and
1114 Langmuir 1998) but having different $\delta^{18}\text{O}$ values: carbonate-bearing sediments (28‰), pelagic
1115 clays (20‰) and average continental crust (15‰) (Eiler, 2001).

1116

1117 Figure 14. (a) $\epsilon^{\text{Nd}}_{\text{initial}}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}_{\text{initial}}$ plot for composite dykes and Aigoual granite. Initial ratios
1118 calculated at 311 Ma. Lithospheric mantle melting field from Davis and Blanckenburg (1995);
1119 Vaugnerites French Massif Central (FMC) from Moyen et al. (2017); Variscan lamprophyres
1120 from Soder and Romer (2018). The mixing calculations used the Variscan mantle composition
1121 from Pin and Paquette (1997), GLOSS from Plank and Langmuir (1998) and FMC paragneiss
1122 compositions from Moyen et al. (2017); (b) $\epsilon^{\text{Hf}}_{\text{initial}}$ vs. $\epsilon^{\text{Nd}}_{\text{initial}}$ plot for composite dykes and
1123 Aigoual granite. Data from MORB and OIB are taken from Stracke (2012) and mantle array
1124 regression line from Chauvel et al. (2008). Global kimberlites field from Tappe et al. (2018). DM
1125 = depleted mantle; OIB = ocean island basalt.

1126

1127 Figure 15. (a) La/Yb vs. Yb plot with batch melting models for phlogopite-bearing spinel and
1128 garnet peridotite. The calculated sources correspond to Depleted Mantle (Workam and Hart
1129 2005) mixed with 20% to 30% GLOSS (Plank and Langmuir 1998). Modal proportions in the
1130 residue are recalculated from Condamine and Médard (2014) for spinel peridotite (olivine 0.69,
1131 orthopyroxene 0.15, clinopyroxene 0.08, phlogopite 0.07, spinel 0.01) and Thibault et al. (1992)
1132 for garnet peridotite (olivine 0.25, orthopyroxene 0.50, clinopyroxene 0.15, phlogopite 0.03,
1133 garnet 0.07). Partition coefficients for mafic and intermediate rocks from Laurent (2012). (b)
1134 Multielement patterns of lamprophyres and granites and calculated compositions generated by 5,
1135 10 and 20% of partial melting of a phlogopite-bearing spinel peridotite source. Black and gray
1136 continuous lines represent partial melting of a calculated source with DM + 20% GLOSS and
1137 blue dashed lines of DM + 35% GLOSS. Partition coefficients from mafic and intermediate rocks
1138 from Laurent (2012). The melt composition from calculated source replicates the multi-element
1139 patterns exhibited by lamprophyres and granites well, except for the LREE.

1140

1141 **Supplementary Material**

1142 Supplementary Figure 1. BSE images and X-ray maps of three different magmatic textures of
1143 carbonate.

1144

1145 Supplementary Figure 2. (a) Ternary diagram of classification of feldspars; (b) Ternary diagram
1146 of classification of clinopyroxene; (c) (d) Amphibole classification.

1147

1148 Supplementary Figure 3. Selected CL images of zircon from composite dykes and Aigoual
1149 granite with U–Pb age and ϵ^{Hf} values.

1150

1151 Supplementary Data 1. Analytical methods.

1152 Supplementary Data 2. Coordinates and lithologies of the samples taken from composite dykes in
1153 the Aigoual granite.

1154

1155 Supplementary Data 3

1156 Supplementary Data 3 Table S1. Whole-rock major- (wt.%) and trace-element (ppm)
1157 compositions of composite dykes from Aigoual granite, French Massif Central.

1158 Supplementary Data 3 Table S2. Whole-rock Sr, Nd, Hf isotopes.

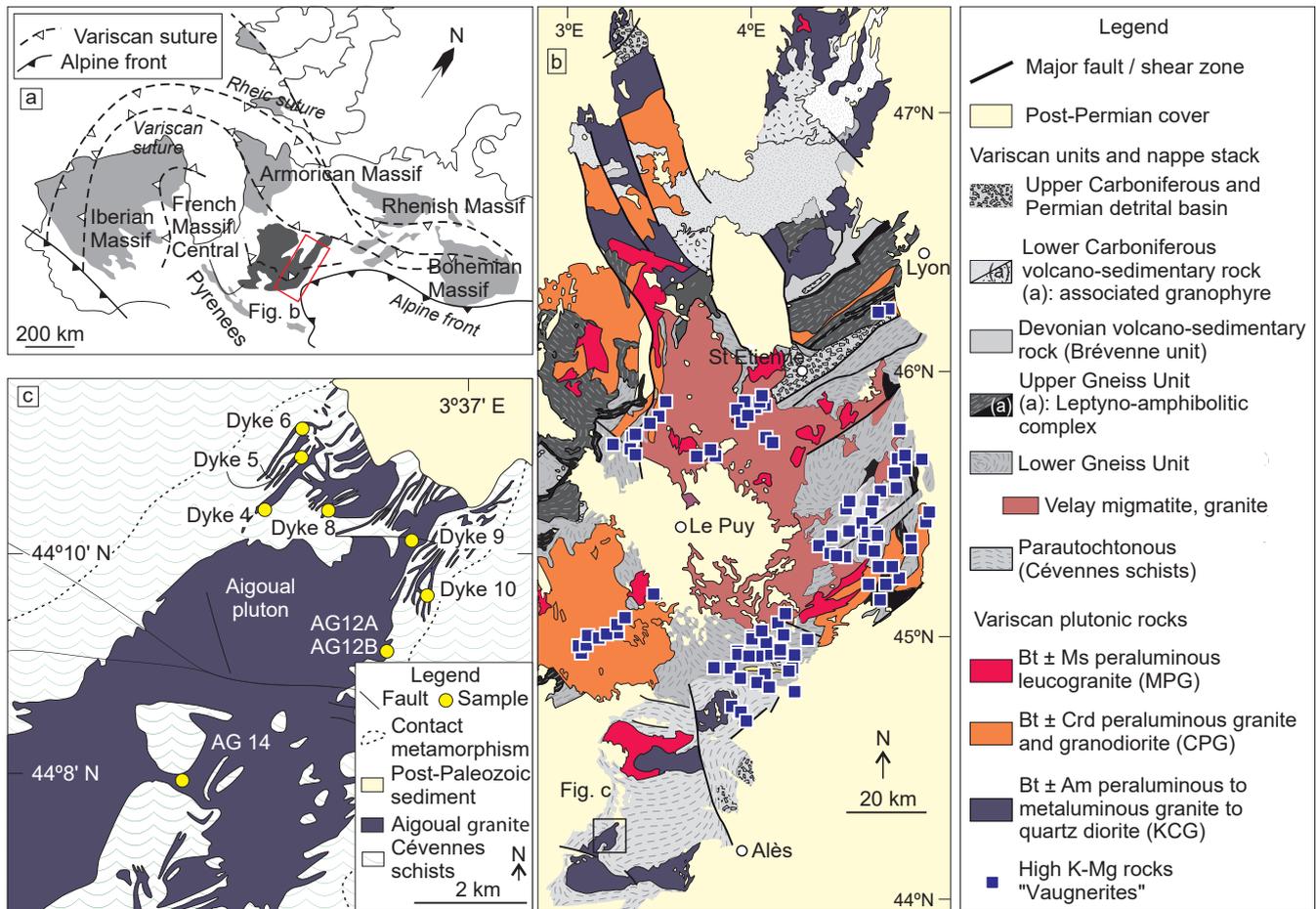
1159 Supplementary Data 3 Table S3. $\delta^{18}\text{O}$ values from quartz and feldspar from lamprophyre and
1160 granite of composite dykes and Aigoual granite (AG14A).

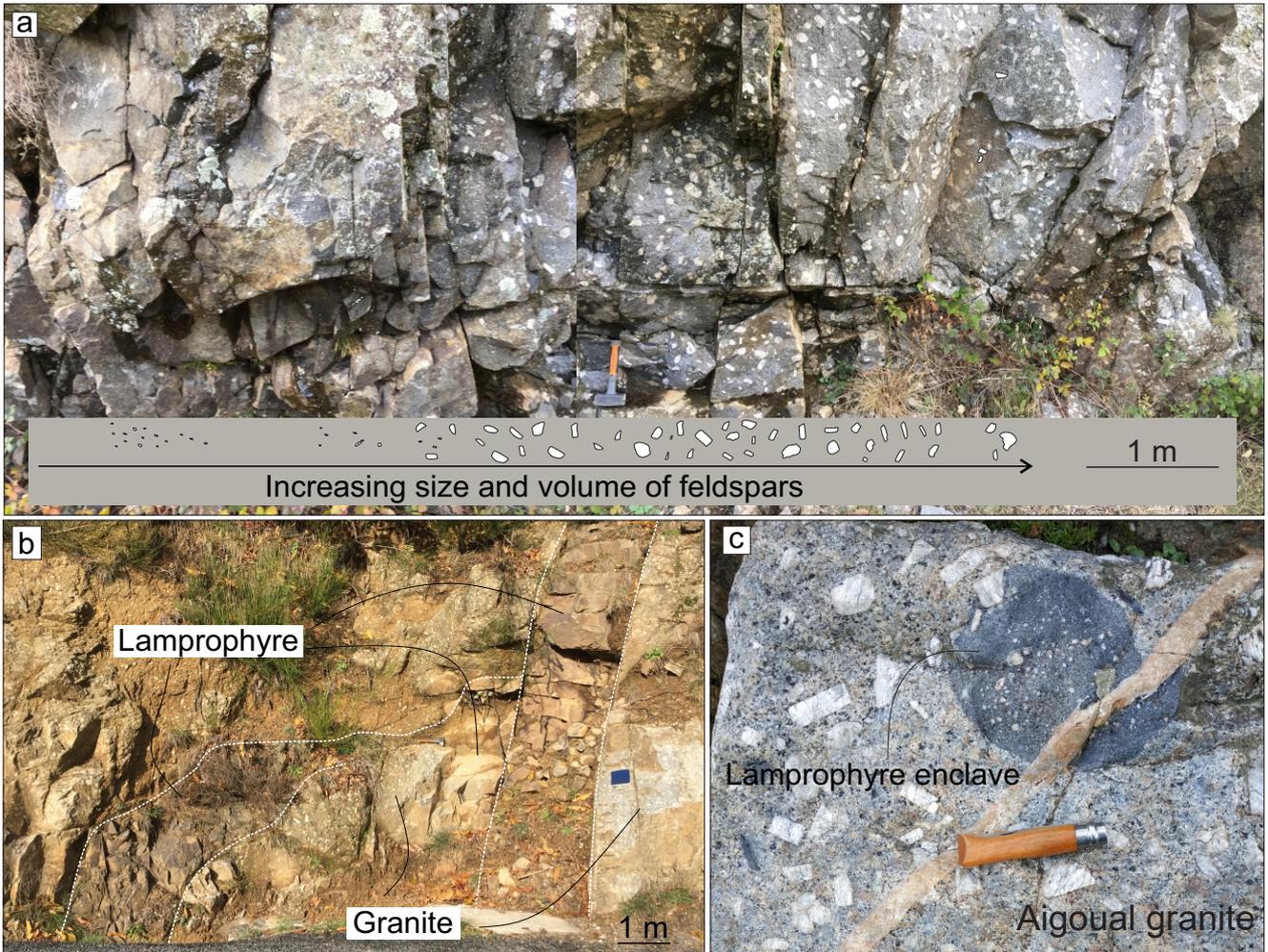
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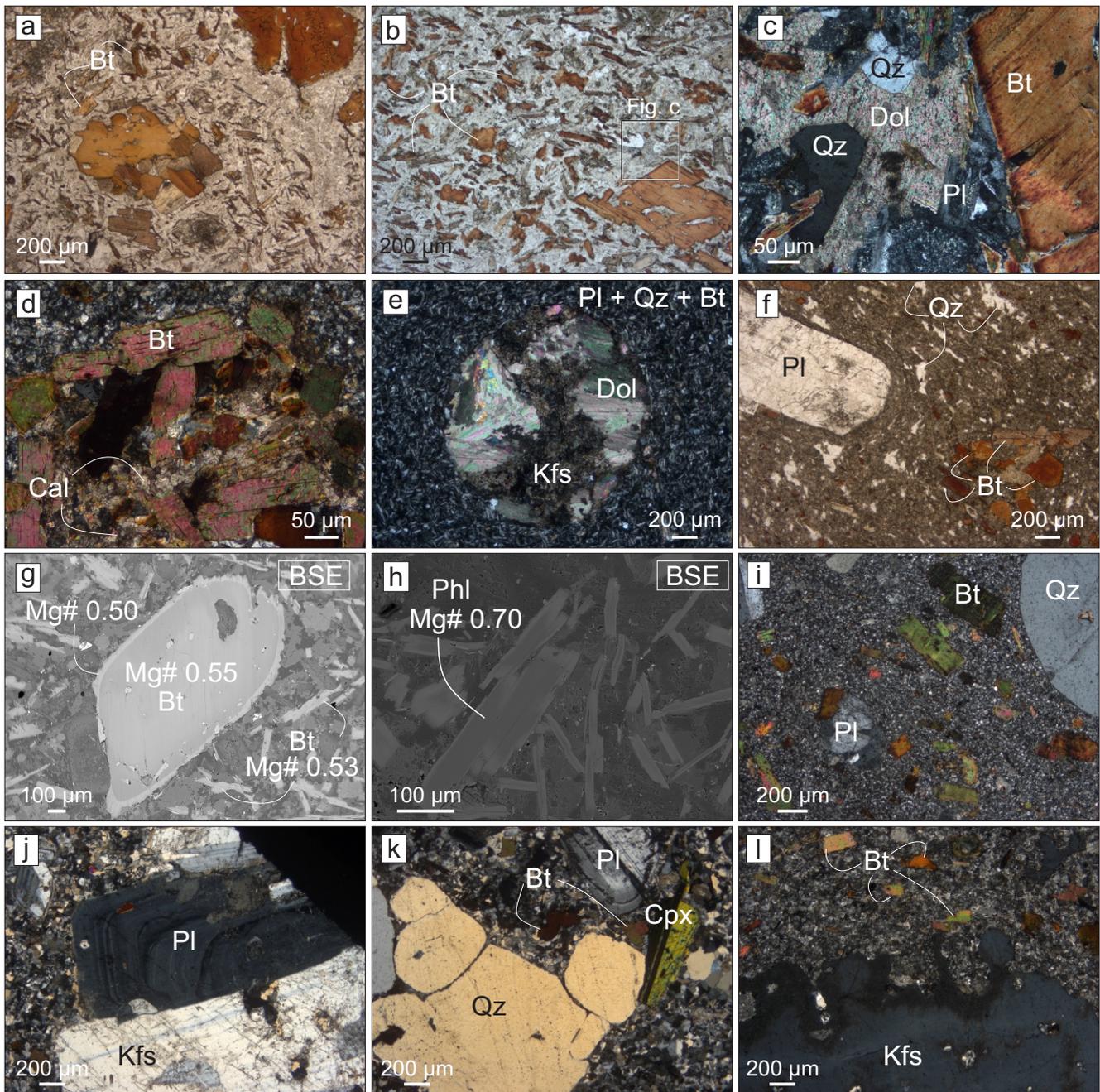
1162 Supplementary Data 4

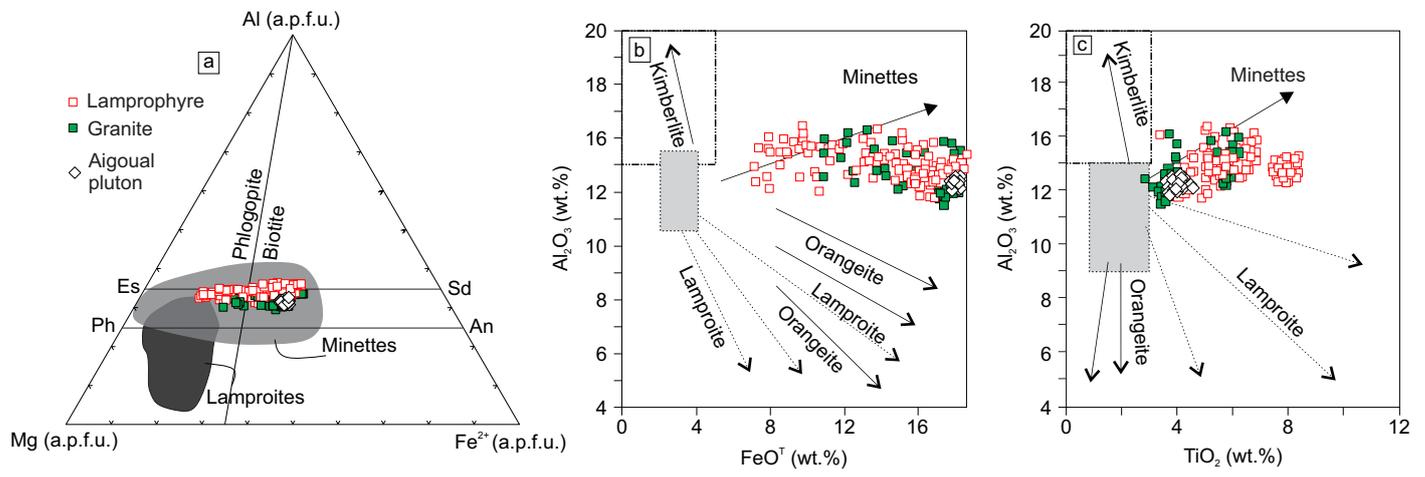
1163 Supplementary Data 4 Table S1. Zircon LA-SF-ICP-MS U–Pb geochronology data.

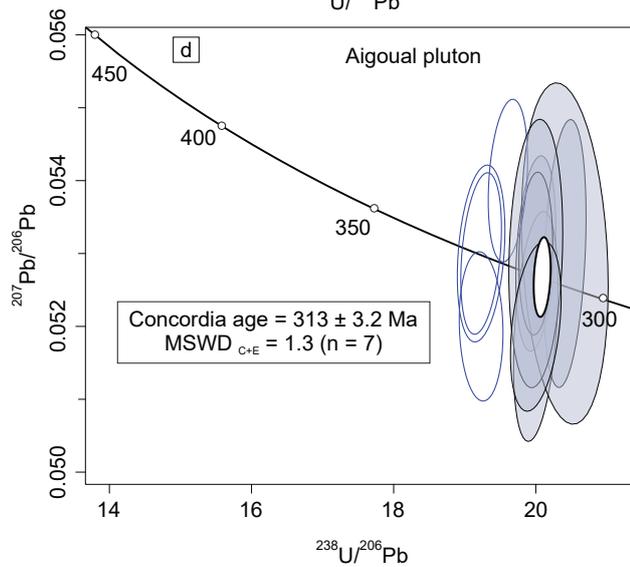
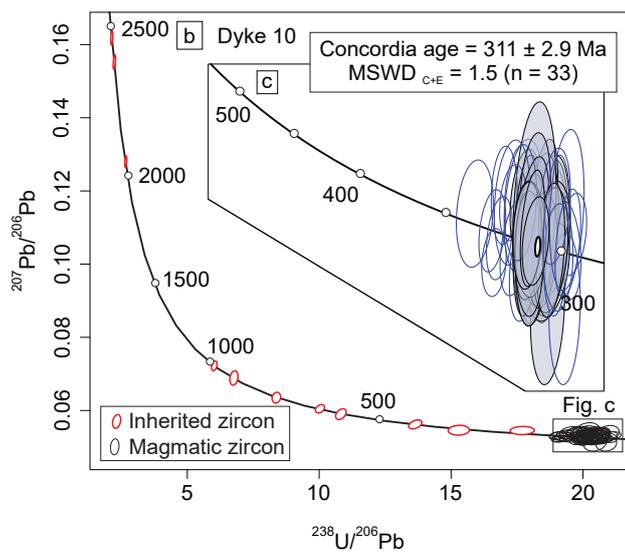
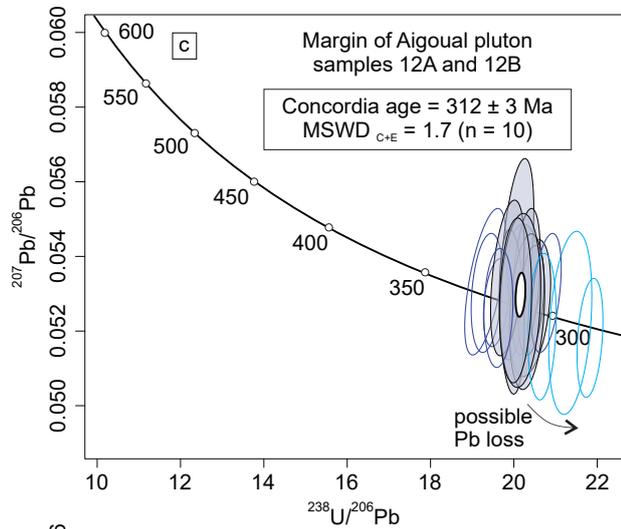
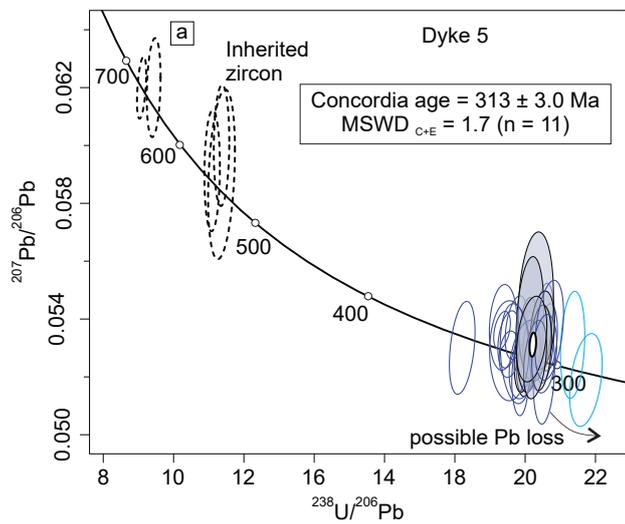
- 1164 Supplementary Data 4 Table S2. Analytical results for zircon Lu–Hf isotopes determinations.
- 1165
- 1166 Supplementary Data 5
- 1167 Supplementary Data 5 Table S1. Mineral chemistry of biotite/phlogopite
- 1168 Supplementary Data 5 Table S2. Mineral chemistry of feldspars
- 1169 Supplementary Data 5 Table S3. Mineral chemistry of clinopyroxene
- 1170 Supplementary Data 5 Table S4. Mineral chemistry of amphibole
- 1171

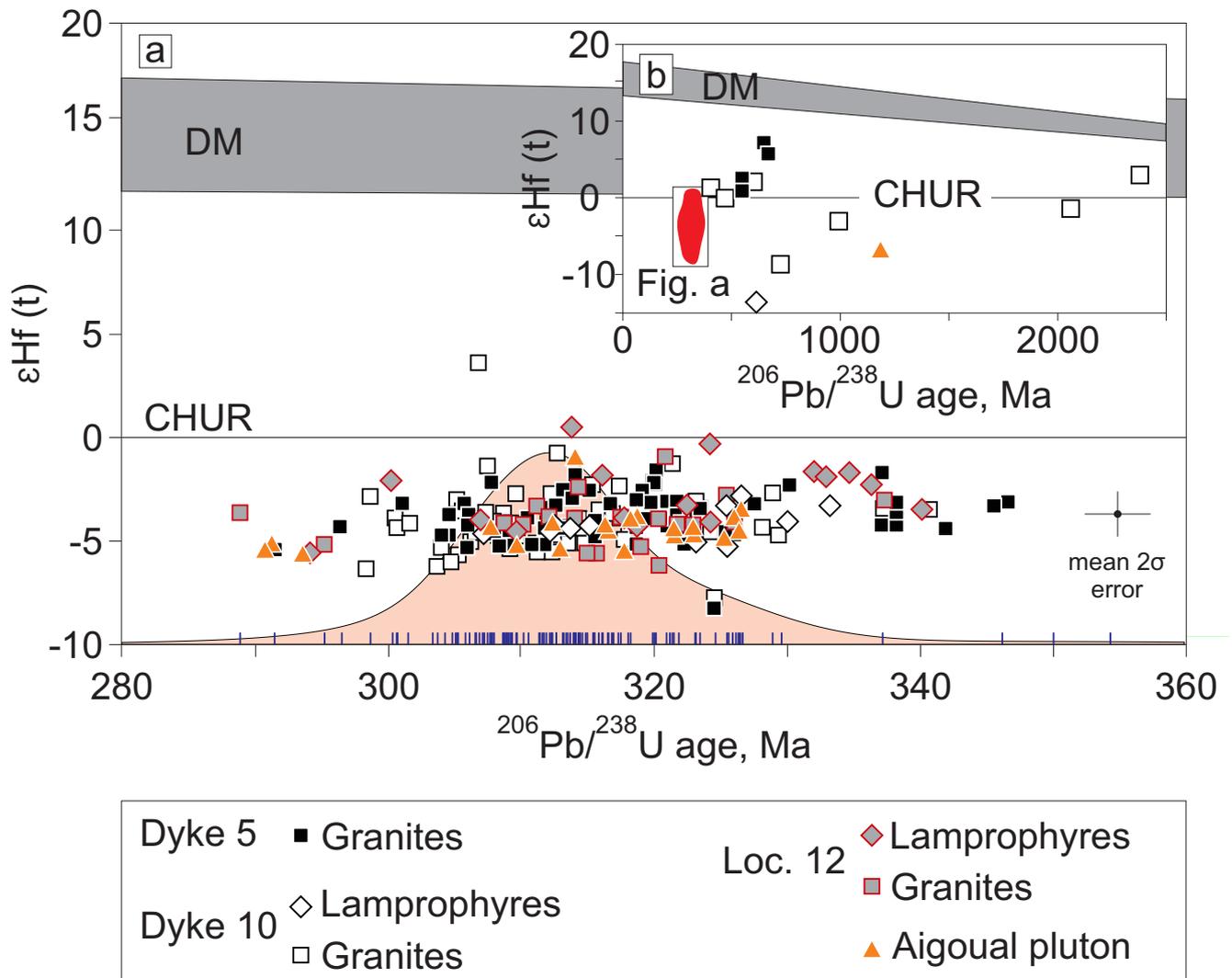


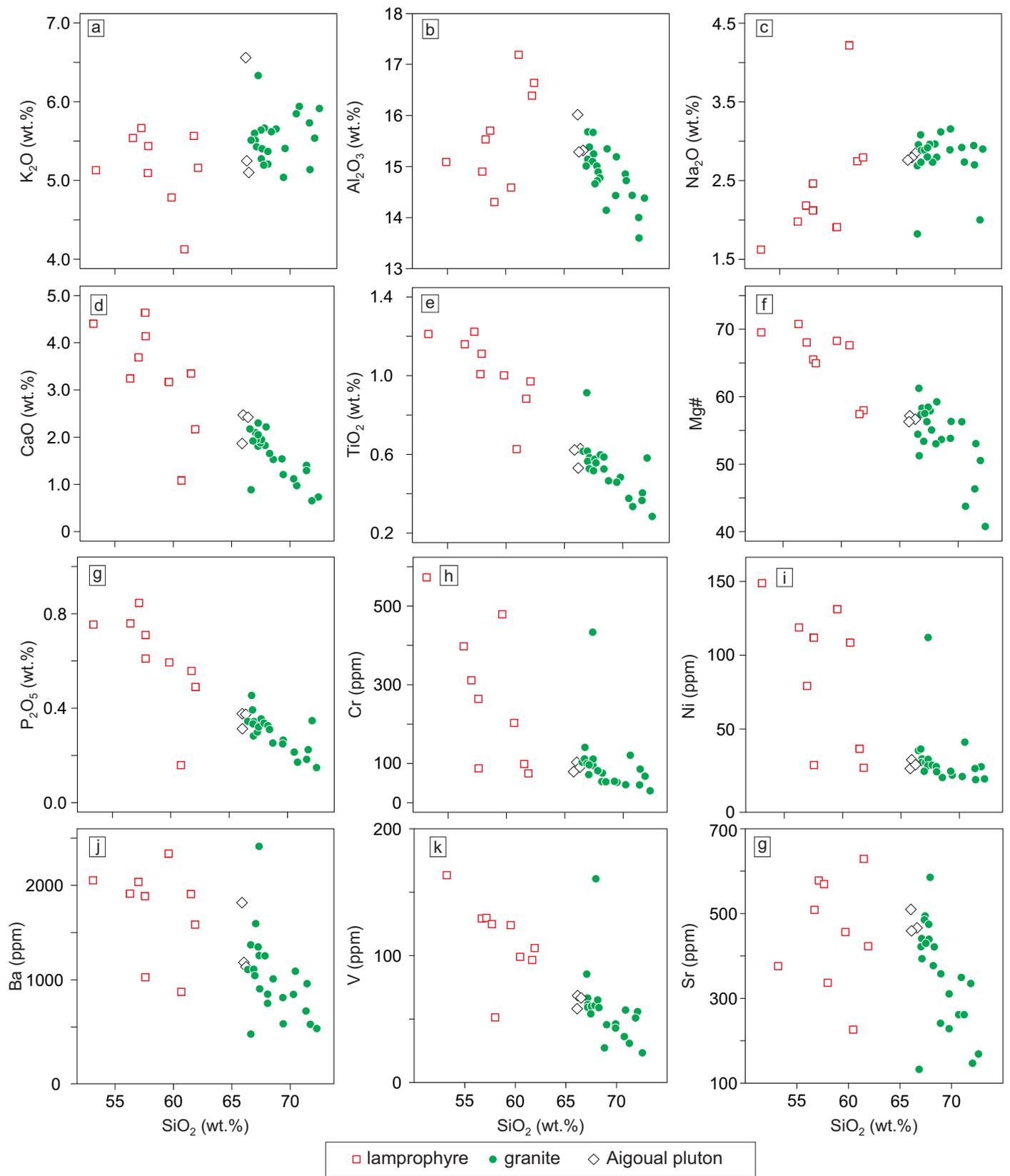


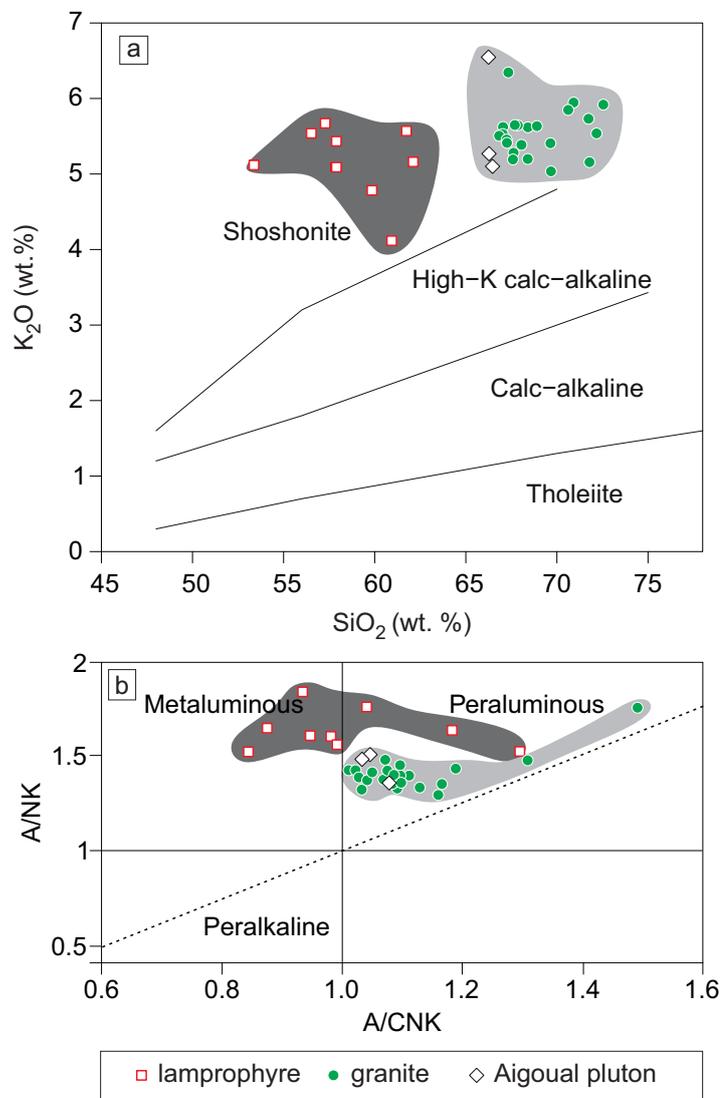


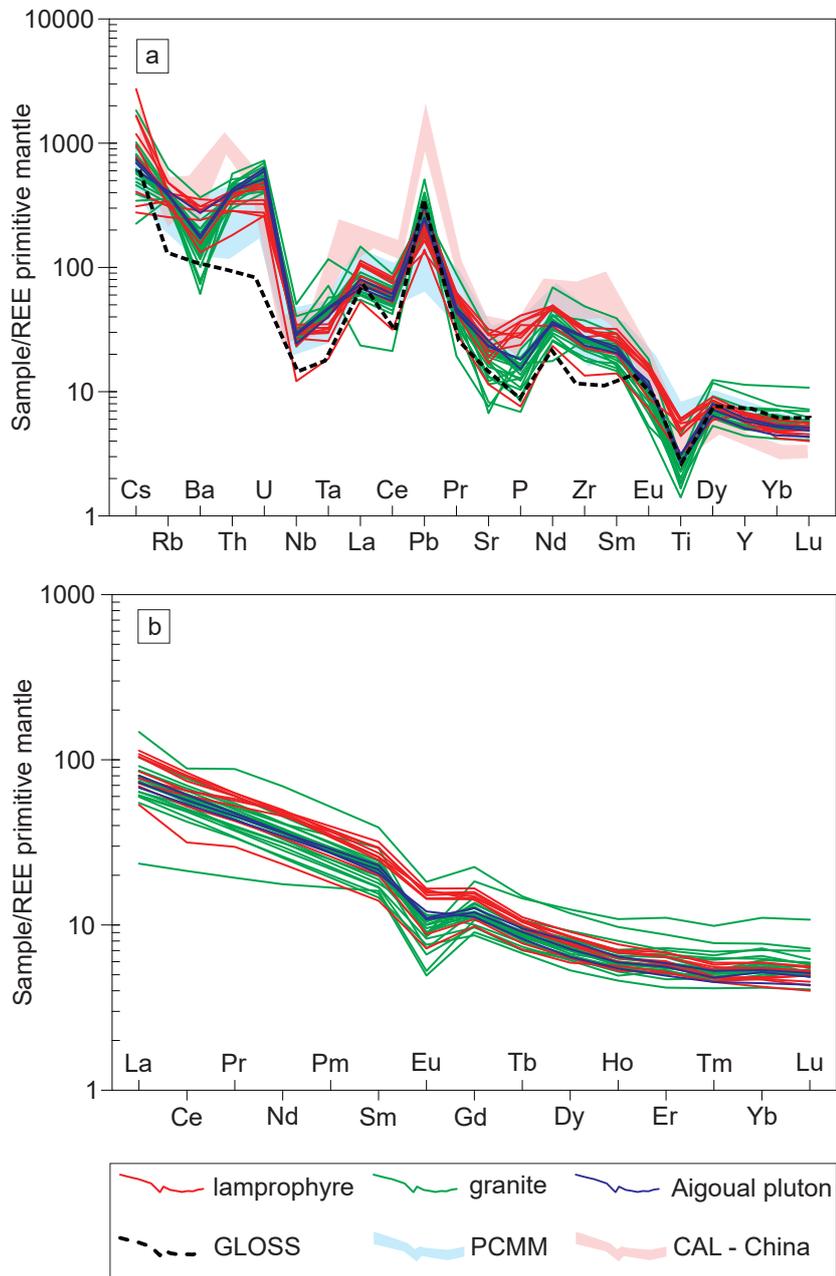


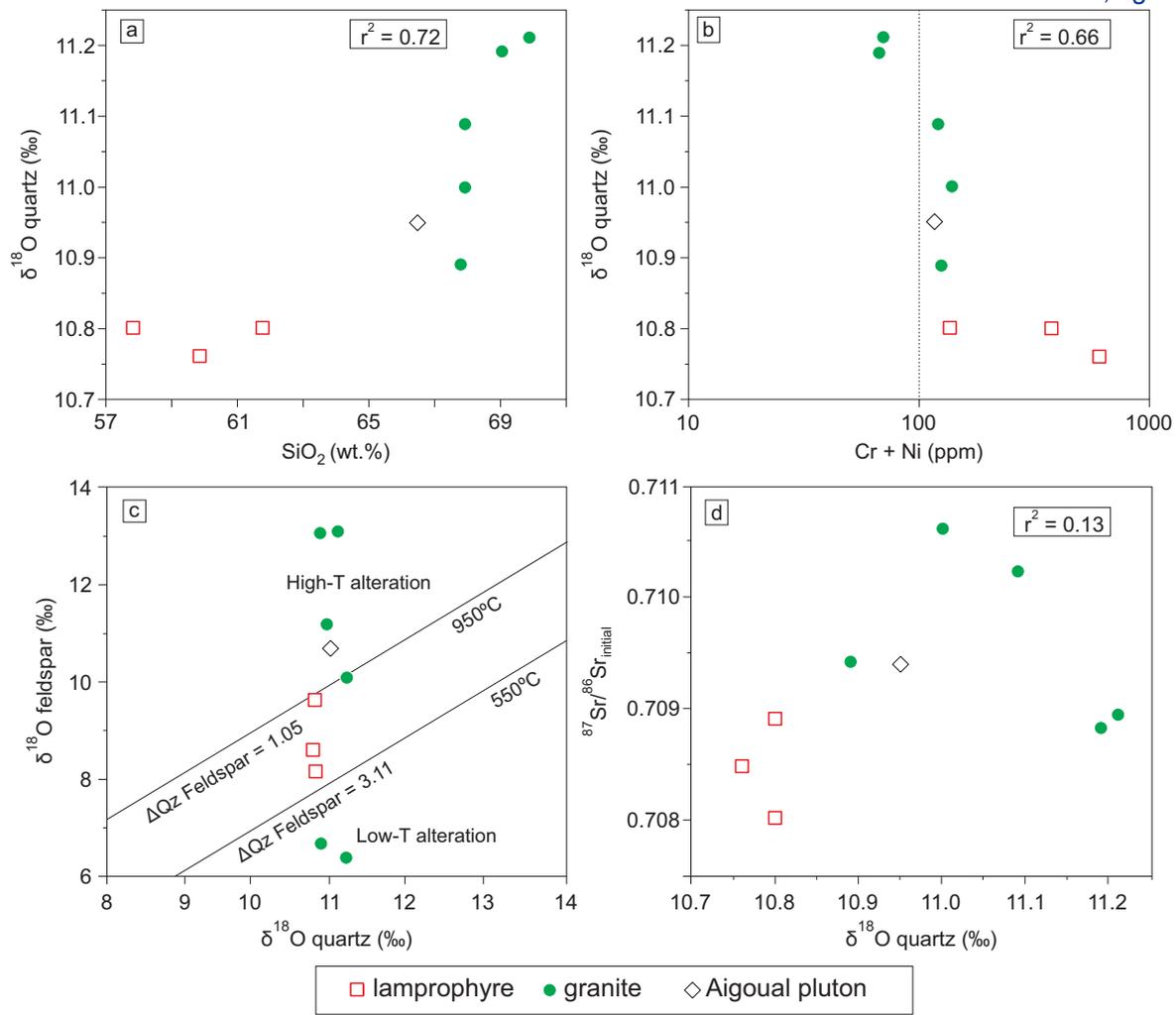


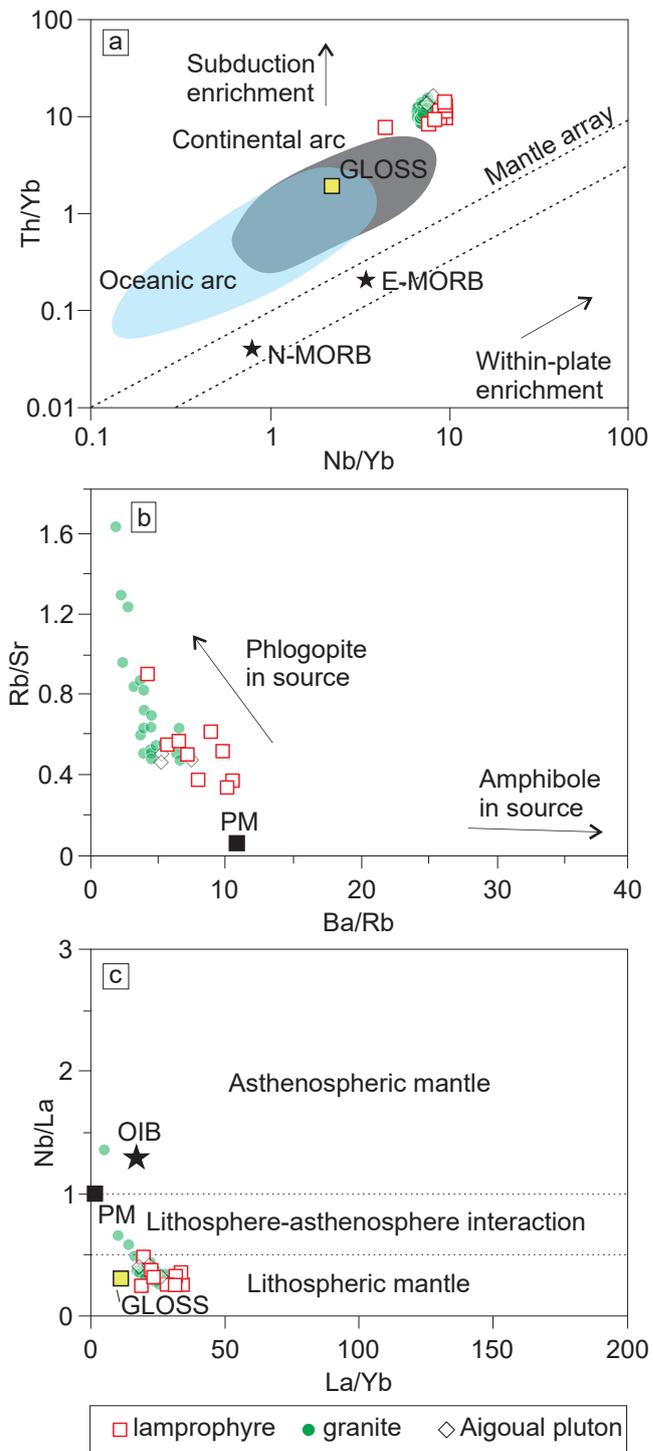


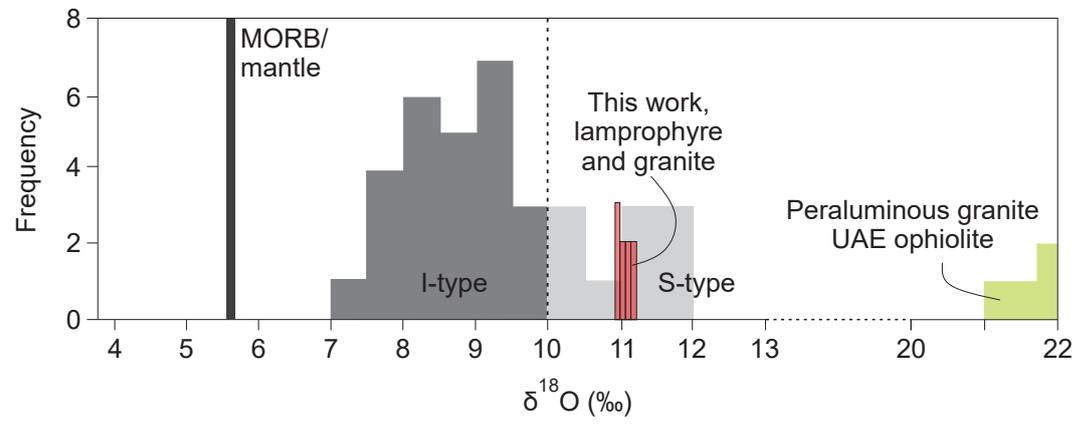


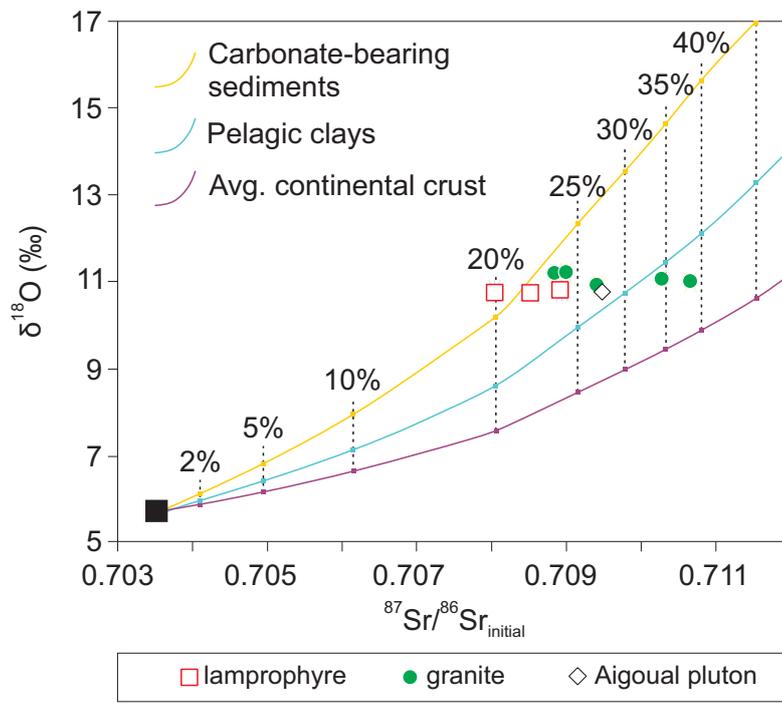


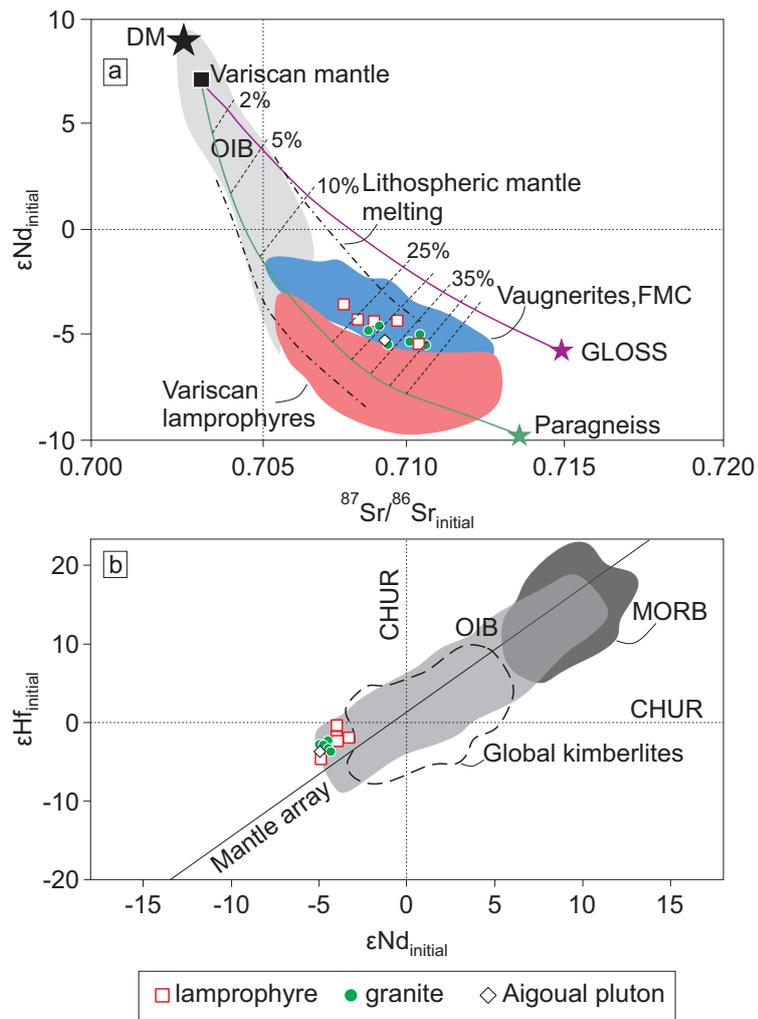


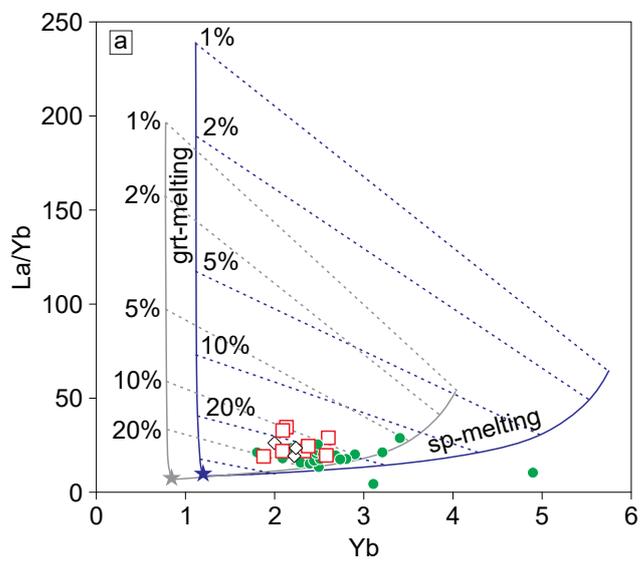




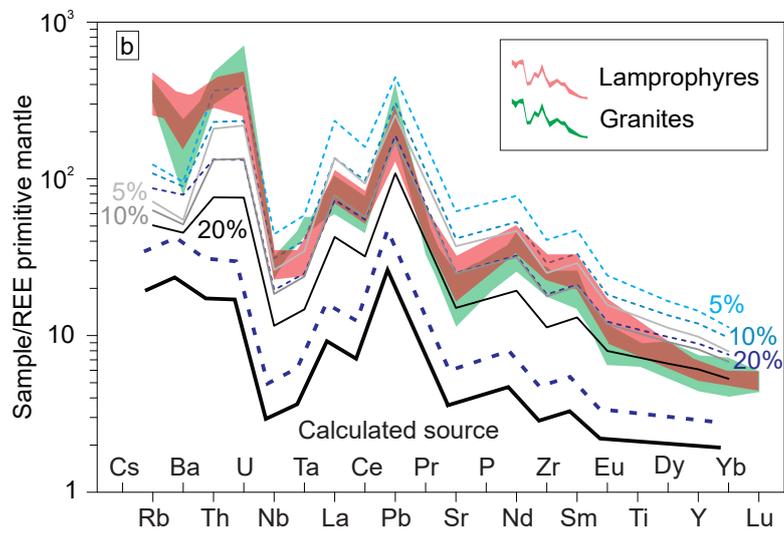








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Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: