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1 **When zircon drowns: elusive geochronological record of water-fluxed**
2 **orthogneiss melting in the Velay dome (Massif Central, France)**

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15 **Keywords:** zircon U–Pb geochronology, water-fluxed melting, anatexis, Variscan
16 orogeny, Velay dome

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Abstract

Zircon U–Pb geochronology is routinely performed to unravel the timing and duration of melting events within the continental crust. A comprehensive understanding of the zircon behavior during anatexis is therefore paramount to accurately assess the impact of crustal melting on orogenic processes. We report on an anatectic system from the southern part of the Velay dome (Variscan French Massif Central) encompassing late Carboniferous migmatites developed at the expense of peraluminous orthogneisses and associated with (leuco)granite bodies intrusive at higher structural levels. Our zircon U-Pb geochronological survey (494 analyzed grains) evidences a very scant record of the Variscan anatectic event. Indeed, oscillatory-zoned melt-precipitated zircon grains extracted from metatexites, leucosomes and (leuco)granites almost systematically yielded the late Neoproterozoic crystallization age of the felsic igneous protolith. Field and petrographic observations, whole-rock geochemical signatures, phase equilibrium and thermodynamically-constrained trace element modelling collectively indicate that partial melting of the orthogneisses took place at $T \sim 700^\circ\text{C}$ and was fluxed by ingress of external water. The lack of new zircon growth is ascribed to a range of processes ultimately related to the low melting temperature including limited zircon solubility in the melt phase, occlusion of zircon in non-reacting source biotite and sluggish zircon dissolution kinetics. Peraluminous orthogneisses, yet highly fusible due to their “eutectic” subsolidus mineral assemblage, exhibit a very limited zircon record of low temperature water-fluxed melting. Thus, zircon has little potential to provide reliable chronological constraints on water-fluxed melting episodes affecting such lithologies.

1. Introduction

57 Crustal melting has a profound impact on the mechanical and rheological behavior of
58 the continental crust (Rey et al., 2009; Rosenberg and Handy, 2005; Vanderhaeghe
59 and Teyssier, 2001). Deciphering orogenic processes thus requires an assessment of
60 the timing and duration of anatexis events. For that purpose, zircon, an accessory
61 mineral commonly crystallizing in intermediate to felsic magmas, has proven a very
62 efficient tool as it can be accurately and precisely dated using the U–Pb method.
63 Accordingly, many studies concluded that anatexis melt can be present for long
64 periods of time within the crust based on the zircon U–Pb age distribution of migmatites
65 or anatexis granites (Guergouz et al., 2018; Harley, 2016; Korhonen et al., 2013;
66 Laurent et al., 2018; Laurent et al., 2017; Lemirre et al., 2019; O'Brien and Miller, 2014;
67 Rubatto et al., 2013; Rubatto et al., 2009; Taylor et al., 2019).

68 Low temperature ($T < 750^{\circ}\text{C}$) anatexis generates significant melt fractions and greatly
69 impacts crustal rheology only if melting is fluxed by the addition of external water, a
70 more common process than generally thought (see Weinberg and Hasalová (2015)
71 and references therein). The zircon record of low temperature water-fluxed melting
72 remains unclear, as studies conducted on anatexis systems having experienced such
73 melting conditions are scarce and yielded contrasting results. Rubatto et al. (2009)
74 report extensive new zircon crystallization in migmatites, allowing an accurate
75 assessment of the timing of partial melting. In contrast, other authors (e.g. Brouand et
76 al. (1990), Oliver et al. (1999)) document cases of migmatites nearly lacking any new
77 zircon growth related to the anatexis event, but without a clear assessment of the role
78 of water-fluxed melting.

79 We address this issue based on a case study from the Velay dome (French Massif
80 Central), an archetypal granite–migmatite dome formed at the end of the late Paleozoic
81 Variscan orogeny, which has been a foundation stone for many concepts regarding
82 the role of partial melting in orogen dynamics (Burg and Vanderhaeghe, 1993; Costa
83 and Rey, 1995; Ledru et al., 2001; Vanderhaeghe et al., 2020; Vanderhaeghe and
84 Teyssier, 2001). Recently, it has been proposed that the supply of heat and H_2O from
85 granitic magmas emplaced at mid-crustal levels played an important role in the
86 development of the Velay dome migmatites (Barbey et al., 2015; Villaros et al., 2018),
87 making them an ideal target for this study. We document the field relationships,

88 petrography, whole-rock geochemistry and zircon U–Pb systematics of migmatites and
89 associated anatectic (leuco)granites from the Masméjean antiform which is located at
90 the southwesternmost part of the Velay dome (Fig. 1). We unravel their petrogenesis
91 and discuss the scant zircon record of crustal melting as observed from our dataset.
92 Our results highlight the limitations of zircon in deciphering the duration of water-fluxed
93 crustal melting events and bear implications on the zircon record of anatexis.

94

95 **2. Geological setting and sampling**

96 **2.1 The Velay dome in the French Massif Central**

97 The eastern French Massif Central exposes igneous and metamorphic lithologies
98 belonging to the inner part of the Variscan belt (Fig. 1), a Devonian–Carboniferous
99 collisional orogen which resulted from the closure of several oceanic domains and
100 culminated in the amalgamation of Gondwana, Laurussia and intervening
101 microcontinents (Matte, 1991). In the eastern French Massif Central, three main
102 metamorphic nappes have been described (the following description is based on Faure
103 et al. (2009); Vanderhaeghe et al. (2020) and references therein). The high-grade
104 *Upper Gneiss Unit* encompasses ortho- and paragneisses, locally anatectic, along with
105 the so-called Leptynite-Amphibolite Complex interpreted as dismembered ophiolite
106 fragments. This nappe preserves evidence for an early high-pressure evolution,
107 records late Devonian to early Carboniferous metamorphic ages and would represent
108 a crustal segment having experienced continental subduction. It was thrust over the
109 *Lower Gneiss Unit*, composed of meta-igneous and meta-sedimentary lithologies
110 which have been pervasively affected by partial melting and intruded by peraluminous
111 granitic magmas in the late Carboniferous, delineating the Velay migmatite–granite
112 dome. To the South, the dome is in contact with the *Parautochthonous Unit*, composed
113 of greenschist to lower amphibolite-facies metasediments. High-grade rocks from the
114 three units are unconformably sealed by late Carboniferous–early Permian coal-
115 bearing sediments.

116 The southern Velay dome (Fig. 1) has been the subject of numerous petrological and
117 geochronological studies reviewed in Barbey et al. (2015). Partial melting took place
118 under biotite-stable conditions at T of ~720°C and mid-crustal pressures of 0.3–0.6
119 GPa, locally reaching T ≥ 800°C at P ~ 0.4 GPa (Montel et al., 1992). These higher-T

120 migmatites commonly crop out as diatexites and are spatially related to the Velay
121 granite s.s (Ledru et al., 2001), dated at 310–300 Ma (Chelle-Michou et al., 2017;
122 Couzinié et al., 2014; Laurent et al., 2017), and to the main Velay dome emplacement
123 (Bouilhol et al., 2006; Vanderhaeghe et al., 2020). Other manifestations of this
124 magmatic event include (Barbey et al., 2015): (i) cross-cutting late intrusive
125 (micro)granites; (ii) K-rich diorites (locally called "vaugnerites") originating from melting
126 of the orogenic mantle (Couzinié et al., 2016). Both yielded zircon and monazite U(–
127 Th)–Pb intrusion ages (ca. 310–295 Ma) that are indistinguishable within uncertainty
128 from the Velay granite s.s (Couzinié et al., 2014; Didier et al., 2013; Laurent et al.,
129 2017).

130 Pervasive melting of the middle crust in the Velay dome is tied to the late orogenic
131 evolution of the Variscan belt (Ledru et al., 2001) and has been attributed to the thermal
132 maturation of the thickened orogenic crust upon radioactive heating supplemented by
133 advective heat transfer from mantle- ("vaugnerites") and lower crust-derived magmas
134 (Barbey et al., 2015; Costa and Rey, 1995; Villaros et al., 2018). The exhumation of
135 the mid-crustal Velay migmatites and granites was accommodated by: (i) a diapiric
136 component, revealed by the strain pattern in the dome core; (ii) a strong mechanical
137 decoupling between the migmatites and the unmelted cover at the roof of the dome,
138 evidenced by high strain ductile shear zones (Fig. 1) such as the Pilat detachment to
139 the North or the Mylonitic Metamorphic Vellave Zone (MMVZ) to the South (Bouilhol et
140 al., 2006).

141 **2.2 The Masméjean antiform**

142 2.2.1 General outline

143 The Masméjean antiform is a conical structure showing gently dipping southwestern
144 and northwestern flanks which contrast with the steep eastern flank (Fig. 2). It is built
145 up by an anatectic orthogneiss core mantled by an association of unmelted
146 greenschist- to amphibolite-facies orthogneisses and metasediments with very scarce
147 and small amphibolite bodies. Near Labastide-Puylaurent, the steep N–S foliation of
148 the eastern flank gradually rotates to an E–W trend, matching that displayed by the
149 Velay migmatites further East (Bouilhol et al., 2006; Briand et al., 1994; Lapadu-
150 Hargues, 1947) therefore demonstrating that both domains are in petrological–
151 structural continuity. Investigation of mineral assemblages in metasediments (Briand

152 et al., 1994; Négron, 1979) led to the recognition of Grt–Chl, And–Bt–Crd and Sil
153 metamorphic zones (all mineral abbreviations are after Whitney and Evans, 2010), the
154 close juxtaposition of which defines a low-pressure high-temperature metamorphic
155 field gradient in the area.

156 Orthogneisses include a range of fine-grained to coarse-grained varieties, some facies
157 containing K-feldspar porphyroclasts defining an augengneiss texture. They belong to
158 the Velay Orthogneiss Formation, a peraluminous meta-granite suite regionally defined
159 as part of the Lower Gneiss Unit, which emplacement was dated at c. 542 Ma
160 (Ediacaran–Cambrian boundary, Fig. 2, Couzinié et al. (2017)). The metasediments
161 structurally overlie the orthogneisses or constitute metre- to kilometre-sized rafts
162 embedded within them. U–Pb dating of detrital zircons constrains their maximum
163 depositional age to be Ediacaran (568.3 ± 4.9 Ma, Couzinié et al. (2019)) supporting an
164 original intrusive contact between the orthogneiss protolith and the (meta)sedimentary
165 series, later on tectonically transposed. Metamorphic rocks have been intruded: (i) to
166 the NW by the Margeride granite, dated at 312.9 ± 2.0 Ma (LA–ICP–MS U–Pb zircon,
167 Laurent et al. (2017)); (ii) to the SE by the Borne granite, dated at 309.5 ± 3.6 Ma (LA–
168 ICP–MS U–Pb zircon, Couzinié (2017)). Finally, the Masméjean antiform was
169 unconformably overlain by Lower Jurassic shallow marine limestones upon which
170 metasediments from the Grt–Chl zone were overthrust at Cenozoic times (Briand et
171 al., 1994).

172 2.2.2 The anatectic core of the antiform

173 One striking field observation in the Masméjean antiform is that migmatites are
174 exclusively developed at the expense of orthogneisses (Fig. 3a to e). Contrastingly,
175 paragneisses do not show clear evidence for in situ partial melting. The migmatites
176 predominantly consist of stromatic metatexites featuring: (i) cm-sized granitic
177 leucosomes with an assemblage Qz+Pl+Kfs and rare Bt; and (ii) narrow mm-sized
178 melanosomes consisting of Bt±Ms (Fig. 3f). Biotite is often altered to chlorite and
179 plagioclase to sericite. Accessory phases include apatite grains (locally mm-sized),
180 ilmenite, monazite and zircon. Relevant microstructures include (Fig. 3f,g): (i)
181 interstitial quartz, often dispersed along grain boundaries and locally showing cusped
182 contacts with adjacent feldspar grains; (ii) well equilibrated quartz–feldspar and
183 feldspar–feldspar triple junctions. Leucosome concentration within fold hinges (Fig. 3b)
184 reflects the collection of anatectic liquids in low-strain domains during syn-anatectic

185 deformation. Besides, melt mobility and coalescence is further indicated by the
186 presence of larger (a few 10's of cm-thick), concordant to discordant granitic
187 leucosomes (Fig. 3c,d). Those are primarily composed of $Qz+Pl+Kfs+Bt\pm Ms\pm Sil$ (Fig.
188 3i). Most migmatites experienced subsolidus deformation as evidenced by undulose
189 quartz extinction. Finally, scarce vaugnerite bodies have been observed within the
190 migmatites (Fig. 3e) and are intimately associated with coarse-grained Qz -feldspar
191 domains.

192 2.2.3 The unmelted parts of the antiform and the contact with the 193 migmatites

194 Metamorphic lithologies from the southwestern flank of the antiform show evidence for
195 strong ductile deformation involving a combination of pure shear and simple shear (Fig.
196 2). Marked flattening is supported by: (i) the well-developed foliation associated with
197 grain-size reduction in augen gneisses and paragneisses, locally grading to S-
198 mylonites; (ii) the conspicuous occurrence of symmetrical folds the axial planes of
199 which match the trend of the regional foliation (Fig. 4a). When present, stretching
200 lineations are underlined by ellipsoidal biotite aggregates and K-feldspars clasts
201 showing a strong preferential orientation in the YZ plane. They plunge towards the SW
202 with a maxima at N230 and 21° dip. Sections in the XZ plane reveal C to C' shear
203 bands, asymmetric drag folds and sigmoid mica "fish" consistently indicative of top-to-
204 the NE shearing (Fig. 4b). Besides, injected granite bodies in the form of metre- to
205 hectometre-sized sills are frequently observed in Bt - Sil paragneisses from the lower
206 part of the high strain zone (Fig. 4c), right above the migmatitic orthogneisses
207 (corresponding to the so-called "epibolite formation" of Briand et al. (1994), Fig. 2).
208 Field observations indicate that they intruded before the last increment of top-to-the
209 NE ductile deformation (Fig. 4d). Finally, non-coaxial deformation along ESE-oriented
210 (maxima at N110) shear planes evidences a top-to-the S displacement at the brittle-
211 ductile transition (Fig. 4c). Collectively, these observations strikingly recall the main
212 features of the Metamorphic Mylonitic Vellave Zone (MMVZ, Bouilhol et al., 2006)
213 described to the East at the interface between the Velay migmatites and their unmelted
214 cover (Fig. 1). We therefore conclude that the high-strain zone observed in the SW
215 flank of the Masméjean antiform represents the western continuation of the MMVZ.

216 In the eastern flank of the antiform, the regional foliation is N-S trending, steeply
217 dipping towards the E (on average at 55°) and the metamorphic field gradient is even

218 more condensed (Fig. 2). The migmatites are most often overlain by Bt–Sil
219 paragneisses here again intruded by numerous concordant granite bodies (Lapadu-
220 Hargues, 1947) grouped together as “Villefort granites” in the following (Fig. 4e). The
221 And–Bt–Crd zone is notably narrow and its constituent lithologies pervasively affected
222 by intense ductile deformation, locally resulting in S- and L-mylonites (Fig. 4f).
223 Stretching lineations are predominantly horizontal and N–S trending, an apparent left-
224 lateral sense of shear being evidenced by asymmetric quartz aggregates and shear
225 bands. Ductile structures are overprinted by brittle–ductile and brittle deformation (Fig.
226 4h and i) along N–S steeply dipping fractures and fault planes with complex kinematics
227 including both strike-slip and vertical displacements. This high-strain zone is related to
228 the activity of the so-called “Villefort Shear Zone” which would have been active since
229 the late Carboniferous (Deroin and Prost, 1993; Lapadu-Hargues, 1947).

230 Finally, in the northwestern part of the antiform, the transition from migmatites to
231 unmelted rocks appears gradual and there is no evidence for intense ductile shearing
232 as identified in the other flanks. Stretching lineations are very scarce and, when
233 present, gently plunge to the NW. Locally, a top-to-the SE sense of shear in unmelted
234 orthogneisses can be inferred based on asymmetric Kfs clasts. Several metre-thick
235 granite sills and dykes have been observed both in this area and in the highest
236 structural levels of the SW flank (in the And–Bt–Crd zone, Fig. 2). Those granites
237 generally cut across the foliation of unmelted orthogneisses or form concordant bodies
238 at the interface between meta-igneous and metasedimentary rocks (Fig. 5a to e).

239 2.2.4 Investigated samples

240 A total of 25 samples were studied in this work. The sampling locations (GPS
241 coordinates) are reported in Supplementary Table S1 and the dated samples are
242 located on figure 2.

243 Four leucocratic orthogneiss samples (CHA-15-35, -36, -44, -52) were collected from
244 the And–Bt–Crd zone, two of which (CHA-15-35 and -52) were selected for zircon U–
245 Pb geochronology. They typically show a planar fabric defined by Ms+Bt layers
246 alternating with fine- to medium-grained Qz and Pl+Kfs layers, cm-sized Kfs
247 porphyroclasts being locally present.

248 Seven migmatitic orthogneiss samples were also collected. Sample CHA-15-43 is a
249 20 cm-thick leucosome within flat-lying metatexites from the antiform core. Sample

250 CHA-15-63 was taken from a decametre-sized granite unit showing gradual contacts
251 with the surrounding migmatites in the northern part of the antiform. CHA-15-17, CHA-
252 15-65M and CHA-15-65L (Fig. 3a,d) were all sampled at the Puylaurent dam and
253 respectively correspond to two bulk metatexites and one leucosome. Besides, two
254 samples of the Velay migmatites were taken at Saint-Laurent-les-Bains, 6 km east of
255 the Masméjean antiform (Fig. 2): CHA-15-47 corresponds to a bulk metatexite and
256 CHA-15-48 a 10 cm-thick leucosome embedded within it.

257 Collected granite samples (13 in total) fall into three groups based on textures,
258 mineralogy and field relationships:

- 259 - Seven medium-grained, two-mica leucogranites have been collected from
260 decametre-sized bodies intruding Bt–Sil paragneisses right above the anatexitic
261 core. Five samples (SC14-40, CHA-15-24, -25, -41, -49) correspond to Villefort
262 granites (from the eastern flank of the antiform). Sample -41 notably displays a
263 vertical magmatic foliation defined by layers rich in centimetre-sized K-feldspar
264 crystals alternating with finer-grained leucocratic domains (Fig. 4g). Two
265 additional samples (CHA-15-05 and -19) have been taken from the SW flank of
266 the antiform (Fig. 4c). All samples underwent sub-solidus high-temperature
267 deformation witnessed by sub-grain rotation in quartz and the presence of Ms-
268 and Sil-filled shear bands. Locally, sample -41 features mm-scale cataclastic
269 bands showing marked grain-size reduction and subsequent alteration (Fig. 4h).
270 Retrogression is evidenced by the replacement of biotite by chlorite and
271 plagioclase by sericite (Fig. 4i).
- 272 - Samples CHA-15-29, -34, -51 and -59 are pink two-mica granite dykes cutting-
273 across the foliation of unmelted orthogneisses and metasediments within the
274 Crd–Bt–And zone (Fig. 5a,b,e). Sample CHA-15-51 contains garnet. All
275 samples experienced subsolidus deformation, from weak (Qz undulose
276 extinction; CHA-15-34, -51) to marked (Qz subgrain rotation; CHA-15-29, -59).
- 277 - Samples CHA-15-33 and -07 have been taken from metre-sized sills concordant
278 within the foliation of the orthogneisses and metasediments in the And–Bt–Crd
279 zone. They are fine-grained porphyritic biotite granites displaying a magmatic
280 fabric defined by the planar orientation of mm-sized embayed quartz and
281 feldspar phenocrysts along with biotite aggregates, particularly abundant in

282 sample -07 (Fig. 5c,d). CHA-15-33 also contains magmatic muscovite
283 phenocrysts.

284

285 **3. Analytical and modelling techniques**

286 **3.1 Whole-rock geochemistry**

287 Whole-rock chemical compositions of the selected samples were obtained from the
288 *ALS Global* company, combining ICP–AES and ICP–MS analyses for major and trace
289 elements, respectively, and plotted using the GCDkit software (Janoušek et al., 2006).
290 Results are presented in Supplementary Table S1 and summarized in figures 6 and 7.

291 **3.2 Zircon U–Pb dating**

292 Zircon grains were separated from 17 crushed rock samples at Saint-Etienne
293 University using conventional techniques (sieving, panning, magnetic and heavy
294 liquids separation followed by handpicking). Selected grains were subsequently cast
295 into epoxy mounts and polished down to an equatorial section. Back-scattered electron
296 (BSE) and cathodoluminescence (CL) imaging were performed at the Central
297 Analytical Facility of Stellenbosch University (SUN, South Africa) using a Zeiss
298 MERLIN scanning electron microscope (SEM) and at the Laboratoire Magmas et
299 Volcans (Clermont-Ferrand) using a Jeol JSM-5910 SEM. Analyses of U–Pb isotopes
300 were carried out by laser ablation – inductively coupled plasma – sector-field – mass
301 spectrometry (LA–ICP–SF–MS) at Goethe Universität, Frankfurt-am-Main (GUF,
302 Germany) and at ETH Zurich (Switzerland), in both cases via a RESOLUTION
303 (ASI/Applied Spectra, Australia/USA) excimer (ArF, 193 nm) laser ablation system
304 attached to an Element XR (Thermo, Germany) ICP-MS. Detailed information about
305 the analytical methods are presented in the Supplementary Text, including the
306 methodology followed to propagate systematic errors to the calculated Concordia
307 dates. The full analytical datasets (standards and samples) are reported in
308 Supplementary Tables S2 to S5. The CL image of each zircon grain along with the
309 location of the laser spots and associated U–Pb data are displayed Supplementary
310 Figures S1 to S5. A summary of the main texture–date results is provided Figure 8 and
311 results of U-Pb analyses summarized in Figure 9 (for orthogneisses and migmatites)
312 and Figure 10 (for granites).

313 **3.3 Phase relations and trace-element modelling**

314 Forward thermodynamic modelling was performed in order to investigate the P–T–X_{H₂O}
315 conditions of anatexis in the Masméjean antiform. Field observations and migmatite
316 U–Pb zircon data (see section 4.2) indicate that augen gneisses from the Velay
317 Orthogneiss Formation arguably constitute the protolith of the migmatites. Accordingly,
318 a representative unmelted augen gneiss composition was chosen as starting
319 composition (SC-14-33, from Couzinié et al. (2017)). Pseudosections were calculated
320 by Gibbs energy minimization using PerpleX (Connolly, 2009) and the self-consistent
321 thermodynamic dataset of Holland and Powell (1998) (updated version of 2004) in the
322 system Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O–TiO₂ (NCKFMASHT). The
323 selected solution models are listed in the Supplementary Text. The starting
324 composition was recalculated anhydrous and the H₂O concentration subsequently set
325 at 0.53 wt.%, based on a T–X_{H₂O} pseudosection, to obtain a fully hydrated subsolidus
326 system at 0.5 GPa. The retained chemical composition is displayed Table 1. Following
327 the methodology of Janoušek and Moyen (2019), the Rb, Ba, Sr and Nb contents of
328 modelled melts and solid residues were estimated at 63 points within the modelled P–
329 T–X_{H₂O} sections (between 680–800°C at 0.4–0.6 GPa and for bulk water contents of
330 0.53, 1.6 and 2.6 wt.%, Fig. 11a,b) using: (i) the calculated major element melt
331 compositions and phase proportions; (ii) a set of partition coefficients for felsic systems
332 compiled by Laurent et al. (2013) and listed Supplementary Table S6.

333

334 **4. Results**

335 **4.1 Geochemistry**

336 4.1.1 Orthogneisses

337 Geochemical data from investigated orthogneisses were merged with the existing
338 dataset available for the Velay Orthogneiss Formation (Couzinié et al., 2017) in the
339 study area. The constituent augen gneisses (n=10) and leucogneisses (n=11)
340 correspond to metamorphosed porphyritic granites and leucogranites, respectively,
341 with SiO₂ contents ranging from 69.4 to 78.6 wt.%. Together, they define a
342 peraluminous, potassic (K/(Na+K): 0.49–0.59) and ferriferous association (sensu
343 Debon and Le Fort (1988), Fig. 6). Augen gneisses show moderate LILE (Sr: 60–141
344 ppm; Ba: 356–1140 ppm, Fig. 7) and elevated HFSE (Zr: 110–234 ppm; Nb: 7.9–13.5

345 ppm) and REE (Sum of REE: 88–150 ppm) contents along with moderately
346 fractionated REE patterns (La_N/Yb_N : 4–9) and negative Eu anomalies (Eu_N/Eu^* : 0.33–
347 0.82, Fig. 7). Detailed description and interpretation of the geochemical signature of
348 this meta-igneous suite are presented in Couzinié et al. (2017).

349 4.1.2 Migmatites

350 The three metatexites mostly overlap in terms of major and trace element composition
351 with unmelted augen gneisses (Fig. 6, 7). The melanocratic sample CHA-15-65M is
352 less silicic but more potassic, mafic and richer in Nb (up to 14.9 ppm) which reflects a
353 higher modal abundance of biotite (Bea et al., 1994; Debon and Le Fort, 1988).

354 The leucosomes are highly silicic (SiO_2 : 72.6–76.3 wt.%), peraluminous and potassic
355 ($K/(Na+K)$: 0.52–0.61, Fig. 6). They show LILE contents (Sr: 54–85 ppm; Ba: 147–526
356 ppm) similar to those of augen gneisses but are markedly depleted in HFSE (Zr <42
357 ppm, Nb <7.5 ppm) and REE (Sum_{REE} <46 ppm, Fig. 7a). REE patterns are poorly
358 fractionated (La_N/Yb_N : 1.7–5.3) and Eu anomalies ranging from slightly negative to
359 slightly positive (Fig. 7, Eu_N/Eu^* : 0.79–1.15) apart from sample 43 (Eu_N/Eu^* = 0.35).

360 4.1.3 Granites

361 Granites are peraluminous and define a trend with a positive slope between the felsic
362 and high peraluminous fields in the B–A diagram. They overall share similar major
363 element compositions with meta-granites from the Velay Orthogneiss Formation but
364 are slightly more magnesian (Fig. 6). The granite dataset can be divided into two
365 groups based on whole-rock geochemistry.

366 A first group termed “low-Ba granites” encompasses three dykes (CHA-15-29, -34, -
367 51) from the And–Crd–Bt zone and two Villefort granites (CHA-15-24, -25). These
368 samples are leucogranites s.s., highly silicic (SiO_2 : 72.6–75 %), sodi-potassic to sodic
369 ($K/(Na+K)$: 0.39–0.50, Fig. 6b), and strongly depleted in LILE (Sr <60 ppm; Ba <170
370 ppm) compared to the leucosomes (Fig. 7). HFSE and REE contents are markedly low
371 (Zr <30 ppm; Sum of REE <50 ppm) apart from Nb (6.8–16.3 ppm). They display poorly
372 fractionated REE patterns (La_N/Yb_N : 3–7) with marked Eu negative anomaly (Eu_N/Eu^* :
373 0.35–0.56, Fig. 7).

374 A second group termed “high-Ba granites” includes dyke CHA-15-59, the fine-grained
375 porphyritic granite sills (CHA-15-07 and -33) and remaining bodies intrusive within Bt–
376 Sil-paragneisses (CHA-15-05, -19, -41, -49, -59, SC-14-40). They show a broader

377 range of SiO₂ contents (66–74 wt.%) and are more potassic than the “low-Ba” granites
378 (K/(Na+K): 0.52–0.69). They are richer in LILE (Sr: 103–279 ppm; Ba: 390–1310 ppm)
379 and Zr–REE (Zr: 81–228 ppm; Sum_{REE}: 80–178 ppm) than low-Ba granites and
380 leucosomes. Nb contents are similar to low-Ba types (11–17.4 ppm). REE patterns are
381 fractionated (La_N/Yb_N: 9–29) and feature a moderate negative Eu anomaly (Eu_N/Eu*:
382 0.41–0.73).

383 **4.2 Zircon U–Pb dates**

384 4.2.1 Orthogneisses

385 Zircon grains extracted from orthogneiss samples are euhedral to sub-euhedral, range
386 in size from 80 to 400 μm, display aspect ratios between 1.4:1 and 3:1 and show well-
387 developed {2 1 1} pyramids (Fig. 8a). Core–rims relationships are common and
388 typically feature CL-dark cores (most often poorly zoned and sometimes recrystallized)
389 surrounded by rims with well-developed magmatic oscillatory zoning. Grains devoid of
390 cores generally show a similar, oscillatory zoning pattern. Such rims and
391 homogeneous grains yielded Concordia dates of 544.1 ± 4.9 Ma (MSWD_{C+E}=0.79,
392 n=11) and 544.4 ± 5.2 Ma (MSWD_{C+E}=0.9, n=9) for CHA-15-35 and CHA-15-52,
393 respectively (Fig. 9), together with low Th/U ratios (average of 0.15). In both samples,
394 few measurements yielded a range of younger, nearly concordant, ²⁰⁶Pb/²³⁸U dates
395 down to 514 ± 7 Ma. Analyses performed on zircon cores and additional core-free
396 grains mostly yielded concordant Neoproterozoic (700–560 Ma) dates. Some
397 discordant analyses in both samples showed similar ²⁰⁶Pb/²³⁸U as concordant spots
398 but higher ²⁰⁷Pb/²⁰⁶Pb (Fig. 9), attributed to the presence of common Pb.

399 4.2.2 Migmatites

400 Zircon crystals extracted from the metatexites and leucosomes are very similar in
401 sizes, shapes and internal textures to orthogneiss zircons. In metatexite CHA-15-17,
402 zircon grains tend to be more elongated (aspect ratios up to 5.5:1, Fig. 8) and
403 commonly lack inherited cores. Besides, zircons are notably smaller (mostly <90 μm)
404 in leucosome CHA-15-63 (Fig. 8). Results of U–Pb dating were very consistent from
405 one sample to another. Concordant ²⁰⁶Pb/²³⁸U dates obtained on oscillatory-zoned
406 core-free grains and rims clustered in the range 550–540 Ma (Fig. 9) and showed low
407 Th/U ratios (average 0.12). Concordia dates of 545.1 ± 4.8 Ma (MSWD_{C+E}=1.02, n=16),
408 539.7 ± 5.7 Ma (MSWD_{C+E}=1.3, n=4) and 549.0 ± 4.8 Ma (MSWD_{C+E}=1.04, n=14) can

409 be calculated for metatexites from the Velay (CHA-15-48) and the Masméjean antiform
410 (CHA-15-17 and -65M), respectively. Cores and few core-free crystals yielded older,
411 predominantly Cryogenian–Ediacaran dates (>560 Ma). No concordant zircon date
412 was retrieved for leucosome sample CHA-15-65L.

413 In the leucosomes, two groups of zircon grains deserve special attention as they did
414 not yield any reliable U–Pb date because they exhibited complex and irregular isotope
415 fractionation patterns during ablation (labelled “ns” in Supplementary Figures, for which
416 no date could be calculated), or showed a scattered range of highly (>50%) discordant
417 Permian to Jurassic $^{206}\text{Pb}/^{238}\text{U}$ dates. In both cases, the spotted grains were very rich
418 in common Pb. Such zircon grains belong to two textural types: short CL-dark to bright
419 oscillatory zoned core-free grains, particularly abundant in sample CHA-15-65L; and
420 CL-dark porous metamict rims wrapping 650–540 Ma CL-bright zoned cores (Fig. 8,
421 rather present in CHA-15-43 and CHA-15-47).

422 4.2.3 Granites

423 Zircon grains from both low- and high-Ba granite samples are similar. They are
424 generally euhedral and show a broad range of sizes (60–350 μm) and aspect ratios
425 (1.1:1–3.6:1), up to 6:1 in sample CHA-15-19. Three zircon types can be recognized
426 (Fig. 8). A first, scarce type corresponds to grains featuring a CL-dark or bright
427 occasionally zoned core, truncated by a new growth zone with distinct magmatic
428 oscillatory zoning, similar to zircons extracted from orthogneisses. Concordant
429 $^{206}\text{Pb}/^{238}\text{U}$ dates obtained from cores and rims were almost exclusively pre-Variscan,
430 predominantly Ediacaran. A second, abundant type encompasses grains with
431 oscillatory zoned cores wrapped by CL-dark porous rim, often <15 μm -large and
432 sometimes displaying faint oscillatory zoning. Analyses of the largest rims revealed
433 high common Pb contents, low Th/U (<0.1) and very few gave concordant Variscan U–
434 Pb dates, often after a common Pb correction was performed. A third group
435 corresponds to zircon grains devoid of core–rim relationships and showing well-
436 developed magmatic oscillatory zoning. Most of them also yielded pre-Variscan
437 (mostly Ediacaran) dates but concordant Variscan were obtained on some grains
438 typified by elevated Th/U ratios (mostly in the range 0.4–1.2).

439 Most analyses performed on the four “low-Ba” granites (CHA-15-24, -25, -34 and -51)
440 yield $^{206}\text{Pb}/^{238}\text{U}$ dates clustered around ca. 550–540 Ma, with few (notably from cores)

441 older (mostly Ediacaran–Cryogenian) dates (Fig. 10). Only three analyses yielded
442 concordant Variscan dates, all from sample CHA-15-24 for which the Concordia date
443 of 305.8 ± 4.4 Ma ($\text{MSWD}_{\text{C+E}}=1.5$, $n=3$) can be calculated. The same situation is
444 observed for the “high-Ba” granites with a dominant Ediacaran population (except in
445 sample CHA-15-19 and -49) and scarce Variscan dates, albeit more frequent than in
446 the “low-Ba” types. Concordia dates of 304.0 ± 3.9 Ma ($\text{MSWD}_{\text{C+E}}=1.4$, $n=5$), 305.3 ± 3.9
447 Ma ($\text{MSWD}_{\text{C+E}}=0.6$, $n=5$) and 302.2 ± 3.7 Ma ($\text{MSWD}_{\text{C+E}}=1.6$, $n=6$) can be calculated
448 for CHA-15-41, CHA-15-19 and CHA-15-33, respectively. One single grain from
449 sample CHA-15-49 yielded a Variscan concordant $^{206}\text{Pb}/^{238}\text{U}$ date of 305 ± 4 Ma.

450 5. Discussion

451 5.1 Interpretation of U–Pb dates

452 For *orthogneisses*, identical (igneous) zircon textures, Th/U ratios and U–Pb dates
453 have been reported by Couzinié et al. (2017) on similar rocks from the southern part
454 of the Velay dome. Following this, we interpret the Concordia dates of 544.1 ± 4.9 Ma
455 and 544.4 ± 5.2 Ma obtained for CHA-15-35 and CHA-15-52 as the crystallization ages
456 of their granitic protoliths. As they were obtained on texturally and compositionally
457 similar crystals, the range of slightly younger $^{206}\text{Pb}/^{238}\text{U}$ dates would represent isotopic
458 resetting possibly related to the Variscan thermal event and/or recent Pb loss. Zircon
459 domains showing older dates (Paleoproterozoic and Neoproterozoic) would
460 correspond to source-inherited grains or xenocrysts incorporated from Ediacaran
461 sedimentary country-rocks during magma ascent.

462 For *granites*, we interpret the Concordia dates of 305.8 ± 4.4 , 304.0 ± 3.9 , 305.3 ± 3.9 ,
463 and 302.2 ± 3.7 Ma obtained on (rare) rims as the crystallization ages of samples CHA-
464 15-24 (“low-Ba”), -41, -19 and -33 (“high-Ba”), respectively. Notably, Variscan dates
465 were obtained from zircon rims and crystals having textural characteristics (CL-dark,
466 faintly zoned) contrasting with zircon grains (domains) with pre-Variscan dates (Fig.
467 8d). The latter would have been inherited from the source or incorporated during ascent
468 from country-rocks. Importantly, inherited grains constitute the bulk of the zircon
469 budget of four granite samples (CHA-15-24, -34, -49 and -51), for which no
470 crystallization age can be estimated.

471 The zircon grains extracted from the *metatexites* show similar CL characteristics, Th/U
472 ratios and $^{206}\text{Pb}/^{238}\text{U}$ dates than those of the local unmelted orthogneisses. The

473 Ediacaran (549.0 ± 4.8 to 539.7 ± 5.7 Ma) Concordia dates are therefore interpreted
474 as reflecting the crystallization age of the granitic protolith of the migmatites. Similarly,
475 the distribution of zircon dates in the **leucosomes** clusters in the range 550–540 Ma
476 and remarkably mimics that of the orthogneisses and metatexites (Fig. 11a). Even
477 though such dates were obtained on grain domains showing well-developed oscillatory
478 zoning, we argue that they do not correspond to the leucosome crystallization ages.
479 Indeed, Bouilhol et al. (2006) report the occurrence of magmatic monazite grains and
480 overgrowths of late Carboniferous age in leucosomes (as constrained by “chemical”
481 U–Th–Pb dating), which indicates that melt production and coalescence took place at
482 Variscan times, as observed throughout the Velay dome. Consequently, in the
483 Masméjean migmatites, the age of the anatectic event is not recorded by zircon.

484 Many zircon analyses in the leucosomes and granites yielded highly discordant dates
485 or unusual laser-induced Pb/U fractionation patterns. Such disturbance of the U–Pb
486 system likely results from Pb loss and/or common Pb incorporation related either to
487 radiation-induced damage for the metamict domains (Mezger and Krogstad, 1997;
488 Nasdala et al., 1998) or crystallization from volatile-rich melts (Watson et al., 1997).

489 Two main conclusions can be drawn from this dataset. First, both “low-Ba” and “high-
490 Ba” granites would be coeval within uncertainties (± 4 Ma) and were emplaced at c. 304
491 Ma. Their abundant inherited/xenocrystic zircon grains show a unimodal distribution of
492 $^{206}\text{Pb}/^{238}\text{U}$ dates featuring a marked peak at c. 545 Ma (Fig. 11b and c) as already
493 described in several Velay granites (Chelle-Michou et al., 2017; Couzinié et al., 2014;
494 Laurent et al., 2017). Assuming that most xenocrysts were incorporated in the magma
495 at source level, the perfect match with the emplacement ages of orthogneisses
496 indicates that the latter constitute the source of both “low-Ba” and “high-Ba” granite
497 magmas. The presence of 680–560 Ma-old dates may record the subordinate
498 involvement of metasediments since this age range represents the main detrital zircon
499 population in the latter (Couzinié et al., 2019). However, most of the pre-545 Ma zircon
500 domains actually correspond to cores wrapped by c. 545 Ma-old rims, suggesting they
501 were already incorporated as inherited grains in the granite protolith of the
502 orthogneisses.

503 Second, the zircon record of Variscan crustal melting in the Masméjean antiform is
504 strikingly limited. Notably, none of the studied metatexites ($n = 3$) and leucosomes (n
505 $= 4$) record any Variscan date (even discordant), and only half of the dated granites

506 yield unambiguous evidence for zircon crystallization during the regional Carboniferous
507 thermal event. In total, out of 494 analyzed zircon grains, only 20 yielded concordant
508 Variscan $^{206}\text{Pb}/^{238}\text{U}$ dates, of which 11 after a common Pb correction only. In other
509 words, <2% of our dataset yielded pristine dates corresponding to the Variscan partial
510 melting event. The significance of this is discussed section 5.5.

511 **5.2 P–T– $X_{\text{H}_2\text{O}}$ conditions of melting**

512 Following petrographic observations (see section 2), the paragenesis inferred at peak
513 metamorphic conditions for the migmatites is: melt+Qz+Kfs+Pl+Bt \pm Sil \pm Ms.
514 Examination of the pseudosection calculated for a representative orthogneiss
515 composition fully hydrated at the solidus, i.e. with 0.53 % H_2O (see section 3.3),
516 indicates that the stability field of this mineral assemblage is bounded at low
517 temperature by the solidus and at high temperature (and/or low pressure) by biotite-
518 breakdown and associated appearance of garnet and/or cordierite (Fig. 12a). As these
519 two minerals have not been observed in the migmatites, partial melting in the
520 Masméjean dome would have taken place at $0.3 < P < 0.7$ GPa and $650 < T < 720^\circ\text{C}$,
521 consistent with the estimates of Montel et al. (1992). Importantly, at these P–T
522 conditions muscovite dehydration-melting would be the main melt-producing reaction,
523 generating sillimanite and K-feldspar as peritectic phases (Clemens and Vielzeuf,
524 1987; Patiño Douce and Harris, 1998). However, with the low $X_{\text{H}_2\text{O}}$ considered in the
525 protolith (0.53 wt.% at near-solidus conditions), as little as 2–5 vol.% melt would have
526 been produced by muscovite-dehydration melting only (Fig. 12a).

527 Clearly, such estimates are inconsistent with field observations which evidence melt
528 mobility at the metre scale (Fig. 3), and likely over 100's of metres or about a kilometre,
529 considering that the granite sheets intruding the lower-grade metasediments and
530 orthogneisses stem from deeper parts of the same migmatite zone (as supported by
531 zircon U–Pb data). This indicates that the likely melt fraction was between the melt
532 percolation threshold under strain of 7 vol.% (Rosenberg and Handy, 2005) and the
533 static melt escape threshold of 20–25 vol.% (Vigneresse et al., 1996). A T– $X_{\text{H}_2\text{O}}$
534 pseudosection (Fig. 12b) highlights that the melt volumes inferred (roughly 10–20
535 vol.%) can be readily explained by a bulk H_2O content of c. 1–2 wt.%. This is
536 significantly higher than the amount of water required to fully hydrate the orthogneiss
537 protolith (0.53 wt.%), which therefore requires water ingress at near- or supra-solidus

538 conditions. Elevated X_{H_2O} in the bulk composition does not affect the main mineral
539 assemblages at given T but results in a dramatic increase of the melt productivity (Fig.
540 12b) due to the "eutectic" composition of the orthogneiss source, with quartz and
541 feldspars being dissolved in the melt phase via the reaction $Qz + Kfs + Pl + H_2O =$
542 melt, only limited by water availability (Johannes and Holtz, 1996).

543 Further evidence for low-temperature water-fluxed melting in the Masméjean
544 orthogneisses is provided by the examination of the overlying Bt–Sil paragneisses from
545 the high-strain zone (MMVZ). These rocks are in direct contact with the anatectic
546 orthogneisses (cf. "melt-in" domain, Fig. 2), regionally experienced the same peak
547 temperatures of 680–710°C (Bouilhol et al., 2006) and yet do not show clear evidence
548 for partial melting. A T– X_{H_2O} pseudosection calculated for a representative paragneiss
549 (SC-14-41) shows that melt productivity in such rocks is not affected by water ingress
550 at $T < 730^\circ\text{C}$ (Fig. 12c,d). This feature likely reflects the near absence of K-feldspar in
551 the paragneisses at peak metamorphic conditions (Négron, 1979), hampering
552 substantial melt production by the "eutectic" melting reaction. In the paragneisses, K-
553 feldspar is therefore the limiting reactant, and not H_2O as in the peraluminous
554 orthogneiss which were able to produce 10–20 vol% of melt as a result of water fluxing
555 in the same conditions.

556 **5.3 Interpretation of granite geochemistry**

557 Field relationships and zircon U–Pb systematics of Variscan leucosomes, "low-Ba" and
558 "high-Ba" granites from the Masméjean antiform indicate that all rock types represent
559 magmas produced by melting of an orthogneiss source. In this section, we unravel the
560 petrogenetic processes accounting for their contrasting whole-rock major and trace
561 element compositions.

562 Comparison between equilibrium melt compositions predicted by thermodynamic
563 modelling and natural samples are consistent for the "low-Ba" granites and leucosome
564 CHA-15-43 (Fig. 13a). Both their normative feldspar–quartz proportions and Ba–Sr–
565 Rb–Nb concentrations match that of modelled primary melts extracted from an average
566 augen gneiss at $T < 760^\circ\text{C}$ (based on Ba–Nb systematics) and under a large range of
567 X_{H_2O} (Fig. 13b and c). Low melting temperatures are also supported by their low
568 contents in REE, Th and Zr because these chemical elements are hosted in accessory
569 minerals (monazite, zircon) which solubilities in melts are positively correlated with

570 temperature (Rapp and Watson, 1986; Watson and Harrison, 1983). Monazite
571 saturation thermometry, which is more robust than zircon saturation thermometry due
572 to the higher solubility of monazite in peraluminous melts compared to zircon (Kelsey
573 et al., 2008), points to melt temperatures of 645–705°C (using the model of Montel
574 (1993)). Collectively, modelling results and saturation thermometry indicate that the
575 “low-Ba” granites would adequately represent equilibrium anatectic melts extracted
576 from the migmatites exposed in the core of the Masméjean antiform.

577 Most leucosomes depart from the composition of “low-Ba” granites, being richer in Or–
578 Q components (Fig. 13a), displaying lower Nb and higher Ba–Sr (Fig. 13b,c) contents
579 and lacking significant Eu anomaly (Fig. 7a). A simple model of 20% accumulation of
580 K-feldspar (Fig. 13c) can account both for their potassic character and the excess Ba–
581 Sr–Eu (elements hosted in the accumulated feldspar). However, their highly variable
582 and distinctly lower Nb contents, inconsistent with values predicted by equilibrium
583 melting of the orthogneisses, cannot be explained solely by K-feldspar accumulation.
584 One way to achieve such low concentrations would be to consider the role of
585 disequilibrium processes during partial melting (Johannes, 1983). When mineral
586 reactants in the melting reaction are consumed, they release their trace element
587 budget to the melt. Then, the trace elements equilibrate with the solid residue by
588 diffusion based on partition coefficients. If equilibration is not attained, the melt leaves
589 the source with a trace element composition governed by the stoichiometry of the
590 mineral reactants. In the Masméjean migmatites, biotite is not involved in the melting
591 reaction and the very first melts formed are likely to be strongly depleted in Nb (Fig.
592 13b). Considering the results of monazite saturation thermometry (632–697°C), the
593 leucosomes would therefore represent low temperature melts having failed to fully
594 equilibrate with the biotite-rich residuum (explaining their low Nb contents) and
595 encompassing a load of relict source crystals (mostly K-feldspar).

596 The “high-Ba” granites are poorer in SiO₂, richer in K₂O and slightly more mafic than
597 leucosomes and “low-Ba” granites (Fig. 6, 13). Results of peraluminous orthogneiss
598 melting experiments (Castro et al. (1999); García-Arias et al. (2012); Fig. 6) indicate
599 that such features can be primarily ascribed to higher melting temperatures ($T \geq 780^\circ\text{C}$),
600 reaching those required for substantial biotite breakdown. Elevated melting
601 temperature are also supported by their high LREE contents with monazite saturation
602 thermometry yielding 760–822°C. Sample CHA-15-07 is more mafic than experimental

603 melts (Fig. 6) which can be related to the presence in the sample of abundant mm-
604 thick biotite aggregates (see section 2.2.4) of possible xenocrystic origin. Importantly,
605 Ba–Sr contents are markedly higher than predicted by the high-T melting model (Fig.
606 13b) and cannot result from the accumulation of K–feldspar and plagioclase as this
607 would require unrealistic high amounts of cumulative feldspar (>30 wt.%, Fig. 13c).
608 Considering a diffusional disequilibrium melting model (as advocated for the
609 leucosomes), the positive correlation between Nb and Ba contents (Fig.13b) could
610 result from the concomitant dissolution of plagioclase, K-feldspar and biotite into the
611 melt phase without diffusive equilibration with the solid residue, the least equilibrated
612 melts showing the highest Ba–Nb. This trend is anchored in the compositional field of
613 high-T melts generated under elevated X_{H_2O} suggesting that the “high-Ba” granite
614 magmas could also have formed through water-fluxed melting but at temperatures
615 markedly higher than the “low-Ba” granites and leucosomes (800°C instead of 700°C).

616 **5.4 Timing of anatexis and origin of the fluxing water**

617 From East to West, the architecture of the contact between the dome and the overlying
618 metamorphic series evolves from overturned to the south (Burg and Vanderhaeghe
619 (1993) to nearly vertical in the Saint-Laurent-les-Bains cross-section (Fig. 2, Bouilhol
620 et al., 2006). This geometry reflects the dome emplacement dynamics at different
621 structural levels (deeper to the East). As such, a schematic pre-doming restored
622 section across the SW part of the Velay dome is depicted figure 14. From this
623 perspective, the MMVZ can be regarded as a relatively flat-lying shear zone the
624 development and activity of which was, based on field observations (see section 2.2),
625 contemporaneous with water-fluxed melting and granite magmatism. The fact that both
626 “low-Ba” and “high-Ba” granite types coexist and yield crystallization ages that are
627 identical within uncertainty suggests that they are the products of the same melting
628 event but sample different depths within the crust (Fig. 14). Specifically, the “low-Ba”
629 granites represent anatectic melts extracted from low-T migmatites (700°C)
630 corresponding to those observed at the present exposure level, while the “high-Ba”
631 hotter (800°C) granites stem from maximum a few km deeper. The timing of magma
632 production, segregation from the source, ascent and crystallization is generally in the
633 order of a few million years (Harris et al., 2000; Petford et al., 2000), i.e. comparable
634 to or smaller than the typical dating uncertainties obtained here for granites (about ± 4
635 Ma). Therefore, the crystallization ages obtained for both the “low-Ba” and “high-Ba”

636 granites can be reasonably considered as dating the anatexis event. Hence, water-
637 fluxed melting in the Masméjean antiform most likely took place within the time interval
638 310–300 Ma and was followed by fast cooling as evidenced by the crystallization age
639 of 302.2 ± 3.7 Ma obtained on the fine-grained granite CHA-15-33 which arguably was
640 emplaced at shallow crustal levels in an already cooled basement.

641 The origin of water which fostered orthogneiss melting in the Masméjean antiform
642 remains elusive. Three sources may be considered: (i) meteoric water, percolating
643 down to the middle crust (Wickham and Taylor, 1985); (ii) metamorphic water
644 originating from the dehydration of deeper-lying or underthrust rocks (Sawyer, 2010),
645 (iii) magmatic water conveyed by hydrous melts (Weinberg and Hasalová, 2015).
646 Evidence for case (i) is very scarce in the geological record (see review in Weinberg
647 and Hasalová (2015)). Meteoric water infiltration down to depths of at least 3 km has
648 been documented within late orogenic shear zones elsewhere in the Variscan belt
649 (Dusséaux et al., 2019), but the melting pressure of >0.3 GPa (i.e. >11.5 km,
650 considering an average crustal density of 2700 kg.m^{-3}) inferred for the Masméjean
651 antiform (see section 5.2) makes such a fluid source unlikely. Considering case (ii), the
652 local thermal structure implies that crustal rocks underlying the Masméjean antiform
653 were also at supra-solidus conditions. Therefore, any metamorphic water released
654 upon heating at deeper crustal levels would have been instantaneously dissolved in a
655 granitic melt phase (Sawyer, 2010) and was thus unable to trigger melting at higher
656 crustal levels.

657 Ascent of hydrous granitic melts followed by diffusive transfer of magmatic H_2O along
658 chemical potential gradients at the emplacement site has been recently invoked to
659 explain water-fluxed melting (Tafur and Diener, 2020; Weinberg and Hasalová, 2015).
660 Villaros et al. (2018) argued that the intrusion of H_2O -rich granitic melts ($X_{\text{H}_2\text{O}} > 8$ wt.%)
661 generated at pressures in excess of 1 GPa was responsible for water-fluxed melting of
662 orthogneisses in the Velay dome. Such a model is plausible but difficult to prove based
663 on field observations in the Masméjean antiform. Indeed, the only evidence for the
664 percolation of granitic melts in the molten domain is represented by the discordant
665 leucosomes which, based on petrographic, geochemical and geochronological data,
666 most likely represent magmas derived from the local orthogneisses. However, it
667 remains possible that the “wet” granitic melts of Villaros et al. (2018) were emplaced
668 and induced melting below the current level of exposure (Fig. 14).

669 Water-fluxed orthogneiss melting could also result from the ingress of magmatic water
670 exsolved following the crystallization of the high-K diorites (“vaugnerites”) exposed in
671 the Masméjean antiform (and throughout the Velay dome). Indeed, the solidus of the
672 high-K diorites lies at about 800°C and their parental magmas were H₂O-rich (Montel
673 and Weisbrod, 1986; Parat et al., 2009) meaning that they were able to release a
674 supercritical H₂O-rich fluid at temperatures close to that inferred for the genesis of the
675 Masméjean granites–migmatites. The volumes of high-K diorites at the current
676 exposure level are limited but the presence of larger bodies at depth can be expected
677 since dioritic magmas are involved in the petrogenesis of the two nearby and coeval
678 granite batholiths (the c. 310 Ma Margeride and Borne plutons, Moyen et al. (2017)).

679 **5.5 The scant zircon record of water-fluxed orthogneiss melting**

680 As shown earlier (section 5.1), a striking feature of the U–Pb zircon dataset is the
681 complete absence of Variscan dates in the migmatites (bulk metatexites and
682 leucosomes). Some grain rims may have crystallized from the newly formed melt
683 fraction but this cannot be ascertained, either because these rims were too small for
684 analysis with routine dating procedures (specialized sample preparation and analytical
685 conditions allowing depth profiling, e.g. Campbell et al. (2005), may allow to date such
686 thin rims), or because they did not yield reliable U–Pb dates (see section 4). Only one
687 “low-Ba” granite sample, arguably corresponding to an anatectic melt extracted from
688 the migmatites, yielded three Variscan dates. In this section, we examine several
689 parameters that could explain the near lack of Variscan record of partial melting.

690 New zircon growth during anatexis is considered to exclusively result from the
691 dissolution of pre-existing protolith zircon grains and re-precipitation of new grains (or
692 overgrowths) from the melt phase (Kelsey et al., 2008; Mintrone et al., 2020; Watson,
693 1996; Yakymchuk and Brown, 2014). Experimental studies have determined the
694 solubility of Zr in silicate melts, which is positively correlated with temperature and also
695 depends on melt composition (Boehnke et al., 2013; Watson and Harrison, 1983).
696 Figures 15a and b depict the Zr solubility of orthogneiss-derived melts calculated based
697 on the major-element compositions retrieved from thermodynamic modelling.
698 Estimated values strongly depend on temperature but are independent of the bulk
699 water content of the protolith (Fig. 15b). The Zr budget of the orthogneiss is almost
700 entirely hosted in zircon, since the Zr concentration of the other mineral phases is

701 negligible (Bea et al., 2006). It is therefore possible to calculate the proportion of the
702 orthogneiss zircon budget that has to be dissolved in order to saturate the melt phase
703 (Fig. 15c, d). At low water content (Fig. 15c), a 20 vol.% melt fraction is only reached
704 for temperatures >850 °C and dissolution of 50 wt.% of the protolith zircon budget is
705 required to saturate the melt. In contrast, at high H₂O contents, the same melt fraction
706 is reached at much lower temperatures and is associated with very limited zircon
707 dissolution. Considering values relevant to the Masméjean case, for a bulk H₂O content
708 of 1.5 wt.%, the migmatite contains 18 vol.% melt at c. 700°C which can be saturated
709 by dissolution of as little as 7 wt.% of the orthogneiss zircon budget (Fig. 15d).

710 Additional constraints related to the low melting temperature may further hinder zircon
711 dissolution and suggest that this value of 7 wt.% should be considered as a maximum.
712 First, since biotite hosts substantial amounts of zircon grains in metamorphic rocks
713 (Clemens, 2003), its stability during low temperature melting would prevent protolith
714 zircon grains from physically interacting with the melt phase (Bea, 1996; Yakymchuk
715 and Brown, 2014). Second, the kinetics of zircon dissolution are expected to be
716 particularly sluggish at the range of melting temperatures inferred for the Masméjean
717 migmatites because this process is controlled by Zr diffusion in the melt (Watson,
718 1996). This author modelled the dissolution kinetics of zircon immersed in a granitic
719 melt at 650–700°C (relevant to the Masméjean migmatites) for a period of 1 Ma and
720 showed that for grains having radii of 90–120 µm (as for the Masméjean
721 orthogneisses), resorption would only affect a maximum 20 µm-thick rim. This
722 thickness would even be smaller if the Zr solubility calibration of Boehnke et al. (2013)
723 is considered.

724 The combined effects of all the processes outlined above collectively point to limited
725 zircon dissolution and hence new crystallization upon cooling. All processes relate to
726 the low-T, water-fluxed melting conditions of the orthogneiss source and readily explain
727 the scarcity of Variscan zircon grains or overgrowths in the Masméjean metatexites,
728 leucosomes and “low-Ba” granites. The fact that Variscan zircon dates were obtained
729 in the “high-Ba” granites (Fig. 10) is consistent with their inferred higher temperature
730 of formation, favoring higher solubility and faster diffusion of Zr in the melt; and the
731 release of biotite-occluded grains owing to biotite breakdown. The main implication of
732 these results is the very scant zircon record (<2% of the dated grains) of a yet non-

733 negligible melting event (which produced as a whole 10–20 vol.% melt on the local
734 scale).

735 In contrast with the zircon behavior documented in this study, Rubatto et al. (2009)
736 report extensive new zircon crystallization in Alpine migmatites formed through water-
737 fluxed melting at similar temperatures to those of the Masméjean antiform. We note
738 that the migmatites studied by Rubatto et al. (2009) developed at the expense of K-
739 feldspar-poor protoliths (amphibole-bearing metasediments, metatonalites) and that
740 melt-producing reactions, even in the presence of water, most likely involved the
741 breakdown of biotite forming peritectic hornblende (Berger et al., 2008). This suggests
742 that the elusive zircon record of water-fluxed melting as observed in our dataset is a
743 feature of peraluminous granitic protoliths.

744 **6. Conclusion**

745 Anatectic orthogneisses exposed in the core of the Masméjean antiform experienced
746 water-fluxed melting at 310–300 Ma during the late-orogenic evolution of the Variscan
747 belt. Anatexis took place at low temperature which produced leucogranitic melts with
748 low Sr–Ba, REE and Zr contents. Those melts were collected within the partially molten
749 domain (leucosomes, which retain a significant crystal load of source-inherited K-
750 feldspar) or extracted to higher crustal levels. At the same time, partial melting of the
751 orthogneisses at greater depths and higher temperature (~800°C) generated granitic
752 melts showing elevated Sr–Ba, REE and Zr contents. The external water which fluxed
753 melting of the orthogneisses most likely originated from H₂O-rich dioritic or granitic
754 magmas ponding in the middle crust. Zircon U-Pb dating overwhelmingly yielded the
755 crystallization ages of the granite protolith of the orthogneiss source (c. 542 Ma),
756 whereas Variscan zircon grains or overgrowths formed in response to the melting event
757 represent <2% of the data. This lack of Variscan zircon is intimately tied to the low
758 melting temperature which is shown to restrict the amount of zircon that can be
759 dissolved in the melt phase (in terms of melt solubility; dissolution kinetics; and physical
760 occlusion of zircon in restitic biotite). Our case study highlights that zircon has little
761 potential to record the timing and duration of low-T, water-fluxed anatectic episodes in
762 collisional orogens, especially those involving granitic (K-feldspar-rich) protoliths.

763

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774 **Figures and Tables captions**

775 Figure 1: (a) Sketch map showing exposed Variscan domains in Western Europe along
776 with the main inferred suture zones (MZ: Moldanubian Zone, RHZ: Rheno-Hercynian
777 Zone, STZ: Saxo-Thuringian Zone); (b) Geological map of the Eastern French Massif
778 Central centered on the Velay dome, adapted from Chantraine et al. (1996).

779 Figure 2: Geological map at scale 1/80,000 of the Masméjean unit and selected cross-
780 sections depicting the main lithotectonic features and the location of the dated
781 samples. Drawn after our field survey and Briand et al. (1994); Lapadu-Hargues
782 (1947); Négron (1979). The white stars outline available geochronological data on the
783 emplacement and maximum depositional age of the orthogneiss (C'17: Couzinié et al.
784 (2017)) and metasedimentary (C'19: Couzinié et al. (2019)) protoliths, respectively.

785 Figure 3: Representative photo(micro)graphs of migmatites from the core of the
786 Masméjean antiform, developed at the expense of augen gneisses: (a) metatexite with
787 segregated neosome in the form of leucocratic quartzofeldspathic layers (leucosome)
788 and biotite-rich melanocratic layers; (b) syn-anatectic folding of the metatexites
789 evidenced by the leucosome distribution within fold hinges, reflecting the collection of
790 anatectic liquids in low-strain domains; (c and d) 10 to 20 cm thick leucosomes
791 discordant with the anatectic foliation of metatexites; (e) paragneiss resister within
792 augen gneiss-derived migmatites and rounded intrusion of K-rich diorite ("vaugnerite")
793 intricately associated with coarse-grained Qz-feldspar domains; (f) contact between a
794 Ms-Bt melanosome and a biotite-bearing granitic leucosome showing thin quartz films
795 distributed along phase boundaries (cross-polarized light); (g) microstructure of a

796 biotite-bearing granitic leucosome with interstitial quartz locally showing cusped
797 contacts with adjacent feldspar grains and well-equilibrated feldspar–feldspar triple
798 junctions (cross-polarized light); (h) Bt–Sil leucosome (plane-polarized light).

799 Figure 4: The high-strain zone located at the interface between the anatectic domain
800 and the unmelted cover in the southwestern flank of the Masméjean antiform
801 (“Mylonitic Metamorphic Vellave Zone” of Bouilhol et al. (2006)): (a) intense folding in
802 Bt–Sil paragneisses, with the axial plane parallel to the regional S_n foliation; (b) top-to-
803 the NE C' shear bands within mylonitic augen gneisses; (c) granite sill injected within
804 Bt–Sil paragneisses and subsequently affected by post-magmatic top-to-the S
805 shearing at the brittle–ductile transition; (d) drag folds affecting Bt–Sil paragneisses
806 and intrusive granite veins indicating that top-to-the NE shearing was coeval to granitic
807 magmatism. The high-strain zone along the eastern flank of the antiform (Villefort
808 Shear Zone): (e) steeply-dipping granite sills concordant within the foliation of Bt–Sil
809 paragneisses; (f) quartzite layer showing a nearly vertical foliation and horizontal
810 mineral lineation marked by scarce Bt aggregates; (g) magmatic layering in a
811 decametric granite body evidenced by the accumulation of K-feldspar megacrysts; (h)
812 polished rock slab of sample CHA-15-41 with a mm-scale shear band (arrow) showing
813 marked grain-size reduction and subsequent alteration; (i) fractured K-feldspar
814 phenocryst and sericitization of the plagioclase-rich matrix within the same sample
815 (cross-polarized light).

816 Figure 5: Representative photo(micro)graphs of intrusive granites from the And–Bt–
817 Crd zone in the Masméjean antiform: (a, b) post-tectonic pink leucogranite dykes
818 cutting across the foliation of unmelted augen gneisses; (c) fine-grained porphyritic
819 granite sill concordant within the foliation of paragneisses themselves embedded as a
820 decametre-sized raft in augen gneisses; (d) texture of granite sample CHA-15-33
821 showing the planar orientation of biotite and muscovite and the occurrence of mm-
822 sized embayed quartz phenocrysts suggesting emplacement at shallow crustal levels
823 (cross-polarized light); (e) polished slab of granite sample CHA-15-29 collected from a
824 pink granite dyke.

825 Figure 6: Major element composition of the investigated metatexites, leucosomes and
826 granites. (a) P–Q, (b) B–A, (c) $K/(K+Na)$ vs. B and (d) B vs. $Mg/(Fe+Mg)$ cationic
827 classification diagrams of Debon and Le Fort (1988). The subdivisions of Villaseca et
828 al. (1998) are indicated in plot (b) along with the experimental melt compositions

829 generated by high-temperature melting of a peraluminous orthogneiss . The grey field
830 depicts the compositional range of local orthogneisses affiliated to the Velay
831 Orthogneiss Formation. Data from Couzinié et al. (2017) and this study.

832 Figure 7: Rare Earth Elements patterns normalized to the chondrite values (Boynnton,
833 1984) and spider diagrams for key trace element normalized to a representative augen
834 gneiss composition (SC-14-33). The grey fields depict the compositional range of local
835 augen gneisses (data from Couzinié et al. (2017)).

836 Figure 8: Representative cathodoluminescence images of zircon grains from
837 metatexite, leucosome and granite samples of the Masméjean unit. The locations of
838 laser spots are indicated along with the spot name (zXX). The letters a, b, c refer to
839 the analytical session number (see Supplementary Text and Tables 2 to 5). The
840 corresponding $^{206}\text{Pb}/^{238}\text{U}$ dates are quoted with $\pm 2\sigma$ uncertainty, in Ma. All displayed
841 analyses are concordant at >95% (except those in italic).

842 Figure 9: Tera-Wasserburg diagrams ($^{238}\text{U}/^{206}\text{Pb}$ vs. $^{207}\text{Pb}/^{206}\text{Pb}$) for zircon grains from
843 the orthogneiss, metatexites and leucosome samples. Error ellipses and dates are
844 displayed at 2σ level of uncertainty. Analyses were not corrected from common Pb.
845 Yellow ellipses are those considered for Concordia date calculations.

846 Figure 10: Tera-Wasserburg diagrams ($^{238}\text{U}/^{206}\text{Pb}$ vs. $^{207}\text{Pb}/^{206}\text{Pb}$) for zircon grains
847 from granite samples. Error ellipses and dates are displayed at 2σ level of uncertainty.
848 Analyses were not corrected from common Pb except those with dotted error ellipses.
849 Yellow ellipses are those considered for Concordia date calculations.

850 Figure 11: Histogram showing the $^{206}\text{Pb}/^{238}\text{U}$ dates measured on zircon grains from
851 leucosomes and granites. Bin width is 15 Ma. Available zircon $^{206}\text{Pb}/^{238}\text{U}$ dates
852 (concordance: 95–105%) for the Velay Orthogneiss Formation (in red) and regional
853 metasediments (in blue, N=107 and 445, respectively) are from Chelle-Michou et al.
854 (2017), Couzinié et al. (2019); Couzinié et al. (2017) and this study, represented as
855 Kernel Density Estimates. The latter were calculated using the DensityPlotter program
856 of Vermeesch (2012) with a bandwidth set at 15 Ma.

857 Figure 12: (a, b) P–T and T– $X_{\text{H}_2\text{O}}$ pseudosections calculated using a representative
858 orthogneiss composition (SC-14-33, from Couzinié et al. (2017)). White lines are melt
859 modes (with numbers indicating melt vol.%). Initial water content of 0.53 wt.% for the

860 P–T pseudosection was adjusted to avoid subsolidus water saturation at 0.5 GPa.
861 Mineral abbreviations after Whitney and Evans (2009). The white dots connected by
862 black dashed lines refer to the location in the P–T– X_{H_2O} space of the melts for which
863 equilibrium Ba–Sr–Rb–Nb compositions have been computed. These calculations rely
864 on the thermodynamically-constrained phase proportions and a set of partition
865 coefficients for felsic systems compiled by Laurent et al. (2013) and listed in
866 Supplementary Table S6. (c, d) Comparison of respective melt productivities (in vol.%)
867 for representative orthogneiss and paragneiss samples (samples SC-14-33 and -41,
868 respectively) at $T < 800^\circ\text{C}$ and variable water contents (for a pressure of 0.5 GPa). The
869 pseudosection for SC-14-41 (whole-rock composition available in Supplementary
870 Table 1) was calculated using the same procedure and solution models as mentioned
871 in section 3.3.

872 Figure 13: Petrogenesis of the Masméjean antiform leucosomes and granites.
873 Comparison between leucosomes/granites and calculated melt compositions: (a)
874 CIPW-normative ternary diagram; (b) Rb/Sr vs. Ba diagram; (c) Nb vs. Ba diagram.
875 The Ba–Rb–Sr composition of K-feldspar from the local augen gneisses is from Négron
876 (1979) while that of plagioclase is estimated from Bea et al. (1994).

877 Figure 14: Sketch cross-section across the southern Velay mid-crust (c. 10–15 km
878 depth) at c. 305 Ma, right before the onset of doming. The Mylonitic Metamorphic
879 Vellave Zone (MMVZ) corresponds to a high-strain shear zone developed at the
880 interface between the anatectic and unmelted domains. Granite magmas formed at the
881 expense of the orthogneisses were either emplaced within the MMVZ or extracted to
882 higher crustal levels where they crop out as dykes. In the partially molten domain,
883 water-fluxed melting possibly resulted from (i) diffusive transfer of magmatic H_2O along
884 chemical potential gradients between orthogneisses and water-rich granitic melts
885 ($X_{H_2O} > 8$ wt.%) generated at pressures in excess of 1 GPa (>35 km depth, considering
886 and average crustal density of $2700 \text{ kg}\cdot\text{m}^{-3}$, left part of the figure); (ii) the exsolution of
887 magmatic water following the crystallization of high-K dioritic bodies (“vaugnerites”,
888 right part of the figure).

889 Figure 15: (a) and (b): calculated maximal Zr concentration (in ppm) in the melt phase
890 for a fully hydrated (at 0.5 GPa) orthogneiss composition (SC-14-33) in the P–T space
891 and along a T– X_{H_2O} section set at 0.5 GPa, based on the solubility equation of Boehnke
892 et al. (2013). The Zr concentration of the stoichiometric zircon was set at 497644 ppm

893 and the average Zr content of the protolith at 115 ppm (average value for the local
894 augen gneisses, data from Couzinié et al. (2017)). (c) and (d): calculated proportion
895 (white lines, in wt.%) of the orthogneiss zircon budget to be dissolved in order to
896 saturate the melt phase. Black dotted lines and bold numbers are melt modes in vol.%.

897 Table 1: Chemical composition of augen gneiss sample SC14-33 used in the modelling
898 (in wt.% for major oxides, in ppm for trace elements). H₂O content was adjusted to
899 reach full hydration (but no water saturation) at the solidus for a pressure of 0.5 GPa.

900

901 Supplementary Figure SF1: CL image, location of laser spots and associated
902 ²⁰⁶Pb/²³⁸U dates (discordant if in italics) for orthogneisses CHA-15-52 and -35 and
903 metatexites CHA-15-17 and -47. «ns» means that no reliable date was retrieved.

904 Supplementary Figure SF2: CL image, location of laser spots and associated
905 ²⁰⁶Pb/²³⁸U dates (discordant if in italics) for leucosome CHA-15-48 and metatexite
906 CHA-15-65M. «ns» means that no reliable date was retrieved.

907 Supplementary Figure SF3: CL image, location of laser spots and associated
908 ²⁰⁶Pb/²³⁸U dates (discordant if in italics) for leucosomes CHA-15-63, 65L and -43. «ns»
909 means that no reliable date was retrieved.

910 Supplementary Figure SF4: CL image, location of laser spots and associated
911 ²⁰⁶Pb/²³⁸U dates (discordant if in italics) for granites CHA-15-19, -24, -34 and -41. «ns»
912 means that no reliable date was retrieved.

913 Supplementary Figure SF5: CL image, location of laser spots and associated
914 ²⁰⁶Pb/²³⁸U dates (discordant if in italics) for granites CHA-15-25, -33, -51, -49. «ns»
915 means that no reliable date was retrieved.

916

917 Supplementary Table S1: Whole-rock major and trace element compositions obtained
918 on the Masméjean antiform samples.

919 Supplementary Table S2: Results of LA-ICP-MS U–Pb analyses of zircon standards
920 performed during the sessions at GUF, 2nd and 3rd of May, 2016

921 Supplementary Table S3: Results of LA-ICP-MS U–Pb analyses of zircon standards
922 performed during the session at ETH Zürich, 12th of February, 2017

923 Supplementary Table S4: Results of LA-ICP-MS zircon U–Pb analyses performed
924 during the sessions at GUF, 2nd and 3rd of May, 2016

925 Supplementary Table S5: Results of LA-ICP-MS zircon U–Pb analyses performed
926 during the session at ETH Zürich, 12th of February, 2017

927 Supplementary Table S6: Set of partition coefficients for felsic systems (compiled by
928 Laurent et al. (2013)).

929

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Table 1: Chemical composition of augen gneiss sample SC14-33 used in the modelling (in wt.% for major oxides, in ppm for trace elements). H₂O content was adjusted to reach full hydration (but no water saturation) at the solidus for a pressure of 0.5 GPa.

SiO ₂	Al ₂ O ₃	FeO	CaO	MgO	Na ₂ O	K ₂ O	TiO ₂	H ₂ O	Ba	Nb	Rb	Sr
72.56	14.90	2.02	0.76	0.40	3.12	5.52	0.19	0.53	576	8	184	87

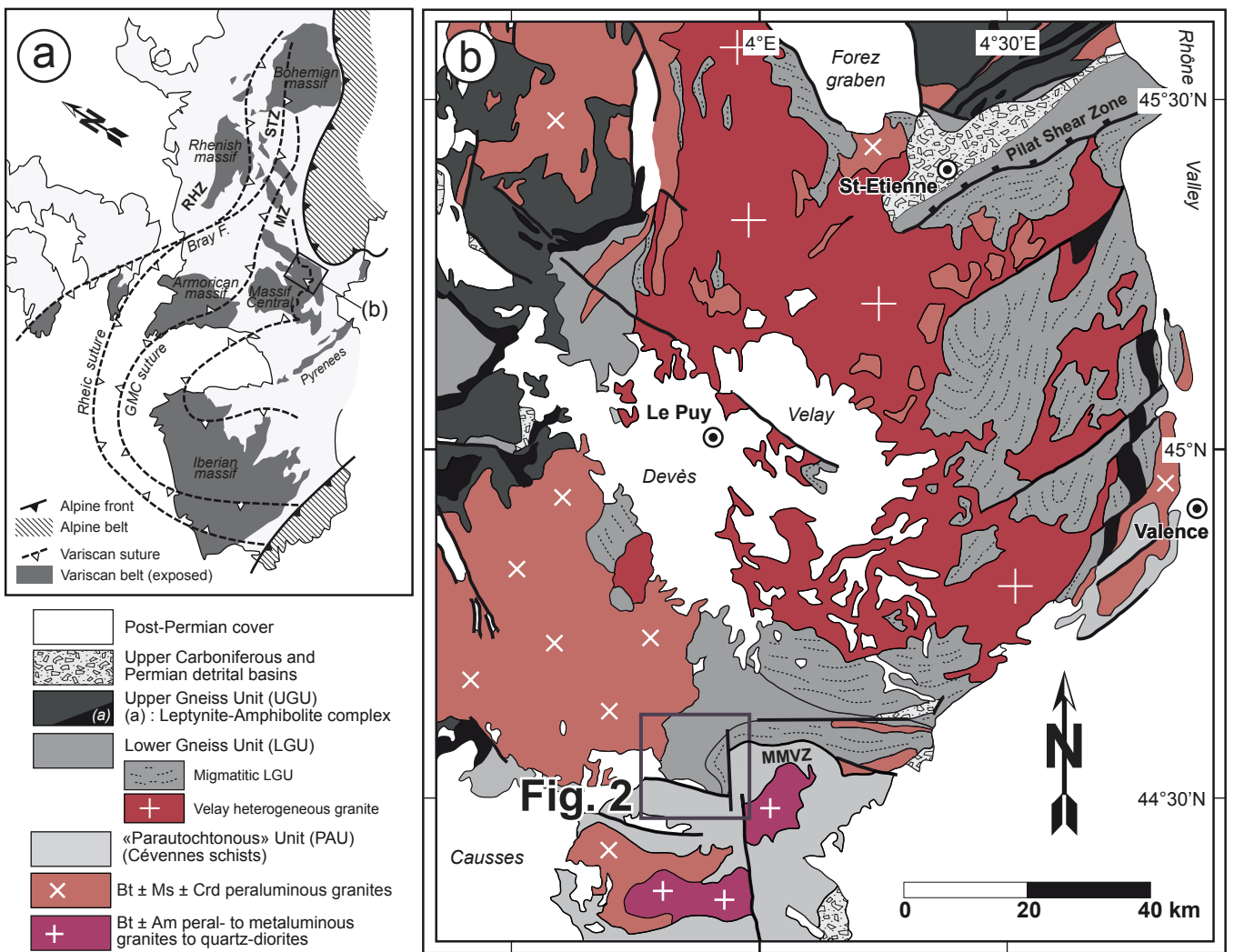


Figure 1: (a) Sketch map showing the repartition of exposed Variscan domains in Western Europe along with the main inferred suture zones (MZ: Moldanubian Zone, RHZ: Rheno-Hercynian Zone, STZ: Saxo-Thuringian Zone); (b) Geological map of the Eastern French Massif Central centered on the Velay dome, adapted from Chantraine et al. (1996).

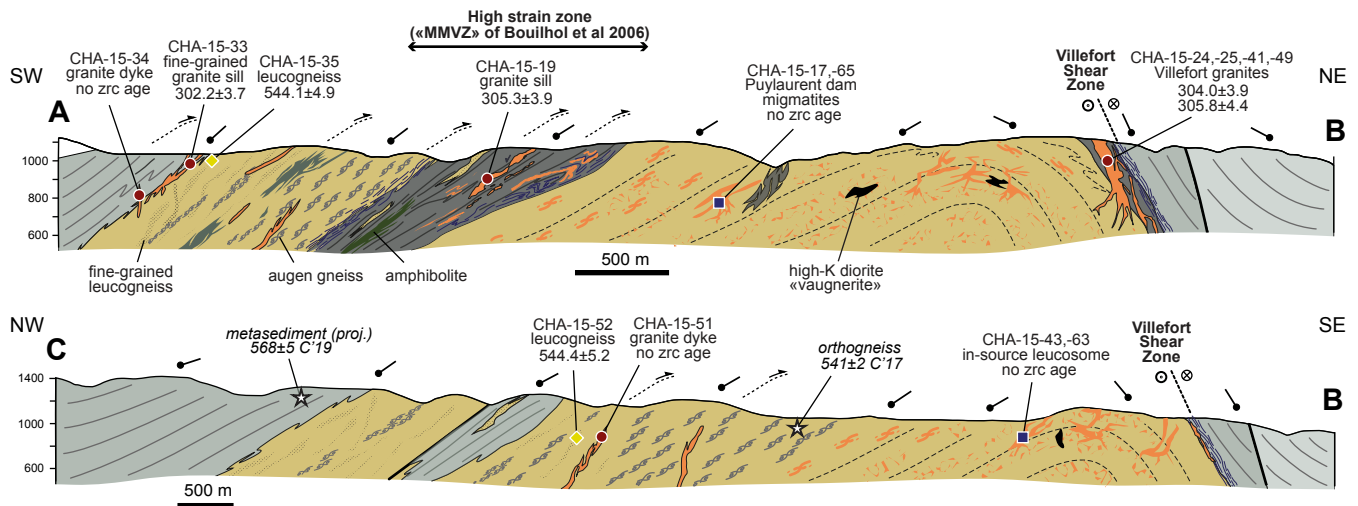
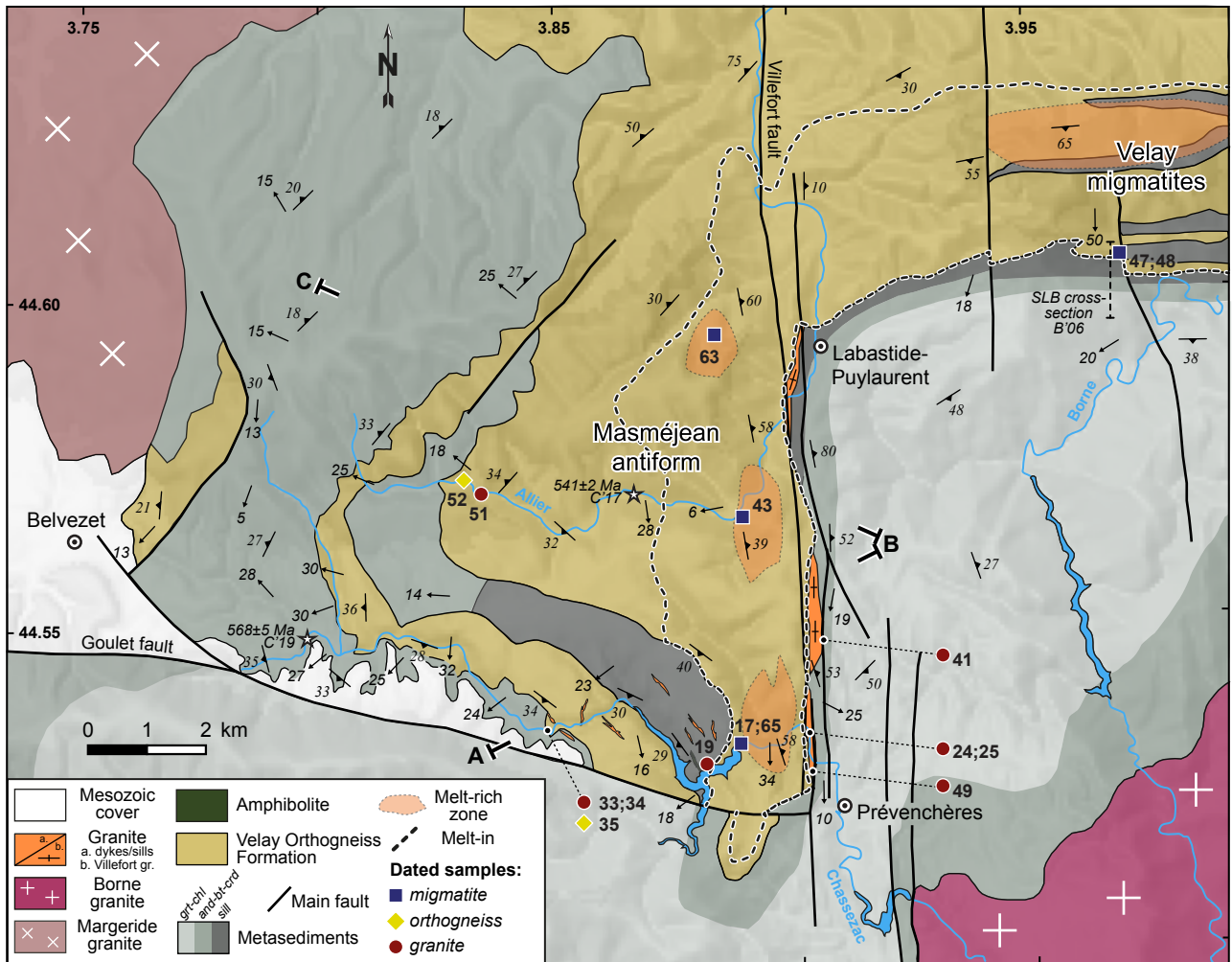


Figure 2: Geological map at scale 1/80,000 of the Masméjean unit and selected cross-sections depicting the main lithotectonic features and the location of the dated samples. Drawn after our field survey and the works of Briand et al. (1994); Lapaduhargues (1947); Négron (1979). The white stars outline available geochronological data on the emplacement and maximum depositional age of the orthogneiss (C'17: Couzinié et al. (2017)) and metasedimentary (C'19: Couzinié et al. (2019)) protoliths, respectively.

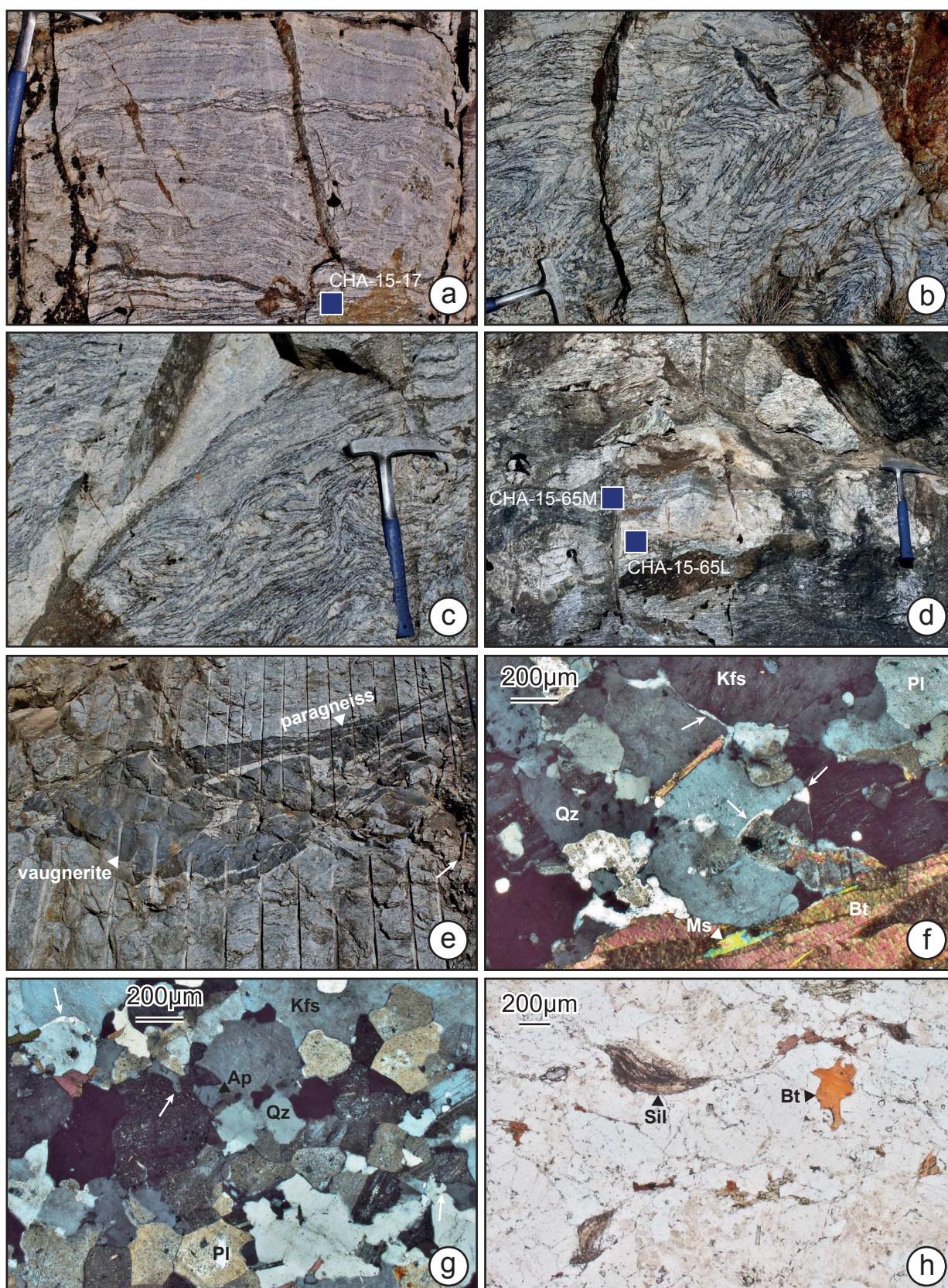


Figure 3: Representative photo(micro)graphs of migmatites from the core of the Masméjean antiform, developed at the expense of augen gneisses: (a) metatexite with segregated neosome in the form of leucocratic quartzofeldspathic layers (leucosome), locally folded, and biotite-rich melanocratic layers; (b) syn-anatectic folding of the metatexites evidenced by the leucosome distribution within fold hinges, reflecting the collection of anatectic liquids in low-strain domains; (c and d) 10 to 20 cm thick leucosomes discordant with the anatectic foliation of metatexites; (e) paragneiss resister within augen gneiss-derived migmatites and rounded intrusion of K-rich diorite ("vaugnerite") intricately associated with coarse-grained Qz-feldspar domains; (f) contact between a muscovite-biotite melanosome and a biotite-bearing granitic leucosome showing thin quartz films distributed along phase boundaries (cross-polarized light); (g) microstructure of a biotite-bearing granitic leucosome with interstitial quartz locally showing cusped contacts with adjacent feldspar grains and well-equilibrated feldspar-feldspar triple junctions (cross-polarized light); (h) biotite- and sillimanite-bearing leucosome (plane-polarized light).

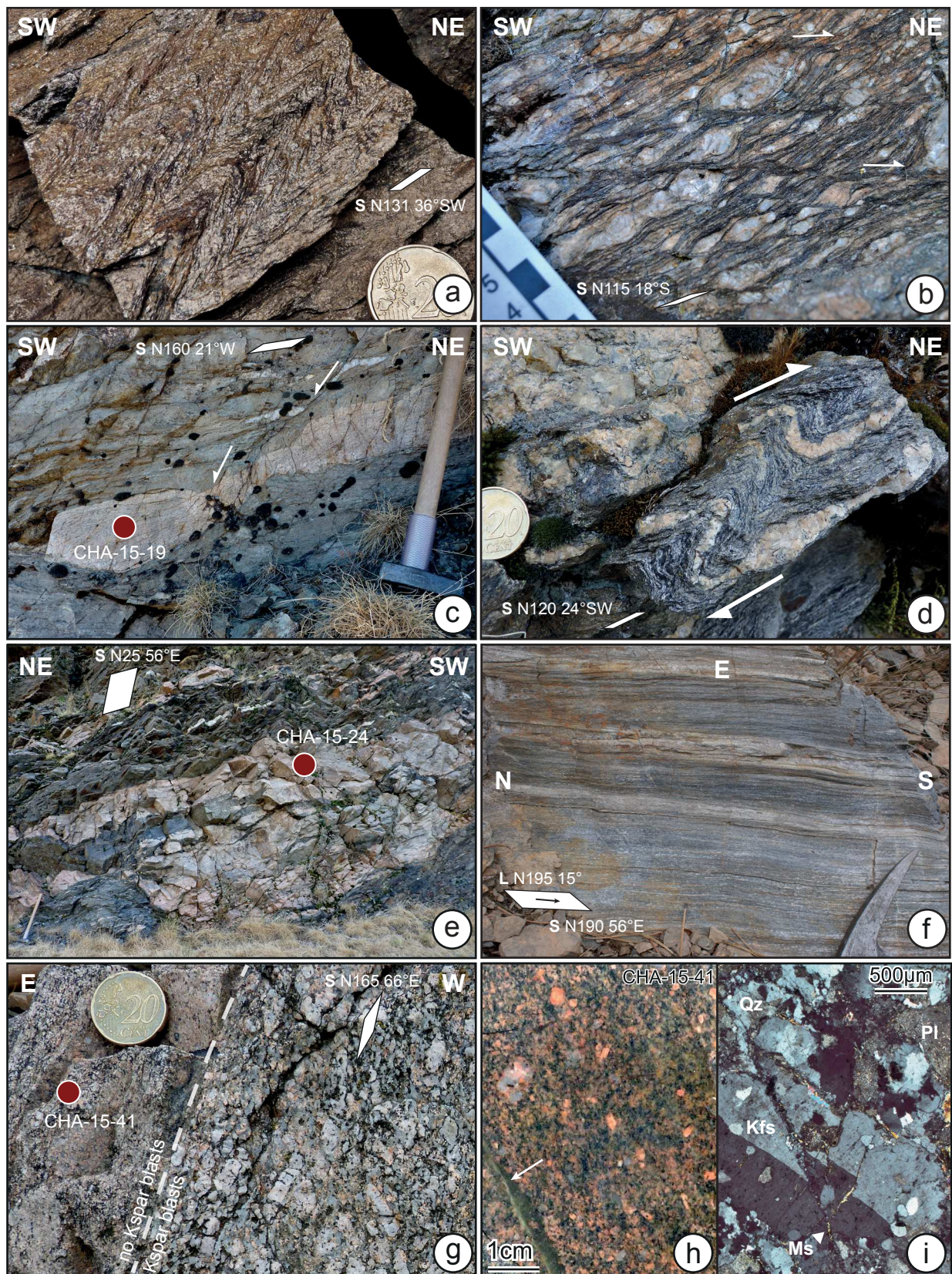


Figure 4: The high-strain zone located at the interface between the anatexitic domain and the unmolten cover in the southwestern flank of the Masméjean antiform (“Mylonitic Metamorphic Vellave Zone” of Bouilhol et al. (2006)): (a) intense folding in Bt–Sil-bearing paragneisses, with the axial plane parallel to the regional S_n foliation; (b) top-to-the NE C' shear bands within mylonitic augen gneisses; (c) granite sill injected within Bt–Sil-bearing paragneisses and subsequently affected by post-magmatic top-to-the S shearing at the brittle–ductile transition; (d) drag folds affecting Bt–Sil-bearing paragneisses and intrusive granite veins indicating that top-to-the NE shearing was coeval to granitic magmatism. The high-strain zone along the eastern flank of the antiform (Villefort Shear Zone): (e) steeply-dipping granite sills concordant within the foliation of Bt–Sil-bearing paragneisses; (f) quartzite layer showing a nearly vertical foliation and horizontal mineral lineation marked by scarce Bt aggregates; (g) magmatic layering in a decametric granite body evidenced by the accumulation of K-feldspar megacrysts; (h) polished rock slab of sample CHA-15-41 with a millimeter-scale shear band (arrow) showing marked grain-size reduction and subsequent alteration; (i) fractured K-feldspar phenocryst and sericitization of the plagioclase-rich matrix within the same sample (cross-polarized light).

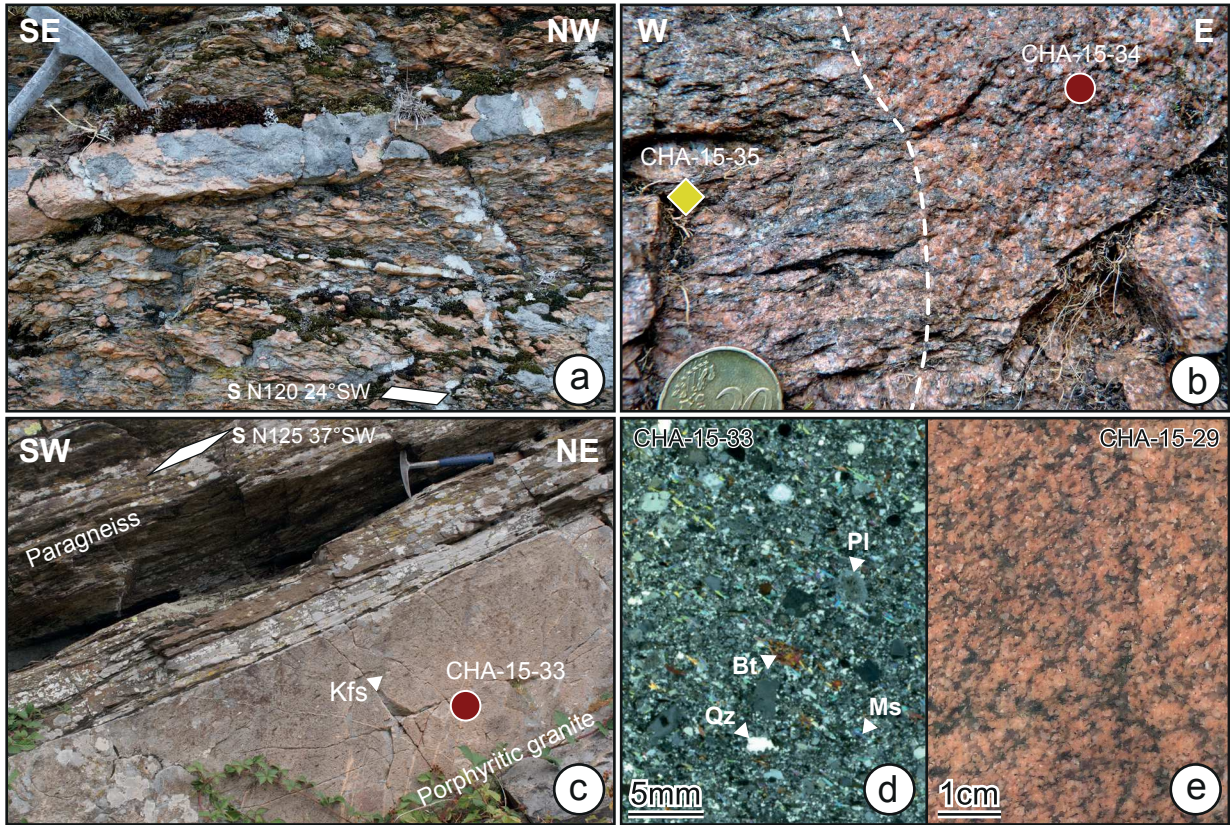


Figure 5: Representative photo(micro)graphs of intrusive granites from the And–Bt–Crd zone in the Masméjean antiform: (a, b) post-tectonic “pink” leucogranite dykes cutting across the foliation of unmolten augen gneisses; (c) fine-grained porphyritic granite sill concordant within the foliation of paragneisses themselves embedded as a decametric raft in augen gneisses; (d) texture of granite sample CHA-15-33 showing the planar orientation of biotite and muscovite and the occurrence of millimetric embayed quartz phenocrysts suggesting emplacement at shallow crustal levels (cross-polarized light); (e) polished slab of granite sample CHA-15-29 collected from a pink granite dyke.

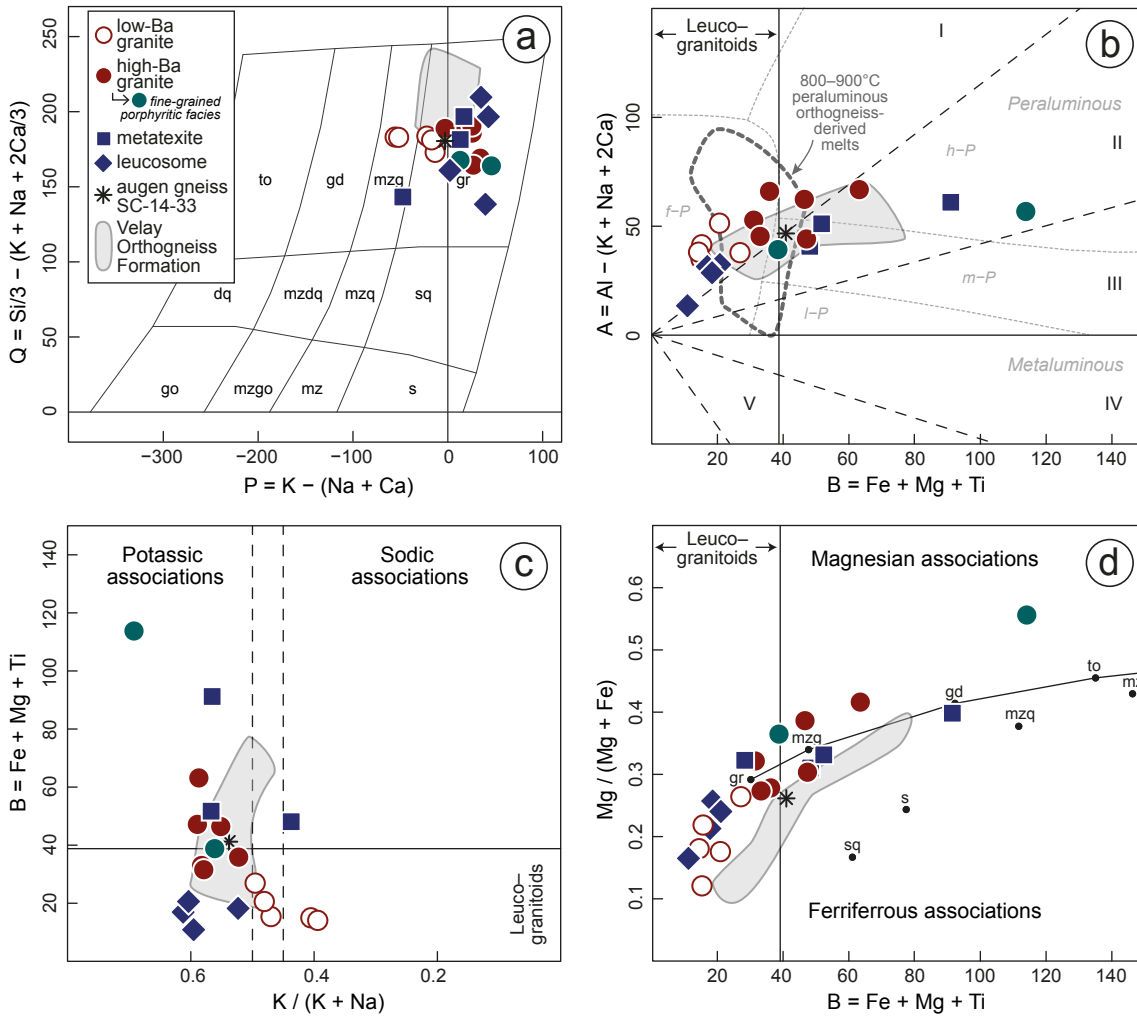


Figure 6: Major element composition of the investigated metatexites, leucosomes and granites from the Masméjean unit. (a) P–Q, (b) B–A, (c) K/(K+Na) vs. B and (d) B vs. Mg/(Fe+Mg) cationic classification diagrams of Debon and Le Fort (1988). The subdivisions of Villaseca et al. (1998) are indicated in plot (b) along with the experimental melt compositions generated by high-temperature melting of a peraluminous orthogneiss (García-Arias et al., 2012). The grey field depicts the compositional range of local orthogneisses affiliated to the Velay Orthogneiss Formation. Data from Couzinié et al. (2017) and this study.

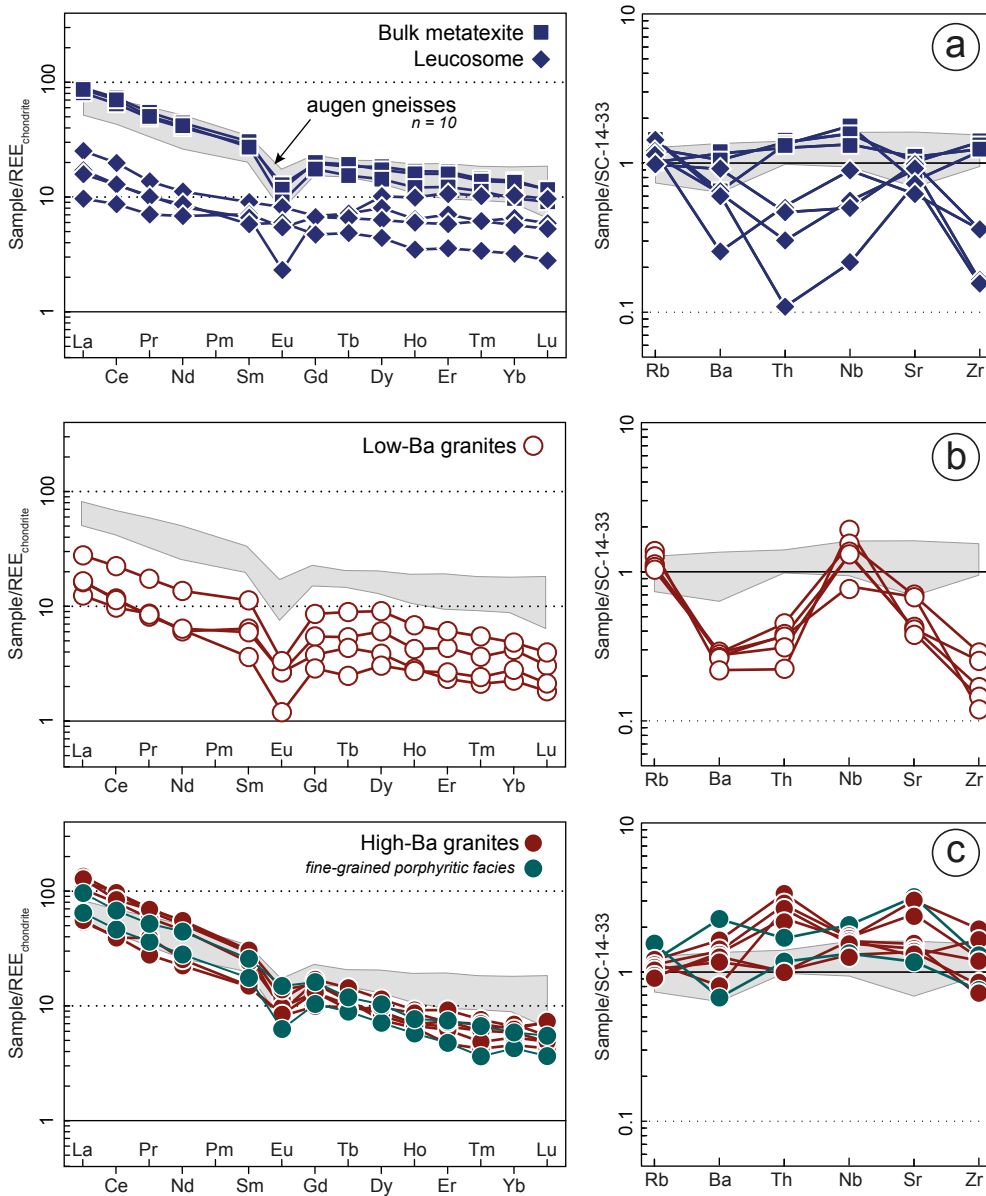


Figure 7: Rare Earth Elements patterns normalized to the chondrite values (Boynton, 1984) and spider diagrams for key trace element normalized to a representative augen gneiss composition (SC-14-33). The grey fields depicts the compositional range of local augen gneisses (data from Couzinié et al. (2017)).

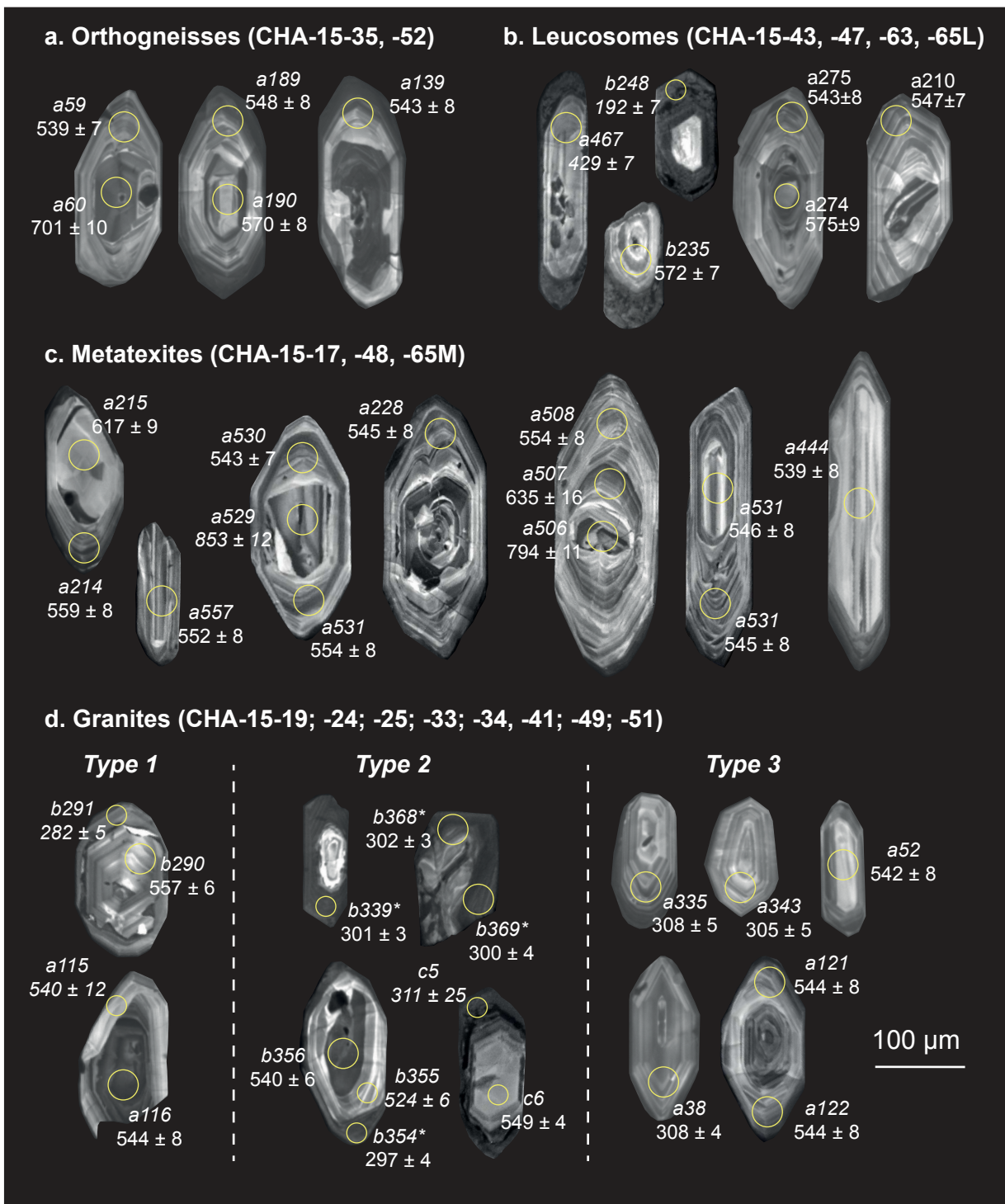


Figure 8: Representative cathodoluminescence images of zircon grains from metatexite, leucosome and granite samples of the Masméjean unit. The locations of laser spots are indicated along with the spot name (zXX). The letters a, b, c refer to the analytical session number (see Supplementary Text and Table). The corresponding $^{206}\text{Pb}/^{238}\text{U}$ dates are quoted with $\pm 2\sigma$ uncertainty, in Ma. All displayed analyses are concordant at $>95\%$ (except those in italic).

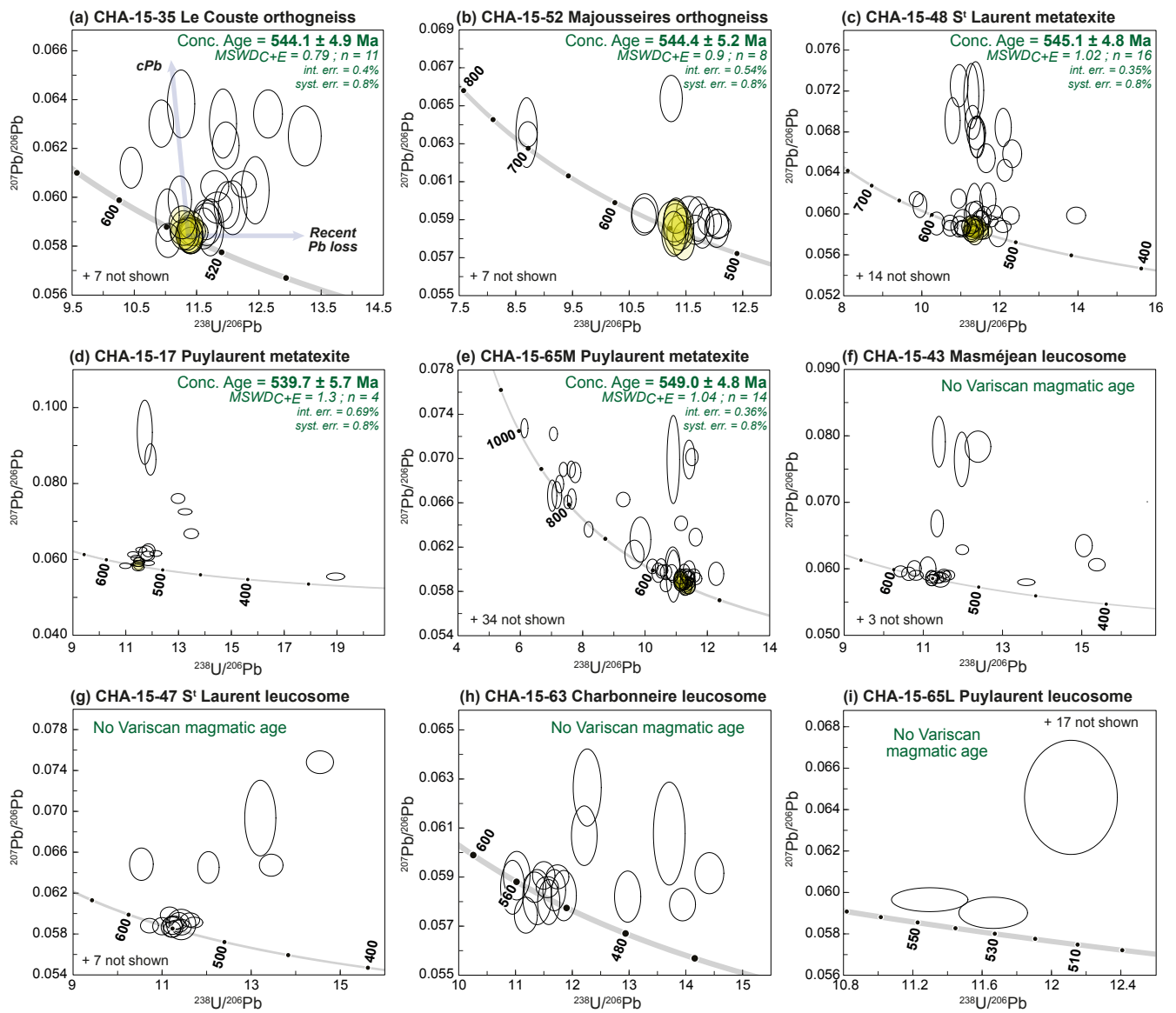


Figure 9: Tera-Wasserburg diagrams ($^{238}\text{U}/^{206}\text{Pb}$ vs. $^{207}\text{Pb}/^{206}\text{Pb}$) for zircon grains from the orthogneiss, metatexites and leucosome samples. Error ellipses and dates are displayed at 2σ level of uncertainty. Yellow ellipses are those considered for Concordia date calculations.

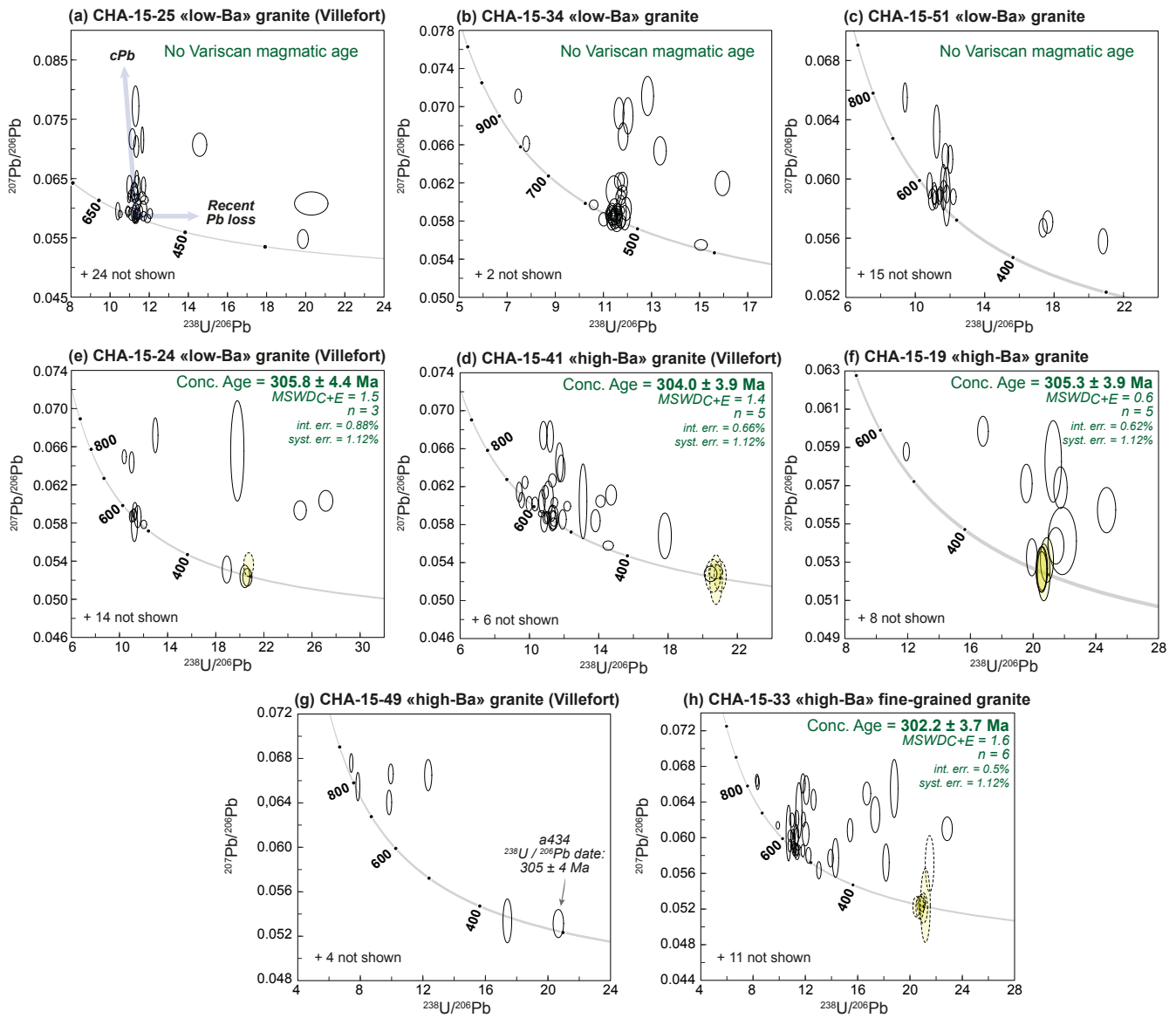


Figure 10: Tera-Wasserburg diagrams ($^{238}\text{U}/^{206}\text{Pb}$ vs. $^{207}\text{Pb}/^{206}\text{Pb}$) for zircon grains from granite samples. Error ellipses and dates are displayed at 2σ level of uncertainty. Yellow ellipses are those considered for Concordia date calculations.

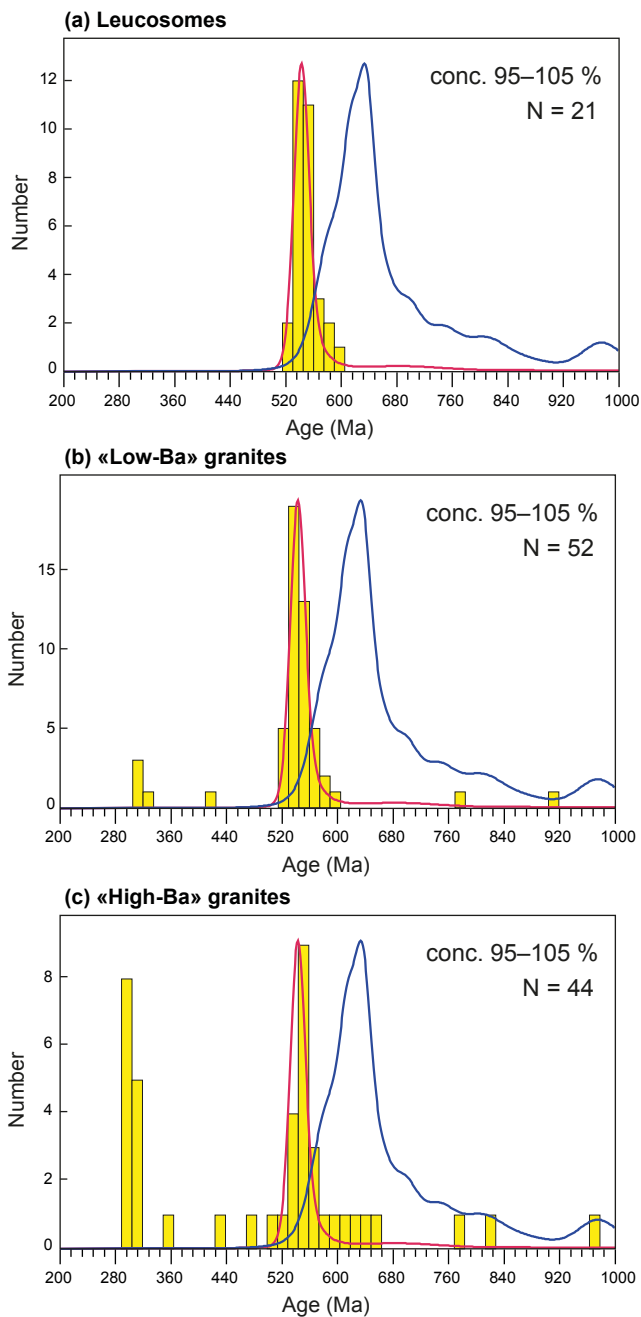


Figure 11: Histogram showing the $^{206}\text{Pb}/^{238}\text{U}$ dates measured on zircon grains from leucosomes and granites. Bin width is 15 Ma. Available zircon $^{206}\text{Pb}/^{238}\text{U}$ dates (concordance: 95–105%) for the Velay Orthogneiss Formation (in red) and regional metasediments (in blue, N=107 and 445, respectively) are from Chelle-Michou et al. (2017), Couzinié et al. (2019); Couzinié et al. (2017) and this study, represented as Kernel Density Estimates. The latter were calculated using the DensityPlotter program of Vermeesch (2012) with a bandwidth set at 15 Ma.

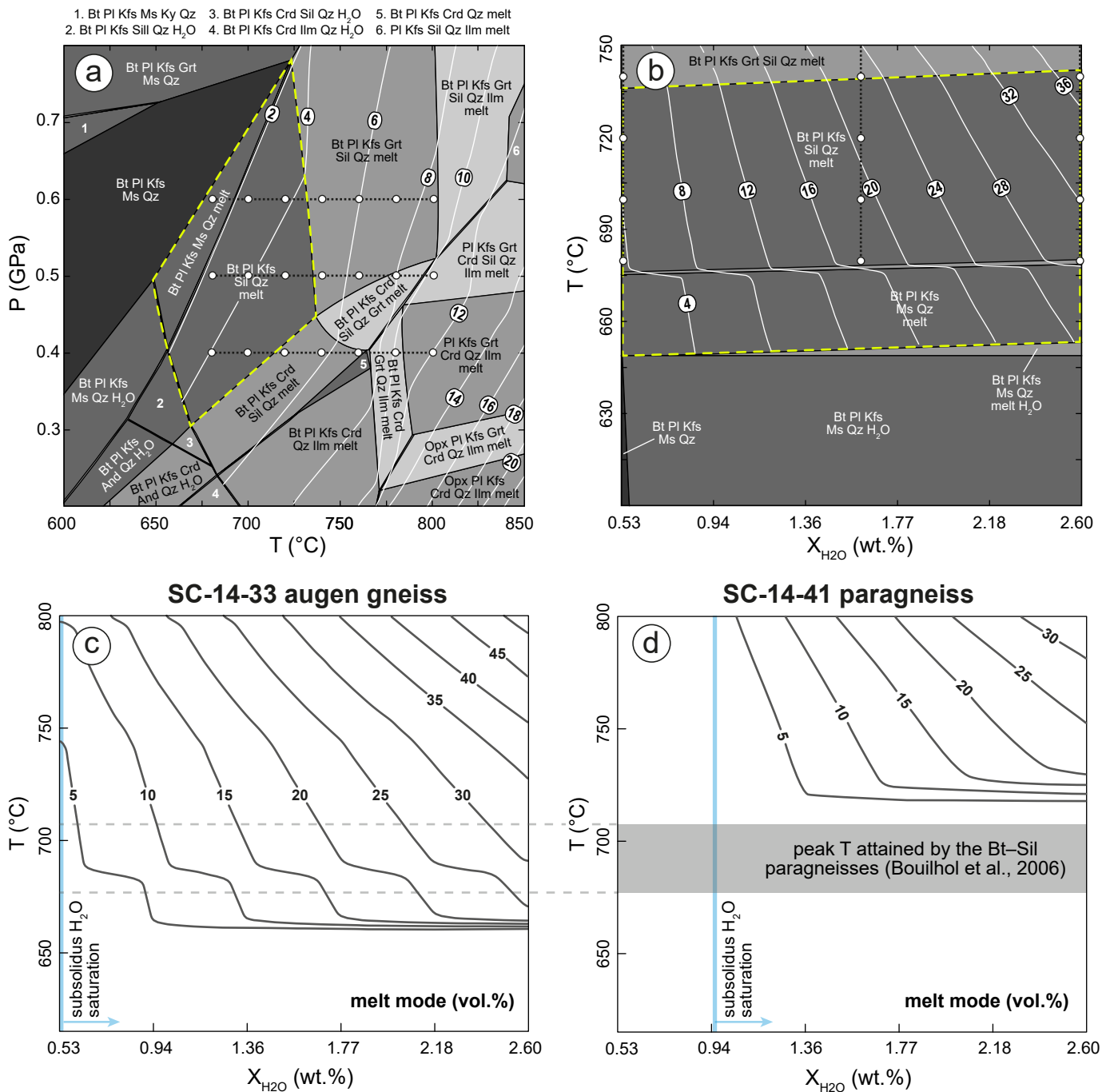


Figure 12: (a, b) P–T and T–X_{H2O} pseudosections calculated using a representative orthogneiss composition (SC-14-33, from Couzinié et al. (2017)). White lines are melt modes (with numbers indicating melt vol.%). Initial water content of 0.53 wt.% for the P–T pseudosection was adjusted to avoid subsolidus water saturation at 0.5 GPa. Mineral abbreviations after Whitney and Evans (2009). The white dots connected by black dashed lines refer to the location in the P–T–X_{H2O} space of the melts for which equilibrium Ba–Sr–Rb–Nb compositions have been computed. These calculations rely on the thermodynamically-constrained phase proportions and a set of partition coefficients for felsic systems compiled by Laurent et al. (2013) and listed in Supplementary Table S6. (c, d) Comparison of respective melt productivities (in vol.%) for representative augen gneiss and paragneiss samples (samples SC-14-33 and -41, respectively) at T < 800°C and variable water contents (for a pressure of 0.5 GPa). The pseudosection for SC-14-41 (whole-rock composition available Supplementary Table 1) was calculated using the same procedure and solution models as mentioned in section 3.3.

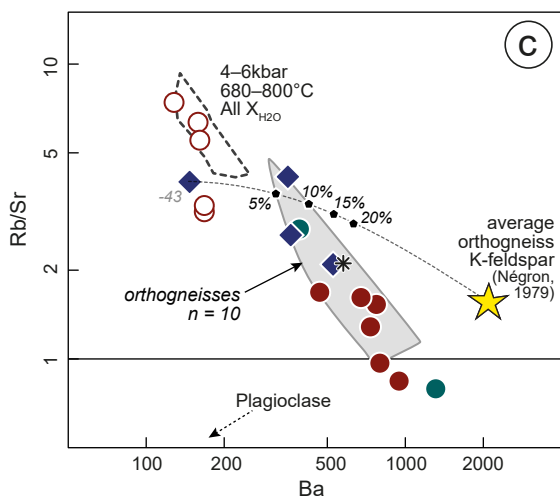
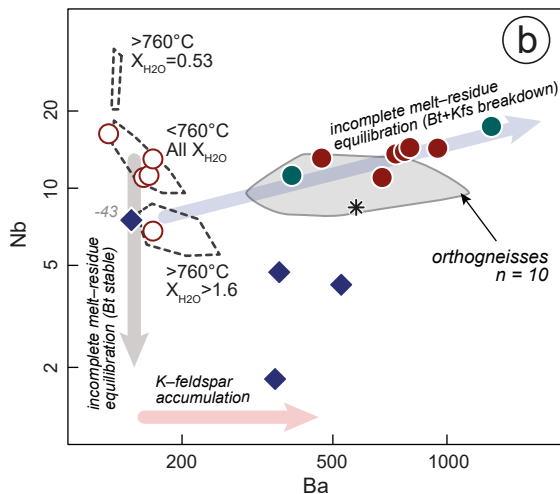
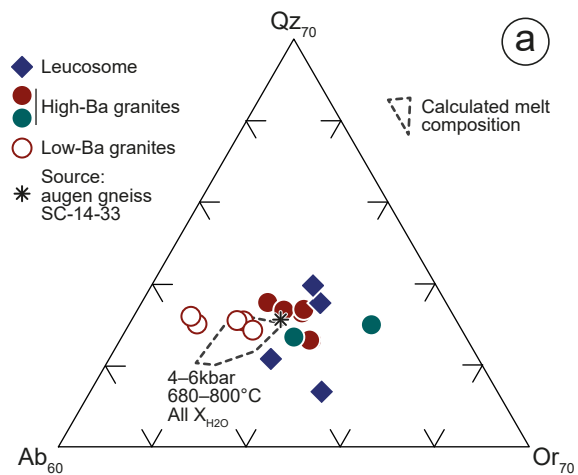


Figure 13: Petrogenesis of the Masméjean antiform leucosomes and granites. Geochemical comparison between leucosomes/granites and calculated melt compositions: (a) CIPW-normative ternary diagram; (b) Rb/Sr vs. Ba diagram; (c) Nb vs. Ba diagram. The Ba–Rb–Sr ortho composition of K-feldspar from the local augen gneisses is from Négron (1979) while that of plagioclase is estimated from Bea et al. (1994).

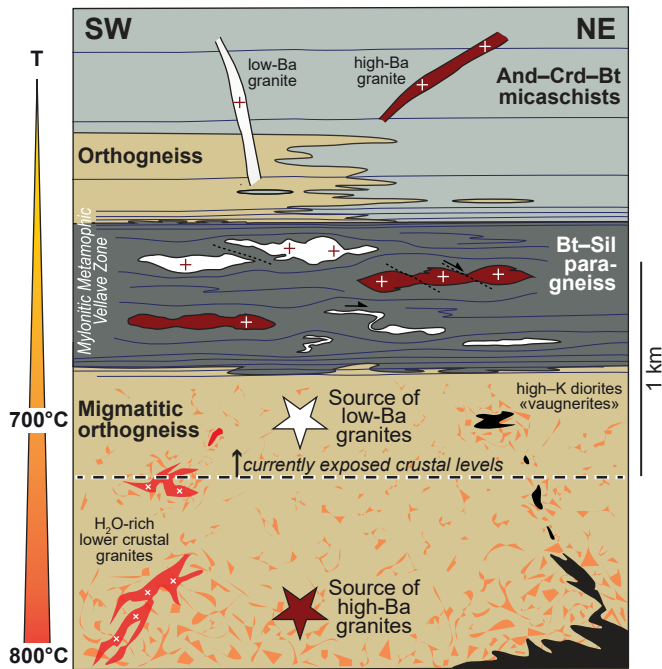


Figure 14: Sketch cross-section through the southern Velay mid-crust (c. 10–15 km depth) at c. 305 Ma, right before the onset of doming. The Mylonitic Metamorphic Vellave Zone (MMVZ) corresponds to a high-strain shear zone developed at the interface between the anatexitic and unmolten domains. Granite magmas formed at the expense of the orthogneisses were either emplaced within the MMVZ or extracted to higher crustal levels where they crop out as dykes. In the partially molten domain, water-fluxed melting possibly resulted from (i) diffusive transfer of magmatic H₂O along chemical potential gradients between orthogneisses and water-rich granitic melts ($X_{H_2O} > 8$ wt.%) generated at pressures in excess of 1 GPa (>35 km depth, considering average crustal density of 2700 kg.m⁻³, left part of the figure); (ii) the exsolution of magmatic water following the crystallization of high-K dioritic bodies (“vaugnerites”, right part of the figure).

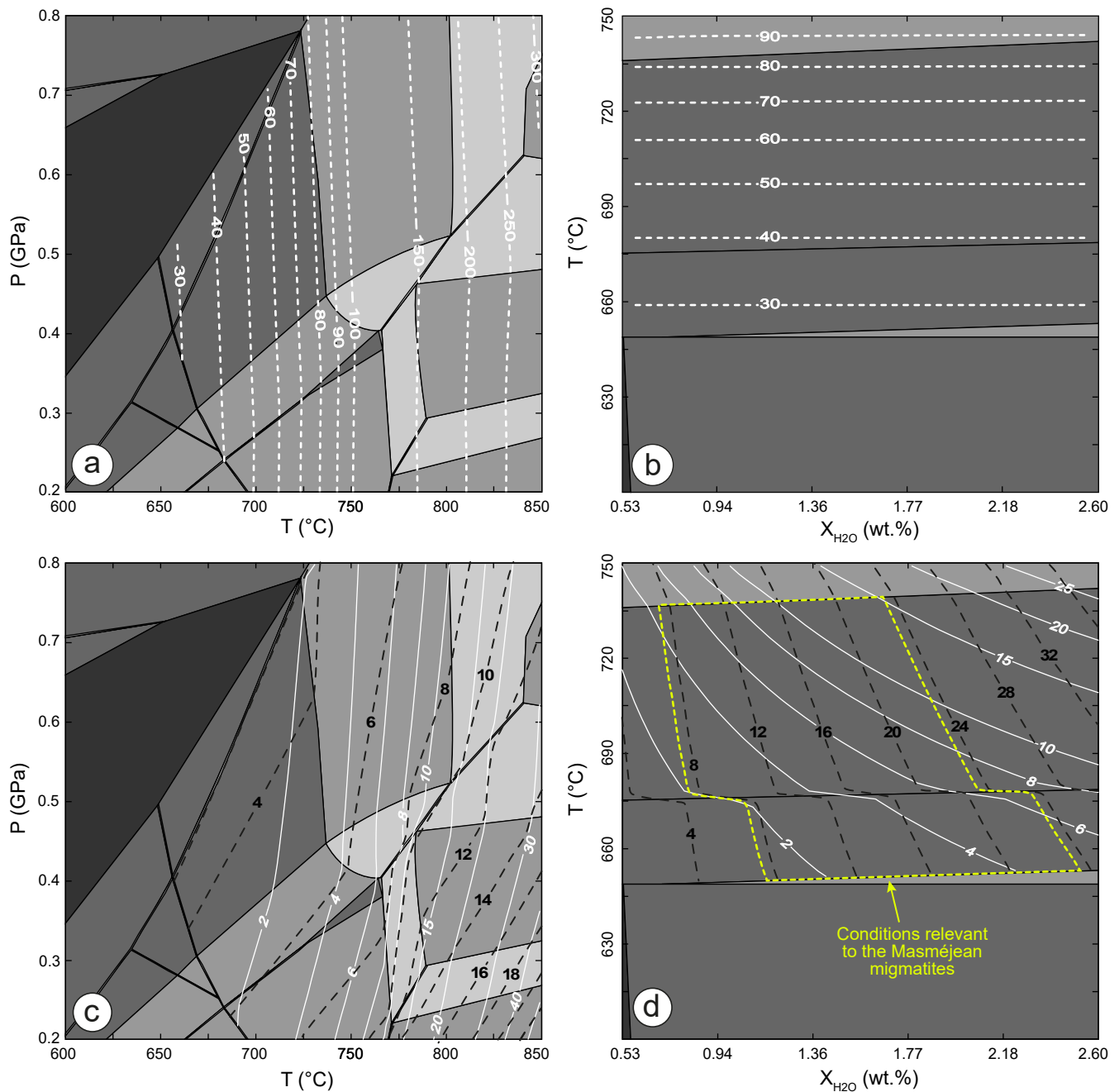


Figure 15: (a) and (b): calculated maximal Zr concentration (in ppm) in the melt phase for a fully hydrated (at 0.5 GPa) orthogneiss composition (SC-14-33) in the P–T space and along a T– X_{H_2O} section set at 0.5 GPa, based on the solubility equation of Boehnke et al. (2013). The Zr concentration of the stoichiometric zircon was set at 497644 ppm and the average Zr content of the protolith at 115 ppm (average value for the local augen gneisses, data from Couzinié et al. (2017)). (c) and (d): calculated proportion (white lines, in wt.%) of the orthogneiss zircon budget to be dissolved in order to saturate the melt phase. Black dotted lines and bold numbers are melt modes in vol.%.