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1 **Timescales of ultra-high temperature metamorphism and crustal differentiation: zircon**  
2 **petrochronology from granulite xenoliths of the Variscan French Massif Central**

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13

14 **Abstract**

15

16 Lower crustal ultra-high temperature (UHT) metamorphism and the resulting production and  
17 transfer of granitic magmas represent key processes of intracrustal differentiation. The  
18 timescales of these phenomena are debated, due to the complexity of the granulite zircon U-  
19 Pb record and because direct links between granulites and granites are difficult to establish.  
20 To address these issues, we present the results of zircon petrochronology (coupled U-Pb  
21 dating, trace element and Lu-Hf isotopic analyses) for lower crustal felsic and mafic granulite  
22 xenoliths from the Variscan eastern French Massif Central and compare them with data from  
23 well-characterized mid-/upper crustal migmatites and granites. The felsic and mafic granulites  
24 represent pre-Variscan meta-sedimentary and meta-igneous mafic rocks, respectively; which  
25 experienced Variscan UHT peak metamorphism at 940–970 °C and  $8 \pm 2$  kbar. Zircons from  
26 the felsic granulites show U-Pb dates spreading over ~50 Myr between ~315 and ~265 Ma,  
27 correlated to Ti-in-zircon temperatures decreasing from 940–970 °C to 800 °C and REE  
28 contents consistent with growth in equilibrium with garnet at (U)HT conditions. This is best  
29 explained by continuous crystallization of zircon upon cooling from the thermal peak owing to  
30 decreasing Zr solubility of residual melt, despite significant prograde melt loss. Zircons from  
31 the mafic granulite only record the last stage of this time-temperature evolution (295–265 Ma;

32 800–900 °C) due to later zircon saturation. Upper crustal granite emplacement started at 340  
33 Ma and culminated at the age of the lower crustal thermal peak of  $313 \pm 3$  Ma (defined by the  
34 highest-temperature zircons from felsic granulites), reflecting melt extraction along the ~27 Myr  
35 prograde path of the lower crust. In turn, the crystallization ages of mid-crustal migmatites  
36 (315–300 Ma) and the lower crustal granulites (315–265 Ma) are consistent with slow cooling  
37 in the presence of melt. These results provide a direct assessment of the timescales of melt  
38 production and residence at the crustal scale; and validate the granulite-granite connection in  
39 the framework of the melt loss theory in migmatitic systems.

40

41 **Keywords:** UHT metamorphism; partial melting; crustal differentiation; granulite; zircon;  
42 Variscan orogeny

43

## 44 1. Introduction

45

46 The formation and differentiation of the Earth's continental crust is a two-step process.  
47 New crust produced mainly in magmatic arcs (e.g. Collins et al., 2020; Jagoutz and Klein,  
48 2018) is reworked during orogenic events through tectonic, metamorphic and/or magmatic  
49 processes to produce the continental crust with bulk andesitic composition and layered  
50 structure (Cawood et al., 2013; Hacker et al., 2015; Rudnick and Gao, 2014). In particular,  
51 (ultra-)high-temperature [(U)HT], granulite-facies metamorphism and partial melting of the  
52 lower crust, followed by the segregation and ascent of the resulting granitic magmas, is a key  
53 process of intracrustal differentiation (Harley, 2016; Moyen et al., 2017a; Sawyer et al., 2011;  
54 Zheng and Gao, 2021). Mid- to upper crustal granitic plutons and lower crustal granulites would  
55 represent the two complementary products of this process i.e., the extracted magma and  
56 residual solid respectively (Brown, 2013; Cawood et al., 2013; Clemens et al., 1990, 2020;  
57 Vielzeuf et al., 1990; White and Powell, 2002). This concept has been validated by phase  
58 equilibria modelling (White et al., 2001; Yakymchuk and Brown, 2014a), experimental  
59 petrology (Gao et al., 2016) and the investigation of melt inclusions in metamorphic minerals  
60 (Cesare et al., 2009).

61 Melt segregation and extraction from the lower crust can theoretically happen at the melt  
62 connectivity threshold of 5 to 7 vol.% (Rabinowicz and Vigneresse, 2004; Rosenberg and  
63 Handy, 2005). Therefore, melt loss is expected to occur progressively as it is produced along

64 the prograde P–T path of the lower crust (Brown, 2013; Kelsey and Hand, 2015; White and  
65 Powell, 2002; Yakymchuk and Brown, 2014a). Because melt segregation and transport are  
66 geologically fast (<100 kyr and down to <1 kyr in duration; Harris et al. 2000; Petford et al.,  
67 2000; Scaillet et al., 2006), granite formation should pre-date or at least be synchronous with  
68 the lower crustal (U)HT metamorphic peak. However, this hypothesis has not been  
69 systematically tested yet, mainly because direct links between granulites and granites are  
70 difficult to establish as these represent two different paleo-crustal levels.

71 Moreover, even when a field connection between granulites and granites is made possible  
72 by tectonic reworking or the presence of lower crustal xenoliths, the comparison is entangled  
73 by the complex geochronological record of granulites. Zircon U-Pb dates in granulites  
74 commonly span several 10s to 100s of Myr (Kelly and Harley, 2005; Kelsey and Hand, 2015;  
75 Harley, 2016; Harley and Black, 1987; Taylor et al., 2016 and references therein).  
76 Interpretations of this protracted zircon record vary from pulsed mineral growth in response to  
77 metamorphic reactions along different parts of the P–T path (Baldwin et al., 2007; Hermann  
78 and Rubatto, 2003; Rubatto et al., 2013; Vavra et al., 1996); continuous crystallization through  
79 melt Zr saturation upon cooling (Guergouz et al., 2018; Kelsey et al., 2008; Taylor et al.,  
80 2015a); igneous crystallization in the case of mafic granulites (Costa and Rey, 1995; Moyen  
81 et al., 2017b); or even post-crystallization disturbance of the U-Pb system (Taylor et al., 2014;  
82 Kunz et al., 2018).

83 In this context, better understanding the processes and timescales of crustal differentiation  
84 requires to (i) robustly assess the significance of zircon U-Pb dates in lower crustal granulites;  
85 and (ii) directly compare these with the age and origin of co-genetic, mid-/upper crustal granite  
86 plutons. For this purpose, we present here the results of zircon petrochronology (coupled U-  
87 Pb dating, trace element quantification and Lu-Hf isotopic analyses) from lower crustal felsic  
88 and mafic granulite xenoliths of the Variscan eastern French Massif Central (E-FMC). These  
89 are interpreted to record (U)HT metamorphism, partial melting and magmatic underplating  
90 related to the late stages of the Variscan orogeny (Downes and Leyreloup, 1986; Downes et  
91 al., 1990; Costa and Rey, 1995; Leyreloup et al., 1977), whose mid- to upper crustal expression  
92 is well constrained from the study of exposed migmatites and granites (Barbey et al., 1999;  
93 Laurent et al., 2017; Montel et al., 1992; Moyen et al., 2017a; Villaros et al., 2018, Williamson  
94 et al., 1997). They provide therefore an ideal case study to investigate the nature of the

95 parameters controlling U-Pb age distributions in granulites; and how such distribution relates  
96 to melt production in, and extraction from the lower crust to eventually differentiate the crust.

97

## 98 **2. Crustal history and structure of the eastern French Massif Central**

99

100 The E-FMC is part of the Variscan Belt of Western Europe. The E-FMC crust consists  
101 of Ediacaran (605–555 Ma) sedimentary rocks (now paragneisses) deposited in a back-arc  
102 basin along the northern Gondwana margin which were intruded by igneous rocks (now  
103 orthogneisses) during back-arc inversion in the late Ediacaran–Cambrian (545–525 Ma) and  
104 subsequent Ordovician (485–450 Ma) extension (Chelle-Michou et al., 2017; Couzinié et al.,  
105 2017, 2019). These lithologies were reworked in the Devonian–Carboniferous through  
106 metamorphism and magmatism, associated with convergence and collision between  
107 Laurussia and Gondwana (see review in Vanderhaeghe et al., 2020). Cenozoic rifting was  
108 associated with abundant alkaline volcanism that peaked in Miocene–Pliocene times (Michon  
109 and Merle, 2001).

110 This study focuses on the area of the Velay migmatite-granite dome in the E-FMC (Fig.  
111 1a). The local metamorphic rocks belong to several nappes (Vanderhaeghe et al., 2020): (i)  
112 the allochthonous Upper Gneiss Unit, typified by a bimodal meta-igneous association at its base  
113 (the so-called “Leptynite-Amphibolite complex”) and a Devonian HP-HT metamorphic  
114 evolution; (ii) the Lower Gneiss Unit, composed of upper amphibolite-facies, migmatitic ortho-  
115 and paragneisses devoid of HP record; (iii) the Paurautochthonous Unit, comprising lower-grade  
116 metasediments. Migmatitic rocks from the Lower Gneiss Unit define a broad structural dome  
117 ~100 km in diameter (Ledru et al., 2001), decorated by 340–310 Ma granite massifs/dykes and  
118 cored by the 310–300 Ma Velay diatexite and associated granites (Montel et al., 1992; Chelle-  
119 Michou et al., 2017; Didier et al., 2013; Laurent et al., 2017) (Fig. 1a). Petrological and  
120 geochemical data demonstrate that most 340–310 Ma granites were sourced in the lower crust  
121 (Downes et al., 1997; Villaros et al., 2018). In contrast, the 310–300 Ma Velay diatexite and  
122 granites formed from both the accumulation of lower crustal melts/magmas and anatectic melts  
123 formed nearly *in situ* at mid-crustal depths, largely at the expense of orthogneisses (Montel et  
124 al., 1992; Williamson et al., 1997; Barbey et al., 1999; Chelle-Michou et al., 2017; Villaros et  
125 al., 2018; Couzinié et al., 2021).

126           The studied xenolith suite originates from the Bournac site (Fig. 1a) where they are  
127 hosted in a phreatomagmatic breccia pipe of Upper Miocene age (Couthures et al., 1991). The  
128 basaltic breccia contains numerous enclaves of amphibolite- to granulite-facies crustal rocks  
129 (Leyreloup et al., 1977). The granulite-facies xenoliths reached peak metamorphic conditions  
130 of up to 6–15 kbar and 800–1000°C (Fig. 1b, Dostal et al., 1980; Downes and Leyreloup, 1986;  
131 Leyreloup, 1992) and were thus interpreted as sampling the lower crust (Leyreloup, 1992).  
132 Among these, two distinct lithologies were identified by previous studies: (i) mafic granulites,  
133 of igneous origin (47%); (ii) felsic granulites corresponding to meta-sedimentary (42%) or  
134 meta-igneous rocks (11%) (Costa and Rey, 1995; Dostal et al., 1980; Downes et al., 1990;  
135 Downes and Leyreloup, 1986; Leyreloup, 1992; Leyreloup et al., 1977).

136           Whole-rock Nd model ages of the granulite xenoliths range from 1.0 to 1.8 Ga (Downes  
137 and Leyreloup, 1986; Downes et al., 1990), similar to those of exposed Variscan rocks in the  
138 FMC (Downes et al., 1997). Published geochronology includes thermal ionization mass  
139 spectrometry (TIMS) U-Pb dating of monazite and zircon that yielded a dominant 320–280 Ma  
140 range in both meta-sedimentary and mafic meta-igneous xenoliths, interpreted as reflecting  
141 granulite-facies metamorphism and coeval crystallization of underplated basaltic magmas,  
142 respectively (Costa and Rey, 1995; Downes et al., 1990). Further secondary ion mass  
143 spectrometry (SIMS) zircon U-Pb dating in a meta-sedimentary granulite revealed two  
144 populations of Neoproterozoic to early Paleozoic, detrital grains (630–430 Ma) and  
145 Carboniferous to Permian (360–250 Ma) metamorphic grains (Rossi et al., 2006).

146           The Bournac xenoliths would sample the lower crust formed during the Variscan  
147 orogeny and underwent minimal disturbance since then (Costa and Rey, 1995; Downes and  
148 Leyreloup, 1986; Downes et al., 1990), notably as this crust segment remained less affected  
149 by Tethyan oceanization than other Variscan domains like Corsica (Rossi et al., 2006) or the  
150 Ivrea Zone (Guergouz et al., 2018). The felsic granulites would represent lower crustal  
151 residues after extraction of magmas that rose and accumulated in mid-crustal migmatite units  
152 and upper crustal, peraluminous granite bodies (Fig. 1b) (Vanderhaeghe et al., 2020; Villaros  
153 et al., 2018). The mafic meta-igneous rocks are generally interpreted as under/intraplated  
154 mantle-derived magmas (Costa and Rey, 1995; Downes et al., 1990) or alternatively as a  
155 constituent lithology of the local pre-Variscan crust (Dostal et al., 1980; Downes and Leyreloup,  
156 1986).

157

158 **3. Materials and methods**

159

160 We have investigated two meta-sedimentary, felsic granulites (HS15-BOU01 and  
161 HS15-BOU17) and one meta-igneous, mafic granulite (HS15-BOU07). These were selected  
162 on petrological and geochemical grounds as to be best representative of the two dominant  
163 lithological groups identified among the lower crustal xenoliths throughout the French Massif  
164 Central (see sections 2 and 4.1). We have worked on zircon separates instead of directly on  
165 thin sections as to provide a statistically meaningful dataset for each sample. For this purpose,  
166 we targeted samples among the largest xenoliths (spheroids 10–15 cm in diameter), which  
167 allowed to analyze 80 to 142 grains per sample (in comparison, thin sections contain about 10  
168 useable grains). After sawing out the outermost 2–3 cm-thick edge of the samples, affected by  
169 interactions with the host (Leyreloup, 1992), the innermost part was selected for thin section  
170 preparation and  $\geq 500$  g powdered down to  $<80$   $\mu\text{m}$  for bulk rock elemental analysis. The  
171 remaining material was crushed and sieved to  $<500$   $\mu\text{m}$  for heavy mineral separation. Zircon  
172 grains were concentrated using standard techniques (panning, magnetic and heavy liquids  
173 separation, handpicking), cast in epoxy resin and polished down to a sub-equatorial section.

174 Bulk rock major and trace element compositions were determined at the Service  
175 d'Analyse des Roches et des Minéraux (CNRS–CRPG, Nancy, France). Major element  
176 compositions of minerals were measured in thin sections by electron microprobe analysis  
177 using a CAMECA SX100 instrument at Service Microsonde Sud (Université Montpellier 2,  
178 France). Thin section cathodoluminescence imaging was performed using a “cold-cathode”  
179 Technosyn 8200 Mk2 at CEREGE (Marseille, France) and a “hot-cathode” Lumic HC6-LM at  
180 GET (Toulouse, France). Zircon cathodoluminescence imaging was performed at the  
181 Laboratoire Magmas et Volcans (Clermont-Ferrand, France) using a Jeol JSM-5910 scanning  
182 electron microscope. Simultaneous zircon U-Pb dating and trace element analyses were  
183 carried out at ETH Zürich (ETHZ), Switzerland, by laser ablation – inductively coupled plasma  
184 – mass spectrometry (LA-ICP-MS) using a RESOLUTION S-155 (ASI, Australia) 193 nm (ArF)  
185 excimer laser system coupled to an Element XR (Thermo Scientific, Germany) sector-field  
186 ICP-MS. Analyses of Lu-Hf isotopes in zircon were performed by laser ablation – multi-  
187 collection – inductively coupled plasma – mass spectrometry (LA-MC-ICP-MS) at ETHZ using  
188 the same laser ablation system as for U-Pb dating and trace elements, attached to a Nu  
189 Plasma 2 (Nu Instruments, UK) MC-ICP-MS. Garnet and rutile trace element data were

190 obtained by LA-ICP-MS at the Service ICP-MS of Observatoire Midi-Pyrénées in Toulouse  
191 (France), using a NWRfemto (ESI, USA) solid-state laser ablation system coupled to an  
192 Element XR sector-field ICP-MS. Phase equilibrium modelling was undertaken using the  
193 software Perple\_X (version 6.9.1, Nov. 2021, Connolly, 2009) in the system Na<sub>2</sub>O–CaO–K<sub>2</sub>O–  
194 FeO–MgO–Al<sub>2</sub>O<sub>3</sub>–SiO<sub>2</sub>–H<sub>2</sub>O–TiO<sub>2</sub>(–O<sub>2</sub>), i.e. NCKFMASHT(O).

195 Details about the analytical and modelling procedures are provided in the  
196 Supplementary Material. The data on samples and reference materials are provided in the  
197 Data Tables.

198

## 199 **4. Results**

200

### 201 *4.1. Petrography and geochemistry*

202

203 Felsic granulites HS15-BOU-01 and HS15-BOU-17 (Fig. 2a,b) are composed of a  
204 seriate interlobate to polygonal assemblage of quartz (15–25 vol.%), plagioclase (An<sub>30-38</sub>Ab<sub>52-</sub>  
205 <sub>60</sub>Or<sub>9-16</sub>; 20–28 vol.%), slightly perthitic alkali feldspar (An<sub>1-5</sub>Ab<sub>24-33</sub>Or<sub>63-74</sub>; 25–32 vol.%),  
206 chemically unzoned garnet (alm<sub>51-56</sub>pyr<sub>39-45</sub>grs<sub>4</sub>sp<sub>0-1</sub>; 22–27 vol.%) (Supplementary Figures  
207 S1 and S2) and prismatic sillimanite, which occurs associated with plagioclase in HS15-BOU-  
208 17 (making up 5–6 vol.% of the rock) and only as inclusions within garnet in HS15-BOU-01.  
209 Both hand specimens exhibit mm-scale banding with alternating plagioclase vs. alkali  
210 feldspar–quartz layers (Fig. 2). Accessory minerals are rutile, apatite, and zircon. Zircon is  
211 mainly located at grain boundaries and rarely as inclusion in garnet (Fig. 2). Whole-rock  
212 compositions match those of meta-sedimentary granulite xenoliths from the E-FMC (Fig. 3a,b).  
213 The felsic granulites are selectively enriched in Fe, Mg, Ti and depleted in alkalis at a given  
214 SiO<sub>2</sub> content with respect to lower-grade metasediments of the local crust (Fig. 3a,b), which  
215 likely results from melt loss during anatexis (Leyreloup et al., 1977). The same hypothesis was  
216 proposed to explain similar chemical differences between granulite-facies (“stronalites”, to  
217 which our samples are akin, Fig. 3) and amphibolite-facies (“kinzigites”) meta-sediments from  
218 the deep crustal section of the Ivrea-Verbano Zone, NW Italy (Redler et al., 2013; Schnetger,  
219 1994). From a trace element perspective, the felsic granulites are chemically complementary  
220 to peraluminous granites from the E-FMC (Fig. 3c).

221 Mafic granulite sample HS15-BOU-07 consists of a granoblastic assemblage of  
222 plagioclase (~60 vol.%), orthopyroxene (~22 vol.%), clinopyroxene (~14 vol.%), ilmenite (~2  
223 vol.%) with accessory apatite and zircon, the latter being either included in major minerals  
224 (most often plagioclase) or located along grain boundaries. Both ortho- and clinopyroxene  
225 grains appear corroded by a microcrystalline, dark-coloured (in plane-polarized light), Fe-rich  
226 but alkali-free alteration rim, possibly representing a former melt film that infiltrated the xenolith.  
227 Whole-rock geochemical data indicate that the xenolith has a quartz-normative, basic (51.7  
228 wt.% SiO<sub>2</sub>), Al-rich (17.9 wt.% Al<sub>2</sub>O<sub>3</sub>) and K<sub>2</sub>O-poor (0.3 wt.%) composition, lacks typical  
229 features of igneous cumulates (e.g. positive Eu anomaly) and shows an affinity to calc-alkaline  
230 suites (Fig. 3c). Such composition is consistent with that of a high-Al basaltic liquid (Crawford  
231 et al., 1987). These characteristics make HS15-BOU-07 representative of the most common  
232 group (~70%) of mafic granulites in the E-FMC (Dostal et al., 1980; Downes et al., 1990).

233 The lower crustal origin of the three xenoliths is supported by their high  $V_p$  values of  
234 6.5–6.9 km.s<sup>-1</sup> (calculation procedure in the Supplementary Material), matching those of the  
235 lowermost crust (6.5–6.8 km.s<sup>-1</sup> at 20–28 km depth) in the E-FMC based on seismic data  
236 (Perrier and Ruegg, 1973).

237

#### 238 *4.2. Thermobarometry and phase equilibrium modelling*

239

240 The results of pseudosection modelling indicate that the dry (biotite-free), mineralogy  
241 of felsic granulite samples entails temperatures >840°C, while the coexistence of garnet,  
242 sillimanite and rutile constrains pressures between 6.5 and 10–13 kbar (Fig. 1d and  
243 Supplementary Figure S3). It is worth noting that solidus temperatures are about 200 °C higher  
244 than the wet solidus of fertile meta-sedimentary rocks. This observation and the preservation  
245 of dry mineral assemblages imply that the rocks have experienced significant melt loss (Redler  
246 et al., 2013; White and Powell, 2002), which is in line with bulk-rock geochemistry (see section  
247 4.1 and Fig. 3). Conventional thermobarometry yields equilibration conditions of 825 ± 29°C at  
248 7.8 ± 0.5 kbar and 833 ± 11°C at 6.9 ± 0.7 kbar for HS15-BOU01 and HS15-BOU17,  
249 respectively (calculated with the two-feldspar thermometer of Putirka, 2008, and the GASP  
250 barometer of Holdaway, 2001, quoted “internal” uncertainties correspond to 2 standard  
251 deviations [SD]). The pressure estimated for HS15-BOU01 must be regarded as a minimum

252 as sillimanite is only present as inclusions in garnet, hence possibly not in equilibrium with  
253 matrix plagioclase. These conditions fall on the solidus for both samples.

254 Based on the calculated pseudosection, the absence of garnet and amphibole in mafic  
255 granulite HS15-BOU07 indicates temperatures exceeding 820°C for maximum pressures of  
256 8.5–10 kbar (Fig. 1d). Quartz is predicted in low modal amounts (<4 vol.%) at such conditions  
257 but was not observed in thin section. The two-pyroxene thermobarometer of Putirka (2008)  
258 yielded temperature–pressure conditions of  $904 \pm 20^\circ\text{C}$  and  $9.4 \pm 1.4$  kbar which although  
259 affected by large uncertainties are consistent with pseudosection analysis.

260

#### 261 4.3. Zircon data

262

263 In all three samples, zircon forms transparent, yellowish to colorless crystals ranging in  
264 size from 50 to 350  $\mu\text{m}$ . In felsic granulites, they generally show rounded, nearly equant  
265 shapes, with the exception of a few elongated grains (Fig. 4a,b). Crystals are more irregularly  
266 shaped to anhedral in mafic granulite HS15-BOU07 (Fig. 4c). Two main zircon domains are  
267 identified based on CL images (Fig. 4): (i) entire grains or, more frequently, rounded cores  
268 showing well-developed zoning, most commonly oscillatory in felsic granulites and patchy in  
269 the mafic granulite; and (ii) entire grains or rims mantling the former, generally showing  
270 brighter, homogeneous luminescence and no to weak zoning. The cores show characteristics  
271 of resorbed zircon with internal zoning clearly truncated by irregular boundaries, which are  
272 either sharp or consist of a dark-CL seam, generally <10  $\mu\text{m}$ - but up to 50  $\mu\text{m}$ -thick (Fig. 4).  
273 Wherever visible, internal structures in the CL-bright, homogeneous rims and grains consist of  
274 sector, fir-tree and/or radial zoning, imparting a multifaceted (“soccerball”) texture (Fig. 4). The  
275 volume ratio between homogeneous rims/grains and zoned cores/grains varies from a grain  
276 to another but, at the scale of the sample, homogeneous rims/grains dominate.

277 The results of LA-(MC-)ICP-MS analyses reveal clear differences between zoned  
278 cores/grains and homogeneous rims/grains, in terms of U-Pb dates, Hf isotopic ratios and trace  
279 element composition. In all samples, zoned cores/grains show a spread of mainly (sub-)  
280 concordant U-Pb dates (Fig. 5). In mafic granulite HS15-BOU07, most analyses (7 out of 11)  
281 range from Ordovician to Carboniferous ( $481 \pm 9$  to  $307 \pm 7$  Ma) with a few younger spots and  
282 define a trend of decreasing  $\epsilon\text{Hf}(t)$  with time from +5.0 at  $\sim 480$  Ma to -2.3 at  $\sim 180$  Ma (Fig. 6a).  
283 Zoned cores/grains in felsic granulites show dates mainly from Neoproterozoic to Permian (807

284  $\pm 25$  to  $249 \pm 8$  Ma) (Fig. 5) and a wide range of  $\epsilon\text{Hf}_{(t)}$  between +5 and -16 (Fig. 6a), with few  
285 outliers at very negative values (down to -58). In contrast, homogeneous rims/grains in all  
286 samples show more clustered, Carboniferous to Permian U-Pb dates between 320 and 270  
287 Ma (Fig. 5) with an apparent mode at 280–290 Ma. The  $\epsilon\text{Hf}_{(t)}$  values are also more  
288 homogeneous in a given sample than results from zoned cores/grains (Fig. 6a), with negative  
289 values in felsic granulites from  $-10.0 \pm 3.3$  (2 S.D.;  $n = 55$ ) in HS15-BOU17 to  $-5.7 \pm 3.0$  (2  
290 S.D.;  $n = 43$ ) for HS15-BOU01; and positive values of  $+4.1 \pm 1.2$  (2 S.D.;  $n = 54$ ) in mafic  
291 granulite HS15-BOU07.

292 While most zoned cores/grains show Ti contents between 10 and 35 ppm regardless  
293 their U-Pb date (Fig. 6b), homogeneous rims/grains have on average higher Ti (20–70 ppm).  
294 In felsic granulite samples, homogeneous rims/grains show a clear correlation between Ti  
295 content and U-Pb date, with >300 Ma spots having >50 ppm Ti, decreasing to 20–30 ppm at  
296 260–280 Ma (Fig. 6b). Data from the mafic granulite do not show such correlation, yet overlap  
297 with the trend defined by homogeneous rims/grains in felsic granulites, at 270–300 Ma and  
298 25–40 ppm Ti (Fig. 6b). In all samples, zoned cores/grains show more scattered and generally  
299 higher concentrations of most trace elements (Nb, Y, HREE, U, Th) than homogeneous  
300 rims/grains (Fig. 7). In felsic granulites the latter are further characterized by flat HREE profiles  
301 ( $\text{Gd}_N/\text{Yb}_N \sim 1$ ) and pronounced negative Eu anomalies ( $\text{Eu}_N/\text{Eu}^* < 0.1$ ) compared with zoned  
302 cores/grains ( $\text{Gd}_N/\text{Yb}_N < 0.2$  and  $\text{Eu}_N/\text{Eu}^* \sim 0.05\text{--}1$ ) (Fig. 7), as well as less pronounced positive  
303 Ce anomalies (Supplementary Figure S7).

304 In all samples, single homogeneous rims/grains analyzed multiple times generally show  
305 identical U-Pb dates and Hf isotopic ratios within uncertainties (Fig. 4), as well as only modest  
306 variations in trace element contents. Whenever analyzable, dark seams separating cores and  
307 rims show U-Pb dates and elemental/isotopic compositions that are similar to those of the rims  
308 (see examples from HS15-BOU07 in Fig. 4) and are therefore considered together with  
309 homogeneous rims/grains in the following.

310

## 311 **5. Discussion**

312

### 313 *5.1. Zircon petrochronology of the Bournac granulites*

314

#### 315 *5.1.1. Felsic granulites*

316           The dichotomy between zoned cores/grains vs. homogeneous grains/rims is typical of  
317 zircon from (U)HT felsic granulites and reflects inheritance from the protolith vs. metamorphic  
318 growth (Kelly and Harley, 2005; Kunz et al., 2018; Taylor et al., 2016; Vavra et al., 1996). The  
319 range of inherited U-Pb ages is consistent with a former detrital zircon population. This range  
320 is skewed towards younger (mainly Paleozoic) ages (Fig. 5) compared with the detrital zircon  
321 age distribution in exposed meta-sedimentary rocks in the E-FMC, showing a Neoproterozoic  
322 (550–800 Ma) peak (Chelle-Michou et al., 2017; Couzinié et al., 2019), certainly reflecting Pb  
323 loss during the Variscan (U)HT event (see Supplementary Figure S8). The crystallization of  
324 homogeneous rims/grains during Variscan metamorphism is confirmed by U-Pb dates  
325 overlapping with the Carboniferous to early Permian age of this event in the E-FMC (review in  
326 Vanderhaeghe et al., 2020) and their distinctive trace element signature, in line with that of  
327 metamorphic zircon from (U)HT granulites (Fig. 7) (e.g. Kelly and Harley, 2005; Kunz et al.,  
328 2018; Rubatto et al., 2013; Taylor et al., 2015a, 2016, 2017).

329           Trace element data indicate that the homogeneous zircon grains/rims grew at supra-  
330 solidus, and up to UHT conditions. The observed low  $\text{Eu}_N/\text{Eu}^*$  ratios (Fig. 7) generally reflects  
331 zircon growth in the presence of K-feldspar produced by biotite dehydration melting reactions  
332 (Rubatto et al., 2013). The  $>1$  order of magnitude lower Nb values of homogeneous grains/rims  
333 relative to zoned cores/grains (Fig. 7) is explained by the presence of rutile throughout zircon  
334 crystallization, as this mineral controls most, if not all of the Nb budget of the rocks (bulk  $\text{TiO}_2$   
335 and Nb contents of ~1 wt.% and 13–18 ppm respectively are balanced by 1 wt.% rutile at  
336 1400–1800 ppm Nb; see Data Table 8). This, together with quartz saturation (see  
337 Supplementary Figure S3), indicates that the obtained Ti-in-zircon temperatures of 800–970  
338 °C (Fig. 6) are robust (Ferry and Watson, 2007). Lastly, the high  $\text{Gd}_N/\text{Yb}_N$  and low Y–HREE  
339 contents (Fig. 7) argue for zircon crystallization in the presence of garnet, which is confirmed  
340 by an evaluation of zircon/garnet REE partitioning. As garnet shows homogeneous M–HREE  
341 concentrations (Fig. 8a), apparent partition coefficients were calculated for each zircon  
342 analysis relative to the median garnet composition. These show a good match to the  
343 experimentally determined zircon/garnet partition coefficients at 900–1000 °C and 7 kbar (Fig.  
344 8b), pointing to zircon growth at UHT conditions in equilibrium with garnet (Taylor et al., 2017).

345           Nevertheless, K-feldspar, rutile and garnet are stable over most of the supra-solidus  
346 P–T space (see Fig. 9 and Supplementary Figure S3), which does not elucidate along which  
347 portion of the P–T path the homogeneous zircon grains/rims were formed. Prograde zircon

348 growth has been previously proposed to form at the onset of partial melting (Hermann and  
349 Rubatto, 2003; Rubatto et al., 2013) and/or in response to the breakdown of Zr-bearing  
350 minerals (Baldwin et al., 2007; Vavra et al., 1996). However, none of these processes appears  
351 likely in our case. The minimum Ti-in-zircon temperatures (ca. 800 °C) exceed the prograde  
352 solidus by >100 °C (Fig. 9) and trace element data point to zircon growth in equilibrium with,  
353 not at the expense of other Zr-bearing minerals (rutile and garnet). Most importantly,  
354 homogeneous grains/rimms show a clear trend of decreasing Ti contents with U-Pb dates,  
355 entailing continuous zircon crystallization during progressive cooling of the system, from ~940–  
356 970 °C to ~800 °C (Fig. 6). This trend is best explained by zircon formation through decreasing  
357 Zr solubility of the silicate melt along the retrograde path (Kelsey et al., 2008; Mintrone et al.,  
358 2020; Taylor et al., 2015a; Yakymchuk and Brown, 2014). This interpretation is further  
359 supported by (i) a rough increase of zircon/garnet HREE partition coefficients towards younger  
360 ages (Fig. 8c), consistent with a progressive temperature decrease (Taylor et al., 2017); and  
361 (ii) limited  $\epsilon\text{Hf}_{(t)}$  scatter (~5  $\epsilon\text{Hf}$ -units) compared to the cores (>20  $\epsilon\text{Hf}$ -units) and no variation  
362 of  $\epsilon\text{Hf}_{(t)}$  with age (Fig. 6), arguing for continuous crystallization from a relatively homogeneous  
363 melt phase.

364 The correlations between U-Pb dates and key zircon trace element indicators make the  
365 ~50 Myr spread of U-Pb dates shown by homogeneous grains/rimms between 315 and 265 Ma  
366 (Figs. 5, 6) indicative of a truly protracted period of zircon crystallization. Analytical artifacts,  
367 such as ablation craters mixing cores and rims, can be ruled out as multiple analyses in single  
368 domains show identical dates (Fig. 4) and mixing trends between cores and rims are absent  
369 from the data (Figs. 6, 7). The low  $\alpha$ -doses and Ti-in-zircon temperatures below their specific  
370 closure temperatures (Supplementary Figures S8 and S9) make post-crystallization  
371 disturbance of the U-Pb system equally unlikely. Therefore, we collectively interpret the zircon  
372 trace element and U-Pb data to reflect continuous, supra-solidus cooling from the thermal  
373 peak at UHT conditions (940–970 °C, highest-temperature zircons; Fig. 6) down to the solidus  
374 (~800 °C; lowest-temperature zircons).

375

### 376 5.1.2. *Mafic granulite*

377 The dominant population of dates from homogeneous rims/grains at  $284.7 \pm 2.7$  Ma in  
378 sample HS15-BOU07 (Fig. 5 and Supplementary Figure S6) could be interpreted as reflecting  
379 either crystallization of a mafic magma or metamorphism of a pre-existing crustal rock, as is

380 generally the case in mafic granulites (e.g. Moyen et al., 2017b). In fact, the Bournac mafic  
381 granulites were originally interpreted as metamorphosed, pre-Variscan mafic rocks (Dostal et  
382 al., 1980) but subsequent zircon U-Pb dating led most authors to conclude that they represent  
383 underplated, Permian mafic magmas (Costa and Rey, 1995; Féménias et al., 2003, Rossi et  
384 al., 2006). The latter interpretation would be apparently consistent with widespread Permian  
385 magmatism recorded throughout the Variscan Belt (Boscaini et al., 2020), notably the ~283  
386 Ma Mafic Complex of the Ivrea-Verbano Zone in NW Italy (Karakas et al., 2019). However, the  
387 trace element composition of homogeneous rims/grains in HS15-BOU07 is distinct from that  
388 of typical igneous zircon and plots off the magma differentiation trend defined by the Ivrea-  
389 Verbano zircons (Fig. 7). In particular, Permian zircons in HS15-BOU07 are richer in Ti (25–  
390 40 ppm) than magmatic zircons from the Mafic Complex of the Ivrea-Verbano Zone (5–30  
391 ppm) (Fig. 6), pointing to different environments and/or conditions of zircon formation.

392 Furthermore, if the rock derives from a Permian mafic magma, the zoned zircon cores  
393 with older U-Pb dates and distinct trace element compositions must be interpreted as  
394 incompletely dissolved xenocrysts. Such xenocrysts would show a random sampling of  
395 compositionally diverse zircons from the country rocks, comparable to zoned cores/grains from  
396 the felsic granulites. In contrast, the zircon cores from HS15-BOU07 define a single trend of  
397 closed-system  $^{176}\text{Hf}/^{177}\text{Hf}$  radiogenic ingrowth (Fig. 6a) and a tight compositional cluster in  
398 terms of trace element concentrations (Fig. 7). Hence, these cores likely represent a coherent,  
399 single zircon population crystallized at the same time, whose present-day range in U-Pb dates  
400 only results from variable extent of Pb loss (Laurent and Zeh, 2015).

401 Therefore, we interpret zoned cores from HS15-BOU07 as reflecting former igneous  
402 zircon crystallized from the protolith, whereas the homogeneous rims/grains correspond to  
403 Variscan metamorphic zircon. The Ordovician age of the oldest zircon core ( $481 \pm 8$  Ma)  
404 thereby places a minimum age for the mafic protolith. Following Dostal et al. (1980), the mafic  
405 granulite would thus correspond to a pre-Variscan, meta-igneous rock that underwent the  
406 same UHT metamorphic and partial melting history as the meta-sedimentary granulites. The  
407 homogeneous grains/rims in HS15-BOU07 likely formed through the same process as in the  
408 felsic granulites, i.e. crystallization from a cooling melt. This is supported by their  
409 homogeneous trace element compositions and  $\epsilon\text{Hf}$  (Figs. 6a, 7) and overlap with  
410 homogeneous zircon rims/grains from felsic granulites in the Ti vs. U-Pb date plot (Fig. 6b).  
411 Although the mafic granulite contains neither quartz, nor rutile, the Ti-in-zircon temperatures

412 are probably not significantly different from those obtained from the felsic granulites at given  
413 Ti content (using  $a_{\text{SiO}_2} = 0.9$  and  $a_{\text{TiO}_2} = 0.7$  would shift temperature up by 25 °C, which roughly  
414 corresponds to the uncertainty). In this case, the shorter range of U-Pb dates in the mafic  
415 granulite may reflect the lower Zr content of the sample (124 ppm) compared to felsic granulites  
416 (222–234 ppm), which would have the effect of delaying zircon saturation along the cooling  
417 path (Kelsey et al., 2008; Mintrone et al., 2020).

418

## 419 *5.2. Melt production and loss during Variscan UHT metamorphism*

420

421 Our new data demonstrate that the investigated granulites represent pre-Variscan  
422 metasedimentary and meta-igneous mafic rocks metamorphosed at UHT conditions during the  
423 late Variscan orogeny. Mineralogy (Fig. 2) and whole-rock compositions (Fig. 3) indicate that  
424 the rocks must have lost melt to preserve anhydrous mineral assemblages, which is generally  
425 considered to happen along the prograde branch of the P–T path (Brown, 2013; Kelsey and  
426 Hand, 2015; White and Powell, 2002; Yakymchuk and Brown, 2014a). However, the  
427 homogeneous zircon rims/grains in all samples must have formed through melt saturation  
428 upon cooling from the thermal peak to the solidus (see section 5.1), meaning that at least some  
429 melt was present along the retrograde path.

430 To estimate the amount of melt required to explain the crystallization of homogeneous  
431 zircon rims/grains, we have applied the melt-o-meter method of Mintrone et al. (2020). This  
432 method uses the Zr mass balance between the bulk sample composition, the relative  
433 proportion of zircon newly crystallized from melt and the theoretical Zr solubility of such melt  
434 at inferred peak temperatures (from the model of Boehnke et al., 2013), i.e. 940–970 °C here  
435 (see Supplementary Material for details). The method does not intend to precisely quantify the  
436 melt fraction but only give a first-order estimate to be compared to theoretical melt  
437 productivities at peak conditions. The calculated fractions of newly precipitated zircon (FPZ)  
438 are 67 and 83 wt.% for HS15-BOU01 and HS15-BOU17 respectively, which corresponds to  
439 11–13 wt.% (ca. 15 vol.%) melt. For the mafic granulite, the calculated FPZ is 97 wt.%, which  
440 yields ca. 7 wt.% (ca. 10 vol.%) melt at 940–970 °C. For the two felsic granulite samples,  
441 pseudosections calculated with bulk compositions adjusted to fit the predicted peak melt  
442 volume fractions (details on the melt reintegration procedure are in the Supplementary  
443 Material) predict a near-solidus (810–840 °C) field characterized by decreasing garnet mode,

444 breaking down to biotite (Fig. 9). This reaction is not observed in the samples and would  
445 contradict zircon/garnet REE partitioning (Fig. 8), which indicates garnet stability throughout  
446 the range of zircon crystallization temperatures. Therefore, retrograde melt loss must have  
447 taken place at temperatures only slightly above 840 °C, preventing hydration of the peak  
448 mineral assemblage yet allowing for zircon crystallization from the melt phase upon cooling.

449 The calculated peak melt proportions (10–15 vol.%) are lower than theoretical  
450 maximum melt productivities at 5–10 kbar and 940–970 °C (30–60 vol.% and 20–40 vol.% for  
451 meta-sedimentary vs. mafic rocks, respectively; Palin et al., 2016; Webb et al., 2015 and  
452 references therein). This confirms that the E-FMC granulites have also lost substantial  
453 amounts of melt along the prograde path. Melt reintegration calculations indicate that up to 18  
454 to 24 vol.% melt would have been lost along the prograde path for HS15-BOU01 and HS15-  
455 BOU17, respectively. This also explains the preservation of protolith zircon despite reaching  
456 UHT conditions, as prograde melt loss tends to shift up-temperature the dissolved zircon  
457 fraction isopleths (Kelsey et al., 2008; Yakymchuk and Brown, 2014b). Melt-reintegrated  
458 compositions also highlight the prograde growth of garnet as a peritectic phase (Fig. 9). Zircon  
459 thus begins to form at peak temperature in the presence of both garnet and residual melt,  
460 explaining the well-equilibrated zircon/garnet REE patterns (Fig. 8).

461 In summary, we propose that the Variscan P–T metamorphic history of the Bournac  
462 granulites involves: (i) prograde heating and partial melting from the solidus up to 940–970 °C  
463 (at  $8 \pm 2$  kbar) with concomitant dissolution of protolith zircon, together with garnet growth and  
464 melt drainage reaching a cumulative loss of minimum 18–24 vol.% melt in the felsic granulites;  
465 followed by (ii) cooling with <15 vol.% residual melt and zircon crystallization, with garnet  
466 remaining stable owing to at least one last melt drainage event close to ca. 840 °C in the felsic  
467 granulites.

468

### 469 *5.3. Timescales of melt production, extraction and crystallization in the crust*

470

471 Data from our granulite xenoliths record prograde melt production and loss, followed  
472 by cooling associated with zircon crystallization and limited further melt drainage. Although  
473 other lower crustal regions might have recorded more complex histories of melt residence and  
474 transfer, the clear timing of zircon crystallization relative to the simple sequence recorded in  
475 our samples allows to temporally benchmark the thermal evolution of the lower crust and

476 intracrustal differentiation (Fig. 10). Indeed, based on these interpretations, the oldest  
477 homogeneous zircon rims/grains analyses from meta-sedimentary granulites should provide a  
478 minimum age for the Variscan lower crustal thermal peak in the E-FMC. The ages obtained  
479 from analyses statistically defining single populations (and comprising the spots with highest  
480 Ti-in-zircon temperatures) are  $313.5 \pm 3.6$  Ma (MSWD = 1.8; n = 9) for HS15-BOU01 and  $312.8$   
481  $\pm 3.1$  Ma (MSWD = 1.5; n = 27) for HS15-BOU17 (Fig. 10a).

482 The co-genetic nature of peraluminous granites from the EFMC with the investigated  
483 felsic granulites has been established by phase equilibria modelling (Villaros et al., 2018),  
484 which is confirmed by the overlapping zircon  $\epsilon\text{Hf}_{(t)}$  between granites and granulites (Fig. 6a).  
485 The latter show on average more negative  $\epsilon\text{Hf}_{(t)}$ , but this is expected whenever zircon remains  
486 a residual phase (Laurent and Zeh, 2015). Based on the theory of melt loss in migmatites, the  
487 granites should have formed prior to or synchronously with the lower crustal thermal peak  
488 (Rosenberg and Handy, 2005; White and Powell, 2002; Yakymchuk and Brown, 2014a). This  
489 is confirmed by the distribution of zircon and monazite U-Pb ages from granites within an area  
490 of ~50 km radius centered on the Bournac pipe (Fig. 10b). The number of U-Pb dates indeed  
491 increases progressively from 340 to 310 Ma, reflecting increasing melt loss along the lower  
492 crustal prograde path, with a maximum at 320-310 Ma that strikingly coincides with the inferred  
493  $313 \pm 3$  Ma age for the thermal peak (Fig. 10b). This is clear evidence for direct genetic and  
494 temporal links between lower crustal granulites and upper crustal granites within the framework  
495 of the melt loss theory.

496 In contrast, the mid-crustal Velay diatexites and granites show a highest density of U-  
497 Pb ages at 310-300 Ma, interpreted as reflecting their crystallization some 5–10 Ma after the  
498 lower crustal thermal peak (Fig. 10b). Numerical models considering the fundamental role  
499 played by melt transfer on the thermal structure of the orogenic crust predict that peak  
500 temperature conditions should be attained synchronously in the lower and the middle crust  
501 (Depine et al., 2008). Following this, it can be inferred that the Velay magmas (of mixed lower  
502 and middle crustal origin) were formed/accumulated at the latest by  $313 \pm 3$  Ma. Their delayed  
503 crystallization would thus indicate the prevalence of a partially molten mid-crustal layer for  
504 several Myr (Vanderhaeghe et al., 2020), in contrast to the faster cooling and crystallization of  
505 granitic plutons at higher structural levels. Additionally, melts extracted along the lower crust  
506 retrograde path possibly fed the mid-crustal partially molten layer. These melts could also have

507 formed the small-volume silicic volcanic rocks/tuffs exposed throughout the EFMC in early  
508 Permian (299–283 Ma), post-orogenic sedimentary basins (Michel et al., 2015).

509         The time lag between the crystallization of the oldest upper crustal granites (ca. 340  
510 Ma) and the thermal peak (ca. 313 Ma) constrains the duration of suprasolidus heating in the  
511 lower crust (700–970 °C) to a maximum of ca. 27 Myr (Fig. 9), from which a minimum heating  
512 rate of 10 °C/Myr can be deduced. Lithospheric mantle delamination has been proposed as a  
513 heat source for Variscan (U)HT metamorphism in the eastern FMC (Laurent et al., 2017;  
514 Vanderhaeghe et al., 2020). Thermal models show that at reasonable mantle heat fluxes (<50  
515 mW m<sup>-2</sup>), the resulting abrupt temperature increase at the Moho would entail crustal melting  
516 and melt-assisted heat advection upwards in the crust for 25–30 Myr until a steady-state  
517 geotherm is reached (Depine et al., 2008), in excellent agreement with our timing constraints.  
518 Conversely, lower crustal cooling from peak down to ca. 800°C is reflected by protracted zircon  
519 crystallization between 315 and 265 Ma (Fig. 10b), corresponding to a slower cooling rate of  
520  $3.0 \pm 1.5$  °C/Myr (Fig. 6b), similar to estimates obtained from other granulite terranes (e.g.  
521 Guergouz et al., 2018; Harley and Black, 1987; Harley, 1989; Kelsey et al., 2008; Mezger et  
522 al., 1993). Such slow cooling rate is consistent with the long-lived presence of melt in the lower  
523 crust. Indeed, melt crystallization releases latent heat and promotes lateral flow of the  
524 weakened lower crust, which hinders cooling associated with tectonic exhumation  
525 (Vanderhaeghe et al., 2020).

526

## 527 **6. Conclusions and implications**

528

529         Granulite xenoliths sampled by the Miocene Bournac pipe in the eastern FMC  
530 represent pre-Variscan lower crustal rocks that underwent the same Variscan UHT  
531 metamorphic episode. Partial melting started at ~340 Ma in response to lithospheric mantle  
532 delamination and was followed by heating up to peak temperatures of 940-970 °C (at 6-10  
533 kbar) reached at  $313 \pm 3$  Ma. Significant melt loss throughout the prograde path and at peak  
534 temperature contributed to form voluminous, 340-310 Ma upper crustal peraluminous granite  
535 plutons and mid-crustal diatexites. However, the lower crust was not completely melt-depleted  
536 at peak conditions, as the cooling (hence Zr saturation) of residual melt (~15 vol.%) is required  
537 to explain zircon formation, in particular the trend of decreasing Ti content throughout the ~50  
538 Myr period of zircon crystallization in felsic granulites. Most of the melt left in the system was

539 eventually drained along this retrograde path before reaching temperatures of ~840 °C (at  
540 ~280 Ma), allowing the preservation of dry assemblages in the lower crust and explaining minor  
541 volumes of late granites/rhyolites in the upper crust.

542         These results have important implications for the interpretation of U-Pb dates in UHT  
543 granulites and the link between granulites and granites in orogenic systems:

- 544         - Although most of the melt loss from the lower crust happens along the prograde path,  
545             some UHT granulites may not be melt-free at peak conditions and non-negligible melt  
546             loss can also take place along the retrograde path, accounting for the generally slow  
547             cooling rates observed in these settings.
- 548         - In addition to the duration spent above a certain temperature (Harley, 2016), the zircon  
549             crystallization time span in these UHT granulites also depends on the amount of melt  
550             still present at peak conditions and the Zr budget of the partly molten system, as melt  
551             cooling and saturation is the dominant zircon-forming reaction in this case.
- 552         - As a corollary, sharp peak of zircon dates in mafic granulites may be mistaken as an  
553             igneous crystallization age, whereas it rather reflects late zircon saturation closer to the  
554             solidus due to lower Zr contents than in meta-sedimentary granulites.
- 555         - The relative timing of zircon crystallization in co-genetic upper crustal granites and  
556             lower crustal granulites is a key information to understand the processes by which they  
557             are linked. In particular, prograde/peak melt loss from the lower crust to form upper  
558             crustal granites, followed by zircon saturation in granulites upon retrograde cooling, is  
559             expected to decouple U-Pb crystallization ages between the upper and lower crust, as  
560             observed here.

561

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563

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572

### 573 **Figure captions**

574

575 **Figure 1.** (a) Sketch map depicting the exposed Variscan domains and inferred suture zones  
576 in W Europe. (b) Geological map of the eastern Massif Central and (c) interpretative cross-  
577 section, adapted from Vanderhaeghe et al. (2020). The crust and mantle structure along with  
578 approximate seismic velocities for the different layers are from Perrier and Ruegg (1973). (d)  
579 Summary of P–T estimates obtained for the three investigated crustal xenoliths; shaded areas  
580 and dotted fields represent the stability fields of mineral assemblages for our samples (derived  
581 from pseudosection analyses – see Supplementary Material) and for similar literature samples  
582 (compiled in Leyreloup, 1992), while the symbols represent average conditions of equilibration  
583 for our samples based on conventional thermobarometry.

584

585 **Figure 2.** Petrography of the studied granulite xenoliths. Each column shows a rock slab  
586 (scale bar is 1 cm) and optical cathodoluminescence images (scale bar is 500  $\mu\text{m}$ ) illustrating  
587 the mineral assemblages and the textural position of zircon grains. Abbreviations: Afs = alkali  
588 feldspar; Cpx = clinopyroxene; Grt = garnet; Ilm = ilmenite; Pl = plagioclase; Opx =  
589 orthopyroxene; Qz = quartz; Sil = sillimanite; Zrn = zircon.

590

591 **Figure 3.** Geochemistry of granulite xenoliths from the FMC. (a)  $\text{SiO}_2$  vs.  $\text{Al}_2\text{O}_3$  plot and (b) B–  
592 A cationic diagram of Villaseca et al. (1998). The dotted lines represent the contours  
593 encompassing 85% of analyses of greenschist-facies metasediments from the FMC (red; n =  
594 80); metabasites from the high-grade allochthonous Upper Gneiss Unit (blue; n = 309);  
595 peraluminous granites from the eastern FMC (black; n = 1533). Data sources are available in  
596 the Supplementary Material. The large crossed dots depict average Ivrea stromalite  
597 compositions (from Schnetger, 1994). The melt loss vector in (b) is deduced from phase  
598 equilibria modelling of Villaros et al. (2018). (c) Ternary La–Nb–Y diagram of Cabanis and  
599 Lecolle (1989) (for mafic rocks only); (d) multi-element diagram normalized to the bulk  
600 continental crust composition of Rudnick and Gao (2014). The pink field shows the variability  
601 of the Ivrea stromalites (Schnetger, 1994).

602

603 **Figure 4.** Representative cathodoluminescence images of zircon grains from the granulite  
604 xenoliths investigated in this study. The white and yellow dotted circles represent the position  
605 of the laser spots for U–Pb dating/trace element (TE) and Lu–Hf isotopic analyses respectively,  
606 with corresponding spot name (S1-XX, M1-XX, M2-XX),  $^{238}\text{U}/^{206}\text{Pb}$  date  $\pm 2\sigma$  uncertainty (in  
607 Ma), Ti content (in blue), and  $\varepsilon\text{Hf}_{(t)} \pm 2\sigma$  uncertainty (calculated using the  $^{238}\text{U}/^{206}\text{Pb}$  date  
608 obtained on the same domain). The value with an asterisk corresponds to the  $^{207}\text{Pb}/^{206}\text{Pb}$  date  
609 as the analysis was discordant.

610

611 **Figure 5.** Results of zircon U-Pb analyses for the Bournac granulite xenoliths shown as Tera–  
612 Wasserburg diagrams ( $^{238}\text{U}/^{206}\text{Pb}$  vs.  $^{207}\text{Pb}/^{206}\text{Pb}$ ) with error ellipses quoted at the 95%  
613 confidence level. The insets selectively represent data from homogeneous grains and rims  
614 with ellipses color-coded based on Ti contents from the same ablated volume.

615

616 **Figure 6.** Plots of zircon  $\varepsilon\text{Hf}_{(t)}$  (a) and Ti contents (b) vs.  $^{238}\text{U}/^{206}\text{Pb}$  date (considering only  
617 concordant data) for the Bournac granulite xenoliths, with a close-up focus on the Variscan  
618 period (240–340 Ma) on the left-hand side diagrams. Note the scale change at 500 Ma in the  
619 right-hand side diagrams. The Ti-in-zircon temperatures were calculated using the calibration  
620 of Ferry and Watson (2007) with  $a(\text{TiO}_2) = a(\text{SiO}_2) = 1$ . The grey and orange fields represent  
621 zircon data respectively from the lower crustal Mafic Complex of the Ivrea-Verbano Zone, NW  
622 Italy (Karakas et al., 2019; Storck et al., 2021) and granites from the E-FMC (Moyen et al.,  
623 2017b, with estimated zircon Ti contents based on magma temperatures proposed by Laurent  
624 et al., 2017). In (a) the orange arrow represents the isotopic evolution of bulk crust with a  
625  $^{176}\text{Lu}/^{177}\text{Hf}$  ratio of 0.015 (yellow); and the green arrows are Pb loss trends for zircon cores and  
626 homogeneous grains/rims from HS15-BOU07 using the average  $^{176}\text{Lu}/^{177}\text{Hf}$  of each group.  
627 Crosses labelled “2 SE” correspond to the typical  $2\sigma$  uncertainties.

628

629 **Figure 7.** Selected trace element plots comparing the compositions of zircons from the  
630 Bournac granulite xenoliths to those of zircon from various parental magmas and geodynamic  
631 environments (as compiled in Laurent et al., 2022), (U)HT migmatites and granulites (Kelly and  
632 Harley, 2005; Kunz et al., 2018; Rubatto et al., 2013; Taylor et al., 2015a) and the Ivrea Mafic

633 Complex (Karakas et al., 2019). For the latter, the arrows depicts the igneous evolution from  
634 the most mafic to the most evolved magmas.

635

636 **Figure 8.** Evaluation of zircon/garnet REE partitioning for the felsic granulites. (a) REE  
637 concentrations of zircon (Zrn) and garnet (Grt) normalized to C1 chondrite composition of  
638 McDonough and Sun (1995). (b) 'Array plot' of Taylor et al. (2017) where  $D_i^{zrn/grt}$  represents  
639 the zircon/garnet partition coefficient of element or ratio  $i$  for each zircon analysis, calculated  
640 using the median garnet composition of the sample. Stars represent experimental data from  
641 Taylor et al. (2015b) at 7 kbar and Rubatto and Hermann (2007) at 20 kbar. (c) Plot of log  
642  $D_{Yb}^{zrn/grt}$  vs.  $^{206}Pb/^{238}U$  date for each zircon analysis.

643

644 **Figure 9.** Simplified P–T pseudosections calculated for the estimated subsolidus (left panels)  
645 and peak (right panels) compositions of felsic granulites HS15-BOU01 and HS15-BOU17,  
646 contoured for garnet mode and where major mineral reactions are indicated. The colored  
647 frames depict the P-T field where the mineral assemblages of the samples are stable. The  
648 inferred P–T path involves (1.) prograde heating and partial melting with concomitant melt  
649 drainage (white dots), up to (2.) peak conditions of 940–970 °C (at  $8 \pm 2$  kbar), followed by (3.)  
650 cooling with  $\leq 15$  vol.% residual melt with at least one last melt drainage event (black dot) at  
651  $\geq 840$  °C. Further details on the melt reintegration procedure are provided in the Supplementary  
652 Material.

653

654 **Figure 10.** Temporal relationships between granulite-facies metamorphism and granite  
655 magmatism in the eastern FMC. (a) Tera-Wasserburg Concordia diagrams for the  
656 homogeneous zircon rims/grains in the Bournac felsic granulites, with calculated Concordia  
657 ages from the oldest population as estimates of the age of the thermal peak. (b) Schematic  
658 crustal column of the eastern FMC along with the histograms and density distributions of zircon  
659 and monazite U-Pb dates obtained from upper crustal granite plutons/dykes, the Velay  
660 diatexite/granites (data from Chelle-Michou et al., 2017; Couzinié et al., 2014; Didier et al.,  
661 2013; Laurent et al., 2017) and lower crustal metasedimentary granulites (this study). The  
662 maximum at 320-310 Ma in the upper crustal rocks coincides with the inferred  $313 \pm 3$  Ma age  
663 for the lower crustal thermal peak and reflect the paroxysm of melt production and extraction

664 in the orogenic crust. In contrast, the highest density of U-Pb ages at 310-300 Ma in the Velay  
665 diatexite/granites point to a delayed crystallization and is consistent with the prevalence of  
666 supra-solidus conditions in the mid-crust 5–10 Ma after the thermal peak. The scattered, even  
667 younger U–Pb dates in metasedimentary granulites is interpreted as reflecting protracted  
668 zircon crystallization in residual lower crustal melts upon slower cooling from peak temperature  
669 down to ~800°C.

670

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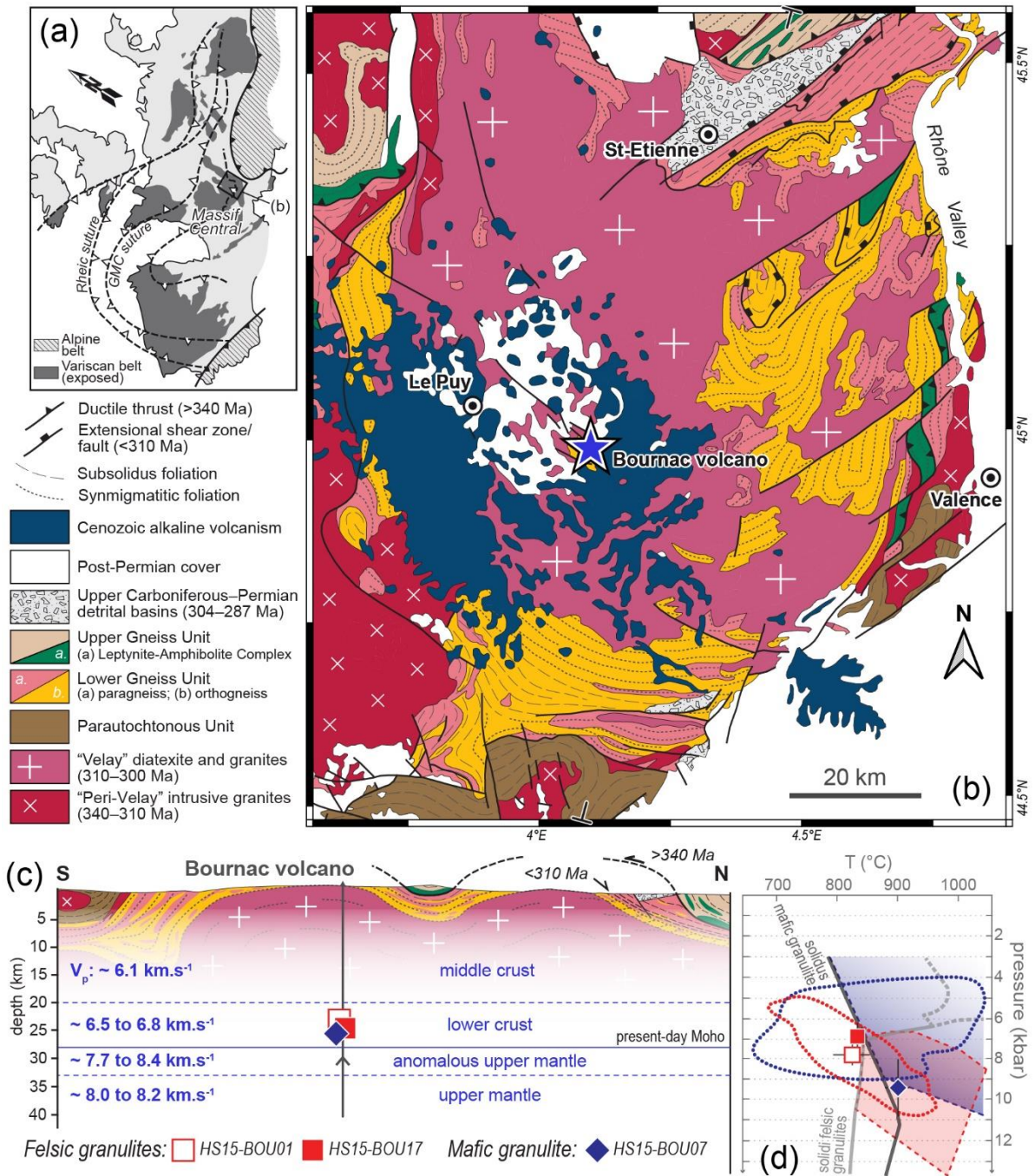


Figure 1

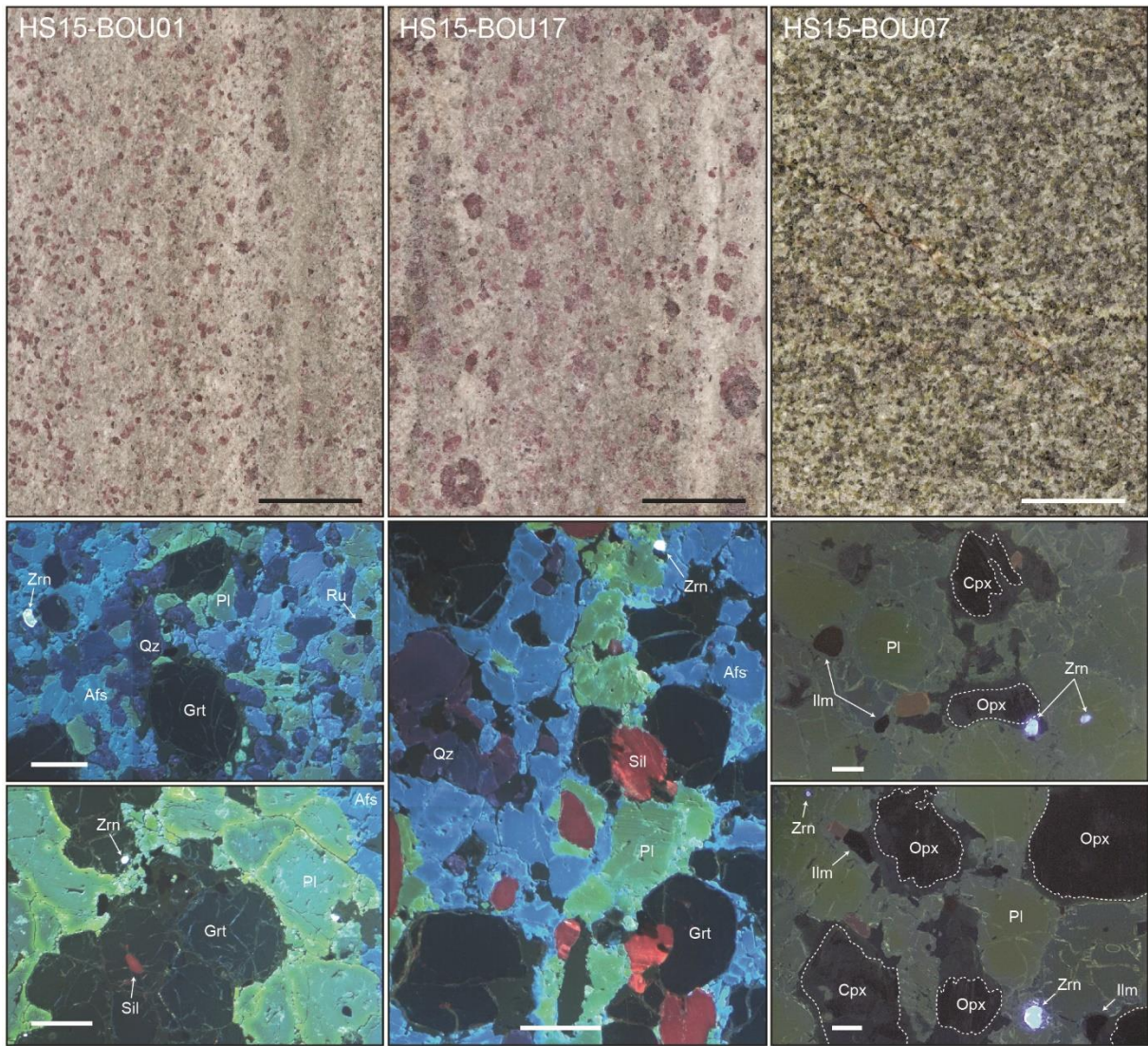


Figure 2

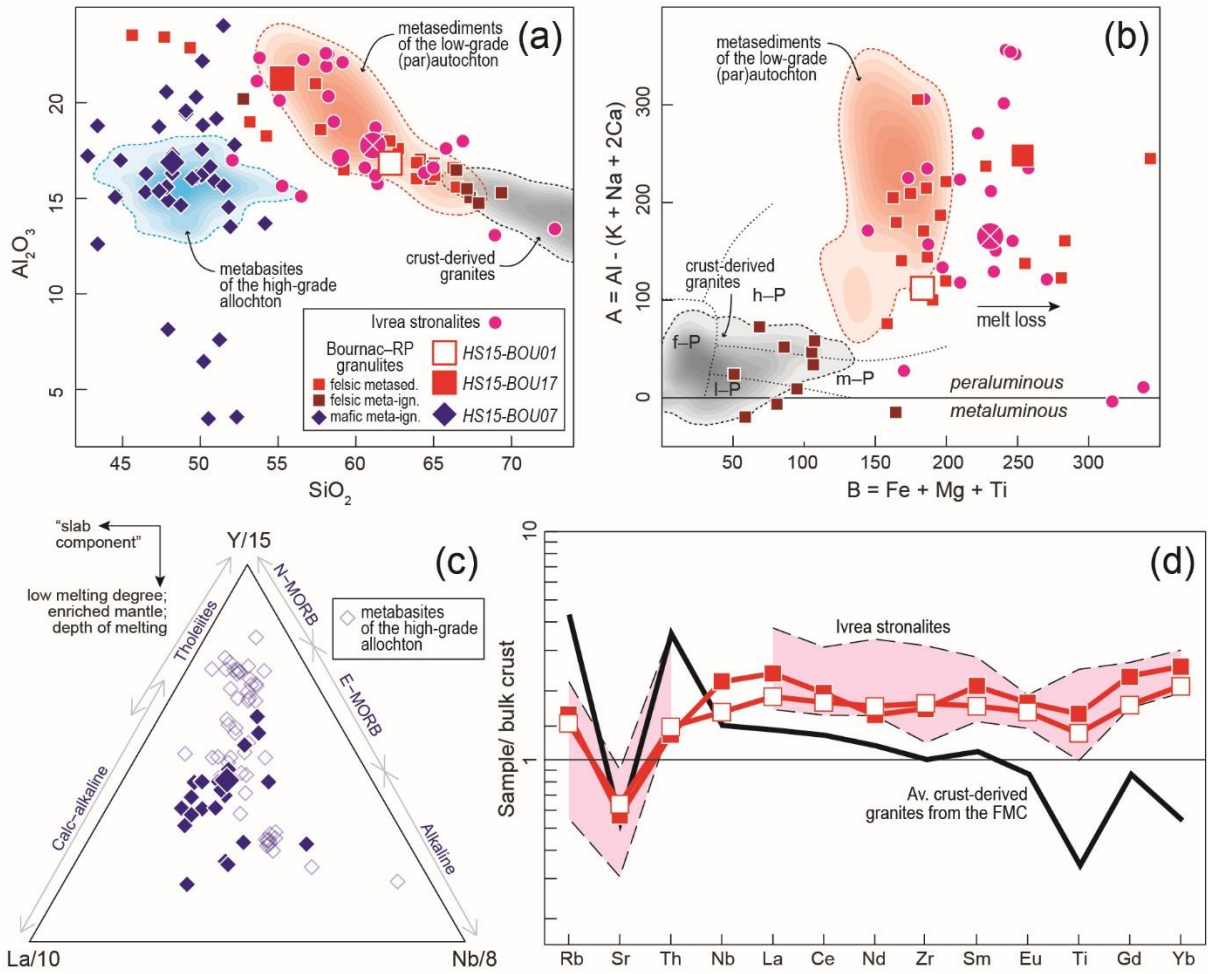


Figure 3

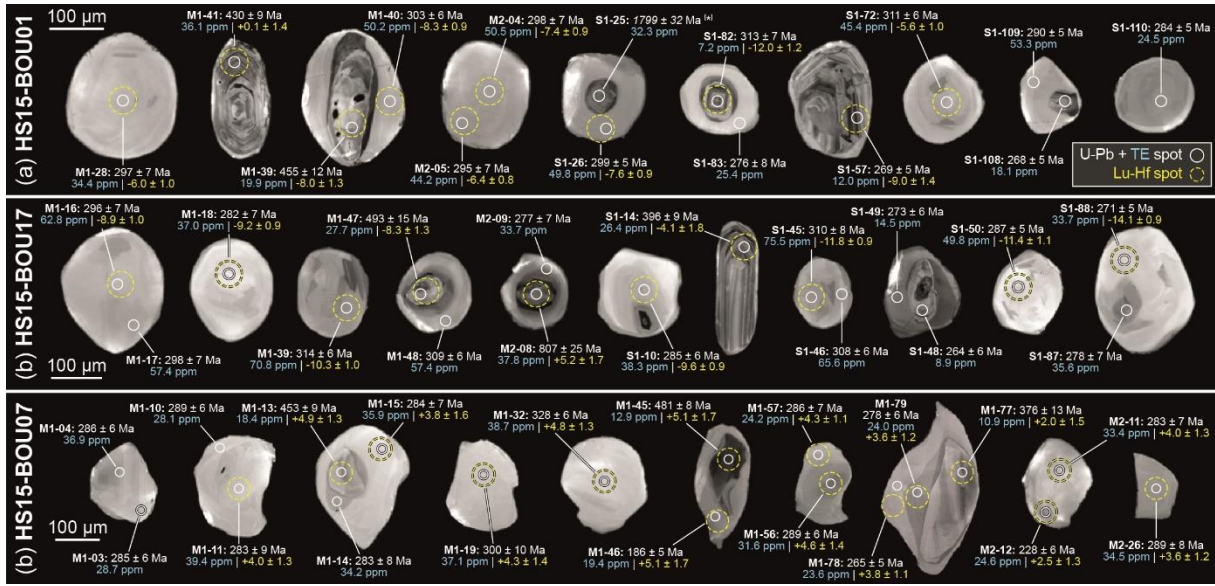


Figure 4

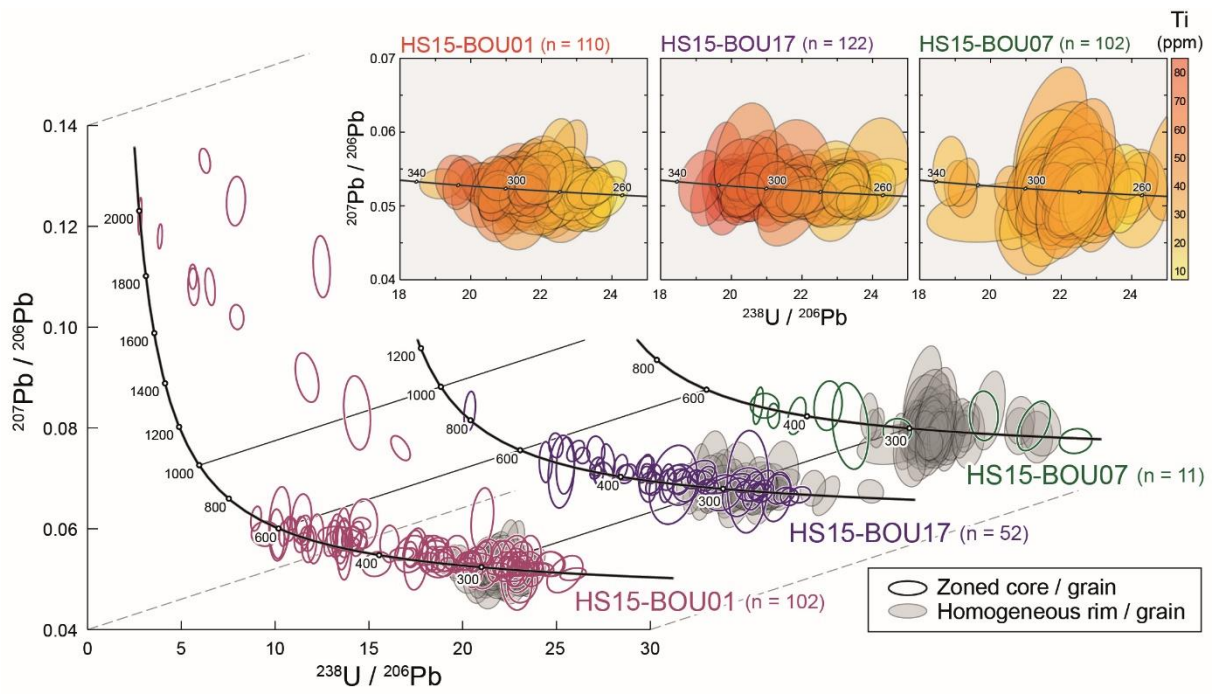


Figure 5



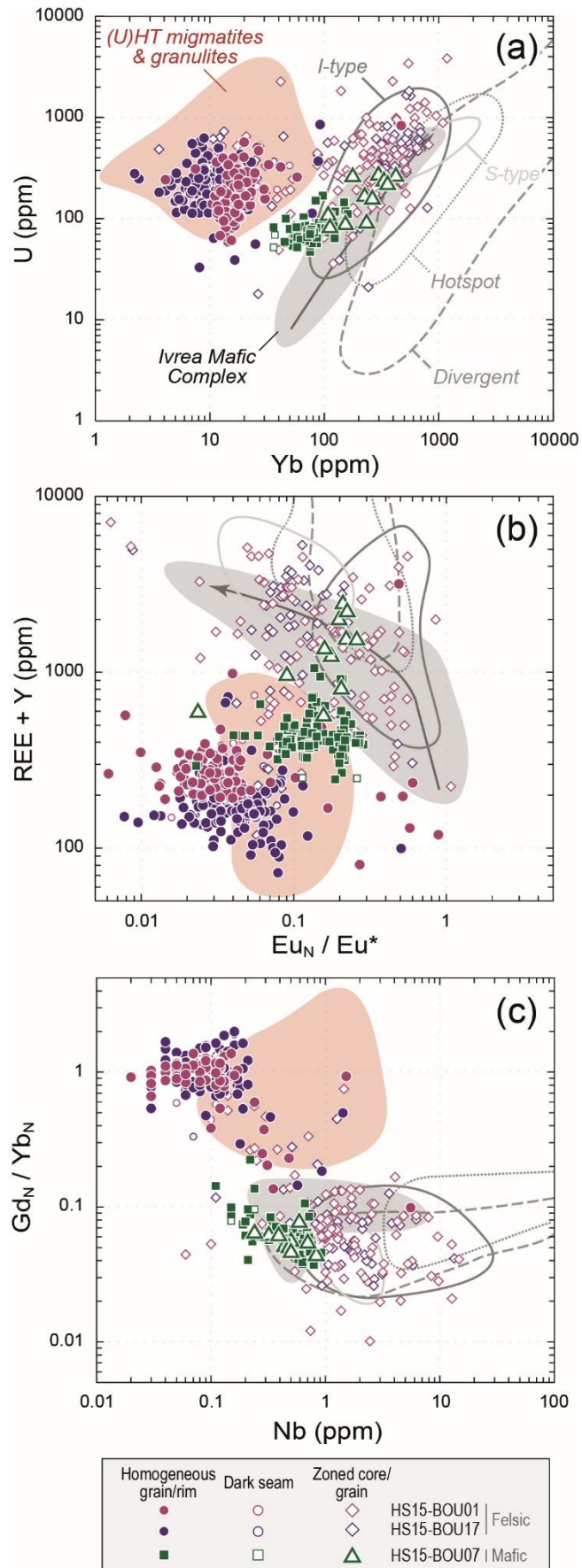


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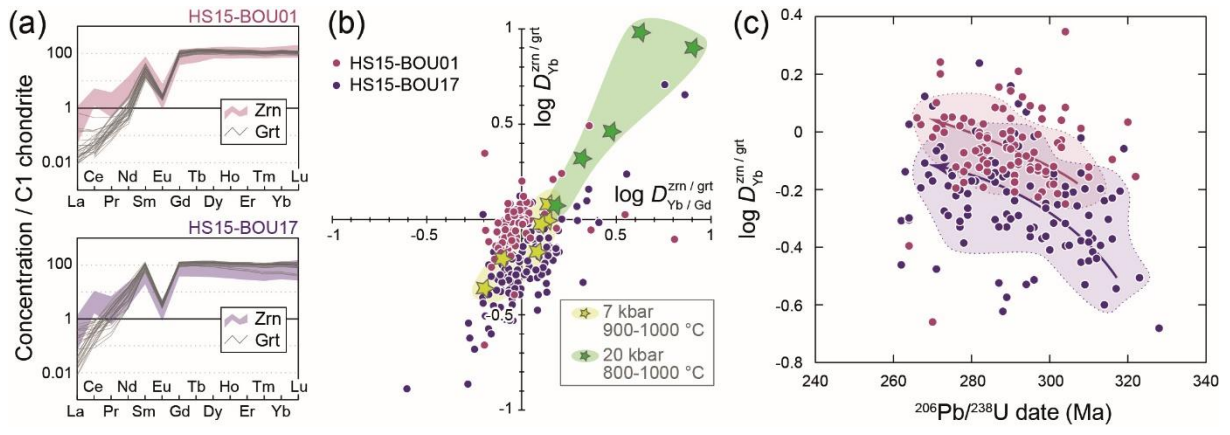
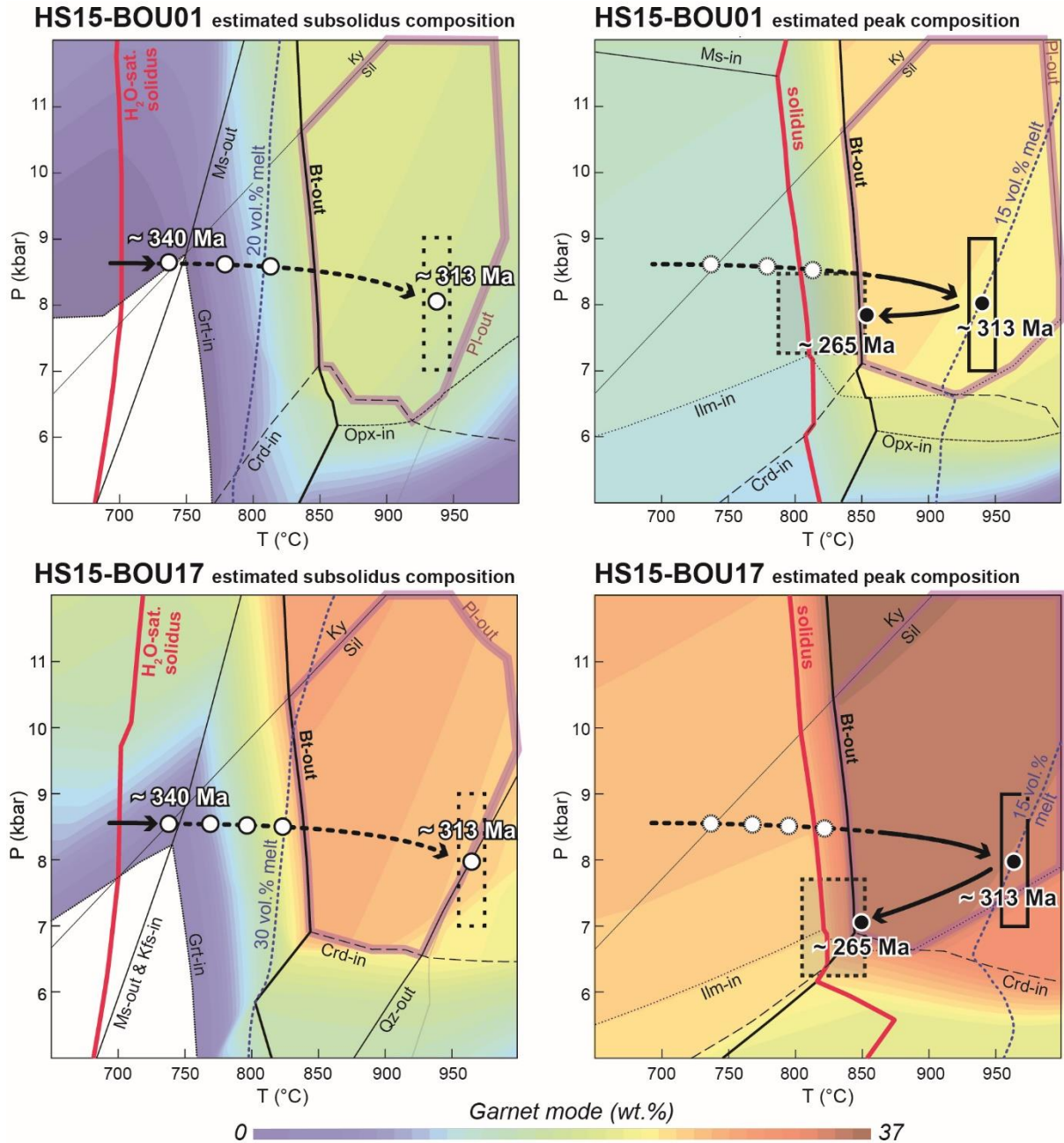


Figure 8



1. melt production & episodic extraction of >18-24 vol.%

2. peak conditions (ca. 15 vol.% melt)  
 3. melt extraction along the retrograde path ●

Figure 9

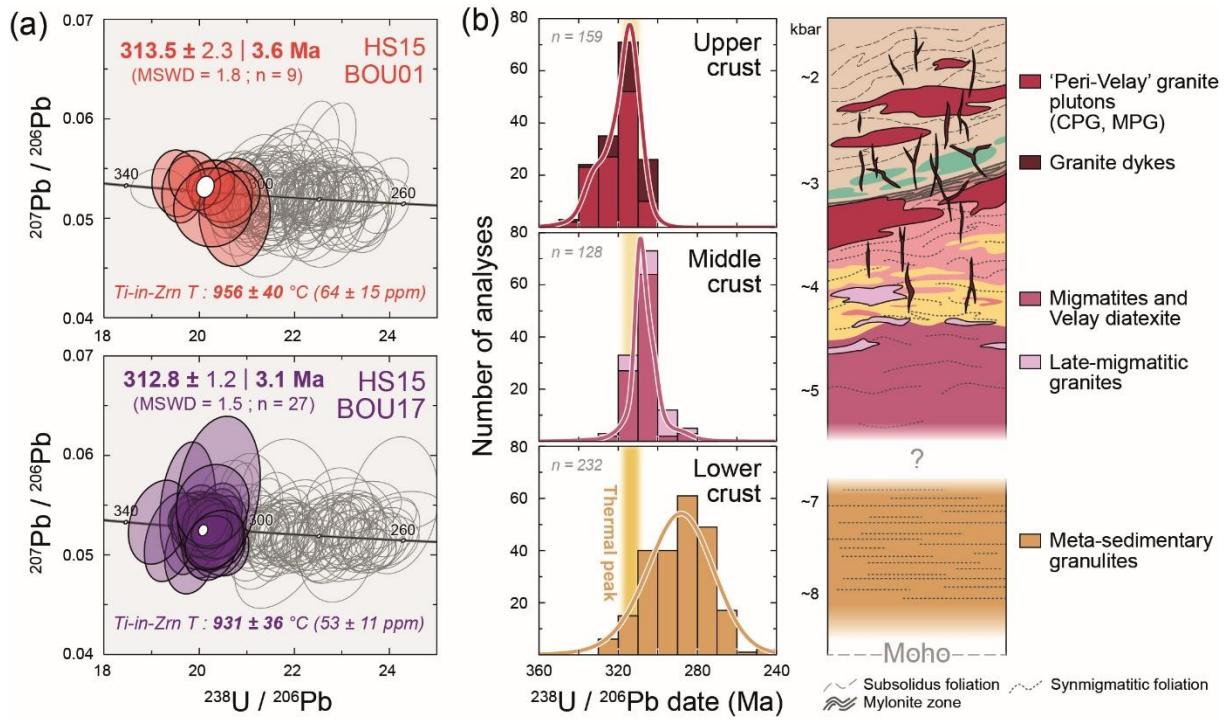


Figure 10