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1 **Insights from mineral trace chemistry on the origin of NYF and Mixed LCT+NYF**
2 **pegmatites and their mineralization at Mangodara, SW Burkina Faso**

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20 Conceptualization, Writing - review & editing. Urbain Wenmenga: Investigation,
21 Conceptualization, Writing - review & editing.

22

23 **Abstract**

24 The Mangodara district (South West of Burkina Faso, West African Craton) consists of a
25 regional-scale Eburnean dome cored by granitoid-gneisses comprising rafts of migmatitic
26 paragneisses and amphibolites of the Paleoproterozoic Birimian series. The occurrence of
27 rare-metal-bearing pegmatites in diffuse contact with these migmatitic and granitoid gneisses
28 suggests that they originated from segregation of a residual melt of these partially molten
29 hosts. In this paper, we constrain the petrogenetic link between pegmatites and their hosts,
30 and the mechanisms of rare metal fractionation in LCT vs NYF petrogenetic signatures by the
31 geochemistry of micas, apatite, columbite-group minerals, garnet and zircon.

32 Titanite-allanite type pegmatites (containing titanite, allanite, epidote, zircon and apatite as
33 accessory minerals) and their evolved equivalent, apatite-zircon type pegmatites (richer in
34 apatite, lower K/Rb and Fe/Mn ratio in biotite but Li-depleted) are poorly enriched
35 metaluminous pegmatites. They display a continuous evolution trend in K/Rb and Fe/Mn in
36 biotite and similar REE pattern in apatite, which favor an origin by segregation of residual
37 melt within tonalitic-trondhjemitic gneiss in the core of the Mangodara dome.

38 Garnet-columbite type pegmatites (hosting REE-bearing phosphates and Zr-U-Th-bearing
39 metamict minerals) are mixed LCT+NYF pegmatites. Their micas slightly enriched in Li,
40 LREE-rich apatite and Nb-Ta-U-rich garnet are consistent with an origin by partial melting of
41 a metasedimentary source, with dehydration of biotite (reservoir of Li, Rb, Nb) and
42 dissolution of apatite-monazite (reservoir of REE). Apatite crystals in one garnet-columbite
43 pegmatite reveal an inherited REE signature typical of apatite-zircon pegmatites, which
44 suggests mingling of the LCT-pegmatite-forming melt with the residual melt derived from
45 crystallization of the metaluminous pegmatites.

46 Garnet-REE type pegmatites are NYF pegmatites (ilmenite-pyrophanite, euxenite-
47 aeschynite), which points to an origin by melt segregation within granodioritic gneiss
48 associated with breakdown/entrainment of amphibole (reservoir of REE, Y) and LREE
49 segregation by allanite and phosphates in the source.

50 These data show that the LCT vs NYF signature of pegmatites of the Mangodara district
51 results primarily from the chemical composition of the partially-molten source and the
52 minerals involved in the partial melting reactions, which vary as a function of depth (micas,
53 phosphates, amphiboles, garnet). The trace-element signature of anatectic peraluminous
54 pegmatite-forming melt might then be affected by mingling with residual Nb-enriched
55 metaluminous melt.

56 **Keywords:** Rare-metal-bearing pegmatite, West Africa Craton, geochemistry, micas,
57 accessory minerals, mixed LCT+NYF pegmatites

58

59 **1. Introduction**

60 Granitic rare-metal-bearing pegmatites are classified in two petrogenetic families based on
61 accessory phase mineralogy and their rare-metal enrichment trends (Černý 1991a, b; Fersman
62 1940). The Lithium-Cesium-Tantalum (LCT) family is characterized by enrichment in alkali
63 elements, Li, Cs, Ta over Nb, Be, P and F, typically hosted by spodumene/petalite, lepidolite,
64 pollucite, and columbite-group minerals (CGM) (Černý 1991a). The pegmatites belonging to
65 the LCT family typically show internal zoning of fabric and mineralogy, and regional zoning
66 of mineralogy (Černý 1991a, b; Ercit et al. 2005). The Niobium-Yttrium-Fluorine (NYF)
67 family is marked by enrichment in Nb > Ta, F, Y, rare earth elements (REE), Ti, Zr, Be and
68 U. This group possess a more exotic mineralogy with rare-metal-bearing minerals like
69 gadolinite, euxenite, aeschynite-Y, samarskite, or fergusonite (Černý 1991b; Černý and Ercit
70 2005; Martin and De Vito 2005), and rarely exhibit a regional zoning pattern (Simmons et al.
71 1987; Smeds 1994). A third family, the mixed LCT+NYF family, designates pegmatites that
72 show composite properties of the two other families, but they are less frequently described in
73 literature (Martin and De Vito 2005; Müller et al. 2017; Turlin et al. 2017).

74 Despite a growing number of studies in recent years, the genesis of granitic rare-metal
75 enriched pegmatites is still a subject of debate. Several authors have proposed an origin from
76 extreme differentiation of a fertile (rare-metal-rich) magma (Černý 1991a; Černý et al. 2012;
77 London 1996; Martin and De Vito 2005). For LCT pegmatites, differentiation of a granitic
78 magma with a peraluminous signature derived from metasediments is invoked (e.g., Breiter et
79 al. 2018; Černý 1991a; Icenhower and London 1996; Roda-Robles et al. 2012; Shearer et al.
80 1987), whereas NYF pegmatites are interpreted as deriving from mantle-sourced anorogenic
81 magmas with a peralkaline signature (e.g., Černý 1991a; Černý and Ercit 2005; Estrade et al.
82 2014; Martin and De Vito 2005; Miller 1996; Schmitt et al. 2002). Mixing of magmas from
83 different sources, or metasomatic overprinting might explain the geochemical features of

84 mixed LCT + NYF pegmatites (Martin and De Vito 2005; Müller et al. 2017). However, the
85 origin of rare-metal-bearing pegmatites from granitic melt fractionation is challenged in case
86 of (i) the absence of a fertile granitic pluton or the rare-metal deficiency of the magmatic
87 source relative to the exposed pegmatites (Müller et al. 2017; Norton and Redden 1990; Shaw
88 et al. 2016; Stewart 1978), (ii) a discontinuity or disparity of geochemical trend between
89 identified plutonic source and pegmatites (Martins et al. 2012; Roda-Robles et al. 1999), and
90 (iii) a temporal gap between the ages of granite and pegmatite crystallization (Goodenough et
91 al. 2014; Melleton et al. 2012; Müller et al. 2017).

92 Alternative propositions to the granite-related hypothesis are an origin of LCT pegmatites
93 either by direct partial melting of metasediments, as supported by Roda-Robles et al. (1999),
94 Shaw et al. (2016), and Müller et al. (2017), or by different degrees of fractionation of a
95 magma that migrated from its partially molten source (Martin and De Vito 2005; Shearer et
96 al. 1985; Stewart 1978). The major drawback of the anatectic model is that trace elements are
97 expected to be diluted in an increasing volume of melt during partial melting (London et al.
98 2012). Also, the strong partitioning dependence of rare-metals on residual and peritectic
99 minerals might further contribute to rare-metals depletion in the anatectic melt (Bea 1996;
100 Bea et al. 1994; Bea and Montero 1999; Brown et al. 2016). Despite these limitations, the
101 anatectic model of pegmatite genesis has recently been invoked in several studies (Bonzi et al.
102 2021; Müller et al. 2015, 2017; Shaw et al. 2016; Turlin et al. 2019).

103 Petrogenetic studies of LCT and NYF pegmatites are mostly based on the chemical
104 composition of their minerals, because these reflect the physical conditions of crystallization
105 (pressure, temperature) and can record the chemical evolution of the pegmatite-forming melt
106 (Barnes et al. 2012; Breiter et al. 2018; Deveaud et al. 2015; Müller et al. 2018; Novák et al.
107 2011; Pichavant et al. 2016; Roda-Robles et al. 1999; Simmons et al. 1987; Thomas et al.

108 2000; Van Lichtenvelde et al. 2008). The composition of micas is used to define enrichment
109 trends in trace elements for evolved granites and pegmatites (Hulsbosch et al. 2016; Martins
110 et al. 2012; Stepanov et al. 2014; Stepanov and Hermann 2013; Van Lichtenvelde et al. 2008).
111 The Nb/Ta and K/Rb ratios in micas are known to decrease with melt fractionation following
112 a Rayleigh type fractionation model (Černý et al. 1985; Černý and Burt 1984; Foord et al.
113 1995; Martins et al. 2012; Roda-Robles et al. 2006; Stepanov et al. 2014; Stepanov and
114 Hermann 2013). Finally, trace elements in minerals such as phosphates, garnet or CGM are
115 proven tracers of the petrogenetic environment of their parental rocks (Belousova et al. 2002;
116 Graupner et al. 2010; Groulier et al. 2020; Linnen and Cuney 2005; Melcher et al. 2015,
117 2017; Samadi et al. 2014; Turlin et al. 2017; Van Lichtenvelde et al. 2010; Villaros et al.
118 2009).

119 In the Paleoproterozoic West African Craton, previously described rare-metal-bearing
120 pegmatites of Saraya in Senegal, Winneba in Ghana, Issia in Ivory Coast and Goulamina in
121 Mali consist of LCT pegmatites mineralized in spodumene and CGM (Allou 2005; Allou et
122 al. 2005; Dampare et al. 2005; Ndiaye et al. 1997; Nude et al. 2011; Wilde et al. 2021). In a
123 previous paper (Bonzi et al. 2021), we have presented the geological context of the rare-
124 metal-bearing pegmatites of the Mangodara district (southwestern Burkina Faso). Based on
125 structural, petrological, geochemical and geochronological data on the pegmatites and their
126 hosts, we have proposed an origin by partial melting of paragneiss/amphibolite and melt
127 segregation in tonalitic-trondhjemitic-granodioritic gneisses. In order to test this hypothesis,
128 in this new contribution, we further describe both mixed (LCT+NYF) features for the
129 pegmatite field of the Mangodara district and document mineralogy and geochemical
130 composition of micas and accessory minerals from pegmatites and host rocks. These data are
131 used to identify the petrogenetic family of the pegmatites, as well as to discuss the sources of
132 rare-metals and their behavior during crystallization.

134 **2. Geological background of the Mangodara district**

135 **2.1. Regional setting of the Paleoproterozoic West African Craton**

136 The Paleoproterozoic domain of the West African Craton (WAC), named Baoule-Mossi
137 domain, was accreted to the Archean nucleus of the WAC during the Eburnean orogeny
138 (Milési 1989; Vidal and Alric 1994; Kouamelan 1996; Tshibubudze et al. 2009; Wane et al.
139 2018; Grenholm et al. 2019a, b; McFarlane et al. 2019) (Fig. 1). The Baoule-Mossi domain is
140 composed of Birimian greenstone belts made of metavolcanic, metavolcano-sedimentary and
141 metasedimentary rocks, emplaced between 2220 Ma and 2160 Ma (Baratoux et al. 2011;
142 Giovenazzo et al. 2018; Hirdes et al. 1996; Hirdes and Davis 2002; Lüdtkke et al. 1998), and
143 Bandamian volcanic rocks aged between 2120 and 2080 Ma (Gasquet et al. 2003; Grenholm
144 et al. 2019a; Hirdes et al. 1996; Hirdes and Davis 2002; Mériaud et al. 2020). The greenstone
145 belts alternate with gneiss, migmatite and granitoids that are referred to as granitoid gneiss
146 complexes. Greenstone belts are affected by greenschist grade metamorphism during the
147 Eburnean orogeny, rising to amphibolite grade in granitoid gneiss complexes. The
148 greenschist-amphibolite facies transition has been interpreted to represent metamorphic
149 aureoles around intrusions (Debat et al. 2003; Doumbia et al. 1998; Feybesse et al. 2006;
150 Ganne et al. 2014; Gueye et al. 2008; Pons et al. 1992), or regional metamorphic gradient
151 (Bonzi et al. 2021; Vidal et al. 2009; Vidal and Alric 1994). Crystallization ages of magmatic
152 zircon from volcanic and plutonic rocks of greenstone belts and granitoid gneiss complexes,
153 in South-Western Burkina Faso and Northern Ivory Coast, range between 2200 Ma and 2000
154 Ma (Baratoux et al. 2011; Gasquet et al. 2003; Hirdes et al. 1996; Metelka 2011; Parra-Avila
155 et al. 2017, 2019). Two peaks of magmatic activity around 2150–2130 Ma and around 2100
156 Ma are referred to respectively as ME1 and ME2 (Baratoux et al. 2011; Hirdes et al. 1996).
157 These peaks are coeval with major deformation phases D1 and D2 that are expressed by folds
7

158 and penetrative foliation in greenstones, synmigmatitic foliation in migmatites and magmatic
159 layering in granitoids (Baratoux et al. 2011; Bonzi et al. 2021; Metelka 2011). These features
160 have been interpreted as recording a combination of gravitational instabilities with diapirs
161 cored by granitoid gneiss complexes deformed during lateral flow of the orogenic crust in a
162 context of transpression (Allibone et al. 2002; Baratoux et al. 2011; Bonzi et al. 2021;
163 Chardon et al. 2020; Feybesse and Milési 1994; Ganne et al. 2012; Lompo 2009, 2010; Milési
164 1989; Milési et al. 1992; Vidal et al. 2009).

165

166 **2.2. Geology of the Mangodara district**

167 The Mangodara district is situated in the 100 km wide and 500 km long Sideradougou Gneiss
168 and Granitoid Complex (SGGC), located between the Banfora greenstone belt and the
169 Houndé-Ouango Fitini greenstone belt (Fig. 1). The SGGC is dominantly composed of
170 hornblende and biotite-bearing gneiss and granitoids, which contain rafts of amphibolites and
171 schists (Castaing et al. 2003; Gasquet et al. 2003; Giovenazzo et al. 2018; Hirdes et al. 1996).

172 The Mangodara district displays a dome structure mantled by a granodioritic gneiss and cored
173 by a tonalitic-trondhjemitic gneiss (Fig. 2, Bonzi et al. 2021). These gneisses contain rafts of
174 amphibolite, schist, and paragneiss that are variably migmatitic. In the eastern part of the
175 district, some potassic hornblende and biotite granitoids (named GG-granitoids) form plutons
176 in gradual contact with the granodioritic gneiss (Fig. 2). The gneisses and granitoids,
177 subdivided into a Na-rich series (tonalitic-trondhjemitic gneiss and two-mica trondhjemite)
178 and a K-rich series (granodioritic gneiss, hornblende-biotite granodiorite with mica-rich
179 enclaves, monzogabbro, GG-granitoids), are interpreted to result from partial melting of the
180 Birimian series (supplementary Fig. S1). Granodioritic gneiss is interpreted as a diatexite
181 resulting from partial melting dominantly of paragneisses whereas the tonalitic-trondhjemitic

182 gneiss is interpreted as a plagioclase-rich mush originating from partial melting of
183 amphibolite. The mineral-geochemical composition and texture of the tonalitic-trondhjemitic
184 gneiss and its intrusive contact relative to the mantling granodioritic gneiss, suggests that the
185 magma has migrated out of its source and has left a residue of garnet and pyroxene at depth
186 (Bonzi et al. 2021). Rafts of amphibolite and paragneiss, preserved within the migmatitic
187 gneiss, attest for a MP-HT metamorphism (sillimanite-garnet/staurolite +/- melt assemblage
188 in paragneiss and plagioclase-rich leucosome in amphibolites) delineating a solid-state to
189 synmigmatitic foliation resulting from progressive D1-D2 deformation. The presence of
190 magmatic epidote, and the estimated pressure of crystallization of hornblende (4.0–6.6 kbar)
191 point to a crystallization of gneiss-granitoids at a depth of about 12 to 20 km (Bonzi et al.
192 2021).

193 Four distinct types of pegmatites are identified in the Mangodara district, based on their
194 regional distribution (Fig. 3a), mineralogy (Fig. 3b, Fig. 4) and petrological-structural
195 relationships with their host rocks (see Bonzi et al. 2021 for a more detailed description).

196 (1) Titanite-allanite pegmatites occur exclusively in tonalitic-trondhjemitic gneiss (Fig. 3).
197 They form veins that are mostly inferior to 2 meters in thickness and are locally folded, with
198 diffuse contacts concordant to sharply discordant to the magmatic foliation of the gneiss (Fig.
199 S2. The veins are interpreted to originate from syntectonic segregation of residual melt within
200 the tonalitic-trondhjemitic gneiss.

201 (2) Apatite-zircon pegmatites are less common (4 dykes formally identified); they form veins
202 of less than one meter thickness that display diffuse contacts, concordant to discordant to the
203 foliation of the granodioritic gneiss (Fig. S3). This pegmatite type is considered as a cogenetic
204 facies of the titanite-allanite type based on their similar mineralogy and higher LILE and REE
205 content.

206 (3) Garnet-columbite pegmatites are hosted in rafts of amphibolite and paragneiss, in
207 granodioritic gneiss, hornblende-biotite granodiorite with mica-rich enclaves, and in two-mica
208 trondhjemite (Fig. S4). They are thick dykes of several meters in thickness (can reach 300 m
209 in apparent width). They are considered as the segregation product from an anatectic melt that
210 migrated out of the partially molten paragneiss.

211 (4) The garnet-REE pegmatites dykes have thicknesses of the order of one to several meters
212 and are distributed around the interface between granodioritic gneiss and GG-granitoids (Fig.
213 S5). They are interpreted as the product of melt segregation within the granodioritic gneiss
214 during its crystallization.

215 The regional distribution of the Mangodara pegmatites depicts a zonation evolving from the
216 core of the Mangodara dome to the eastern part of the district (Fig. 3). The crystallization of
217 titanite-allanite pegmatite is dated at 2094.3 ± 8 Ma using *in situ* U-Pb dating in zircon by
218 laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS), which
219 corresponds to the end of the ME2 (Bonzi et al. 2021). Apatite ages (*in situ* U-Pb LA-ICP-
220 MS) of 2094 ± 21 Ma on tonalitic-trondhjemitic gneiss, 2055 ± 20 Ma on apatite-zircon type
221 pegmatite, and 2041 ± 33 Ma on granodioritic gneiss record the regional cooling under 500
222 °C, which occurred after 2100 Ma (Bonzi et al. 2021).

223

224 **3. Sampling and analytical methods**

225 **Pegmatites and surrounding rocks of the Mangodara district were sampled to obtain**
226 **representative thin sections of the different lithologies. Pegmatites were specifically**
227 **sampled in internal parts of the dykes which exhibit fine (~0.5–2 mm) to medium-sized**
228 **(2–5 mm) crystals and preferentially contain mica (see Tables**

229 **Table 1). Aplitic samples (grain size < 0.5 mm) were avoided to limit modal and**
230 **geochemical biases linked with late-albitization (e.g., cleavelandite (Dewaele et al. 2011;**
231 **Hulsbosch and Muchez 2020; Müller et al. 2018). Also, very coarse-grained samples**
232 **(grain size > 0.5 mm) were avoided to guarantee a homogeneous representativity of the**
233 **mineral phases existing in each sample. Micas, feldspars, garnet and accessory phases (<**
234 **5% in modal composition) were identified *in situ* on thin section using optical**
235 **microscopy, and backscattered electrons imaging (BSE) with a Scanning Electronic**
236 **Microscope (SEM) Jeol 6360LV coupled with an energy dispersive X-ray spectroscope**
237 **Silicon Drift Detector from Bruker (GET, Toulouse, France). The major element**
238 **compositions of micas, apatite, CGM, garnet and zircon were obtained *in situ* on thin**
239 **sections using an electron probe microanalyser (EPMA) Cameca SXFive at the Center**
240 **Raimond Castaing (Toulouse, France), and a Cameca SX100 at GeoRessources (Nancy,**
241 **France). Standards used are albite (Na), wollastonite (Si, Ca), Al₂O₃ (Al), sanidine (K),**
242 **MnTiO₃ (Mn,Ti), Fe₂O₃ (Fe), topaze (F), MgO (Mg), tugtupite (Cl), Cr₂O₃ (Cr), LaPO₄**
243 **(La), CePO₄ (Ce), UO₂ (U), zircon (Zr), HfZr (Hf), graffonite (P), Ta and Nb.**
244 **Specifically for apatite analysis, a Durango apatite was used as standard for F and P.**
245 **Detection limits for major element oxides are under 0.2 wt%. Detection limits for F, La**
246 **and Ce oxides are under 0.5 wt%. Li content in micas was estimated by correlation with**
247 **Mg content for biotite and Al content for muscovite following the empirical formulae**
248 **compiled by Tischendorf et al. (2004), only for the purpose of mica identification and**
249 **representation in Tischendorf diagram (Tischendorf et al. 1997, 1999, 2001). Structural**
250 **formulae of micas, garnet, apatite, zircon and monazite were calculated based on 11, 12,**
251 **13, 2 and 4 oxygen atoms, respectively. Mineral compositions of sampled rocks and**
252 **analyzed mineral species are given in Tables**
253 **Table 1).**

254 Minor and trace elements (including Li) in micas, apatite, CGM, garnet and zircon were
255 analyzed *in situ* on thin sections (30 μm thick) by laser ablation – inductively coupled plasma
256 – mass spectrometry (LA-ICP-MS) using Geolas Pro coupled to an Agilent 7500 at
257 GeoRessources (Nancy, France). Data were reduced using NIST-610 and NIST-612 as
258 internal standard, and EPMA major element composition as external standard.

259

260 **4. Results**

261 **4.1. Petrography of pegmatites**

262 **4.1.1. Titanite-allanite pegmatites**

263 Titanite-allanite pegmatites are dominated by medium to coarse-grained plagioclase and
264 quartz (crystal size from millimeter to decimeter), which commonly display graphic texture,
265 particularly the coarser grains. These pegmatites are generally unzoned, but small dykes,
266 discordant to the foliation of their host, show quartz-rich cores and margins with plagioclase
267 crystals oriented perpendicular to the contacts.

268 Major mineral phases in titanite-allanite pegmatites are oligoclase, quartz, rare to abundant
269 biotite, whereas muscovite and microcline are rare (Fig. 4, supplementary Fig. S6). Muscovite
270 occurs as secondary crystals, intercalated between biotite sheets, or as radiating acicular laths
271 in inter-grain fractures, resulting in a miarolitic – like texture (supplementary Fig. S6a).
272 Accessory minerals are magnetite, apatite, epidote, commonly metamict allanite, titanite,
273 ilmenite and zircon. Apatite is found as inclusions in micas, plagioclase, quartz, and altered
274 allanite (supplementary Fig. S6b). Zircon is frequently present as inclusions with size $< 5 \mu\text{m}$
275 in biotite, plagioclase, apatite or titanite (Fig. 5a, supplementary Fig. S6b). SEM images of
276 zircons reveal oscillatory zoning partially overprinted by patchy metamict zones. Titanite is
277 subhedral and can be associated with allanite and zircon (Fig. 5). Euhedral epidote can host

278 exsolutions of titanite in fissures and crystal rims (Fig. S6c). Ilmenite is usually rimmed by a
279 corona of late titanite, or late symplectitic epidote (supplementary Fig. S6d).

280

281 **4.1.2. Apatite-zircon pegmatites**

282 Apatite-zircon pegmatites are dominantly coarse-grained (centimeter scale) with
283 hypidiomorphic to cumulatic texture marked by subhedral plagioclase and intergranular
284 anhedral biotite (Fig. 5b, supplementary Fig. S7a). These pegmatites are rich in plagioclase,
285 poor in quartz and K-feldspar, with biotite as the unique primary mica and the main
286 ferromagnesian mineral, rare epidote, and euhedral magnetite (Fig. 5b, supplementary Fig.
287 S7b). Apatite is abundant as millimeter sized euhedral prisms (supplementary Fig. S7a) that
288 are locally aggregated in centimeter-wide masses. Zircon grains with size up to 200 μm occur
289 in biotite, typified under SEM by a patchy texture that overprints the primary oscillatory
290 zoning.

291

292 **4.1.3. Garnet-columbite pegmatites**

293 Garnet-columbite pegmatites are characterized by a diversity of textures such as decimeter-
294 scale coarse grained texture, graphic texture of K-feldspar-quartz, medium-grained texture
295 with muscovite, quartz and feldspars (supplementary Fig. S8a), quartz-mica intergrowths,
296 internal or margin-controlled directional growth of quartz-feldspar crystals, and rhythmic
297 layering in cleavelandite-rich aplitic zones (supplementary Fig. S4b). Muscovite is the
298 dominant mica, which ranges from millimetric interstitial flakes to centimetric tabular stacks,
299 and occasionally showing kink folds suggesting some deformation during pegmatite
300 crystallization (supplementary Fig. S8b). Biotite and muscovite can coexist in the same
301 pegmatite body, but biotite is rare to absent in garnet-rich sections (Fig. 4).

302 Garnet-columbite pegmatites have a richer variety of accessory minerals relative to the other
303 pegmatite types (Fig. 4). Apatite is observed as millimetric euhedral grains, more common in
304 pegmatites of the northern part of the studied zone (Mangodara and Farakorosso). Orange to
305 reddish garnet forms clusters or strings of millimetric to centimetric euhedral grains (Fig.
306 S8c), generally abundant in aplitic layers. This garnet frequently hosts inclusions ($< 200 \mu\text{m}$)
307 of zircon, monazite (EPMA composition in supplementary Table S1), CGM and REE-
308 metamict minerals (Fig. 5c). Tourmaline occurs as prisms of several mm to cm in size,
309 arranged in glomeromorphic or interstitial texture in some pegmatites. Monazite occurs as
310 subhedral crystals inferior to $200 \mu\text{m}$, included in feldspar or in muscovite, frequently cracked
311 or altered. Xenotime is rarely present in feldspar and garnet, as xenomorph crystals
312 intergrown with zircon (supplementary Fig. S8d), or subhedral minerals altered by
313 metamictization, with size inferior to $100 \mu\text{m}$. Columbite-group minerals occur as elongated
314 to acicular subhedral to euhedral accessory crystals, with size inferior to $200 \mu\text{m}$. It forms
315 micro-inclusions in garnet, feldspar and muscovite of pegmatites near Farakorosso and at
316 Massadeyirikoro-West (Fig. 2, Fig. 5c, supplementary Fig. S8c, e and f). Metamict thorite
317 (Fig. S8g), and relict of altered minerals bearing Zr, U, Th and REE (possibly cheralite, Fig.
318 S8e, f, g and h) are found in garnet, and rarely in plagioclase, sometimes associated with
319 zircon, xenotime or CGM (Fig. 5c, supplementary Fig. S8d and f). Scanning electron
320 microscopy reveals that they contain traces of Zr, U, Th, Hf, Ce, Ca.

321 A dyke of garnet-columbite pegmatite subparallel to a dyke of apatite-zircon pegmatite shows
322 an intrusive contact at a shallow angle. The contact between the two pegmatites is sharp at
323 macroscale, but irregular at the microscopic scale and delineated by euhedral crystal borders
324 depicting a continuous magmatic texture (illustrated in Bonzi et al. 2021).

325

326 **4.1.4. Garnet-REE pegmatites**

327 Garnet-REE pegmatites are characterized by the alternation of aplitic and medium to coarse-
328 grained layers. Medium to coarse-grained layers (crystal size from millimeter to centimeter)
329 are mostly unzoned, but locally display a graphic texture. They are mainly composed of
330 feldspar, quartz (supplementary

331 Fig. S9a and b), biotite as millimetric to centimetric layers, rare muscovite, and accessory
332 garnet, epidote and magnetite (Fig. 4). Garnet occurs as euhedral centimetric minerals.
333 Tourmaline is rare and was only identified in one dyke. Ilmenite-pyrophanite (Fe-Ti-Mn-
334 bearing oxide) is found as subhedral crystals included in quartz, feldspar, and garnet (Fig. 5d,
335 supplementary

336 Fig. S9b and c). Some pegmatite dykes contain altered metamict accessory minerals with
337 elongated shapes that display a euxenite-aeschynite-like composition (supplementary

338 Fig. S9a, d, e and f, data available in supplementary Table S2). These REE-oxides are
339 replaced by clay-hematite assemblages (supplementary

340 Fig. S9d, e and f). Garnet-REE pegmatites do not contain apatite nor monazite. Epidote grains
341 are common in the pegmatite veins linked with leucosomes of a granodioritic gneiss (Bonzi et
342 al. 2021). Garnet is rare in these leucosome-related pegmatite veins.

343

344 **4.2. Rare-metal-bearing accessory minerals in pegmatite host rocks**

345 *Paragneiss and metatexitic paragneiss*

346 Garnet from paragneiss and metatexitic paragneiss occurs as synkinematic porphyroblasts of
347 up to 2 cm size containing numerous inclusions of lobate quartz-feldspar, retrograde biotite,

348 and rare opaque minerals (supplementary Fig. S10). Contrarily to their inclusion-rich cores,
349 their rims are quite homogeneous. Garnet in metatexite is homogeneous in texture, and is
350 interpreted as a peritectic phase in equilibrium with the melt segregated into leucosome veins
351 (Bonzi et al. 2021). Zircon is found as small inclusions ($< 40 \mu\text{m}$) in biotite from paragneiss,
352 mostly identified by dark aureoles of metamict alteration. In a micaschist enclave enclosed in
353 garnet-columbite pegmatite (station BMS6, Fig. 2), larger zircon (up to $200 \mu\text{m}$) is
354 characterized by regular oscillatory zoning, and hosts multiple solid inclusions, which is in
355 favor of its metamorphic origin. Zircon is also identified in metatextitic paragneiss as small
356 inclusions ($< 40 \mu\text{m}$) hosted by biotite, garnet or ilmenite (supplementary Fig. S10a). Apatite
357 is widespread in this micaschist enclave, as anhedral and lumpy crystals in micas, whereas it
358 is not present in paragneiss. Monazite is found in the metatextitic paragneiss, as inclusions ($<$
359 $50 \mu\text{m}$) in micas, quartz and garnet (supplementary Fig. S10a).

360

361 *Gneiss and plutonic rocks of Na-rich and K-rich Mangodara series*

362 Allanite-epidote, apatite, zircon and magnetite are ubiquitous minerals in both granodioritic
363 gneiss, tonalitic-trondhjemitic gneiss, and the plutonic rocks attributed to the Na-rich and K-
364 rich series of Mangodara district. Allanite can be found as non-altered crystals rimmed by
365 epidote, or altered metamict grains replaced by amorphous epidote or subhedral apatite (e.g.,
366 supplementary Fig. S10b). Subhedral to euhedral epidote is associated with biotite,
367 plagioclase, and can exhibit concentric internal zonation in GG-granitoids. Zircon mostly
368 occurs as small haloed inclusions in biotite ($< 40 \mu\text{m}$). Apatite forms subhedral to euhedral
369 crystals with size up to $300 \mu\text{m}$, intersecting biotite and amphibole, or barely included in
370 altered allanite (supplementary Fig. S10b). Magnetite occurs as euhedral to anhedral crystals
371 that reach up to 1 cm in size in the tonalitic-trondhjemitic gneiss. In granodioritic gneiss and

372 plutonic rocks, it is mostly associated with mafic minerals (biotite, amphibole) as opaque
373 inclusions (< 200 μm). Titanite is commonly seen in units of the K-rich series (granodioritic
374 gneiss, GG-granitoids), as lozenge-shaped euhedral crystals, with size up to 1 cm in length.

375

376 **4.3. Mineral geochemistry**

377 **4.3.1. Geochemical signatures of micas**

378 Major element compositions (supplementary Table S3 and Table S4) and trace element
379 compositions (supplementary Table S5 and Table S6) were collected in muscovite and biotite
380 from paragneiss, metatextitic paragneiss, two-mica trondhjemite, pegmatites, and a micaschist
381 enclave enclosed in garnet-columbite pegmatite (supplementary Fig. S8a). In the Tischendorf
382 diagram (Tischendorf et al. 1997, 2004), muscovite composition corresponds to muscovite
383 *sensu stricto* whereas biotite plots in the annite-phlogopite domain (Fig. 6). The lowest values
384 of mgli parameter (Mg - Li) in this diagram indicate the enrichment in Li (and F), which is
385 observed in muscovite from garnet-columbite pegmatite, and biotite from garnet-REE and
386 garnet-muscovite pegmatite.

387

388 *Major element composition*

389 Muscovite composition evolves through the following substitution: $\text{Al[VI]} \leftrightarrow \text{Si} + \text{Mg} + \text{Fe}^{2+}$
390 $+ \text{Ti}$ (Fig. 7a). Muscovite in biotite-bearing pegmatites (titanite-allanite pegmatites and some
391 garnet-columbite pegmatites) are richer in Fe+Mg than muscovite-only-bearing samples
392 (higher mgli parameter in Fig. 6, $\text{Fe} + \text{Mg} > 0.3$ apfu in Fig. 7a and c). Muscovite from the
393 micaschist enclave is particularly enriched in F (Fig. 7e).

394 Correlations of major elements suggest that biotite follows the substitution $\text{Si} + \text{Al}[\text{VI}] + \text{Mg}$
395 $\leftrightarrow \text{Al}[\text{IV}] + \text{Fe} + \text{Li}$ (Fig. 7b) with negligible Li contribution. Manganese in biotite follows a
396 general Rayleigh differentiation trend relative to Fe (Fig. 7g), which is continuous between
397 titanite-allanite and apatite-zircon pegmatites, and also between garnet-columbite pegmatites
398 and the two-mica trondhjemite.

399 **Titanite-allanite and apatite-zircon pegmatites host the most Mg-rich biotite among**
400 **pegmatites ($0.43 < \text{Mg}\# < 0.54$, Fig. 7d, Tables**

401 Table 1). Biotite is richer in F in apatite-zircon pegmatites ($< 0.3 \text{ wt}\%$) than in titanite-allanite
402 pegmatites ($\text{F} < 0.2 \text{ wt}\%$, Fig. 7f). Iron-magnesium-poor biotite outliers from apatite-zircon
403 pegmatites originate from the dyke crosscut by a garnet-columbite pegmatite (Fig. 7d).

404 Biotite of garnet-columbite pegmatites overlaps that of the two-mica trondhjemite, with
405 intermediary Mg content ($0.38 < \text{Mg}\# < 0.43$), but higher Ti, Mn and Fe content than in other
406 pegmatites (Fig. 6, Fig. 7d and g). Fe-rich biotite outliers of garnet-columbite pegmatites
407 (sample BMS139B, Fig. 2) correspond to a dyke containing coexisting biotite and garnet (Fig.
408 7b, d and g). Biotite from the micaschist enclave differs from biotite of paragneiss, granitoids
409 and pegmatites by its low Fe+Mg (Fig. 7d) and its remarkably high F content (0.26–0.79
410 wt%, see Fig. 7f). Biotite of garnet-REE pegmatites has low Mg# (around 0.31), F content
411 below detection limits, and plots out of the Fe/Mn Rayleigh trend drawn by biotite from other
412 pegmatites (Fig. 7g).

413

414 *Trace element composition*

415 Muscovite of garnet-columbite pegmatites follows a Rayleigh-type K/Rb fractionation trend,
416 with high K/Rb ratios (84–41) and high Rb contents (1034–2203 ppm) relative to some LCT

417 and NYF pegmatites occurrences (see Fig. 8a). Its Li content (65–580 ppm) is uncorrelated
418 with the K/Rb ratio (Fig. 8c). Niobium and Ta are the richest in muscovite of garnet-
419 columbite pegmatites (97–438 ppm Nb, Nb/Ta > 5). Nb/Ta ratios do not show a particular
420 Rayleigh fractionation trend, and rather seem to be controlled by CGM crystallization, as
421 spots corresponding to constant Nb/Ta ratios belong to muscovite that is spatially close to
422 CGM (found in the same thin section), or contains inclusions of garnet and CGM (example in
423 supplementary Fig. S8c and e).

424 Muscovite from paragneiss is depleted in Nb and Ta, with high Nb/Ta ratio (19–26).
425 However, muscovite from the micaschist enclave is particularly rich in Li (860–996 ppm), Nb
426 (138–173 ppm) and Ta (20–44 ppm).

427 Biotite of tonalitic-trondhjemitic gneiss and related pegmatites follows a K/Rb Rayleigh-type
428 trend characterized by high K/Rb ratios (81–600) and high Rb (<858 ppm), which is topped
429 by titanite-allanite pegmatites and bottomed by apatite-zircon pegmatites (Fig. 8b). Biotite
430 from titanite-allanite pegmatites is rich in Li (485–1281 ppm) but poor in Nb (< 32 ppm) and
431 strongly depleted in Ta (< 2.7 ppm). Biotite in apatite-zircon pegmatites is poorer in Li (~360
432 ppm) than in titanite-allanite pegmatites and tonalitic-trondhjemitic gneiss (~450 ppm, Fig.
433 8d). It has average content in Ta (2.4–14.4 ppm) and Nb (30.4–75.7 ppm), whereas its Nb/Ta
434 ratio vs Ta follows an apparent Rayleigh-type fractionation trend.

435 Biotite of paragneiss and metatexitic paragneiss has a low Nb content (~20 ppm) and a high
436 Nb/Ta ratio. On the contrary, biotite of the micaschist enclave is extremely enriched in Li
437 (1313–2274 ppm), Rb (1530–2544 ppm), with a very low K/Rb ratio. It is also enriched in Nb
438 and Ta (123–142 ppm Nb, 50–61 ppm Ta).

439

440 **4.3.2. Geochemical signatures of apatite**

441 Apatite from granodioritic gneiss, tonalitic-trondhjemitic gneiss, two-mica trondhjemitic,
442 titanite-allanite pegmatites, apatite-zircon pegmatites and garnet-columbite pegmatites
443 consists of fluorapatite, with an F content ranging from 2.1 to 3.7 wt% (Fig. 9a,
444 supplementary Table S7). In apatite-zircon pegmatites, in the micaschist enclave and in the
445 two-mica trondhjemitic, the MnO content of apatite is below 0.72 wt%. The highest Mn
446 contents (MnO~2.5 wt%) are reached in F-rich apatite from a garnet-columbite pegmatite.

447 Distinctions between apatite from different host rocks can be established based on trace
448 element composition (supplementary Table S8). Titanite-allanite and apatite-zircon apatite
449 have identical REE enrichment trends (Fig. 9b). They are depleted in light rare earth elements
450 (LREE) over a large magnitude, whereas medium rare earth elements (MREE) and heavy rare
451 earth elements (HREE) follow a slight negative slope. Europium anomaly is very pronounced
452 ($0.18 < \text{Eu}/\text{Eu}^* < 0.05$, Fig. 9b).

453 Chondrite-normalized REE patterns (McDonough and Sun 1995) of apatite in garnet-
454 columbite pegmatites is characterized by a general negative slope from LREE to HREE, a
455 relative peak in MREE (from Sm to Tb), and a weak Eu anomaly (Eu/Eu^* , 0.57–0.96), (Fig.
456 9c). The HREE, Sr and Y are strongly depleted and Mn and Pb are higher relative to the other
457 pegmatites. Apatite from the micaschist enclosed in garnet-columbite pegmatite has a
458 negative shallow dipping REE pattern with a weak Eu anomaly, but higher REE content than
459 apatite from garnet-columbite pegmatites (Fig. 9c). It is also depleted in Ti compared to
460 pegmatitic apatite. Apatite from two- mica trondhjemitic is depleted in REE and other trace
461 elements as Mn, Pb, Sr and Y, with a pronounced negative Eu anomaly (Eu/Eu^* , 0.25–0.31).

462

463 4.3.3. Geochemical signatures of CGM

464 Nb-Ta-oxides are exclusively observed in garnet-columbite pegmatites, in the central and in
465 the south-western parts of the district (Fig. 3b). Their composition plots mainly in the
466 ferrocolumbite domain, but very close to the mid-limit of the Mn/(Mn +Fe) ratio (Mn#) of 0.5
467 (Fig. 10a). The highest Ta concentrations are found in a pegmatite situated in the central part
468 of the district, (sample BMN13, Fig. 2), with a $Ta\# = Ta/(Ta + Nb)$ ratio up to 0.6, whereas in
469 other samples, $Ta\#$ is inferior to 0.2 (supplementary Table S9).

470 Columbite-group minerals in SEM imagery generally reveal homogeneous cores surrounded
471 by sharp or corroding overgrowths, with oscillatory and patchy textures related to variations
472 in $Ta\#$ and Mn# (supplementary Fig. S11). The cores represent the primary magmatic stage of
473 crystallization, and have a higher Fe content relative to Mn. Overgrowths are characterized by
474 higher Nb contents ($Ta\# < 0.28$), a slight Ta decrease, and a sharp Mn# increase. These
475 fractionation paths reflect complex chemical fractionation or late-stage processes that leads to
476 reverse and oscillatory zoning in most pegmatitic CGM (Fosso Tchunte et al. 2018; Fuchsloch
477 et al. 2019; Neiva et al. 2015; Van Lichtervelde et al. 2007, 2018).

478 The total REE content in CGM ranges from 160 ppm to 2079 ppm (supplementary Table
479 S10). Chondrite-normalized patterns (McDonough and Sun 1995) are characterized by a
480 relative enrichment in HREE compared to LREE (Fig. 10b). Lanthanum, Ce, Pr and Nd cover
481 a large range (two orders of magnitude) of compositions, inferior to 1 ppm up to 311 ppm,
482 whereas the HREE range from 10 ppm to 208 ppm. The REE patterns have very strong
483 negative Eu anomalies. Compared with the global columbite group minerals (CGM) average
484 of Melcher et al. (2015), our columbite-tantalite are strongly enriched in Mg, Sc, Y, Yb and
485 Th, and depleted in Sn (Fig. 10c).

486

487 **4.3.4. Geochemical signatures of garnet**

488 Garnet from paragneiss and metatextitic paragneiss has a composition of metamorphic garnet
489 from metapelitic rocks (Aubrecht et al. 2009; Krippner et al. 2014; Mange and Morton 2007;
490 Suggate and Hall 2014), with almandine (Alm) component between 74% and 80%, pyrope
491 (Pyr) component between 9% and 20%, grossular (Grs) component inferior to 7%, and
492 spessartine (Sps) component between 2.9% and 9.0% (supplementary Table S11). This garnet
493 is depleted in LREE, Rb, U, Th, Nb and U, and shows a slight negative Eu anomaly (Fig. 11b,
494 supplementary Table S12). Its budget in HREE is inferior to the average REE content of
495 pegmatitic garnet (Fig. 11b).

496 Garnet from garnet-columbite pegmatites is dominated by Sps (43–73%) whereas Alm
497 component ranges between 22% and 49%. Pyrope and Grs components remain low (Pyr <
498 5.8%, Grs < 4.3%). Internal Fe/Mn zoning is noticed, with a decrease of Mn compared to Fe
499 from core to rim, which is referred to as “normal zoning” (Dahlquist et al. 2007; Leake 1967;
500 Manning 1983; Miller and Stoddard 1981). This garnet is depleted in LREE, and contains
501 high contents in MREE and HREE, but HREE are depleted relative to MREE (Dy to Lu). It is
502 relatively rich in U (< 21 ppm), Nb (< 237 ppm), Ta (<124 ppm), Zr (< 118 ppm) and Hf (<
503 20 ppm) and has a strong negative anomaly in Eu ($\text{Eu}/\text{Eu}^* < 0.9$ for garnet-columbite
504 pegmatite garnet). Garnet in sample BMS139B (Fig. 2) shows a different feature as it is more
505 Mn-depleted and Fe-rich ($51\% < \text{Alm} < 53\%$, $35\% < \text{Sps} < 38\%$). It has a Grs component
506 around 4% and a slightly higher Mg content ($7\% < \text{Pyr} < 8\%$). Also, it shows a flat pattern for
507 MREE and HREE, which distinguishes it from common garnet-columbite pegmatite trends.

508 Garnet from garnet-REE pegmatites mainly differs from that of garnet-columbite pegmatites
509 by its higher Ca content ($2.9 \text{ wt}\% < \text{CaO} < 3.7 \text{ wt}\%$ in garnet-REE pegmatites, while $\text{CaO} <$
510 $1.48 \text{ wt}\%$ in garnet-columbite pegmatites, see Fig. 11a and supplementary Table S11).

511 Grossular component ranges between 8% and 11%. Its REE content is depicted by strong
512 depletion in LREE, a positive pattern toward HREE, and a strong Eu anomaly ($0.05 < \text{Eu}/\text{Eu}^*$
513 < 0.1). Yttrium, Ti and Li are higher in garnet-REE pegmatites (2903–17240 ppm, up to 3930
514 ppm and 113–518 ppm, respectively) than in garnet-columbite pegmatites (18–9910 ppm Y,
515 $\text{Ti} \leq 1380$ ppm, 49–292 ppm Li, supplementary Table S12).

516 Overall, Nb, Ta, U, Zr and Hf concentrations in garnet are higher in garnet-columbite
517 pegmatites whereas Li, Y, Ti, Rb and REE contents are higher in garnet from garnet-REE
518 pegmatites (Fig. 11d).

519

520 **4.3.5. Geochemical signatures of zircon**

521 Chondrite-normalized REE patterns of zircon from titanite-allanite and apatite-zircon
522 pegmatites, as well as micaschist enclave in garnet-columbite pegmatite, have a positive
523 slope, which is shallow for LREE, and steep for HREE, with a slight negative Eu anomaly
524 and a slight positive Ce anomaly (Fig. 12). Other trace elements in zircon attain 1.9 wt.% Hf,
525 62 ppm Nb, 7450 ppm U, and 2980 ppm Y (supplementary Table S13). Interestingly, the
526 highest Nb, Ta, Th and Be contents are observed in zircon from a micaschist enclave (Fig. 12
527 and supplementary Table S13). Titanium (992–2220 ppm) is exceptionally elevated compared
528 to available data on magmatic zircon (Fu et al. 2008). Such high Ti content in zircon has been
529 attributed to contamination along cracks (Harrison and Schmitt 2007).

530 Some zircon analyses in a titanite-allanite sample (BMN28, Fig. 2) show low LREE and Th
531 contents (Fig. 12), accompanied by low contents in Al, Fe, Ba, Sr, Y, and Bi. These spots are
532 located in non-metamict zones in the core or internal part of zircons, with homogeneous to
533 oscillatory texture according to contrast of SEM imaging.

534 Zircon from apatite-zircon pegmatites has higher content in HREE, Y, U and Th than that of
535 titanite-allanite pegmatites (Fig. 12, supplementary Table S13 and Table S14). Zircon
536 metamictization in apatite-zircon pegmatites induced loss of major elements like Zr and Si,
537 and an increase in Ca.

538

539 **5. Discussion**

540 **5.1. Classification of the Mangodara pegmatites**

541 **5.1.1. Titanite-allanite and apatite-zircon pegmatites**

542 Titanite-allanite and apatite-zircon pegmatites are very similar in mineralogy, except that in
543 the latter type, ilmenite and titanite are absent, and apatite is more abundant (Fig. 4). The
544 profusion of minerals rich in Fe-Mg-Ti-Ca (biotite, magnetite, ilmenite, oligoclase, titanite)
545 indicates a metaluminous composition of the pegmatite, and a low differentiation degree
546 (Broska et al. 2000; Gieré and Sorensen 2004; Neiva et al. 2002; Villaros and Pichavant
547 2019). Allanite, an accessory phase in titanite-allanite pegmatites, is typically found in NYF
548 pegmatites related to metaluminous granites, defined as an allanite-subtype (enrichment in
549 LREE, Nb over Ta, Ti and Zr, Wise 1999). Apatite, which is common in LCT pegmatites but
550 is also reported as a primary phosphate in NYF pegmatites (Nizamoff et al. 1999), constitutes
551 a LREE-host in addition to allanite in titanite-allanite, and the main REE host in apatite-
552 zircon pegmatites.

553 **Titanite-allanite and apatite-zircon pegmatites lack the characteristic complex**
554 **mineralogy of mineralized pegmatites, and do not exhibit mineralization in Li, Nb or Ta**
555 **(Tables**

556 Table 1). Accordingly, they are classified as poorly enriched metaluminous pegmatites, with
557 apatite-zircon pegmatite as the most differentiated member of the group.

558

559 **5.1.2. Garnet-columbite pegmatites**

560 **Garnet-columbite pegmatites exhibit a typical mineralogy for peraluminous LCT**
561 **pegmatites (muscovite > biotite, garnet, tourmaline, apatite, see Fig. 4), but both LCT**
562 **and NYF signatures are revealed by the geochemistry of their minerals (Nb > Ta in**
563 **micas, CGM and garnet, REE-rich accessory minerals ...), (Tables**
564 Table 1). The high Nb contents in muscovite (up to 438 ppm Nb), are typical of white micas
565 from LCT pegmatites (up to 490 ppm, Alfonso et al. 2003, Viana et al. 2007). Remarkably
566 high Nb contents are recorded in garnet. Although experiments show that Nb is incompatible
567 in garnet (Fulmer et al. 2010, and references therein), Nb-rich garnet has also been reported in
568 CGM-bearing LCT pegmatites (e.g., Samadi et al. 2014). Columbite-group minerals has
569 intermediate Mn# and Ta# compositions (Fig. 10) that are typical of the beryl-columbite
570 pegmatite subtype (Černý 1989; Novák et al. 2003; Uher et al. 2010), or NYF pegmatites
571 (Melcher et al. 2017). Although beryl is missing in garnet-columbite type pegmatites, Be is
572 contained in micas up to 10 ppm (supplementary Table S5 and Table S6).

573 **The occurrence of REE-bearing minerals such as monazite, xenotime, and relicts of**
574 **metamict minerals rich in U, Th and Y (thorite/cheralite) is evidence of a moderate**
575 **enrichment in REE of this pegmatite type (Tables**

576 Table 1). Columbite composition compared with average global CGM (Melcher et al. 2015) is
577 enriched in Y, HREE, Th and Sc, and depleted in Sn (Fig. 10c), which is typical for CGM
578 from NYF and mixed LCT+NYF pegmatites. Its high REE content and its medium to low
579 U/Th ratio also match the CGM composition of NYF and mixed LCT+NYF families
580 (Melcher et al. 2017). Only the depletion in W is incoherent with general tendencies observed
581 in the NYF type (Melcher et al. 2017).

582 **Although garnet-columbite pegmatites do not host Ta, Cs or Sn mineralization typical of**
583 **LCT pegmatites, geochemical features like Nb>Ta and REE-U-Th enrichment trend in**
584 **accessory minerals (Tables**
585 **Table 1) allow classifying them in the mixed LCT+NYF family.**

586

587 **5.1.3. Garnet-REE pegmatites**

588 **Garnet and tourmaline, which occur in our garnet-REE pegmatites, are characteristic of**
589 **LCT pegmatites But such minerals are also reported from numerous NYF pegmatites,**
590 **where they are stabilized by high Ca, Fe and Ti contents in the rock (Čopjaková et al.**
591 **2013; Hönig et al. 2014; Novák et al. 2011). Metamict Nb-Ti-Th-Y-bearing minerals of**
592 **composition close to euxenite-aeschynite coexist with Ti-oxides (ilmenite-pyrophanite),**
593 **which are typical mineral species of NYF pegmatites (Černý 1991b; Černý et al. 2005;**
594 **Ercit et al. 2005). The high Y and REE contents of garnet are similar to existing data of**
595 **garnet from NYF pegmatites in Sveconorwegian province (Hönig et al. 2010, 2014;**
596 **Müller et al. 2012). High Ca content of garnet is also comparable with the garnet Ca**
597 **content of some NYF pegmatites, mixed LCT+NYF pegmatites or A-type pegmatites**
598 **(Fig. 11a). The Ca-rich garnet, the exotic REE-mineral species and geochemical**
599 **enrichment trend in REE, Y and Ti (Tables**

600 **Table 1) are mostly described from NYF pegmatites (Černý et al. 1985; Müller et al. 2012,**
601 **2018; Novák et al. 2011, 2012; Pieczka et al. 2013). Hence, garnet-REE pegmatites can be**
602 **classified as NYF pegmatites, more specifically, those of the euxenite subtype (Wise 1999).**

603

604 **5.2. Origin of the geochemical diversity of Mangodara pegmatites**

605 **5.2.1. Petrogenesis of metaluminous pegmatites by melt segregation within the tonalitic** 606 **gneiss**

607 Titanite-allanite and apatite-zircon pegmatites were proposed to derive from a residual melt
608 segregated during syntectonic crystallization of the tonalitic-trondhjemitic gneiss, based on
609 field relationships. Indeed, titanite-allanite type pegmatites occur as a network of veins
610 concordant to discordant relative to the foliation of the tonalitic-trondhjemitic gneiss and in
611 continuity with quartz-feldspar veins with a magmatic texture interpreted as syntectonic
612 segregated melts (Bonzi et al. 2021). The apatite-zircon type of pegmatites is characterized by
613 a very similar mineralogy to that of the titanite-allanite type, and was interpreted as an
614 equivalent that crystallized out of the tonalitic-trondhjemitic gneiss. (Bonzi et al. 2021). The
615 mineral chemistry results presented in this work, particularly the overlapping biotite
616 compositions (Fig. 6, Fig. 7), the continuity in the K/Rb fractionation trend of biotite (Fig. 8b)
617 and the unique REE profiles of apatite (Fig. 9b) from tonalitic-trondhjemitic gneiss, titanite-
618 allanite and apatite-zircon pegmatites corroborate this hypothesis (Hulsbosch et al. 2014;
619 Villaros and Pichavant 2019).

620 In titanite-allanite pegmatites, allanite crystallization precedes the stabilization of other
621 LREE-bearing phases such as titanite and apatite. Titanite or epidote coronae around ilmenite
622 nuclei and crystallization of apatite on allanite crystal borders could reflect late reactions
623 induced by a late P- and Ca-rich pegmatitic melt or fluid, which remobilized Fe (Angiboust
624 and Harlov 2017; Broska et al. 2007) and Ti, leading to the reaction: ilmenite (Fe, Ti) +
625 Allanite (Fe, REE) + Ca + P → Titanite (Ti, Ca) + Epidote (Fe, Ca) + apatite (Ca, REE).
626 Zircon from titanite-allanite pegmatites is characterized by LREE-depleted cores that could be
627 attributed to metamict alteration, as suggested by the increase in Ca, Al, Fe, Ba and LREE of
628 the rims compared to the nuclei (Geisler et al. 2003; Hoskin and Schaltegger 2003), or
28

629 alternatively by concurrent segregation of LREE in allanite. The metaluminous pegmatite-
630 forming melt firstly evolved through primary segregation of REE in allanite and zircon, and
631 then evolved toward phosphorus saturation materialized by late crystallization of apatite.

632 From titanite-allanite to apatite-zircon pegmatites, the increase in F, Mn and Rb in biotite as
633 well as the continuous K/Rb, Nb/Ta and Fe/Mn fractionation trends in micas (Fig. 6, Fig. 7g,
634 and Fig. 8b and f) can be explained by fractional crystallization. However, the decreasing Li
635 trend from titanite-allanite to apatite-zircon biotite suggests the exsolution of a Li (+ minor
636 F)-rich aqueous fluid early during apatite-zircon pegmatite crystallization. Apatite tends to be
637 richer in F and Mn in apatite-zircon pegmatites compared to titanite-allanite pegmatites (Fig.
638 9a). This behavior could reflect magmatic differentiation coupled to a lower content of mafic
639 minerals in apatite-zircon pegmatite (Belousova et al. 2001), and/or increasing aluminosity in
640 the residual melt during fractional crystallization (Chu et al. 2009; Miles et al. 2014).

641

642 **5.2.2. Petrogenesis of mixed NYF-LCT garnet-columbite pegmatites**

643 *Partial melting of metasediments*

644 Garnet-columbite pegmatites have been proposed as the product of partial melting of
645 metasediments, based on field relationships. Indeed, garnet-columbite type pegmatite dykes
646 are in textural continuity with leucosome veins of metatextitic paragneiss (Bonzi et al. 2021).
647 Their peraluminous composition is typical of S-type granites and supports a metasedimentary
648 source (Chappell and White 2001; Clemens 2003). Biotite and muscovite from paragneiss and
649 metatextitic paragneiss, despite their minor content in Nb and trace contents in Li and Ta,
650 could be the sources of Nb, Li, fluxing elements such as F and H₂O, and LILE such as Rb
651 (Neves 1997; Stepanov et al. 2014; Stepanov and Hermann 2013). Apatite and monazite
652 (reservoirs for REE, U and Th) are found in paragneiss, and more abundantly in the
29

653 micaschist enclave enclosed in garnet-columbite pegmatite and in the metatextitic paragneiss
654 (supplementary Fig. S10). These phosphates are likely to be consumed by partial melting
655 reactions in the presence of Al-rich silicate melt, during heating and/or retrograde
656 metamorphic paths (Bea et al. 1992; Harrison and Watson 1984; Pichavant et al. 1992; Rapp
657 and Watson 1986; Watson 1980; Watson and Capobianco 1981; Wolf and London 1995;
658 Yakymchuk 2017). Garnet of paragneiss and metatextitic paragneiss, which has high contents
659 in HREE and Y but is depleted in LREE (Fig. 9c), would act, together with ilmenite, as a
660 HREE sink during prograde metamorphism of paragneiss, but can supply the anatectic melt
661 with HREE and Y during the retrograde path.

662 A possible alternative granitic source for the peraluminous garnet-columbite pegmatites is the
663 two-mica trondhjemite. The two-mica trondhjemite is interpreted as the product of melt
664 accumulation in the upper part of the tonalitic-trondhjemitic gneiss, formed by fractionation
665 of plagioclase, biotite and amphibole within the crystallizing gneiss that enhanced its
666 aluminosity (Bonzi et al. 2021). Its mineral geochemistry reveals high Li and Nb (Bonzi et al.
667 2021), moderate K/Rb in micas (Fig. 8b and c), and F-rich apatite. However, one would
668 expect a gradational evolution in texture of the two-mica trondhjemite into a pegmatitic
669 texture, as documented for the apical parts of some granitic intrusions (Černý 1991a, c; Černý
670 et al. 2005), and this was not observed in the field at Mangodara. Also in this case, micas
671 from garnet-columbite pegmatites should be richer in Li, and at least follow a geochemical
672 continuity such as the K/Rb or Nb/Ta Rayleigh trend (Černý et al. 1985; Černý and Burt
673 1984; Foord et al. 1995; Martins et al. 2012; Roda-Robles et al. 2006; Stepanov et al. 2014;
674 Stepanov and Hermann 2013), or a compositional evolution path in major elements
675 (Tischendorf et al. 2004) with the micas from the two-mica trondhjemite. Likely, garnet-
676 columbite pegmatite essentially derives from partial melting of the paragneiss, preferably

677 during the retrograde decompression path associated with exhumation, implying breakdown
678 of micas and entrainment of phosphates and garnet in the melt.

679

680 *Mixing of peraluminous and metaluminous melt*

681 Interaction between apatite-zircon and garnet-columbite pegmatites is suggested by the
682 observation of a garnet-columbite pegmatite cross-cutting at a shallow angle an apatite-zircon
683 pegmatite, showing continuous magmatic texture at the microscale (station BMS99A, Fig. 2,
684 see Bonzi et al. 2021). At the contact with garnet-columbite pegmatite, apatite-zircon
685 pegmatite displays biotite with higher Mn, F, and lower Fe + Mg contents than the average
686 biotite in any apatite-zircon pegmatite (Fig. 7d, f, and g), which might represent an overprint
687 by the garnet-columbite pegmatite-forming melt. Likewise, apatite crystals of a garnet-
688 columbite pegmatite (sample BMN13, Fig. 2) reveal a typical signature of apatite from
689 apatite-zircon pegmatite (Fig. 9b). These crystals would represent an early magmatic phase
690 inherited from an apatite-zircon pegmatite and assimilated into the peraluminous LCT-
691 affiliated melt, but which conserved its REE geochemical signature. Given the continuous
692 textural relation between the two pegmatite types, their interaction might have occurred
693 before the end of crystallization of the apatite-zircon pegmatite, implying mingling of residual
694 metaluminous pegmatite-forming melt with LCT pegmatite-forming melt. Then,
695 crystallization of the garnet-columbite occurred almost synchronously, or shortly after the
696 apatite-zircon pegmatite crystallization.

697 In this situation, the incompatible elements (i.e., Nb and REE) accumulated in the residual
698 metaluminous pegmatite-forming melt might have been incorporated in the LCT pegmatite-
699 forming melt during magma mixing. Alternatively, dissolution of biotite and apatite from
700 apatite-zircon pegmatite could have liberated Nb and REE (together with P) into the LCT

701 pegmatite-forming melt. In any case, CGM, which has a higher solubility in NYF-type (i.e.,
702 peralkaline) melts than in peraluminous melts (Chevychelov et al. 2010, 2020; Linnen and
703 Keppler 1997), attained its saturation in the peraluminous LCT-type melt, hence provoking
704 the crystallization of low-Ta# and REE-rich columbite.

705

706 *Magmatic fractionation of rare-metals*

707 The occurrence of tourmaline and the high Li and Rb contents in micas are evidence for
708 concentration of volatile elements in the pegmatite-forming melt (e.g., Černý et al. 1985;
709 Hulsbosch et al. 2014; Jolliff et al. 1987, 1992; London 2011; Viana et al. 2007; Villaros and
710 Pichavant 2019). Occurrence of phosphates indicates that P saturation is reached during
711 pegmatite crystallization (Duc-Tin and Keppler 2015; Van Lichtervelde et al. 2021). The
712 steeply increasing HREE trend in chondrite-normalized patterns of zircon confirms the early
713 removal of HREE from the melt by zircon crystallization. Early segregation of HREE and Y
714 in zircon and xenotime is indicated by their depletion in garnet, CGM and apatite (Fig. 11.c).
715 The presence of CGM as inclusions in major minerals (quartz, feldspar, muscovite) shows
716 that saturation in Nb-Ta minerals was reached in the early stage of crystallization. The
717 absence of Rayleigh type fractionation between Nb and Ta in muscovite suggests that the Nb-
718 Ta content was buffered by CGM crystallization, which is in agreement with the spatial
719 association between CGM and muscovite (Fig. 8e). Nb-rich rims of CGM crystals reflect the
720 late Nb enrichment in the pegmatite-forming melt. This enrichment could characterize the
721 transition of the melt to critical fluxing-element-rich fluids (Kaeter et al. 2018; Van
722 Lichtervelde et al. 2007; Zraisky et al. 2010; Zhang et al. 2004) or the mixing with another
723 rare-metal rich melt such as the residual metaluminous melt at the origin of the apatite-zircon
724 pegmatites.

725 The crystallization of garnet-columbite pegmatites obeys the trend of sequential rare-metal
726 saturation, i.e., Zr (zircon), REE-U-Th (xenotime, monazite, apatite, thorite) and Nb-Ta
727 (CGM). This sequence stands apart from the regular differentiation trend of peraluminous
728 melts, in which REE and Zr are rather segregated in cumulated crystals (zircon, phosphates,
729 micas), leaving a residual melt with decreasing REE, Zr, and increasing Ta and Sn (Ballouard
730 et al. 2016, 2020; Linnen and Cuney 2005). Instead, the primary enrichment in REE, U, Th
731 and Zr might reflect the primitive composition of the anatectic pegmatite-forming melt, which
732 crystallized close to its partially molten source without facing significant magmatic
733 fractionation (Fuchsloch et al. 2018). In other words, the primary trace element geochemistry
734 of garnet-columbite pegmatite is closer to the geochemistry of an undifferentiated melt than to
735 a residual melt from magmatic fractionation. Other authors highlight multiple processes such
736 as contamination by host rocks (e.g. with Fe, Mg, Ti, Sn, Ca...), or melt mixing (Galliski et
737 al. 2008; Novák et al. 2012; Pieczka et al. 2013). Melt mixing might indeed affect the trace-
738 element geochemistry of the pegmatite-forming melt, but more investigation is needed to test
739 contamination of pegmatite-forming melt by host rocks.

740

741 *The case of F-Li rich micas in micaschist enclosed in garnet-columbite pegmatite*

742 A micaschist enclave in a garnet-columbite pegmatite (station BMS99A) contains micas that
743 are more enriched in F, Li, Rb, Nb and Ta, than the micas in the host pegmatite (Fig. 7e and f,
744 Fig. 8). The compositional gap between micaschist micas and micas from paragneiss shows
745 that the former micas underwent a specific enrichment in incompatible elements, unrelated to
746 metamorphism. Two hypotheses might be formulated to explain the formation of these
747 enriched micas. (1) The micaschist is a residue left after partial melting of an enriched
748 paragneissic protolith. This proposal implies that high amounts of LILE (Li, Rb), fluxing

749 elements (F) and HFSE (Nb, Ta) are partitioned in residual micas, which would challenge
750 their incompatible character. This proposition is consistent with the presence of paragneiss
751 and migmatitic paragneiss as country rocks of the sampled pegmatites (e.g. BMS96, also
752 BMS7B in Bonzi et al. 2021). (2) The micaschist is an enclave of paragneiss metasomatized
753 by highly fractionated fluids exsolved from the host garnet-columbite pegmatite. In this case,
754 the presence of Li-rich mica might reflect the metasomatism occurring in the rocks
755 surrounding the pegmatite dyke. Exomorphic micas and hornblende formed during
756 metasomatism, or by reaction of existing minerals to produce secondary minerals, are
757 reported in some pegmatite wall rocks (Morgan and London 1987; Shearer et al. 1986).

758

759 **5.2.3. Petrogenesis of the NYF garnet-REE pegmatites by melt segregation within the** 760 **granodioritic gneiss**

761 Bonzi et al. (2021) interpreted garnet-REE pegmatites to be a product of melt segregation
762 within the granodioritic gneiss, based on textural continuity with leucosomes. The distinct
763 composition of their micas compared to other pegmatites (higher Fe content, F-depletion; Fig.
764 6 and Fig. 7) and the absence of phosphates confirm that their source is clearly different from
765 that of the other pegmatite types. The whole-rock composition of garnet-REE pegmatites
766 shows a depletion in LREE relative to the granodioritic gneiss, but a similar content in HREE,
767 which is not consistent with extreme fractionation during crystallization. The genesis of such
768 pegmatite-forming melts would require a selective segregation of LREE in the parental
769 gneiss, whereas HREE and Y are released in the melt. The low P content in garnet-REE
770 pegmatite excludes the dissolution of phosphates as a provider of REE to the melt. On the
771 contrary, phosphates and allanite might retain LREE in the parental gneiss. Hornblende, by
772 virtue of its high modal composition in the granodioritic gneiss, could supply the pegmatite-

773 forming melt with Ti, Zr, REE and Y (Bea 1996; Bernard et al. 2020; Green 1994) if it is
774 destabilized during melt segregation, for example through the replacement by biotite (low-
775 REE-bearing phases).

776 Based on these considerations, we propose that the genesis of garnet-REE pegmatite parental
777 melt corresponds to the segregation of residual melt within the granodioritic gneiss, enriched
778 in HFSE, but depleted in LREE preferentially trapped in allanite and apatite that remained in
779 the granodioritic gneiss. A probable explanation for the low P content in these pegmatites is
780 the primary compatibility of phosphate in the high-Si granodioritic gneiss, which prevents the
781 release of P in the residual melt (Bea et al. 1992; Broska and Petřík 2008). In the absence of
782 P, REE and Y are incorporated in euxenite-aeschynite-like minerals and garnet (Fig. 11c).
783 The source of the Li enriched in garnet-REE pegmatites is unclear. The most probable
784 reservoir in the granodioritic gneiss might be hornblende (Shaw et al. 1988) that we expect to
785 be destabilized during the pegmatite-forming melt segregation. However, Li is generally
786 concentrated in calcic amphibole from peralkaline magmatism (e.g., Bernard et al. 2020;
787 Hawthorne et al. 2001; Oberti et al. 2003; Tiepolo et al. 2007), but it might also be hosted by
788 biotite which is a constituent of granodioritic gneiss. Li enrichment in garnet may be
789 explained by a preferential partitioning in this pegmatite type induced by limited amount of
790 micas which are the main Li host (Dutrow et al. 1986), or by preferential substitution with Y
791 in the garnet structure (Cahalan et al. 2014).

792

793 **5.3. Petrogenetic model of LCT+NYF Mangodara pegmatite formation**

794 It has been proposed in several studies of Precambrian terranes that the generation of REE-
795 rich LCT and NYF pegmatites was related to partial melting in the middle crust occurring at
796 different stages of the orogenic cycle (Müller et al. 2012; Turlin et al. 2018, 2019). It has been

797 highlighted that the petrogenetic family of a rare-metal-bearing pegmatite is linked to the
798 metamorphic grade of its source, supracrustal for LCT pegmatites (metasediments, $P < 15$
799 km), deeper for NYF pegmatites (amphibolite/granulite domains, $P > 15$ km) (Müller et al.
800 2017).

801 The Mangodara pegmatite field displays a different succession of petrogenetic families of
802 pegmatites. At the lowest structural level (tonalitic-trondhjemitic gneiss), titanite-allanite
803 pegmatites are interpreted to represent a residual melt segregated during crystallization of the
804 tonalitic-trondhjemitic gneiss coring the Mangodara dome and intruding the granodioritic
805 gneiss. The plagioclase-rich assemblage of this gneiss and its relatively low Fe, Mg, Nb, Ta
806 and REE contents are attributed to accumulation of residual pyroxene and garnet in its deeper
807 source (Bonzi et al. 2021) (Fig. 13). This unit represents a LREE-depleted section of the
808 middle crust (Laurent et al. 2020; Moyen 2011), deriving from HP/HT conditions of partial
809 melting, in which melt segregation resulted in poorly enriched metaluminous pegmatites.

810 The intermediate structural level (granodioritic gneiss, deriving from LP/HT conditions of
811 partial melting) hosts apatite-zircon pegmatites that crystallized from the metaluminous melt
812 that migrated out of the tonalitic gneiss and intruded at higher structural level. This level also
813 hosts garnet-columbite pegmatites deriving from partial melting of a mica-rich source, which
814 might correspond to the paragneiss rafts included in the granodioritic diatexite (Bonzi et al.
815 2021). Partial melting can be achieved in these rafts by dehydration of micas (mainly biotite)
816 (Fig. 13). Additional REE and Nb are supplied to these pegmatites by the mingling of
817 metaluminous pegmatite-forming melt with peraluminous pegmatite-forming melt.

818 Segregation of residual melt within the granodioritic diatexite, protolith of the granodioritic
819 gneiss, resulted in the generation of NYF pegmatite-forming melt that intruded plutonic rocks
820 crystallized at a higher structural level (Fig. 13). We explain the distinct geochemistry of

821 garnet-REE pegmatites by LREE segregation by allanite and apatite within the gneiss (Fig.
822 13) and amphibole breakdown that supplied the melt in HFSE. Consequently, garnet-REE
823 pegmatites are not the end-members of a regional zoning related to the other pegmatite types,
824 but the differentiated member of the K-rich magmatic series.

825 Overall, the petrogenesis of Mangodara pegmatites is related to the mineral composition of
826 the partially molten source that is in turn controlled by (i) the chemical composition of the
827 protoliths (ii) the PT conditions of partial melting and the minerals involved in the melting
828 reactions and (iii) the degree of partial melting and fractional crystallization of the partially
829 molten rock. The varying composition of the protolith between paragneiss and amphibolite is
830 reflected by the abundance of micas and amphiboles, respectively, which are sources of
831 fluxing elements (Li, F, H₂O) favoring growth of pegmatitic crystals (London 2018; London
832 and Morgan 2017; Sirbescu et al. 2017; Webber et al. 1999). Micas, prevalent in
833 metasediments, are reservoirs for LCT-affiliated rare elements (Rb, Nb, Ta) whereas
834 amphiboles, prevalent in amphibolites, are reservoirs for NYF-affiliated rare elements (REE,
835 Zr, Ti). Entrainment of garnet and phosphates in melt during partial melting/segregation
836 would provide REE, Y, U, Th, and P shifting the melt composition toward a mixed
837 LCT+NYF signature. Moreover, the hybrid LCT+NYF signature is enhanced by the poor
838 differentiation of the melt fraction that directly crystallized after leaving its source, and by the
839 mingling of the residual metaluminous melt with the peraluminous melt.

840

841 **6. Conclusions**

842 The mineralogy and the geochemical signatures of micas, garnet and accessory minerals from
843 Mangodara pegmatites permit to define their petrogenetic families, discuss their sources by

844 partial melting of paragneiss/amphibolite and segregation within granitoid gneiss, and identify
845 fractionation of rare-metals during crystallization.

846 (1) Poorly enriched titanite-allanite pegmatites correspond to residual metaluminous melt
847 segregated within the tonalitic-trondhjemitic gneiss. Their rare-metal content is
848 initially controlled by allanite, then by zircon and apatite. Residual melt migrated out
849 of the tonalitic-trondhjemitic massif and intruded at higher structural level in the
850 granodioritic gneiss, forming apatite-zircon pegmatites, which are slightly richer in P
851 and REE, but poorer in Li.

852 (2) Garnet-columbite type pegmatites are classified in the LCT+NYF family, with
853 enrichment in Li, Nb > Ta, Th, U and REE. It derives from partial melting of
854 paragneiss, in which micas are sinks of Li, Nb and Ta, phosphates are sinks of REE, U
855 and Th, and garnet is a sink of REE. The mingling of the garnet-columbite parental
856 melt with residual melt originating from the crystallization of metaluminous
857 pegmatites is suggested by field relationships and the inherited signature of apatite in a
858 garnet-columbite pegmatite. Locally, strong melt fractionation led to the exsolution of
859 a Li-bearing fluid, as suggested by the presence of micas highly enriched in rare-
860 metals in an enclave of metasediment in garnet-columbite pegmatite.

861 (3) Garnet-REE-bearing pegmatites are NYF pegmatites showing enrichment trends in Li,
862 Ti, Y and HREE, and strong depletion in P. Their origin by melt segregation within
863 the granodioritic gneiss requires that P and LREE are segregated in the parental rocks
864 by allanite and apatite, whereas HFSE are supplied by breakdown/replacement of
865 hornblende.

866 (4) The LCT versus NYF nature of rare-metal-bearing pegmatites is explained by the
867 mineral composition of their partially molten source, which is in turn controlled by

868 lithology-chemistry of the protoliths and by PT conditions of partial melting: mica-
869 rich for LCT pegmatites (metasediments), and amphibole-rich for NYF pegmatites
870 (e.g., a granodioritic gneiss in Mangodara). The mixed LCT + NYF signature of
871 garnet-columbite pegmatites reflects the poor differentiation of anatectic pegmatite-
872 forming melt after extraction from the source, and also the mingling with residual
873 metaluminous melt coming from the deeper tonalitic-trondhjemitic gneiss. Then,
874 partial melting and melt segregation in the middle to lower crust could generate
875 pegmatite-forming melt, whose rare metals content depends on the composition of the
876 rock involved in these processes.

877

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894

895 **Conflicts of interest/Competing interests**

896 The authors declare that they have no known competing financial interests or personal
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898

899 **Availability of data and material**

900 The datasets generated during and/or analysed during the current study are available from the
901 corresponding author on reasonable request.

902

903 **Code availability**

904 Not applicable

905

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- 1522
- 1523

1524 **Figure captions**

1525

1526 Fig. 1. Geological map of the West African Craton, modified from Block et al. (2015), Milési
1527 et al. (2004). Abbreviations: Ba= Banfora greenstone belt, Di = Diebougou gneiss-granitoid
1528 complex, Ho= Hounde-Ouango Fitini greenstone belt, KT = Koudougou-Tumu gneiss-
1529 granitoid complex, Si= Sideradougou gneiss-granitoid complex, OFSZ: Ouango-Fitini shear
1530 zone. The area investigated in this study is indicated by the red rectangle.

1531

1532 Fig. 2. Lithostructural map of the Mangodara district; lithofabrics are compiled after aerial
1533 photographs and measures of fabrics in the field. Interpreted shear zones were identified from
1534 aeromagnetic data, using the method proposed by Chardon et al. (2020). Sample location is
1535 indicated in the map.

1536

1537 Fig. 3. (a) Regional distribution of pegmatite dykes in the Mangodara district. The symbols
1538 are oriented parallel to strike, black arrows indicate plunge dips. Size is not to scale. (b)
1539 Regional distribution of occurrence of some accessory minerals in pegmatite dykes. Large
1540 white contours delineate the delimited zone of pegmatite types or mineral occurrences.
1541 Internal white contours delineate subzone with abundant outcrops of pegmatite.

1542

1543 Fig. 4. Abundance of selected major minerals and accessory minerals in Mangodara
1544 pegmatites and their host rocks. Mineral abundance ranges from rare (thinnest line) to minor
1545 (> 5 vol.%, thickest line). Affinity of minerals with metaluminous magma, peraluminous
1546 magmas or NYF-style magma are indicated by a background color.

1547

1548 Fig. 5. Scanning electron microscope back-scattered electron (SEM-BSE) images of
1549 Mangodara pegmatites. (a) Image of the sample BMN43 showing the association of accessory
1550 minerals (allanite, zircon, titanite, epidote, apatite, magnetite). (b) Textural association of
1551 apatite and magnetite with biotite folia in apatite-zircon type pegmatite (sample BMN82B).
1552 (c) Inclusion of columbite and Zr-metamict crystals in garnet (sample BMS1). (d) Inclusion of
1553 ilmenite-pyrophanite in garnet, with metamict Ti-Ca-Nb-bearing mineral (sample BB18).
1554 Abbreviations as in Table 1.

1555

1556 Fig. 6. Composition of micas from pegmatites and host rocks in the Tischendorf diagram
1557 (Tischendorf et al. 2007). Li contents in the calculation of the Tischendorf parameters were
1558 obtained with the formulas of Tischendorf (2004).

1559

1560 Fig. 7. Major element composition of muscovite and biotite of pegmatites and host rocks.
1561 Values are given in atom per formula unit, calculated based on 11 oxygen atoms.

1562

1563 Fig. 8. K/Rb and Nb/Ta ratios compared with Rb, Li, Nb and Ta contents in muscovite and
1564 biotite of pegmatites and various host rocks. Data from different pegmatite families are
1565 plotted for comparison: representative compositions of micas from Li-rich LCT pegmatites
1566 from the Fregeneda area, Spain (Roda-Robles et al. 1999), representative compositions of
1567 micas from the LCT Tanco pegmatite, Canada (Van Lichtervelde et al. 2008), and average
1568 composition of Li-rich micas and siderophyllite/phlogopite from the NYF Evje-Iveland
1569 pegmatites, Sweden (Rosing-Schow et al. 2018).

1570

1571 Fig. 9. (a) Fluorine versus MnO content in apatite from pegmatites and their host rocks. (b)
1572 REE spectra of apatite from titanite-allanite pegmatite and apatite-zircon pegmatite
1573 normalized to chondrite (McDonough and Sun 1995). (c) REE spectra of apatite from two-
1574 mica trondhjemite, garnet-columbite pegmatite and enclosed micaschist, normalized to
1575 chondrite (McDonough and Sun 1995).

1576

1577 Fig. 10. (a) Composition of CGM from garnet-columbite pegmatites in the columbite
1578 quadrilateral $Ta\# = Ta/(Nb+Ta)$ vs $Mn\# = Mn/(Fe + Mn)$. Filled symbols correspond to
1579 crystal cores. Enrichment trends of Nb and Ta are extracted from Černý (1989). (b) REE
1580 spider diagram of CGM normalized to chondrite (McDonough and Sun 1995). Black arrows
1581 indicate HREE depletion. (c) Trace elements composition of CGM, normalized to world
1582 median CGM composition (Melcher et al. 2015).

1583

1584 Fig. 11. (a) MnO versus CaO contents of garnet from paragneiss, metatextitic paragneiss and
1585 pegmatites. Compositional fields are reproduced from Samadi et al. (2014). Data from
1586 different pegmatite families and rare-metal granites are plotted for comparison (Breiter et al.
1587 2018; Dahlquist et al. 2007; Feng et al. 2017; Hönig et al. 2014; Müller et al. 2012, 2018;
1588 Pieczka et al. 2013). (b) REE distribution in garnet from pegmatites, normalized to chondrites
1589 (McDonough and Sun 1995). (c) REE distribution in garnet from paragneiss and garnet-
1590 bearing migmatitic gneiss, normalized to chondrites (McDonough and Sun 1995). Brown and
1591 green arrows indicate the trace element enrichments in garnet from garnet-columbite type

1592 pegmatites and garnet-REE type pegmatites, respectively. (d) Multi-element spider diagram
1593 of garnet, normalized to chondrites (McDonough and Sun 1995).

1594

1595 Fig. 12. REE spider diagram of zircon from titanite-allanite pegmatite, apatite-zircon
1596 pegmatite and enclave of micaschist in garnet-columbite pegmatite, normalized to chondrite
1597 (McDonough and Sun 1995).

1598

1599 Fig. 13. Conceptual cross-section showing the source of Mangodara pegmatites (modified
1600 from Bonzi et al. 2021): late melt segregation within gneisses for titanite-allanite pegmatite
1601 (blue arrows), apatite-zircon pegmatite (red arrows) and garnet-REE pegmatite (green
1602 arrows), and partial melting of paragneiss for garnet-columbite pegmatite (yellow arrow).
1603 BGB = Banfora greenstone belt, SGGC = Sidéradougou gneiss and granitoid complex.

1604

1605 **Tables**

1606 Table 1. Mineralogy of pegmatite samples and their host rocks investigated using scanning
 1607 electronic microscopy (SEM), electron probe microanalyser (EPMA) and (laser ablation –
 1608 inductively coupled plasma – mass spectrometry) LA-ICP-MS.

	Mineralogy	Analyzed minerals (EPMA)	Analyzed minerals (LA-ICP-MS)
Paragneiss			
BMS49	Bt-Mu-Grt-St-Sil-Chl	Bt-Mu-Grt	Bt-Mu-Grt
BMS96	Bt-Mu-Grt-Sil-Chl	Bt-Mu-Grt	Bt-Mu-Grt
Metatextitic paragneiss			
BMN21	Bt-Grt-Sil-Chl-Ilm-Mnz-Zrn	Bt-Grt-Zrn	Bt-Grt
Micaschist enclave in garnet-columbite pegmatite			
BMS6	Bt-Mu-Ap-Zrn-Mnz-Rt	Bt-Mu-Ap	Bt-Mu-Ap
Granodioritic gneiss			
BMN24	Bt-Hbl-Mag-Aln-Ep-Ttn-Ap-Zrn	Bt-Ap	Bt
Tonalitic-trondhjemitic gneiss			
BMN11	Bt-Mag-Ep-Ap-Zrn	Bt	
BMN27	Bt-Mag-Ep-Ap-Zrn	Bt-Ap	Bt
BMN164H (hornblende tonalite vein)	Bt-Hbl-Mag	Bt	Bt
Hornblende-biotite granodiorite with mica-rich enclaves			
BMS46	Bt-Hbl-Mag-Aln-Ep-Ttn-Ap	Bt	
Two-mica trondhjemitic			
BMS104	Mu-Bt-Mag-Ep-Ap	Bt-Mu-Ap	Bt-Mu-Ap
Titanite-allanite type pegmatite (2 samples investigated on 63 stations)			
BMN28	Bt-Mu-Mag-Ep-Zrn-Ap	Mu-Bt-Ap-Zrn	Mu-Bt-Ap-Zrn
BMN43	Bt-Mag-Ilm-Aln-Ep-Ttn-Zrn-Ap	Bt-Ap-Kfs-Pl	Bt-Ap
Apatite-zircon type pegmatite (3 samples investigated on 4 stations)			
BMN20	Bt-Mag-Ap-Zrn	Bt-Ap-Zrn	Bt-Ap-Zrn
BMN82B	Bt-Mag-Ap-Zr		
BMS99A	Bt	Bt	
Garnet-columbite type pegmatite (15 samples investigated on 254 stations)			
BMN8 (coarse-grained, devoid of mica)	Grt-Ap-(P, Th, Ca, U, REE)		
BMN13 (fine to medium- grained)	Mu-Ap-Mnz-Xtm-Grt-CGM-(U, Th, Zr)	Mu-Grt-Mnz-CGM	Mu-Grt-CGM

BMN47	Mu-Grt-Zrn-Xtm-Ttn		
BMN86 (aplite)	Bt-Mu-Mag-Grt-CGM-(U, Th, Zr)	CGM	
BMN131 (hosting large apatite crystals, polished section)	Mu-Ap-Grt-Tur	Ap	Ap
BMS1 (fine to medium-grained)	Mu-Grt-Mnz-CGM-Zrn-(U, Th, Zr)	Mu-Grt-CGM	Mu-Grt-CGM
BMS5B	Mu-Grt-CGM-Zrn-Mnz	Mu-Grt-CGM-Kfs-Pl	Mu-Grt-CGM
BMS6	Mu-Ap-Grt-Tur	Mu-Ap	Mu
BMS12	Mu-Grt-Zrn-CGM	CGM	
BMS15	Bt-Mu-Grt	Grt	
BMS48	Mu-Grt-CGM	Mu-Grt-CGM	
BMS90B (mica-rich)	Mu	Mu	
BMS99A	Mu-Grt	Mu-Grt	
BMS136	Bt-Mu-Mag-Ap-Zrn-Grt-Thr	Mu-Bt	
BMS139B	Bt-Mu-Grt	Bt-Grt	Grt

Garnet-REE type pegmatite (4 samples investigated on 36 stations)

BB6A	Bt-Ilm/Pph-Ep-Eux/Aes-(REE, Y, Th)	Pl	
BB15	Bt-Grt-Ilm/Pph-Eux/Aes	Eux/Aes	
BB18	Bt-Grt-Eux/Aes-Pph	Grt	Grt
BB24C	Bt-Mu-Ep-Grt	Mu-Bt	Mu-Bt

- 1609 Abbreviations: Aes = aeschynite, Aln = allanite, Ap = apatite, Bt = biotite, Chl = chlorite,
1610 CGM = columbite-group mineral, Ep = epidote, Eux = euxenite, Grt = garnet, Hbl =
1611 hornblende, Ilm = ilmenite, Kfs = K-feldspar, Mag = magnetite, Mnz = monazite, Mu =
1612 Muscovite, Mu2 = secondary muscovite, Pl = plagioclase, Pph = pyrophanite, Qz = quartz, Sil
1613 = sillimanite, Rt = rutile, St = staurolite, Ttn = titanite, Thr = thorite, Tur = tourmaline, Zrn =
1614 zircon, Xtm = xenotime (principally after Whitney and Evans 2010).

1615 Table 2. Geochemical features of mineral phases in pegmatite, tonalitic-trondhjemitic gneiss, and micaschist enclave in garnet-columbite

1616 pegmatite.

Host rock	Tonalitic-trondhjemitic gneiss	Titanite-allanite type pegmatite	Apatite-zircon type pegmatite	Garnet-columbite type pegmatite	Micaschist in garnet-columbite type pegmatite	Garnet-REE type pegmatite
Biotite	0.48 < Mg# < 0.57 Medium K/Rb	0.46 < Mg# < 0.54 High to medium K/Rb	0.43 < Mg# < 0.49 Medium K/Rb	0.26 < Mg# < 0.42 (no data)	0.43 < Mg# < 0.48 Low K/Rb, low Nb/Ta, Ta fractionation	0.31 < Mg# < 0.32 Medium K/Rb
	(Ta, Nb)-- Li--	(Ta, Nb)-- Li-	(Ta, Nb)- Li--	(no data)	(Ta, Nb)+, Li+++ , Be	Li+
Muscovite		0.42 < Mg# < 0.55 (secondary) High K/Rb Nb-Ta depleted		0.11 < Mg# < 0.40 Low K/Rb Medium to low K/Rb Li+	0.34 < Mg# < 0.39 Low K/Rb High Nb/Ta, Ta fractionation Li++	0.32 < Mg# < 0.33 Medium K/Rb High Nb/Ta Li+
Apatite		LREE depleted Sr+, Y+	LREE depleted	High LREE, HREE depleted (zircon-monazite-xenotime fractionation) F, Mn-rich	High REE Ti-	
Garnet				LREE-, Nb, Ta, U, Zr HREE relatively depleted (zircon-monazite-xenotime fractionation)		Ca, Ti, Li, Y, HREE
Zircon		LREE-poor core (allanite fractionation) HREE, Y, Nb, U, Th	HREE, Y, Nb, U, Th		Nb, Ta, Th, Be	
Columbite-group minerals				Columbite-tantalite Mn# ~0.5, Ta# < 0.6 Enriched in Mg, Sc, Y, Yb, Th, U, depleted in Sn, W		Euxenite-aeschynite Ti, Nb > Ta, REE, Y, Th, U
Else				B (tourmaline), Th (thorite)		Low B (tourmaline)

1617

1618 The relative elements enrichment in minerals are indicated by + and – symbols.

1619 – – / -: Highly depleted / depleted. + / ++ / +++ : Enriched / highly enriched / very highly enriched

1620 **Electronic Supplementary Material 1**

1621

1622 Fig. S1. Petrogenetic relationships between rafts of amphibolites, schist, paragneiss,
1623 gneiss/granitoids of Na-rich and K-rich series and pegmatites of Mangodara district (Bonzi et
1624 al. 2021).

1625

1626 Fig. S2. Field photographs of titanite-allanite type pegmatites. (a) Titanite-allanite type
1627 pegmatite veins in continuity with segregation veins in the tonalitic-trondhjemitic gneiss. (b)
1628 Folded titanite-allanite type pegmatite with a margin in grading contact with the tonalitic-
1629 trondhjemitic gneiss. (c) Titanite-allanite type pegmatite in the tonalitic-trondhjemitic gneiss
1630 with sharp contacts.

1631

1632 Fig. S3. Field photographs of apatite-zircon type pegmatites. (a) Garnet-columbite type
1633 pegmatite cross-cutting an apatite-zircon type pegmatite in the granodioritic gneiss (station
1634 BMS99, see Bonzi et al. 2021). (b) Delimitation of an apatite-zircon type pegmatite intrusive
1635 in the granodioritic gneiss (station BMN20). (c) Small vein of apatite-zircon type pegmatite in
1636 the granodioritic gneiss, with larger sheets of biotite.

1637

1638 Fig. S4. Field photographs of garnet-columbite pegmatites. (a) Outcrop of a garnet-columbite
1639 type pegmatite. (b) Complex texture in a garnet-columbite type pegmatite: aplitic
1640 cleavelandite-rich layers alternating of coarse-grained layers. (c) Pinched garnet-columbite
1641 type pegmatite in garnet-staurolite paragneiss. (c) Pinched garnet-columbite type pegmatite in
1642 the granodioritic gneiss (see Bonzi et al. 2021).

1643

1644 Fig. S5. Field photographs of garnet-REE type pegmatites. (a) Garnet-REE type pegmatite
1645 intrusive in a GG-granitoid. (b) Garnet-REE type pegmatite showing scattered crystals of
1646 garnet. (c) Garnet-REE type pegmatites connected with leucocratic layers in a banded
1647 granodioritic gneiss. (d) Stretched dykes of garnet-REE type pegmatite in the granodioritic
1648 gneiss.

1649

1650 Fig. S6. Photomicrograph (a) and Scanning electron microscope back-scattered electron
1651 (SEM-BSE) images (b to d) of mineral assemblages from titanite-allanite type pegmatites. (a)
1652 Quartz, plagioclase and biotite hosting zircons. Secondary muscovite forms interlayers in
1653 biotite, or crystals interjoin with miarolitic-like structure (sample BMN28). (b) BSE image of
1654 altered allanite with overgrowths of apatite and zircon crystals (sample BMN43). (c)
1655 Exsolutions of titanite in a subhedral epidote (sample BMN43). (d) BSE image of sample
1656 BMN43 showing a symplectitic-like corona of epidote around ilmenite nuclei, at the contact
1657 between two large plagioclase crystals.

1658

1659 Fig. S7. (a) Photomicrograph of mineral assemblages in apatite-zircon type pegmatite (sample
1660 BMN82B), showing cumulatic texture, made of accumulation of plagioclase surrounded by
1661 biotite crystals, containing euhedral apatite and magnetite. (b) Subhedral magnetite
1662 surrounded by sheets of secondary muscovite (sample BMN82B).

1663

1664 Fig. S8. Photomicrographs (a, b) and BSE images (c to h) of mineral assemblages from
1665 garnet-columbite type pegmatites. (a) Contact (blue line) between a garnet-columbite
66

1666 pegmatite to the right and a micaschist enclave to the left. In the micaschist, subhedral
1667 muscovite contains lobate quartz and subhedral apatite, and crosscuts biotite (sample BMS6).
1668 (b) Folded to kinked muscovite in garnet-columbite pegmatite, bordered in the upper part of
1669 the image by dynamically recrystallized quartz (sample BMS6). (c) Inclusions of CGM in
1670 muscovite, and quartz-feldspar, in a garnet-rich portion of a garnet-columbite pegmatite
1671 (sample BMS1). (d) Assemblage of xenotime and zircon (sample BMN47), illustrative of the
1672 zircon-xenotime substitution (Grimes et al. 2015; Hoskin and Schaltegger 2003). (e)
1673 Columbite needle in muscovite and quartz, and Zr-bearing metamict mineral next to a
1674 muscovite sheet (sample BMS1). (f) Inclusion of metamict zircon and columbite in a garnet
1675 (sample BMN13). (g) Zircon crystal associated with metamict thorite (sample BMS136). (h)
1676 Metamict mineral, showing traces of P, Ca, U, REE, and an aureole of Fe-rich oxide (sample
1677 BMN8). In parenthesis are chemical elements detected under SEM.

1678

1679 Fig. S9. Photomicrographs (a, b) and BSE images (c to f) of mineral assemblages in garnet-
1680 REE pegmatites. (a) Inclusion in graphic plagioclase-quartz of a REE-bearing mineral (traces
1681 of Th, Ce, Gd, Si, Y, Al identified with SEM), framed by radial cracks indicating damages by
1682 radioactive irradiation (sample BB6A). (b and c) Euhedral ilmenite-pyrophanite included in
1683 graphic plagioclase-quartz (sample BB6A). (d) Altered minerals with relicts containing P, Th,
1684 Y and REE (sample BB6A). (e) and (f) Typical elongated and altered metamict minerals,
1685 bearing Ti, Ca, Nb, Y, Th, P and REE (sample BB18 for e, and sample BB15 for f). They are
1686 associated with phyllite alteration, and surrounded by an aureole of iron-oxide also filling
1687 cracks. In parenthesis are chemical elements detected under SEM.

1688

1689 Fig. S10. Backscattered electron images of accessory minerals in (a) garnet-bearing
1690 metatextitic paragneiss (sample BMN21), and (b) granodioritic gneiss (sample BMN24).

1691

1692 Fig. S11. Backscattered electronic imagery of CGM in garnet-columbite type samples
1693 BMN13 (a) and BMS1 (b). Spots indicate EPMA analysis points. Related Ta# and Nb# are
1694 outlined.

1695

1696 Fig. S12. Backscattered electronic images of zircon in a titanite-allanite pegmatite sample
1697 (BMN28). Circles indicate crater of LA-ICP-MS analyses. Concentrations in two LREE (La,
1698 Ce) and a HREE (Yb) are displayed for every spot.

1699

1700 **Electronic Supplementary Material 2**

1701 Table S1. EPMA compositions of monazite in garnet-columbite type pegmatite (sample
1702 BMN13).

1703

1704 Table S2. EPMA compositions of euxenite-aeschynite like minerals in garnet-REE type
1705 pegmatites (sample BB15).

1706

1707 Table S3. EPMA compositions of muscovite in pegmatite types and some host rocks.

1708

1709 Table S4. EPMA compositions of biotite in pegmatite types, and some host rocks.

1710

1711 Table S5. LA-ICP-MS compositions of muscovite in pegmatites type, micaschist enclave in
1712 garnet-columbite pegmatite and tonalitic-trondhjemitic gneiss.

1713

1714 Table S6. LA-ICP-MS compositions of biotite in pegmatites type, micaschist enclave in
1715 garnet-columbite pegmatite and tonalitic-trondhjemitic gneiss.

1716

1717 Table S7. EPMA compositions of apatite in pegmatites and host rocks.

1718

1719 Table S8. LA-ICP-MS compositions of apatite in pegmatites garnet-columbite type and
1720 garnet-REE type pegmatite.

1721

1722 Table S9. EPMA compositions of CGM in garnet-columbite type pegmatite.

1723

1724 Table S10. LA-ICP-MS compositions of CGM in garnet-columbite type pegmatite.

1725

1726 Table S11. Representative EPMA compositions of garnet in garnet-columbite and garnet-REE
1727 type pegmatite.

1728

1729 Table S12. Representative LA-ICP-MS compositions of garnet in garnet-columbite and
1730 garnet-REE type pegmatite.

1731

1732 Table S13. LA-ICP-MS compositions of zircon in titanite-allanite type pegmatite, apatite-
1733 zircon type pegmatite and in micaschist enclave in garnet-columbite type pegmatite.

1734

1735 Table S14. Representative EPMA compositions of zircon in titanite-allanite type pegmatite,
1736 apatite-zircon type pegmatite and in micaschist enclave in garnet-columbite type pegmatite.