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1 The role of melting on the geochemical evolution and
2 isotopic variability of an anatectic complex in the Iberian
3 Variscides

4

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16

17 Abstract: Formation and evolution of migmatite-granite terranes usually involve several
18 complex petrological and geochemical processes that leave their imprint on elemental
19 and isotopic signatures. In this paper, we assess the role of melting reactions on the
20 genesis of the Variscan Figueira de Castelo Rodrigo-Lumbrals Anatectic Complex
21 (FCR-LAC) at the Central Iberian Zone, culminating in the generation of abundant S-
22 type granites. At odds with the proposed for several anatectic complexes elsewhere, it
23 is demonstrated that no isotopic (Sr-Nd-Hf-Pb) disequilibrium occurred in the FCR-
24 LAC, in the transitions from metatexites to diatexites and, finally, to granites. However,

25 variable contribution of the source minerals generated batches of melts characterized
26 by distinct parent/daughter ratios, which explain the significant heterogeneity of
27 present-day isotopic signatures. While Rb/Sr ratios and the Pb budget of the
28 successively generated lithotypes were mainly controlled by major minerals, such as
29 muscovite, K-feldspar and plagioclase, the accessory phases played a major control on
30 Sm/Nd and Lu/Hf ratios. Our study demonstrates the existence of two distinct diatexite
31 groups produced by different reactions. Type-1 diatexites, having high Rb contents and
32 fractionated HREE, are geochemically more akin to the associated S-type granites.
33 They were produced via fluid-absent reactions during dehydration-melting of muscovite
34 with production of peritectic K-feldspar, sillimanite and melt. The less abundant type-2
35 diatexites required influx of external fluids during melting reactions, which consumed
36 more plagioclase than muscovite.

37 Isotopic data reveal the existence of two groups with distinct initial compositions (ϵNd_{320}
38 = -5.05 to -6.03 and ϵHf_{320} = -3.42 to -4.45 vs. ϵNd_{320} = -7.30 to -8.89 and ϵHf_{320} = -6.45
39 to -8.47), both composed of metatexites, diatexites and granites, which is explained by
40 source heterogeneity. The main source of the anatexitic complex was the
41 metasedimentary Neoproterozoic – Lower Cambrian Douro-Beiras Supergroup, with
42 minor contribution of the Ordovician Ollo de Sapo magmatic rocks.

43

44 Keywords: Anatexitic complex; Melting reactions; Isotope geochemistry; Variscan
45 Orogeny

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47

48

49 1. Introduction

50 Late-stage mountain building is typically characterized by crustal anatexis and
51 generation of granitic bodies, a common feature in the continental tectonic framework

52 since the Archean (Whitney et al., 2004). Upper crust emplacement of granitic magmas
53 leaving refractory residues leads to intracrustal differentiation, which explains the
54 compositional distinction between lower and upper crust (Sawyer et al., 2011).

55 The production of large volumes of melt through partial melting of a source rock
56 depends on protolith fertility, on the presence of fluids, and on the temperature-
57 pressure conditions, namely the magnitude of the thermal anomaly (Brown, 2013;
58 Clemens, 2006). Metapelitic and metapsammitic rocks containing large amounts of
59 hydrous phases, such as muscovite and biotite (30% to 50%), are fertile protoliths and
60 potential sources of peraluminous melts through metamorphic incongruent melting
61 reactions (fluid-present or fluid-absent) at 700 – 800 °C (Brown, 2013; Sawyer et al.,
62 2011). Indeed, fluid-absent melting of micas in metapelites and metagreywackes can
63 yield up to 50 vol% of melt (Clemens and Vielzeuf, 1987; Sawyer et al., 2011; Bento
64 dos Santos et al., 2011b; Brown, 2013). For these reasons, migmatite-granite
65 complexes have been the target of several studies with the objective of exploring links
66 between high-grade metamorphism, partial melting processes, the origin of the related
67 granitic bodies and crustal differentiation (e.g. Johannes et al., 2003; Bento dos Santos
68 et al., 2011a; Brown et al., 2016; Sola et al., 2013).

69 In the Iberian Variscan Belt (the southwestern sector of the European Variscan
70 orogen), more precisely in the Central Iberian Zone (CIZ), several authors have
71 emphasized the relationship between some granites and migmatitic rocks (Areias et al.,
72 2014; Carrington da Costa and Teixeira, 1957; Ferreira et al., 2014; Pereira et al.,
73 2017; Ribeiro et al., 2011; Vanderhaeghe, 2009). The Figueira de Castelo Rodrigo-
74 Lumbrals Anatectic Complex (FCR-LAC) is one of such anatectic complexes and has
75 the particularity of showing a clear spatial relationship between metatexites, diatexites
76 and granites. In this way, this complex offers the opportunity to assess the melting
77 reactions that genetically link migmatites (metatexites and diatexites), granites, and
78 their metasedimentary sources. Links between the FCR-LAC granites and the hosting
79 pre-Ordovician metasedimentary rocks have already been suggested by Ferreira et al.

80 (2019), based on the study of inherited zircon grains, a question that will be
81 readdressed in this paper.

82 Radiogenic isotopes are widely applied to identify source characteristics of
83 magmatic rocks (e.g. White, 2010). Indeed, it is usually assumed that the isotopic
84 composition of a magma is similar to that of the source rock, a hypothesis implying that
85 no isotopic fractionation occurs during melting and that almost the same happens with
86 the parent (P)/daughter (D) element pairs. This is true if each of the source mineral
87 phases is not a significant repository of any of the P or D elements for the isotopic
88 systems considered, i.e. when the P and D elements are highly incompatible. However,
89 in silica-rich magmatic systems, such as S-type granites and related anatectic rocks,
90 isotopic disequilibrium becomes more common (e.g. Himalayan leucogranites - Ayres
91 and Harris, 1997; Anatectic Complex of Toledo - Barbero et al., 1995; Sierra Nevada
92 Batholith - Tommasini and Davies, 1997; Southern Sierra Nevada - Zeng et al., 2005b;
93 see also Wolf et al., 2019;). Indeed, residue/melt isotopic disequilibrium can take place
94 due to a) insufficient temperature to reach the closure temperature of some of the
95 accessory phases, with which P and/or D elements are compatible, resulting in a
96 significant fractionation (up to $> 20 \text{ } \epsilon_{\text{Hf}}$ units; Tang et al., 2014); and b) chemical
97 diffusivity being sluggish compared with the time frame necessary for melt extraction
98 from the residue (Ayres and Harris, 1997; Farina and Stevens, 2011; Zeng et al., 2005;
99 Tang et al., 2014; Wolf et al., 2019). These factors can lead to the perception that
100 isotopic disequilibrium during anatexis is ubiquitous (e.g. Tommasini and Davies,
101 1997). However, in some cases, even for situations specifically investigated for isotopic
102 disequilibrium, no significant evidence—was found (e.g. Wolf et al., 2019). In this
103 perspective, the isotopic variability of an anatectic complex must also be addressed in
104 light of regional source heterogeneities (e.g. Yakymchuk et al., 2015, 2013), and on the
105 role of melting reactions leading to distinct P/D pairs with the consequent development
106 of distinct isotopic signatures due to variable rates of radiogenic ingrowth through time.

107 For these reasons, the importance of isotopic disequilibrium during anatexis is still
108 inconclusive.

109 Distinct isotopic systems are expected to behave differently during anatexis, with
110 some of them responding variably to diverse melting conditions (e.g. Wolf et al., 2019).
111 This confers to multi-isotopic studies the potential to investigate the hypothetical role of
112 isotopic disequilibrium and source heterogeneity, as well as to better decipher the role
113 of distinct melting reactions during the different stages of anatexis.

114 Therefore, in this study, we use Sr, Nd, Hf and Pb isotopes, the elemental
115 whole-rock compositions, and zircon and apatite trace-element compositions from the
116 FCR-LAC lithologies to: 1) assess the melting reactions behind the different lithotypes;
117 2) evaluate the potential role of isotopic disequilibrium during anatexis; 3) provide
118 insights onto the causes of isotopic variability characterizing the anatectic complex; and
119 4) explore the geochemical link between the different stages of anatexis, from
120 migmatites to granites, and also to infer potential metapelitic protoliths.

121

122 2. Geological Setting

123 The European Variscan Belt is the result of a complex evolution culminating with the
124 polyphasic collision of the continents Laurentia and Gondwana during the Devonian –
125 Carboniferous periods (e.g. Dias and Ribeiro, 1995; Nance et al., 2010). In the internal
126 zones of the orogen, the continental collision produced the rapid formation of
127 metamorphic core complexes, characterised by the exhumation of migmatites, large
128 volumes of granitic magmas, and LP-HT metamorphism (Burg et al., 1994; Schulmann
129 et al., 2002).

130 The FCR-LAC is located within the autochthonous terranes of the variscan Central
131 Iberian Zone (CIZ) and it is an example of a migmatite-granite complex formed during
132 the Variscan Orogeny (e.g. Ferreira et al., 2019) (Fig. 1). This anatectic complex
133 preserves significant field macroscopic evidence of the gradual anatectic evolution from

134 metatexites, to diatexites and to syn-tectonic granites (Fig. 2). The FCR-LAC contacts
135 with low-grade (biotite and chlorite zone) metamorphic units of Ediacaran-Cambrian (to
136 the North and South) and Ordovician age (to the South) through the Huebra and the
137 Juzbado-Penalva do Castelo shear zones, respectively. Their sinistral kinematics were
138 active at least during the intracontinental collision stage (D_3 , third phase of the Iberian
139 variscan deformation; Pereira et al., 2017 and references therein), although it has been
140 suggested that it represents a much older tectonic discontinuity (Iglesias and Ribeiro,
141 1981). The Juzbado-Penalva do Castelo shear zone (JPCSZ) is a 200 km long, 5 to 15
142 km wide first-order structure, with a proposed horizontal displacement between 65 and
143 100 km (Iglesias and Ribeiro, 1981; Villar Alonso et al., 2000). Geothermobarometric
144 estimates using mineral equilibria point to peak metamorphic conditions of $T = 761 \pm 50$
145 $^{\circ}\text{C}$ and $P = 5.0 \pm 1.0$ kbar (Pereira et al., 2017). Exhumation mechanisms of this
146 anatexitic complex have been widely discussed and two major models have been
147 proposed. Díez Fernández and Pereira (2016) invoked extensional tectonics and
148 orogenic collapse as the mechanism responsible for anatexis and upper crust
149 emplacement, followed by capture by the strike-slip shear zone. However, the
150 interpretation of the complete exhumation process of this granitic-migmatitic
151 association must consider that this complex occurs in contact with low grade
152 metamorphic rocks to which is juxtaposed by first-order high-angle shear zones, also
153 evidenced by recent magnetotelluric studies (Alves Ribeiro et al., 2017). Migmatites
154 and the associated granites preserve a low dipping, non-horizontal transport lineation
155 (6° – 12° ; Pereira et al., 2017), clearly indicative of a significant net vertical mass
156 transfer during the 65 – 100 km horizontal displacement of the JPCSZ, caused by
157 simple shear-dominated transpression during the D_3 stage (see Pereira et al., 2017).
158 Petrochronological constrains recently published estimated cooling rates as high as 35
159 $^{\circ}\text{C}.\text{Ma}^{-1}$ and fast exhumation rates (max of $0.84 \text{ mm}.\text{a}^{-1}$), which confirm a tectonically-
160 assisted exhumation process (Ferreira et al., 2019).

161

162 3. Field and petrographic observations

163 The low-grade (chlorite-biotite zones) metamorphic units are composed of phyllites
164 and quartzphyllites, affected by a nearly E-W schistosity (Fig. 3a). Phyllites in the
165 vicinity of granites show post-deformation andalusite porphyroblast growth, usually
166 pseudomorphosed, formed by contact metamorphism (Fig. 2a and Fig. 3b). The
167 mineral assemblage of these phyllites also includes quartz, muscovite, biotite, and
168 minor plagioclase and chlorite, as well as accessory titanite, zircon, apatite, rare
169 staurolite, and opaque minerals.

170 Metatexites exhibit stromatic structures still preserving a pre-melting banded
171 orientation inherited from the regional E-W schistosity (i.e. inherited fabric). Sometimes,
172 centimetric or millimetric lenses of peritectic sillimanite associated to muscovite are
173 found (Fig. 2b). Occasionally, metatexites are intersected by centimetric leucosome,
174 sometimes boudinated, and filled with granitic to pegmatitic material (Fig. 2c).
175 Leucosome patches up to 20-35 cm, also occur in association with these veins (Fig.
176 2d). The large majority of leucosomes are concordant to the pre-migmatization
177 structures (schistosity) of the hosting rocks. The melanosome of the foliated metatexite
178 is predominantly composed of biotite and muscovite. In contrast, leucosomes are
179 predominantly composed of a quartz-feldspar assemblage (Fig. 3c). Fibrolite appears
180 as relic grains in late muscovite formed during retrogression (Fig. 3d). The most
181 common accessory phases are zircon, apatite, tourmaline, staurolite and ilmenite.

182 Diatexites show structures such as restitic nodules (Fig. 2e), schlieren (Fig. 2f), and,
183 occasionally, ptygmatic folding. Closer to the shear zones, diatexites reveal E-W
184 oriented shear deformation planes, i.e. the same orientation as the shear zone. Locally,
185 there are pegmatitic and leucosome veins forming vein-structured diatexites. Some of
186 these diatexites show a near-nebulitic texture (only slightly more coarser-grained and
187 biotite-enriched than the examples described in Sawyer (2008)). As a whole, diatexites
188 comprise quartz, plagioclase, K-feldspar, biotite, minor fibrolitic sillimanite and

189 secondary muscovite (Fig. 3f). Chemically, two different groups of diatexites can be
190 distinguished (see section 6.2), which are not macroscopically distinguishable.
191 However, thin section petrography reveals that the mineralogical composition is slightly
192 different, with type-1 diatexites having more fibrolitic sillimanite and secondary
193 muscovite, but lower amounts of K-feldspar than type-2 diatexites (Table 1).

194 Additional evidence of partial melting in these migmatites can be found in the
195 occurrence of corroded grains of biotite and melt films (Fig. 3e), mainly in quartz-
196 feldspar boundaries.

197 In general, the contact between the metatexites, diatexites and granites is marked
198 by a gradual transition, but there are also injections of granitic material found in the
199 migmatites, feeding the S-type plutons.

200 Granites are essentially two mica-bearing, with biotite and muscovite appearing in
201 different proportions and with variable grain sizes. These variations led to their
202 grouping onto 10 different granite facies (Silva and Ribeiro, 2000; Fig. 2g and Fig. 2h).
203 Some granitic outcrops reveal deformation structures like those found in the
204 migmatites, compatible with the JPCSZ movement (Fig. 3h). These granites are mainly
205 composed of quartz, plagioclase, K-feldspar, biotite, muscovite (Fig. 3g) and minor
206 fibrolitic sillimanite. Accessory mineral assemblage includes zircon, apatite, rutile
207 needles and ilmenite.

208 Apatite from these granites yield 288 – 307 Ma, while newly grown zircon grains
209 provide magmatic U-Pb crystallisation ages between 300 and 317 Ma (Ferreira et al.,
210 2019). However, a significant part of the zircon population is inherited (54%), yielding
211 ages in the range 400 – 650 Ma, having survived the melting and segregation stages.

212

213 4. Analytical methods

214 We report 55 analyses of whole-rock geochemistry of granites, migmatites and
215 phyllites from the Douro-Beiras Supergroup that were conducted in the Activation Lab

216 in Ontario, Canada. Major elements were measured by Fusion Inductively Coupled
217 Plasma Optical Emission Spectrometry and trace elements by Fusion Inductively
218 Coupled Plasma Mass Spectrometry (see details in the Supplementary Material 1).

219 Whole-rock isotopic analyses of Sr, Nd, Hf and Pb were performed in 13 samples,
220 including 8 migmatites (5 metatexites, 2 type-1 diatexites and 1 type-2 diatexite), 5
221 granites and 2 metasedimentary units from the Douro-Beiras Supergroup at the
222 Laboratoire G-Time of the Université Libre de Bruxelles (ULB, Belgium) using a Nu
223 Plasma I Multi-Collector Inductively Coupled Mass Spectrometer (MC-ICP-MS)
224 instrument (for more details see the Supplementary Material 1).

225 Zircon and apatite grains separated from the anatexitic units were analysed for their
226 trace element content using an ASI RESOLUTION 193 nm ArF excimer laser coupled to
227 the ANALYTIK JENA Plasma Quant Elite quadrupole ICP-MS at the University of
228 Portsmouth (UoP) (see more analytical details in the Supplementary Material 1).

229

230 5. Results

231 5.1. Whole-rock elemental geochemistry

232 Whole-rock elemental compositions are presented in Table 2, where the studied
233 samples are grouped as phyllites, metatexites, diatexites and granites.

234 Phyllites and metatexites are enriched in Al_2O_3 , FeO^{\dagger} , MgO and TiO_2 in comparison
235 to diatexites and granites, evidencing a negative correlation with silica (Fig. 4). This
236 reflects the larger abundance of refractory minerals (e.g. biotite and titanite) in the
237 metasedimentary protoliths and in the metatexite's melanosomes, representative of an
238 incipient degree of partial melting. On the other hand, diatexites and granites
239 commonly yield higher contents of SiO_2 , Na_2O , K_2O and P_2O_5 when compared to
240 metatexites. This difference is consistent with higher degree of partial melting and,
241 eventually the occurrence of some crystal fractionation (e.g. Sawyer, 2008), and, thus,
242 with the production of quartz-plagioclase-, K-feldspar- and apatite-rich melts. Two types

243 of diatexites were identified mainly based on their composition (Fig. 4): type-1
244 diatexites ($\text{SiO}_2 = 68 - 74 \text{ wt}\%$) and type-2 diatexites ($\text{SiO}_2 = 74 - 75 \text{ wt}\%$).
245 Additionally, type-2 diatexites are distinguishable by a wider variation of Na_2O
246 concentration, slightly lower average $\text{Fe}_2\text{O}_3^{\text{t}}$ content and lower concentrations of P_2O_5 ,
247 which are similar to the depicted by metatexites and phyllites.

248 All lithotypes display high Al_2O_3 content ($\text{Al}_2\text{O}_3 > 12 \text{ wt}\%$), being strongly
249 peraluminous (phyllites: $\text{ASI} = 2.0 - 3.3$, quartzphyllites: $\text{ASI} = 1.2 - 2.4$, metatexites:
250 $\text{ASI} = 1.5 - 2.7$, diatexites: $\text{ASI} = 1.1 - 1.5$, and granites: $\text{ASI} = 1.1 - 1.4$).

251 The resemblance between metatexites and phyllites, revealed by major elements, is
252 also generally depicted by trace elements (Fig. 5). Type-2 diatexites exhibit lower Rb
253 contents compared to the other anatectic rocks, but similar to metasediments. Phyllites,
254 quartzphyllites, and metatexites have significantly similar patterns in the multi-element
255 diagrams, showing negative anomalies of Nb, K, Pb, Sr and Ti, and positive anomalies
256 of U, Ta, La, Ce and Nd (Fig. 6a). Zr and Hf positive anomalies are evident in the
257 quartzphyllites (Fig. 6a), as the result of their higher zircon contents (observed under
258 the microscope), in agreement with their more psammitic characteristics in comparison
259 with the remaining metasediments. Multi-element patterns of diatexites and granites
260 overlap within error (Fig. 6b), exhibiting a wide range of values for some elements,
261 such as Th, REE and Zr-Hf. Both lithotypes show negative anomalies for Nb, Pb, Sr
262 and Ti. There are some similarities between metasediments/metatexites and
263 diatexites/granites, as all yield Nb, Pb, Sr and Ti anomalies (Fig. 6a and 6b).

264 REE composition data are summarized in Table 2 and Figure 7. Phyllites,
265 quartzphyllites and metatexites have almost identical normalised patterns (Fig. 7a),
266 with all but one sample being included in a very limited concentration range (e.g. La =
267 41 to 61 ppm; Yb = 2.2 to 3.5 ppm). The exception is one metatexite sample (JTJ-60A)
268 which stands out of the group, because of its lower REE contents (La = 18; Yb = 1.4),
269 but yet yielding similar $(\text{La}/\text{Yb})_{\text{N}}$ ratios (7.43 to 13.35). The Heavy Rare Earth Elements
270 (HREE) are weakly fractionated ($\text{Dy}/\text{Yb}_{\text{N}} = 1.07 - 1.35$) and the Eu anomalies are small

271 (Eu/Eu* = 0.5 – 0.7). Diatexites and granites display a wide range of REE
272 concentrations (Σ = 8 – 777 ppm) (Fig. 7b and 7c), despite the studied granites having
273 similar normalised patterns across the different facies. At odds, diatexites have a wider
274 variability, showing different degrees of HREE fractionation (Dy/Yb_N = 0.8 to 5) and Eu
275 anomalies, either negative (down to 0.2; type-1 diatexites) or positive (up to 4; type-2
276 diatexites). See section 6.2 for more details on the chemical differences between these
277 two types of diatexites.

278

279 5.2. Accessory minerals: trace element composition

280

281 5.2.1. Apatite

282 The trace element compositions of apatite grains from the FCR-LAC and their host
283 rocks (phyllites) are remarkably distinct in Sr, Y, Th and REE compositions (Table 3;
284 Fig. 8a-8e). The highest Sr content in apatite was found in phyllites (317 – 1471 ppm),
285 decreasing gradually in metatexites (113 – 487 ppm), diatexites (86 – 144 ppm) and
286 granites (74 – 99 ppm). Thorium behaves similarly. On the other hand, Y contents in
287 apatite are low in phyllites (480 – 854 ppm), are quite similar between metatexites (837
288 – 1973 ppm) and diatexites (847 – 2026 ppm) and show a large range in granites (212
289 – 1546 ppm). Concerning their U composition, phyllites reveal lower values (8 – 34
290 ppm), while metatexites exhibit a larger variation (17 – 223 ppm), identical to that of
291 diatexites (33 and 185 ppm). The maximum U concentration in the diatexites
292 corresponds to a type-2 diatexite. The variation of U in granites is smaller, but within
293 the same range of values as the migmatites (49 and 81 ppm).

294 Apatite grains are highly enriched in REE with Σ REE ranging from 1397 to 5829.
295 They all show a marked enrichment in LREE compared to HREE, as depicted by
296 (La/Yb)_n ratios (Fig. 8f; phyllites = 5.1 – 32.6, metatexites = 0.6 – 3.7, diatexites = 0.6 –
297 6.8 and granites = 3.7 – 19.1, yet (La/Sm)_n ratios are lower than the unity, except in Ily

298 granite. At odds, apatite (Tb/Lu)_n ratios are clearly higher than 1 for all the lithotypes.
299 The Ily granite has a REE fractionation higher than the remaining granite facies,
300 because of its higher LREE contents. In general, Eu anomalies are negative with the
301 exception of the late-tectonic granite (granite Ily: 1.06), which lacks an anomaly.

302

303 5.2.2. Zircon

304 Contents of Hf, Y, U, Th and HREE in zircon grains exhibit positive correlations with
305 those from the whole rock, indicating that zircon controls the budget of these elements
306 in migmatites and granites (Table 4; Fig. 13b and Fig. 13c). Hf contents are similar
307 between metatexites (8180 – 14410 ppm), diatexites (11060 – 12750 ppm) and
308 granites (10280 – 12670 ppm). Lu values decrease in zircon grains from metatexites
309 (48 – 309 ppm), to diatexites (48 – 91 ppm), and to granites (14 – 58 ppm). We were
310 only able to determine Y contents for half the measured points (11 out of 23). For the
311 available measurements, metatexites yield 120 – 2074 ppm, diatexites 610 ppm, and
312 granites range between 469 – 1199 ppm. U and Th contents are highly variable, with
313 Th/U ratios in zircon from diatexites (0.06 – 0.11) and granites (0.11 – 5.88) being
314 broadly higher than 0.1 (10 spots out of 12), whereas metatexites yield a wide range
315 (0.003 – 0.845), with half the analysed grains showing ratios lower than 0.1 (6 of the 11
316 spots).

317 Zircon from the different units exhibit high and similar Σ HREE concentrations, here
318 considered from Eu to Lu (Fig. 8g), ranging from 490 to 2430 ppm in metatexites, 488
319 to 1679 ppm in diatexites, and 261 to 2540 ppm in granites. They also show Eu
320 negative anomalies (metatexites 0.16 – 0.52, diatexites 0.14 – 0.33 and granites 0.02 –
321 0.16). On the other hand, zircon from the different FCR-LAC rocks show Ce positive
322 anomalies, suggesting high amounts of Ce in 4⁺ valence, which allowed its greater
323 incorporation (Trail et al., 2012). Most zircon in the metatexites have La below the
324 detection limit, so to a better estimation of the REE fractionation, Ce/Yb_N is used. REE

325 fractionation values (Ce/Yb_N) in zircon retrieved from the metatexites are the most
326 fractionated ($Ce/Yb_N = 0.00001 - 0.05$), while zircon from granites and diatexites
327 clearly yield more elevated values ($Ce/Yb_N = 0.01 - 0.4$ and $0.001 - 0.04$,
328 respectively) (Fig. 8g). Overall, Lu/Hf ratios in zircon from the different units are
329 variable, decreasing from metatexites (0.015), to diatexites (0.006) and granites
330 (0.003).

331

332 5.3. Isotopic geochemistry

333 The radiogenic isotopes presented in this study were recalculated to 320 Ma, which
334 corresponds to the migmatization age of the autochthonous domain of the CIZ (318 –
335 325 Ma, Martínez Catalán et al., 2014; Ferreira et al., 2019) (Table 5). While in the
336 previous section, granites were presented as a single lithological group, in this section
337 they are presented as syn-tectonic granites and late-tectonic granites (Ferreira et al.,
338 2019). Likewise, phyllites are presented as phyllite N and phyllite S, depending if they
339 were collected to the North (N) or South (S) of the anatectic complex.

340

341 5.3.1. Sr, Nd and Hf isotopic compositions

342 All migmatitic and granitic samples, except two, plot in the fourth quadrant
343 (radiogenic Sr and less radiogenic Nd) of the $^{87}Sr/^{86}Sr_{320}$ vs. ϵNd_{320} diagram (Fig. 9a),
344 implying time-integrated evolution characterized by enrichment of Rb and Nd relatively
345 to the less incompatible Sr and Sm, respectively.

346 Granites share similar ϵNd_{320} values (-6.03 to -8.89), but significantly different
347 $^{87}Sr/^{86}Sr_{320}$ ratios (0.7022 to 0.7126). These compositions are compatible with the
348 upper continental crust (Chauvel et al., 2014; Hart et al., 1999). Overall, metatexite
349 compositions are similar to those of granites with $^{87}Sr/^{86}Sr_{320}$ values ranging between
350 0.7039 and 0.7124 and ϵNd_{320} between -5.21 to -8.86.

351 The metasedimentary units, phyllite N ($^{87}\text{Sr}/^{86}\text{Sr}_{320} = 0.7153$; $\epsilon\text{Nd}_{320} = -10.07$) and
352 phyllite S ($^{87}\text{Sr}/^{86}\text{Sr}_{320} = 0.7089$; $\epsilon\text{Nd}_{320} = 3.28$), have distinct compositions but both plot
353 in the field defined for the Douro Group (see Teixeira, 2008). The phyllite S is also
354 strikingly different from the FCR-LAC in terms of its Nd composition. Most granites and
355 metatexites show similar compositions to that of the syn-tectonic Aguiar da Beira
356 muscovite-biotite leucogranite (see Costa et al., 2014), while two samples of diatexite
357 and one sample of metatexite have similarities with the Olo de Sapo gneisses (see
358 Montero et al., 2017; Fig. 9b).

359 The anatectic units have enriched Hf isotopic compositions that result in ϵHf_{320}
360 ranging from -3.42 to -9.25. Except for facies Ily ($\epsilon\text{Hf}_{320} = -3.99$), granites have similar
361 ϵHf_{320} compositions (-7.70 to -8.43). The peculiar composition of Ily facies can be
362 related to the fact that this granite is a 300 Ma late-tectonic granite, whereas the
363 remaining granites are syn-tectonic, with ages between 313 and 317 Ma (Ferreira et al.
364 2019). Hf isotopic compositions of the metasedimentary units show distinct
365 compositions (phyllite N with $\epsilon\text{Hf}_{320} = -13.68$ and phyllite S with $\epsilon\text{Hf}_{320} = -7.16$). The
366 ϵNd_{320} vs. ϵHf_{320} diagram shows an overall positive correlation defined by the granites,
367 metatexites and some diatexites (Fig. 9c). Despite some isotopic variability, two tight
368 clusters are noticeable in the ϵNd_{320} vs. ϵHf_{320} diagram, each including granites,
369 metatexites and type-1 diatexites. Type-2 diatexite reveals a ϵNd_{320} composition
370 between the two clusters.

371 Phyllites, N and S, yield two-stage Nd model ages (TDM_2) of 1.83 Ga and 779 Ma,
372 respectively. The range of Nd TDM_2 ages for the metatexites is 1.45 to 1.73 Ga, for the
373 diatexites is 1.43 to 1.69 Ga, and for the granites is 1.51 to 1.73 Ga. The Hf TDM_2 ages
374 are older, displaying the following values: phyllite N (2.16 Ga), phyllite S (1.75 Ga),
375 metatexites (1.57 to 1.83 Ga), diatexites (1.51 to 1.88 Ga), and granites (1.54 to 1.83
376 Ga). These Nd and Hf model ages are similar for most of the anatectic complex
377 lithologies (granites, diatexites, and metatexites). Nd TDM_2 for the phyllite S sample is

378 different and younger than the remaining analysed lithologies, and also somewhat
379 different regarding its Hf TDM₂.

380

381 5.3.2. Pb isotopic composition

382 When recalculated to 320 Ma, the Pb isotopic ratios retrieved from phyllites become
383 clustered onto one group with the remaining lithologies, in opposition to what was
384 shown for the Sr, Nd, and Hf isotopic systems in section 5.3.1. The ²⁰⁶Pb/²⁰⁴Pb₃₂₀ and
385 ²⁰⁷Pb/²⁰⁴Pb₃₂₀ isotopic ratios tend to be almost identical for all lithologies (²⁰⁶Pb/²⁰⁴Pb₃₂₀
386 = 17.24 to 18.27 and ²⁰⁷Pb/²⁰⁴Pb₃₂₀ = 15.61 to 15.67), with the exception of one
387 metatexite outlier (Fig. 10a). In comparison, the ²⁰⁸Pb/²⁰⁴Pb₃₂₀ isotopic ratio (36.67 to
388 39.39) is slightly more variable (Fig. 10b).

389

390 6. Discussion

391

392 6.1. Granite systematics

393 The intimate field relationship between granites, migmatites and metasediments in
394 the FCR-LAC suggests a genetic link between these lithotypes. This is supported by
395 the granites plotting close to ternary minimums in the Ab-Or-Q ternary diagram (Fig.
396 11c; Tuttle and Bowen, 1958; Winter, 2014). This points out to the S-type character of
397 the studied granites, which is also supported by their peraluminous character (ASI >
398 1.1) (Fig. 11a) and by their positioning in the A-B granite classification diagram of
399 Villaseca et al. (1998) (Fig. 11b). Low CaO (0.2 – 1 wt.%), Na₂O (in general < 3.2
400 wt.%), and high K₂O (≈ 5 wt.%) contents, as well as ⁸⁷Sr/⁸⁶Sr₃₂₀ ratios up to 0.7124 are
401 also typical of S-type granites (Chappell and White, 1974). ⁸⁷Sr/⁸⁶Sr₃₂₀ ratios are
402 somewhat variable (0.7039 – 0.7124), reflecting variability between different granite
403 facies. This variability, typical of S-type granites (Chappell and White, 2001), can be a

404 consequence of heterogeneous protoliths or reflect isotopic disequilibrium (see below).
405 The tight clustering of granites and migmatites in the ϵNd_{320} vs. ϵHf_{320} plot (Fig. 9c) puts
406 in evidence the cogenetic association amongst these units, which reinforces their
407 characterisation as S-type granites.

408 The abundant presence of perthites indicates the hypersolvus character of these
409 granites, compatible with relatively low H_2O pressure, which is also suggested by the
410 positioning between the ternary minimum defined at 0.1 and 0.2 GPa in the Ab-Or-Q
411 ternary diagram (Fig. 11c).

412

413 6.2. The diatexite-granite link

414 In an anatectic complex, diatexites are regarded as the result of melting processes
415 which by melt aggregation, migration, and emplacement give rise to granitic intrusive
416 bodies (Milord et al., 2001). Diatexites in the FCR-LAC are characterized by a
417 significant compositional variability, which can provide clues on the shared genetic link
418 with granites.

419 Major and trace element concentrations were used to geochemically separate
420 distinct types of diatexites: type-1 ($\text{SiO}_2 = 68 - 74$ wt%) and type-2 ($\text{SiO}_2 = 74 - 75$
421 wt%), whereas granites ($\text{SiO}_2 = 70 - 75$ wt%) are generally less silica-enriched than
422 type-2 diatexites. Using the Frost et al. (2001) diagram, type-1 diatexites and granites
423 are alkali-calcic to alkalic, whereas type-2 diatexites range between the alkali-calcic
424 and calc-alkalic (Fig. 11d). Regarding trace element compositions, type-2 diatexites
425 also stand out by their lower Rb (145 – 238 ppm) (Fig. 5a) and higher Yb and Lu
426 concentrations (Fig. 6b), and flat Tb to Yb patterns ($\text{Tb}/\text{Yb}_N = 0.8 - 1.5$) (Fig. 7b), at
427 odds with what can be observed in the other diatexites and granites (Fig. 7b and 7c)
428 ($\text{Tb}/\text{Yb}_N > 2.64$). In addition, type-1 diatexites exhibit fractionated HREE ($\text{Dy}/\text{Yb}_N = 2 -$
429 5) and Eu negative anomalies (0.2 – 0.7), in contrast with type-2 diatexites, which

430 reveal unfractionated HREE ($Dy/Yb_N = 0.8 - 1$) and Eu anomalies from slightly negative
431 to strongly positive (0.6 – 4).

432 Comparing both types of diatexites with granites, it is evident the similarity between
433 granites and type-1 diatexites. Taking into account the differences between these two
434 types of diatexites, type-2 could correspond to cumulates of quartz and feldspar that
435 evolved isolated from type-1 diatexites and granites through an earlier fractional
436 crystallization process (e.g. Brown et al., 2016; Morfin et al., 2014; Sawyer, 1987). The
437 Hf isotopic signatures of these two types of diatexites are clearly distinct, with ϵHf_{320} of
438 the type-2 diatexites being significantly less radiogenic (up to 6 units; Table 5) than the
439 reported for type-1 diatexites and also for the studied granites. Thus, the Hf isotopic
440 data reinforces the idea of existing two distinct types of diatexites and of a genetic link
441 between type-1 diatexites and granites, which is not so obvious when type-2 diatexites
442 are considered. The role of melting reactions in this variability is potentially significant
443 and is discussed in the following section.

444 Taking into account the referred isotope similitude with granites (see above), type-1
445 diatexites are here considered as related to the granites by melt-residuum separation.
446 This process is evident in the Harker diagrams (Fig. 4), where this type of diatexites are
447 tendentially more rich in FeO^t , MgO , TiO_2 and Al_2O_3 , but comparatively depleted in K_2O ,
448 Na_2O and SiO_2 than the associated granites.

449

450 6.3. Melting reactions

451 The formation of migmatite-granite complexes is the culmination of high-grade
452 metamorphism during orogenic events. Partial melting is a sequential process where
453 the production and accumulation of melt gradually increases with time. As prograde
454 reactions take place, melting of the protolith lead to the formation of metatextitic and
455 diatextitic migmatites, culminating with a granitic melt at higher melt fractions (Brown,
456 2013). Melting reactions can take place at different temperatures, pressures and water

457 contents. Typically, pelitic and greywacke protoliths begin to melt at about 650-700 °C
458 in the presence of an aqueous fluid (fluid-present reactions), producing up to 60 vol.%
459 of melt (Brown, 2013; Milord et al., 2001; Vielzeuf and Holloway, 1988; Weinberg and
460 Hasalová, 2015). During the prograde metamorphic path, as temperature increases to
461 ca. 800-1000 °C, partial melting can occur in the absence of an external fluid, giving
462 place to a series of fluid-absent reactions by dehydration of hydrous minerals, such as
463 micas and amphiboles, which progressively increases melt production up to 70 modal
464 percent melt (Brown, 2013; Bucher and Grapes, 2011; Yakymchuk and Brown, 2014).

465 Field observations combined with major element geochemistry of the FCR-LAC
466 units show that the most probable protolith of the granitic melts (s.l.) is of sedimentary
467 origin (see sections 2 and 6.1). Moreover, the major elements geochemistry, in
468 particular the CaO/Na₂O ratio, reinforces the idea that FCR-LAC granites are derived
469 from pelitic melts (CaO/Na₂O < 0.3; Jung and Pfänder, 2007) (Fig. 12a). The type of
470 melting reactions can be inferred based on trace-element modelling using Rb, Ba, Sr
471 (Inger and Harris, 1993) and taking into account the mineral assemblages of the FCR-
472 LAC rocks (section 3). Yet, the exact reaction cannot be accurately determined given
473 its high dependence on the chemical and modal composition of the protolith, which in
474 the present study is thought to be significantly heterogeneous on basis of several
475 evidences (sections 6.5 and 6.6; Teixeira, 2008). In this way, we can infer that melting
476 of the pelitic source was mainly controlled by two consecutive reactions leading to the
477 progressive increase of melt production.

478 It has been shown that fluid-present and fluid-absent reactions produce melts with
479 distinct characteristics for some key trace-elements, which are constrained by relative
480 mass proportions of consumed feldspar and mica (Inger & Harris, 1993; Gao et al.,
481 2017; Dou et al., 2019). The role of these two reaction types on the evolution of the
482 studied anatectic complex (FCR-LAC) is well evident on diagrams involving elements
483 compatible with plagioclase or muscovite (Fig. 12b; Fig. 12c).

484 The operation of fluid-present reaction is characterized by the development of
485 evolution trends with almost constant Rb/Sr ratios but highly variable Sr and Ba
486 contents (e.g. Gao et al., 2017; Dou et al., 2019), which at FCR-LAC (Fig. 12b and
487 12c) encompass phyllites and most of the metatexites.

488 Petrographic observations indicate that the metatexite melanosomes are
489 characterized by a lower plagioclase/muscovite proportion than the phyllites from
490 where they are considered to have been derived. This suggests that the following fluid-
491 induced key reaction (Patiño Douce and Harris, 1998; Weinberg and Hasalová, 2015)
492 occurred during the initial steps of melting:

493



495

496 Increase of Rb/Sr ratios from phyllite, to metatexite, and to diatexite/granite and the
497 higher Sr and Ba in metatexites in comparison with diatexite/granite (Fig. 12b) seem to
498 support that early partial melting and melt segregation (phyllites to metatexite
499 formation) resulted from such congruent fluid-present reaction (1), consuming
500 plagioclase in larger quantities than muscovite (see also Dou et al., 2019; Gao et al.,
501 2017; Martini et al., 2019).

502 As temperature increases during prograde metamorphism, fluid-absent melting can
503 take place by the consumption of hydrous minerals, such as micas. A significant
504 increase of Rb/Sr ratios accompanied by a decrease of Sr and Ba is observed for
505 granites and most type-1 diatexites (Fig. 12b and 12c), suggesting dehydration-melting
506 involving the significant consumption of muscovite and crystallization of peritectic
507 minerals, such as K-feldspar and sillimanite, in a reaction that could have been similar
508 to (2) (Patiño Douce and Harris, 1998; see also Gao et al., 2017; Inger and Harris,
509 1993; Yang et al., 2019):

510



512

513 This reaction also involves the crystallization of peritectic minerals, such as K-
514 feldspar and sillimanite. This agrees with the absence of primary muscovite in type-1
515 diatexites and granites, as well as the abundance of K-feldspar and sillimanite.
516 Reaction (2) also explains the negative Eu anomalies characterizing most of type-1
517 diatexites and granites (Eu/Eu^* down to 0.2), which can be explained by the
518 crystallization of peritectic K-feldspar (Fig. 12d; Dou et al., 2019; Gao et al., 2017).

519 After fluid-absent reactions involving the consumption of muscovite, it is commonly
520 described the dehydration-melting of biotite as a response to temperature increase
521 along the metamorphic path. There is some petrographic evidence stemming from
522 migmatites for the breakdown of biotite, such as the occurrence of biotite crystals with
523 corroded borders. However, reactions involving the melting consumption of biotite are
524 accompanied by the peritectic production of garnet (> 0.4 GPa) or cordierite (< 0.4
525 GPa) (Spear et al., 1999; Bento dos Santos et al., 2011). Such minerals were not
526 detected in the studied samples, although garnet has been reported (Díez Fernández
527 and Pereira, 2016), and therefore some incipient melting of biotite should have
528 occurred.

529 We demonstrated before the existence of two geochemically distinct types of
530 diatexites (see section 6.2) and that type-1 diatexites were derived from fluid absent
531 reactions (2) associated to the dehydration-melting of muscovite. Type-2 diatexites
532 exhibit positive Eu anomalies (Eu/Eu^* up to 4), but low CaO contents and Rb/Sr ratios
533 (Fig. 12b; 12c), compared to the remaining granites and type-1 diatexites. This implies
534 that peritectic K-feldspar was not formed or only crystallized in small proportions (Dou
535 et al., 2019; Gao et al., 2017), and, instead, another process must have been
536 important. Fluid-fluxed melting of muscovite (reaction 1) is a more likely mechanism to
537 generate melts with compositions similar to type-2 diatexites. The melt productivity of
538 such reaction is dependent on the amount of fluid present in the rock. The lower
539 abundance of type-2 diatexites and the lack of granites with compositions compatible

540 with this type of diatexites suggest that the amount of melt generated by that fluid-
541 present reaction was not enough to promote the melt segregation/extraction needed to
542 granite generation. This suggests that the amount of fluid in the source rock was small.
543 Yet, as mentioned in section 6.2, the relative higher silica content in type-2 diatexites
544 suggests that their composition probably reflect, in addition, the occurrence of some
545 fractional crystallization.

546 Ti-in-zircon temperatures from young zircon grains within the syn-tectonic granites
547 range between 783 – 836 °C (Ferreira et al., 2019), implying that the metamorphic
548 peak reached such temperatures. Additionally, this range of estimated melt
549 temperatures are consistent with the minimum metamorphic peak temperature
550 previously calculated from mineral equilibria for the FCR-LAC ($T = 761 \pm 50$ °C; Pereira
551 et al., 2017). At this temperature range, up to 60% volume of melt is produced by the
552 referred partial melting reactions (Vielzeuf and Holloway, 1988), thus allowing
553 significant melt extraction given that the critical melt percentage for melt migration (20
554 to 35% volume; e.g. Clemens and Vielzeuf, 1987 and Bento dos Santos et al., 2011a;
555 $\leq 7\%$; Rosenberg and Handy, 2005 and Brown, 2013) is clearly surpassed.

556

557 6.4. The role of accessory minerals

558 Accessory minerals have an important role controlling the composition of trace
559 elements and isotopic ratios in migmatites and granites (e.g. Ayres and Harris, 1997;
560 Bea, 1996a, 1996b; Bea et al., 1994; Yakymchuk and Acosta-Vigil, 2019; Zeng et al.,
561 2005a). Apatite and zircon are the main accessory minerals that occur in the FCR-LAC
562 rocks and their trace element compositions are described in section 5.2.

563 Results demonstrate that apatite is the main carrier of LREE and Y in the FCR-LAC
564 (Table 3) and exhibits a flat REE pattern (more evident for LREE; Fig. 8a), typical of
565 peraluminous rocks (Bea, 1996a). The moderate LREE and P_2O_5 covariation observed

566 in the whole-rock geochemistry (Fig. 13a) confirm the control that apatite has in the
567 variation of these elements.

568 Conversely, the HREE composition of the anatectic rocks is mainly influenced by
569 zircon (Table 4). This influence is also confirmed by the covariation between HREE and
570 Zr/Hf from the whole-rock geochemistry (Fig. 13b-d). Granites show depletion in the
571 HREE and lower Lu/Hf ratios when compared to metatexites and the country rocks,
572 which is due to the role of zircon. Indeed, Lu is preferentially retained in restitic zircon
573 grains during partial melting, resulting in a constant decrease of the Lu concentration
574 and Lu/Hf ratio in subsequent crystallised zircons.

575 Inherited zircon grains are quite common in all these granites, indicating survival of
576 significant amounts of zircon, in agreement with partial melting experimental data.
577 Indeed, considering an average Zr bulk-rock composition of 150 ppm (Table 2), an
578 estimated maximum melting temperatures of ~840 °C (section 6.3) and an approximate
579 1:1 proportion of metapelite and greywacke in the protolith, only about 15% of zircon is
580 expected to dissolve into the melt (Yakymchuk et al., 2017).

581 Apatite grains found in the metasedimentary units are highly enriched in LREE (La
582 up to 1595 ppm) and Th (up to 97 ppm), which is a typical feature of magmatic apatite
583 (Henrichs et al., 2018). Therefore, it is plausible that these correspond to detrital
584 magmatic-derived apatite grains. The recrystallization of apatite during prograde
585 metamorphism can explain the relative depletion of LREE (La ≤ 405 ppm) in the
586 metatextitic apatite, typical of metamorphic apatites (Henrichs et al., 2018).

587

588 6.5. Isotope variability

589 The studied rocks form two tight clusters in the ϵHf_{320} vs. ϵNd_{320} diagram (Fig. 9c) ,
590 each of them comprising metatexites, diatexites and granites. Sm, Nd, Lu and Hf, the
591 parent and daughter elements of the Nd and Hf isotopic systems, are all characterized
592 by high field strengths, which confer them a significant degree of immobility during

593 meteoric and metamorphic processes. Considering this, such clusters clearly suggest
594 the existence of two somewhat distinct sources for the FCR-LAC anatectic complex, in
595 agreement with the significant isotopic variability of the DBSG phyllites (Teixeira,
596 2008). Additionally, the fact that each of the clusters comprise metatexites, diatexites
597 and granites suggests that during the metatexite-diatexite-granite evolution partial
598 melting was not accompanied by significant isotopic disequilibrium, like those caused
599 by the incomplete melting of zircon and the consequent variable release of zircon
600 unradiogenic ^{177}Hf and non-zircon Hf during melting (the zircon effect; Tang et al.,
601 2014). Considering the very long half-life of ^{176}Lu (37.1 Ga) and ^{147}Sm (106 Ga) and the
602 lack of evidence for isotopic disequilibrium, protolith minerals must have had a
603 comparatively short time period for developing distinct isotopic signatures.
604 Alternatively, their isotopic compositions may have been homogenised during the high-
605 grade metamorphism path preceding anatexis itself (Wolf et al., 2019). The occurrence
606 of restricted isotopic disequilibrium during the first melting stages that led to the
607 formation of the metatexites should not be discarded. However, given the natural
608 isotopic variability of the probable source rocks (e.g. Teixeira, 2008) the most likely
609 explanation for the isotopic variability in the migmatite-granites is source heterogeneity.
610 Below we will focus on the metatexite-diatexite-granite compositional transitions.

611 It should be noticed that the present-day values (ϵ_0) show a distinct behaviour for the
612 isotope systems in consideration. Indeed, inside of each of the two groups, enclosing
613 metatexites/diatexites/granites, a significantly larger variation is observed for the ϵHf_0
614 values than for ϵNd_0 (Fig. 14a). Considering this and the fact that in the ϵHf_{320} vs.
615 ϵNd_{320} diagram (Fig. 9c) two clusters are observed comprising granites and migmatites,
616 we propose that a significant fractionation between Hf and Nd isotopic systems
617 occurred during melting reactions from metatexites to diatexites and finally to granites,
618 caused by a distinct behaviour of their parent/daughter pairs. As shown before (6.3),
619 the melting process involved essentially micas and plagioclase, phases with no
620 capability to significantly fractionate Sm from Nd or Lu from Hf, thus suggesting the role

621 of an accessory phase to explain this different behaviour. Apatite in peraluminous
622 systems has $D^{\text{Sm/Nd}} \approx 0.3$ and $D^{\text{Lu/Hf}} \approx 70$ (Bea et al., 1994), making it a good
623 candidate. However, ϵNd_0 of granite-metatexite pairs is similar or only slightly varying,
624 which would not be expected, and apatite dissolution during melting reactions would
625 induce an increase of $^{176}\text{Hf}/^{177}\text{Hf}$ by radiogenic ingrowth of ^{176}Hf with time, opposite to
626 what it is observed.. Therefore, decoupling of Lu and Hf cannot be attributed to apatite
627 dissolution. The observed decoupling of Hf and Nd isotopic systems can be achieved
628 by the involvement of zircon. Zirconium and Hf are usually considered as two
629 geochemically coherent elements, which explains the very high D_{Hf} characterizing
630 zircon ($D_{\text{Hf}} \approx 2420$; Wang et al., 2010). The consequent very low Lu/Hf ratios
631 characterizing zircon will allow, that by incomplete melting of zircon, Lu/Hf in the melt to
632 be distinct of that of the whole source rock, and that granites would evolve with time to
633 lower $^{176}\text{Hf}/^{177}\text{Hf}$ ratios than the less evolved metatexites. Moreover, variable zircon
634 contribution to melt will contribute to the production of liquid batches characterized by
635 significantly different Lu/Hf ratios. This would be facilitated by incomplete zircon
636 dissolution, a process which has been demonstrated to frequently occur in
637 peraluminous granitic magmas where zircon saturation occurs at relatively low Zr
638 concentrations (~ 78 ppm; Harrison and Watson, 1983; Hogan and Sinha, 1991; Tang
639 et al., 2014) and, also, by the fact that zircon is preferentially located inside biotite
640 which can protect it from dissolution (Bea, 1996) (see also 6.4). In opposition to the
641 observed for Lu/Hf ratios, zircon only have a limited capability to fractionate Sm from
642 Nd ($D_{\text{Sm/Nd}} \approx 1$; Bea et al 1994), thus explaining the contrasting behaviour of Hf and Nd
643 present day isotope ratios depicted on Fig. 14a).

644 Each of the FCR-LAC metatexite/diatexite/granite groups defined on the ϵHf_{320} vs.
645 ϵNd_{320} diagram (Fig. 9c) display a significant present-day $^{87}\text{Sr}/^{86}\text{Sr}$ range at almost
646 constant ϵNd (Fig. 14b). This suggests that Sr isotope signatures evolved under distinct
647 $^{87}\text{Rb}/^{86}\text{Sr}$, allowing radiogenic ingrowth of ^{87}Sr at different rates. At odds with the Nd

648 and Hf systems, which are mainly controlled by the solubility of accessory phases, in
649 the Rb/Sr system the melt composition is constrained by the major mineral phases
650 involved in the melting reactions (Tang et al., 2014; Wolf et al., 2019). In fact, an
651 increase of $(^{87}\text{Sr}/^{86}\text{Sr})_0$ from metatexites to diatexites and then to granites is observed
652 (Fig. 14b), which indicates an increase of $^{87}\text{Rb}/^{86}\text{Sr}$ in the melts progressively
653 generated at higher temperatures. If we take into account the melting reactions inferred
654 for the FCR-LAC (as in section 6.3), it is expected firstly a significant contribution of
655 muscovite to the melt (reactions 1 and 2) and later a minor contribution of biotite, a
656 mineral usually characterized by significantly higher Rb/Sr than coexisting muscovite
657 (e.g. Neiva et al., 2002), which together can explain these isotopic ranges and
658 evolutions.

659 Nonetheless, the range in Sr isotopes is observed even when the initial values
660 $(^{87}\text{Sr}/^{86}\text{Sr})_{320}$ are considered (Fig. 9a). There are several ways to explain variation on
661 the initial Sr isotopic composition: isotopic disequilibrium, hydrothermal alteration,
662 assimilation/fractional crystallization (AFC), and source heterogeneity.

663 Isotopic disequilibrium taking place during the melting stages involving various
664 minerals characterized by significantly distinct $^{87}\text{Rb}/^{86}\text{Sr}$ (plagioclase, muscovite,
665 biotite) is a plausible mechanism, especially in these settings (crustal anatexis).
666 However, these isotopic signatures had to be developed previously to melting to
667 account for variable mineral $^{87}\text{Sr}/^{86}\text{Sr}$ during radiogenic ingrowth (Farina and Stevens,
668 2011; Zeng et al., 2005b). Yet, at odds with the commonly observed in situations of
669 isotopic disequilibrium (e.g. Wolf et al., 2019), no systematic isotopic variation is
670 observed when considered the different lithotypes (metatexites, diatexites, and
671 granites), but only a significant overlapping of the $^{87}\text{Sr}/^{86}\text{Sr}_{320}$ characterizing metatexites
672 and granites. In this perspective, the $^{87}\text{Sr}/^{86}\text{Sr}_{320}$ variability could likely just be the result
673 of open-system processes, such as hydrothermal alteration and/or
674 assimilation/fractional crystallization (AFC). A hydrothermal overprint on these units
675 would result in a lack of correlation between mobile and high field strength immobile

676 elements, such as LILE and Ti, respectively (e.g. Huang et al., 2019). This is not
677 supported by our data, which show a clearly linear correlation between K, Sr, Ba vs.
678 TiO_2 (Fig. 15a-c). Therefore, $^{87}\text{Sr}/^{86}\text{Sr}_{320}$ variability as a result of post-magmatic
679 overprint is not plausible.

680 AFC is not a very probable process in constraining the chemical variability of S-type
681 granitic systems (Clemens and Stevens, 2012; Yakymchuk, 2019). Indeed, as
682 suggested by Castro (2014), the small difference in density between melt and minerals
683 precociously crystallised, as well as the high viscosity of high silica melts, lead to
684 consider as probably small and inefficient the role of fractional crystallization on granitic
685 magmas. In addition, it has been considered, based on thermodynamics, that granitic
686 melts have a limited capability of assimilation of crustal rocks, which is more evident for
687 relatively low-temperature S-type magmas like those in this study (e.g. Glazner, 2007).
688 In agreement, no positive correlation is observed for the studied rocks between La/Sm
689 and Th/Sc (Fig. 15d), both ratios being expected to increase during AFC processes
690 (Huang et al., 2019).

691 Given the lack of evidence for Sr isotopic disequilibrium and for the occurrence of
692 hydrothermal alteration or AFC processes, the large variability of the $^{87}\text{Sr}/^{86}\text{Sr}_{320}$ was
693 most probably caused by source heterogeneity. This is in agreement with geochemical
694 studies of the DBSG phyllites which have shown a much larger heterogeneity of the Sr
695 isotope ratios than the observed for $^{143}\text{Nd}/^{144}\text{Nd}$ (Teixeira, 2008), and in line with
696 Clemens and Stevens (2012) and Yakymchuk (2019) findings, who concluded that
697 source heterogeneity is an important factor on isotopic variations in granites in as much
698 it controls the type of peritectic minerals that will form.

699 On the $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Fig.14c), the studied rocks define a linear
700 array close to the evolution line modelled for the upper continental crust by Zartman
701 and Doe (1981). However, a significantly larger dispersion is observed in the
702 $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ (Fig. 14d and Fig. 10b), suggesting decoupling of ^{206}Pb and
703 ^{207}Pb (the final products of the decay chains of ^{238}U and ^{235}U , respectively) from ^{208}Pb ,

704 the final radiogenic product of ^{232}Th . In granitic systems, the elemental Pb budget is
705 mainly controlled by K-feldspar, whereas accessory phases are the main repositories
706 of U and Th (e.g. Bea, 1996), which have the ability to strongly influence the long-term
707 lead isotope evolution (Hogan and Sinha, 1991; Wolf et al., 2019). Decoupling of
708 uranogenic lead from thorogenic lead (Fig. 10b) may then reflect the variable modal
709 proportions of these phases in the source materials and also the variable contribution
710 to melts of U-rich (e.g.: zircon, apatite and monazite) and Th-rich (e.g.: monazite and
711 apatite) minerals.

712

713 6.6. Source rocks: some considerations

714 Based on the whole-rock geochemistry (e.g. Harker diagrams; Fig. 4 and 5), we can
715 distinguish the geochemical evolution due to partial melting and the progressive
716 residuum-melt separation, with the metapelitic rocks (phyllites) plotting at the origin of
717 the metatexite-diatexite-granite sequence. This relationship is in clear agreement with
718 field-based observations (as seen in section 3).

719 Despite the isotopic differences/variability between the phyllite N and phyllite S
720 (Douro-Beiras Supergroup; Fig. 9b; Teixeira, 2008), and even among the anatectic
721 complex lithologies, the Sr, Nd, and Hf isotopic compositions show a clear affinity
722 between migmatites and phyllites. This is more evident for the phyllite N sample with Sr
723 and Nd signatures within the range of the FCR-LAC units (Fig. 9a and Table 5), than
724 for the phyllite S sample, where only the Hf initial composition is within analytical
725 uncertainty of the remaining anatectic units (Fig. 9c and Table 5). TDM_2 provenance
726 ages of the metasedimentary protoliths of the anatectic complex broadly agree with the
727 phyllite units TDM_2 ages (Table 5), which reinforces the Douro-Beiras Supergroup as
728 representing the most probable protolith.

729 While we can not go too far beyond our current small dataset to draw further links
730 between a definitive source to this anatectic complex (either phyllite N or phyllite S), it

731 seems reasonable to propose a stronger affinity with the Douro Group, sampled to the
732 N. However, we cannot fully discard the involvement of the southern Beiras Group
733 units (Tassinari et al., 1996) nor a minor contribution from the Olo de Sapo magmatic
734 rocks (Montero et al., 2017), particularly when the Nd initial isotopic compositions are
735 taken into account (Fig. 9b). This is particularly the case of the Ily late-tectonic granite,
736 which exhibits isotopic compositions significantly different from the syn-tectonic
737 granites (Fig. 9a and 9c; see Ferreira et al., 2019 for granite ages).

738 The contrasting Nd and Hf isotopic results between phyllites sampled from the
739 northern ($\epsilon\text{Nd}_{320} = -10.07$) and southern ($\epsilon\text{Nd}_{320} = +3.28$) domains are striking, but in
740 agreement with previously published data for the CIZ metasediments (Villaseca et al.
741 (2014). This suggests source heterogeneities between both domains, with variable
742 detrital inputs (more and less evolved crustal sources), which influenced the isotopic
743 composition of the FCR-LAC units.

744 Further work is needed to clearly demonstrate: a) a systematic isotopic variability
745 between the metasedimentary units to the North and South of the FCR-LAC; and b) the
746 affinity of the anatectic complex with the Douro Group. This will further elucidate if the
747 JPCSZ was a major paleogeographic boundary during the Precambrian/Cambrian, as
748 proposed by Villaseca et al. (2014) and Iglesias and Ribeiro (1981).

749 7. Conclusions

750 This study shows the links between the distinct lithologies of the variscan Figueira
751 de Castelo Rodrigo – Lumbrales Anatectic Complex of the Central Iberian Zone and
752 evaluates the occurrence of isotopic disequilibrium during melting reactions from
753 metatexites to diatexites and granites. The main conclusions are summarized below:

754 1) Based on whole-rock elemental and isotopic geochemistry a genetic link is
755 definitively established between metatexites, diatexites and granites, which
756 represent different evolutionary stages of the anatectic complex, culminating on
757 the genesis of the S-type granites.

- 758 2) We demonstrate the existence of two geochemically distinct diatexite groups
759 (type-1 and type-2) generated at different stages of the partial melting process.
760 We show that type-1 diatexite, with high Rb concentration and fractionated
761 HREE is geochemically similar to the regional S-type granites, whereas type-2
762 diatexites are not linked to any of the outcropping granites.
- 763 3) Based on the mineral assemblage and whole-rock geochemistry we
764 demonstrated that type-1 diatexites—and granites were produced during
765 dehydration-melting of muscovite and production of peritectic K-feldspar and
766 sillimanite plus melt. On the contrary, Rb-poor, unfractionated HREE type-2
767 diatexites required influx of externally derived fluids during melting, which
768 consumed more plagioclase than muscovite. The type-2 diatexites represent a
769 more evolved melt composition than the outcropping granites, being mainly
770 composed by quartz and feldspar.
- 771 4) Evidence of incipient biotite melting and Ti-in-zircon thermometry allow to
772 establish peak temperatures ranging from 783 to 836 °C. These conditions
773 resulted in the production of large volumes of melt, which explains the profusion
774 of S-type granites within the CIZ.
- 775 5) There are no signs of significant isotopic disequilibrium during melting in the
776 transition from metatexites to diatexites and, finally, to granites. This means that
777 the source minerals were in isotopic equilibrium at the onset of the melting
778 process, either reflecting that they were formed shortly before melting, or,
779 alternatively, that isotopic homogenization during the high-grade metamorphism
780 preceding the melting process took place. However, the distinct contributions of
781 mineral phases to the melt produced batches of melt with distinct
782 parent/daughters' ratios, which resulted in distinct radiogenic ingrowth, and
783 explain the present-day isotopic heterogeneity found in these units.
- 784 6) Rb/Sr ratios and the Pb budget of the successively generated lithotypes were
785 mainly controlled by major minerals such as muscovite, K-feldspar and

786 plagioclase. The accessory phases were the main minerals controlling the
787 Sm/Nd, Lu/Hf ratios and the U and Th contents.

788 7) Two distinct groups, both comprising metatexites, diatexites and granites, were
789 identified when comparing the initial isotopic ratios, testifying for the existence of
790 source heterogeneity. This is consistent with the isotopic variability of
791 metasediments considered to be the sources of the anatectic complex and may
792 also explain the differences in the initial Hf and Nd isotopic compositions
793 between late-tectonic granite-migmatite cluster and the syn-tectonic granite-
794 migmatite cluster.

795 8) The genetic links with probable protoliths of the anatectic complex were also
796 explored. We showed that anatectic complex is sourced on the
797 Neoproterozoic/Lower Cambrian Douro-Beiras Supergroup, with possible minor
798 contribution of the Ordovician Ollo de Sapo magmatic rocks.

799

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809

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1117 Figure Captions

1118 Fig.1 – (a) Figueira de Castelo Rodrigo-Lumbrals Anatectic Complex in the Iberian
1119 Variscan Belt setting (adapted from Dias et al., 2016); (b) Geological map of the
1120 Figueira de Castelo Rodrigo-Lumbrals Anatectic Complex (modified from Silva and
1121 Ribeiro, 2000).

1122 Fig. 2 – Field observations: (a) phyllites with andalusite porphyroblasts; (b) metatexite
1123 with bands of peritectic sillimanite; (c) metatexite with boudinated leucosome veins; (d)
1124 leucosome pocket associated to metatexite; (e) restitic nodule in diatexite; (f) schlieren
1125 structures in diatexite; (g) nebulitic texture in diatexite; (h) and (i) two-mica granites with
1126 variable grain size.

1127 Fig. 3 – Petrographic observations: (a) phyllite with foliated texture; (b) pseudomorph of
1128 andalusite in the phyllite; (c) alternation of leucosome and melanosome in the
1129 metatexite; (d) fibrolite included in the retrograde secondary muscovite in metatexite;
1130 (e) Corroded biotite and melt films evidences (red arrows) in diatexite; (f) secondary
1131 muscovite including fibrolite in diatexite; (g) granite texture and typical mineralogy; (h)
1132 granite with evidences of deformation.

1133 Fig. 4 – Harker diagrams for major elements for the host rocks (phyllites and
1134 quartzphyllites) and FCR-LAC lithologies (metatexites, diatexites and granites).

1135 Fig. 5 – Harker diagrams for trace elements for the host rocks (phyllites and
1136 quartzphyllites) and FCR-LAC lithologies (metatexites, diatexites and granites).

1137 Fig. 6 – (a) Multi-element diagram for phyllites, quartzphyllites and metatexites; (b)
1138 Multi-element diagram for type-1 diatexites, type-2 diatexites and granites. Chondrite
1139 normalization values after Palme and O’Neil (2003).

1140 Fig. 7 – (a) REE diagram for phyllites, quartzphyllites and metatexites; (b) REE
1141 diagram for type-1 diatexites and type-2 diatexites; (c) REE diagram for granites.
1142 Chondrite normalization values after Palme and O’Neil (2003).

1143 Fig. 8 – (a), (b), (c), (d) and (e) Sr, Y, Th, U and REE (ppm) *versus* SiO₂ WR (Whole-
1144 Rock; wt. %), respectively; (f) Apatite REE pattern for phyllites, quartzphyllites,
1145 metatexites, type-1 diatexites, type-2 diatexites and granites; (g) Zircon REE pattern for
1146 metatexites, type-1 diatexites, type-2 diatexites and granites. Chondrite normalization
1147 values after Palme and O’Neil (2003).

1148 Fig. 9 – (a) $^{87}\text{Sr}/^{86}\text{Sr}_{320}$ *versus* ϵNd_{320} diagram for the host rocks (phyllites) and FCR-
1149 LAC lithologies (metatexites, diatexites and granites); (b) $^{87}\text{Sr}/^{86}\text{Sr}_{320}$ *versus* ϵNd_{320}
1150 diagram for the host rocks (phyllites), FCR-LAC lithologies (metatexites, diatexites and
1151 granites) and comparison with other geological domains of the CIZ: Douro-Beiras
1152 Supergroup (Douro Group – Teixeira, 2008; Beiras Group – Tassinari et al., 1996;
1153 South CIZ – Villaseca et al., 2014), Aguiar da Beira granitoids (biotite granodiorite-
1154 granite and muscovite-biotite leucogranite; Costa et al., 2014), and Ollo de Sapo
1155 gneisses (Montero et al., 2017). The Aguiar da Beira granitoids were chosen because
1156 of its proximity to the FCR-LAC and their syn-tectonic character; (c) ϵNd_{320} *versus*
1157 ϵHf_{320} diagram for the host rocks (phyllites) and FCR-LAC lithologies (metatexites,
1158 diatexites and granites). The green and the purple ellipses correspond to groups
1159 defined by very close ϵNd_{320} vs. ϵHf_{320} compositions.

1160 Fig.10 – (a) $^{206}\text{Pb}/^{204}\text{Pb}_{320}$ *versus* $^{207}\text{Pb}/^{204}\text{Pb}_{320}$ for the host rocks (phyllites) and FCR-
1161 LAC lithologies (metatexites, diatexites and granites); (b) $^{206}\text{Pb}/^{204}\text{Pb}_{320}$ *versus*

1162 $^{208}\text{Pb}/^{204}\text{Pb}_{320}$ for the host rocks (phyllites) and FCR-LAC lithologies (metatexites,
1163 diatexites and granites).

1164 Fig. 11 – (a) Aluminium Saturation Index (ASI) classification for granitic and diatexitic
1165 rocks proposed by (Frost et al., 2001); (b) A-B granite classification diagram of
1166 Villaseca et al. (1998); (c) Ab-Or-Q ternary diagram (Tuttle and Bowen, 1958; Winter,
1167 2014) for the FCR-LAC granites; (d) $\text{Na}_2\text{O}+\text{K}_2\text{O}-\text{CaO}$ (wt. %) versus SiO_2 (wt. %)
1168 classification diagram (Frost et al., 2001) showing the ranges between alkalic and
1169 calcic composition for the FCR-LAC diatexites and granites.

1170 Fig. 12 – (a) $\text{CaO}/\text{Na}_2\text{O}$ versus $\text{Al}_2\text{O}_3/\text{TiO}_2$ (wt. %) distinguish between pelite-derived
1171 melts ($\text{CaO}/\text{Na}_2\text{O} < 0.3$) and melts derived from greywackes or igneous sources (Jung
1172 and Pfänder, 2007) for FCR-LAC granites and diatexites; (b) Rb/Sr versus Sr (ppm)
1173 and (c) Rb/Sr versus Ba (ppm) for the host rocks (phyllites) and FCR-LAC lithologies
1174 (metatexites, diatexites and granites). The vectors represent the evolution of partial
1175 melting reactions (Inger and Harris, 1993): $\text{Mu}(\text{VP})$ – vapour-present muscovite
1176 melting; $\text{Mu}(\text{VA})$ – vapour-absent muscovite melting; $\text{Bi}(\text{VA})$ – vapour-absent biotite
1177 melting; (d) CaO (wt. %) versus Eu/Eu^* for the host rocks (phyllites) and FCR-LAC
1178 lithologies (metatexites, diatexites and granites).

1179 Fig. 13 – Apatite and zircon role during the partial melting process in the FCR-LAC: (a)
1180 P_2O_5 versus LREE (ppm); (b) Zr versus Hf (ppm); (c) Zr versus HREE (ppm); (d) Hf
1181 versus HREE (ppm).

1182 Fig. 14 – (a) ϵNd_0 versus ϵHf_0 diagram for the host rocks (phyllites) and FCR-LAC
1183 lithologies (metatexites, diatexites and granites); (b) $^{87}\text{Sr}/^{86}\text{Sr}_0$ versus ϵNd_0 diagram for
1184 the host rocks (phyllites) and FCR-LAC lithologies (metatexites, diatexites and
1185 granites); (b) and (c) Plumbotectonic model after Zartman and Doe (1981) applied to
1186 the host rocks (phyllites) and FCR-LAC lithologies (metatexites, diatexites and
1187 granites): A – mantle; B – orogene; C – upper crust; D – lower crust.

1188 Fig. 15 – Covariation between LILE (K, Sr and Ba) and HFSE (TiO₂), (a), (b) and (c),
1189 respectively; (d) AFC assessment – La/Sm *versus* Th/Sc (ppm).

1190

1191 Table 1 – Modal mineralogical proportion for the host rocks (phyllites and
1192 quartzphyllites) and rocks from the Figueira de Castelo Rodrigo-Lumbrales Anatectic
1193 Complex (metatexites, diatexites and granites).

1194 Table 2 – Whole-rock chemical composition of phyllites, quartzphyllites (qtzphyllites)
1195 and FCR-LAC rocks (metatexites, diatexites, and granites). Major elements in weight
1196 percentage (wt. %) and the trace elements expressed in parts per million (ppm).

1197 Table 3 – Apatite trace element composition (ppm) from the host rocks (phyllites) and
1198 FCR-LAC rocks (metatexites, diatexites, and granites).

1199 Table 4 – Zircon trace element composition (ppm) from FCR-LAC rocks (metatexites,
1200 diatexites, and granites).

1201 Table 5 – Whole-rock Sr, Nd, Hf and Pb isotopic data for the host rocks (phyllites) and
1202 FCR-LAC rocks (metatexites, diatexites, and granites).

Table 1

	Phyllites	Quartzphyllites	Metatexites	Type-1 Diatexites	Type-2 Diatexites	Granites
Quartz	36	48	32	32	40	33
Plagioclase	10	-	14	22	28	23
K-feldspar	-	-	7	12	14	17
Biotite	30	30	22	9	8	12
Muscovite	19	19	21	20	9	12
Sillimanite	-	-	3	4	<1	3
Chlorite	4	2	-	-	-	-
Accessory minerals	<1	<1	<1	<1	<1	<1

Table 2

Lithology	Phyllite N		Phyllite S		Qtzphyllite N		Qtzphyllite S		Metatexite									
	JTJ-2	JTJ-6	JTJ-18	JTJ-19	JTJ-49	JTJ-41b	JTJ-8	JTJ-20	JTJ-22B	JTJ-24B	JTJ-25	JTJ-37	JTJ-48B	JTJ-53	JTJ-54B	JTJ-56A	JTJ-60A	
SiO ₂	61.02	54.64	63.82	66.63	68.39	65.71	58.57	66.85	63.74	74.73	60.94	69.23	59.31	64.82	58.51	69.93	71.14	
Al ₂ O ₃	18.02	23.40	17.68	14.94	14.18	16.15	20.42	16.58	17.78	12.47	18.57	15.08	18.69	15.58	20.42	14.10	15.38	
Fe ₂ O ₃ ¹	6.95	7.34	4.91	5.36	4.55	6.24	8.10	5.51	5.61	3.78	6.52	4.76	7.32	5.90	6.79	4.47	2.50	
FeO ¹	6.25	6.60	4.42	4.82	4.09	5.61	7.29	4.96	5.05	3.40	5.87	4.28	6.59	5.31	6.11	4.02	2.25	
MnO	0.06	0.07	0.05	0.04	0.06	0.08	0.09	0.06	0.05	0.12	0.04	0.07	0.10	0.10	0.09	0.09	0.08	
MgO	2.40	2.06	1.75	1.67	1.65	2.00	2.43	1.74	2.01	1.25	2.24	1.84	2.70	2.16	2.73	1.69	0.73	
CaO	1.36	0.30	0.10	0.15	2.36	0.44	0.55	0.79	0.35	0.76	0.51	1.13	0.54	0.23	0.69	0.74	0.60	
Na ₂ O	1.75	1.14	1.18	1.35	2.52	1.75	1.57	2.30	1.18	2.18	1.40	2.65	1.21	1.45	1.82	2.80	3.04	
K ₂ O	3.40	4.49	4.34	2.86	3.68	3.03	3.92	3.48	3.84	3.41	4.16	3.27	4.42	2.84	3.67	3.41	4.34	
TiO ₂	0.87	0.82	0.76	0.86	0.76	0.89	0.84	0.86	0.70	0.58	0.78	0.64	0.82	0.70	0.74	0.60	0.32	
P ₂ O ₅	0.14	0.13	0.11	0.07	0.15	0.15	0.13	0.11	0.14	0.17	0.17	0.19	0.15	0.14	0.17	0.16	0.24	
LOI	3.65	5.09	4.22	5.02	1.01	3.60	3.72	2.25	3.69	1.23	4.51	1.53	4.43	4.72	4.75	1.63	1.37	
Total	99.62	99.45	98.92	98.95	99.31	100.00	100.30	100.50	99.08	100.70	99.84	100.40	99.68	98.63	100.40	99.62	99.74	
Sc	15.00	18.00	15.00	14.00	11.00	15.00	18.00	15.00	15.00	8.00	16.00	12.00	17.00	15.00	17.00	13.00	7.00	
V	91.00	115.00	123.00	94.00	67.00	98.00	112.00	92.00	111.00	59.00	111.00	79.00	109.00	136.00	124.00	72.00	36.00	
Ba	737.00	780.00	764.00	608.00	660.00	608.00	735.00	497.00	615.00	683.00	921.00	664.00	738.00	1038.00	1005.00	361.00	286.00	
Sr	209.00	110.00	64.00	49.00	243.00	82.00	118.00	149.00	79.00	156.00	114.00	189.00	111.00	74.00	169.00	107.00	79.00	
Y	32.00	29.00	28.00	34.00	28.00	33.00	29.00	22.00	24.00	21.00	26.00	29.00	28.00	33.00	29.00	26.00	14.00	
Zr	203.00	133.00	211.00	218.00	386.00	285.00	142.00	188.00	162.00	274.00	142.00	200.00	140.00	155.00	140.00	240.00	114.00	
Cr	90.00	110.00	90.00	80.00	60.00	90.00	100.00	80.00	80.00	50.00	90.00	70.00	100.00	100.00	90.00	70.00	40.00	
Ni	40.00	40.00	20.00	30.00	20.00	20.00	50.00	30.00	20.00	< 20	30.00	< 20	40.00	50.00	30.00	30.00	< 20	
Ga	25.00	32.00	23.00	20.00	19.00	21.00	29.00	22.00	23.00	16.00	25.00	19.00	26.00	23.00	28.00	21.00	22.00	
Rb	154.00	184.00	174.00	116.00	154.00	110.00	189.00	167.00	180.00	186.00	170.00	150.00	234.00	170.00	281.00	279.00	260.00	
Nb	15.00	14.00	11.00	12.00	13.00	12.00	15.00	13.00	11.00	8.00	12.00	10.00	14.00	11.00	12.00	16.00	19.00	
Cs	4.90	9.90	7.60	4.80	8.80	2.10	16.30	12.60	33.30	20.60	9.30	9.40	28.50	13.30	44.10	58.40	42.00	
La	51.00	53.90	42.80	46.40	50.90	38.60	61.10	34.70	34.50	43.60	46.40	38.20	45.90	44.70	46.90	40.80	17.70	
Ce	92.40	104.00	83.00	93.20	103.00	77.60	114.00	73.00	69.20	91.10	92.10	78.50	90.60	88.30	92.80	82.70	35.00	
Pr	11.30	11.90	10.30	10.70	11.40	9.28	13.20	8.22	8.26	9.82	10.90	8.68	10.00	10.30	10.40	9.44	4.03	
Nd	40.10	43.30	37.90	40.80	41.60	34.30	46.60	31.00	30.00	35.50	40.90	31.60	37.50	38.40	38.40	34.00	14.00	
Sm	7.60	8.40	7.30	8.10	7.60	7.10	8.50	6.20	6.10	6.60	7.90	6.40	7.00	7.40	7.40	6.80	2.90	
Eu	1.55	1.69	1.41	1.65	1.67	1.42	1.67	1.23	1.22	1.10	1.50	1.36	1.39	1.47	1.53	1.10	0.54	
Gd	6.30	6.80	6.00	7.40	6.30	6.70	6.60	5.20	5.40	5.50	6.60	5.60	6.10	6.80	6.10	5.50	2.50	
Tb	1.00	1.00	0.90	1.20	1.00	1.10	1.00	0.80	0.90	0.80	1.00	0.90	1.00	1.00	1.00	0.90	0.50	
Dy	6.00	5.90	5.50	6.90	5.70	6.30	6.00	4.50	5.00	4.50	5.60	5.60	5.40	6.00	5.70	5.00	2.70	
Ho	1.10	1.10	1.10	1.30	1.10	1.20	1.10	0.80	0.90	0.80	1.00	1.10	1.00	1.20	1.10	1.00	0.50	
Er	3.30	3.20	3.10	3.70	3.10	3.70	3.10	2.50	2.80	2.30	2.90	3.40	3.00	3.40	3.10	2.80	1.40	
Tm	0.47	0.48	0.46	0.54	0.45	0.53	0.46	0.37	0.41	0.33	0.40	0.51	0.43	0.49	0.43	0.41	0.21	
Yb	3.10	3.20	3.10	3.50	3.10	3.50	2.90	2.60	2.90	2.20	2.70	3.40	2.90	3.20	2.90	2.80	1.40	
Lu	0.46	0.47	0.46	0.51	0.47	0.55	0.45	0.40	0.45	0.34	0.42	0.53	0.43	0.51	0.44	0.43	0.22	
Hf	5.40	3.60	5.20	5.50	9.30	7.00	3.70	4.80	4.20	6.00	3.50	5.00	3.80	3.80	3.80	6.10	2.80	
Ta	1.30	1.20	1.10	1.30	1.20	1.00	1.20	1.50	2.10	1.00	0.90	1.10	1.40	0.90	1.00	2.40	4.90	
Pb	23.00	21.00	10.00	11.00	22.00	15.00	25.00	20.00	17.00	28.00	21.00	29.00	22.00	22.00	25.00	17.00	21.00	
Th	17.10	16.90	12.00	10.20	16.70	13.40	17.20	10.80	10.50	15.30	12.70	11.90	13.20	12.20	13.20	12.80	5.50	
U	3.50	6.50	3.60	3.80	3.20	3.80	4.20	5.10	6.60	3.50	4.50	3.90	6.40	43.70	9.40	7.10	3.80	
REE	225.68	245.34	203.33	225.90	237.39	191.88	266.68	171.52	168.04	204.49	220.32	185.78	212.65	213.17	218.20	193.68	83.60	

LREE	202.40	221.50	181.30	199.20	214.50	166.88	243.40	153.12	148.06	186.62	198.20	163.38	191.00	189.10	195.90	173.74	73.63
HREE	23.28	23.84	22.03	26.70	22.89	25.00	23.28	18.40	19.98	17.87	22.12	22.40	21.65	24.07	22.30	19.94	9.97
LREE/HREE	8.69	9.29	8.23	7.46	9.37	6.68	10.46	8.32	7.41	10.44	8.96	7.29	8.82	7.86	8.78	8.71	7.39
La/Y _N	11.08	11.34	9.30	8.93	11.06	7.43	14.19	8.99	8.01	13.35	11.57	7.57	10.66	9.41	10.89	9.81	8.51
La/Sm _N	4.22	4.03	3.69	3.60	4.21	3.42	4.52	3.52	3.56	4.15	3.69	3.75	4.12	3.80	3.98	3.77	3.84
Dy/Y _N	1.26	1.20	1.15	1.28	1.19	1.17	1.34	1.12	1.12	1.33	1.35	1.07	1.21	1.22	1.28	1.16	1.25
Tb/Y _N	1.42	1.38	1.28	1.51	1.42	1.38	1.52	1.35	1.37	1.60	1.63	1.16	1.52	1.38	1.52	1.41	1.57
Eu/Eu*	0.68	0.68	0.65	0.65	0.74	0.63	0.68	0.66	0.65	0.56	0.63	0.69	0.65	0.63	0.70	0.55	0.61

Table 2 continued

Lithology	Type-1 Diatexite								Type-2 Diatexite							
	Sample	JTJ-27	JTJ-30	JTJ-31C	JTJ-43	JTJ-48A	JTJ-52	JTJ-55	JPC-48	JTJ-23A	JTJ-24A	JTJ-26	JTJ-28A	JTJ-51	JTJ-56B	JTJ-60B
SiO ₂	72.92	72.29	74.46	68.32	72.75	72.33	73.30	69.83	73.53	74.24	74.82	75.14	75.24	74.98	72.95	74.55
Al ₂ O ₃	13.92	14.17	14.07	15.60	15.37	14.94	14.72	15.62	14.35	13.93	13.27	14.96	12.87	12.91	15.25	14.11
Fe ₂ O ₃	0.97	1.38	0.90	2.90	1.29	1.52	1.13	2.15	1.12	0.75	0.81	0.18	2.67	1.01	0.98	0.81
FeO ¹	0.87	1.24	0.81	2.61	1.16	1.37	1.02	1.93	1.01	0.67	0.73	0.16	2.40	0.91	0.88	0.73
MnO	0.01	0.01	0.02	0.03	0.03	0.02	0.03	0.03	0.03	0.03	0.01	0.01	0.05	0.03	0.02	0.02
MgO	0.21	0.31	0.25	0.84	0.27	0.41	0.29	0.67	0.18	0.17	0.16	0.06	0.90	0.36	0.49	0.44
CaO	0.59	0.63	0.54	0.95	0.61	0.74	0.88	0.62	0.46	0.59	0.60	0.42	0.65	0.48	0.44	0.21
Na ₂ O	3.07	3.12	3.13	2.80	3.39	2.98	3.56	2.91	3.57	3.66	3.15	2.57	2.31	3.16	3.89	2.96
K ₂ O	5.31	5.64	5.50	5.71	4.82	5.44	4.18	5.93	4.28	4.97	4.50	6.16	3.77	4.55	3.91	5.27
TiO ₂	0.12	0.20	0.13	0.75	0.19	0.24	0.15	0.40	0.07	0.07	0.11	0.02	0.39	0.12	0.10	0.12
P ₂ O ₅	0.36	0.47	0.30	0.44	0.42	0.39	0.25	0.40	0.24	0.25	0.18	0.34	0.14	0.17	0.24	0.20
LOI	1.10	0.93	0.83	1.41	1.05	1.15	1.21	1.23	1.35	0.72	0.98	0.81	1.47	1.50	1.20	1.19
Total	98.58	99.14	100.10	99.74	100.20	100.20	99.68	101.72	99.18	99.38	98.58	100.70	100.50	99.25	99.46	100.61
Sc	4.00	2.00	1.00	3.00	2.00	3.00	3.00	<7	4.00	3.00	5.00	<1	8.00	3.00	3.00	<7
V	<5	<5	<5	29.00	6.00	8.00	<5	13.00	<5	<5	<5	<5	38.00	9.00	9.00	5.00
Ba	196.00	217.00	232.00	640.00	205.00	282.00	468.00	353.00	146.00	203.00	200.00	358.00	582.00	482.00	187.00	220.00
Sr	57.00	56.00	61.00	150.00	50.00	74.00	139.00	89.00	43.00	79.00	63.00	111.00	128.00	115.00	69.00	82.00
Y	12.00	11.00	4.00	11.00	8.00	12.00	9.00	14.00	9.00	8.00	12.00	4.00	24.00	9.00	7.00	10.00
Zr	47.00	79.00	45.00	415.00	74.00	100.00	48.00	163.00	28.00	42.00	35.00	25.00	159.00	42.00	39.00	37.00
Cr	<20	<20	<20	<20	<20	<20	<20	19.00	<20	<20	<20	<20	40.00	<20	<20	6.00
Ni	<20	<20	<20	<20	<20	<20	<20	7.00	<20	<20	<20	<20	<20	<20	<20	<7
Ga	18.00	21.00	19.00	26.00	24.00	20.00	17.00	23.00	16.00	12.00	16.00	17.00	16.00	12.00	15.00	14.00
Rb	239.00	303.00	275.00	331.00	315.00	282.00	143.00	290.00	238.00	170.00	183.00	234.00	165.00	154.00	145.00	205.00
Nb	6.00	7.00	4.00	6.00	10.00	7.00	6.00	10.37	9.00	4.00	3.00	<1	7.00	5.00	5.00	4.98
Cs	10.20	8.60	6.60	7.80	20.30	17.10	14.00	-	34.80	9.70	5.00	6.00	10.10	8.60	6.00	-
La	7.20	15.60	5.90	145.00	14.70	20.40	21.00	36.96	3.50	3.60	4.60	1.10	36.60	7.70	6.80	3.79
Ce	15.40	35.10	13.40	368.00	32.20	44.70	41.70	79.12	7.80	7.20	9.60	2.20	74.20	15.20	13.10	7.77
Pr	1.85	4.23	1.65	46.50	3.78	5.19	4.80	9.79	0.88	0.79	1.11	0.28	8.21	1.64	1.49	0.95
Nd	6.70	16.00	6.10	175.00	14.40	19.50	17.10	36.54	3.40	2.90	4.10	1.10	29.20	6.60	5.20	3.46
Sm	1.80	4.10	1.80	25.40	3.90	5.10	4.10	8.99	0.80	0.70	1.10	0.40	5.90	1.30	1.00	0.98
Eu	0.41	0.45	0.39	1.06	0.33	0.55	0.79	0.61	0.25	0.40	0.44	0.55	1.08	0.56	0.34	0.43
Gd	1.90	3.60	1.80	9.50	3.50	4.70	3.30	6.86	1.00	0.80	1.30	0.50	4.80	1.30	0.90	1.12
Tb	0.40	0.60	0.30	0.80	0.50	0.70	0.50	0.85	0.30	0.20	0.30	0.10	0.80	0.20	0.20	0.24
Dy	2.20	2.90	1.10	3.10	2.10	3.10	2.10	3.52	1.70	1.30	2.00	0.60	4.70	1.40	1.10	1.70

Ho	0.40	0.40	0.20	0.40	0.20	0.40	0.30	0.49	0.30	0.30	0.40	0.10	0.90	0.20	0.20	0.35
Er	1.10	0.70	0.30	0.90	0.50	0.80	0.60	1.10	0.90	0.80	1.20	0.30	2.50	0.70	0.70	1.07
Tm	0.14	0.08	0.05	0.12	0.06	0.11	0.08	0.15	0.14	0.14	0.20	0.05	0.37	0.10	0.11	0.18
Yb	0.90	0.40	0.20	0.70	0.40	0.60	0.50	1.00	1.00	1.10	1.40	0.30	2.30	0.80	0.90	1.27
Lu	0.12	0.01	0.01	0.09	0.05	0.10	0.06	0.14	0.15	0.17	0.22	0.04	0.36	0.13	0.13	0.20
Hf	1.40	2.10	1.30	10.00	2.10	2.80	1.50	2.32	0.80	1.20	0.90	0.90	4.10	1.00	1.30	<0.06
Ta	0.70	0.80	0.60	0.50	1.90	0.90	1.10	1.03	1.90	1.30	0.30	< 0.1	0.90	1.00	1.40	0.37
Pb	32.00	33.00	33.00	41.00	27.00	36.00	36.00	39.00	36.00	41.00	31.00	43.00	28.00	37.00	28.00	38.00
Th	2.50	7.50	2.90	133.00	6.90	8.10	6.90	20.62	0.90	0.90	1.30	< 0.1	12.30	2.30	1.80	1.54
U	5.50	9.30	6.10	10.30	10.00	7.80	5.90	13.48	6.80	12.80	3.30	1.50	5.40	3.00	7.10	3.38
REE	40.52	84.17	33.20	776.57	76.62	105.95	96.93	186.13	22.12	20.40	27.97	7.62	171.92	37.83	32.17	23.52
LREE	32.95	75.03	28.85	759.90	68.98	94.89	88.70	171.40	16.38	15.19	20.51	5.08	154.11	32.44	27.59	16.96
HREE	7.57	9.14	4.35	16.67	7.64	11.06	8.23	14.73	5.74	5.21	7.46	2.54	17.81	5.39	4.58	6.57
LREE/HREE	4.35	8.21	6.63	45.58	9.03	8.58	10.78	11.64	2.85	2.92	2.75	2.00	8.65	6.02	6.02	2.58
La/Yb _N	5.39	26.27	19.87	139.50	24.75	22.90	28.29	24.80	2.36	2.20	2.21	2.47	10.72	6.48	5.09	2.01
La/Sm _N	2.51	2.39	2.06	3.59	2.37	2.51	3.22	2.58	2.75	3.23	2.63	1.73	3.90	3.72	4.27	2.43
Dy/Yb _N	1.59	4.71	3.57	2.88	3.41	3.36	2.73	2.28	1.10	0.77	0.93	1.30	1.33	1.14	0.79	0.87
Tb/Yb _N	1.96	6.60	6.60	5.03	5.50	5.13	4.40	3.72	1.32	0.80	0.94	1.47	1.53	1.10	0.98	0.83
Eu/Eu*	0.68	0.36	0.66	0.21	0.27	0.34	0.66	0.24	0.85	1.63	1.12	3.76	0.62	1.32	1.10	1.26

Table 2 continued

Lithology	Granite																
	Sample	JTJ-31A	JTJ-32	JTJ-34	JTJ-40	JPC-49	JPC-56	JPC-57	JTJ-9	JTJ-10	JTJ-44	JTJ-45	JTJ-47	JTJ-50	JTJ-62	JPC-51	JPC-52
SiO ₂	73.04	71.18	70.97	74.07	71.66	72.13	71.99	73.37	71.55	71.54	73.60	70.78	72.76	73.31	71.53	71.37	
Al ₂ O ₃	15.07	14.88	15.14	14.33	14.95	14.88	15.02	14.56	14.70	15.20	14.40	14.89	14.50	15.40	14.92	15.19	
Fe ₂ O ₃ ^I	1.04	1.91	2.36	0.51	1.94	1.56	1.50	1.28	1.45	1.56	1.36	1.90	1.32	0.92	1.48	1.41	
FeO ^I	0.94	1.72	2.12	0.46	1.75	1.40	1.35	1.15	1.30	1.40	1.22	1.71	1.19	0.83	1.33	1.27	
MnO	0.02	0.03	0.02	0.01	0.02	0.02	0.02	0.02	0.03	0.02	0.02	0.03	0.03	0.04	0.02	0.02	
MgO	0.26	0.46	0.57	0.09	0.60	0.65	0.65	0.25	0.29	0.43	0.32	0.48	0.23	0.10	0.71	0.60	
CaO	0.55	0.71	0.88	0.34	0.45	0.36	0.36	0.53	0.62	0.60	0.55	0.71	0.57	0.44	0.34	0.33	
Na ₂ O	2.90	2.51	2.61	1.99	2.67	3.14	3.03	2.98	3.30	3.04	2.79	2.83	3.15	3.92	3.87	3.04	
K ₂ O	5.24	5.43	5.98	7.57	5.61	5.38	5.54	4.99	5.02	5.34	5.28	5.56	5.01	4.05	5.47	6.09	
TiO ₂	0.15	0.29	0.54	0.06	0.32	0.27	0.25	0.19	0.19	0.27	0.21	0.33	0.16	0.08	0.22	0.27	
P ₂ O ₅	0.38	0.48	0.41	0.55	0.33	0.37	0.39	0.36	0.40	0.44	0.41	0.35	0.35	0.45	0.38	0.33	
LOI	0.94	1.28	1.20	0.90	1.34	1.09	1.13	1.61	0.96	1.25	1.30	1.11	0.96	1.23	0.98	1.19	
Total	99.59	99.15	100.70	100.40	101.64	101.25	101.23	100.10	98.52	99.67	100.20	98.96	99.01	99.95	101.25	101.11	
Sc	1.00	3.00	2.00	< 1	< 7	< 7	< 7	2.00	2.00	2.00	3.00	2.00	2.00	2.00	< 7	< 7	
V	8.00	10.00	17.00	< 5	11.00	7.00	8.00	7.00	5.00	11.00	7.00	13.00	< 5	< 5	7.00	8.00	
Ba	211.00	279.00	357.00	143.00	258.00	235.00	217.00	243.00	225.00	262.00	206.00	263.00	139.00	46.00	213.00	261.00	
Sr	60.00	68.00	84.00	54.00	69.00	67.00	62.00	57.00	57.00	64.00	54.00	80.00	48.00	27.00	62.00	79.00	
Y	5.00	17.00	10.00	9.00	12.00	10.00	12.00	7.00	7.00	10.00	7.00	7.00	6.00	6.00	10.00	9.00	
Zr	57.00	137.00	287.00	23.00	129.00	108.00	104.00	64.00	70.00	106.00	84.00	117.00	64.00	40.00	96.00	114.00	
Cr	< 20	< 20	< 20	< 20	8.00	8.00	7.00	< 20	< 20	< 20	< 20	< 20	< 20	< 20	5.00	5.00	
Ni	< 20	< 20	< 20	< 20	< 7	< 7	< 7	< 20	< 20	< 20	< 20	< 20	< 20	< 20	< 7	< 7	
Ga	23.00	24.00	27.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	27.00	23.00	26.00	20.00	23.00	
Rb	325.00	316.00	376.00	401.00	284.00	343.00	311.00	311.00	342.00	330.00	338.00	416.00	419.00	521.00	339.00	371.00	

Nb	5.00	9.00	7.00	3.00	10.54	11.26	11.68	9.00	9.00	7.00	9.00	9.00	14.00	15.00	10.08	8.87
Cs	20.30	18.40	11.50	13.60	-	-	-	13.30	27.00	15.20	18.90	17.80	26.60	37.50	-	-
La	10.80	31.80	97.30	2.90	37.13	24.76	22.92	15.50	15.90	21.40	19.80	36.00	15.00	8.00	21.82	31.72
Ce	25.00	75.50	248.00	7.80	85.76	54.74	52.13	34.00	34.60	50.80	47.50	84.10	34.10	17.90	50.56	71.00
Pr	3.08	9.37	31.50	1.05	11.01	6.94	6.70	4.01	3.98	6.28	5.85	10.30	4.08	2.13	6.57	9.03
Nd	12.40	35.40	119.00	4.30	41.49	26.35	25.79	14.80	15.20	24.10	22.40	38.50	14.50	7.80	25.37	34.11
Sm	3.10	8.10	20.50	1.50	8.33	6.07	6.00	3.70	3.80	5.80	4.50	8.50	3.60	1.90	5.77	7.68
Eu	0.41	0.54	0.78	0.33	0.42	0.42	0.40	0.34	0.36	0.51	0.38	0.62	0.33	0.13	0.37	0.54
Gd	2.30	6.20	9.30	1.50	5.12	4.50	4.59	3.40	3.50	4.60	3.20	5.40	2.80	1.60	3.96	5.16
Tb	0.30	0.90	0.80	0.30	0.63	0.54	0.63	0.50	0.50	0.60	0.40	0.60	0.40	0.30	0.51	0.58
Dy	1.20	4.00	2.90	1.70	2.75	2.18	2.82	1.90	2.00	2.40	1.80	2.00	1.80	1.30	2.29	2.18
Ho	0.20	0.60	0.30	0.30	0.39	0.29	0.38	0.20	0.20	0.30	0.20	0.20	0.20	0.20	0.32	0.29
Er	0.40	1.30	0.70	0.70	0.84	0.58	0.78	0.40	0.40	0.60	0.60	0.60	0.50	0.50	0.70	0.56
Tm	0.05	0.15	0.09	0.08	0.09	0.07	0.10	0.06	0.05	0.07	0.07	0.06	0.06	0.06	0.08	0.07
Yb	0.30	0.80	0.50	0.50	0.60	0.38	0.55	0.30	0.30	0.40	0.40	0.30	0.30	0.40	0.47	0.43
Lu	0.01	0.10	0.07	0.06	0.08	0.05	0.06	0.04	0.04	0.06	0.06	0.04	0.04	0.06	0.06	0.06
Hf	1.50	3.70	7.00	1.00	1.55	<0.06	1.50	1.70	1.90	2.80	2.20	3.10	1.80	1.50	1.31	3.02
Ta	0.90	1.10	0.80	0.40	0.58	1.14	1.14	1.80	1.90	1.20	1.50	1.00	3.30	4.90	1.20	0.86
Pb	28.00	36.00	39.00	43.00	38.00	35.00	35.00	27.00	27.00	31.00	29.00	36.00	32.00	16.00	36.00	43.00
Th	6.80	22.20	107.00	1.20	39.25	22.16	20.21	6.90	7.20	13.90	15.60	28.50	9.10	5.10	20.86	33.69
U	6.70	12.60	10.70	3.50	16.30	11.93	14.03	5.10	9.40	8.70	11.10	15.00	13.70	4.50	10.69	5.84
REE	59.55	174.76	531.74	23.02	194.63	127.86	123.85	79.15	80.83	117.92	107.16	187.22	77.71	42.28	118.85	163.41
LREE	54.38	160.17	516.30	17.55	183.71	118.86	113.54	72.01	73.48	108.38	100.05	177.40	71.28	37.73	110.09	153.54
HREE	5.17	14.59	15.44	5.47	10.92	9.00	10.31	7.14	7.35	9.54	7.11	9.82	6.43	4.55	8.76	9.88
LREE/HREE	10.52	10.98	33.44	3.21	16.83	13.20	11.01	10.09	10.00	11.36	14.07	18.07	11.09	8.29	12.57	15.55
La/Yb _N	24.24	26.77	131.06	3.91	41.48	43.79	27.82	34.80	35.69	36.03	33.34	80.82	33.67	13.47	31.35	49.25
La/Sm _N	2.19	2.47	2.98	1.22	2.80	2.56	2.40	2.63	2.63	2.32	2.77	2.66	2.62	2.65	2.38	2.59
Dy/Yb _N	2.60	3.25	3.77	2.21	2.96	3.72	3.30	4.11	4.33	3.90	2.92	4.33	3.90	2.11	3.18	3.27
Tb/Yb _N	4.40	4.95	7.04	2.64	4.57	6.24	5.00	7.33	7.33	6.60	4.40	8.80	5.87	3.30	4.82	5.83
Eu/Eu*	0.47	0.23	0.17	0.67	0.20	0.24	0.24	0.29	0.30	0.30	0.31	0.28	0.32	0.23	0.24	0.26

Table 2 continued

Lithology	Granite														
	Sample	JPC-53	JTJ-35	JPC-50	JTJ-11	JTJ-12	JPC-55	JTJ-59	JTJ-13	JPC-19	JPC-24	JPC-59	JPC-54	JTJ-36	JTJ-58
SiO ₂	71.62	70.63	71.41	71.13	71.39	70.32	71.96	71.78	70.64	72.31	75.24	73.33	72.64	73.41	
Al ₂ O ₃	15.20	15.10	15.05	14.71	15.01	15.51	14.50	14.60	15.15	14.64	13.95	14.72	14.91	14.81	
Fe ₂ O ₃ ^I	1.72	2.14	1.51	1.64	1.84	2.15	1.66	1.63	1.73	1.89	0.70	1.19	1.34	0.82	
FeO ^I	1.55	1.93	1.36	1.48	1.66	1.93	1.49	1.47	1.56	1.70	0.63	1.07	1.21	0.74	
MnO	0.02	0.02	0.02	0.02	0.02	0.03	0.04	0.02	0.02	0.03	0.02	0.02	0.02	0.03	
MgO	0.70	0.59	0.55	0.43	0.41	0.75	0.54	0.36	0.63	0.64	0.36	0.48	0.31	0.19	
CaO	0.40	0.72	0.26	0.73	0.61	0.58	1.01	0.58	0.46	0.37	0.09	0.24	0.63	0.55	
Na ₂ O	3.02	2.57	2.99	3.06	2.72	3.16	3.54	2.66	2.63	2.80	3.48	2.98	3.25	3.27	
K ₂ O	5.53	5.37	5.93	5.43	5.39	5.32	4.86	5.71	6.91	5.39	4.33	5.20	4.97	4.59	
TiO ₂	0.31	0.44	0.22	0.28	0.28	0.42	0.28	0.28	0.40	0.31	0.10	0.19	0.19	0.10	
P ₂ O ₅	0.37	0.48	0.47	0.37	0.40	0.39	0.29	0.49	0.31	0.36	0.30	0.30	0.44	0.35	
LOI	0.95	1.51	1.44	0.94	1.06	1.21	1.02	1.21	0.84	1.01	1.32	1.21	1.11	1.67	

Total	101.39	99.58	101.21	98.71	99.14	101.77	99.70	99.31	101.28	101.45	100.52	100.93	99.82	99.78
Sc	<7	3.00	<7	2.00	2.00	<7	3.00	2.00	<7	<7	<7	<7	2.00	3.00
V	8.00	19.00	<5	11.00	9.00	18.00	17.00	10.00	8.00	9.00	<5	5.00	7.00	< 5
Ba	240.00	376.00	183.00	283.00	235.00	287.00	491.00	268.00	407.00	172.00	134.00	146.00	274.00	201.00
Sr	76.00	93.00	47.00	80.00	63.00	74.00	142.00	76.00	106.00	49.00	42.00	46.00	74.00	61.00
Y	10.00	10.00	14.00	9.00	11.00	10.00	8.00	8.00	9.00	11.00	5.00	6.00	9.00	7.00
Zr	123.00	211.00	86.00	112.00	116.00	198.00	130.00	113.00	193.00	130.00	35.00	72.00	74.00	46.00
Cr	9.00	< 20	6.00	< 20	< 20	11.00	< 20	< 20	6.00	7.00	<5	5.00	< 20	< 20
Ni	<7	< 20	<7	< 20	< 20	<7	< 20	< 20	<7	<7	<7	<7	< 20	< 20
Ga	24.00	25.00	23.00	22.00	22.00	25.00	23.00	25.00	23.00	24.00	20.00	22.00	23.00	18.00
Rb	371.00	358.00	403.00	334.00	357.00	369.00	276.00	381.00	305.00	378.00	277.00	293.00	288.00	281.00
Nb	11.61	7.00	14.99	8.00	10.00	9.94	10.00	8.00	4.32	15.97	12.91	7.92	10.00	9.00
Cs	-	10.60	-	14.80	18.30	-	19.10	13.00	-	-	-	-	17.00	18.90
La	33.65	65.40	16.95	26.60	30.40	64.65	38.80	31.20	61.40	37.77	5.03	14.98	12.40	8.50
Ce	75.62	162.00	39.29	60.00	72.70	151.70	77.60	72.00	135.11	87.83	11.04	33.71	26.90	17.70
Pr	9.74	20.40	5.00	7.08	8.82	19.91	8.80	8.88	16.41	11.35	1.36	4.24	3.27	2.09
Nd	36.91	75.60	18.49	26.50	33.50	75.41	30.80	33.80	61.23	43.58	5.09	15.65	12.40	7.50
Sm	8.17	12.90	3.94	6.30	7.00	12.56	5.50	7.40	13.18	9.40	1.22	3.56	3.30	2.20
Eu	0.50	0.74	0.25	0.57	0.45	0.59	0.75	0.56	0.79	0.46	0.11	0.23	0.34	0.51
Gd	5.30	6.40	2.91	4.50	5.10	6.18	3.70	4.80	7.47	5.84	1.20	2.62	3.00	2.10
Tb	0.61	0.70	0.50	0.60	0.70	0.68	0.40	0.60	0.70	0.72	0.19	0.34	0.50	0.30
Dy	2.43	2.80	2.66	2.20	2.80	2.60	1.90	2.10	2.39	2.92	0.95	1.46	2.30	1.80
Ho	0.30	0.40	0.41	0.30	0.40	0.38	0.30	0.30	0.30	0.43	0.15	0.18	0.30	0.30
Er	0.60	0.80	0.91	0.60	0.80	0.82	0.60	0.60	0.69	0.95	0.32	0.40	0.60	0.60
Tm	0.07	0.11	0.11	0.07	0.10	0.10	0.09	0.07	0.06	0.11	0.04	0.04	0.08	0.08
Yb	0.39	0.60	0.64	0.40	0.50	0.62	0.60	0.30	0.46	0.66	0.27	0.26	0.50	0.50
Lu	0.05	0.09	0.08	0.06	0.07	0.08	0.09	0.05	0.05	0.07	0.04	0.03	0.06	0.07
Hf	2.21	5.10	0.48	2.80	3.10	5.09	3.40	2.90	2.18	2.09	1.06	<0.06	2.10	1.50
Ta	1.05	0.80	1.61	1.10	1.60	1.27	1.40	0.90	0.52	1.23	2.09	1.07	2.10	2.50
Pb	39.00	34.00	34.00	36.00	33.00	36.00	34.00	37.00	53.00	34.00	21.00	33.00	26.00	42.00
Th	32.41	57.10	21.08	14.00	22.30	66.13	17.00	24.60	38.52	36.34	2.48	14.02	4.90	3.40
U	19.28	11.60	12.44	9.90	10.70	13.69	10.30	10.60	10.11	13.20	3.67	13.15	5.70	5.80
REE	174.34	348.94	92.13	135.78	163.34	336.28	169.93	162.66	300.24	202.10	27.00	77.70	65.95	44.25
LREE	164.08	336.30	83.67	126.48	152.42	324.23	161.50	153.28	287.33	189.93	23.74	72.14	58.27	37.99
HREE	10.26	12.64	8.46	9.30	10.92	12.05	8.43	9.38	12.91	12.17	3.27	5.56	7.68	6.26
LREE/HREE	15.99	26.61	9.89	13.60	13.96	26.90	19.16	16.34	22.27	15.61	7.27	12.98	7.59	6.07
La/Yb _N	58.29	73.41	17.92	44.79	40.95	69.91	43.55	70.04	90.16	38.33	12.66	38.24	16.70	11.45
La/Sm _N	2.59	3.19	2.70	2.65	2.73	3.23	4.43	2.65	2.93	2.53	2.60	2.64	2.36	2.43
Dy/Yb _N	4.06	3.03	2.71	3.57	3.64	2.71	2.06	4.55	3.39	2.86	2.31	3.58	2.99	2.34
Tb/Yb _N	6.94	5.13	3.45	6.60	6.16	4.82	2.93	8.80	6.72	4.78	5.69	3.12	4.40	2.64
Eu/Eu*	0.23	0.25	0.23	0.33	0.23	0.20	0.51	0.29	0.24	0.19	0.28	0.23	0.33	0.73

Table 3

Lithology	Phyllite			Metatexite					Type-1 Diatexite				Type-2 Diatexite	Granite X	Granite IX	Granite VIII	Granite V	Granite III	Granite II
Sample	M1_2_1	M1_2_2	M1_2_3	M1_20_1	M2_25_1	M3_8_1	M4_60A_1	M5_53_1	M2_43_1	M2_43_2	M3_55_1	M3_55_2	M1_24A_1	M4_34_1	M3_10_1	M2_35_1	M1_12_1	M5_13_1	M4_36_1
Sr	317	1471	333	167	243	487	113	183	112	115	144	131	86	78	99	96	74	76	133
Y	854	480	697	1973	1915	837	1618	1457	847	852	1569	1634	2026	908	542	1026	1546	1083	212
Th	27	97	18	1	1	50	1	0	15	21	1	0	1	7	5	5	19	5	4
U	8	34	18	223	32	17	91	46	33	47	135	123	185	49	51	62	74	75	81
La	386	1585	897	162	167	405	159	126	496	538	192	169	214	391	330	344	380	340	276
Ce	1124	2019	1672	547	559	921	503	425	1848	1924	654	557	574	1404	879	1230	1289	1129	561
Pr	175	174	179	91	97	107	83	74	329	336	114	96	79	241	120	214	215	181	66
Nd	931	630	747	513	565	455	457	435	1760	1808	643	533	348	1282	514	1115	1114	899	259
Sm	253	94	143	216	245	120	190	195	491	502	321	261	142	432	168	377	414	363	67
Eu	21	21	19	17	26	29	16	17	10	10	14	14	11	10	17	13	8	11	22
Gd	252	98	158	301	350	145	255	278	315	321	444	361	194	334	160	318	392	345	59
Tb	33	16	22	60	67	27	50	52	39	39	82	69	48	45	28	47	66	54	10
Dy	169	100	130	386	410	166	316	320	181	184	404	378	341	205	124	226	327	255	46
Ho	29	18	25	77	78	31	61	60	29	30	55	58	67	31	15	35	49	36	6
Er	67	42	60	199	190	78	155	146	68	68	101	130	192	64	31	75	104	73	12
Tm	8	5	8	28	25	11	22	19	9	9	11	17	32	8	4	9	13	8	2
Yb	51	33	46	183	148	73	145	120	52	53	64	111	231	41	24	53	70	46	10
Lu	7	4	7	25	19	9	19	16	7	7	8	14	31	5	3	6	8	5	1
LREE	2870	4503	3638	1529	1632	2008	1391	1255	4924	5108	1923	1615	1357	3750	2011	3279	3413	2911	1229
HREE	638	338	474	1275	1311	569	1039	1028	709	720	1183	1153	1147	742	407	783	1036	833	168
REE	3508	4841	4112	2804	2943	2577	2431	2283	5633	5829	3106	2768	2504	4493	2418	4062	4448	3744	1397
Eu/Eu*	0.25	0.67	0.39	0.20	0.27	0.67	0.22	0.22	0.07	0.08	0.12	0.14	0.19	0.08	0.33	0.11	0.06	0.09	1.06
La/Yb _N	5.1	32.6	13.0	0.6	0.8	3.7	0.7	0.7	6.4	6.8	2.0	1.0	0.6	6.5	9.1	4.3	3.7	5.0	19.1
La/Sm _N	1.0	10.6	3.9	0.5	0.4	2.1	0.5	0.4	0.6	0.7	0.4	0.4	0.9	0.6	1.2	0.6	0.6	0.6	2.6
Dy/Yb _N	2.1	2.0	1.8	1.4	1.8	1.5	1.4	1.7	2.3	2.2	4.1	2.2	1.0	3.3	3.3	2.8	3.0	3.6	3.1

Table 4

Lithology	Metatexite										Type-1 Diatexite		Type-2 Diatexite	Granite X								Granite IX	Granite V
Sample	M1T_20_11c-1	M1T_20_25c-1	M1T_20_46c-1	M1T_20_57c-1	M1T_20_78c-1	M6T_25_1c-1	M6T_25_2c-1	M6T_25_3c-1	M6T_25_4c-1	M6T_25_5c-1	M6T_25_7c-1	M6T_55_10c-1	M6T_55_12c-1	M1T_24A_17c-1	M6T_34_10c-1	M6T_34_15c-1	M6T_34_23c-1	M6T_34_24c-1	M6T_34_26c-1	M6T_34_27c-1	M6T_34_3c-1	M6T_45_12_1c-1	M6T_12_1c-1
Hf	10140	11080	8180	12530	10000	14300	13290	13260	13300	14410	14360	12750	11060	11810	12850	11660	11990	12070	12430	12670	12650	10280	11950
U	1152	341	97	2171	851	1310	1137	1261	1378	1213	1239	475	1514	321	734	1258	1182	1603	1480	1000	190	898	2770
Th	419	288	57	520	253	4	3	6	4	4	5	28	117	37	84	314	284	398	7400	980	68	214	16300
Th/U	0.36	0.85	0.58	0.24	0.30	0.003	0.003	0.005	0.003	0.003	0.004	0.06	0.08	0.12	0.11	0.25	0.24	0.25	5.00	0.98	0.36	0.24	5.88
Y	-	-	677	-	120	-	-	-	-	-	2074	-	-	610	469	855	797	1199	985	592	799	-	-
La	56	0.2	<LOD	14	3	<LOD	<LOD	<LOD	<LOD	<LOD	1	0.1	11	0.2	0.1	2	4	22	731	64	<LOD	2	-
Ce	95	23	22	70	29	0.1	0.1	0.1	0.1	0.1	5	1	54	15	2	17	38	260	-	64	5	17	-
Pr	8	0.2	0.1	12	3	<LOD	<LOD	0.02	<LOD	<LOD	1	0.1	11	0.3	0.3	4	9	33	377	115	0.1	4	-
Nd	34	3	2	79	21	0.2	0.2	0.2	0.1	0.2	6	1	68	2	3	31	70	209	1920	950	1	27	-
Sm	18	6	3	57	18	1	1	1	1	1	5	2	47	2	6	26	50	90	1170	153	2	25	3440
Eu	4	1	1	15	4	0.2	0.2	0.2	0.2	0.2	2	0.2	9	0.5	0.1	1	2	5	7	2	0.1	1	18

Gd	57	30	15	131	54	12	10	12	12	10	24	14	144	12	23	49	70	106	450	89	13	75	1380
Tb	18	10	5	33	16	7	6	8	8	7	11	7	48	4	6	12	14	20	40	13	5	19	115
Dy	201	128	58	327	182	139	111	143	143	127	177	98	489	54	58	106	111	158	196	97	76	168	457
Ho	69	48	21	94	62	64	52	65	66	59	73	38	120	20	15	28	26	39	38	24	29	45	71
Er	309	230	105	391	288	388	320	394	407	354	419	188	369	101	55	104	91	136	124	89	131	180	200
Tm	63	49	23	77	62	112	94	112	118	103	115	44	55	23	10	19	16	24	21	17	26	35	31
Yb	534	439	215	647	552	1305	1090	1284	1368	1193	1307	434	381	225	80	146	118	178	157	132	214	296	231
Lu	104	92	48	122	116	297	248	290	309	269	295	91	64	48	14	25	20	30	27	23	39	58	38
LRE	212	32	27	232	73	1	1	1	1	1	19	4	190	19	11	80	170	614	4198	1346	8	75	3440
HR	1357	1026	490	1837	1334	2324	1930	2308	2430	2122	2423	915	1679	488	261	490	468	696	1058	486	533	877	2540
EE	1569	1059	517	2069	1407	2325	1931	2309	2432	2124	2442	919	1869	507	272	570	638	1310	5256	1832	542	952	5980
REE	1569	1059	517	2069	1407	2325	1931	2309	2432	2124	2442	919	1869	507	272	570	638	1310	5256	1832	542	952	5980
Yb/ Gd _N	11.64	18.19	17.93	6.11	12.73	138.1 4	140.3 8	127.7 1	144.8 1	141.1 5	67.50	38.07	3.27	23.07	4.25	3.65	2.08	2.08	0.43	1.83	19.91	4.89	0.21
Ce/ Yb _N	0.05	0.01	0.03	0.03	0.01	0.000 03	0.000 02	0.000 02	0.000 01	0.000 01	0.001	0.001	0.04	0.02	0.01	0.03	0.08	0.4	-	0.1	0.01	0.01	-
Eu/ Eu*	0.34	0.30	0.51	0.52	0.43	0.17	0.17	0.16	0.17	0.17	0.44	0.14	0.33	0.26	0.03	0.09	0.12	0.16	0.03	0.04	0.04	0.07	0.02
Ce/ Ce*	1.09	26.09	-	1.30	2.49	-	-	-	-	-	1.15	3.77	1.22	14.55	2.43	1.39	1.52	2.34	-	0.18	-	1.43	-

Table 5

Lithology	Phyllite N	Phyllite S	Metatexite					Type-1 Diatexite		Type-2 Diatexite	Granite X	Granite IX	Granite VIII	Granite V	Granite II
Sample	JTJ-2	JTJ-19	JTJ-20	JTJ-25	JTJ-48B	JTJ-53	JTJ-60A	JTJ-43	JTJ-55	JTJ-24A	JTJ-34	JTJ-45	JTJ-35	JTJ-12	JTJ-36
⁸⁷ Sr/ ⁸⁶ Sr	0.7250	0.7402	0.7245	0.7321	0.7318	0.7403	0.7489	0.7364	0.7340	0.7447	0.7648	0.7893	0.7636	0.7774	0.7578
±2σ	0.000013	0.000016	0.000019	0.000016	0.000013	0.000017	0.000034	0.000018	0.000065	0.000020	0.000015	0.000024	0.000025	0.000013	0.000021
⁸⁷ Sr/ ⁸⁶ Sr ₃₂₀	0.7153	0.7089	0.7097	0.7124	0.7039	0.7099	0.7053	0.7072	0.7204	0.7162	0.7055	0.7062	0.7126	0.7022	0.7063
¹⁴³ Nd/ ¹⁴⁴ Nd	0.5119	0.5126	0.5122	0.5121	0.5120	0.5121	0.5121	0.5120	0.5123	0.5122	0.5120	0.5121	0.5120	0.5120	0.5123
±2σ	0.000011	0.000006	0.000009	0.000010	0.000007	0.000009	0.000009	0.000009	0.000011	0.000007	0.000009	0.000008	0.000009	0.000012	0.000012
εNd	-13.42	0.15	-8.30	-10.91	-12.28	-10.57	-10.99	-12.77	-7.16	-8.28	-12.66	-11.06	-12.18	-11.51	-7.49
εNd ₃₂₀	-10.07	3.28	-5.21	-7.65	-8.86	-7.30	-8.08	-8.32	-5.05	-6.21	-8.89	-7.99	-8.36	-8.64	-6.03
Nd T _{DM2} (Ma) ³	1826	779	1445	1637	1731	1609	1670	1524	1690	1433	1714	1734	1693	1664	1510
¹⁷⁶ Hf/ ¹⁷⁷ Hf	0.2823	0.2824	0.2825	0.2824	0.2824	0.2825	0.2824	0.2824	0.2825	0.2824	0.2823	0.2824	0.2824	0.2824	0.2825
±2σ	0.000005	0.000005	0.000004	0.000009	0.000005	0.000004	0.000005	0.000005	0.000005	0.000007	0.000006	0.000006	0.000005	0.000004	0.000005
εHf	-18.14	-11.40	-8.98	-11.41	-12.09	-9.44	-12.82	-14.92	-9.25	-12.02	-15.16	-13.99	-14.21	-14.06	-10.16
εHf ₃₂₀	-13.68	-7.16	-4.45	-7.99	-8.47	-6.45	-8.16	-8.16	-3.42	-9.25	-8.43	-7.78	-7.71	-7.70	-3.99

Hf T _{DM2} (Ma) ^b	2160	1747	1574	1800	1830	1701	1810	1880	1810	1508	1781	1828	1781	1786	1544
²⁰⁸ Pb/ ²⁰⁴ Pb	38.7956	38.6236	38.6820	38.8925	39.2638	38.6856	38.4898	39.9730	38.4068	38.2375	42.1802	38.6802	39.2789	38.9877	38.3497
±2σ	0.0021	0.0023	0.0023	0.0021	0.0018	0.0013	0.0022	0.0022	0.0026	0.0024	0.0022	0.0021	0.0017	0.0014	0.0023
²⁰⁶ Pb/ ²⁰⁴ Pb	18.3550	19.0401	18.7160	18.6505	18.6196	19.0338	18.5718	18.5110	18.4268	18.2206	19.0352	18.6156	18.6437	18.6888	18.6227
±2σ	0.0009	0.0010	0.0011	0.0010	0.0009	0.0005	0.0011	0.0010	0.0011	0.0009	0.0009	0.0010	0.0007	0.0006	0.0009
²⁰⁷ Pb/ ²⁰⁴ Pb	15.6924	15.6746	15.6434	15.6434	15.6485	15.6848	15.6502	15.6483	15.6407	15.6246	15.6830	15.6541	15.6583	15.6631	15.6618
±2σ	0.0008	0.0008	0.0009	0.0008	0.0007	0.0005	0.0009	0.0008	0.0010	0.0008	0.0008	0.0008	0.0007	0.0005	0.0008
²⁰⁸ Pb/ ²⁰⁶ Pb	2.1136	2.0285	2.0668	2.0853	2.1087	2.0325	2.0724	2.1594	2.0843	2.0986	2.2159	2.0778	2.1068	2.0862	2.0593
±2σ	0.000029	0.000033	0.000037	0.000046	0.000034	0.000027	0.000037	0.000041	0.000032	0.000037	0.000030	0.000046	0.000039	0.000045	0.000086
²⁰⁷ Pb/ ²⁰⁶ Pb	0.8549	0.8232	0.8358	0.8388	0.8404	0.8241	0.8427	0.8453	0.8488	0.8575	0.8239	0.8409	0.8399	0.8381	0.8410
±2σ	0.000011	0.000011	0.000013	0.000016	0.000010	0.000008	0.000011	0.000014	0.000010	0.000010	0.000010	0.000015	0.000010	0.000015	0.000024
²⁰⁶ Pb/ ²⁰⁴ Pb ₃₂₀	17.8770	17.9549	17.9149	17.9774	17.7057	12.7938	18.0033	17.7218	17.9120	17.2398	18.1733	17.4132	17.5719	17.6702	17.9340
²⁰⁷ Pb/ ²⁰⁴ Pb ₃₂₀	15.6671	15.6173	15.6011	15.6079	15.6002	15.3554	15.6201	15.6067	15.6135	15.5728	15.6375	15.5907	15.6017	15.6093	15.6254
²⁰⁸ Pb/ ²⁰⁴ Pb ₃₂₀	38.0388	37.6798	38.1323	38.2769	38.6531	38.1212	38.2232	36.6712	38.2117	38.2151	39.3876	38.1326	37.5695	38.2999	38.1579

$${}^a T_{DM2} = 1/\lambda \times \ln\left(\frac{{}^{143}\text{Nd}/{}^{144}\text{Nd} - (e^{-\lambda t} - 1) \times ({}^{147}\text{Sm}/{}^{144}\text{Nd} - {}^{147}\text{Sm}/{}^{144}\text{Nd}_{CC}) - {}^{143}\text{Nd}/{}^{144}\text{Nd}_{DM}}{({}^{147}\text{Sm}/{}^{144}\text{Nd}_{CC} - {}^{147}\text{Sm}/{}^{144}\text{Nd}_{DM}) + 1}\right)$$

$${}^b T_{DM2} = 1/\lambda \times \ln\left(\frac{{}^{176}\text{Hf}/{}^{177}\text{Hf} - (e^{-\lambda t} - 1) \times ({}^{176}\text{Lu}/{}^{177}\text{Hf} - {}^{176}\text{Lu}/{}^{177}\text{Hf}_{CC}) - {}^{176}\text{Hf}/{}^{177}\text{Hf}_{DM}}{({}^{176}\text{Lu}/{}^{177}\text{Hf}_{CC} - {}^{176}\text{Lu}/{}^{177}\text{Hf}_{DM}) + 1}\right)$$

λRb - Steiger and Jäger (1977)

λSm - Lugmair and Marti (1978)

λLu - Söderlund et al. (2004)

λU and λTh - Steiger and Jäger (1977)

¹⁴⁷Sm/¹⁴⁴Nd_{CHUR} and ¹⁴³Nd/¹⁴⁴Nd_{CHUR} - Jacobsen and Wasserburg (1980)

¹⁴⁷Sm/¹⁴⁴Nd_{DM}, ¹⁴³Nd/¹⁴⁴Nd_{DM}, and ¹⁴⁷Sm/¹⁴⁴Nd_{CC} - Liew and Hofmann (1988)

¹⁷⁶Lu/¹⁷⁷Hf_{CHUR} and ¹⁷⁶Hf/¹⁷⁷Hf_{CHUR} - Blichert-Toft and Albarède (1997)

¹⁷⁶Lu/¹⁷⁷Hf_{DM} and ¹⁷⁶Hf/¹⁷⁷Hf_{DM} - Vervoort and Blichert-Toft (1999)

¹⁷⁶Lu/¹⁷⁷Hf_{CC} - Griffin et al. (2002)

Fig.1

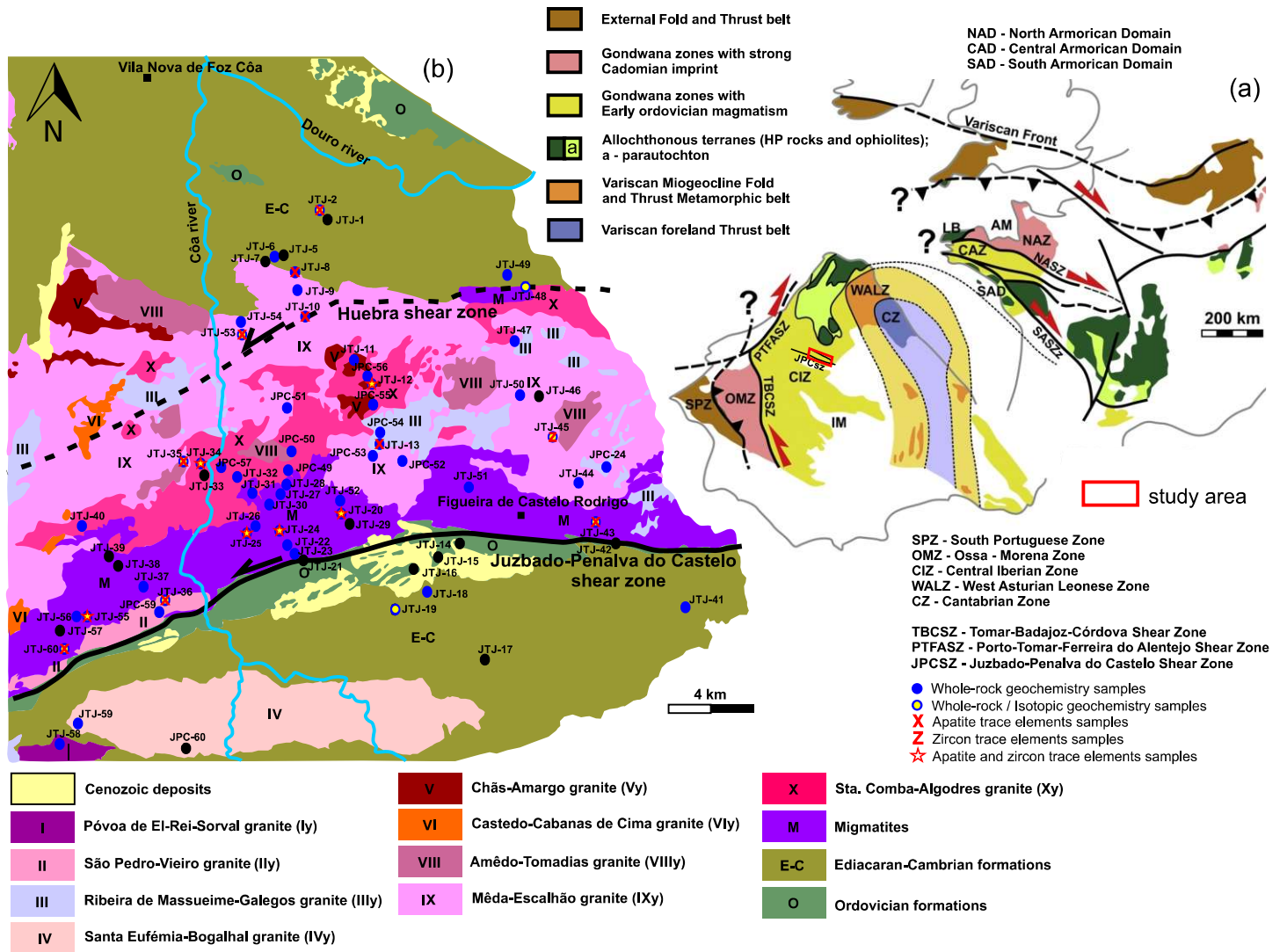


Fig.2

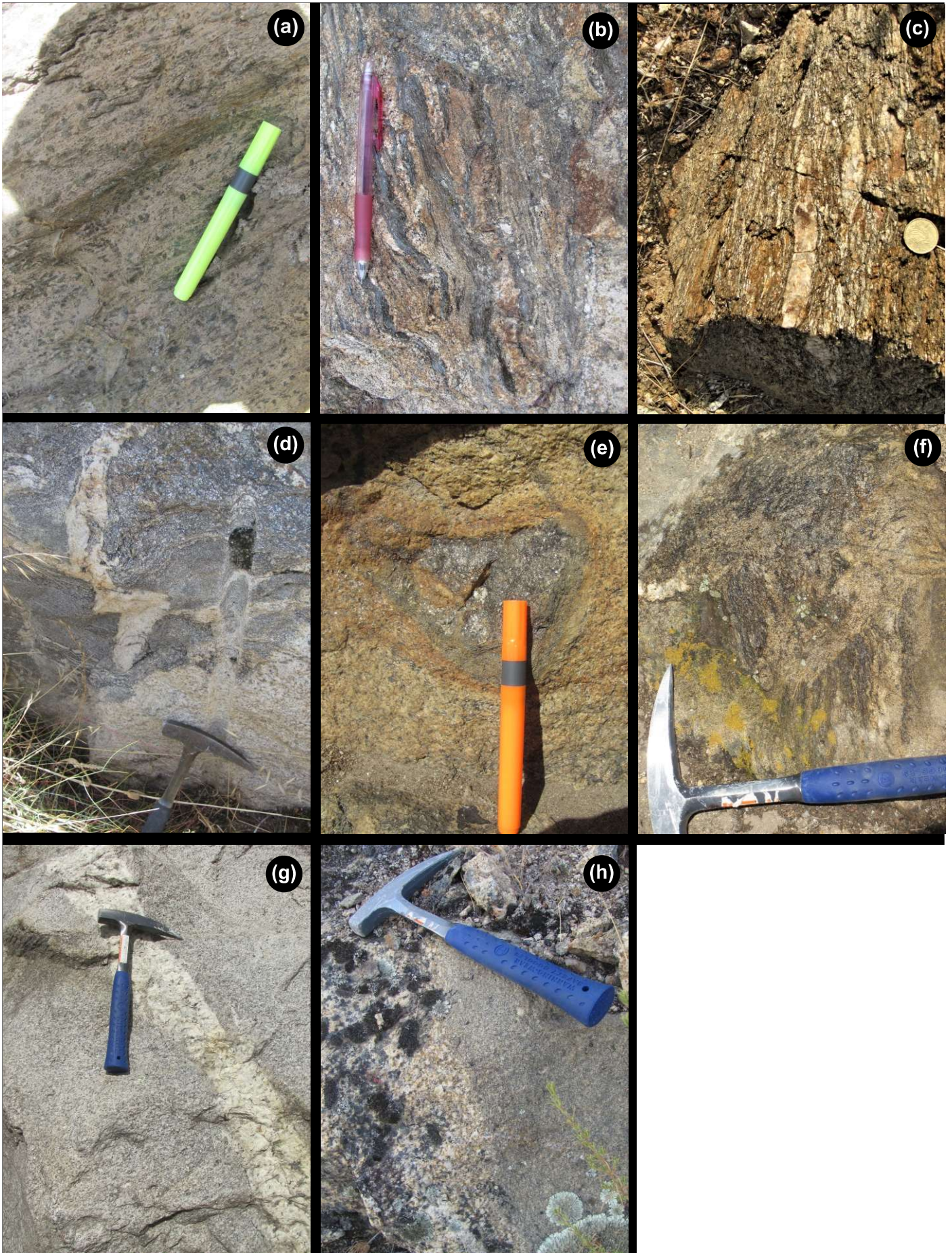


Fig.3

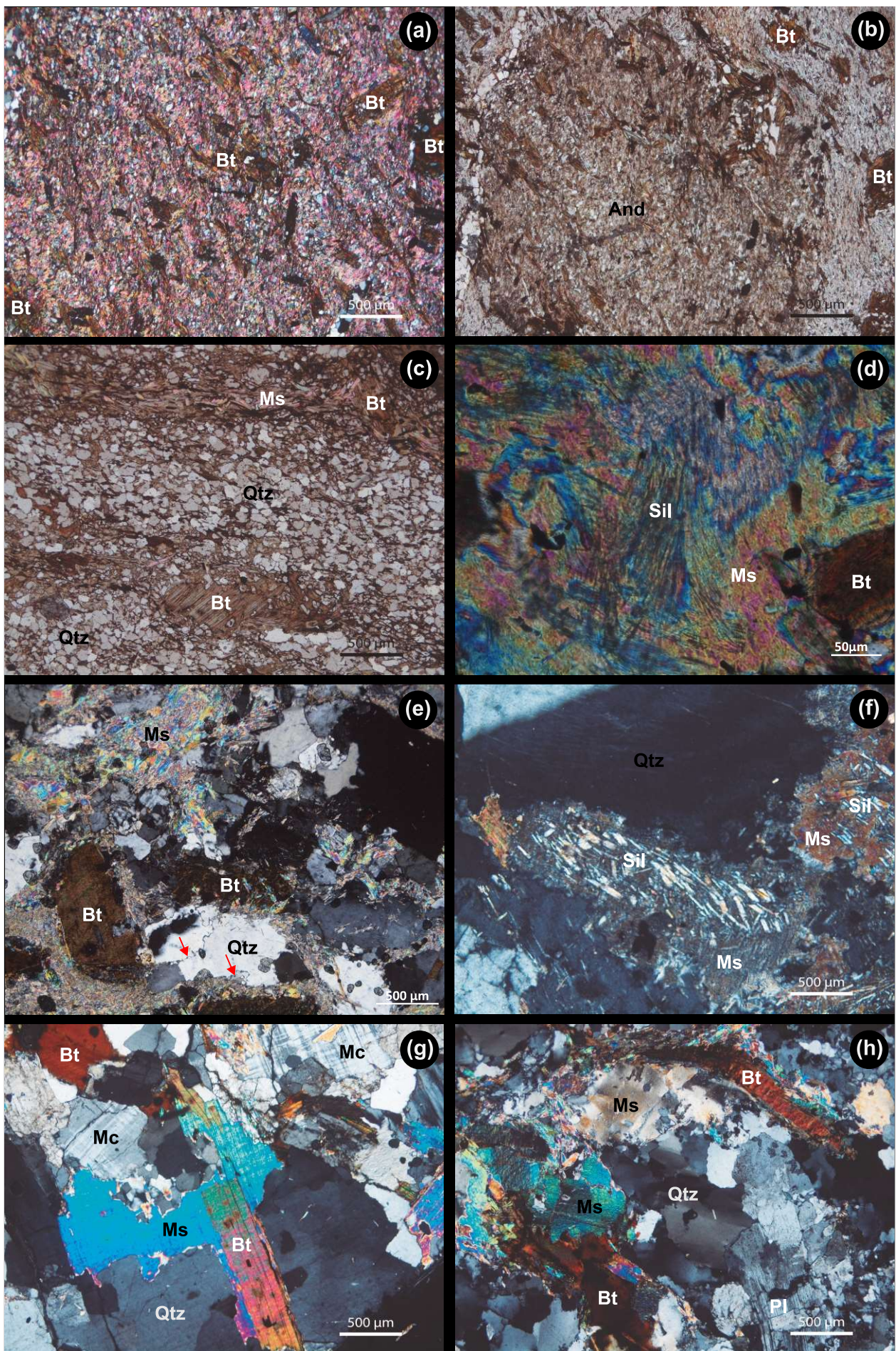


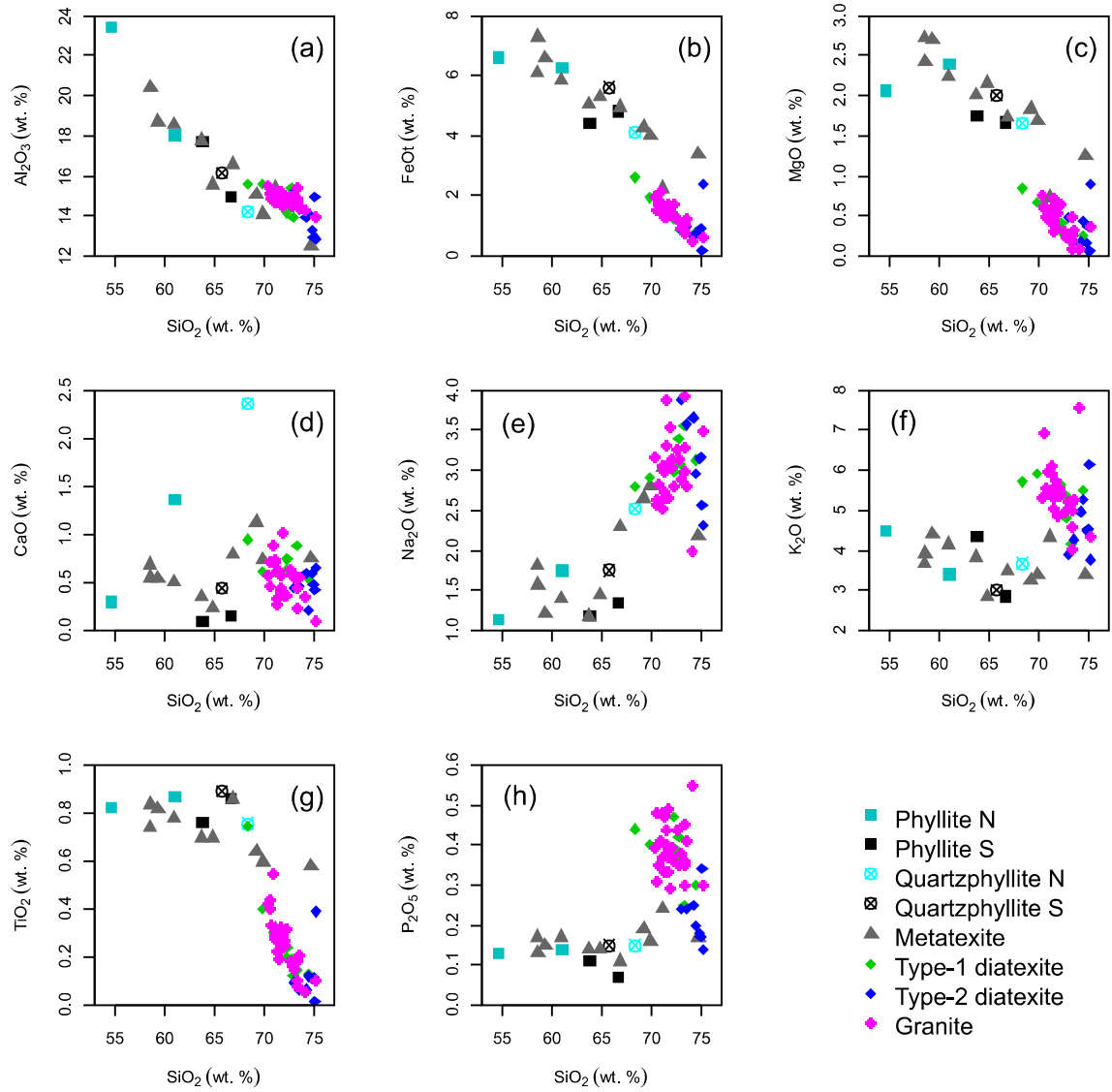
Fig.4

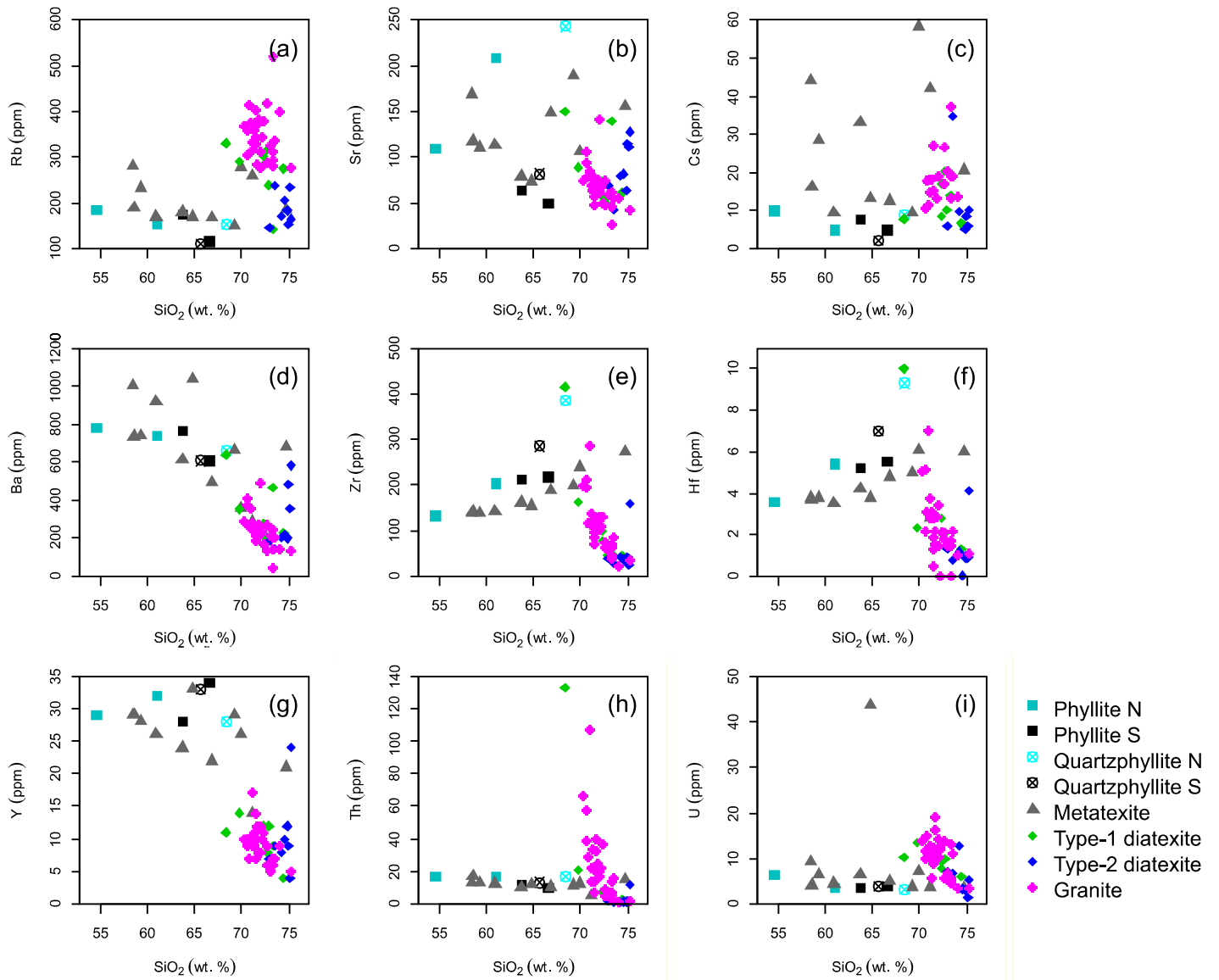
Fig.5

Fig.6

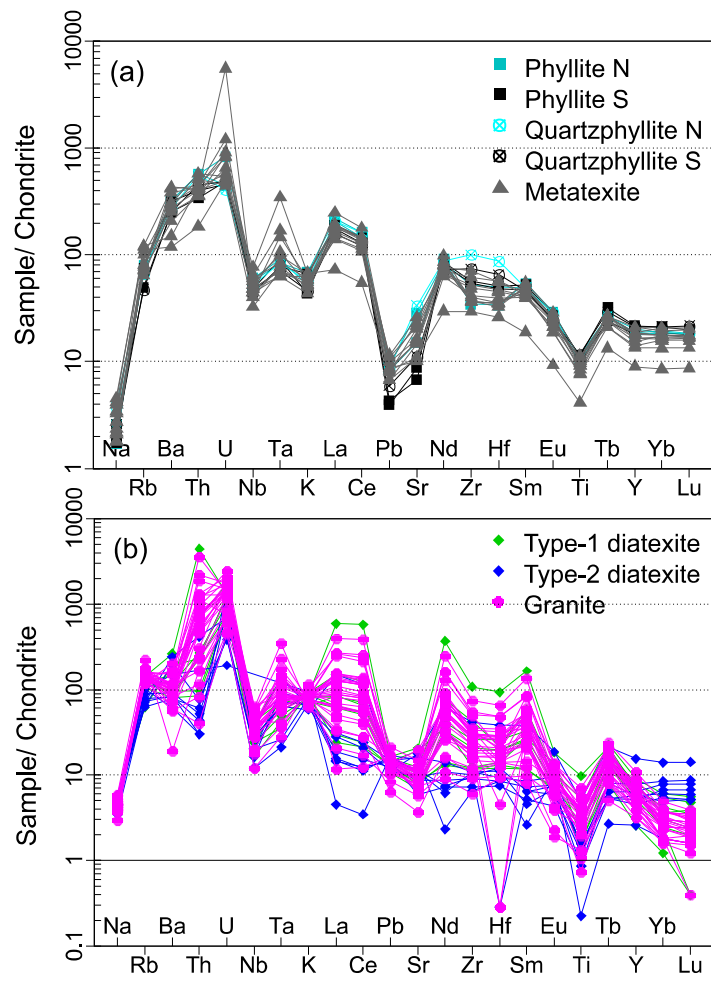


Fig.7

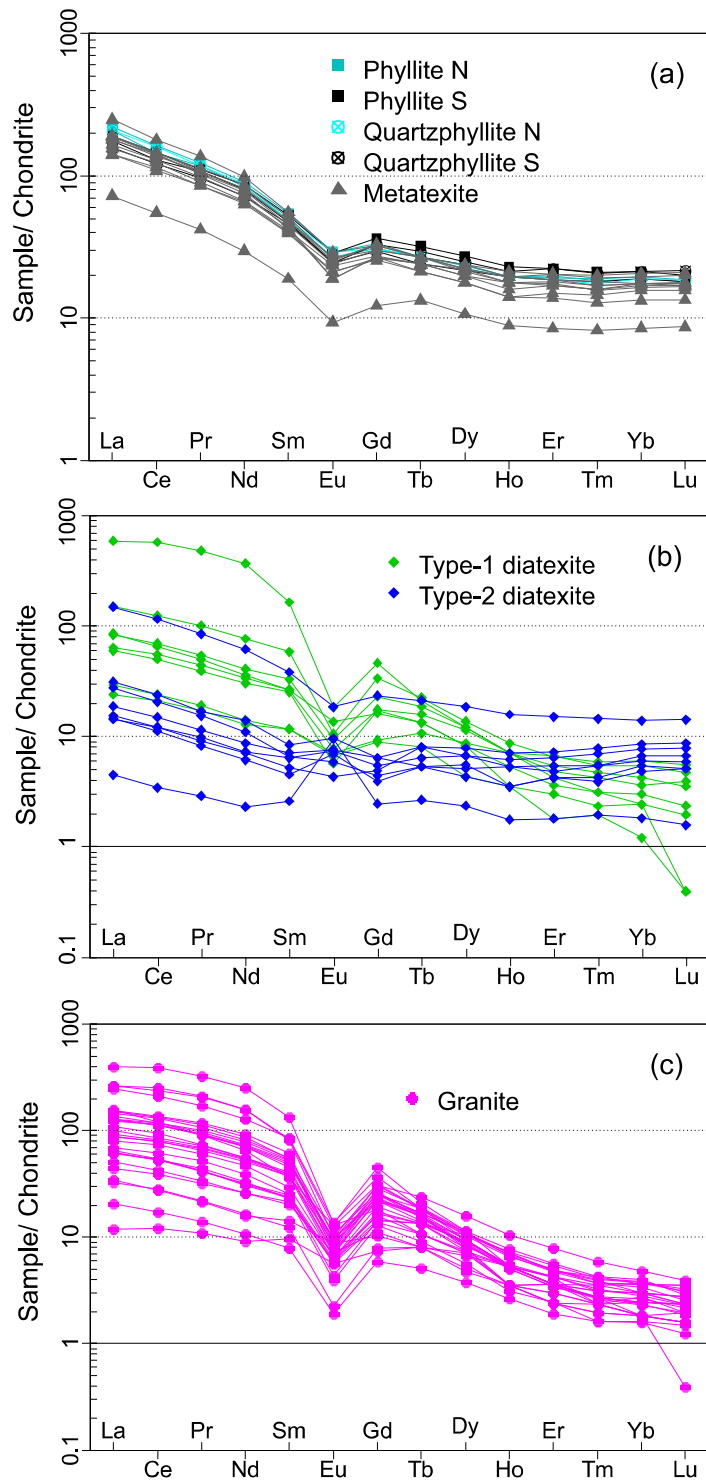


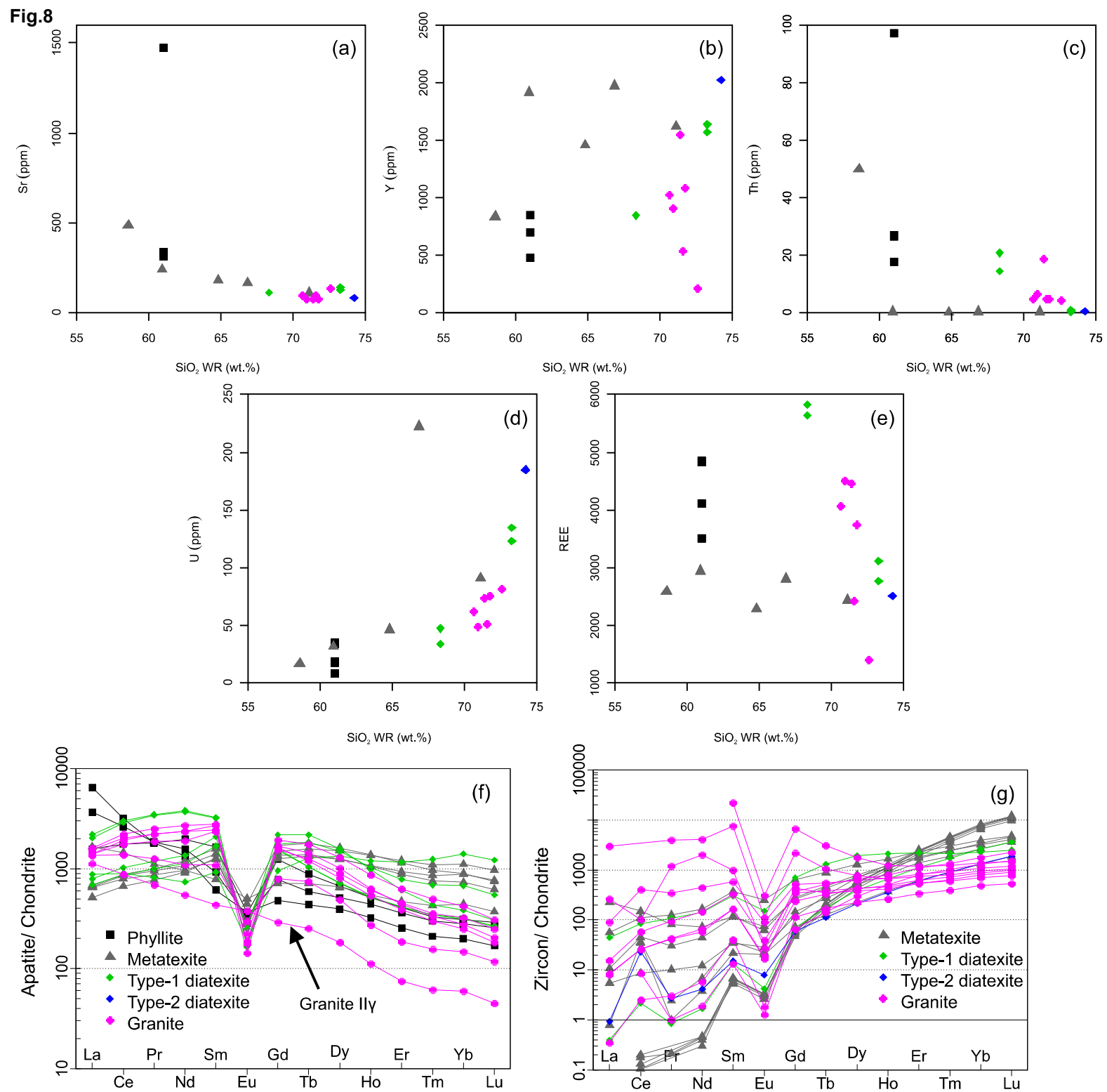
Fig.8

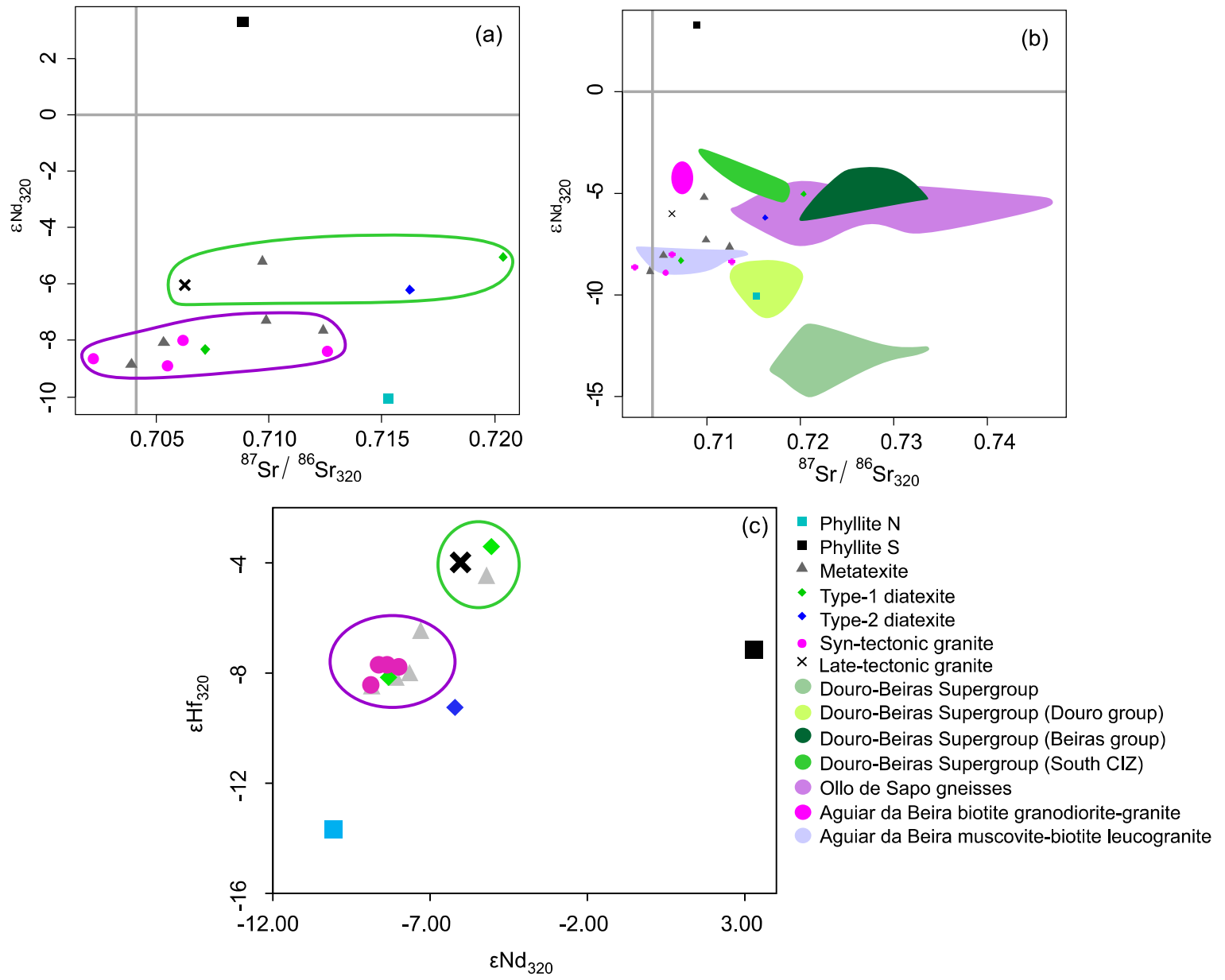
Fig.9

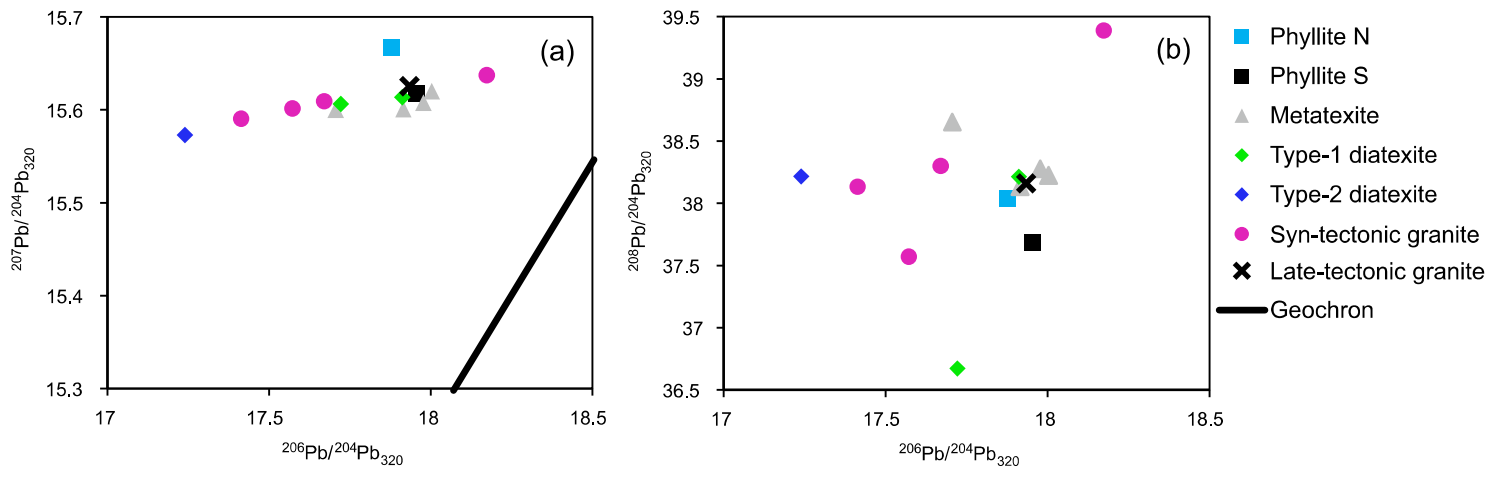
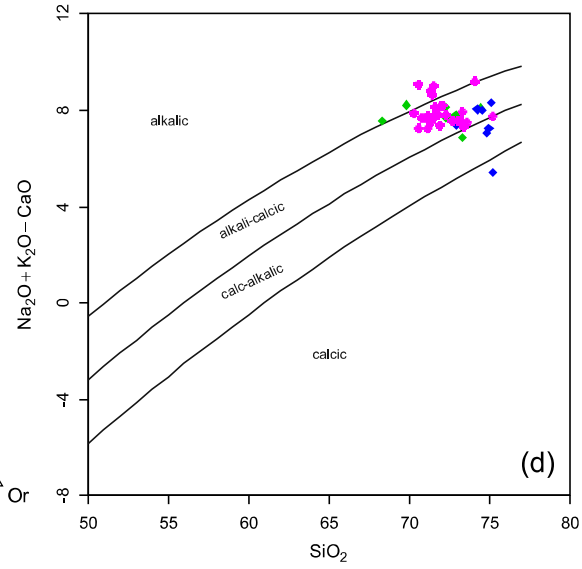
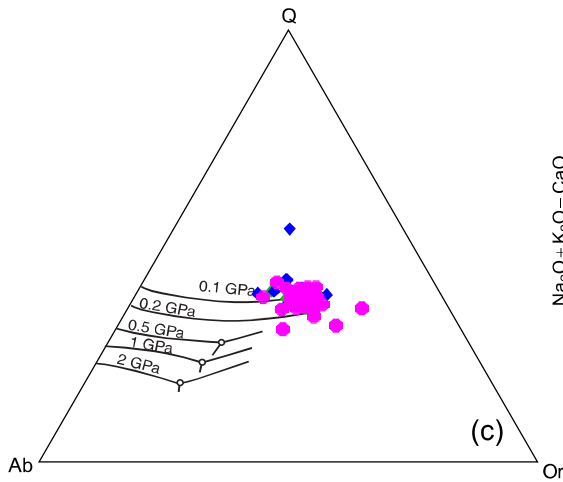
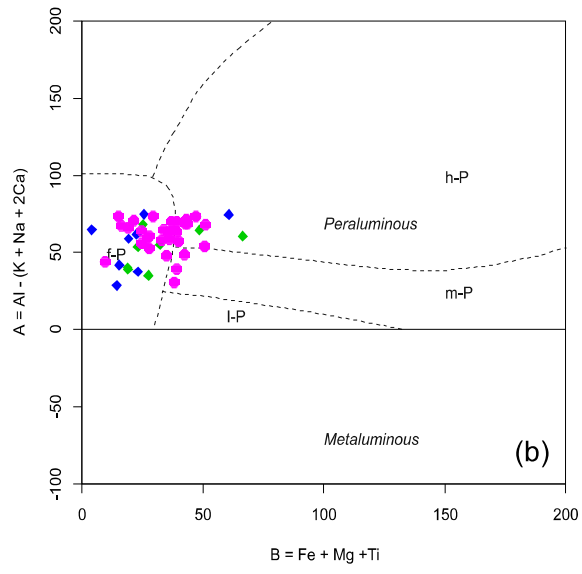
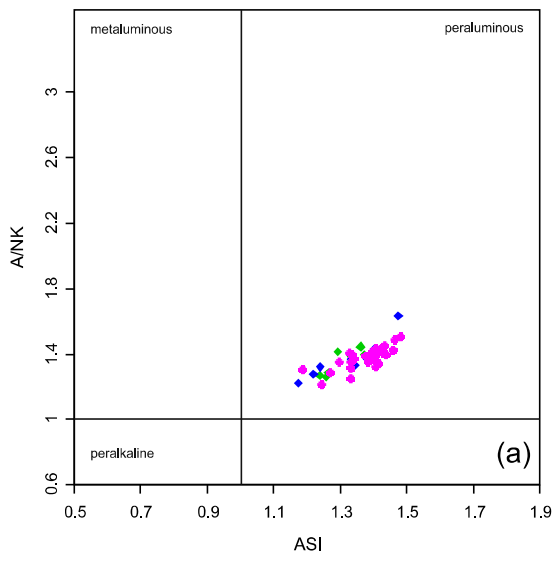
Fig.10

Fig.11

- Phyllite N
- Phyllite S
- ⊠ Quartzphyllite N
- ⊠ Quartzphyllite S
- ▲ Metatexite
- ◆ Type-1 diatexite
- ◆ Type-2 diatexite
- Granite

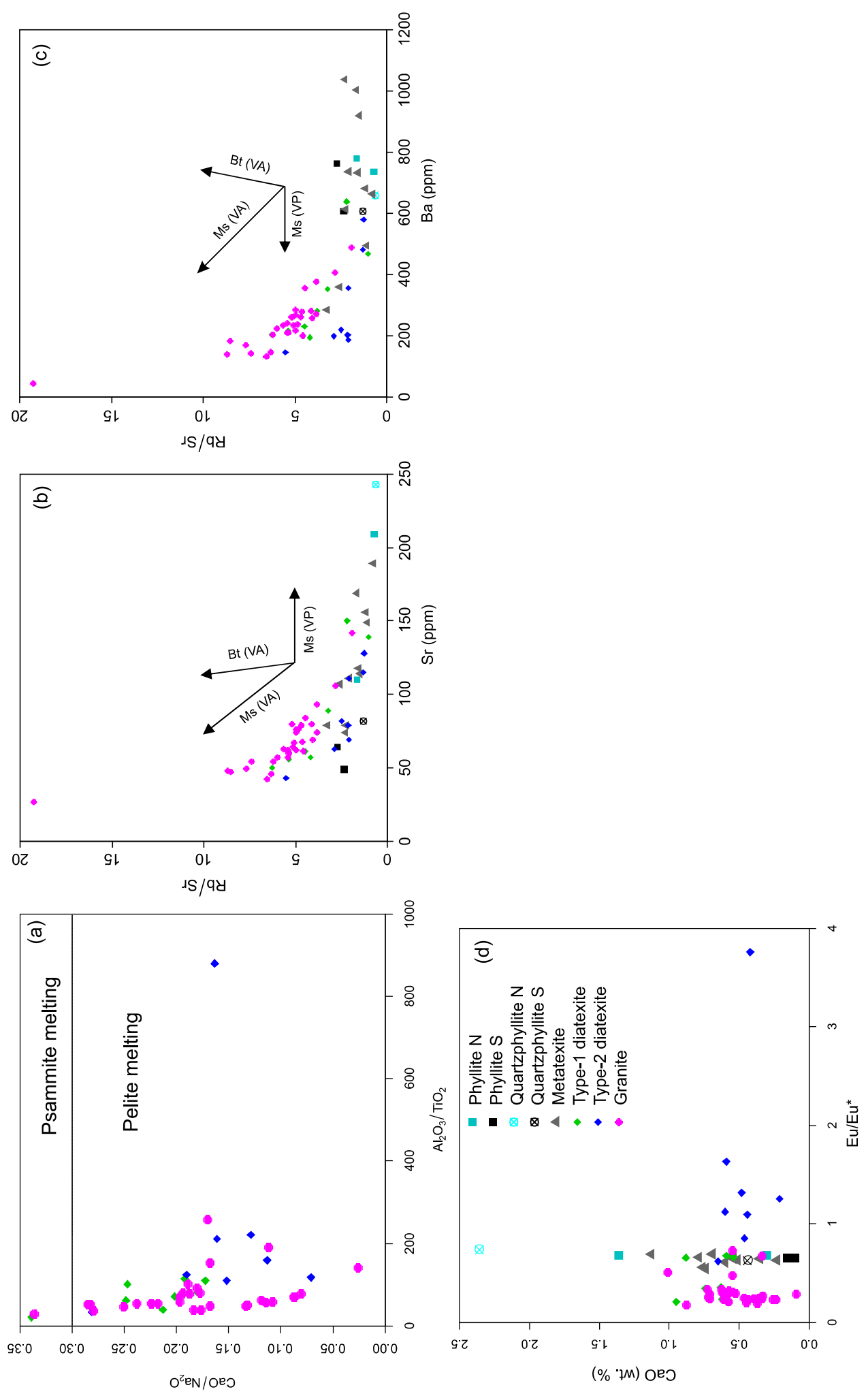


Fig.12

Fig.13

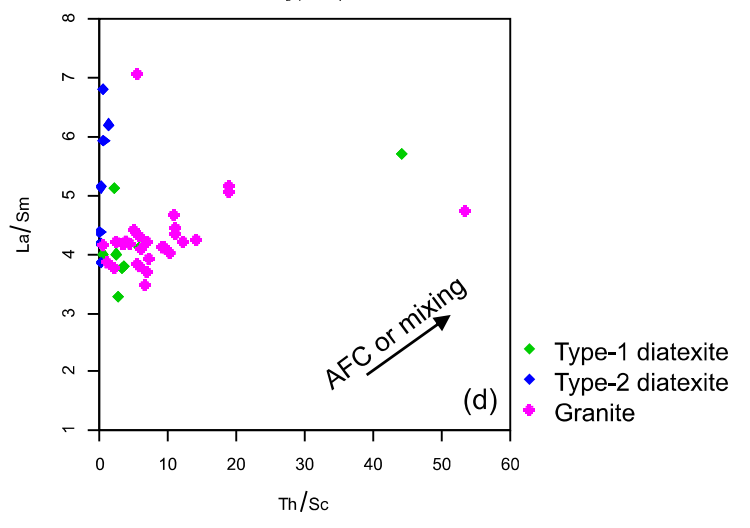
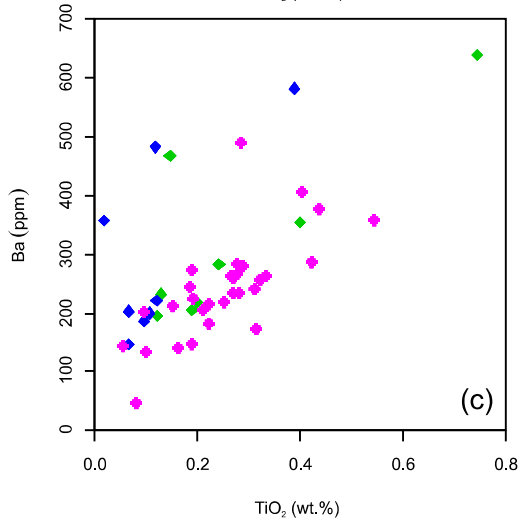
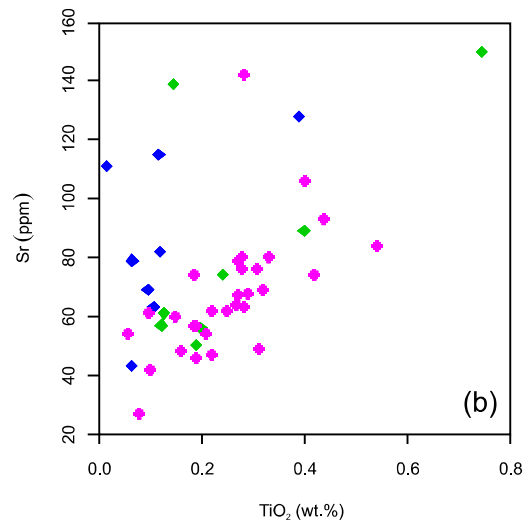
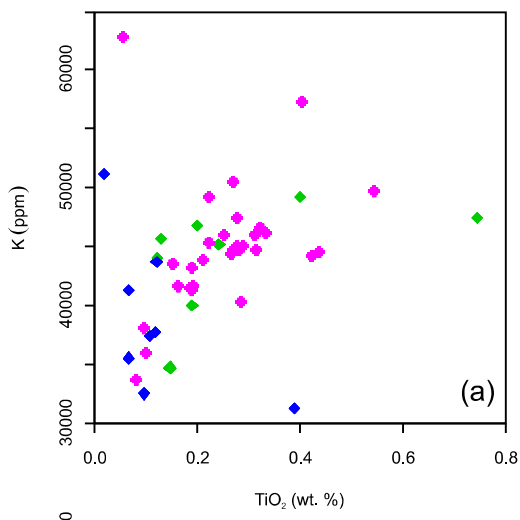


Fig.14

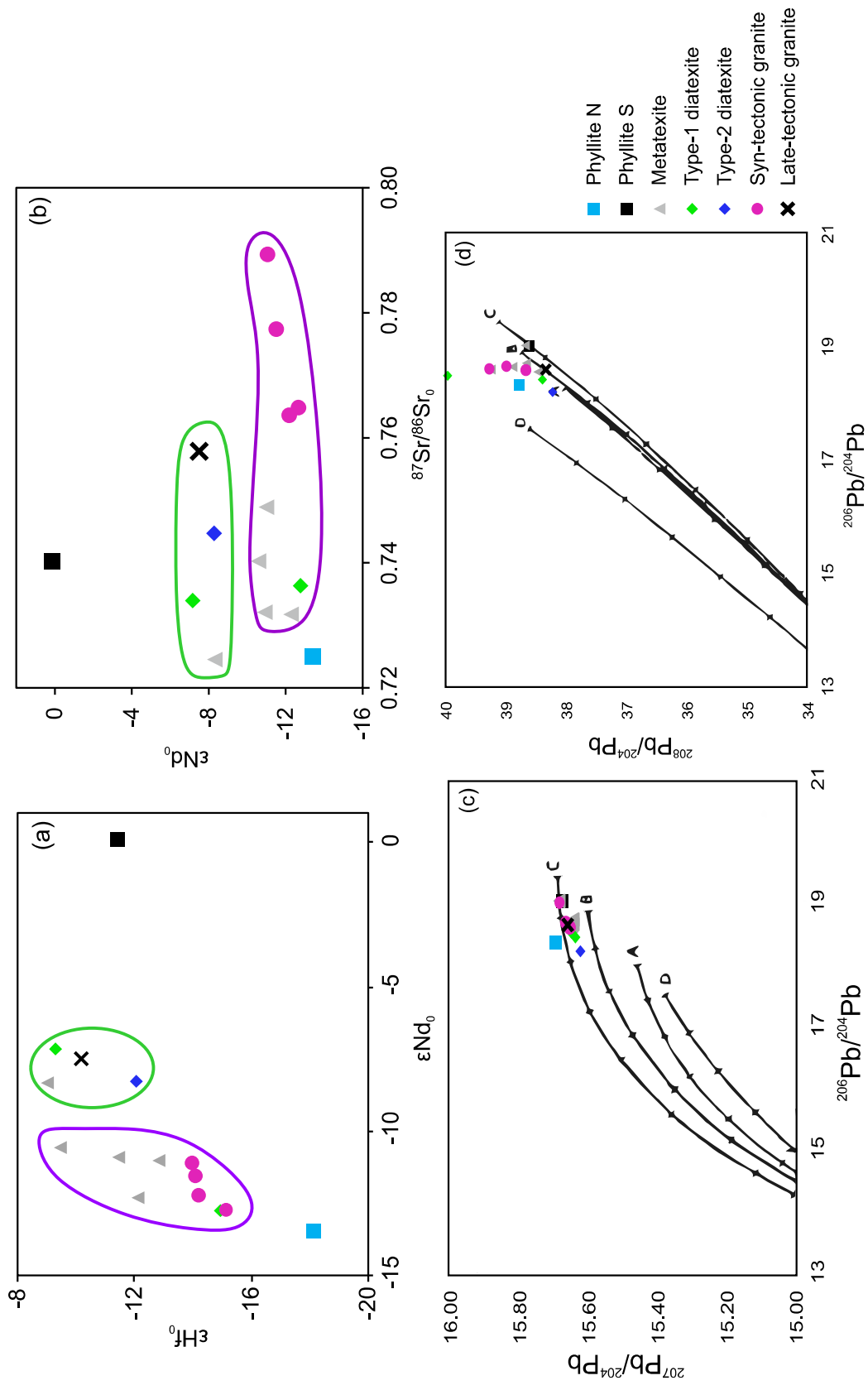
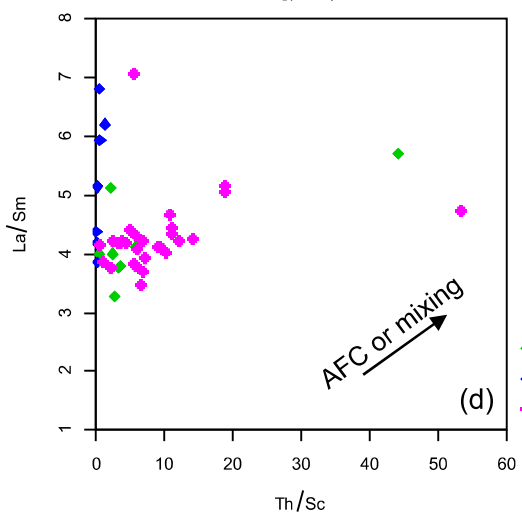
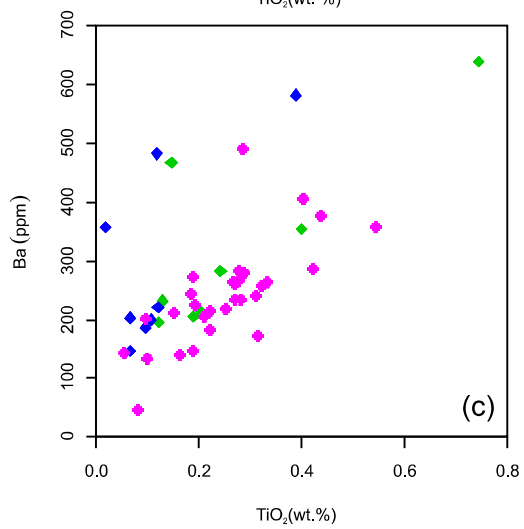
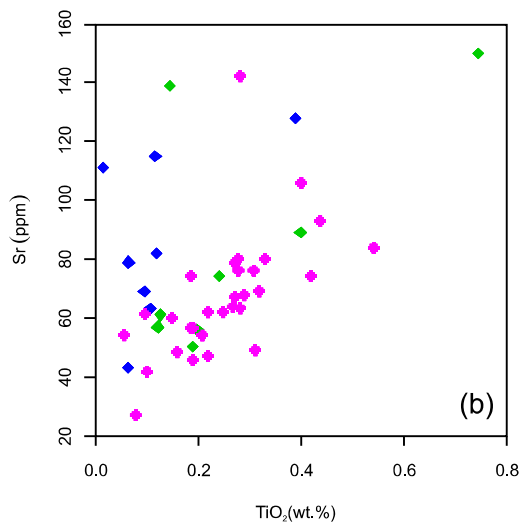
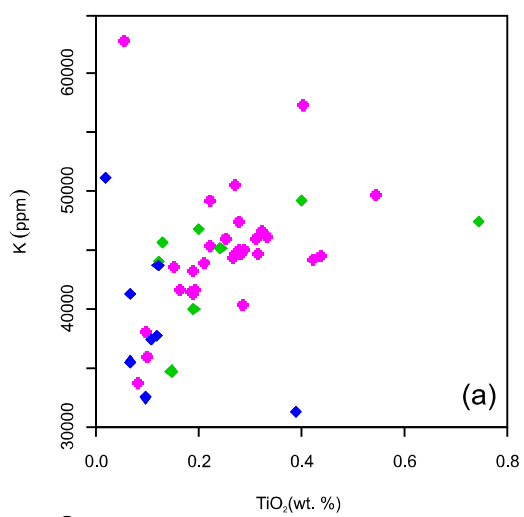


Fig.15



- ◆ Type-1 diatexite
- ◆ Type-2 diatexite
- ◆ Granite