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**Garnet as a monitor for melt-rock interaction: textural, mineralogical and compositional evidence of partial melting and melt-driven metasomatism**

Abridged title : Garnet as a monitor for melt-rock interaction

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28

29 **ABSTRACT**

30 In this study, we focus on a partially melted garnet-bearing granulite from the Salvador-  
31 Esplanade belt (Salvador da Bahia, Brazil), and examine the behavior of major and trace  
32 elements during partial melting and melt-driven metasomatism. Phase equilibria modeling and  
33 U-Th-Pb geochronology shows that the sample underwent partial melting during the heating  
34 segment of the decompression path from ~ 1.2 GPa and 675-700°C to ~ 0.8 GPa and 790°C at ca.  
35 2.06 Ga. During the last stage of decompression, from 0.8 to ~ 0.5 GPa, physical segregation of  
36 melt resulted in the establishment of chemical potential gradients and mass transfer between  
37 the host-granulite and the leucosome. Modelling shows that H<sub>2</sub>O, CaO, K<sub>2</sub>O and Na<sub>2</sub>O diffuse from  
38 the melt into the host residue, whereas SiO<sub>2</sub> is transferred from the host-granulite into the  
39 adjacent leucosome. Opposed senses of diffusional transfer resulted in the formation of a  
40 quartz-rich anhydrous leucosome and a quartz-undersaturated selvage in the host-granulite.  
41 Compositional maps show that garnets exhibit unusual and contrasting major and trace element  
42 distribution depending on their textural position. The largest garnets located in the quartz-  
43 undersaturated selvage preserve their original Ca and trace element growth zoning. A transition  
44 from bell-shaped profiles for Y and HREEs to bowl-shape profiles for LREEs is consistent with a  
45 typical Rayleigh fractionation model, fast intergranular element mobility and rock-wide  
46 equilibrium during prograde partial melting. In contrast, smaller garnets away from the  
47 leucosome show evidence of prograde growth zoning modified by intragranular diffusion, as  
48 evidenced by a network of open channels and healed-cracks that act as connecting pathways  
49 between the matrix and garnet core. This results in either subtle modification of both major and  
50 trace elements adjacent to the inner core inclusions, or a complete re-equilibration. The  
51 recognition that the initial Ca growth zoning, including the inner core compositions, was later  
52 modified by intragranular diffusion implies that misleading thermobarometric results and  
53 tectonic interpretations would be obtained if the garnet core composition was used to  
54 fingerprint the early garnet nucleation stage. This study further demonstrates that at high

55 temperature ( $> 750^{\circ}\text{C}$ ) and in the presence of melt, REE are not less vulnerable to diffusive  
56 resetting than divalent cations like  $\text{Ca}^{2+}$ .

57 keywords : garnet zoning, partial melting, chemical potential, diffusion metasomatism, LA-ICP-

58 MS

59

## 60 1. INTRODUCTION

61 Partial melting is a fundamental process for differentiating (Brown, 2010) and weakening  
62 the deep crust (Piazolo, Daczko, Silva & Raimondo, 2020; Rosenberg & Handy, 2005), which in  
63 turn exerts a first-order control on deformation style, lateral redistribution of mass, crustal  
64 thickness and surface topography (Beaumont, Jamieson, Nguyen & Medvedev, 2004; Hollister &  
65 Crawford, 1986; Jamieson, Unsworth, Harris, Rosenberg & Schulmann, 2011; Rabin, Trap, Carry,  
66 Fréville & Cenki-Tok, 2015; Royden et al., 1997; Vanderhaeghe, 2012). Exhumed migmatites  
67 preserve, through their mineralogy and structure, a wealth of information that can be used to  
68 decipher the thermo-mechanical behavior of orogenic continental crust. With this motivation,  
69 the petrological analysis of partially melted felsic rocks is now commonplace in metamorphic  
70 petrology, thanks to extended experimental work (e.g. Clemens, 1984; Grant, 1985; Le Breton,  
71 1988; Thompson, 1982; Vielzeuf & Holloway, 1988; Vielzeuf & Montel, 1994) and thermodynamic  
72 phase relation modeling (Holland & Powell, 2001a; Spear, Kohn & Cheney, 1999; Vielzeuf &  
73 Schmidt, 2001; White, Powell & Holland, 2001; White & Powell, 2010). Despite this progress,  
74 modeling phase relations in migmatites is not trivial because they are subject to redistribution  
75 and transfer of the highly mobile melt component at various scales, from grain- to crustal-scale  
76 (Etheridge, Daczko, Chapman & Stuart, 2020; Sawyer, 2001). Therefore, phase relation modeling  
77 in migmatites requires the consideration of not only changes in environmental variables like  
78 pressure and temperature, but also the variation of the equilibrium volume and composition of  
79 the system (i.e. open system processes).

80 Modeling the role of melt transfer (loss and gain) on phase relations has been extensively  
81 investigated (Alessio, Hand, Morrissey, Kelsey & Payne, 2017; Bartoli, 2017; Dumond, Goncalves,  
82 Williams & Jercinovic, 2015; White et al., 2001; Yakymchuk & Brown, 2014; ). Among the many  
83 benefits of open system forward modeling is the application of a melt-reintegration approach  
84 that consist of recovering the bulk composition prior to any melt transfer. This method allows  
85 investigating the prograde evolution of melt-bearing systems that tend to be obliterated by  
86 continuous re-equilibration up to peak metamorphic conditions. Another complexity arising from

87 the segregation of melt from the solid residue is that the system becomes physically and  
88 chemically differentiated into melt-rich domains (leucosome) and solid-rich domains  
89 (melanosome). The thermodynamic system is therefore composed of two sub-systems that will  
90 interact, with changing pressure and temperature conditions, due to the establishment of  
91 chemical potential gradients of components at the interface between both sub-systems (White &  
92 Powell, 2010). The diffusion of components ( $H_2O$ ,  $SiO_2$ ,...) between the two sub-systems  
93 produces significant variations in composition, mineral reactions and assemblages (Fitzsimons,  
94 1996; White & Powell, 2010). Despite this conceptual framework being generally accepted, melt-  
95 rock interaction in segregated migmatites remains rarely identified via textural, geochemical  
96 and/or mineralogical evidence, nor investigated via thermodynamic modeling (Kriegsman &  
97 Hensen, 1998; Lin & Sawyer, 2019; White & Powell, 2010).

98 In migmatites, garnet is very common as a peritectic phase and has the ability to  
99 preserve zoning of divalent cations (Fe, Mg, Ca and Mn) that reflects changing P-T conditions  
100 and/or bulk rock chemistry, either due to garnet fractionation or mass transfer during melt  
101 segregation, transport and interaction with the garnet-bearing residual rock (e.g. Caddick,  
102 Konopásek & Thompson, 2010; Konrad-Schmolke, O'Brien, de Capitani & Carswell, 2008). To  
103 retrieve the maximum information related to garnet chemical evolution, the characterization of  
104 major element zoning can be complemented with trace elements (Hyppolito et al., 2019;  
105 Raimondo et al., 2017). Trivalent cations, like rare-earth elements, have lower diffusion  
106 coefficients than major divalent cations. They are therefore more resistant to diffusional  
107 modification and allow the preservation of prograde and high temperature processes (Carlson,  
108 2012; Gaidies et al., 2020; George & Gaidies, 2017; Moore, Carlson & Hesse, 2013; Rubatto et  
109 al., 2020 ), including those involved during partial melting (Dumond et al., 2015).

110 The trace element characteristics of garnet can be linked to the evolution of accessory  
111 phases used for geochronology (e.g. monazite, zircon and rutile) to better constrain  
112 tectonometamorphic histories (Dumond et al., 2015; Hermann & Rubatto, 2003; Mahan,  
113 Goncalves, Williams & Jercinovic, 2006; Prent et al., 2019; Pyle & Spear, 2003; Tomkins &  
114 Pattison, 2007; Weller, Jackson, Miller, St-Onge & Rayner, 2020). They also provide insight into  
5

115 the mechanisms controlling the uptake (interface- vs diffusion-controlled growth) and mobility  
116 of trace element during metamorphism and metasomatism (George, Gaidies & Boucher, 2018;  
117 Konrad-Schmolke, Zack, O'Brien & Jacob, 2008; Moore et al., 2013; Raimondo et al., 2017; Skora  
118 et al., 2006). With the recent development of in-situ laser ablation - inductively coupled plasma  
119 - mass spectrometry (LA-ICP-MS), it is now possible to rapidly and quantitatively map the 2D  
120 distribution of trace elements across major and accessory minerals of interest (George et al.,  
121 2018; Raimondo et al., 2017; Ubide, McKenna, Chew & Kamber, 2015). Yet, despite the known  
122 benefits of analyzing trace elements in garnet for the purposes listed above, there remains a  
123 limited number of studies using this approach to understand the grain-scale controls on garnet  
124 crystallization (Gaidies et al. 2020; George et al., 2018; Rubatto et al. 2020).

125 In this contribution, we report the integrated petrological and geochemical  
126 characterization of a high-grade rock that preserves evidence of changes in P-T conditions,  
127 partial melting and melt-driven metasomatism. The selected sample, from the São Francisco  
128 craton in Brazil, is a layered composite rock that consists of a garnet-bearing quartz-saturated  
129 host granulite, a quartzo-feldspatic leucosome, that may correspond to crystallized segregated  
130 melt or differentiated melt residue (Guernina & Sawyer, 2003), and a quartz-undersaturated  
131 garnet-bearing selvage formed at the interface between the host-granulite and leucosome.  
132 Petrological analysis and thermodynamic modeling (using P-T- $\mu$ -X phase diagrams, where  $\mu$  =  
133 chemical potential and X = composition) shows that the distinct mineralogy of the selvage is the  
134 result of mass transfer driven by chemical potential gradients established between the host-  
135 granulite and the segregated melt during decompression. In addition, garnets show a consistent  
136 change in grain size, morphology, major and trace element composition and zoning patterns  
137 across the gneissic layering, as revealed by quantified LA-ICP-MS raster maps. The strong  
138 correlation between textural position and zoning pattern in garnet gives insight into the grain-  
139 scale petrogenetic processes involved in partially-molten granulites. We distinguish the role of  
140 (1) intergranular mobility of elements and accessory phases during metamorphic garnet growth;  
141 and (2) post-growth intragranular diffusion of major and trace elements during melt-driven  
142 metasomatism.



## 143 2. GEOLOGICAL SETTING

### 144 2.1. São Francisco craton and the Salvador-Esplanade belt

145 The São Francisco Craton, located in the east-central part of Brazil, consists of an Archean to  
146 Paleoproterozoic crystalline basement mostly overlain by Meso to Neoproterozoic sediments in  
147 its western part (Figure S1). The craton is bounded by Neoproterozoic marginal fold belts (Figure  
148 S1). The eastern part of the craton, where most of the metamorphic rocks are exposed, is  
149 subdivided into four major geological units: three Archean blocks (Gavião, Jequié and Serrinha)  
150 and two Paleoproterozoic belts: the Itabuna-Salvador-Curaçá Orogen (ISCO) and the Salvador-  
151 Esplanada belt (SEB) (Figure S1) (Barbosa & Sabaté, 2004, see references therein). The  
152 southernmost part of the Salvador-Esplanada belt, where this study has been conducted, trends  
153 broadly North-East and consists mostly of mafic to felsic granulites and migmatites.

154

155

### 156 2.2. Farol da Barra: metamorphism and age constraints

157 In the Salvador da Bahia city, high grade metamorphic and magmatic rocks of the SEB are  
158 exposed in a horst that is bound on the western side by the Recôncavo sedimentary basin and on  
159 the eastern side by the Atlantic coastal margin. The studied area is located at the Barra lighthouse  
160 (Farol da Barra; 13°00'41"S; 38°31'57"W, Figure 1a). Four distinct lithologies are identified  
161 (Souza Barbosa & Gomes, 2010; Souza-Oliveira et al., 2014) (Figure 1a): (1) orthopyroxene-  
162 bearing granulites that constitute most of the basement of the city of Salvador da Bahia and are  
163 interpreted to be derived from late Archean tonalitic magmatic rocks (2) garnet-bearing  
164 granulites and sapphirine-bearing Al-Mg granulites interpreted as meta-sediments (3)  
165 granulitized ultramafic and mafic rocks that occur as centimeter to meter-scale boudins within  
166 the paraderived granulites; and (4) monzonitic to syenitic intrusives dykes, oriented ~N10°  
167 (Figure 1a), that are either deformed or undeformed with sharp intrusive contacts.

168 Metamorphic conditions have been estimated at 0.75-0.9 GPa and 840-900°C in the  
169 sapphirine-bearing granulites (Fujimori, 1988). The age of the granulitic metamorphism has been

170 constrained by SHRIMP U-Pb dating at  $2089 \pm 11$  Ma on zircon overgrowths from the  
171 orthopyroxene-bearing granulites (Da Silva, McNaughton, 1997). On the same samples, zircon  
172 cores yielded an age of  $2561 \pm 7$  Ma interpreted as the crystallization age of the protolith  
173 magmatic rocks. The later syenitic to monzonitic magmatism has been dated by Pb-Pb  
174 evaporation on zircon at  $2064 \pm 6$  Ma (Souza et al., 2010).

175

### 176 2.3. Farol da Barra: structural relationships

177 The Barra lighthouse outcrop is characterized by a complex ductile finite strain pattern  
178 that results from the superposition of two main phases of deformation ( $D_1$  and  $D_2$ ) (Figure 1b).  
179 The overall geometry consists of meter to decameter lenses, in which  $D_1$  structures are best  
180 preserved, surrounded by  $D_2$  -East-West striking high strain zones (Figures 1b and S2). The first  
181 deformation phase ( $D_1$ ) is characterized by a gently dipping gneissic foliation ( $S_1$ ) that strikes NE-  
182 SW (Figure 1b and S2). This foliation is associated with isoclinal folds with axial planes parallel  
183 to  $S_1$  (Figure 1b). The  $S_1$  foliation bears a sub-horizontal  $L_1$  stretching and mineral lineation  
184 plunging  $0-20^\circ$ , marked by the preferred orientation of feldspars, quartz ribbons, mafic mineral  
185 and elongated ultramafic to mafic boudins. The second deformation phase ( $D_2$ ) is characterized  
186 by heterogeneous deformation that progressively steepens and transposes the previous  $D_1$  fabric  
187 along upright  $D_2$  folds and steeply dipping high strain zones ( $> 55^\circ$ ) that forms an anastomosing  
188 network of conjugated shear zones trending approximately east-west (Figures 1b and S2). An  $L_2$   
189 lineation is defined by stretched quartz, feldspars and by the preferred orientation of biotite  
190 with a moderate plunge of  $10$  to  $35^\circ$  towards the NE.

191 Kinematic analysis in the low  $D_2$  strain domains show that the two sets of ductile  
192 conjugate shear zones consist of a dominant  $108/72$ SSW dextral shear zone and minor sinistral  
193  $021/79$ ESE shear zones (Figure S2). This pattern of  $D_2$  finite strain is consistent with a NNW-SSE  
194 shortening and a bulk dextral sense of shear. The NW-SE cross-section from the Santa Maria Forte  
195 towards the Cristo hill (Figure 1b) shows that the studied area (the Barra lighthouse) corresponds  
196 to a major localized  $D_2$  high strain zone within a preserved  $D_1$  low strain domain characterized by  
197 gently dipping foliation. Whereas the  $D_1$  domains (Santa Maria fort, Hispaniola hospital, and  
8

198 Cristo Hill) consist mainly of orthopyroxene-bearing granulites, the Barra lighthouse D<sub>2</sub> high  
 199 strain zone is characterized by the abundance of garnet-bearing granulite. This apparent  
 200 relationship between deformation and lithology is beyond the scope of this contribution and will  
 201 not be  
 202 discussed further.

203

204

### 205 3. TEXTURAL AND MINERALOGICAL CHARACTERISATION

#### 206 3.1. Outcrop and sample descriptions

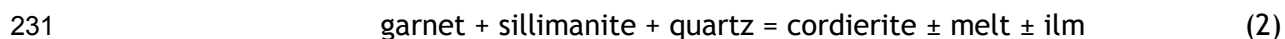
207 The studied sample is located in a S<sub>2</sub> high strain zone that surrounds a large decametric  
 208 lense of low strain rocks where an earlier fabric (S<sub>1</sub>) is preserved (Figure 1). The S<sub>2</sub> granulite-  
 209 grade gneissic foliation is steeply dipping (072/80S) and contains a gently plunging mineral  
 210 lineation (09—>075) defined by the preferred orientation of biotite and sillimanite. The gneissic  
 211 foliation is marked by a compositional banding with centimeter-wide quartzo-feldspathic  
 212 leucosomes in a garnet-bearing granulite (Figure 2a). Leucosomes are usually less than a three  
 213 centimeters in width and are systematically surrounded by a centimeter thick dark and coarse-  
 214 grained quartz-undersaturated selvage developed at the expense of the surrounding garnet-  
 215 bearing granulite (Figure 2b).

216 The studied sample 06SB06 encompasses the three lithologies described above: the host  
 217 quartz-saturated granulite, a 1 cm-wide quartz-undersaturated selvage and a thin leucosome  
 218 (Figure 2b). The peak-metamorphic assemblage of the quartz-saturated granulite consists of  
 219 garnet, sillimanite, plagioclase, biotite, quartz and K-feldspar (Figure 3a). Garnet is interpreted  
 220 as a peritectic phase produced during biotite dehydration melting via the reaction:



222 Spinel is locally observed in the matrix but it is always surrounded by a thin corona of sillimanite  
 223 and it is never in contact with quartz. The preferred orientation of large sillimanite blades (up  
 224 to 5 mm), K-feldspar and biotite, that occurs either as small flakes or large resorbed grains,

225 defines the main  $S_2$  foliation observed at the outcrop scale. Garnet contains numerous quartz  
 226 inclusions and polyphase aggregates of plagioclase + quartz with minor biotite inclusions that all  
 227 show a preferred orientation at high angle with respect to the  $S_2$  foliation (Figure 3a). This  
 228 alignment is interpreted as a relict of the former shallow dipping  $S_1$  foliation. Peak metamorphic  
 229 garnet in contact with sillimanite is partially resorbed and surrounded by a moat of retrograde  
 230 cordierite (Figure 3b), consistent with the classic reaction



232 which proceeds during decompression and/or cooling.

233 The selvage, in direct contact with the leucosome, is composed of the same peak  
 234 metamorphic assemblage, consisting of prismatic sillimanite, plagioclase, garnet, biotite and K-  
 235 feldspar, except it lacks quartz. Biotite is more abundant than in the quartz-saturated granulite.  
 236 Garnet cores contain numerous quartz inclusions, in association with plagioclase, suggesting that  
 237 the onset of partial melting and crystallization of peritectic garnet occurred under quartz-  
 238 saturated conditions. Garnet porphyroclasts are partially broken-down into cordierite and  
 239 sillimanite is strongly resorbed and replaced by cordierite + spinel symplectites (Figure 3c) via  
 240 the reaction



242 The leucosome consists mainly of quartz and thin elongated K-feldspar ribbons.

243 Plagioclase is very rare and occurs as late myrmekite intergrowths next to K-feldspar.

244 Mineral compositions are available in Dataset S1. Biotite has  $X_{Mg}$  of 0.67-0.69 and 0.71-  
 245 0.73 for the small flakes and large grains respectively, with a constant Ti content of 0.25 to 0.30  
 246 pfu. Plagioclase in the matrix has a variable range in anorthite content from 0.43 to 0.52. Spinel  
 247 is very homogeneous in composition with a  $X_{Mg}$  of 0.31. Garnet compositions are described in  
 248 Section 4.

249

### 250 3.2. Evolution of garnet morphology and inclusions

251 One of the most conspicuous observations is the evolution of garnet size, distribution of

252 inclusions and composition across the quartz-undersaturated selvage and the granulite  
253 perpendicular to the gneissic layering. Although no quantitative textural analysis has been  
254 conducted, it appears that garnet grain size decreases and the number of garnet grains increases  
255 away from the leucosome, with the biggest garnet grains observed in the quartz-undersaturated  
256 selvage (Figures 4 and 5). Furthermore, the amount, shape and composition of inclusions in  
257 garnet varies greatly and continuously across the gneissic foliation. To describe this evolution,  
258 seven garnets (A to G) were selected (see location in Figure 4).

259 In the quartz-undersaturated selvage, most garnets contain a few rounded inclusions of  
260 quartz in their core (garnet A, Figure 5a). Away from the leucosome, but still in the selvage, the  
261 largest garnet porphyroblast (b) contains in its core a small polymineralic inclusion, that consists  
262 of plagioclase and quartz, surrounded by rounded isolated quartz inclusions (Figure 5b).  
263 Inclusion-rich garnet cores are surrounded by a large rim free of inclusions except for numerous  
264 and homogeneously distributed crystallized granitic melt inclusions that are less than 20 microns  
265 in diameter (Figures 5a-b and 6). The inclusions have been identified as crystallized melt based  
266 on the main microstructural criteria summarized in Cesare, Acosta-Vigil, Bartoli & Ferrero (2015)  
267 : (1) the development of a perfect negative crystal shape (Figure 6a); (2) its polycrystalline  
268 nature that consists of biotite, plagioclase, quartz and apatite for the most common phases  
269 (nanogranitoids) (Figure 6b-c); and (3) the development of cracks (decrepitation) during  
270 decompression filled with minerals similar to those in the melt inclusion (Figure S6). The  
271 presence of these crystallized melt inclusions attest to the growth of the garnet rim in the  
272 presence of melt.

273 In the quartz-saturated granulite, the abundance of polymineralic inclusions of  
274 plagioclase and quartz observed in garnets increases away from the leucosome (Figures 4 and 5c-  
275 g). The garnet core, defined as the garnet containing plagioclase-quartz polymineralic inclusions  
276 surrounded by isolated quartz grains can be as large as 1mm and represent up to 50% of the  
277 garnet volume. Polymineralic inclusions have rounded (Figure 5c-d-f) to very irregular shapes  
278 with straight boundaries (Figure 5e). These facets are inferred to be parallel to crystallographic  
279 orientations of the garnet host and result from post-entrapment readjustment of the inclusion  
11

280 and host by dissolution-reprecipitation processes (e.g. Frezotti, 2001). Plagioclase in the  
281 polymineralic inclusions are chemically zoned with a composition varying almost continuously  
282 from pure anorthite  $An_{95}Ab_{05}$  to  $An_{58}Ab_{41}$ . Highest amounts of albite are measured in the vicinity  
283 of rounded quartz inclusions within plagioclase and the garnet host (Figure S3). All garnets have  
284 an inclusion-free rim without any melt-inclusions, as for the garnet located in the quartz-  
285 undersaturated selvage (Figure 5). This difference in the ability for garnet rims to entrap melt  
286 inclusions may be related to a slowdown of radial growth rate as garnet continuously grows.

287 X-ray computed microtomography and sequential cuts through a single garnet crystal has  
288 been performed to highlight the volume, shape and potential connectivity of the polyphased  
289 inclusions with the matrix. The selected garnet is located in the quartz-saturated granulite  
290 where the amount of polyphase quartz-plagioclase is most abundant. The X-ray tomography  
291 reveals the very irregular shape of the polyphased inclusions as well as the “spongy” texture of  
292 the garnet core (Figure 7). Although all thin-section photomicrographs of garnet presented in  
293 Figure 5 suggest that polyphase inclusions are completely enclosed in the garnet, the 2D cuts  
294 rather suggest that they are connected to the matrix through small pathways (Figure 7d-f and  
295 Movie S1).

296

### 297 3.3. Microtextures in the quartz-undersaturated selvage

298 Microtextures in the quartz-undersaturated selvage have been revealed using a cold  
299 cathode cathodoluminescence detector on an optical microscope. The matrix consists mainly of  
300 K-feldspar grains (in blue in the CL images) that form a fine-grained mosaic texture with straight  
301 boundaries and triple junctions at  $120^\circ$  (Figure 8a-b). K-feldspar is characterized by a core with  
302 fine-grained perthite textures surrounded by a K-feldspar rim (Figure 8a). These boundaries are  
303 coated with thin plagioclase films or elongated grains (brown color in Figure 8a-b) that can be  
304 connected to form a branching array. Terminations of the films and elongated grains are  
305 characterized by small dihedral angles ( $< 20^\circ$ ). The connection of these films also leads to the  
306 formation of plagioclase pools along grain boundaries (Figure 8a-b) or at K-feldspar triple

307 junctions. Where plagioclase films are abundant, K-feldspar grains show very irregular resorbed  
308 grain boundaries (Figure 8b) and plagioclase films are connected to myrmekites pools (Figure 8a-  
309 b). Finally, a large plagioclase ( $\pm$  quartz) moat surrounds sillimanite and garnet porphyroblasts  
310 (Figure 8b).

311

## 312 4. GARNET COMPOSITION

313 To investigate the link between the textural and mineralogical features of melt-rock  
314 interaction and their compositional characteristics, major element data from multiple garnet  
315 grains were obtained by EMPA, and trace element data by LA-ICP-MS. X-ray maps and linear spot  
316 traverses were acquired to reveal the evolution of major element zoning patterns across the  
317 quartz-undersaturated selvage and granulite. Trace element mapping was performed to better  
318 inform the paragenetic context, particularly with regard to accessory minerals, and evaluate the  
319 record of geochemical mobility during garnet growth and dissolution. Major and trace element  
320 data are provided in Datasets S1 and S2, respectively, and presented in Figures 9, 10, 11, 12 and  
321 13. . Raw trace element maps from which the profiles were extracted are available as “.csv”  
322 text file images (Dataset S3-S5). LA-ICP-MS analytical methods are provided in Appendix S1.

### 323 4.1. Major element zoning

324 Garnets are almandine - pyrope - grossular solid solutions with minor amounts of  
325 spessartine ( $< 0.015$ ) (Dataset S1). All garnets are zoned and their composition varies with  
326 respect to their location and distance from the leucosome. In the quartz-undersaturated  
327 selvage, a polymineralic inclusion-free garnet porphyroblast in direct contact with the  
328 leucosome (Garnet A; Figure 5a) is characterized by bell-shaped zoning in grossular, with a  
329 maximum content of 0.12 in the core that decreases continuously towards the rims down to 0.03  
330 (Figure 9a). Almandine and pyrope content increase towards the rim from 0.56 to 0.61 and 0.30  
331 to 0.35, respectively, which results in a constant  $X_{Mg} [=Mg/(Fe+Mg)]$  of 0.34-0.35. Garnet B has  
332 equivalent zoning, except that close to the polymineralic inclusions, grossular content decreases

333 from a 700  $\mu\text{m}$  wide and uniform plateau at 0.12 down to 0.08 (Figure 9b). This decrease in  
334 grossular content is mimicked by an increase of XMg from 0.35 to 0.38. It is noteworthy that the  
335 depleted grossular core also corresponds to the domain containing smoothed and rounded  
336 inclusions of quartz (light gray area in Figure 9b). The outer rim of Garnet B is characterized by  
337 grossular content and XMg value of 0.03 and  $\sim$ 0.35-0.36, respectively (Figure 9b). The resorbed  
338 outer left rim, in contact with cordierite and biotite, shows a sharp decrease in XMg down to  
339 0.28 over a distance not greater than 100  $\mu\text{m}$  (Figure 5b and 9b).

340 Garnets from the quartz-saturated granulite (Garnets C-G) show zoning profiles  
341 comparable in shape to Garnet B from the quartz-undersaturated selvage, except that the  
342 endmember modal abundances are significantly different (compare Figure 9b and Figure 9c-g).  
343 Garnet cores in direct contact with the polymineralic inclusions are characterized by low  
344 grossular values (0.03-0.04) that are similar in composition to garnet rims in contact with the  
345 matrix. Grossular content then increases towards the rim to a maximum value that is  
346 systematically located at the transition between the inclusion-bearing and inclusion-free garnet  
347 (Figure 9c-g). The maximum grossular content varies from grain to grain, from 0.095 (Garnet E)  
348 down to 0.045 (Garnet F). Finally, the inclusion-free rims show a decrease in grossular content  
349 towards the matrix to a minimum value of 0.03, equivalent to the core compositions. XMg ratio  
350 profiles for all samples are relatively flat, with values evolving from 0.36-0.37 (Garnet C) to  
351 0.41-0.44 (Garnet G) with increasing distance from the leucosome. The smallest garnet grain  
352 analyzed (Garnet F) does not follow this trend, showing a lower XMg value of 0.36 and a near-  
353 constant grossular content of 0.03-0.04. Garnet E profile shows on its left side a sharp decrease  
354 in XMg from 0.40 to 0.36, over a distance less than 100  $\mu\text{m}$ , that is interpreted as late Fe-Mg  
355 exchange during partial garnet breakdown into cordierite, comparable to Garnet B.

356

357

#### 358 4.2. Trace element zoning

359 Similar to their major element characteristics, the analyzed garnets show significant



360 variations in trace element composition and zoning patterns depending on their location with  
361 respect to the leucosome. We will focus on Garnets A and B, located in the quartz-  
362 undersaturated selvage, and Garnet G, located in the quartz-present host granulite.

363 Garnet A is located directly adjacent to the leucosome, and contains the least inclusions  
364 in its core except for a few rounded quartz grains. Figures 10a, 11 and Figure S4 illustrate the  
365 key features of the trace element zoning. Y, Zr, V and heavy rare-earth elements (HREEs), from  
366 Dy to Lu, show a symmetrical bell-shaped profile without sharp discontinuities except for an  
367 abrupt increase at the rim. In contrast, light rare-earth elements (LREEs), from La to Gd, show a  
368 bowl-shaped zoning profile with a broad flat core that is mimicked by Li, P and Sc. Notably, the  
369 radial positions of broad peaks along the zoning profile are correlated for all REEs and Y (Figure  
370 11), and there is no difference in gradient with decreasing atomic number.

371 Garnet B is the largest analyzed grain, with a small polyphase plagioclase + quartz  
372 inclusion in its core. It is located a few millimeters from Garnet A within the quartz-  
373 undersaturated selvage but further displaced from the leucosome. Trace element map and  
374 profile are shown in Figure 10b, 12 and S4. Its zoning pattern shows the same symmetrical bell-  
375 shape profile for Y, Zr and the heaviest HREE (from Yb to Er), but with elevated concentrations  
376 compared to Garnet A (e.g. Y ~360 vs 270 ppm; Yb ~75 vs 45 and Lu ~15 vs 6 ppm). In contrast,  
377 rim compositions are strictly identical for all elements in both samples (compare Figures 11 and  
378 12 and S4). Because Garnets A and B are located in the same petrographic setting, we suggest  
379 that the difference in absolute concentrations reflects a sectioning effect (i.e. Garnet A is not  
380 sectioned through its core) rather than a geological process. The outer rim of Garnet B is  
381 characterized by a sharp increase in Y, Zr and HREEs, and LREEs and P have the same overall  
382 bowl-shaped zoning as Garnet A despite a more irregular profile that includes several narrow  
383 peaks. Concentrations in La, Ce and P are significantly enriched next to the polyphase inclusion  
384 and the matrix on the left side of the profile (Figure 12). The major difference in comparison to  
385 Garnet A corresponds to a broad depletion in Dy, Ho and Tb adjacent to the polyphase inclusion  
386 located at the garnet core. This feature closely matches the grossular zoning profile (Figure 9b),

387 and corresponds to the domain containing rounded inclusions of quartz.

388 Garnet G is derived from the quartz-present host granulite and is located furthest from  
389 the leucosome (Figures 4 and 5). It is the smallest analyzed grain and contains a large irregular  
390 polyphase inclusion in its core. Trace element map and profile are shown in Figure 10c, 13 and  
391 S4. Y and HREEs, from Lu to Tb, show a broad bell-shaped profile, but with maximum  
392 concentrations on average two to three times lower than the larger Garnets A and B (Figure 13).  
393 In contrast, rim compositions for all elements are once again similar for all samples. La, Ce and  
394 P show bowl-shaped profiles with a large flat core surrounded by an enriched rim. Sm (and to a  
395 lesser extent Gd) are no longer characterized by a typical bowl-shaped profile but rather mimic  
396 the grossular zoning, with a depleted core next to the polyphase inclusion (Figure 13). This  
397 similarity continues along the profile, with an identical location for the maximum Sm  
398 concentration and a core composition adjacent to the polyphase inclusion that is equivalent to  
399 the rim composition (2 to 3 ppm on average; Figure 13).

400

## 401 5. PETROLOGICAL MODELING

402 We use computed phase diagram sections to constrain (1) the pressure and temperature  
403 of garnet nucleation and growth, partial melting, and retrogression; and (2) the most likely  
404 processes responsible for the formation of the quartz-undersaturated selvage. For this purpose,  
405 the model system  $\text{Na}_2\text{O}-\text{CaO}-\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}$  was chosen to calculate phase  
406 relationships using Perple\_X 6.6.6 software (Connolly, 2005), with the thermodynamic database  
407 of (Holland & Powell (1998), as revised in 2002).  $\text{TiO}_2$  and  $\text{Fe}_2\text{O}_3$  were not considered in the  
408 modeling because (1) the bulk rock content of  $\text{TiO}_2$  is less than 0.6 wt%; and (2) there is no  $\text{Fe}^{3+}$ -  
409 bearing phase like hematite or magnetite and the amount of  $\text{Fe}^{3+}$  in garnet based on structural  
410 formulae is negligible. Solution models and end-member phases considered in the modeling are  
411 listed in Table S1.

412

## 413 5.1. P-T conditions of garnet nucleation and growth

414 Deciphering the P-T conditions of the early garnet growth and prograde path relies  
415 essentially on the observation of garnet inclusions and the valid interpretation of garnet  
416 composition and zoning. Garnet cores are characterized by the occurrence of quartz and  
417 polyphase plagioclase-quartz inclusions, whereas the inner rim of the largest garnet grains show  
418 abundant crystallized melt inclusions. We therefore suggest that the onset of garnet growth may  
419 have occurred under sub-solidus conditions.

420 Garnet core compositions are typically used to retrieve the approximate P-T conditions of  
421 garnet nucleation. However we will show later that major and trace element zoning patterns  
422 and more particularly garnet core compositions have been modified, most likely as a result of  
423 post-growth re-equilibration during partial melting and retrogression. The biggest garnets located  
424 in the quartz under-saturated selvage (garnet A and B, Figure 4-5) show the least modified  
425 zoning profile, with typical bell-shape zoning in grossular (Figure 9a,b) and trace elements  
426 (Figure 11-12). Although these large garnets A and B are located in the quartz-undersaturated  
427 selvage, the presence of quartz inclusions within their cores suggests that the onset of garnet  
428 growth occurred under quartz-saturated conditions. We therefore suggest that the loss of  $\text{SiO}_2$   
429 and the formation of the quartz-undersaturated layer is a post-garnet growth process. Hence, to  
430 retrieve the P-T conditions of early garnet growth and constrain the prograde path, we have  
431 computed a P-T phase diagram section with a XRF bulk composition corresponding to the quartz-  
432 saturated domain composition (Table 1 and Figure 14). The phase diagram section has been  
433 contoured for grossular content,  $X_{\text{Mg}} (= \text{MgO}/\text{MgO}+\text{FeO})$  in garnet and garnet modes (Figure 14).

434 The amount of water is known to be a critical variable when modeling phase relations in  
435 melt-bearing systems (e.g. White et al., 2001). A classic strategy to estimate the amount of  
436 water in melt-bearing systems consists of adding enough water to the system such that the  
437 solidus is water-saturated. However, if we follow this approach, high-grossular garnet core  
438 compositions cannot be reproduced. Therefore, the system was kept water-undersaturated, such  
439 that measured garnet core compositions (i.e. the highest grossular values) are predictable, and

440 the amount of major rock-forming phases are consistent with observed mineral modes. The  
441 amount of water was adjusted and fixed at 1.67 mol% (0.025 mol or 0.46 wt%). With this low  
442 water condition, partial melting was restricted to temperatures higher than  $\sim 750^{\circ}\text{C}$  (Figure 14)

443 The maximum grossular content of  $\sim 0.12$ , coupled with the corresponding XMg (0.35-  
444 0.38), allow estimating the minimum P-T conditions of garnet growth at  $\sim 1.2$  GPa,  $675^{\circ}\text{C}$  (black  
445 circles with letters in Figure 14). This corresponds to sub-solidus conditions where garnet is in  
446 equilibrium with plagioclase, biotite, kyanite and muscovite. Although these last two phases  
447 have never been observed as inclusions in garnet, the predicted plagioclase composition ( $\text{An}_{76}$ -  
448  $\text{Ab}_{23}$ ) is consistent with the high anorthite compositions measured in plagioclase inclusions.  
449 Furthermore, melt-inclusions are absent in the high grossular cores (Figure 5), whereas they are  
450 abundant in the low-grossular rims. This observation is consistent with an early garnet  
451 nucleation and growth under sub-solidus conditions with peak metamorphism and garnet rim  
452 crystallization occurring in melt-present conditions.

453 In Figure 14, we report the composition of Garnet C to G, located in the quartz-saturated  
454 domain, with the highest grossular content and corresponding XMg values (grey circles with  
455 letters in Figure 14). The maximum grossular content, preserved in the inner rim at the  
456 transition between the inclusion-bearing and inclusion-free domains, varies from grain to grain  
457 from 0.095 (Garnet E) down to 0.04 (Garnet F) and XMg from 0.36 (Garnet C) to 0.43 (Garnet G)  
458 (Figure 9). The inferred P-T conditions vary from  $\sim 1.3$  GPa and  $720^{\circ}\text{C}$  to 0.85 GPa and  $780^{\circ}\text{C}$   
459 corresponding approximately to the P-T conditions of garnet nucleation and peak  
460 metamorphism / partial melting, respectively (see Section 5.2 below). We suggest that this  
461 could reflect the progressive and incomplete re-equilibration of garnet during a prograde  
462 (heating/decompression) path from  $\sim 1.2$  GPa and  $675^{\circ}\text{C}$  down to 0.8 GPa and  $790^{\circ}\text{C}$ . During this  
463 P-T evolution minor garnet growth is expected (volume % = 14-16, Figure 14).

464

## 465 5.2. P-T conditions of peak metamorphism, partial melting and retrogression

466 The computed P-T phase diagram in Figure 14 shows the classic phase relations for  
 467 partially melted aluminous-rich pelites (e.g. White et al., 2001), with the divariant muscovite-  
 468 dehydration melting equilibrium at high pressure ( $1.2 > P > 0.9$  GPa) and the biotite-dehydration  
 469 melting equilibrium that produces garnet and cordierite at lower pressure ( $0.8 > P > 0.6$  GPa).  
 470 Between these two divariant narrow fields lies a trivariant assemblage with garnet, biotite,  
 471 sillimanite, K-feldspar, quartz and melt in equilibrium, between 770 and 820 °C (Figure 14). This  
 472 calculated assemblage is consistent with the observed assemblage in the quartz-saturated  
 473 granulite. Garnet core and rim compositions in direct contact with the polyphased plagioclase-  
 474 quartz inclusions (ca. 3% grossular and  $X_{Mg} = 0.36-0.44$ ) give consistent P-T conditions of  $0.8 \pm$   
 475  $0.05$  GPa and  $790 \pm 20$  °C (white circles with letters in Figure 14), with the predicted  $X_{Mg}$  of  
 476 biotite (0.70) consistent with the measured biotite composition. We interpret these estimates to  
 477 reflect partial melting during the development of the steeply dipping  $S_2$  granulite-grade gneissic  
 478 foliation. As discussed in Section 7.2, however, the garnet trace element distribution indicates  
 479 that core compositions may have been affected by post-growth re-equilibration.

480 Retrogression in the quartz-saturated domain is characterized by the local development  
 481 of cordierite at the expense of garnet and sillimanite (Figure 4b), through reaction (1). This local  
 482 and partial retrogression is also evidenced by the sharp decrease in  $X_{Mg}$  from 0.40 to 0.36, at  
 483 constant grossular content ( $\sim 0.30$ ) (Figure 9e, outer rim of Garnet E). Figure 14 shows that the  
 484 development of cordierite and the decrease in  $X_{Mg}$ , at almost constant grossular content, would  
 485 require further isothermal decompression at or below 0.7 GPa in the garnet + cordierite +  
 486 sillimanite + K-feldspar + plagioclase + quartz stability field.

487

## 488 5.3 Formation of the quartz-undersaturated selvage via melt-driven metasomatism

489 In migmatites, the formation of quartz-undersaturated assemblages has been interpreted  
 490 as characteristics of advance quartz-consuming reactions during partial melting coupled with  
 491 local melt extraction (Fitzsimons, 1996; Owen, 1991). Alternative models consist of producing  
 19

492 the quartz-depleted selvage by small-scale silica metasomatism between the host rock and a  
493 quartz vein (Penniston-Dorland & Ferry, 2008) or a silica-rich leucosome formed after melt  
494 segregation and transport (Dunkley, Clarke & Harley, 1999). In the later case, selvage formation  
495 is a post-anatectic process rather than being directly related to partial melting of the host-rock  
496 (Lin & Sawyer, 2019). In the following section, we model and show that element exchange  
497 (diffusion-metasomatism) between the segregated melt (leucosome) and the host granulite is  
498 likely to be responsible for the formation of the quartz-depleted selvage during post-peak  
499 decompression.

500         Following the approach of White and Powell (2010), we investigate through phase  
501 equilibria modeling the melt-driven metasomatism and more particularly the role of diffusion,  
502 between melt-rich domains (leucosome) and the host granulite, on the formation of the selvage.  
503 Our conceptual model is similar to Lin & Sawyer (2019) and requires a first stage of partial  
504 melting and melt segregation to produce two subsystems corresponding to the quartz-saturated  
505 domain (~residue) and the melt domain. At the P-T conditions of partial melting and melt  
506 segregation (i.e. 0.8 GPa and 790°C), chemical potentials are the same in the solid phases and  
507 the melt of both domains, if thermodynamic equilibrium is maintained. The second stage  
508 corresponds to decompression into the cordierite-bearing stability field below 0.7 GPa. As  
509 pressure decreases, chemical potential evolve independently in both subsystems. Therefore,  
510 diffusion of components is driven by the chemical potential gradient established at the interface  
511 between the quartz-saturated residue and the leucosome.

512         In order to determine the direction of diffusion, we have computed the chemical  
513 potentials of H<sub>2</sub>O, SiO<sub>2</sub>, CaO and K<sub>2</sub>O for the quartz-saturated and melt domains along an  
514 isothermal decompression path from 0.8 to 0.5 GPa at 790°C (Figure 15). The chemical  
515 potentials in the melt domain were determined using the chemical composition of melt  
516 produced at 0.8 GPa and 790°C (Table 1). Chemical potential of the selected components were  
517 computed separately for each sub-systems (segregated melt and residue) by assuming that  
518 during decompression these two sub-systems are closed and chemically isolated from each other  
519 (Figure 15). This implies that there is no interaction or mass transfer between the two  
20

520 subsystems. This ideal chemical isolation model is in a strict sense impossible because the two  
 521 sub-systems are physically in contact and therefore chemical interactions are expected at some  
 522 scale and chemical potential will continuously evolve with ongoing decompression due to  
 523 diffusive mass transfer (White & Powell, 2010). However, our approach can be used at first order  
 524 to determine the direction of transport of the four selected species if chemical interaction at  
 525 the interface is permitted. During decompression, the chemical potentials of H<sub>2</sub>O, CaO, K<sub>2</sub>O and  
 526 Na<sub>2</sub>O (Na<sub>2</sub>O not shown in Figure 15) become higher in the melt segregated domain (leucosome)  
 527 than in the residue. In contrast, the chemical potential of SiO<sub>2</sub> is lower in the melt than in the  
 528 residue. This suggests that species such as H<sub>2</sub>O, CaO, K<sub>2</sub>O and Na<sub>2</sub>O will diffuse from the melt  
 529 into the host residue whereas SiO<sub>2</sub> is transferred from the host residue into the adjacent melt  
 530 domain.

531 To model the effect of metasomatism and pressure on phase relations, a P-X phase  
 532 diagram section was computed at 790°C (Figure 16). The initial amounts of CaO and Na<sub>2</sub>O are  
 533 very low (< 1.3 and 0.6 wt%, respectively) and do not significantly influence the phase relations.  
 534 Therefore, variations in Na<sub>2</sub>O and CaO content are not considered. The bulk composition for X =  
 535 0 corresponds the quartz-saturated host-rock composition prior to any mass transfer. The bulk  
 536 composition for X=1 is obtained assuming a gain in H<sub>2</sub>O of 200% (H<sub>2</sub>O = 0.075 mol) and a SiO<sub>2</sub> loss  
 537 of 55% (SiO<sub>2</sub> = 0.342 mol). These values are arbitrary. Bulk compositions used are provided in  
 538 Table 1.

539 Figure 16 shows the location of the two main low variance (V=2) retrogression equilibria  
 540 that have been observed in the quartz-saturated domain and quartz-absent selvage: Grt + Sil +  
 541 Qz ± Bt = Crd + L (2) and Grt + Sil ± Bt = Crd + Spl + L (3) at 0.72 GPa and 0.55 GPa respectively.  
 542 These divariant equilibria together with their related trivariant assemblage are used to locate  
 543 the quartz-out, cordierite-in and spinel-in boundaries (Figure 16). The formation of selvage  
 544 assemblage (Crd + Spl + Sil + L ± Grt/Bt) requires a decompression below 0.55 GPa and a bulk  
 545 composition with X ranging from ~0.65 to 0.8. Taking a value of X = 0.7, mass transfer required  
 546 to produce the selvage from the host quartz-saturated granulite corresponds to a relative gain of

547 about 140 % of H<sub>2</sub>O coupled with a relative loss of 47% of SiO<sub>2</sub>. The P-X phase diagram is also  
548 contoured for garnet mode (Figure 16). As expected, decompression is associated with garnet  
549 breakdown, but at pressures above the cordierite-in boundary ( $P > 0.72$  GPa), mass transfer (SiO<sub>2</sub>  
550 loss coupled with H<sub>2</sub>O gain) induces a slight increase in garnet mode, from 12 to 17 vol%. This  
551 mineralogical evolution is consistent with the observation of a higher modal abundance of garnet  
552 in the metasomatic selvage with respect to the host granulite.

553         Based on the estimation of mass transfer required to produce the quartz-undersaturated  
554 selvage, a P-T phase diagram section has been computed with an estimated selvage composition  
555 (Figure 17 and table 1). This composition corresponds to  $X = 0.70$  in Figure 16. Figure 17 shows  
556 the location in a P-T space of the two main low-variance retrogression reactions (2) and (3) and  
557 the position of the quartz-out, cordierite-in and spinel-in lines. Once again, the selvage  
558 formation is consistent with a decompression from about 0.8 GPa down to 0.5 GPa at  
559 temperatures greater than 750 °C.

560

561

## 562 **6. U-TH-PB MONAZITE AND ZIRCON GEOCHRONOLOGY**

563 Monazite and zircon geochronology were acquired to establish a temporal framework for melt-  
564 rock interactions and the relationship of garnet growth and dissolution to the accessory mineral  
565 evolution. Chemical zoning in monazite and zircon were first characterised via X-ray mapping  
566 and cathodoluminescence respectively. In situ U-Th-Pb monazite and zircon ages were then  
567 acquired by laser ablation inductively coupled plasma-mass spectrometry (LA-ICP-MS). Sample  
568 preparation and operational procedures for all techniques are described in Hurai, Paquette,  
569 Huraiová & Kronečný (2010) and Paquette et al. (2014) and presented in Appendix S2. U-Th-Pb  
570 geochronological data are provided in Dataset S6.

571



## 572 6.1. Textural features and dates

573 Monazite is abundant in both quartz saturated and undersaturated domains but is totally  
574 absent in the leucosome. The analyzed monazites range in size from 50 to 250  $\mu\text{m}$  and occur in  
575 the matrix, at grain boundaries and included in sillimanite and plagioclase (M1-2-13), in the  
576 vicinity of garnet and within the retrograde cordierite moat (M3-6-4-8-9-10-11-12) (Figure 18a).  
577 Monazite grains M3, M6 and M12, located in cordierite moats, have an unusual worm-like shape  
578 that mimics the embayed garnet rim (Figure 18c). Only one grain (M7) was found included in a  
579 garnet rim.

580 All monazite grains are chemically homogeneous (Y-poor) (Figure 18b,c) except for two  
581 grains (M9 and M10) that show a subtle zoning with high Th domains. Grains located in the  
582 cordierite moat show discrete and discontinuous micron-scale Y-rich rims (Figure 18c) whereas  
583 those included in other phases are devoid of such overgrowths (Figure 18b). These observations  
584 suggest that the crystallization of the low Y grains might precede or be coeval with the peak  
585 assemblage garnet (rim), sillimanite, biotite, plagioclase, K-feldspar  $\pm$  quartz. The crystallization  
586 of discrete Y-rich monazite overgrowths in the vicinity of garnet porphyroblasts partially  
587 retrogressed into cordierite has been previously interpreted as an indicator of accessory phase  
588 growth during garnet breakdown (Mahan et al., 2006). We suggest that Y overgrowths are also  
589 contemporaneous with cordierite crystallization and selvage formation.

590 Ten homogeneous Y-poor grains were analyzed for a total of 39 analyses (Dataset S6). The  
591 Y-rich overgrowths were not analyzed due to their small size in order to avoid mixing during  
592 ablation. In a  $^{206}\text{Pb}/^{238}\text{U}$  versus  $^{208}\text{Pb}/^{232}\text{Th}$  diagram all analyses are concordant or sub-concordant  
593 (Figure 19a). Excluding six analyses (see Dataset S6) gives a concordia age of  $2061 \pm 6$  Ma  
594 ( $\text{MSWD}(\text{C}+\text{E}) = 1.3$ ). The weighted mean of  $^{208}\text{Pb}/^{232}\text{Th}$  dates gives an age of  $2069 \pm 8$  Ma ( $\text{MSWD} =$   
595 1.6) (Figure 19b). The  $^{208}\text{Pb}/^{232}\text{Th}$  dates show no systematic variation with textural location (see  
596 Figure 18a and Dataset S6), and more specifically, monazite included in garnet (M7) is not  
597 statistically different from matrix grains (Figure 19). The date of 2.06 Ga is interpreted as the  
598 age of partial melting at  $0.8 \pm 0.05$  GPa and  $790 \pm 20$   $^{\circ}\text{C}$ , before decompression responsible for

599 partial garnet breakdown and crystallization of small Y-rich overgrowths.

600           Zircons are rare and do not exceed 50  $\mu\text{m}$  in diameter. They are translucent with a  
601 rounded shape typical of zircons recrystallized under granulite facies conditions. Fourteen  
602 analyses were performed on fourteen grains (Dataset S6). They do not show significant zoning in  
603 cathodoluminescence. A first group of 7 analyses are concordant and gives a concordia age of  
604  $2072 \pm 7$  Ma (MSWD(C+E) = 2.3) (Figure 19c). A second group of 4 analysis are discordant and  
605 when plotted together with the previous 7 analysis, it defines a discordia with an upper  
606 intercept at  $2073 \pm 11$  Ma (MSWD = 2.0). The last group, consists of 4 analyses that do not match  
607 the previous tow identified clusters (white ellipses in Figure 19), with  $^{207}\text{Pb}/^{206}\text{Pb}$  dates that  
608 range from 2.07 to 2.33 Ga and one concordant analysis at  $2192 \pm 15$  Ma. The 2.07 Ga date is  
609 similar to that obtained for monazite and is interpreted as the age of partial melting. The  
610 remaining three analyses are interpreted as inherited ages.

611

## 612 7. DISCUSSION

613           Petrological and micro-textural analysis carried out across the interface between the  
614 leucosome and the host-granulite shows a systematic and consistent evolution in mineral  
615 paragenesis, micro-textures, garnet grain size and composition (major and trace element) with  
616 the distance from the leucosome. The studied sample is therefore used to discuss small-scale  
617 melt-rock interaction processes. In this section, we will first summarize the petrological model  
618 of formation of the migmatite and more particularly the quartz-undersaturated selvage at the  
619 interface between the segregated melt (leucosome) and the partially melted host granulite.  
620 Based on this petrological evolution, we will discuss the significance of major and trace element  
621 zonation observed in garnet.

622

623 **7.1. Petrological model of migmatite and quartz-undersaturated selvage formation through**  
624 **diffusion metasomatism**

625 The phase relation modeling performed suggests that the onset of garnet crystallization  
626 occurred under subsolidus conditions at high pressure ( $P \sim 1.2$  GPa) and temperatures of 675 to  
627  $700^\circ\text{C}$ . These P-T conditions, that were estimated using the maximum grossular content of  
628 garnet cores and anorthite content of plagioclase inclusions, are consistent with a steady state  
629 geotherm considering average crustal thermal parameters (thermal conductivity  $k = 3 \text{ W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$ ,  
630 heat production  $A = 0.75 \mu\text{W}\cdot\text{m}^{-3}$ , a mantle heat flow  $Q = 30 \text{ mW}^{-2}$ , a density  $\rho = 2.7 \text{ kg}\cdot\text{m}^{-3}$ ) at a  
631 depth of 45 km (1.2 GPa). We suggest that the gently dipping gneissic foliation ( $S_1$ ) that trends  
632 NE-SW (Figures 2b and S1) may be occurred at to these conditions and may correspond to a  
633 crustal thickening event. This interpretation is also supported by the observation of quartz and  
634 plagioclase inclusions trails, defining a  $S_1$  foliation, in the high-grossular garnet cores (Figure  
635 3a).

636 Partial melting occurred at 0.8 GPa and  $790^\circ\text{C}$  at  $\sim 2.06\text{-}2.07$  Ga. The former presence of  
637 melt at peak metamorphic conditions can reliably be inferred from the occurrence of peritectic  
638 garnet, produced by biotite-dehydration melting (1) (Figures 4-5) and the presence of  
639 crystallized melt inclusions within the garnet (Figure 6). Interstitial elongated films along grains  
640 boundaries and pools at the triple junction (Figure 8) are alternative criteria diagnostic of the  
641 former presence of melt and melt movement at the grain scale (Holness & Sawyer, 2008; Sawyer,  
642 2001; Stuart, Piazzolo & Daczko, 2018; Vernon, 2011). The interstitial plagioclase may represent  
643 the plagioclase component crystallized from the melt (Hasalová, Schulmann, Lexa, Štípská &  
644 Hroda, 2008; Stuart et al., 2018). The corroded K-feldspar grains and the crystallization of  
645 plagioclase and myrmekite (figure 8) are interpreted as the result of K-feldspar dissolution  
646 during the interaction between the host rock and a migrating melt (Hasalová et al., 2008). We  
647 suggest that the connection between the leucosome and the host-granulite via the grain  
648 boundary melt array favored efficient small-scale mass transfer in and out of the host-granulite  
649 to produce the quartz-undersaturated selvage.

650 Thermodynamic modeling shows that decompression from 0.8 to 0.5 GPa at 790°C  
651 coupled with melt segregation are responsible for the establishment of chemical potential  
652 gradients between the leucosome and the host-granulite. Chemical potential gradients drive  
653 diffusive metasomatism of species between the two reservoirs and the formation of the quartz-  
654 undersaturated selvage at the interface. The modeling suggests that, during post-peak  
655 decompression from 0.8 to ~0.5 GPa at 790°C, species like H<sub>2</sub>O, CaO and Na<sub>2</sub>O will be transferred  
656 from the leucosome to the host-rock while SiO<sub>2</sub> had an opposite diffusion direction, going from  
657 the host-rock to the leucosome. Phase diagrams in figure 16 and 17 show that mass transfer  
658 coupled with decompression resulted in the partial transformation of the quartz-saturated peak  
659 metamorphic assemblages (Grt, Bt, Sil, Qz + L) into a quartz-undersaturated assemblage with  
660 cordierite and spinel (see location of Qz-out, Crd-in and Spl in Figure 16). Formation of the  
661 quartz-undersaturated selvage via melt-driven metasomatism (ie. SiO<sub>2</sub> loss), is well supported by  
662 the absence of quartz in the matrix and the observation of quartz only as inclusions in garnet  
663 cores. This observation rules out the possibility that the gneissic layering and more particularly  
664 the quartz-absent selvage is inherited from the sedimentary protolith. The phase diagram shows  
665 that further mass transfer and decompression below 0.5 GPa could result in the formation of  
666 corundum-bearing assemblage.

667 Several other petrological and micro-textural observations provide support for the  
668 modeled mass transfer direction and the formation of the selvage:

669 (1) The quartz-undersaturated selvage, where local mass-transfer is inferred as the  
670 dominant process, is characterized by larger garnet grain sizes. This is consistent with modeled  
671 garnet modes in Figure 16, that show an isobaric and isothermal increase in garnet abundance at  
672 0.8 GPa, only due to mass transfer (SiO<sub>2</sub> gain and H<sub>2</sub>O loss).

673 (2) Diffusion of water from the melt into the host-granulite will favor the crystallization  
674 of anhydrous products in the melt, producing almost an anhydrous leucosome, while the  
675 adjacent host granulite is hydrated, as demonstrated by the slight enrichment of the selvage in  
676 hydrous-phases like biotite and cordierite. This observation is very consistent with results from

677 White and Powell (2010) who also predicted that the diffusion of H<sub>2</sub>O into the residue will be  
678 responsible for the formation of anhydrous leucosomes during an isobaric cooling under granulite  
679 facies conditions.

680 (3) Diffusion of other mobile species like CaO, Na<sub>2</sub>O and K<sub>2</sub>O, from the leucosome into the  
681 selvage is indirectly supported by the mineralogical composition of the leucosome that consists  
682 mostly of quartz with very minor K-feldspar and almost no plagioclase. It is well known that  
683 leucosomes in migmatites do not have the expected minimum granitic composition (Ashworth &  
684 McLellan, 1985) due to post-anatectic magmatic differentiation processes. However, following  
685 Olsen and Grant (1991), who performed a mass-balance (isocon) analysis on leucosomes, we  
686 suggest that the quartz-rich and K-feldspar and plagioclase-poor leucosomes most likely formed  
687 by loss of Na<sub>2</sub>O and CaO by diffusion.

688 (4) Diffusion of K<sub>2</sub>O into the host-rock is likely to be responsible for the formation of the  
689 distinct inclusion-free outer-rim of K-feldspar located in the selvage (Figure 8).

690 (5) The introduction of a Na<sub>2</sub>O/CaO-rich fluid or melt into the selvage, is consistent with  
691 the observation of plagioclase films and pools along K-feldspar grain boundaries (Figure 8). The  
692 formation of myrmekite in the selvage is also consistent with a fluid or melt mediated Na-Ca  
693 replacement of K-feldspar by plagioclase (Yuguchi & Nishiyama, 2008).

694 Melt segregation and the development of the selvage defines a steeply deeping S<sub>2</sub> gneissic  
695 layering that transposes the previous S<sub>1</sub> fabric (Figure 3 and Figure S1). Kinematic analysis of the  
696 D<sub>2</sub> high strain is consistent with NNW-SSE shortening and a bulk dextral sense of shear. We  
697 conclude that melt transfer and metasomatism may have been enhanced by D<sub>2</sub> transpressive  
698 deformation.

699

## 700 **7.2. Major and trace element behavior during partial melting and retrograde melt-rock** 701 **interaction**

702 The studied sample preserves mineralogical and textural evidence of prograde partial  
703 melting and melt segregation, near-isothermal decompression and melt-driven metasomatism.

704 With respect to other phases, garnet has the tremendous advantage to preserve records of these  
27

705 various processes via its chemical zoning in major and trace elements, because of the slow rates  
706 of intracrystalline diffusion (e.g. Carlson, 2006). However, in order to interpret accurately these  
707 zoning patterns in terms of large-scale and long-term geological/geodynamic processes, we have  
708 to understand the small-scale processes that control the behavior of major and trace elements  
709 during garnet nucleation, growth, breakdown and re-equilibration (George & Gaidies, 2017 and  
710 references therein). Although the zoning features varies from garnet to garnet, they show  
711 systematic variations and similarities that can be correlated with the P-T-X evolution and  
712 petrographic setting. Via a qualitative analysis of zoning profiles, and previous work by others  
713 cited below, we suggest these features reflect the superposition of at least three grain-scale  
714 processes that control the intra- and intergranular mobility of chemical components (major and  
715 trace elements).

716

717 *Evidence of fast intergranular mobility of elements, including REE, and rock-wide equilibrium*

718 The P-T evolution of the studied sample occurred at high temperature (> 650°C) and in  
719 the presence of melt, from peak-metamorphic conditions (0.8 GPa; 790°C) to the conditions of  
720 selvage formation (~0.5 GPa; 800°C). Therefore, the high mobility of major and to a lesser  
721 extent trace elements is expected. We report below evidence of fast intergranular mobility of  
722 elements, including REE, and rock-wide equilibrium.

723 Rock-wide equilibrium and by extension fast inter-granular transport of major and trace  
724 elements during peak metamorphic conditions and subsequent melt-driven metasomatism is best  
725 demonstrated using garnet rim compositions. Although the analyzed garnets are located in  
726 different petrographic settings (within the quartz-undersaturated selvage or in the host-  
727 granulite displaced from the leucosome) and vary markedly in grain size and morphology, their  
728 outermost rims share consistent major element (grossular content = 3%; Figure 9) and trace  
729 element compositions (Y and REEs; Figure 11-13). This consistency in major element composition  
730 of the rims permits an accurate estimation of peak metamorphic conditions at 0.8 GPa and  
731 790°C (Figure 14).

732           The length scale of equilibration during garnet growth, which is related to the degree of  
733 mobility of major and trace elements through the intergranular matrix, can be evaluated using  
734 the REE zoning pattern (Moore et al., 2013; Skora et al., 2006). In both studies, these authors  
735 have shown a dependency of REE distribution on atomic number in large and early nucleating  
736 garnets. Such garnets are characterized by pronounced central peaks for Y and HREEs (Lu to Er),  
737 central peaks and annular maxima for MREEs (Ho, Dy), and well defined annular maxima for  
738 LREEs (Tb to Nd). This distribution is driven by diffusion-limited uptake that evolves during  
739 garnet growth. With increasing temperature, intergranular diffusivities increase progressively as  
740 garnet grows, but with more rapid diffusion for HREEs compared to LREEs (Moore et al., 2013;  
741 Skora et al., 2006). However, this dependency of REE distribution with atomic number is less  
742 valid at high temperature. Numerical modeling of garnet zoning along a temperature-time path  
743 performed by Moore et al. (2013) demonstrates that a uniform bell-shaped distribution for all  
744 REEs is consistent with garnet nucleation and growth at high temperature, whereas at lower  
745 temperature a sharp central peak with an annular maximum would be expected (see Figure 19  
746 and Figure 20b in Moore et al. (2013)). From these previous studies, we can conclude that a  
747 uniform distribution for all REEs is the result of rapid intergranular diffusion that strongly  
748 minimizes the development of concentration gradients in the matrix surrounding the growing  
749 garnet.

750           All studied garnets are characterized by a bell-shape distribution for Y and REEs, ranging  
751 from Lu to Dy ( $\pm$  Tb), with a strong correlation between the radial positions of all peaks. This  
752 distribution is consistent with a progressive depletion of the matrix in Y and REEs (Lu-Dy) due to  
753 Rayleigh fractionation with an unchanging mineral assemblage (Moore et al., 2013; Otamendi et  
754 al., 2002). Although the inferred P-T path in Figure 14 suggests that the onset of garnet growth  
755 may have occurred under sub-solidus conditions prior to muscovite dehydration, we instead  
756 suggest that the majority of garnet growth occurred via biotite dehydration melting in the  
757 trivariant NaCaKFMASH stability field Bio - L - Gt - Kf - sill - pl - q at a temperature greater than  
758  $-760^{\circ}\text{C}$ . The uniform bell-shaped distribution for REEs, ranging from Lu to Dy, is consistent with

759 rapid intergranular diffusivities.

760 All garnet grains are characterized by (1) the absence of progressive outwards migration  
761 of REE peak locations with decreasing atomic number; and (2) the lack of a central peak for  
762 LREEs between La and Gd. Instead, they show a large LREE-depleted core of constant  
763 composition surrounded by an enriched rim. This transition from bell- to bowl-shaped profiles  
764 with decreasing atomic number is similar to an example reported by Raimondo et al. (2017) from  
765 a metasomatic shear zone in central Australia. Both observations (1) and (2) are once again  
766 indicative of fast intergranular mobility and rock-wide equilibrium during garnet growth, without  
767 significant differences in diffusivity between HREEs and LREEs. However, we will see below that  
768 the bowl-shaped profile could also be related, or at least accentuated, by a post-growth  
769 modification.

770 Following Raimondo et al. (2017) and Konrad-Schmolke et al. (2008), we suggest that the  
771 REE zoning reflects a typical Rayleigh fractionation model. As garnet nucleates and grows at high  
772 temperature ( $T > 750^\circ\text{C}$ ), the matrix is progressively depleted in HREEs on a large scale due to the  
773 presence of melt along grain boundaries that favor fast element mobility and rock-wide  
774 equilibration. Simultaneously, the matrix is passively enriched in LREE on a rock-wide scale due  
775 to the low garnet/matrix partition coefficients for LREEs (Moore et al., 2013). The rapid  
776 intergranular mobility for all REEs with respect to garnet growth, without any dependence on  
777 atomic number, precludes the establishment of diffusion gradients in the matrix, resulting in  
778 smooth core-to-rim zoning and a transition from bell-shaped HREE to bowl-shaped LREE profiles.

779 Small garnets located in the quartz-saturated host granulite (e.g. Garnet G) are  
780 characterized by a significantly HREE-depleted composition compared to large garnets located in  
781 the quartz-undersaturated selvage adjacent to the leucosome (Garnets A and B), whereas their  
782 LREE content remains similar. This discrepancy in REE concentrations could be related to a later  
783 nucleation of garnets located in the quartz-present host granulite. Assuming a rapid  
784 intergranular mobility and rock-wide equilibration for REEs, the early nucleation and enhanced  
785 growth of garnet in proximity to the leucosome may have induced a bulk-rock HREE depletion at



786 the scale of the distance between Garnet A and Garnet G (~7 cm). In contrast, the small  
787 garnet/matrix LREE partitioning coefficient has a negligible effect on the bulk rock composition,  
788 resulting in an approximately constant LREE concentration for garnet regardless its nucleation  
789 and growth history.

790

#### 791 *Post-growth intragranular diffusion of major and trace elements*

792 Calcium maps and profiles from Garnets B-G reveal that their prograde growth zoning is  
793 modified near the inner core inclusions. This post-growth modification resulted in chemical re-  
794 equilibration of the grossular content to varying degrees, depending on the location of the  
795 garnet with respect to the quartz-undersaturated selvage and the host granulite. In the quartz-  
796 saturated host granulite, the smallest garnets (which also contain the biggest polyphase  
797 inclusions) show a complete (Garnet F) or partial re-equilibration (Garnets C, D, E, G) of the  
798 garnet core in contact with the inclusions, with grossular content decreasing to 3%,  
799 corresponding to the rim composition. This suggests that post-growth modification occurred near  
800 peak P-T conditions (0.8 GPa; 790°C). Located in the quartz-undersaturated selvage, the largest  
801 analyzed grain (Garnet B) also shows post-growth modification of bell-shaped Ca zoning, but  
802 chemical re-equilibration at peak-metamorphic conditions was only partial, with a minimum  
803 grossular content of 8% in the garnet core.

804 The modification of garnet cores in contact with polyphase inclusions and their strong  
805 similarity with matrix compositions requires a pervasive connectivity to permit such rock-wide  
806 re-equilibration to proceed. Both X-ray tomography (Figure 7) and LA-ICP-MS mapping (e.g.  
807 linear features highlighted in Li map; Figure 10) reveal the presence of a network of open  
808 channels and re-healed cracks that may have acted as potential connecting pathways. These  
809 features confirm that the garnet core was not isolated or armored from the matrix, but was  
810 rather freely accessible to the extent that fluid access along these discrete open fractures or  
811 larger conduits (Figure 7) may favor cation transport and garnet re-equilibration.

812 Furthermore, we suggest that REEs are also susceptible to the post-growth re-equilibration  
813 process. In contact with the polyphase inclusions in Garnet B, the bell-shaped HREE profile is  
814 modified with a decrease in Ho, Dy and Tb that correlates with the Ca zoning (Figure 12). The  
815 remaining HREEs are apparently unaffected by this modification, whereas the modification of  
816 LREEs such as Sm, Gd, La and Ce is more challenging to interpret. This is largely because the  
817 bowl-shaped profile for Sm and Gd, as described above, could also be enhanced by a depletion  
818 in LREEs next to the inner core inclusions. However, Grain G does show a pronounced depletion  
819 in LREEs at the garnet core, resulting in zoning that closely mimics the Ca distribution (compare  
820 Figs 9g and 13). Unlike Grain B, the absence of significant depletion in Ho, Dy and Tb, coupled  
821 with its lower HREE content and well-preserved bell-shaped HREE profile, may support the  
822 interpretation that Grain G represents a late nucleating garnet in the host granulite at peak  
823 metamorphic condition. Post-growth modification proceeded at a lower temperature where only  
824 LREE remained mobile, in the presence of F, Cl and/or P complexes (Ague, 2017; Migdisov,  
825 Williams-Jones & Wagner, 2009; Williams-Jones, Migdisov & Samson, 2012). Alternatively, it may  
826 suggest that selective REE mobility is coupled to Ca substitution via retrograde garnet-  
827 plagioclase exchange that favors LREE over HREE according to kinetic controls and element  
828 compatibility.

829

### 830 *Role of changes in accessory phases and major rock-forming minerals*

831 In the studied sample, the LREE + P whole rock budget is largely controlled by the  
832 stability of monazite. As such, an approximately constant LREE concentration for garnet  
833 whatever its nucleation and growth history, coupled with a smooth bowl-shaped LREE + P profile  
834 prior to post-growth modification (i.e. Garnet A; Figure 13), suggests that monazite did not grow  
835 on the prograde path. Bulk rock fractionation of this kind is incompatible with garnet-monazite  
836 partitioning relationships if their growth was coeval (Warren et al., 2019). Rather, the apparent  
837 lack of monazite competition for LREE + P throughout the garnet growth history necessitates  
838 that monazite first appeared at peak conditions, consistent with the U-Pb age interpretation

839 documented previously and further explained below.

840 Superimposed on the bowl-shaped LREE profile, the rims of Garnets B and G show a  
841 sharp coincident enrichment in Ce, La and P where garnet is in contact with the matrix  
842 assemblage consisting of plagioclase and biotite. The same sharp enrichment is observed next to  
843 polyphase plagioclase and quartz inclusions located in garnet cores. There are two possibilities  
844 for explaining LREE enrichment. We suggest that these sharp enrichments may reflect  
845 modification of garnet zoning during post-peak breakdown of monazite. This interpretation is  
846 consistent with the pervasive occurrence of thin Y-rich overgrowths on monazite located in the  
847 matrix (Figure 18c), which may represent dissolution-precipitation events during the retrograde  
848 evolution. Plagioclase breakdown may also contribute to the LREE enrichment since garnet rims  
849 are locally enriched in large ion lithophile elements (LILEs) such as Ba and Sr, further suggesting  
850 that post-peak garnet-plagioclase exchange is responsible for the modified grossular zoning and  
851 that monazite ages reflect the timing of partial melting prior to breakdown on the retrograde  
852 path.

853 The outermost rims of garnets are characterized by a sharp increase in Y and HREEs along  
854 grain boundaries (Figures 10-13). This local enrichment is interpreted to be the result of partial  
855 garnet breakdown during decompression via reactions (2) and (3) and back-diffusion of elements  
856 that strongly partition into garnet with respect to the matrix. Notably, the absence of sharp  
857 HREE and Y enrichment in core domains, despite an apparently pervasive matrix connectivity  
858 during post-growth re-equilibration and simultaneous monazite replacement, confirms that  
859 garnet resorption must have post-dated these events.

860 Similar to the HREEs, Zr zoning is radial in all garnets except for sharp spikes adjacent to  
861 grain boundaries (figure S4). There are no satellite peaks or annular maxima indicative of zircon  
862 breakdown, nor are there any pronounced depletions indicative of sudden zircon crystallization.  
863 This suggests that zircon grew on the prograde P-T path, with the bulk rock Zr budget  
864 progressively depleted as zircon and garnet grew under rock-wide equilibrium. The distribution  
865 of U-Pb dates supports this interpretation, with concordant zircon analyses at 2.07 Ga being

866 similar to the 2.06 Ga monazite age inferred to represent peak metamorphism and partial  
867 melting. Rimward spikes in Zr are coincident with the HREE and Y enrichment noted above and  
868 slope away from the grain boundary, thus most likely due to garnet resorption rather than zircon  
869 breakdown.

870

871

## 872 CONCLUSIONS

873 The petrological and geochemical investigations of this partially melted granulite highlight the  
874 complexity of processes involved during partial melting at a scale ranging from a few microns to  
875 centimeters. This study confirms and shows that partially melted rocks are open systems in  
876 which changes in P-T conditions, deformation and mass transfer, that we have defined as melt-  
877 driven metasomatism, are preserved in the mineralogical assemblage and the major and trace  
878 element distribution revealed by X-ray microprobe and LA-ICPMS mapping respectively. One of  
879 the main challenges is to be able to distinguish the mineralogical records of a process like  
880 metasomatism from those related to changes in P-T conditions.

881 A first key observation and interpretation is that melt segregation and extraction coupled with  
882 changes in P-T conditions (due to decompression in this example) induce local disequilibrium  
883 between the segregated melt and the residual host-rock. The establishment of chemical  
884 potential gradients at the interface of these two sub-systems drives diffusive mass transfer. Our  
885 thermodynamic modeling shows that while H<sub>2</sub>O, Na<sub>2</sub>O, CaO and K<sub>2</sub>O are transferred from the  
886 melt to the host rock, SiO<sub>2</sub> diffuses in the opposite direction. This combined mass transfer  
887 results in the formation of a quartz-rich and almost anhydrous leucosome and a quartz-  
888 undersaturated selvage at the expense of the host-granulite. Depending on the initial bulk-rock  
889 composition, P-T conditions of melt extraction and shape of the P-T path, the element that will  
890 diffusive (or not) in or out the melt domain and the residue will vary. We suggest that the variety  
891 of mineralogical compositions encountered in selvages surrounding leucosomes in migmatites

892 relates directly to these environmental variables and the induced melt-driven metasomatism.

893 Garnets show a complex major and trace element distribution that varies consistently with the

894 petrographic setting. We suggest that preserved chemical zoning reflects the superposition of

895 changes in P-T conditions and melt-driven metasomatism. Major and trace element zoning was

896 first acquired during prograde and peak metamorphism garnet growth in a temperature range of

897 675 and 790 °C. Garnet grew mostly as a peritectic phase during biotite-dehydration melting

898 reaction. Calcium, Y and HREEs (from Lu to Dy), are characterized by a bell-shape distribution

899 and LREEs (from La to Gd) show a bowl-shape. The chemical zoning is consistent with a typical

900 Rayleigh fractionation model with fast intergranular element mobility of all elements, including

901 REEs, and rock-wide equilibration. The fast mobility is enhanced by the presence of melt along

902 grain boundaries.

903 The post-peak melt-driven metasomatism event is responsible for the modification of the

904 prograde zoning growth. Both Calcium and REE's are susceptible to the post-growth alteration

905 and results in a complete or partial re-equilibration through diffusion of garnet composition at

906 the P-T-X conditions of metasomatism (0.8 to 0.5 GPa; 790 °C). Garnet cores in contact with

907 polyphased inclusions were affected by the post-growth modification due to the presence of a

908 network of open channels and cracks that act as connecting pathways between the matrix and

909 garnet core. Once again, at these high temperature conditions and in the presence of melt (and

910 fluid), divalent cation ( $\text{Ca}^{2+}$ ) and REE are characterized by fast inter-granular and intragranular

911 element mobility and rock-wide equilibration. We conclude that in this peculiar case (high

912 temperature and melt/fluid-bearing open system), REE are not less vulnerable to diffusive

913 resetting than divalent cations.

914 Besides the interpretations regarding the formation of the quartz-undersaturated selvage and

915 the mobility of major and trace elements at high temperature conditions, this contribution

916 shows that an inaccurate interpretation of chemical zoning in garnet could result in misleading

917 thermobarometric results. Indeed, if the modified garnet core compositions (i.e. low calcium

918 content) were used as the chemical fingerprint of the early garnet nucleation stage it would

919 have resulted in erroneous low-pressure condition estimates for garnet nucleation.

920

921

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## 939 **FIGURE CAPTIONS**

940

941

942 Figure 1. (a) Aerial photograph of the Farol da Barra outcrop in the city of Salvador da Bahia,  
943 with a corresponding structural and geological map (right-hand side). (b) Simplified structural  
944 cross section including the studied area, showing the finite strain pattern consisting of a gently  
945 dipping S1 fabric and steeply dipping D2 high strain zones. On both figures, the location of the  
946 studied sample 06SB06 is shown as a red star.

947

948 Figure 2. (a) Photograph of the studied outcrop showing the steeply dipping S2 gneissic foliation.  
949 The leucosome is surrounded by a melanocratic quartz-undersaturated selvage. (b) Polished slice

45

950 of the studied sample 06SB06 showing the location of the three thin sections.

951

952 Figure 3. (a) Photomicrograph of the metamorphic assemblage in the quartz-saturated granulite.

953 (b) Cordierite development at the contact between garnet and sillimanite in the quartz-

954 saturated granulite. (c) Spinel and cordierite symplectite developed at the expense of garnet

955 and sillimanite in the quartz-undersaturated selvage.

956

957 Figure 4. Thin-section scans under crossed-polarized light showing the location of the seven

958 analyzed garnets (A to G). Garnets A and B are located in the quartz-undersaturated selvage.

959 Notice the change in grain size and proportion of inclusions with increasing distance from the

960 leucosome.

961

962 Figure 5. Photomicrographs of the seven analyzed garnets. (a) Garnet A; (b) Garnet B; (c) Garnet

963 C; (d) Garnet D; (e) Garnet E; (f) Garnet F; (g) Garnet G. The size of the scale bar shown is 50

964  $\mu\text{m}$  for all cases, and dashed lines correspond to the location of major element profiles shown in

965 Figure 9.

966

967 Figure 6. Backscattered electron images of crystallized melt inclusions in the outer rim of Garnet

968 B located in the quartz-undersaturated selvage. (a), garnet negative crystal shape crystallized

969 melt inclusion; (b) polyphase xenomorph crystallized melt inclusion; and (c) radial crack

970 attesting to melt inclusion decrepitation

971

972 Figure 7. (a) Backscattered electron image of a representative garnet in the quartz-saturated

973 granulite. (b) 3D reconstruction from a computed X-ray microtomography of garnet. (c-f)

974 sequential slices of the 3D reconstructed garnet (in red) showing the channels (white arrows)



975 connecting the irregular inclusions (transparent) to the outside matrix.

976

977 Figure 8. (a) Cathodoluminescence images of K-feldspar texture (blue) with plagioclase films and  
978 pool (brown) along grain boundaries. Plagioclase films are connected to myrmekite pools (m).

979 (b) Extensive replacement of K-feldspar by plagioclase and myrmekites.

980

981 Figure 9. Garnet composition (black square = molar amount of grossular and white square =  
982 XMg). The location of each line profile is shown in figure 5. The dark and light gray stripes  
983 correspond to plagioclase-quartz inclusions and domains with minute inclusions of quartz,  
984 respectively.

985

986 Figure 10. LA-ICP-MS trace element maps of (a) garnet A; (b) Garnet B; (c) Garnet G.

987 Concentrations values are in ppm. The line profiles shown in Figure 11-13 are located in each  
988 case on the Phosphorus (P) map.

989

990 Figure 11. LA-ICP-MS trace element line profile of Garnet A. . See Figure 10a for traverse  
991 location.

992

993 Figure 12. LA-ICP-MS trace element line profile of Garnet B. See Figure 10b for traverse  
994 location.

995

996 Figure 13. LA-ICP-MS trace element line profile of Garnet G. See Figure 10c for traverse  
997 location.

998

999 Figure 14. P-T phase diagram section computed for with quartz-saturated granulite composition  
1000 (see Table 1). Circles with letters indicate the P-T conditions corresponding to measured garnet  
1001 compositions. The letters refer to garnet grain labels in Figure 5.

1002

1003 Figure 15. Evolution of chemical potential of H<sub>2</sub>O, SiO<sub>2</sub>, CaO and K<sub>2</sub>O for the quartz-saturated  
1004 domain (black line) and melt (dashed line) along a decompression path from 0.8 to 0.5 GPa at  
1005 790° C.

1006

1007 Figure 16. P-X phase diagram section computed at 790° C with X=0 corresponding to the quartz-  
1008 saturated domain and X=1 corresponding to a 55% loss of SiO<sub>2</sub> and 200% gain of H<sub>2</sub>O. See Table 1  
1009 for compositions.

1010

1011 Figure 17. P-T phase diagram section computed with a modeled quartz-undersaturated selvage  
1012 composition produced after a 47% loss of SiO<sub>2</sub> and 140% gain of H<sub>2</sub>O. See Table 1 for  
1013 compositions.

1014

1015 Figure 18. (a) Full thin-section X-ray Al map of sample 06SB06a showing the petrographic setting  
1016 of dated monazites. (b) Y map of monazite included in garnet and plagioclase. (c) Y map of  
1017 monazite in the cordierite-bearing matrix. Note the presence of thin Y-rich overgrowths.

1018

1019 Figure 19. (a) Monazite U-Th-Pb concordia diagram. (b) Monazite Th-Pb ages. (c) Zircon U-Pb  
1020 concordia diagram

1021

1022 **TABLE CAPTIONS**

1023 Table 1. Bulk geochemical compositions used in phase diagrams shown in figures 14-17

1024

## 1025 DATASETS AND SUPPORTING INFORMATION

1026 Figure S1. Simplified geological map of the São Francisco Craton modified after Barbosa et al.  
1027 (2003).

1028

1029 Figure S2. (a) Simplified structural map of the southwestern part of the city of Salvador da  
1030 Bahia. (b) Simplified structural cross-section. (c) Equal-area Schmidt lower hemisphere  
1031 projection of poles to  $S_1$  and  $S_2$  foliations (dashed lines = conjugated shear zones associated the  
1032  $S_2$  mylonitic foliation).

1033

1034

1035

1036 Figure S3. (a) Backscattered image of a polyphased plagioclase and quartz inclusion in a garnet.  
1037 (b) anorthite vs albite content in mole of plagioclase. Grey and black diamonds are plagioclase  
1038 composition in contact with quartz and garnet respectively. (c) Profile showing the evolution of  
1039 plagioclase composition in contact with quartz and garnet. See location of the profile in figure a.

1040

1041 Figure S4. LA-ICP-MS Zr line profile of garnet A, B and G

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1044

1045

1046 Movie S1. 3D reconstruction from computed X-ray microtomography of garnet (see Figure 7).

1047

1048 Table S1. Solid solution models used in Figures 14-17

1049

1050 Dataset S1. Electron microprobe analysis of biotite, plagioclase, spinel and garnet. Garnet  
1051 compositions include the seven profiles collected from Garnets A-G in Figure 9 (see Figure 5 for  
1052 the line profile locations).

1053

1054 Dataset S2. LA-ICP-MS trace element data of line profiles collected from Garnets A, B and G (see  
1055 Figures 11-13).

1056

1057 Dataset S3. Raw LA-ICPMS trace element data of garnet A (see map in Figure 10a)

1058

1059 Dataset S4. Raw LA-ICPMS trace element data of garnet B (see map in Figure 10b)

1060

1061 Dataset S5. Raw LA-ICPMS trace element data of garnet A (see map in Figure 10c)

1062

1063 Dataset S6. LA-ICP-MS isotopic U-Th-Pb data of monazite and zircon

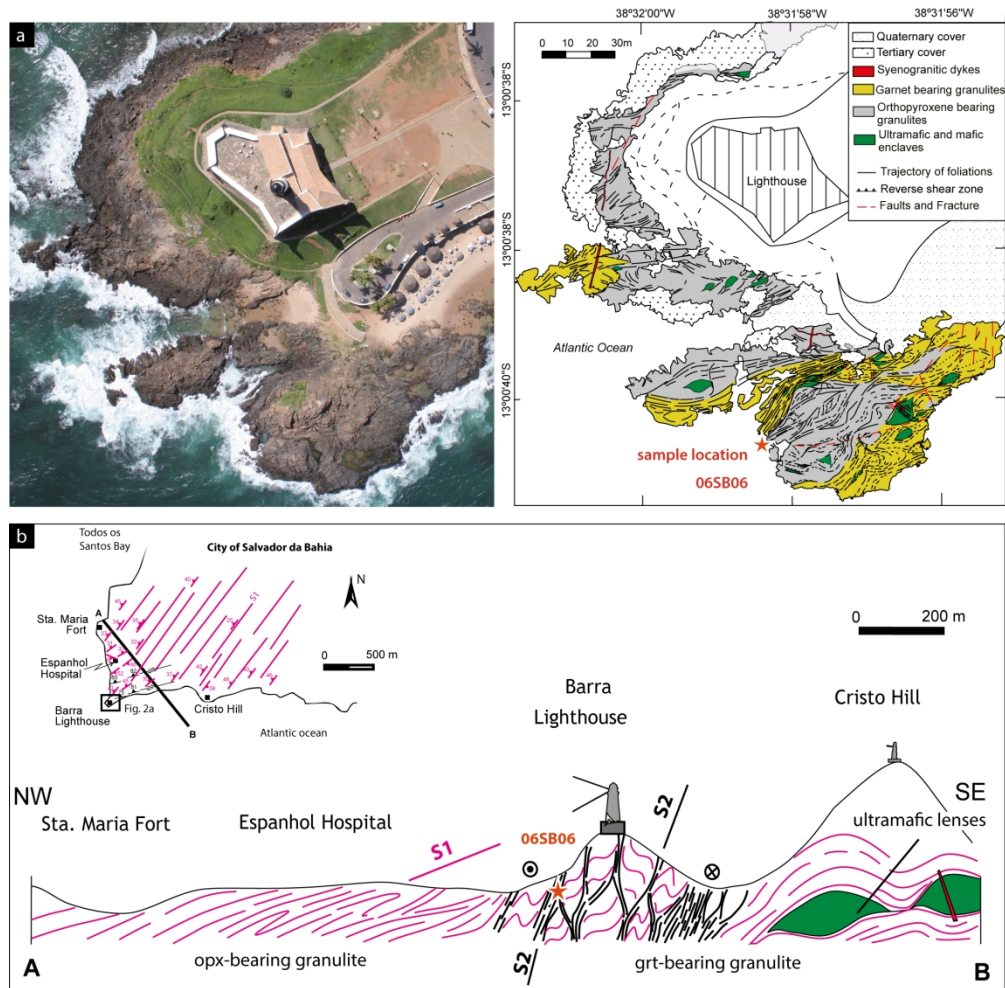


Figure 1

Figure 1. (a) Aerial photograph of the Farol da Barra outcrop in the city of Salvador da Bahia, with a corresponding structural and geological map (right-hand side). (b) Simplified structural cross section including the studied area

180x188mm (300 x 300 DPI)

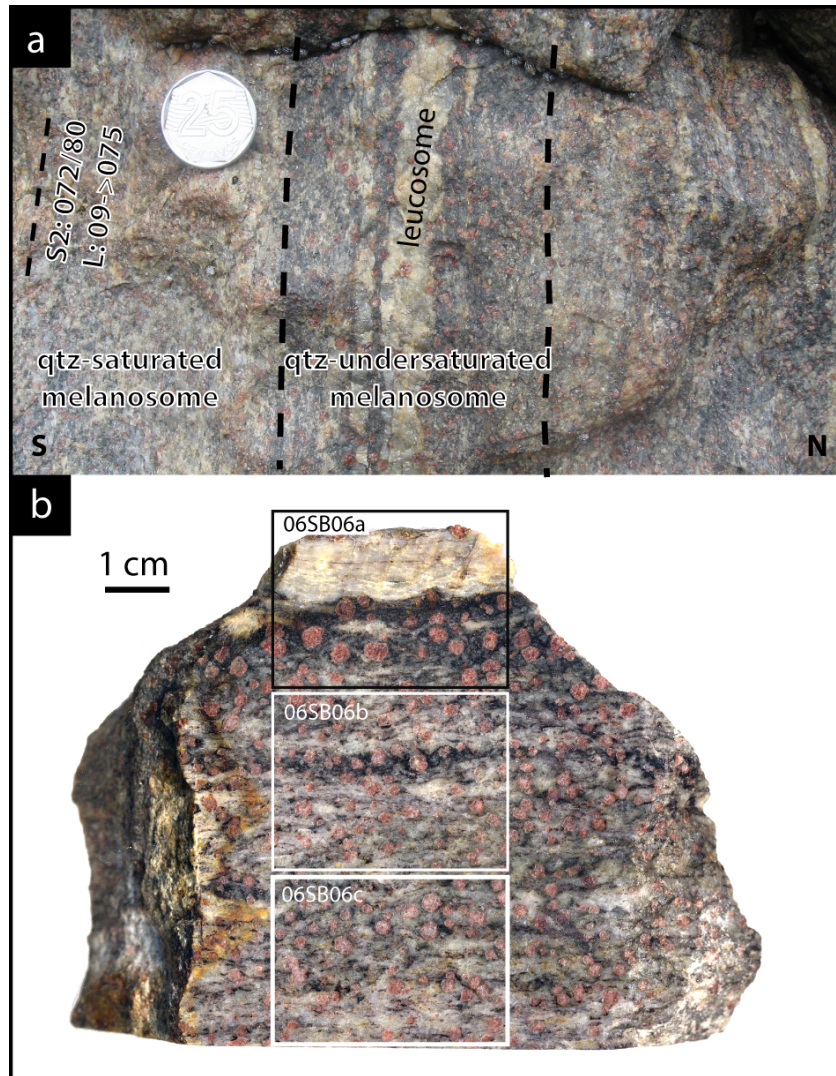


figure 2

Figure 2. (a) Photograph of the studied outcrop showing the steeply dipping S2 gneissic foliation. The leucosome is surrounded by a melanocratic quartz-undersaturated selvage. (b) Polished slice of the studied sample 06SB06 showing the location of the three thin sections.

80x115mm (300 x 300 DPI)

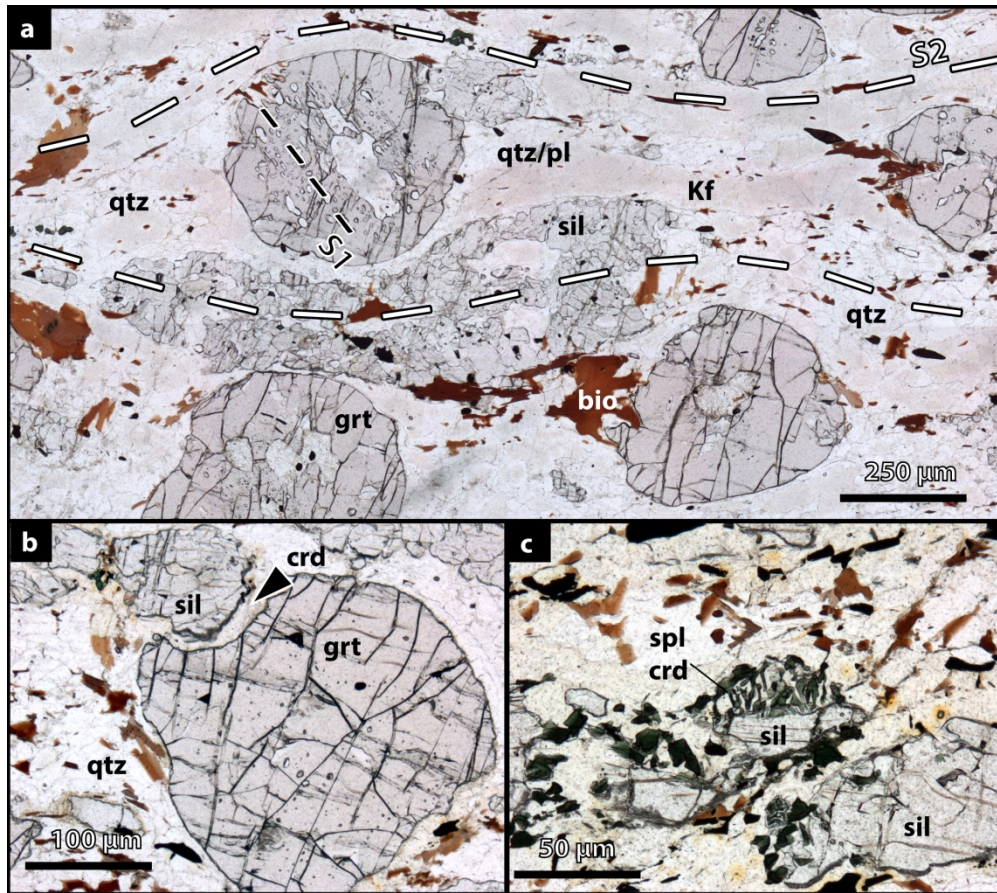


Figure 3

Figure 3. (a) Photomicrograph of the metamorphic assemblage in the quartz-saturated granulite. (b) Cordierite development at the contact between garnet and sillimanite in the quartz-saturated granulite. (c) Spinel and cordierite symplectite developed at the expense of garnet and sillimanite in the quartz-undersaturated selvage.

180x173mm (300 x 300 DPI)

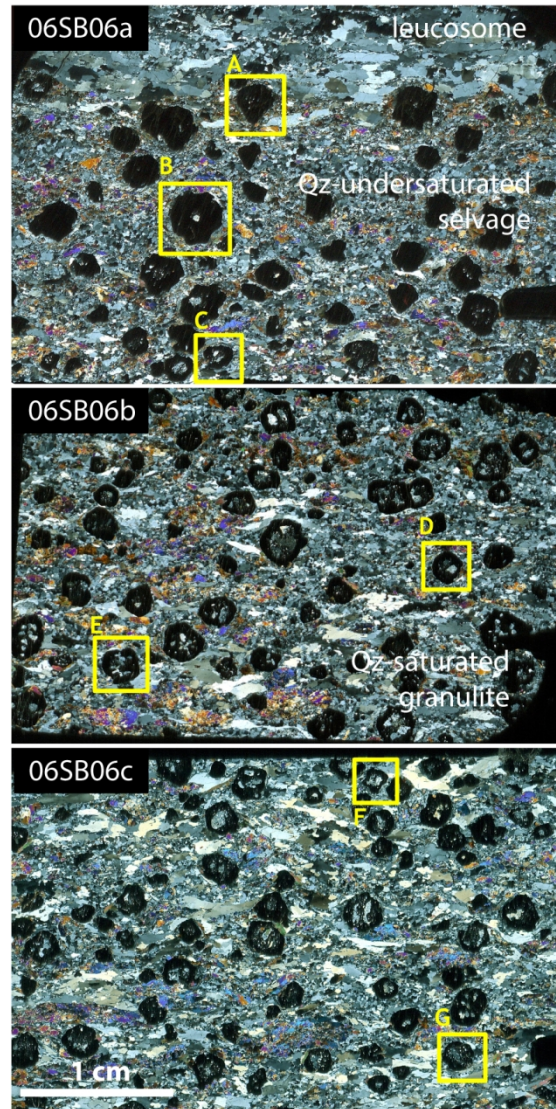


Figure 4

Figure 4. Thin-section scans under crossed-polarized light showing the location of the seven analyzed garnets (A to G).

85x160mm (300 x 300 DPI)



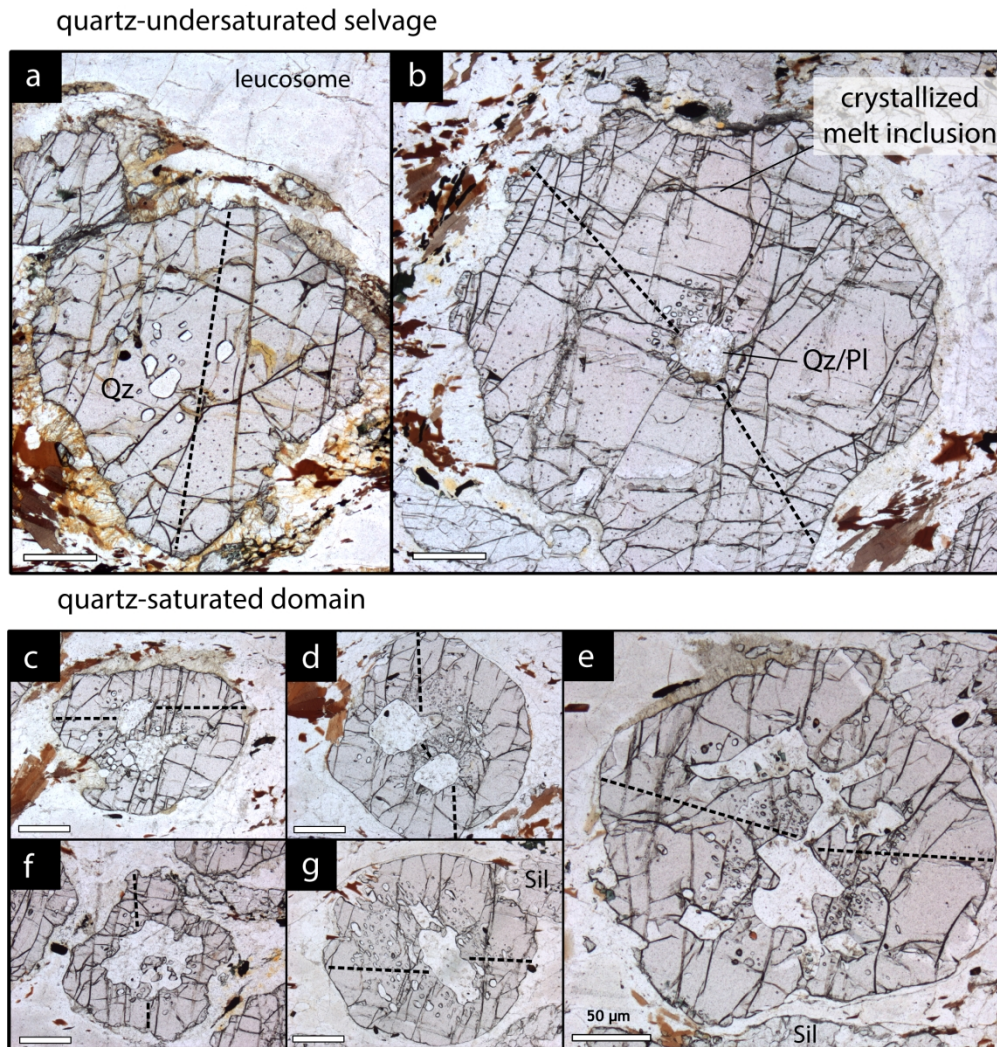


Figure 5

Figure 5. Photomicrographs of the seven analyzed garnets. (a) Garnet A; (b) Garnet B; (c) Garnet C; (d) Garnet D; (e) Garnet E; (f) Garnet F; (g) Garnet G. The size of the scale bar shown is 50  $\mu$ m for all cases, and dashed lines correspond to the location of major element profiles shown in Figure 9.

180x206mm (300 x 300 DPI)

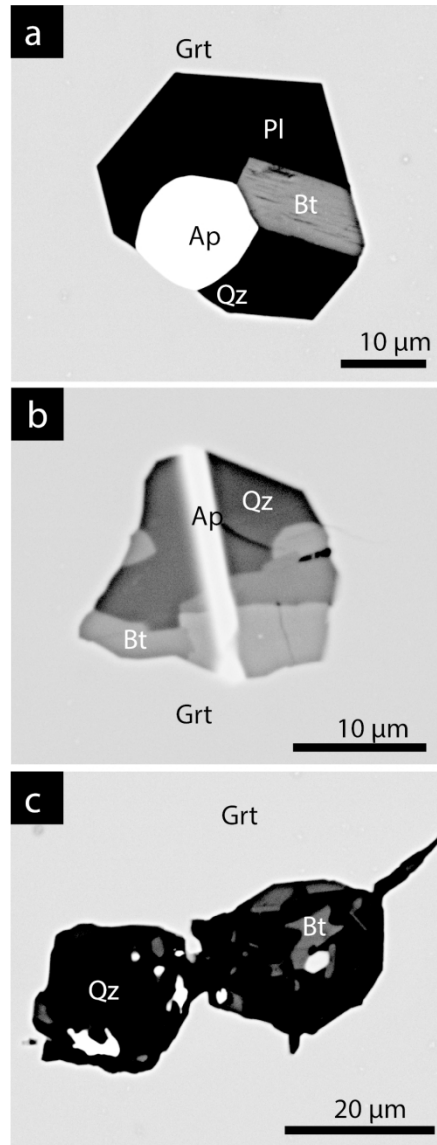


Figure 6

Figure 6. Backscattered electron images of crystallized melt inclusions in the outer rim of Garnet B located in the quartz-undersaturated selvage. (a), garnet negative crystal shape crystallized melt inclusion; (b) polyphase xenomorph crystallized melt inclusion; and (c) radial crack attesting to melt inclusion decrepitation

80x220mm (300 x 300 DPI)

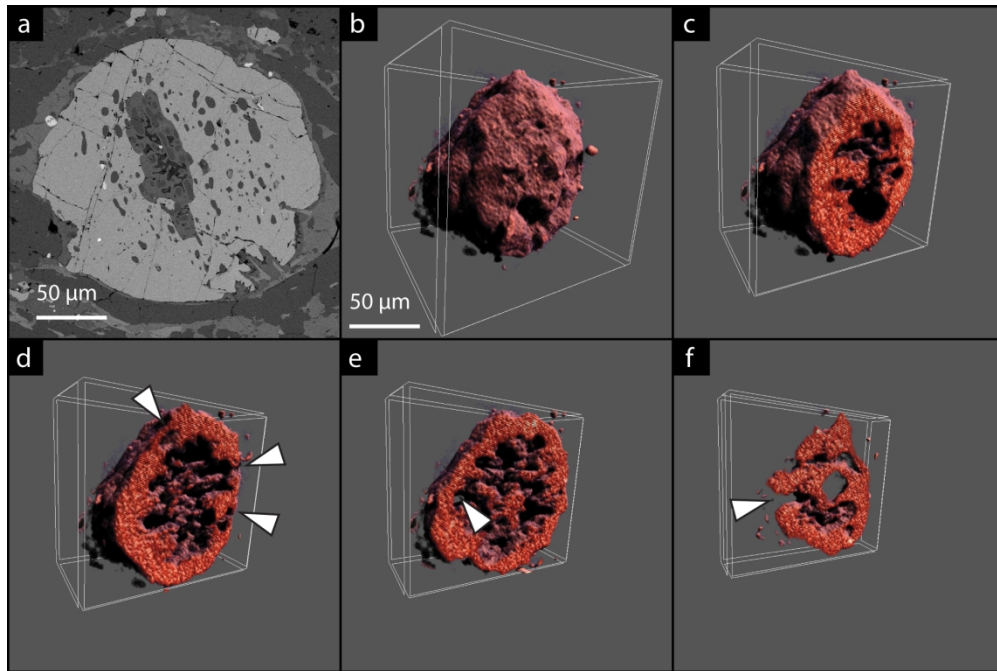


Figure 7

Figure 7. (a) Backscattered electron image of a representative garnet in the quartz-saturated granulite. (b) 3D reconstruction from a computed X-ray microtomography of garnet. (c-f) sequential slices of the 3D reconstructed garnet (in red) showing the channels (white arrows) connecting the irregular inclusions (transparent) to the outside matrix.

180x140mm (300 x 300 DPI)

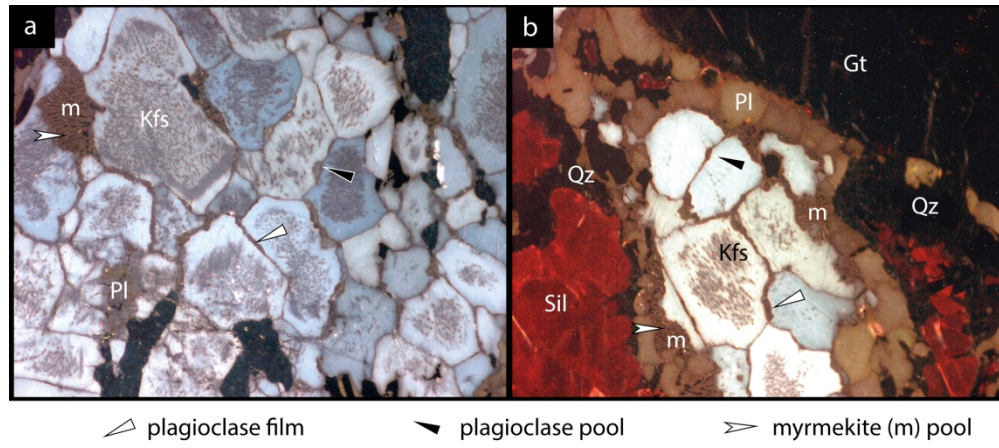


Figure 8

Figure 8. (a) Cathodoluminescence images of K-feldspar texture (blue) with plagioclase films and pool (brown) along grain boundaries. Plagioclase films are connected to myrmekite (m). (b) Extensive replacement of K-feldspar by plagioclase and myrmekites.

180x92mm (300 x 300 DPI)

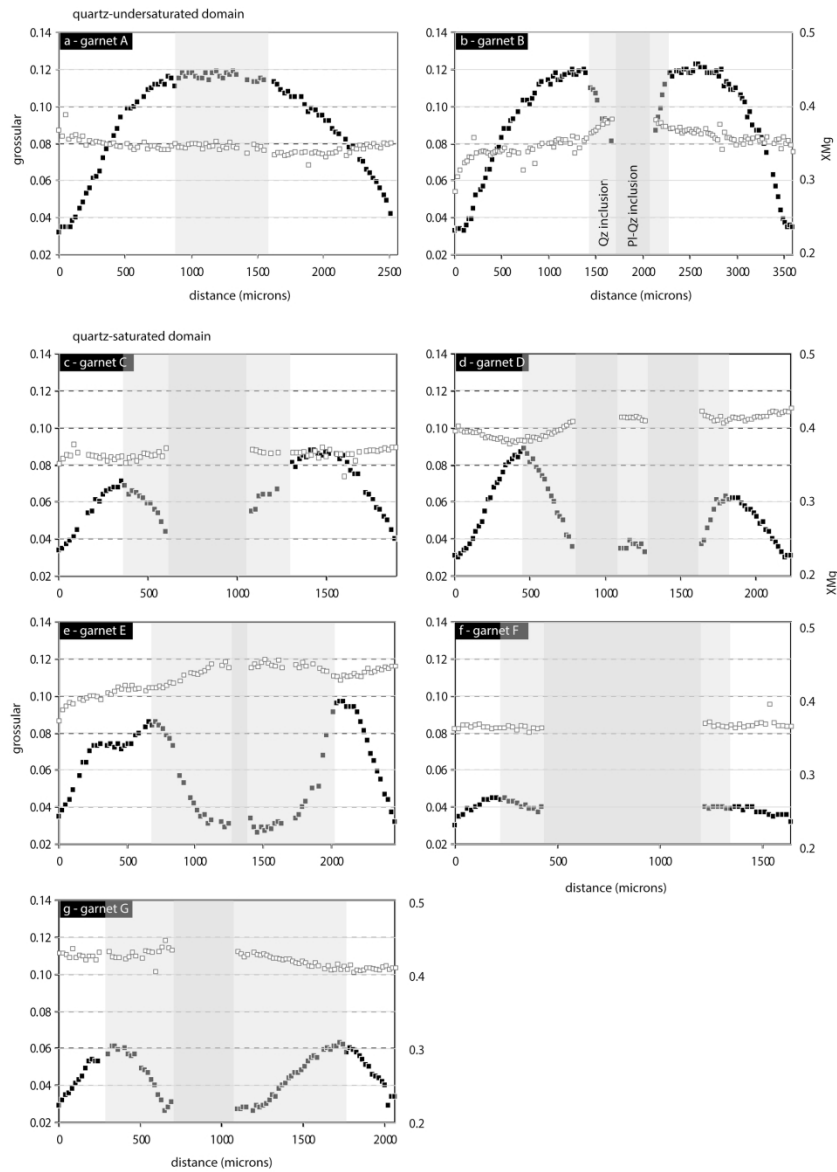


Figure 9

Figure 9. Garnet composition (black square = molar amount of grossular and white square = XMg). The location of each line profile is shown in figure 5. The dark and light gray stripes correspond to plagioclase-quartz inclusions and domains with minute inclusions of quartz, respectively.

180x260mm (300 x 300 DPI)

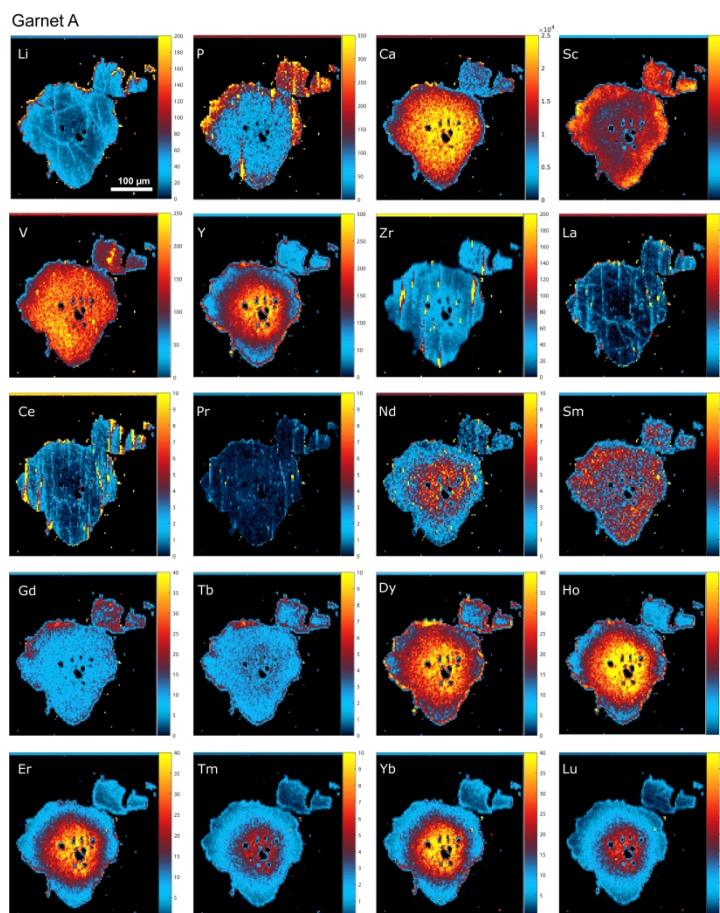


figure 10a

Figure 10. LA-ICP-MS trace element maps of (a) garnet A; (b) Garnet B; (c) Garnet G. Concentrations values are in ppm. The line profiles shown in Figure 11-13 are located in each case on the Phosphorus (P) map.

210x297mm (300 x 300 DPI)

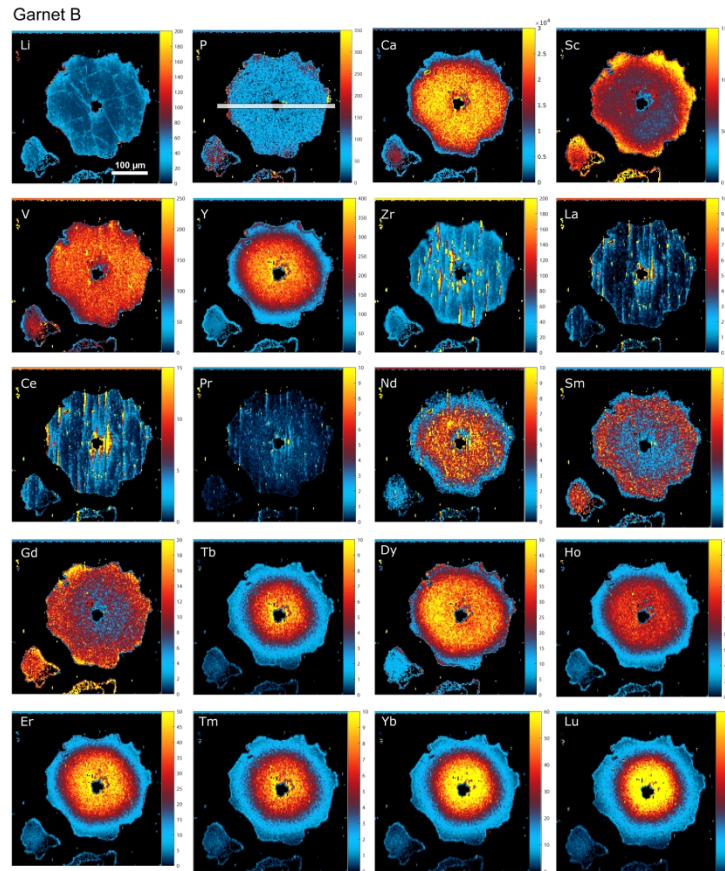


Figure 10b

Figure 10. LA-ICP-MS trace element maps of (a) garnet A; (b) Garnet B; (c) Garnet G. Concentrations values are in ppm. The line profiles shown in Figure 11-13 are located in each case on the Phosphorus (P) map.

210x297mm (300 x 300 DPI)

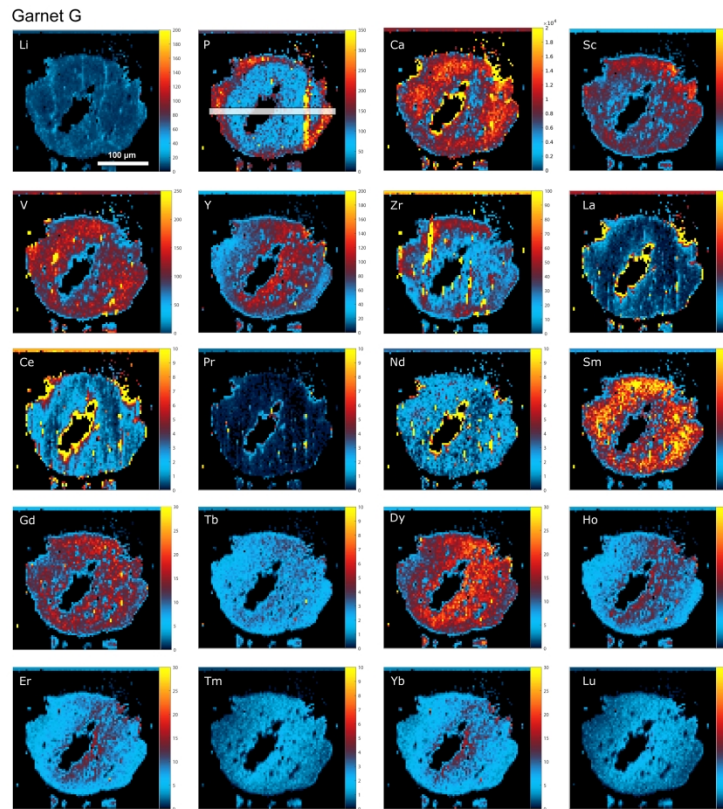


Figure 10c

Figure 10. LA-ICP-MS trace element maps of (a) garnet A; (b) Garnet B; (c) Garnet G. Concentrations values are in ppm. The line profiles shown in Figure 11-13 are located in each case on the Phosphorus (P) map.

210x297mm (300 x 300 DPI)



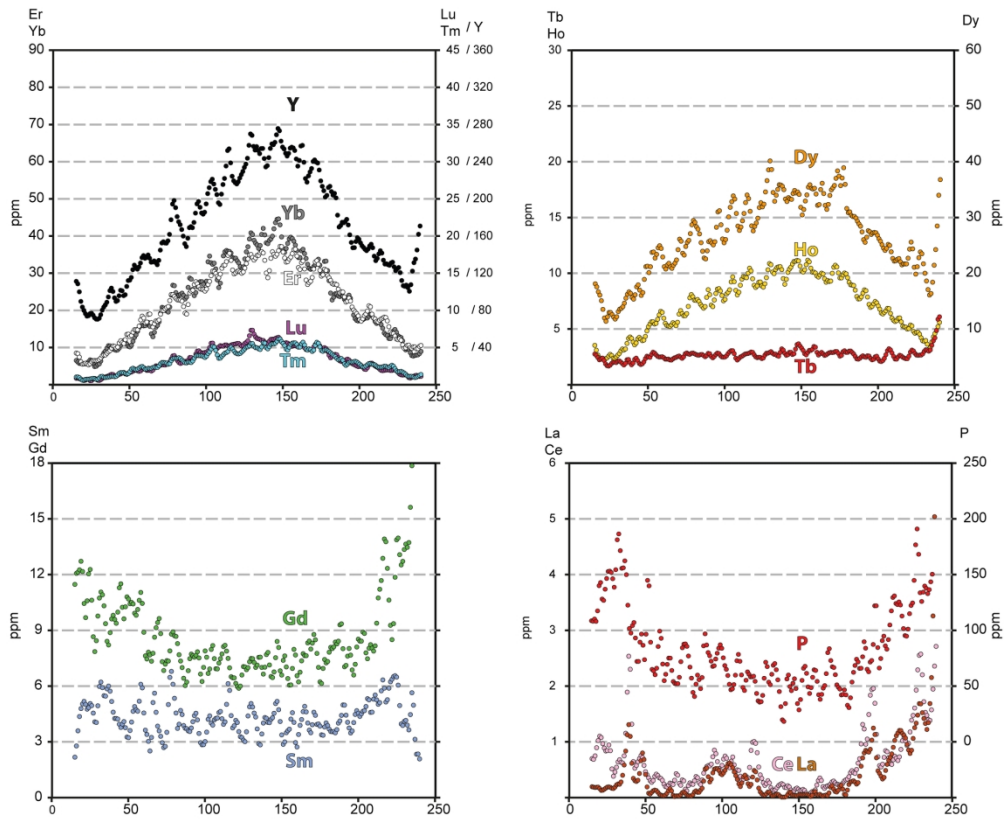


Figure 11

Figure 11. LA-ICP-MS trace element line profile of Garnet A. . See Figure 10a for traverse location.

176x158mm (300 x 300 DPI)

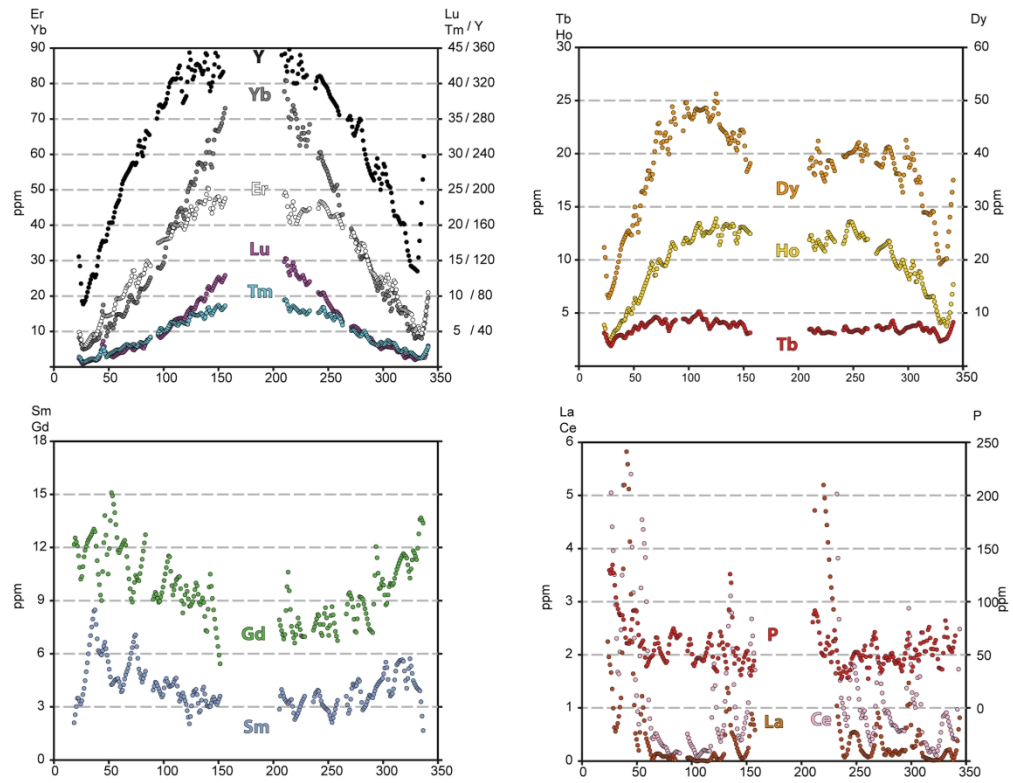


Figure 12

Figure 12. LA-ICP-MS trace element line profile of Garnet B. See Figure 10b for traverse location.

185x165mm (300 x 300 DPI)

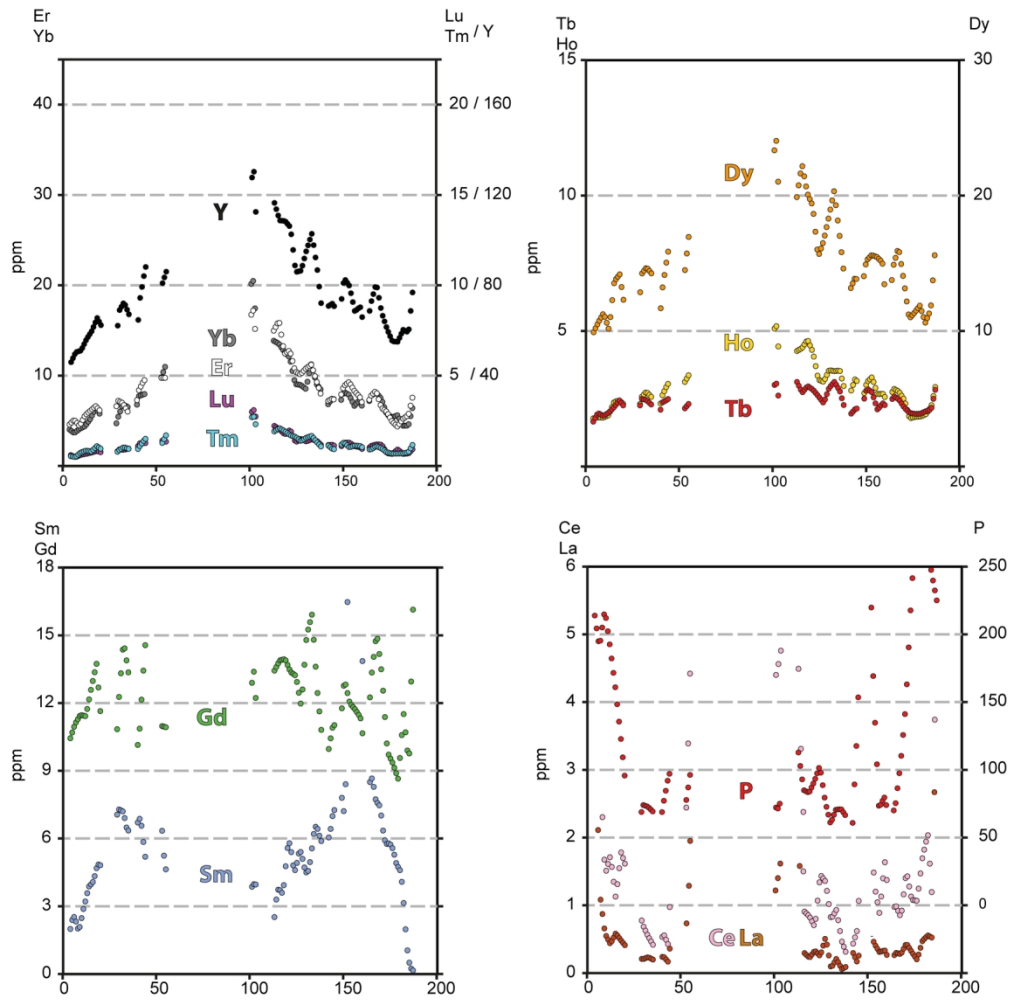


Figure 13

Figure 13. LA-ICP-MS trace element line profile of Garnet G. See Figure 10c for traverse location.

145x161mm (300 x 300 DPI)

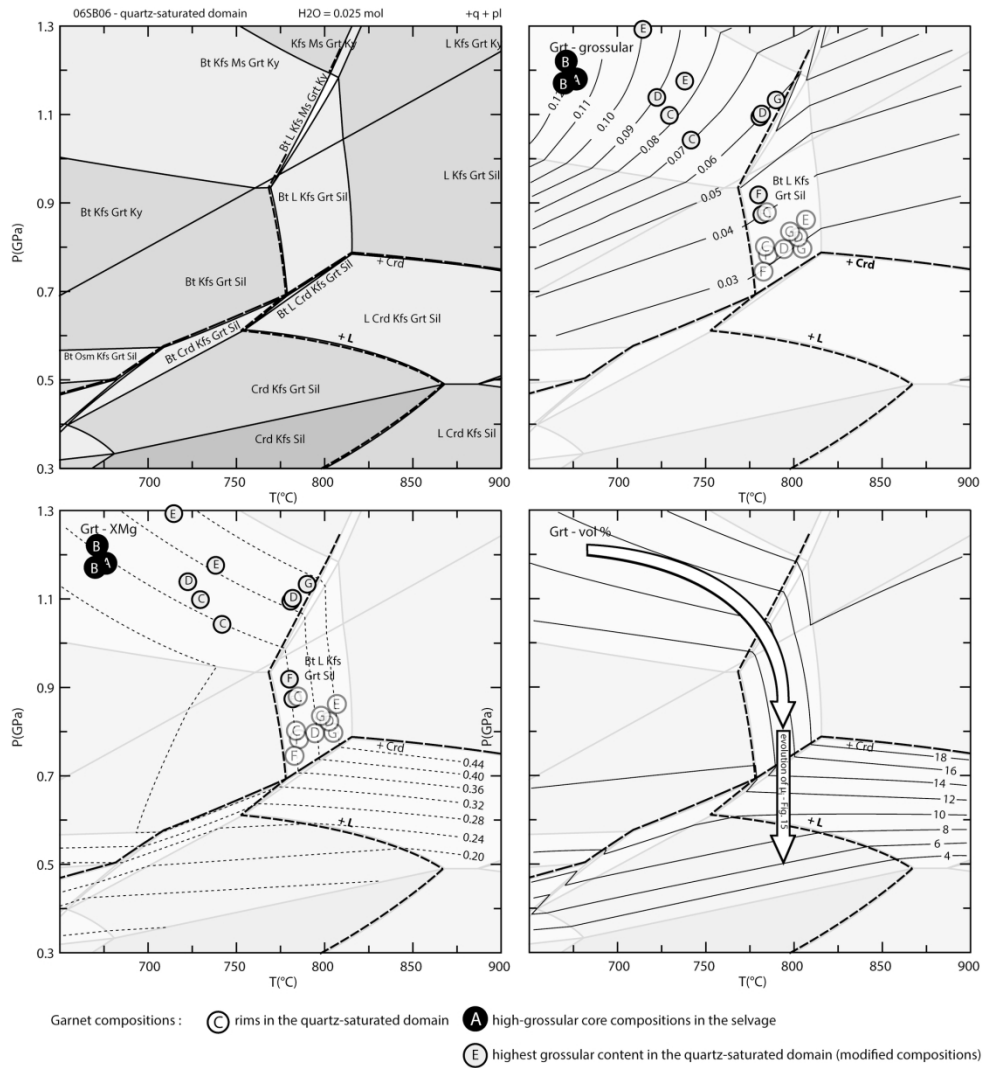


Figure 14

Figure 14. P-T phase diagram section computed for with quartz-saturated granulite composition (see Table 1). Circles with letters indicate the P-T conditions corresponding to measured garnet compositions. The letters refer to garnet grain labels in Figure 5.

180x204mm (300 x 300 DPI)

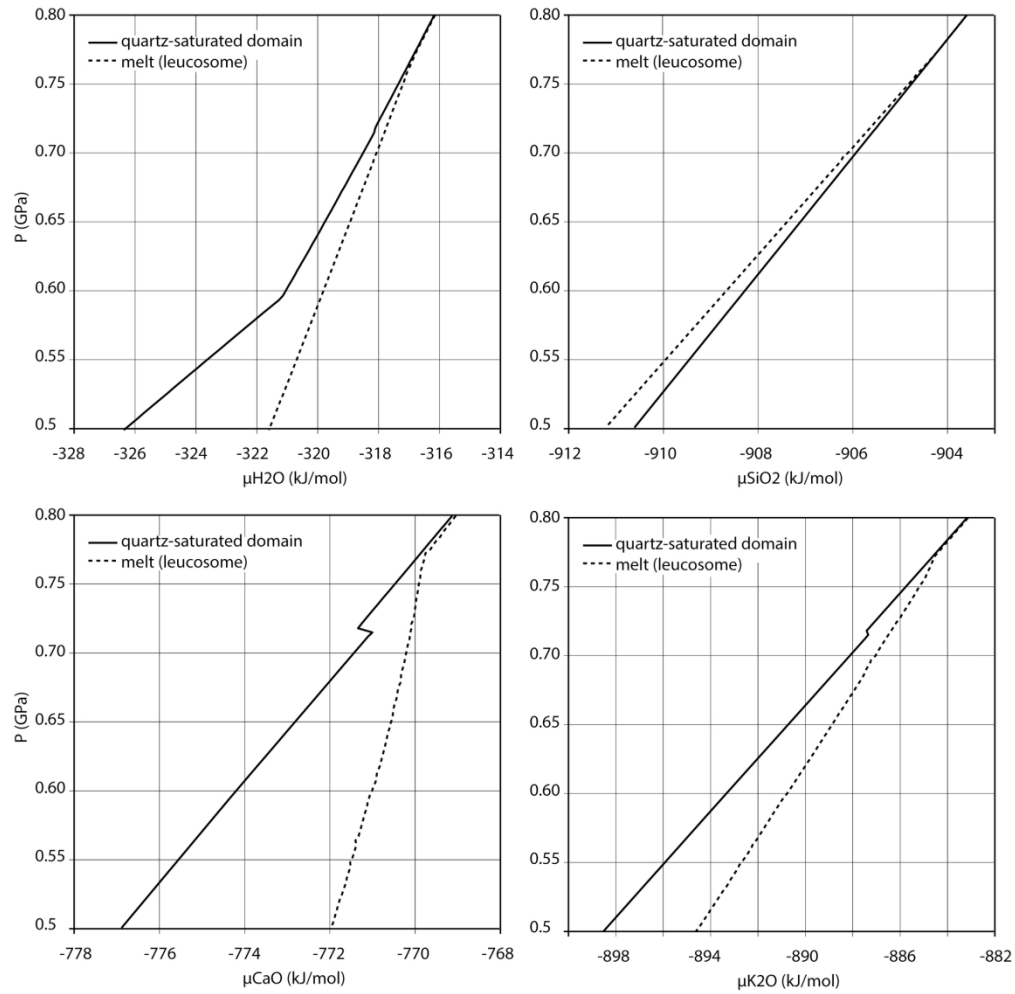


Figure 15

Figure 15. Evolution of chemical potential of H<sub>2</sub>O, SiO<sub>2</sub>, CaO and K<sub>2</sub>O for the quartz-saturated domain (black line) and melt (dashed line) along a decompression path from 0.8 to 0.5 GPa at 790°C.

180x191mm (300 x 300 DPI)

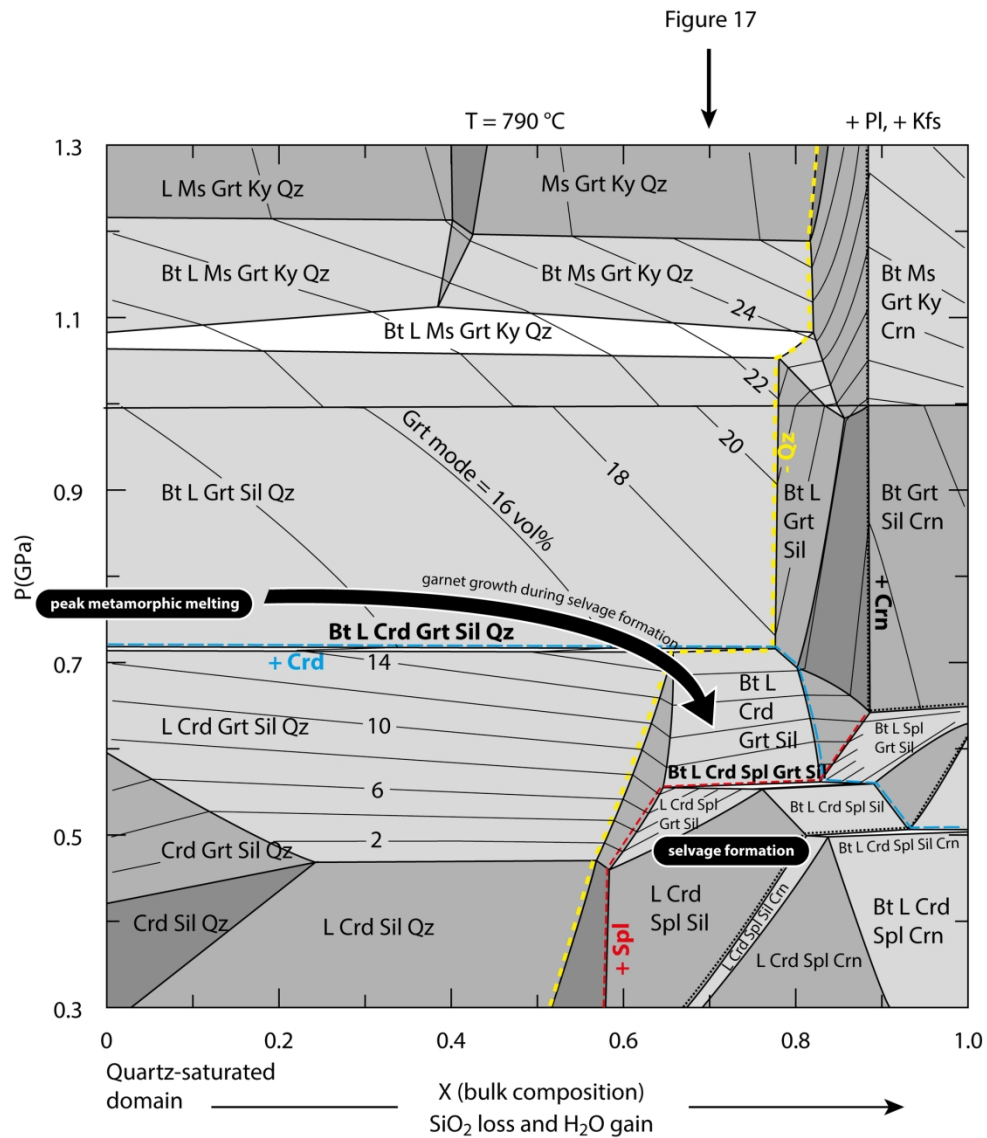


Figure 16

Figure 16. P-X phase diagram section computed at 790°C with X=0 corresponding to the quartz-saturated domain and X=1 corresponding to a 55% loss of SiO<sub>2</sub> and 200% gain of H<sub>2</sub>O. See Table 1 for compositions.

171x212mm (300 x 300 DPI)

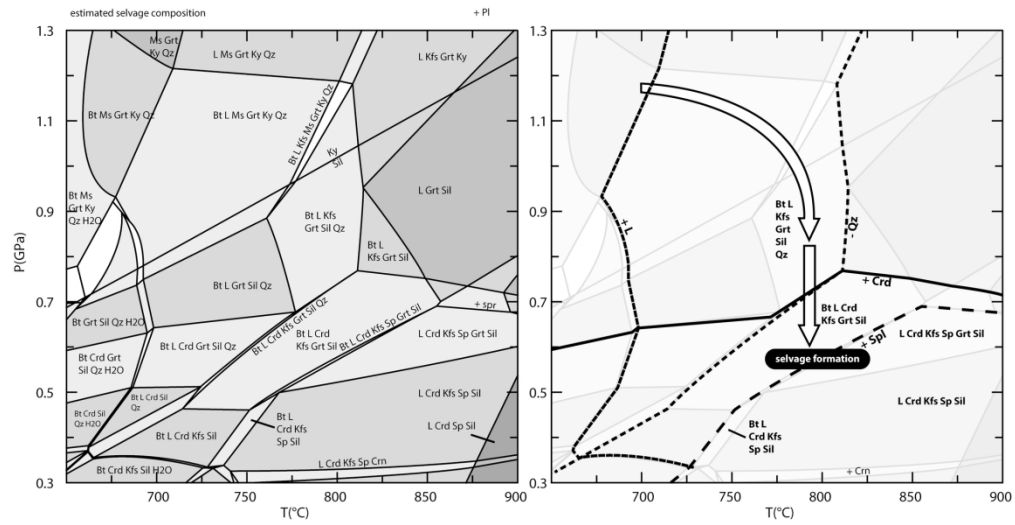


Figure 17

Figure 17. P-T phase diagram section computed with a modeled quartz-undersaturated selvage composition produced after a 47% loss of SiO<sub>2</sub> and 140% gain of H<sub>2</sub>O. See Table 1 for compositions.

179x103mm (300 x 300 DPI)

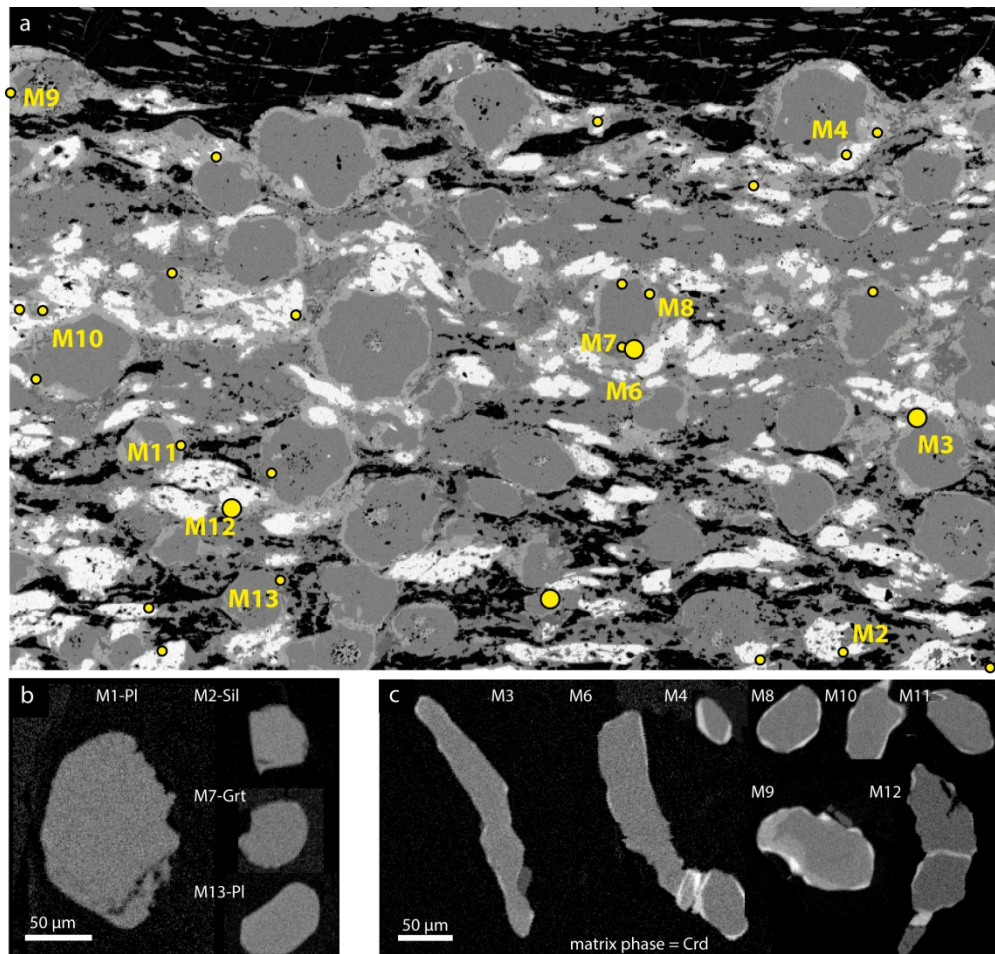


Figure 18. (a) Full thin-section X-ray Al map of sample 06SB06a showing the petrographic setting of dated monazites. (b) Y map of monazite included in garnet and plagioclase. (c) Y map of monazite in the cordierite-bearing matrix. Note the presence of thin Y-rich overgrowths.

180x171mm (300 x 300 DPI)



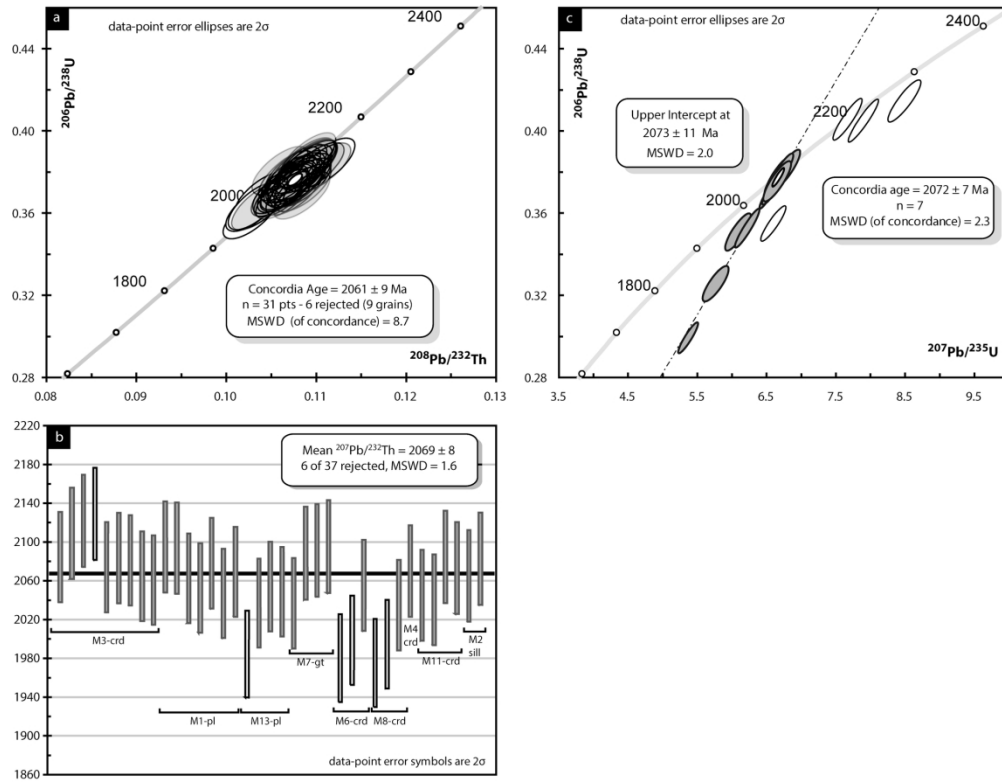


Figure 19. (a) Monazite U-Th-Pb concordia diagram. (b) Monazite Th-Pb ages. (c) Zircon U-Pb concordia diagram

180x139mm (300 x 300 DPI)

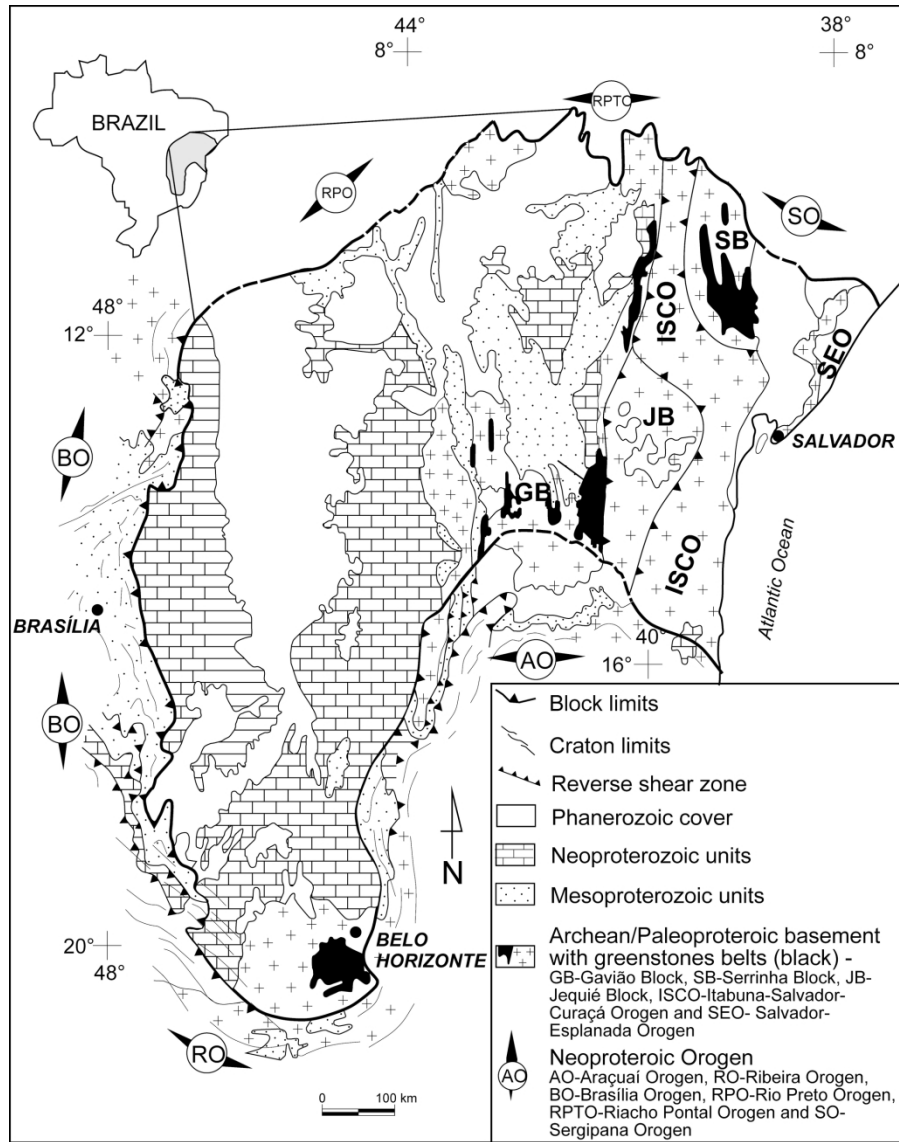


Figure 1

Figure S1. Simplified geological map of the São Francisco Craton modified after Barbosa et al. (2003).

180x240mm (300 x 300 DPI)

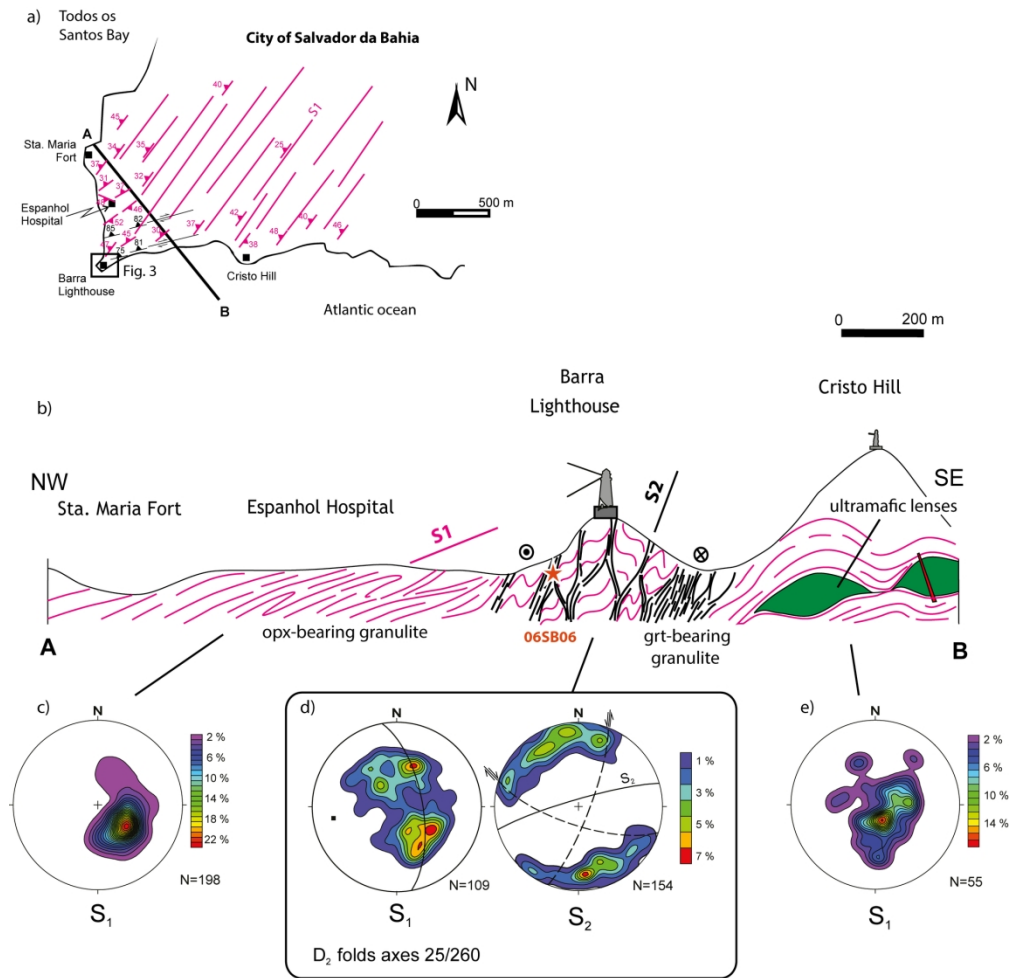


Figure S1

Figure S2. (a) Simplified structural map of the southwestern part of the city of Salvador da Bahia. (b) Simplified structural cross-section. (c) Equal-area Schmidt lower hemisphere projection of poles to S1 and S2 foliations (dashed lines = conjugated shear zones associated the S2 mylonitic foliation).

180x197mm (300 x 300 DPI)

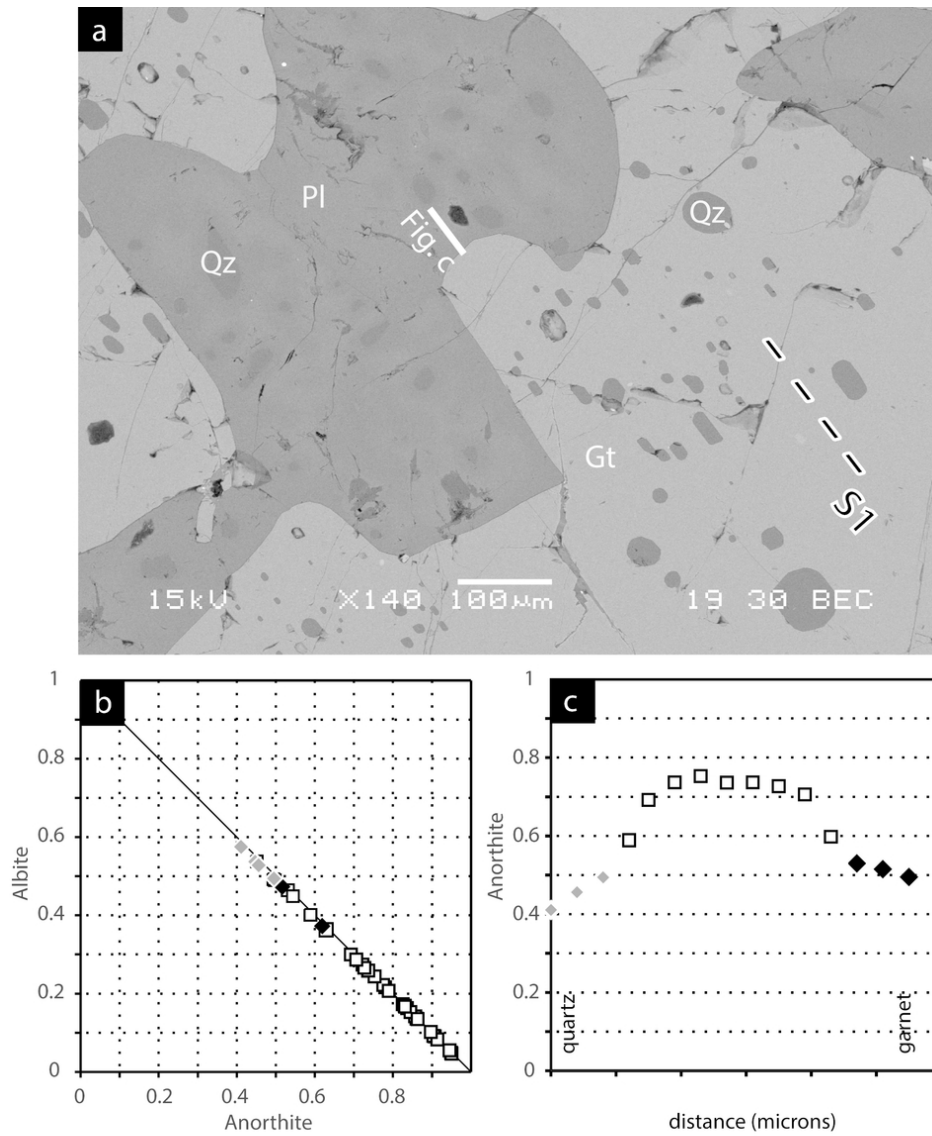


Figure S3

Figure S3. (a) Backscattered image of a polyphased plagioclase and quartz inclusion in a garnet. (b) anorthite vs albite content in mole of plagioclase. Grey and black diamonds are plagioclase composition in contact with quartz and garnet respectively. (c) Profile showing the evolution of plagioclase composition in contact with quartz and garnet. See location of the profile in figure a.

80x102mm (300 x 300 DPI)

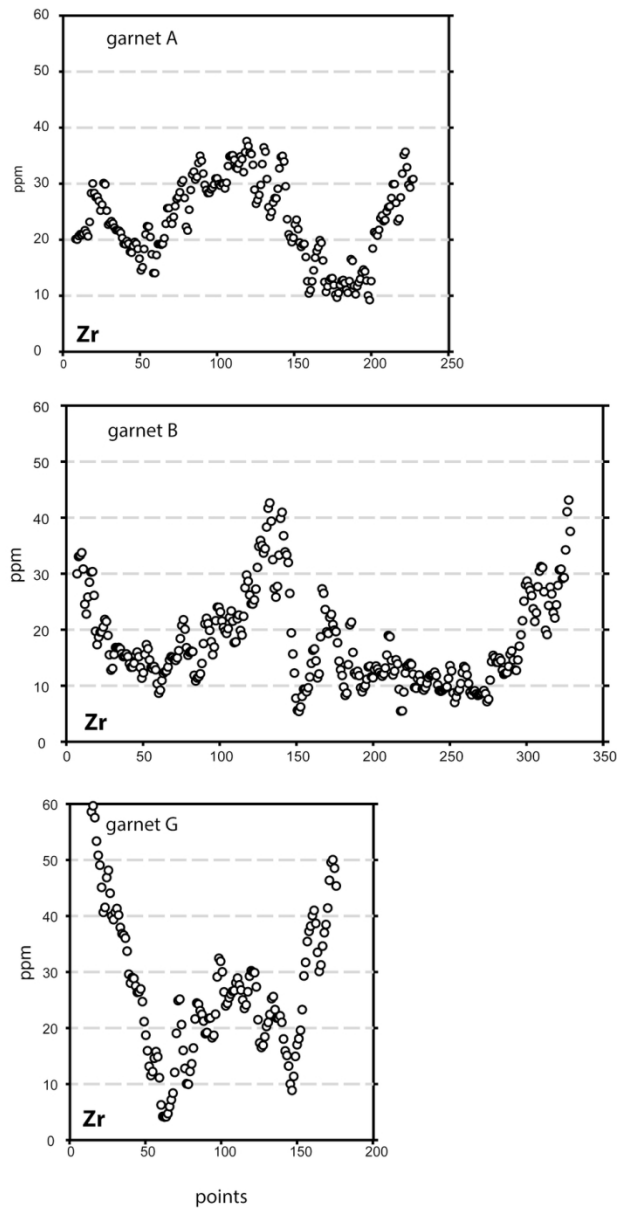


Figure S4. LA-ICP-MS Zr line profile of garnet A, B and G

80x156mm (300 x 300 DPI)

Table 1

	P-T pseudosection Figure 14	P- $\mu$ diagrams in Figure 15	P-X diagram in Figure 16		P-T pseudosection in Figure 17
	06SB06 – qtz-saturated granulite	melt produced @ 0.8 GPa and 790°C	X = 0	X = 1	selvage composition after H <sub>2</sub> O and SiO <sub>2</sub> transfer (X=0.70)
	wt%	wt%	wt%	wt%	wt%
SiO <sub>2</sub>	64.22	69.12	64.22	36.25	48.21
Al <sub>2</sub> O <sub>3</sub>	22.93	13.94	22.93	39.84	32.61
FeO	5.76	0.42	5.76	10	8.19
MgO	2.89	0.15	2.89	5.02	4.11
CaO	1.25	0.74	1.25	2.17	1.77
K <sub>2</sub> O	1.86	6.51	1.86	3.23	2.64
Na <sub>2</sub> O	0.64	2.28	0.64	1.11	0.91
H <sub>2</sub> O	0.46	6.85	0.46	2.38	1.56
	mol	mol	mol	mol	mol
SiO <sub>2</sub>	1.053	1.565	1.053	0.342	0.556
Al <sub>2</sub> O <sub>3</sub>	0.222	0.186	0.222	0.222	0.222
FeO	0.079	0.008	0.079	0.079	0.079
MgO	0.071	0.005	0.071	0.071	0.071
CaO	0.022	0.018	0.022	0.022	0.022
K <sub>2</sub> O	0.019	0.094	0.019	0.019	0.019
Na <sub>2</sub> O	0.010	0.050	0.010	0.010	0.010
H <sub>2</sub> O	0.025	0.517	0.025	0.075	0.060

Table S1

solid solution models	abbreviation in Perple_X	abbreviation used	references
Biotite	Bio(TCC)	Bio	Tajcmanova et al. (2009)
Orthopyroxene	Opx(HP)	Opx	Powell & Holland (1999)
Melt	melt(HP)	L	White et al. (2001)
K-feldspar	San	Kf	Waldbaum & Thompson (1968)
Plagioclase	Pl(h)	pl	Newton et al. (1981)
Garnet	Gt(WPH)	Gt	White et al. (2007)
Spinel	Sp(HP)	Sp	
Sapphirine	Sapp(KWP)	Spr	Kelsey et al. (2014)
White mica	Mica(CHA)	Mu	Auzanneau et al. (2009)
Cordierite	hCrd	Crd	Holland et al. (1996)

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White, R. W., Powell, R., & Holland, T. J. B. (2007). Progress relating to calculation of partial melting equilibria for metapelites. *Journal of Metamorphic Geology*, 25(5), 511-527. <https://doi.org/10.1111/j.1525-1314.2007.00711.x>





profil gamet D

Table with 34 columns (Pt 1-34, Dist 0) and rows for elements: SiO2, Al2O3, FeO, MgO, MnO, CaO, Total, Si, Al, Fe, Mg, Mn, Ca, MgRat, Alm, Pyr, Sps, Grs.

profil gamet E

Table with 39 columns (Pt 1-39, Dist 0) and rows for elements: SiO2, Al2O3, FeO, MgO, MnO, CaO, Total, Si, Al, Fe, Mg, Mn, Ca, MgRat, Alm, Pyr, Sps, Grs.

profil gamet F

Table with 35 columns (Pt 1-35, Dist 0) and rows for elements: SiO2, Al2O3, FeO, MgO, MnO, CaO, Total, Si, Al, Fe, Mg, Mn, Ca, MgRat, Alm, Pyr, Sps, Grs.











71	72	73	74	75	76	77	78	79	80
2047	2067	2088	2108	2129	2149	2170	2190	2211	2231
38.56	39.54	39.40	38.82	39.29	38.72	39.64	39.70	39.25	39.40
22.12	22.06	21.94	21.98	21.89	22.28	22.35	21.88	22.00	21.98
26.08	26.51	26.21	26.18	26.48	26.70	26.56	26.35	26.59	26.31
10.38	10.54	10.64	10.73	10.81	10.77	10.80	10.79	10.87	10.94
0.50	0.58	0.59	0.56	0.57	0.50	0.48	0.55	0.61	0.60
1.80	1.69	1.63	1.48	1.42	1.36	1.23	1.10	1.16	1.14
99.44	100.93	100.41	99.75	100.46	100.32	101.07	100.37	100.47	100.36
2.989	2.998	3.000	2.978	2.993	2.958	2.995	3.018	2.989	2.998
2.007	1.971	1.969	1.987	1.965	2.006	1.990	1.960	1.974	1.971
1.679	1.681	1.669	1.679	1.687	1.706	1.678	1.675	1.694	1.674
1.192	1.191	1.208	1.227	1.227	1.226	1.216	1.222	1.233	1.240
0.033	0.038	0.038	0.036	0.037	0.032	0.031	0.035	0.039	0.038
0.148	0.138	0.133	0.121	0.116	0.111	0.1	0.09	0.095	0.093
0.42	0.42	0.42	0.42	0.42	0.42	0.42	0.42	0.42	0.43
0.55	0.55	0.55	0.55	0.55	0.56	0.56	0.55	0.55	0.55
0.39	0.39	0.40	0.40	0.40	0.40	0.40	0.40	0.40	0.41
0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
0.05	0.05	0.04	0.04	0.04	0.04	0.03	0.03	0.03	0.03

83	84	85	86	87	88	89	90	91	92	93
2214	2239	2265	2290	2315	2340	2365	2391	2416	2441	2466
39.91	39.06	39.25	39.25	39.78	38.97	39.58	39.71	39.53	39.00	39.67
22.15	22.34	22.25	22.11	22.33	22.08	22.24	22.13	22.44	22.18	22.44
24.90	25.10	24.45	25.33	25.26	25.46	25.53	25.33	25.41	25.52	26.02
10.37	10.73	10.63	10.75	10.81	11.10	11.01	11.12	11.07	11.26	11.48
0.46	0.53	0.49	0.43	0.54	0.48	0.54	0.51	0.46	0.57	0.48
3.23	3.08	2.83	2.59	2.43	2.22	2.08	1.75	1.65	1.39	1.23
101.02	100.83	99.91	100.46	101.16	100.32	100.98	100.57	100.55	99.93	101.31
3.010	2.960	2.989	2.982	2.996	2.967	2.989	3.004	2.990	2.975	2.983
1.969	1.996	1.997	1.980	1.982	1.982	1.979	1.973	2.000	1.994	1.989
1.571	1.591	1.557	1.610	1.591	1.621	1.612	1.603	1.607	1.628	1.636
1.166	1.212	1.207	1.217	1.214	1.260	1.239	1.254	1.248	1.281	1.286
0.029	0.034	0.031	0.028	0.034	0.031	0.035	0.033	0.03	0.037	0.03
0.261	0.25	0.231	0.211	0.196	0.181	0.168	0.142	0.134	0.113	0.099
0.43	0.43	0.44	0.43	0.43	0.44	0.44	0.44	0.44	0.44	0.44
0.52	0.52	0.52	0.53	0.52	0.52	0.53	0.53	0.53	0.53	0.54
0.39	0.39	0.40	0.40	0.40	0.41	0.41	0.41	0.41	0.42	0.42
0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
0.09	0.08	0.08	0.07	0.07	0.06	0.06	0.05	0.04	0.04	0.03



76 1884	77 1904	78 1924	79 1943	80 1963	81 1983	82 2003	83 2023	84 2042	85 2062
38.96	39.43	38.72	38.81	39.49	39.12	39.34	35.88	39.26	39.11
22.22	22.06	22.04	22.28	22.40	22.17	22.34	22.28	22.26	22.03
26.50	26.81	26.97	26.73	27.05	26.94	26.74	26.60	27.10	26.88
10.30	10.42	10.43	10.31	10.37	10.51	10.50	10.23	10.50	10.42
0.51	0.55	0.57	0.49	0.52	0.50	0.57	0.45	0.53	0.49
1.90	1.88	1.75	1.67	1.64	1.56	1.51	1.06	1.29	1.26
100.39	101.15	100.48	100.29	101.47	100.80	100.99	96.50	100.93	100.20
2.974	2.989	2.962	2.968	2.983	2.976	2.982	2.867	2.982	2.990
1.999	1.971	1.987	2.008	1.994	1.988	1.996	2.098	1.992	1.985
1.692	1.700	1.726	1.709	1.709	1.714	1.695	1.778	1.721	1.719
1.172	1.178	1.190	1.175	1.167	1.192	1.187	1.219	1.189	1.188
0.033	0.035	0.037	0.032	0.033	0.032	0.037	0.03	0.034	0.032
0.155	0.153	0.144	0.137	0.132	0.127	0.123	0.091	0.105	0.103
0.41	0.41	0.41	0.41	0.41	0.41	0.41	0.41	0.41	0.41
0.55	0.55	0.56	0.56	0.56	0.56	0.56	0.57	0.57	0.57
0.38	0.38	0.38	0.39	0.38	0.39	0.39	0.39	0.39	0.39
0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
0.05	0.05	0.05	0.05	0.04	0.04	0.04	0.03	0.03	0.03

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127	128
3561	3586

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38.84	38.54
21.80	21.87
29.21	29.49
8.75	8.48
0.57	0.62
1.33	1.27
100.50	100.27

2.995	2.985
1.962	1.997
1.884	1.910
1.006	0.979
0.037	0.041
0.11	0.105

0.35	0.34
0.62	0.63
0.33	0.32
0.01	0.01
0.04	0.04

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## APPENDIX S1 : Trace element LA-ICP-MS analytical procedure

LA-ICP-MS trace element maps were acquired using a Resonetics M-50-LR 193 nm excimer laser coupled to an Agilent 7700x Quadrupole ICP-MS housed at Adelaide Microscopy, University of Adelaide. Instrument conditions and mapping protocols similar to that employed in this study are outlined by (Raimondo et al., 2017). Mapping was performed on standard 30  $\mu\text{m}$ -thick polished thin sections by ablating a series of parallel rasters across the sample surface to form a square or rectangular grid. A beam diameter of 16  $\mu\text{m}$  coupled with a laser repetition rate of 10 Hz produced an energy density of  $\sim 3.5 \text{ J/cm}^2$  at the target, and a scan speed of 22  $\mu\text{m/s}$  and line spacing of 16  $\mu\text{m}$  were used.

Data acquisition was performed in time-resolved analysis mode as a single continuous experiment. Each analysis comprised a suite of 39 elements (Figures S4-S6 and Datasets S1 and S2); dwell times for major elements were 0.005 s (to reduce excessive count rates) and all other masses were 0.008 s, giving a total sweep time of 0.31 s. Standards were analysed in duplicate at the start and end of each mapping run, and included reference glasses NIST 610 and 612 (Jochum et al., 2011; Pearce et al., 1997) and garnet standard MON-GT (C. Harris & Vogeli, 2010; Chris Harris et al., 2000). Data reduction was performed using the software *lolyte* (Hellstrom et al., 2008; Paton et al., 2011; Woodhead et al., 2007), with quantification via the Trace Elements DRS applied using  $^{29}\text{Si}$  for internal calibration, applicable to garnet pixels only. Finally, image and data processing was completed using *XMapTools* 2.3.4 (Lanari et al., 2014; Raimondo et al., 2017) and *ImageJ* (Abramoff, Magalhães & Ram, 2004).

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## Appendix S2 : U-Th-Pb monazite and zircon geochronology analytical procedures

The identification, distribution and characterization of chemical zoning of monazite was performed in-situ, following the approach of Williams, Jercinovic, Goncalves & Mahan (2006), on a Cameca SX100<sup>®</sup> electron microprobe at the Laboratoire Magmas et Volcans in Clermont-Ferrand (France). All monazite grains were identified through full-thin-section X-ray mapping for Ce L $\alpha$ , P K $\alpha$  along with Al K $\alpha$ , Mg K $\alpha$  to obtain a mineralogically and texturally map in order to reveal the monazite distribution with respect to the leucosome, quartz saturated and undersaturated domains (Figure 18). Analytical conditions were a 15 keV accelerating voltage with a 200 nA beam current, a defocused beam at 35  $\mu$ m, and a counting time of 9 ms per pixel with a 35  $\mu$ m pixel setp size. High resolution compositional maps were performed on all the selected grains for Y L $\alpha$ , Ca K $\alpha$ , Th M $\alpha$  and U MB. Analytical conditions were 15 kV, 200 Na with a rastering focussed beam. Counting time varies from 100 to 150 ms, with a step size ranging from 0.4 to 1  $\mu$ m. Intensities of all raw compositional were normalized in counts per second and processed all together so that gray levels (i.e. composition) and zoning features can be compared from grain to grain. Zircon were identified via a conventional optical microscope.

U-Th-Pb geochronology of zircon and monazite was conducted by laser ablation inductively coupled plasma spectrometry (LA-ICPMS) at the Laboratoire Magmas et Volcans, Clermont-Ferrand (France). The analyses involve the ablation of minerals with a Resonetics Resolution M-50 powered by an ultra short pulse ATL Atlex Excimer laser system operating at a wavelength of 193 nm (detailed description in Müller, Shelley, Miller & Broude (2009)). Spot diameters of 20-26  $\mu$ m and 7  $\mu$ m associated to repetition rates of 3 Hz and 1 Hz were used for zircon and monazite respectively. The ablated material is carried into helium, and then mixed with nitrogen and argon, before injection into a plasma source of an Agilent 7500 cs ICP-MS equipped with a dual pumping system to enhance the sensitivity.

The alignment of the instrument and mass calibration was performed before every analytical session using the NIST SRM 612 reference glass, by inspecting the signal of <sup>238</sup>U and by minimising the ThO<sup>+</sup>/Th<sup>+</sup> ratio (<< 1%). The mean sensitivity on <sup>238</sup>U at the instrumental conditions reported in Table 1 and using a spot size of 44  $\mu$ m is about 15-20,000 cps/ppm. The analytical method for isotope dating of monazite with laser ablation ICPMS is basically similar to that developed for zircon and monazite and reported in Paquette & Tiepolo (2007) and Tiepolo, Bottazzi, Palenzona & Vannucci (2003). The signal of <sup>204</sup>(Pb+Hg), <sup>206</sup>Pb, <sup>207</sup>Pb, <sup>208</sup>Pb, <sup>232</sup>Th and <sup>238</sup>U

masses are acquired. The occurrence of common Pb in the sample can be monitored by the evolution of the  $^{204}\text{Pb}+\text{Hg}$  signal intensity, but no common Pb correction was applied owing to the large isobaric interference from Hg. The  $^{235}\text{U}$  signal is calculated from  $^{238}\text{U}$  on the basis of the ratio  $^{238}\text{U}/^{235}\text{U}= 137.88$ . Single analyses consisted of 20 seconds of background integration with laser off followed by 1 minute integration with the laser firing and a 20 seconds delay to wash out the previous sample (approximately 10 seconds for 6 orders of magnitude) and prepare the next analysis.

Data are corrected for U-Pb and Th-Pb fractionation occurring during laser sampling and for instrumental mass discrimination (mass bias) by standard bracketing with repeated measurements of GJ-1 zircon (Jackson, Pearson, Griffin & Belousova, 2004) or Moacyr monazite (Gasquet, Bertrand, Paquette, Lehmann & Ratzov, 2010) standard. Data reduction was carried out with the software package GLITTER® (Macquarie Research Ltd, 2001; Jackson et al. (2004); Van Achterbergh, Ryan, Jackson & Griffin (2001). For each analysis, the time resolved signal of single isotopes and isotope ratios was monitored and carefully inspected to verify the presence of perturbations related to inclusions, fractures, mixing of different age domains or common Pb. Calculated ratios were exported and Concordia ages and diagrams were generated using Isoplot/Ex v. 2.49 software package by Ludwig (2001). The concentrations in U-Th-Pb were calibrated relative to the certified contents of 91500 zircon standard (Wiedenbeck, Allé, Corfu, Griffin & Meier, 1995) and Moacyr monazite (Seydoux-Guillaume, Wirth, Deutsch & Schärer, 2004).

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