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1 **Experimental crystallization of the Beauvoir granite as a model for** 2 **the evolution of Variscan rare metal magmas**

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10 11 **ABSTRACT**

12 New experiments on the strongly peraluminous, Li-, P- and F-rich Beauvoir granite (Massif
13 Central, France) provide a basis for understanding crystallization and chemical fractionation
14 in Variscan rare metal magmas. Crystallization experiments were performed on two natural
15 granite compositions under H₂O-saturated conditions at 100, 200 and 300 MPa, from 540°C
16 to 700°C and between ~NNO+3 and NNO-1.4. Experimental charges were examined by SEM
17 and their products (glasses and crystals) analyzed for major elements by EMPA. Trace
18 element concentrations in selected glasses were also determined by LA ICP-MS. Despite
19 experimental durations commonly exceeding 1000 h and some up to 4000 h, kinetic problems
20 were encountered in particular in the 100 MPa charges whereas, at 200 and 300 MPa, results
21 consistent with previous melting experiments were obtained. Beauvoir melts crystallize
22 quartz, plagioclase, K-feldspar and mica as major phases. At NNO-1.4, mica is a biotite
23 whereas it is a Li-mica between ~NNO+3 and NNO-1. Apatite, Fe-Ti oxides, either hematite
24 or magnetite, topaz, amblygonite, cassiterite and columbite-tantalite appear as accessory
25 phases between ~NNO+3 and NNO+1. Experimental plagioclases are albitic (An < 4.5 mol%)
26 and more Ca and K-rich than natural albites in the granite whereas experimental K-feldspars
27 are more sodic (Ab < 45 mol%) than the natural crystals. The less evolved starting melt
28 crystallized Li phengites whereas the most evolved yielded Li-, F-rich micas near the
29 polyolithionite-zinnwaldite series, similar to natural micas in the granite. Equilibrium
30 crystallization increases A/CNK, F and P and concentrates Li, Be, B, Rb, Cs, W, U in the
31 melt. Nb and Ta are also enriched, their behavior being controlled by the solubility of
32 columbite-tantalite in the melt. Other elements are either unchanged (Mn, Zn, Ti) or depleted
33 (Sr, Pb) during magmatic fractionation. Sn is concentrated in Li-mica and hematite and it
34 behaves compatibly at high fO₂. Beauvoir melts crystallize at very low temperatures, below
35 670°C for the two compositions studied and solidus temperatures, determined from previous
36 melting experiments and confirmed by the new crystallization experiments, are near 550°C.
37 The experiments demonstrate that most of the mineralogical and geochemical characteristics

38 that make the Beauvoir granite distinctive result from magmatic rather than hydrothermal
39 post-magmatic processes. Albitic plagioclase, Li-mica, topaz and amblygonite are of
40 magmatic origin. Glass major element compositions suggest that the two granite samples
41 represent crystallized liquids. Trace element fractionations for most elements at Beauvoir can
42 be accounted for by magmatic crystallization-differentiation processes. Implications for the
43 mineralogy, fO_2 , volatile concentrations, crystallization and conditions of emplacement,
44 fractionation mechanisms and origin of the Beauvoir granite are discussed.

45

46 **Key words:** rare metal granite, crystallization experiments, mica compositions, magmatic
47 fractionation

48

49 INTRODUCTION

50 The Beauvoir granite (northwest French Massif Central, FMC) is a composite granitic
51 intrusion with mineralogical and geochemical characteristics emblematic of Variscan rare
52 metal granites (Linnen and Cuney, 2005; Breiter, 2012; Romer and Pichavant, 2021). Since
53 Aubert (1969), the Beauvoir granite has been extensively studied. Scientific results obtained
54 from the ~900 m long continuous core drilling Géologie Profonde de la France (GPF)
55 program have led to an integrated petrological, geochemical, geophysical, structural and
56 geochronological description of the Beauvoir granite and its country rocks (Cuney and
57 Autran, 1987). This has formed the basis for metallogenic models of mineralizations
58 associated with peraluminous granites (Cuney et al., 1992; Raimbault et al., 1995). Recently,
59 the Beauvoir granite has attracted a renewed interest, mainly centered on the magmatic-
60 hydrothermal evolution and on W deposition processes. Harlaux et al. (2018) have re-
61 investigated the primary fluid inclusions hosted in quartz and topaz from the granite and the
62 associated La Bosse W-bearing stockwork. Several episodes of wolframite generation and W
63 deposition have been recognized in the Echassières district around the Beauvoir granite
64 (Monnier et al., 2019; 2020).

65 Although the Beauvoir granite can be locally extensively transformed as a result of
66 greisenisation and kaolinization, results from the GPF program have emphasized the
67 magmatic evolution of the granite (Cuney and Autran, 1987). Raimbault et al. (1995) showed
68 that the Beauvoir granite was built as several independent magma batches. Its P- and Ta-rich
69 nature was recognized and also the geochemical similarity between the Beauvoir granite and
70 other enriched crustal magmas such as rare element pegmatites (Raimbault et al., 1995; Cerny
71 et al., 2005a). Subsequently, the Beauvoir granite has served as a reference for the

72 mineralogical and geochemical characteristics of other rare metal granites such as Montebras
73 (Marcoux et al., 2021), Argemela (Michaud and Pichavant, 2020) and Podlesi (Breiter et al.,
74 1997) in the Variscan belt and also Pitinga (Costi et al., 2009) and Yichun (Pollard, 2021)
75 outside the Variscan belt.

76 The Beauvoir granite is specific in many aspects, one of the most unusual being the
77 very low solidus temperatures, < 600°C for H₂O-saturated conditions between 100 and 300
78 MPa (Pichavant et al., 1987a), i.e. more than 100°C lower than the solidus of the haplogranite
79 Qz-Ab-Or system (Tuttle and Bowen, 1958). These low solidus temperatures have been an
80 important anchoring point in models proposed for the origin, magmatic evolution and
81 magmatic-hydrothermal transition of the Beauvoir granite (Aïssa et al., 1987; Cuney and
82 Autran, 1987; Cuney et al., 1992; Fouillac and Rossi, 1991; Raimbault et al., 1995). The
83 solidus temperature determinations were based on melting experiments and only a few
84 crystallization experiments were performed (Pichavant et al., 1987a). Since 1987, additional
85 experimental investigations have been carried out on the Beauvoir granite (Lahlafi, 1997;
86 Pichavant et al., 2016; unpublished) and new constraints have become available on the
87 magmatic evolution. The aim of this paper is to present a synthesis of experiments available
88 on the Beauvoir granite.

89 Pichavant et al. (1987a) pointed out the restricted number of experimental phase
90 equilibria available on natural rare metal granites in comparison with data on equivalent
91 synthetic compositions. Since that time and the works of Kovalenko (1979) on ongonites and
92 of Weidner and Martin (1987) on the Slip leucogranite from the St Austell pluton, several
93 experimental studies have come out on highly evolved granites. Solidus and liquidus
94 temperatures, crystallization sequences and melt fractionation trends have been determined on
95 one sample from the vertically zoned, F-bearing Xianghualing pluton in south China (Xiong
96 et al., 2002). Lukkari and Holtz (2007) worked out the phase relations and mutual stabilities
97 of F-bearing phases for the Kymi topaz granite in south Finland. Aksyuk et al. (2016)
98 presented solidus data for a biotite and a Li-F granite from the Voznesenka area, far-east
99 Russia. More recently, phase equilibria have been determined for the different facies of the
100 Qitianling pluton, south China (Huang et al., 2021). All these experimentally investigated
101 granites have in common to be F-bearing or F-rich and metaluminous to slightly
102 peraluminous and to present A-type affinities. In contrast, rare metal granites in the Variscan
103 belt are mostly strongly peraluminous and P-rich, besides being also high in F and Li
104 (Raimbault et al., 1995; Linnen and Cuney, 2005). This underlines the importance of phase
105 equilibrium data on the Beauvoir granite which, together with the Slip leucogranite (Weidner

106 and Martin, 1987), represents the only experimentally investigated example of peraluminous,
107 P-rich natural rare metal granite composition (PHP type of Linnen and Cuney, 2005). Below,
108 the experimental phase relations for two facies of the Beauvoir granite are presented, focusing
109 on liquidus and solidus temperatures, crystallization sequence, melt and mineral composition
110 trends and rare element enrichment mechanisms. The experimental results constrain the
111 conditions of crystallization of PHP rare metal magmas in the Variscan belt (Linnen and
112 Cuney, 2005; Roda Robles et al., 2018; Michaud and Pichavant, 2020) and provide insight on
113 the behavior of rare metals at the magmatic stage.

114

115 **THE BEAUVOIR GRANITE**

116 The Beauvoir granite is located within the Echassières granitic complex in the northwest part
117 of the FMC which belongs to the northern margin of Gondwana, the southern continent
118 involved in the Variscan collision (Faure et al., 2009). The Echassières granitic complex
119 outcrops in a tectonic window that exposes the Para-Autochthonous and Lower Gneiss Units,
120 structurally the deepest formations of the stack of metamorphic nappes in the FMC (Faure et
121 al., 2009). Granitic rocks from the Echassières complex, including the Colettes and the
122 Beauvoir granite, intrude biotite-muscovite micaschists from the Para-Autochthonous Unit
123 (Cuney and Autran, 1987; see map in Harlaux et al., 2017).

124 The main granite unit in the Echassières complex is the Colettes granite (surface
125 extension 40 km²), which is a biotite-muscovite granite. The Beauvoir granite forms a small
126 (surface extension 0.1 km²) satellite intrusion south of the Colettes. Field relations and
127 previous works demonstrate that the Beauvoir is intrusive into the Colettes granite. However,
128 radiometric ages for the two granites overlap, from 312 +/- 8 Ma (Colettes) to between 308
129 and 317 Ma (Beauvoir, see details in Cuney and Marignac, 2015 and Monnier et al., 2019).
130 Both granites crosscut the quartz-wolframite veins of the La Bosse stockwork hosted in
131 country rock micaschists south of the granitic complex. This early mineralizing episode (335
132 Ma, Harlaux et al., 2018; Monnier et al., 2019) has been interpreted to be genetically related
133 to the La Bosse granite whose existence is postulated at depth (Aubert, 1969; Cuney et al.,
134 1992; Harlaux et al., 2017).

135 The Beauvoir granite is mineralogically and geochemically highly specific. The GPF
136 drill hole intersected both the upper and lower contacts of the granite with metamorphic
137 country rocks at depths of 100 and 880 m respectively below the surface. Petrographic and
138 structural studies show that the granite was emplaced as an inclined sheet-like body in country
139 rock micaschists, remnants of which occur as roof pendants of several m in the upper part and

140 as sheets at the base of the granite. Analysis of the drill core revealed that the granite is
141 composite and consists of 3 main units (Cuney and Autran, 1987; Cuney et al., 1992),
142 including from top to bottom the B1 (from 100 to 450 m depth), B2 (from 450 to 750 m
143 depth) and B3 (from 750 to 880 m depth) units. Several porphyritic microgranites are
144 encountered near the bottom of the drill hole.

145 Mineralogically, the Beauvoir is an albite-lepidolite granite. There are minor but
146 systematic variations in textures and proportions of the main mineral phases (albite, lepidolite,
147 quartz, K-feldspar) between the different facies (Rossi et al., 1987; Cuney et al., 1992;
148 Harlaux et al., 2017). The B1 facies is hololeucocratic, rich in albite and lepidolite and
149 relatively poor in K-feldspar and it contains globular quartz with a snowball texture. The B2
150 facies, slightly darker than the B1, is enriched in K-feldspar and depleted in albite and
151 lepidolite. The B3 facies is characterized by pink perthitic K-feldspars which are more
152 abundant than micas. The granite hosts accessory phases such as topaz, apatite, amblygonite
153 and zircon and a rare element mineral assemblage including cassiterite, columbite-tantalite
154 (C-T), microlite and uraninite (Rossi et al., 1987; Cuney et al., 1992).

155 Geochemically, the Beauvoir granite is moderately silicic (68-72 wt% SiO₂), strongly
156 peraluminous (A/CNK = 1.25-1.45), enriched in Na (up to ~5 wt% Na₂O), P (from 1 to > 2
157 wt% P₂O₅) and F (from 1.5 to > 2.5 wt%) and depleted in Fe (< 0.5 wt% Fe₂O₃), Ca (< 1 wt%
158 CaO), Ti (< 100 ppm) and Mg (< 300 ppm, Raimbault et al., 1995). Li (from ~2000 to > 5000
159 ppm), Sn (up to ~1500 ppm), Nb (up to ~150 ppm), Ta (up to ~450 ppm) are strongly
160 enriched as well as W, Cs, Rb and Be whereas Ba, Pb, La, Zr, REE and Th are depleted
161 (Raimbault et al., 1995). Fe/Mn, Nb/Ta and Zr/Hf are particularly low whereas Fe/Mg is
162 particularly high. These enrichment and depletion patterns become progressively more
163 marked from base to top, i.e., from the B3, B2 to the B1 unit. However, some levels within
164 the B2 and B3 (the B' facies) preserve whole-rock compositions outside the main trends and
165 which have been interpreted as witnesses of a magma batch less evolved than the one at the
166 origin of the B1, B2 and B3 units (Raimbault et al., 1995).

167 The top part of the Beauvoir granite has been hydrothermally altered. Greisenisation
168 develops along fissures and fractures and the granite transforms into a quartz-muscovite
169 assemblage which can cover an important surface in outcrop (15 vol% in the kaolin open pit,
170 Cuney et al., 1992). Both the Beauvoir and the Colettes granites experienced extensive
171 kaolinization. However, hydrothermal alteration is generally inconspicuous in core samples
172 (Cuney et al., 1992) and of negligible importance from a geochemical point of view, except
173 for Sr and Be concentrated in late phosphate minerals (Raimbault et al., 1995; Charoy et al.,

174 2003). Yet, this does not necessarily ensure that granite whole-rocks are equivalent to magma
175 compositions since some elements (in particular F and Li) are enriched in micaschists around
176 the granite. These elements were probably extracted by late-magmatic fluids coming from the
177 Beauvoir magma (Cuney et al., 1992).

178

179 **EXPERIMENTAL METHODS**

180 **Starting materials**

181 Two Beauvoir granite samples were selected from the GPF drill core, one from the B1
182 (depth: 278.20-279.95 m) and the other from the B2 (depth: 606.05-606.95 m) unit. Both
183 samples were crushed and sieved to grain sizes of ~50 μm . The resulting rock powders were
184 analysed for major elements plus FeO, F, CO₂, total H₂O and Li at CRPG, Vandoeuvre les
185 Nancy (Pichavant et al., 1987a). Systematic differences consistent with those noted at the
186 scale of the entire GPF core appear between the two samples. The B1 is more sodic and
187 peraluminous, less potassic and so more evolved than the B2. Total Fe concentrations
188 decrease from the B2 to the B1 sample which is also the most P-, F- and Li-rich (Table 1).
189 Pichavant et al. (1987a) noted that the modal proportions of major phases in the two facies are
190 in the ranges expected; both quartz and K-feldspar increase and albite and lepidolite decrease
191 from the B1 to the B2 sample. No marked change in the major element composition of major
192 phases was found except for lepidolite, higher in Si and F and lower in Fe in the B1 than in
193 the B2 (Pichavant et al., 1987a).

194 The two rock powders were used directly in the melting experiments (Pichavant et al.,
195 1987a). For the crystallization experiments detailed in this study, two types of starting glasses
196 were prepared. First, hydrous glasses were synthesized by melting the two rock powders plus
197 5 wt% H₂O at 800°C, 100 MPa for 4 to 5 days. Second, dry glasses were prepared by melting
198 the rock powders at 1400°C, 0.1 MPa for ~3h. In both cases, two cycles of melting, with
199 grinding in between, were performed to improve the chemical homogeneity of the glasses
200 (Table 1). Both types of glasses, hydrous and dry, were crushed to grain sizes of ~50 μm and
201 used in the crystallization experiments.

202 Compositions of granite whole rocks and starting glasses are compared on Table 1.
203 The correspondence is generally good although some differences are apparent for elements
204 present in low concentrations (e.g., 0.22 vs. 0.40 wt% CaO in B1, Table 1). Minor variability
205 is also noted between glass batches (0.62 vs. 0.40 wt% FeO_t for B2, Table 1). These
206 discrepancies are analytical. They reflect the different methods and uncertainties associated

207 with the analysis of whole rocks and glasses (see below) and are not due to systematic
208 deviations between glass and whole rock compositions.

209

210 **Experiments**

211 Crystallization experiments on the Beauvoir granite were performed over a period of more
212 than 30 years and successively in two laboratories (Nancy and Orléans).

213

214 *Unbuffered (type 1) experiments*

215 These were carried out between 1986 and 1988 in Nancy in Tuttle-type cold-seal pressure
216 vessels working horizontally and pressurized with Ar. Temperature was measured by external
217 type K thermocouples and permanently recorded. Each thermocouple and bomb-furnace
218 assembly were calibrated under pressure using an internal dual thermocouple which served to
219 precisely locate the hot spot zone and to determine the corrections to be applied to the
220 external temperature readings. In these experiments, temperature is known to better than +/-
221 10°C. Pressure, measured with a Heise manometer, is known to better than +/- 2 MPa. Run
222 durations lasted from a minimum of 672 h (28 days) around the liquidus to a maximum of
223 2184 h (91 days) at subliquidus temperatures.

224 Experimental charges consisted of 95 % (~47 mg) of hydrous glass (containing
225 approximately 5% H₂O) plus 5 % doubly distilled demineralized water so that the total
226 amount of H₂O present was ~10 wt%. H₂O solubilities in F-, Li-, B- and P-rich silicic melts
227 such as the Macusani obsidians are ~8.5 wt% at 300 MPa (Holtz et al., 1993). Therefore,
228 Beauvoir experimental charges with 10 wt% H₂O in total are considered H₂O-saturated.
229 Charges were loaded in Au capsules (25 mm length, 2.5 mm internal diameter, thickness 0.2
230 mm) subsequently sealed by arc-welding. B1 and B2 capsules were placed side by side in the
231 same bomb to allow direct comparison of the results. A total of 24 charges were produced (12
232 for the B1, Table 2 and 12 for the B2, Table 3) of which 14 have been reported but only
233 partially described (Pichavant et al., 1987a). Although the fO_2 was not controlled in these
234 experiments, the redox conditions were highly oxidizing because of an Ar pressurizing
235 medium very poor in hydrogen. Thus, H₂ generated by H₂O dissociation inside experimental
236 capsules became lost to the pressurizing medium, hence lowering fH_2 and increasing fO_2 . In
237 practice, this highly oxidizing environment is marked by the common presence of thin red-
238 brick platelets of a hematite solid solution which is particularly abundant in B2 charges (see
239 below).

240

241 *Redox-buffered (type 2) experiments*

242 These experiments were performed to investigate the influence of fO_2 on mica crystallization
243 in evolved silicic melts (Lahlafi, 1997; Pichavant et al., 2016; Pichavant, unpublished). Most
244 were carried out between 1994 and 1996 in Orléans in horizontal externally heated rapid-
245 quench cold-seal hydrothermal pressure vessels (Pichavant 1987; Schmidt et al. 1995) and at
246 300 MPa. Total pressure was read with a Bourdon-tube manometer with an uncertainty
247 estimated to ± 20 bar. Temperatures were measured permanently with a type-K
248 thermocouple located in a well drilled at the bottom of the vessel. Temperature readings,
249 calibrated against measurements using an internal thermocouple working under pressure, are
250 accurate to better than $\pm 10^\circ\text{C}$. At the end of the experiments, the vessel was tilted vertically
251 and capsules slid from the hot spot to the cold bottom part of the vessel in a few seconds.
252 Different techniques were employed to control and vary the redox conditions (see Pichavant
253 et al., 2016 for details). The *oxidizing experiments* were all performed using Ar as
254 pressurizing gas yielding a moderately oxidizing fO_2 around $\text{NNO} + 1$ ($\Delta\text{NNO} = +1$, Lahlafi,
255 1997; Pichavant et al., 2016). They include one experiment on the B1 (Table 2) and six on the
256 B2 (four at 620°C and two at 700°C , both regrouped in Table 3) which have been previously
257 partially described (Lahlafi, 1997; Pichavant et al., 2016). They lasted from 1008 h (42 days)
258 to 4032 h (168 days). The *reducing experiments*, all on the B2 (three at 620°C regrouped in
259 Table 3, one at 630°C and one at 700°C) have been previously partially described (Lahlafi,
260 1997; Pichavant et al., 2016) except the one at 630°C . Most employed the double capsule
261 method of Eugster (1957). A Co-CoO buffer assemblage was used, yielding a ΔNNO of
262 around -1.4 (Table 3). The buffer was loaded into a 0.4 mm wall thickness external Au
263 capsule to minimize H_2 loss toward the Ar gas and experimental durations were limited to 360
264 h (15 days). The 630°C experiment was performed in 2017 in an internally-heated pressure
265 vessel (Scaillet et al., 1995) pressurized with an Ar- H_2 gas mixture to control the fO_2 (Table
266 3). Total pressure was permanently recorded and is known to better than ± 20 bar.
267 Temperature was continuously measured by three thermocouples with an uncertainty of \pm
268 5°C . The experiment lasted for 1123 h (48 days) under an experimental fO_2 of $\text{NNO}-1$
269 ($\Delta\text{NNO} = -1$) as determined from CoPd solid sensors (Michaud et al., 2021). At the end of the
270 experiment, the power to the furnace was switched off and temperature dropped by about
271 300°C during the first 4 mn. Last, one *two-stage reducing – oxidizing experiment* was
272 performed on the B2 (Pichavant et al., 2016).

273 All these experiments used Au capsules as containers (inner diameter: 2.5 mm, length:
274 20 mm, wall thickness: 0.2 mm). Charges consisted of about 50 mg of the dry glass plus about

275 5 mg of doubly distilled demineralized water so that the proportion of H₂O was kept at ~10
276 wt.% of the total charge which, from above, corresponds to H₂O-saturated conditions.

277 Following quench, pressure vessels were opened. Experimental capsules were
278 weighted to check for leaks, then opened and charges recovered to be analysed.

279

280 **Analytical methods**

281 Analysis of experimental charges focused on solid phases, glasses and crystals. No attempt
282 was made to recover and analyze the fluid phase. Initially, Pichavant et al. (1987a) used
283 optical microscopy and X-ray diffraction for charge characterization. Later, scanning electron
284 microscopy (SEM) became the main phase identification tool (Lahlafi, 1997; Pichavant et al.,
285 2016). In this study, experimental phase assemblages have been determined by SEM and
286 optical microscopy, and confirmed from electron microprobe data and mass balance
287 calculations. Several charges previously analysed by X-ray diffraction were examined by
288 SEM to check the consistency of phase identification between the two methods. Most type 2
289 experimental charges were examined with a JEOL 6100 scanning electron microscope
290 (BRGM, Orléans) under conditions summarized in Lahlafi (1997) and Pichavant et al. (2016).
291 More recently, a Zeiss Merlin Compact (ISTO, Orléans) instrument operating under either
292 secondary or back-scattered electron (BSE) mode and fitted with a Bruker energy dispersive
293 system (EDS) was used for the type 1 and some type 2 experimental charges.

294 Building on the rare analyses of feldspars from crystallization experiments in
295 Pichavant et al. (1987a), experimental charges were analysed by electron microprobe (EMPA)
296 and the major element compositions of experimental glasses and crystals systematically
297 determined. Most type 2 experimental charges were analysed using a CAMECA SX 50
298 instrument (BRGM-CNRS) working under 15 kV acceleration voltage, 12 nA sample current
299 and 10 and 5 s counting time on peak and background respectively (Lahlafi, 1997; Pichavant
300 et al., 2016). For the analysis of glasses, the beam was defocussed to 10 µm (vs. 1-2 µm for
301 crystals) and the raw Na₂O and K₂O concentrations were corrected using a Macusani glass
302 (Pichavant et al., 1987b) secondary standard following procedures detailed in Pichavant
303 (1987). More recently, type 1 and some type 2 experimental charges were analysed using a
304 CAMECA SX Five instrument (ISTO-BRGM) working under 15 kV acceleration voltage, 6
305 nA sample current and 10 and 5 s counting time on peak and background respectively. For the
306 analysis of glasses, the beam was defocused to 10 µm and focused to 1-2 µm for crystals.
307 Silicate analytical programs (Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K with P and F) were used. In
308 some sessions, Rb and Sn were added to the list of elements analysed but the Rb data were

309 discarded because of an interference with Si. Sn was also analysed separately under higher
310 sample current (20 nA) and counting times (60 and 30 s on peak and background). Attempts
311 were made to circumvent the problem of Na₂O and K₂O migration in glass. Time series data
312 established that alkali concentrations in experimental Beauvoir glasses strongly decrease with
313 counting time and, for a given counting time, with the sample current. “Zero time”
314 calculations (using number of counts calculated at $t = 0$ s rather than number of counts
315 measured after $t = 10$ s) were tested but this resulted in small insufficient corrections.
316 Combining results from time series measurements with analysis of secondary standards
317 (Pichavant, 1987), correction factors from 15 to 25% and 5 to 7.5% respectively for Na₂O
318 and K₂O were applied to the raw glass data. For glasses, relative analytical errors, determined
319 as standard deviations on multiple analyses on Macusani glass JV2, are: SiO₂ (0.5–1%), TiO₂
320 (80%), Al₂O₃ (1–1.5%), FeO (25%), MnO (100%), MgO (100%), CaO (15%), P₂O₅ (20%), F
321 (5–10%). For Na₂O and K₂O, the relative analytical errors are estimated to 5–10%.

322 Micas were too small to be analysed in all charges with the electron microprobe. This
323 difficulty was circumvented by using the EDS coupled with the Zeiss Merlin Compact SEM
324 to estimate mica compositions. The analyses were performed with an acquisition time of 30 s
325 and a sample current of 600 pA. The internal calibration of the instrument was used
326 unmodified and the data were processed with the ESPRIT2.3 software from Bruker. The SEM
327 data were compared with EMPA compositions of experimental micas analyzed in the same
328 charge. For most elements (Si, Al, Fe, Mg, Na, K, P), relative differences between SEM and
329 EMPA results were lower than the analytical errors of the electron microprobe. F was the only
330 element significantly underestimated by SEM (by an average factor of 2.3) and F
331 concentrations were corrected accordingly.

332 Proportions of glass and major mineral phases present in experimental charges were
333 determined by mass balance, knowing the composition of the bulk charge (Table 1) and of
334 each phase present from electron microprobe data. The calculations were performed on eight
335 major oxides (SiO₂, TiO₂, Al₂O₃, FeO, MgO, CaO, Na₂O, K₂O) simultaneously (i.e., using
336 input matrices with eight rows) and with a variable number of phases (depending on the
337 assemblage present in the charge) using linear least square procedures detailed in Albarède
338 (1995).

339 Experimental glasses and quartz crystals were analysed for trace elements by LA ICP-
340 MS. A RESOLUTION -SE 193 nm excimer laser coupled to an Agilent 8900 QQQ ICP-MS was
341 used (ISTO-IRAMAT). Ablation was performed at 10 Hz (glass) and 7 Hz (Qz) and a fluence
342 of 4 J.cm⁻². The spot diameter was set at 10 μm. SiO₂ from electron microprobe analysis was

343 used as an internal standard, NIST610 as an external standard (Pearce et al. 1997) and
344 Macusani glass JV2 (Pichavant et al. 1987b) as control. Reference glasses (NIST610 and
345 Macusani) were analyzed three times at the beginning and at the end of each run of ten
346 analyses. A total of 28 isotopes (Li7, Be9, B11, Na23, Al27, Si29, P31, K39, Ti49, Mn55,
347 Zn66, Ge72, Rb85, Sr88, Zr90, Nb93, Sn118, Cs133, Ba137, La139, Sm147, Eu153, Gd157,
348 Yb172, Ta181, W182, Pb208, U238) were measured. The raw data were processed off-line
349 using the Glitter software where time-resolved signals were examined one by one. Element
350 concentrations were calculated for the initial ~ 30 s of all ablation signals relative to the gas
351 blank (~ 20 s), with a linear interpolation across standard analyses.

352

353 **RESULTS**

354 **Experimental products and textures**

355 Phase assemblages and proportions determined by mass balance (major phases only) are
356 detailed in Table 2 and 3 for the B1 and B2 experiments respectively. Glass is the most
357 abundant phase in the charges, being readily identified by optical methods and SEM. It
358 frequently contains fluid inclusions of small to medium size (10-30 μm) homogeneously
359 distributed in the charge. All glass-containing charges appeared as solid consolidated beads
360 upon opening the capsule. In contrast, the B1N-16 charge (the only one without glass present,
361 Table 2) was recovered as a soft loose powder.

362 Phases crystallizing from Beauvoir granitic melts include quartz (Qz), an albitic
363 plagioclase (Pl) and K-feldspar (Af). SEM observations show that Qz forms euhedral, faceted,
364 mostly equant and sometimes globular crystals up to 20-30 μm large (Fig. 1a) which can
365 contain inclusions of micrometric micas. In charges close to the liquidus, Pl crystallizes as
366 euhedral elongated laths isolated in glass, typically 10-15 μm long but which can reach ~100
367 μm in some B1 experiments (Fig. 1b). They appear uniformly grey and so chemically
368 homogeneous in BSE mode (Fig. 1b). In contrast, in sub-liquidus charges, feldspars grow to
369 sizes generally larger than under near-liquidus conditions, up to ~100 μm . Small euhedral Qz
370 crystallize at their extremities (Fig. 1c). These large feldspars are chemically heterogeneous.
371 They are made up of white to dark grey domains in BSE mode corresponding to K- and Na-
372 rich zones (Fig. 1c). In some crystals, the transition from one zone to the other is progressive
373 and geometrically complex (Fig. 1c). In others, the two zones are well separated, both
374 geometrically and compositionally, and two chemically homogeneous feldspar phases, Pl and
375 Af can be individualized (Fig. 1d). Af most commonly appears as small (maximum 15 μm)

376 euhedral homogeneous K-rich phases coexisting with Pl in composite crystals (Fig. 1d) and
377 was unfrequently found as isolated crystals isolated in glass.

378 Beauvoir melts have been shown previously to crystallize a mica phase (Pichavant et
379 al., 1987a; Lahlafi, 1997; Pichavant et al., 2016). Under the most reducing conditions
380 corresponding to the CCO buffer (NNO-1.4), biotite (Bt) is the liquidus phase at 300 MPa in
381 B2 (experiments B2O-38/40, Table 3; Lahlafi, 1997; Pichavant et al., 2016). This Bt is a
382 moderately F-rich siderophyllite (Pichavant et al., 2016). It amounts to several wt.% of the
383 charge and reaches a maximum size of 30–40 μm (see Fig. 1b in Pichavant et al., 2016). For
384 strongly oxidizing (NNO+3) to moderately reducing conditions (NNO-1) in B2 (Table 3) as
385 well as between NNO+3 and NNO+1 in B1 (Table 2), Beauvoir melts crystallize a Li-, F-
386 bearing mica. This mica (designated below as Li-mica, Li-Mi) has compositions ranging from
387 relatively Li-, F-poor and mainly dioctahedral to Li-, F-rich and markedly trioctahedral. SEM
388 observations show that Li-Mi appears mostly as platelets, either elongated as in B2 (Fig. 1e)
389 or more rectangular and equant as in B1 charges, sometimes forming aggregates and
390 inclusions in quartz and feldspars. Li-Mi is generally very flat and thin, typically 10-15 μm
391 long and 2-3 μm large in B2 (Fig. 1e) and \sim 2 μm long and 1-2 μm large in B1 charges (Fig.
392 1f; g). Exceptionally, Li-Mi can be up to several tens μm long in some B2 charges
393 (experiments B2O-2/5, Table 3, see Fig. 1a in Pichavant et al., 2016).

394 Several accessory phases were encountered although the accessory assemblage was
395 not systematically determined in all charges. The most systematic include apatite (Ap) in both
396 B1 and B2 charges and Fe-Ti oxides in B2 charges only, either hematite (Hm) or magnetite
397 (Mt, Table 2; 3). Ap forms small subhedral crystals up to 5-10 μm (Fig. 1e), sometimes but
398 rarely larger and often grouped in aggregates. In the type 1 experiments which are the most
399 oxidized (NNO+3), Hm forms thin red-brown losange-shaped platelets up to several tens μm
400 long (Fig. 1h), readily detected by optical microscopic examination and locally abundant in
401 B2 charges (B2N-19, B2N-20, Table 3). In the less oxidized type 2 experiments (NNO+1) on
402 B2, a small (a few μm) sub-hexagonal oxide phase identified as magnetite (Lahlafi, 1997) is
403 present. No Fe-Ti oxide crystallized in B2 experiments performed at NNO-1 and NNO-1.4,
404 (Table 3). Topaz was extensively looked for but only found in one charge, B1N-21, where it
405 appears as small ($< 5 \mu\text{m}$) homogeneously distributed euhedral prisms slightly brighter than
406 quartz and feldspar in BSE mode (Fig. 1f, g). This B1N-21 charge contains an Al-, F-rich
407 phosphate phase of the amblygonite-montebbrasite series present together with apatite. It forms
408 relatively abundant, subhedral 10 μm large crystals darker than the other phases in BSE mode
409 (Fig. 1g). Cassiterite and C-T are conspicuous in B1 as well as in B2 charges.

410

411 **Experimental phase relations**

412 Results of the crystallization experiments are reported in Table 2 and 3 for the B1 and B2
413 starting compositions respectively. Below, we focus on results in the strongly oxidizing to
414 moderately reducing fO_2 range (from NNO+3 to NNO-1, Table 2; 3), i.e., the most reducing
415 (NNO-1.4) B2O-38/40 and B2O-41 charges are excluded. Therefore, only conditions of
416 crystallization of Qz, Af, Pl and Li-Mi are discussed and those for Bt are not considered
417 further.

418 Some inconsistencies were found between experiments in the sense that the same P-T
419 conditions did not always yield the same phase assemblages. For example, at fixed T and P
420 (620°C and 300 MPa) for the B1 composition, Li-Mi is present in B1N-22 (type 1
421 experiment) whereas it is absent in B1O-1 (type 2 experiment) which also lacks Af and Qz
422 (Table 2). For the B2 composition, both Pl and trace Af crystallized in B2O-2/5 (type 2
423 experiment) at 620°C and 300 MPa whereas these two phases are absent in B2N-25 (type 1
424 experiment, Table 3) performed under the same P-T conditions. These inconsistencies are
425 attributed to several experimental factors such as use of different equipment and contrasted
426 experimental durations. By opposite, phase identification issues are not considered
427 responsible for these inconsistencies. Phase assemblages are identical, whether determined
428 from SEM (this study) or X-ray diffraction (Pichavant et al., 1987a). Starting compositions
429 were successfully mass balanced from linear combinations of phases present in the charges
430 (low Σr^2 in Table 2 and 3), implying that our phase identification methods are reliable. In only
431 one case (B1N-1), 2 wt% Qz (undetected by X-ray and SEM) had to be added to the phase
432 assemblage for the charge to be successfully mass balanced. In the same way, the presence of
433 fluid inclusions in most charges confirms that H_2O -saturated conditions (i.e., $a_{H_2O} = 1$) were
434 effectively imposed. Thus, differences in phase assemblages for fixed composition and P, T
435 conditions cannot be due to a_{H_2O} being variable between charges although, in this study, melt
436 H_2O concentrations have not been determined. Beauvoir melt inclusions contain between 3.6
437 and 6.2 wt% H_2O (Kovalenko et al., 1998).

438 These inconsistencies have been considered in the construction of the P-T diagrams
439 (Fig. 2; 3). For the B1, saturation curves at 300 MPa exclude two experimental charges, B1O-
440 1 and B1N-15 (Table 2). SEM observations and mass balance calculations show that these
441 charges are much less crystallized than the others at the same pressure (B1N-21 and B1N-22,
442 Table 2). Moreover, Qz, Af and Li-Mi are absent in B1O-1 and B1N-15 whereas all are
443 present in B1N-22 (Table 2). These inconsistencies are attributed to differences in

444 experimental durations, much shorter for B1O-1 (1008 h) and B1N-15 (1104 h) than for B1N-
445 21 and B1N-22 (2184 h, Table 2). For the B2, the Pl and Af saturation curves pass at
446 temperatures slightly lower (-10 and -20°C respectively) than the 300 MPa B2O-2/5 type 2
447 charges which contain Pl and trace Af. In other words, results from the B2N-25 and B2N-24
448 type 1 charges (Table 3) were preferred.

449 For the B1 composition, the 100 MPa solidus is bracketed by the 540°C glass absent
450 (B1N-16) and the 560°C trace glass present (B1N-11) charges. The 300 MPa solidus is half-
451 bracketed by the 580°C glass bearing (B1N-21) charge (Table 2). Although solidus
452 temperatures determined from melting experiments are preferred because of the possibility of
453 supercooling problems in crystallization experiments (Pichavant et al., 1987a and see below),
454 the latter are only ~20°C lower than the former at 100 MPa (Fig. 2). At this pressure, Qz, Pl,
455 Af and Li-Mi all saturate from 590°C to > 600°C. Remarkably, temperatures of Qz, Pl and Af
456 saturation are higher at 300 MPa (from 625°C to > 640°C) than at 100 MPa (Fig. 2). This is
457 symptomatic of kinetic difficulties (melt supercooling) at 100 MPa since, under water-
458 saturated conditions, anhydrous phases such as Qz and feldspars must saturate at
459 progressively lower temperatures upon increasing pressure (Tuttle and Bowen, 1958; Luth,
460 1976). Consequently, the Qz, Pl and Af saturation curves are drawn separated in two parts,
461 one appropriate for 100 MPa and the other for 300 MPa. By opposite, kinetic problems are
462 negligible for Li-Mi (see below) and Li-Mi saturation is represented by a single continuous
463 curve with a positive slope between 100 and 300 MPa (Fig. 2). Qz, Pl and Li-Mi appear all
464 together on the B1 liquidus (at 600°C, 100 MPa and at 640°C, 300 MPa) and so their relative
465 crystallization order cannot be specified. They are followed by Af which is the fourth and last
466 major phase in the sequence.

467 For the B2 composition, the solidus is only approached (and half-bracketed) at 100
468 MPa by the 560°C (B2N-12) charge which contains a trace of glass and it is < 580°C at 300
469 MPa (Fig. 3). The 100 MPa solidus temperature is lower by 20-30°C than the one determined
470 from the melting experiments (Pichavant et al., 1987a). Qz and Pl crystallize at progressively
471 lower temperatures at 300 (615-625°C) than at 200 MPa (> 640°C), as expected under water-
472 saturated conditions (Tuttle and Bowen, 1958; Luth, 1976). In comparison, these two phases
473 saturate below 640°C at 100 MPa which indicates supercooling problems at this pressure.
474 Therefore, the Qz and Pl saturation curves are drawn in two parts, one appropriate for 100
475 MPa and the other for 200 to 300 MPa (Fig. 3). As supercooling problems are less important
476 to negligible for Af and Li-Mi (see below), saturation conditions are represented by single
477 continuous curves between 100 and 300 MPa. One notes the contrast in slope between the Li-

478 Mi (positive) and the Qz, Pl and Af saturation curves (all negative, Fig. 3). Li-mica is the
479 liquidus phase at 300 MPa being joined by Qz and Pl on the liquidus at 200 MPa. At 100
480 MPa, the four major phases crystallize all together on the liquidus (B2N-20 charge, Table 3).
481 The 300 MPa experiments imply that Qz crystallizes before Pl and the 300 and 200 MPa data
482 that Af is the fourth and last major phase in the sequence (Fig. 3).

483

484 **Major element compositions**

485 *Glasses*

486 Major element compositions of experimental glasses corrected for alkali migration are
487 reported in Table 4. Except the few data from Lahlafi (1997) which are normalized to 100
488 wt%, other glass analyses are shown unnormalized. Average standard deviations for SiO₂ and
489 Al₂O₃ are no higher than analytical errors, thus indicating that the experimental glasses are
490 homogeneous. We note that the MnO, MgO, CaO and SnO₂ concentrations are in the same
491 range as standard deviations and in fact often lower than detection limits. They are reported
492 only for indicative purposes in Table 4. Experimental Beauvoir glasses are extremely felsic,
493 being composed of essentially four oxides (Si, Al, Na, K) plus P₂O₅, F and H₂O. This is
494 reflected in the calculated normative Qz, Ab and Or concentrations which sum to more than
495 90-95%. Normative An contents are between 0.2 and 0.7% for the B1 and between 0.2 and
496 1.4% for the B2 glasses. Normative Corundum (from 4.9 to 7% for the B1 and from 2.5 to
497 5.5% for the B2) and A/CNK (1.37-1.60 for the B1 and 1.20-1.44 for the B2) demonstrate
498 that all glasses are strongly peraluminous. Near-liquidus experimental glasses (e.g., B1N-15
499 for B1 and B2N-26 for B2, Table 4) are not markedly more peraluminous than their
500 corresponding starting glasses (Table 1), suggesting that the influence of alkali migration on
501 glass analyses has been satisfactorily counteracted.

502 Both B1 and B2 glasses display the same compositional trends with progressive
503 crystallization and decreasing melt fraction (*f*, Fig. 4). Normative Corundum and A/CNK both
504 increase and residual melts become progressively more peraluminous. P₂O₅ and F
505 concentrations (n.b., not normalized to 100% anhydrous) also increase, indicating their
506 progressive accumulation in residual melts. The most evolved B1 melts reach ~1.70 wt%
507 P₂O₅ and ~3.3 wt% F (for ~1.50 and ~2.3 wt% in starting materials) and the most evolved B2
508 melts ~1.30 wt% P₂O₅ and ~2.6 wt% F (for ~1 and ~1.7 wt% in starting materials). By
509 opposite, normative Anorthite are much lower in experimental glasses than in starting
510 materials but the trends (Fig. 4d) are blurred by the large uncertainty on CaO in glasses by
511 electron microprobe because of the low concentrations involved.

512 The major element compositions of starting materials and experimental glasses are
513 plotted in the Qz-Ab-Or diagram (Fig. 5). Globally, the B1 are more Ab-rich and Qz-poor
514 than the B2 compositions. The former plot left to the H₂O-saturated haplogranite minimum
515 liquidus temperature trend (Tuttle and Bowen, 1958) whereas the latter plot mostly on the 300
516 MPa minimum point (Fig. 5). Both glass groups have Qz higher than H₂O-saturated minimum
517 liquidus temperature points at 100 MPa with added F (Manning, 1981). Compositions of
518 several glasses (see captions to Fig. 5) and starting materials overlap as expected for charges
519 above or near the liquidus and in fact demonstrates the internal consistency of experiments
520 and analyses. Glasses from sub-liquidus charges do not systematically vary with the phase
521 assemblage, being tightly grouped and little dispersed around their respective starting
522 compositions. The 300 MPa B1 glasses (B1N-23, -22, -21) show slight variations in Ab/Or at
523 constant Qz contents whereas the 100 MPa glasses (B1N-1, B1N-17) evolve toward distinctly
524 higher Qz contents (Fig. 5). In the same way, the sub-liquidus B2 glass compositions have
525 constant Qz, plotting close to the starting composition (Fig. 5). Two glasses have very high
526 Qz contents in comparison with the other B2 compositions, B2N-3 (which come from a 100
527 MPa experiment) and B2O-38/40 (analysis from Lahlafi, 1997). The B2N-18 glass has Qz
528 and Or respectively lower and higher than the other B2 compositions.

529

530 *Quartz*

531 Representative major element compositions of experimental quartz are reported in Table 5.
532 The complete database can be found in the Supplementary Material (Table S1). In most
533 charges, quartz is very close to pure SiO₂. The only element systematically detected with the
534 electron microprobe is Al with Al₂O₃ concentrations between 0.07 to 0.64 wt%, on average
535 0.23 wt%. Phases with concentrations much richer in Al₂O₃ (2.83-3.01 wt%, Table 5) were
536 found in one low temperature 100 MPa B1 charge (B1N-17, Table 2). SEM examinations
537 show that these Al-rich phases are morphologically and texturally identical to Qz. They
538 coexist with Qz having “normal” Al₂O₃ concentrations in the same charge. The high Al₂O₃
539 analyses do not show elevated alkali contents but are systematically associated with totals in
540 the 97.5-98 wt% range, lower than for the other Qz (Table 5). Since Li is present at elevated
541 concentrations in the B1 starting material (1.28 wt% Li₂O, Table 1), Li incorporation in the
542 quartz lattice according to the coupled Si = Al + Li substitution (Breiter et al., 2014) might
543 account for the analytical results. These peculiar compositions would thus correspond to
544 virgilite solid solutions between quartz and spodumene of stoichiometry Li_xAl_xSi_{3-x}O₆
545 (French et al., 1978; Sirbescu et al., 2017), with x ~0.10 from our data (Table 5; S1).

546

547 *Feldspars*

548 Representative compositions of experimental feldspars are reported in Table 6 and the entire
549 database is available in the Supplementary Material (Table S2). Feldspar compositions were
550 determined in all feldspar-bearing charges including in some previously analysed by
551 Pichavant et al. (1987a). Despite their small size (5 μm) in some charges, overall feldspars
552 were correctly analysed and satisfactory structural formulae ($\text{Si}+\text{Al}+\text{P} = 4.01$; $\text{K}+\text{Na}+\text{Ca} =$
553 0.97 apfu on average) were obtained. Three groups of compositions are distinguished
554 corresponding to Pl, Af and heterogeneous feldspars (Fig. 1c). All three groups contain
555 significant P_2O_5 concentrations, higher in plagioclase (from ~ 0.6 to ~ 1.4 wt% P_2O_5) than in
556 K-feldspar (from ~ 0.4 to ~ 1 wt% P_2O_5 , Table 6). The incorporation of phosphorus in
557 experimental feldspars from this study is consistent with the peraluminous composition and
558 elevated P_2O_5 contents of Beauvoir melts (Simpson, 1977; London, 1992; London et al.,
559 1999).

560 The three groups of compositions are plotted on the An vs. Or diagram of Fig. 6. Use
561 of a binary instead of the more familiar ternary An-Or-Ab representation is warranted since
562 feldspars in this study have very low An ($< 5\%$, Table 6; S2). A minor dispersion is apparent
563 and feldspars in the same charge can have An, Or or Ab contents that differ by 1-2 % from
564 each other. Pl are albitic with on average $\sim 10\%$ Or and An from ~ 0.5 to $\sim 4.5\%$ (Fig. 6). There
565 are slight tendencies for An to increase with temperature and Or to decrease with pressure. Pl
566 compositions from B1 and B2 charges overlap but the least An-rich ($< 1.5\%$ An) mostly
567 correspond to B1 and the most An-rich (An $> 2.5\%$) to B2 charges. For comparison, natural
568 albite from starting rocks ($\text{Or}_{0.8}\text{Ab}_{98.8}\text{An}_{0.4}$, Pichavant et al., 1987a) has An at the low end of
569 the experimental range and Or much lower than Pl from this study (Fig. 6). Af compositions
570 are very close to binary, most being grouped between Or_{67} and Or_{73} with An contents ~ 0
571 although some compositions extend to lower Or and An > 0 . No difference appears between
572 B1 and B2 charges (Fig. 6) but Af from 100 MPa charges have lower Or than those from 300
573 MPa charges. Natural K-feldspar from starting rocks ($\text{Or}_{89}\text{Ab}_{11}$, Pichavant et al., 1987a) plots
574 outside the experimental range, being much more potassic and less sodic than Af from this
575 study. Analyses of heterogeneous feldspars yield a group of compositions intermediate
576 between Pl and Af.

577 Equilibrium between feldspar pairs coexisting in the same charge was tested by using
578 the SOLV CALC software of Wen and Nekvasil (1994) with the feldspar solution model of
579 Nekvasil and Burnham (1987). 14 pairs from 3 B1 and 3 B2 charges were tested. 10 returned

580 equilibration temperatures within 2-32°C of experimental values as illustrated (Fig. 6). 4 pairs
581 were out-of-equilibrium because of Pl compositions slightly outside the solvus. Thus, a
582 majority of feldspar pairs in our experimental charges represent compositions equilibrated on
583 the ternary solvus. The intermediate compositions of heterogeneous feldspars are interpreted
584 as out-of-equilibrium compositions from inside the solvus rather than as mixtures of
585 compositions at equilibrium on the solvus (Fig. 1c; 6).

586

587 *Li-micas*

588 Li-mica compositions were determined by EMPA in B2 charges only. In B1 charges where
589 micas are very small, compositions were estimated by SEM. Li-Mi in charge B2N-25 were
590 analysed both by EMPA and SEM to calibrate the SEM analyses. Li₂O contents in Li-Mi
591 were estimated from F concentrations, using the equation [di 1] of Tischendorf et al. (1997).
592 Results were checked against mineral separate data on Beauvoir micas (Monier et al., 1987).
593 Equation [di 1] performs very well (relative differences between measured and expected
594 concentrations < 5%) for very Li-rich micas (from 5 to > 7 wt% Li₂O). For less Li-rich micas,
595 [di 1] underestimates Li₂O concentrations by 15-30%. Representative EMPA and SEM
596 compositions of experimental Li-Mi are reported in Table 7 and the entire database can be
597 found in Table S3. The data demonstrate a marked contrast in Li-Mi compositions between
598 the B1 and B2 charges. In B2, Li-Mi are grouped around the Li phengite end-member (n.b.,
599 the nomenclature of Tischendorf et al., 1997 is used for consistency with the representation of
600 mica compositions) whereas, in B1, they plot around the cryophyllite end-member and so
601 possess a marked trioctahedral character (Fig. 7a). Combining EMPA and SEM data (Table
602 S3), Li-Mi in B2 charges have 43.6-48.3 wt% SiO₂, 29.6-34.9 wt% Al₂O₃, 1.62-7.43 wt%
603 FeO_t, 0.07-0.60 wt% MgO, 0-0.44 wt% MnO, 0.69-2.23 wt% Na₂O and 8.9-10.2 wt% K₂O. F
604 ranges from 1.37 to 4.91 wt% and Li₂O from 0.60 to 3.25 wt%. SnO₂ concentrations exceed
605 the detection limit (~400 ppm Sn under our electron microprobe operating conditions),
606 reaching values up to ~0.3 wt% SnO₂ in the more oxidized type 1 charges. For example, in
607 charge B2N-25, SnO₂ concentrations range from 0.14 to 0.32 wt% (Table 7; S3) and nine
608 separate analyses (not detailed in Table S3, detection limit ~200 ppm Sn) yielded an average
609 concentration of 0.27 +/- 0.05 wt% SnO₂. Compositions of B2 Li-Mi do not systematically
610 change with experimental P-T conditions. Yet, FeO_t broadly decreases with decreasing melt
611 fraction and we note that Li-Mi in the reduced type 2 B2O-50 charge have lower FeO_t (2.28-
612 2.55 wt%) than in oxidized near-liquidus charges (4.77-5.68 wt% FeO_t in B2N-25) and Sn
613 concentrations below detection (Table 7; S3).

614 In the three B1 charges analyzed, Li-Mi have 50.9-56.9 wt% SiO₂, 18.2-22.0 wt%
615 Al₂O₃, 1.24-5.40 wt% FeO_t, 0.09-2.31 wt% MgO, 1.07-3.00 wt% Na₂O and 5.29-7.42 wt%
616 K₂O. F ranges from 7.13 to 8.79 wt% (applying the constant x 2.3 correction factor
617 determined on B2N-25) and Li₂O from 5.32 to 7.02 wt% with [di 1] (Table 7; S3). SiO₂,
618 Al₂O₃, K₂O, Na₂O, F and Li₂O do not significantly vary between the three charges. However,
619 MgO progressively decreases from B1N-23, -22 to -21 (i.e., with decreasing *f*) and Li-Mi in
620 B1N-21 has FeO_t (1.24-2.70 wt%) lower than in B1N-22 and B1N-23 (3.41-5.40 wt%, Table
621 7; S3) so that the former and the two latter charges plot in separate fields (Fig. 7a). Apart from
622 the first-order differences in SiO₂, Al₂O₃, F and Li₂O between B1 and B2 micas, MgO and
623 Na₂O are enriched and K₂O depleted in B1 compared to B2 whereas FeO_t are similar in the
624 two mica groups (Table 7).

625

626 *Accessory phases*

627 Representative major element compositions of accessory phases are given in Table 8.
628 Although few analyses are available and their quality is relatively poor because of the small
629 size of the crystals, topaz, amblygonite and apatite are all very F-rich. Atomic F/(F+OH) in
630 topaz ranges from 0.89 to 0.99 and in amblygonite from 0.94 to 0.95. Apatite are fluor-
631 apatites with atomic F concentrations commonly exceeding 1 atom per formula unit (Table 8).
632 F concentrations in topaz and glass from charge B1N-21 (Table 4; 8) are consistent with the
633 proposed distribution of F between topaz and silicic melts (Soufi, 2021). In this same charge,
634 the distribution of F between amblygonite and melt broadly follows the experimental
635 calibration of London et al. (2001).

636 Experimental hematites are very Fe-rich and Ti-poor. They have a minimum of 95%
637 Hm (Fe₂O₃, Table 8), consistent with the high *f*O₂ of type 1 experiments (Table 3). Hematites
638 contain very little MnO and MgO but significant Al₂O₃ (0.79-2.71 wt%) and SnO₂ (1.39-1.57
639 wt%). Separate analyses (not detailed in Table 8) yielded an average of 1.81 (+/- 0.22) wt%
640 SnO₂ for Hm in charge B2N-25. Such high Sn concentrations in Hm are consistent with those
641 detected in coexisting Li-Mi (see above).

642

643 **Trace element compositions**

644 Trace elements were analyzed in several highly oxidizing (~NNO+3) B2 charges and in Qz
645 crystals from one charge and concentrations of selected elements are reported in Table 9. The
646 charges analyzed are crystallized except one which is essentially crystal-free. Yet,
647 examination of time-resolved ablation spectra allowed glass signals to be unambiguously

648 identified in all charges. Multiple analyses (from 2 to 6) are available on each glass. Chemical
649 dispersions (defined as standard deviation/mean of analyses) are mostly < 10% (Table 9),
650 globally demonstrating homogeneous trace element distributions in the experimental glasses
651 analyzed. In some charges and for some elements, chemical dispersions are higher (> 20%).
652 This is the case in particular for Ba and La which have low concentrations in the analyzed
653 glasses (~1 ppm for Ba and ~0.1 ppm for La, Table 9), not much above instrumental
654 uncertainties. For Zr and Sn which have glass concentrations respectively of 3-4 ppm and
655 between ~15 and ~35 ppm (Table 9), the relatively high dispersions (20-30%) in some
656 charges could reflect glass contamination by zircon and Sn-bearing phases (cassiterite, micas
657 and hematite). Dispersions > 10% for Ta in two charges might also be due to contamination
658 with C-T and it is possible that hematite contamination explains some Ti dispersions > 10%.
659 Almost all Sm, Eu, Gd and Yb concentrations were below detection and these elements are
660 not reported in Table 9.

661 The evolution of trace element concentrations with progressive crystallization is
662 illustrated on Fig. 9. Experimental melt (i.e., glass) concentrations (C) referenced to those in
663 the initial melt (C_0 , represented by the essentially crystal-free charge B2N-26) are plotted as a
664 function of the melt fraction. For most elements, concentrations increase with decreasing f . Li,
665 Be, B, Cs, U are the most strongly enriched, C/C_0 reaching averages of ~1.4 ($f \sim 75\%$) and
666 ~1.7 ($f \sim 50\%$, Fig. 9). Below 25% crystallization, Li, Be, B, Cs, U concentrations follow
667 Rayleigh fractionation for a bulk mineral/melt partition coefficient (D) of 0.01 (Fig. 9),
668 indicating a strongly incompatible behavior. Upon further lowering f , rates of enrichment
669 decrease and concentrations plot on a Rayleigh fractionation trend for a higher D (0.5),
670 suggesting that these elements become less incompatible as f drops. Rb and W are also
671 concentrated in the melt although less markedly ($C/C_0 \sim 1.2$ for $f \sim 50\%$), most probably
672 because both are more compatible in Li-Mi than Li, B, Cs according to the muscovite/melt
673 partition coefficients summarized by Michaud and Pichavant (2020). We conclude that the
674 behavior of Li, Be, B, Rb, Cs, W, U reflects their globally incompatible character during
675 crystallization of Beauvoir melts. The D values, their variability from one element to the other
676 and their evolution with varying f control the observed levels of enrichment.

677 Nb and Ta are also concentrated in the melt during crystallization. C/C_0 are quite high,
678 ~1.3 for Nb and ~1.5 for Ta for $f \sim 50\%$ (Fig. 9), indicating preferential concentration of Ta
679 rather than Nb. However, at the difference of the group of elements above, Nb and Ta are
680 essential structural components of C-T [(Fe,Mn)(Nb,Ta)₂O₆], an accessory phase present in
681 experimental charges. Therefore, the measured glass Nb and Ta concentrations should

682 represent C-T solubilities provided that chemical equilibrium is established between melt and
683 accessory phases in the experiments. C-T solubilities in silicic melts are complex functions of
684 temperature and compositions of the melt and of the C-T phase (Linnen and Keppler, 1997;
685 Linnen, 1998; Van Lichtervelde et al., 2018). At a temperature of 600°C appropriate for
686 Beauvoir melts, the calculated Nb concentration of a 500 ppm Mn melt at equilibrium with
687 MnNb_2O_6 is 52 ppm (Linnen and Keppler, 1997). At the same temperature, the empirical
688 saturation model of Michaud and Pichavant (2020) yields 55 ppm Nb and 43 ppm Ta for a
689 500 ppm Mn and 2000 ppm Fe melt saturated with C-T with at. $\text{Mn}/(\text{Mn}+\text{Fe}) = \text{Ta}/(\text{Ta}+\text{Nb}) =$
690 0.5. For Beauvoir temperature and melt compositions, Linnen (1998) estimated a
691 concentration of 50 ppm Ta at saturation with MnTa_2O_6 , increasing to 1200-3000 ppm Ta and
692 40-100 ppm Nb at saturation with MnNb_2O_6 if the influence of Li and F melt concentrations
693 on C-T solubility is accounted for. Although the calculations above do not all agree with
694 concentrations measured in experimental Beauvoir glasses (38-49 ppm Nb, 20-34 ppm Ta,
695 Table 9), most, in particular for Nb, are in the same range. Considering the uncertainties in the
696 calculations and simplifying assumptions involved, this is taken as permissive evidence that
697 our Nb and Ta glass concentrations represent C-T solubilities. The observed Nb and Ta
698 enrichments with decreasing f (Fig. 9) would thus reflect a C-T solubility increase resulting
699 from progressive melt fractionation during crystallization (higher F, P, Li, Fig. 4; 9).

700 Sn is an essential structural component in cassiterite (SnO_2), another accessory phase
701 present in the charges and, so, Sn melt concentrations should reflect cassiterite solubilities
702 (Linnen et al., 1995; 1996; Bhalla et al., 2005). Extrapolating to lower temperatures the data
703 of Bhalla et al. (2005) at NNO+2.3 (vs. \sim NNO+3 in our experiments), the solubility of
704 cassiterite at 600°C is 120 ppm Sn, i.e. \sim 4 times higher than the measured Sn glass
705 concentrations (between \sim 15 and \sim 35 ppm, Table 9). We also note the marked Sn depletions
706 in B2N-25 and B2N-18 glasses and the global absence of Sn enrichment in the melt (Fig. 9).
707 Sn alloying with the Au capsule (Linnen et al., 1995; Michaud et al., 2021) is probably
708 responsible for Sn glass concentrations at cassiterite saturation being lower than calculated
709 solubilities. Additionally, the high Li-mica and hematite Sn concentrations indicate that Sn is
710 compatible at \sim NNO+3 (see also below).

711 Several elements have concentrations that do not increase in the melt when decreasing
712 f . This is the case for Mn which, following initial enrichment, returns to melt concentrations
713 not much different from the initial C_0 for $f \sim 50\%$ (Fig. 9). Such a behavior could reflect
714 increased Mn sequestration by Li-Mi at low f (see partition coefficient in Michaud and
715 Pichavant, 2020) although Mn melt concentrations also depend on accessory phase

716 solubilities, as it is an essential structural component in CT. Zn is neither significantly
717 enriched nor depleted and Ti shows a similar although more irregular behavior (Table 9). The
718 only element regularly depleted in the melt upon decreasing f is Sr. C/C_0 is 0.37 ($f \sim 75\%$) and
719 0.27 ($f \sim 50\%$) and the Sr data can be approximately fitted by Rayleigh fractionation for $D = 3$
720 (Fig. 9) thus indicating a marked compatible character during crystallization of Beauvoir
721 melts. In comparison, Pb is more irregular than Sr although a depletion trend appears if the
722 anomalous concentration in charge B2N-18 (Table 9) is discarded. The Ba and La systematics
723 are unclear because of their very low melt concentrations, as discussed above.

724 Experimental Qz crystals were analyzed only in B2N-25 because of size limitations
725 (Fig. 1a). Five analytical spots were performed. Only one yielded a stable signal during
726 almost the entire ablation time interval before finally the underlying glass was hit. This Qz
727 has 81.34 ppm Li, 367.92 ppm Al and 152.94 ppm P (not corrected for possible interferences,
728 Audétat et al., 2014) and it contains Ti, Zn, Sn and Cs concentrations in the few ppm range
729 (Table 9).

730

731 **Mineral/melt partitioning**

732 The distribution of F between mica and melt is strongly concentration dependent (Fig. 10).
733 Below ~ 3 wt% F in melt, an approximately linear dependence is observed between F in mica
734 vs. F in melt with a slope of ~ 1.4 (30 data points). F is therefore preferentially partitioned in
735 Li-Mi than in silicic melt, consistent with previous results ($K_d \text{ Mu/L} = 1.8$, Icenhower and
736 London, 1995; 0.94-1.56, Pichavant et al., 2016). Above ~ 3 wt% F in melt, F concentrations
737 in mica increase drastically. This goes along with important changes in mica crystal chemistry
738 toward a more marked trioctahedral character. The F $K_d \text{ Li-Mi/L}$ reaches values between 2
739 and 3, in the range of biotite-liquid F partitioning data ($K_d \text{ Bt/L} = 2.3$, Icenhower and
740 London, 1997; average of 2.3, Pichavant et al., 2016). Therefore, F partitioning closely
741 follows the major element composition and structure of the mica. No attempt was made to
742 constrain the distribution of Li between mica and melt since in this study Li concentrations in
743 micas are calculated (see above).

744 The trace element data in charge B2N-25 allow the Qz/L partition coefficients for Li,
745 Al, P, Ti, Zn, Sn, Cs, W to be estimated (Table 9). For Li, the experimentally determined
746 partition coefficient (0.027) is significantly lower than calculated from the Yellowstone tuffs
747 (~ 0.073 , Neukampf et al., 2019) and the Macusani volcanics (0.17, Michaud and Pichavant
748 (2020).

749 Sn concentrations in charge B2N-25 yield very high partition coefficients for Li-mica
750 and hematite ($K_d \text{ Li-Mi/L} \sim 170$ and $K_d \text{ Hm/L} \sim 890$ respectively) using the separate Sn
751 analyses of Li-mica and hematite reported above and the glass analysis from Table 9. These
752 partition coefficients represent maximum values because Sn was progressively lost from the
753 melt to the capsule. However, $K_d \text{ Li-Mi/L}$ from this study is markedly higher than previous
754 muscovite/L partitioning data (0.38-0.49, Raimbault and Burnol, 1998; 0.48-0.49, Michaud
755 and Pichavant, 2020) implying that Sn is compatible in micas under our experimental
756 conditions. Given the high fO_2 of the B2N-25 experiment ($\sim \text{NNO}+3$), tin should be present as
757 Sn^{4+} (Linnen et al., 1995; 1996) and the very elevated $K_d \text{ Li-Mi/L}$ from this study suggests
758 that Sn^{4+} preferentially partitions in mica rather than in melt.

759

760 **DISCUSSION**

761 **Evaluation of kinetic issues**

762 In this study, crystallization experiments have been conducted at temperatures uncommonly
763 low (down to $\sim 550^\circ\text{C}$, Table 2; 3) for silica-rich melt compositions and, so, an evaluation of
764 kinetic issues is particularly necessary. Silicic crystal-melt systems have been known for long
765 to be difficult to equilibrate. In crystallization experiments, supercooling is of general concern
766 and, in melting experiments, superheating can be necessary to overcome the sluggish
767 dissolution kinetics of some phases, in particular quartz (Piwinskii and Martin, 1970;
768 Pichavant et al., 1987a and references therein).

769 Several observations indicate kinetic difficulties in our experiments despite the long
770 experimental durations. The most critical concerns saturation temperatures for major phases at
771 100 MPa which are lower than at higher pressures under water-saturated conditions (see
772 above), a behavior symptomatic of supercooled melts. The effect of melt supercooling is to
773 delay phase appearance for a given temperature and/or shift phase appearance to a lower
774 temperature for a given experimental duration. In supercooled charges which contain
775 proportions of crystals lower than equilibrium, melts compositions can be metastable. Results
776 of melting experiments (Pichavant et al., 1987a; Table S4) confirm the very low saturation
777 temperatures in the 100 MPa crystallization experiments. Qz saturation temperatures are
778 lower by $<115^\circ\text{C}$ in crystallization than in melting experiments on B1, and by 85°C on B2
779 (Fig. 2; 3). For Pl, differences between melting and crystallization saturation temperatures are
780 $<80^\circ\text{C}$ for B1 and 55°C for B2 (Fig. 2; 3). Af in crystallization experiments saturate at
781 temperatures 20°C lower than in melting experiments and, for Mu, the two saturation
782 temperatures are within $\pm 5^\circ\text{C}$ of each other (Fig. 2; 3). Therefore, the 100 MPa

783 crystallization experiments underestimate the Qz and Pl saturation temperatures and also
784 slightly the Af. We stress that this does not necessarily imply that saturation temperatures
785 from melting experiments approach equilibrium more closely because melting kinetics of
786 major phases can be sluggish, particularly for Qz as noted above.

787 In contrast, at 300 MPa for B2, Pl, Af and Li-Mi saturation curves in crystallization
788 and melting experiments are within 5°C of each other (Fig. 3). Qz is the only major phase
789 whose saturation temperature could be slightly underestimated since it persists up to 40°C
790 higher in melting than in crystallization experiments (Fig. 3). In the 200 MPa B2 experiments,
791 all major phases saturate at temperatures consistent with those at 300 MPa (Fig. 3). For B1,
792 saturation temperatures are all higher in crystallization than in melting experiments, most
793 probably because of the low sensitivity of the X-ray phase identification method used in the
794 melting experiments (Pichavant et al., 1987a). Therefore, there is no indication that
795 equilibrium is not approached in the 300 MPa B1 crystallization charges. We conclude that
796 melt supercooling issues in this study mainly concern the 100 MPa and are minor or absent in
797 the 200 and 300 MPa charges. Such a marked influence of pressure is most probably due to
798 the significantly higher melt H₂O concentrations at 300 and 200 than at 100 MPa which
799 would speed up crystallization kinetics at high pressures.

800 Major element compositions of major phases provide additional insights on
801 equilibrium states. Although the glasses are chemically homogeneous, some specific
802 compositions might correspond to metastable melts. This includes the 100 MPa glasses which
803 have normative Qz higher than the other glasses (Fig. 5) and could be too SiO₂-rich because
804 of Qz proportions less than equilibrium. Similarly, the distinctive normative Or-rich
805 composition of B2N-18 glass (Fig. 5) could indicate delayed Af appearance. The observation
806 that feldspar pairs mostly correspond to compositions at equilibrium on the ternary solvus
807 (Fig. 6) does not necessarily imply that all phases in our experiments reached equilibrium
808 compositions. The heterogeneous feldspars (and their compositions inside the solvus) and the
809 Al-rich quartz solid solutions are interpreted as remnants of non-equilibrium compositions
810 still present in some charges despite long experimental durations.

811

812 **Natural feldspar and mica compositions**

813 At the difference of the melting experiments where mineral assemblages and compositions are
814 inherited from the starting granites (Pichavant et al., 1987a), experiments from this study
815 provide information on compositions of phases crystallizing from B1 and B2 melts. This
816 allows crystallization of the Beauvoir granite to be reconstructed and compositions of major

817 mineral phases, mainly feldspars and micas, in the granite to be interpreted in the light of
818 experimental results.

819 Averages of natural feldspar compositions along the GPF drill core ($\text{Or}_1\text{Ab}_{98.5}\text{An}_{0.5}$ for
820 plagioclase, $n = 69$ and $\text{Or}_{90}\text{Ab}_{10}$ for K-feldspar, $n = 42$, Johan et al., 1988) are very close to
821 compositions in our starting rocks (Fig. 6). Only 5 plagioclases have $> 1\%$ An (maximum 2.5)
822 and 3 have $> 1.5\%$ Or (maximum 7.4, Johan et al., 1988). In comparison, experimental Pl
823 have An mostly $> 1\%$ and are much more K-rich ($\sim 10\%$ Or) than the natural compositions
824 (Fig. 6). Natural K-feldspars range from Or_{87} up to Or_{100} and only one composition (Or_{78} ,
825 Johan et al., 1988) approaches the Af experimental range (Or_{67-73} , Fig. 6). The natural feldspar
826 pair in the starting rocks yields equilibration temperatures of 391 and 407°C for pressures of
827 100 and 300 MPa respectively (Wen and Nekvasil, 1994) and, therefore, their compositions
828 were frozen $\sim 150^\circ\text{C}$ colder than the Beauvoir solidi (Fig. 2; 3). Clearly, most feldspars do not
829 preserve magmatic compositions. However, a few rare analyses (the most An- and Or-rich Pl
830 and the most Na-rich Af) suggest that primary compositions were close to the experimental
831 ranges. We conclude that feldspars in the Beauvoir granite have been extensively re-
832 equilibrated during the post-magmatic evolution.

833 Several mica crystallization stages have been recognized in the Beauvoir granite. The
834 earliest corresponds to relatively Li-poor Fe-bearing Li-Mi (Li-, Fe-muscovites of
835 Kosakevitch, 1976) subsequently interpreted as near-liquidus phases from relatively
836 unfractionated melts (Monier and Tegye, 1985; Monier et al., 1987). These compositions are
837 quite rare and represented in our dataset by only one EMPA analysis (the right most in Fig.
838 7b). They have not been exactly reproduced in the experiments, and the B2 Li-Mi
839 compositions have Mg-Li slightly higher than early stage micas (Fig. 7a; b). The main stage is
840 characterized by the crystallization of Li-, F-rich zinnwaldites to polyolithionites which
841 postdate the early stage compositions and have been previously interpreted as crystallized
842 from fractionated sub-liquidus melts (Monier et al., 1987). This mica crystallization stage has
843 been adequately simulated by the B1 experiments where Li-Mi compositions are closely
844 similar to the natural Beauvoir micas (Fig. 7).

845 The evolution of Beauvoir micas from early Li-poor Fe-bearing Li-Mi to zinnwaldites
846 and polyolithionites (increasing $\text{Fe}_t + \text{Mn} + \text{Ti} - \text{Al}^{\text{VI}}$ with decreasing Mg – Li) parallels the
847 compositional shift of experimental micas from the B2 to the B1 charges. This trend ((1) in
848 Fig. 7a) is positively correlated with the melt F content (Fig. 8). It describes the compositional
849 evolution of Li-Mi crystallizing from a progressively more F-rich peraluminous granitic melt.
850 The mica becomes more silicic, less aluminous and its trioctahedral character increases, a

851 general evolution which can be understood from the structural role of F in aluminosilicate
852 melts (increase of a_{SiO_2} , complexing between F and Al, Manning, 1981; Xiong et al., 1998;
853 1999). The F control of experimental mica compositions along trend (1) is also marked in
854 PHP rare metal granites. In the Argemela intrusion which contains significantly less F than
855 Beauvoir (0.5-1.0 wt% F, Michaud and Pichavant, 2020), most micas plot at the high Mg-Li
856 end of trend (1), in the same domain as experimental B2 micas (Fig. 7a; b). Thus, Beauvoir
857 melts with F contents intermediate between our two starting compositions would be expected
858 to crystallize Li-Mi filling the gap between the B1 and B2 experimental compositions, in the
859 range inferred for the early stage micas. Another evolutionary trend ((2) in Fig. 7a and 8) is
860 apparent both in natural and experimental micas. $\text{Fe}_t + \text{Mn} + \text{Ti} - \text{Al}^{\text{VI}}$ decreases with
861 decreasing Mg – Li as illustrated by the Beauvoir mica separate data (Fig. 7b) and by the
862 experimental compositions from B1N-23, -22 to -21 (i.e., upon decreasing temperature and
863 melt fraction, Fig. 4; 7a; 8). This trend is a mica fractionation trend, i.e., it reflects the
864 removal and progressive exhaustion of mafic components from the F-rich peraluminous
865 granitic melt as mica crystallizes. The observed zonation of mica compositions across the
866 GPF drill core (zinnwaldites at the base and polyolithionites at the top, Monier et al., 1987) is
867 consistent with crystallization from a more fractionated melt in the latter than in the former
868 case. We conclude that the composition of Beauvoir micas preserves two distinct magmatic
869 evolutionary trends (Fig. 7a).

870 However, several mica chemical features do not fit in an entirely magmatic
871 interpretation. Micas from the B1 unit have FeO_t and MgO concentrations (respectively < 1
872 and < 0.1 wt%, Monier et al., 1987) much lower than in B1 charges (~1-5.5 and ~0-2 wt%
873 respectively, Table 7; S3). Conversely, MnO concentrations in natural micas are higher (on
874 average ~0.8 wt% and up to ~1.5 wt% MnO in the B2 unit, Monier et al., 1987) than in
875 experimental micas (0-0.4 wt%, Table 7; S3). This suggests that FeO_t , MgO and MnO in
876 micas do not represent magmatic concentrations and have been modified by post-magmatic
877 fluid circulations.

878

879 **Redox conditions and magmatic F, Li concentrations**

880 The presence of hematite and magnetite in our experiments constitutes the main point of
881 divergence between the experimental and natural phase assemblages since the Beauvoir
882 granite contains no Fe-Ti oxide except secondary rutile (e.g., Monier et al., 1987; Cuney et
883 al., 1992). Below, we exclude the possibility that Fe-Ti oxides were once present but all
884 completely destroyed during late-stage crystallization processes, either magmatic or post-

885 magmatic. Therefore, the lack of Fe-Ti oxides in the granite can be used to constrain the
886 redox conditions. In our experiments, hematite was found in the most oxidizing (\sim NNO+3)
887 and magnetite in less oxidizing (NNO+1) charges (Lahlafi, 1997). It is recalled that no Fe-Ti
888 oxide crystallized in the NNO-1 and NNO-1.4 experiments. This constrains the fO_2 of the
889 Beauvoir granite to a range more reducing than NNO+1 and more oxidizing than NNO-1.4
890 since Bt would be present under such highly reducing conditions (Table 3), which is not
891 observed (Cuney et al., 1992). Previously, an oxidizing magmatic evolution has been
892 recognized at Beauvoir (Cuney et al., 1992 and see the Ta, Nb, Sn mineralization section
893 below). A progressive oxidizing trend is developed at the scale of the GPF drill core,
894 superimposed on more reducing conditions as demonstrated by lepidolites with low Fe^{3+}/Fe^{2+}
895 at the base, presence of uraninite in B3 and B2 (Cuney et al., 1992) and absence of Fe-Ti
896 oxides in the lower granite units. Therefore, the $< NNO+1$ to $> NNO-1.4$ range inferred from
897 our experiments is interpreted as representative of early magmatic conditions. This implies
898 that most of our crystallization experiments have been performed at an inappropriately
899 oxidizing fO_2 . This has minor consequences for the stability of major phases since varying
900 fO_2 from \sim NNO+3 to NNO-1 has no systematic influence on the location of saturation curves
901 (Fig. 2; 3). However, fO_2 has definitely an influence on phase compositions (see above the Fe
902 and Sn contents of experimental micas). The specific behavior of some trace elements
903 (compatible for Sn and strongly incompatible for U, Fig. 9) also reflect the oxidizing
904 conditions of most of our experiments.

905 Crystallization of topaz and amblygonite in experiments from this study provides
906 unambiguous confirmation that both can be magmatic phases in the Beauvoir granite (Monier
907 and Tegye, 1985; Rossi et al., 1987; Cuney et al., 1992). Previously, topaz has been reported
908 to be present in early crystallization experiments. However, topaz identification was uncertain
909 (see Pichavant et al., 1987a). In this study, topaz was identified by SEM and analyzed by
910 electron microprobe. It appears only in charge B1N-21 at 580°C, 300 MPa which is one of the
911 two most evolved B1 charges. Melt in B1N-21 has the highest F concentration (3.31 wt% if
912 unnormalized) and one of the two highest normative Corundum (6.5%, Fig. 4; Table 4). The
913 rarity of topaz in our experimental charges contrasts with its ubiquitous presence in the
914 Beauvoir granite. Previous work has emphasized F and Li loss from the crystallizing magma
915 and F concentrations as high as 4 wt% have been considered possible for Beauvoir melts
916 (Cuney et al., 1992). An average F concentration of 4.9 wt% has been found in re-
917 homogenized melt inclusions from the B1 unit (Kovalenko et al., 1998). Such F concentration
918 levels are on average above our experimental melt concentrations although the most F-

919 enriched B1 glasses approach the 4 wt% range, if normalized to 100% anhydrous (i.e., 3.70
920 wt% F for B1N-21). However, experimental topazes have elevated F concentrations (at.
921 $F/(F+OH) = 0.89-0.99$, Table 8). This is higher than the old values from the GPF drill core (at.
922 $F/(F+OH) = 0.78-0.92$, Monier and Tegye, 1985) but in the range of recent data for topaz in
923 the B1 unit (0.96-1.04) and in a dike associated with the B2 unit (0.93-0.94, L. Monnier, pers.
924 comm. 2022). In other words, there is no indication that the F concentration in the B1N-21
925 experimental melt was much lower than in natural Beauvoir melts. In comparison,
926 experimental F concentrations in B2 charges were probably insufficiently high given the
927 relatively Li-, F-poor Li-Mi compositions crystallized in B2 (Fig. 7 and see above).

928 Topaz is a rather low temperature and late crystallizing phase in our experiments,
929 appearing later than mica in the B1 300 MPa crystallization sequence. The data suggest that a
930 minimum F content of 3-3.5 wt% in the melt is necessary for topaz to be stable (see also
931 Soufi, 2021). In our experiments, such elevated F contents are reached in the sub-liquidus
932 domain where quartz and feldspar crystallization enrich the residual melt in F (Fig. 4). The
933 low topaz crystallization temperature observed in this study is consistent with at. $F/(F+OH)$
934 ~ 1 in experimental topaz since at higher temperatures F partitions less favorably in topaz than
935 in other phases (Soufi, 2021). In melts with $F \gg 3-3.5$ wt%, topaz with at. $F/(F+OH) < 1$ is a
936 high temperature near-liquidus phase (Xiong et al., 1999; Lukkari and Holtz, 2007). This is
937 not what is observed at Beauvoir where topaz is a sub-liquidus low temperature phase and has
938 at. $F/(F+OH) \sim 1$. We conclude that the rarity of topaz in our experiments is simply a
939 reflection of the small number of low melt fraction charges available for the B1 composition
940 (Fig. 4). At 580°C and 300 MPa, melt fractions are still high ($f=67$ wt% in the topaz-bearing
941 charge B1N-21, Table 3) so that, during the crystallization of the last (>50 wt%) B1 melt
942 fraction, topaz should be present. Under these conditions, the melt F content should also reach
943 the 4 wt% level typical of the B1 unit (Cuney et al., 1992; Kovalenko et al., 1998).

944 Experimental amblygonites from this study are richer in F (~13 wt% F, Table 8) than
945 their natural counterparts in the Beauvoir granite (~10.5-11.5 wt% F, London et al., 2001).
946 Yet, it is possible that amblygonites from the more evolved B1 and B2 units have higher F
947 contents since the data above (from Ph. Rossi) are for the B3. Apatites in experiments and in
948 the GPF drill core (Johan et al., 1988; Cuney et al., 1992) are both F-rich, the natural having
949 generally higher MnO concentrations (~3 to ~6 wt%) than the experimental compositions
950 (Table 8). In conclusion, the range of F melt concentrations generated in our B1 charges is in
951 the range or slightly lower than in the natural magma. However, the B2 experiments probably
952 underestimate the F concentrations in the natural magma. In fact, our starting B2 sample,

953 although stratigraphically from the B2 unit, has geochemical characteristics typical of B'2
954 compositions (high SiO₂ and low Al₂O₃, P₂O₅ and F, Raimbault et al., 1995). This explains
955 our experimental results in particular the relatively Li-, F-poor compositions of Li-Mi
956 crystallized in B2 (see above).

957

958 **Crystallization and conditions of emplacement of the Beauvoir granite**

959 Results from this study confirm the very low magmatic temperatures of the Beauvoir granite.
960 Further to the solidi previously discussed (Fig. 2; 3; Pichavant et al., 1987a), the 100 MPa
961 liquidus temperatures are 715°C for B1 and 720°C for B2 (from the melting experiments
962 because of melt supercooling in the 100 MPa crystallization experiments) and the 300 MPa >
963 640°C for B1 (620°C from the melting experiments) and 640°C for B2 (670°C from the
964 melting experiments). Thus, in the 100-300 MPa range, the two Beauvoir compositions
965 studied begin to crystallize at temperatures lower than the H₂O-saturated haplogranite solidus
966 (Tuttle and Bowen, 1958; Fig. 2; 3). Such magmatic conditions are exceptional and only
967 approached by flux-rich granitic compositions (Xiong et al., 2002; Lukkari and Holtz, 2007;
968 Aksyuk et al., 2016), granitic pegmatites (Burnham and Nekvasil, 1986) and by the Macusani
969 glass (Pichavant et al., 1987b; London et al., 1989).

970 Two types of experimental results suggest that the B1 and B2 granite units have the
971 significance of crystallized liquids. First, the four major phases (Qz, Pl, Af, Li-Mi) saturate in
972 very narrow temperature intervals, both in B1 and B2 (Fig. 2; 3). Saturation temperature
973 intervals are of 20-30°C at 100 MPa, probably a consequence of melt supercooling, and of 30-
974 40°C at 300 MPa. Thus, liquidus temperatures are only a little higher than temperatures of
975 multiple saturation of the melt with the four major phases. This requires that the B1 and B2
976 starting granites are compositionally close to their respective multiply saturated melts.
977 Second, residual melts in both the B1 and B2 charges little depart from their respective
978 starting compositions (Fig. 5). Deviations are within 2-3% normative, at the exception of a
979 few charges (the 100 MPa, B2O-38/40, B2N-18, Fig. 5) discussed previously. Such a
980 behavior again requires that compositions of liquidus and multiple saturation melts are close
981 to each other and, so, the two starting compositions must represent essentially frozen liquids.

982 This result opens perspectives for constraining the crystallization pressure of Beauvoir
983 melts. Experimental data are available on the influence of pressure (under water-saturated
984 conditions, Tuttle and Bowen, 1958) and of additional components at constant pressure (F, Li,
985 P, A/CNK, (F: Manning, 1981; Li: Martin, 1983; A/CNK: Holtz et al., 1992a; b; P: London et
986 al., 1993) on the location of minima and eutectics in the haplogranite system (Fig. 5).

987 However, the combined effect of these different variables is as yet undetermined. As an
988 illustration, the F-rich B1 and B2 compositions plot at normative Quartz contents significantly
989 above the minimum liquidus points with added F at 100 MPa (Manning, 1981; Fig. 5). This is
990 attributed to the influence of very high A/CNK and normative Corundum in Beauvoir melts
991 (Holtz et al., 1992a; b). Therefore, pressures of Beauvoir melt crystallization cannot be
992 constrained directly from the reference liquidus points presently available. Melts multiply
993 saturated with the four major phases at 300 MPa plot closer than at 100 MPa to their
994 respective starting compositions which would tend to suggest that 100 MPa is too low a
995 pressure for early crystallization. However, the 100 MPa melts probably represent metastable
996 compositions (see above). We must conclude that no reliable barometric information can be
997 presently extracted from the major element compositions of Beauvoir granites.

998 Petrographic observations and textural evidence have shown that albite and lepidolite
999 are the two earliest major phases to crystallize in Beauvoir magmas (Kosakevitch, 1976).
1000 Mica is earlier than albite in some samples but the reverse is observed in others and, so, the
1001 two phases can be considered to appear almost together on the liquidus. Quartz, frequently
1002 euhedral and with albite inclusions present in external growth zones (M. Cuney, personal
1003 communication, 2022), is also an early phase, being followed by K-feldspar. By reference to
1004 the P-T diagrams, these textural relations require that crystallization started at pressures > 300
1005 MPa for B1 and at ~200-250 MPa for B2 (Fig. 2; 3). Final solidification of the pluton took
1006 place at its emplacement level (~ 80 MPa from the P-T conditions of exsolution of the
1007 magmatic fluids, Cuney et al., 1992). Therefore, pressures of early (> 200-250 MPa) and final
1008 crystallization (80 MPa) are significantly different and Beauvoir melts started to crystallize at
1009 depths higher than their emplacement level. Since they represent crystallized liquids, this
1010 implies that little melt-crystals segregation took place during final ascent.

1011

1012 **Fractionation in the Beauvoir granite**

1013 The major and trace element zonation observed among the different Beauvoir granite units
1014 has been used as a reference example of chemical fractionation in rare metal granites (Zhu et
1015 al., 2001; Costi et al., 2009; Cuney and Marignac, 2015; Pollard, 2021). Although
1016 mechanisms of fractionation in rare metal granites and pegmatites are still debated, this study
1017 demonstrates that most of the distinctive mineralogical and geochemical characteristics of the
1018 Beauvoir granite result from magmatic rather than hydrothermal post-magmatic processes.
1019 Albitic plagioclase, Li-, F-rich mica, topaz and amblygonite are of magmatic origin. A/CNK,

1020 F and P₂O₅ concentrations increase and trace elements fractionate as a result of progressive
1021 crystallization (Fig. 4; 9).

1022 The major element signature of fractionation at Beauvoir is examined by comparing
1023 granite whole rock and experimental glass compositions (Fig. 11). The granite units globally
1024 define a progressive fractionation trend marked by increasing A (= 1000 (Al – (Na + K +
1025 2Ca) at.) at decreasing B (=1000 (Fe+Mg+Ti) at., Fig. 11a). This natural trend, classical for
1026 rare metal granites (Linnen and Cuney, 2005; Cuney and Marignac, 2015), is well reproduced
1027 by the experimental data (Fig. 11b) and this confirms that Beauvoir granites are reliable
1028 proxies for magmatic liquid compositions. In the Q*³-B*³-F*³-M*³ representation, all
1029 granite compositions plot close to each other but the two groups of experimental glasses
1030 define fractionation trends marked by increasing M*³/F*³ at constant Q*³ (Fig. 11c; d).
1031 These trends match with the evolutions at constant normative Quartz contents in the Qz-Ab-
1032 Or diagram (Fig. 5) and with the increase in normative Corundum with decreasing melt
1033 fraction and progressive fractionation (Fig. 4). The 100 MPa glasses again deviate from the
1034 trends, being distinctly shifted toward Q*³-rich compositions (Fig. 11d). We note that the two
1035 B1 and B2 glass fractionation trends do not coincide (Fig. 11b; d) as observed in the Qz-Ab-
1036 Or diagram where the B1 and B2 data points plot in different fields (Fig. 5). The two B1 and
1037 B2 trends also remain separate from each other when plotted as a function of melt fraction
1038 (Fig. 4). Therefore, the B1 and the B2 samples do not belong to the same evolutionary trend
1039 and this demonstrates that our two Beauvoir starting compositions are not comagmatic.
1040 Reminding that our B2 starting rock has geochemical attributes of B'2 compositions, the
1041 experimental glasses thus confirm previous conclusions from whole rock data that the B and
1042 B' represent two distinct magmatic series (Raimbault et al., 1995). Experimental data are
1043 lacking to discuss the mutual relationships within the B series, i.e., between the B1, B2, B3
1044 compositions which are geochemically close to each other (Raimbault et al., 1995; Fig. 11c).
1045 A progressive decrease in normative Quartz is observed at the scale of the GPF drill core from
1046 the B3, B2 to the B1 units, positively correlated with an increase of the F content (Cuney et
1047 al., 1992). This evolution has been attributed to the role of F on granite phase equilibria
1048 (Manning, 1981; Cuney et al., 1992) although other components (Li, P, A/CNK) must also be
1049 involved (see above, Fig. 5). However, the important point to be emphasized is that this
1050 geochemical variability among the B units is not generated at the emplacement level of the
1051 intrusion because differentiation of Beauvoir melts yields derivative products with almost
1052 constant normative Quartz (Fig. 5; 11). Therefore, our results suggest that the B1, B2 and B3
1053 granite units represent three independent magma batches generated at depth.

1054 Trace element fractionation generated in experiments from this study are compared
1055 with those observed in Beauvoir granites (Fig. 12). For most elements, the data points plot in
1056 the upper right and lower left quadrants and so most elements fractionate (become either
1057 enriched or depleted) similarly in experiments and from B2 to B1. This strongly suggests that
1058 magmatic crystallization is at the origin of the trace element fractionations observed at
1059 Beauvoir. Elements enriched include Li, Be, U, F, P, Rb, W while Sr, Pb are depleted. The
1060 former group plots close to the 1:1 line (except Be, Fig. 12) and so, for these elements, the
1061 experimentally generated enrichments are similar in magnitude to those observed from B2 to
1062 B1. Cs is an exception (enriched in the experiments but depleted from B2 to B1) explained by
1063 the high Cs concentration in our B2 starting sample (Raimbault et al., 1995). For Sr and Pb,
1064 the experimental depletions are respectively higher and lower than from B2 to B1. Ti is
1065 significantly fractionated neither in the experiments nor in the granites (Raimbault et al.,
1066 1995). Zn and Mn plot close to the vertical axis (Fig. 12) and so they do not fractionate in the
1067 experiments (Fig. 9), although both are depleted from B2 to B1 (Raimbault et al., 1995).
1068 However, the significant variations of whole rock Mn concentrations in the top part of B1
1069 (where a 20 m Mn-enriched “fringe zone” overlies a Mn-depleted level of about the same
1070 thickness, Raimbault et al., 1995) and the observation that the natural lepidolites and apatites
1071 are generally richer in Mn than their experimental counterparts (see above) suggests Mn re-
1072 distribution during cooling of the intrusion.

1073

1074 **Origin of the Beauvoir granite**

1075 The constraints placed on the evolution of the Beauvoir granite raise the question of its origin.
1076 There is agreement that it is ultimately related to the tectonic evolution, thermal maturation
1077 and partial melting of the continental crust during late Variscan times (e.g., Gebelin et al.,
1078 2009; Lardeaux, 2014; Augier et al., 2015; Barbey et al., 2015). In this framework, the
1079 genesis of highly fractionated granitic melts such as Beauvoir can be accounted for by two
1080 possible models (1) extreme differentiation from less evolved crustal parental magmas (e.g.,
1081 Raimbault et al., 1995; Cerny et al., 2005b; London, 2008; Hulsbosch, 2019) or (2) direct
1082 anatectic origin (e.g., Stewart, 1978; Christiansen et al., 1986; Cuney and Autran, 1987;
1083 Nabelek et al., 1992a; b; Muller et al., 2017). In fact, results of this study do not favor any of
1084 these two models. The thermal characteristics of Beauvoir melts (liquidus temperatures of
1085 670°C maximum at 300 MPa) make it possible that they represent near-solidus residual melts
1086 from the crystallization of parental peraluminous leucogranitic magmas (e.g., Scaillet et al.,
1087 1995). Alternatively, these liquidus temperatures are within the P-T evolution of metamorphic

1088 units during late Variscan times and appropriate for crustal melting either under fluid-absent
1089 (muscovite-dehydration melting) or fluid-present (Cuney and Autran, 1987; Gardien et al.,
1090 1997; Lardeaux, 2014; Augier et al., 2015; Barbey et al., 2015; Michaud et al., 2021).
1091 Although an origin of rare metal granites by fluid-absent melting of residual lithologies
1092 previously enriched by the removal of a low-temperature melt fraction has been commonly
1093 proposed (see Michaud et al., 2021), this model can be discarded for Beauvoir since it would
1094 imply melt generation at temperatures much higher (at least 100°C higher for biotite-
1095 dehydration melting) than the B1 and B2 liquids.

1096 Geological evidence in support of any of the two models is ambiguous. At Beauvoir,
1097 there is no candidate parental magma exposed (it would likely be represented by granitic units
1098 intermediate between the Beauvoir and Colettes) but other granites are probably present at
1099 depth (Aubert, 1969; Cuney and Marignac, 2015). The anatectic model (Cuney and Autran,
1100 1987) has been ruled out on a geochemical basis (Raimbault et al., 1995). However, updated
1101 versions of the model have been recently elaborated (Cuney and Barbey, 2014). We also note
1102 that the differentiation model called for initial melts slightly enriched in rare metals (Sn and
1103 W) originated from geochemically specialized lower crustal sources (Raimbault et al., 1995).
1104 Differentiation, postulated to start at the anatexis level, generated strongly peraluminous, F-
1105 and P-bearing and rare metal enriched melts (Raimbault et al., 1995). Further differentiation
1106 during ascent led to the individualization of the B and B' magma series, the B1 and B2 units
1107 representing two variably differentiated liquids from the same B magma and the B3 possibly a
1108 slightly cumulative part of the B2 (Raimbault et al., 1995).

1109 Our experimental results confirm that the Beauvoir intrusion was built from at least
1110 two distinct magmatic batches (see above, Fig. 11), all strongly peraluminous, P- and F-rich
1111 and with elevated rare metal contents. They stress the importance of deep-seated fractionation
1112 processes rather than those occurring in situ at the emplacement level. We note that the
1113 fractionation mechanisms supporting the differentiation model are still poorly defined. For
1114 example, amplification of melt structure and viscosity fluctuations during ascent has been
1115 proposed to promote dynamic separation between more evolved less polymerized and less
1116 evolved more polymerized melts (Raimbault et al., 1995). A combination of Bagnold effect
1117 and filter-press processes has been suggested to control fractionation during ascent of rare
1118 metal granitic melts in dykes (Marignac et al., 2009; Cuney and Marignac, 2015). These
1119 mechanisms need further testing and validation and we conclude that the origin of the
1120 Beauvoir melts still requires more work.

1121

1122 Ta, Nb, Sn mineralization

1123 Ta and Nb are both enriched although much more in granites than in experiments as indicated
1124 by their position well above the 1:1 line (Fig. 12). The experimental Ta and Nb contents
1125 (maximum 34 ppm Ta and 49 ppm Nb, Table 9) stay globally in the range of B2
1126 concentrations (~45 ppm Ta and ~50 ppm Nb, Raimbault et al., 1995). They do not reach the
1127 high to very high values observed in B1, ~150 ppm Nb and up to ~450 ppm Ta in the upper
1128 “fringe zone” (Raimbault et al., 1995). C-T solubility data (Linnen and Keppler, 1997;
1129 Linnen, 1998; Van Lichtervelde et al., 2018) have demonstrated higher solubility limits for
1130 the Ta than for the Nb C-T end-members which provides an explanation for the fractionation
1131 of Ta over Nb and the preferential Ta enrichment observed in B1. They also stress the
1132 importance of the melt Li and F concentrations in controlling C-T saturation in rare metal
1133 magmas. In this respect, it is worth emphasizing that our trace element data concern the B2
1134 magma only. Fractionation in B2 charges increases F, P, Li (Fig. 4; 9) which leads to a C-T
1135 solubility increase responsible for the Ta and Nb enrichments observed in residual melts (Fig.
1136 9). Fractionation in B1 generates melts more F-, P- and presumably more Li-rich than in B2
1137 (Fig. 4). Thus, Ta and Nb melt concentrations at C-T saturation higher than in B2 charges can
1138 be expected and it is possible that residual B1 melts would have Ta and Nb approaching the
1139 concentration levels observed in the B1 unit. This interpretation is consistent with a magmatic
1140 control of the Ta and Nb mineralization at Beauvoir (Raimbault et al., 1995).

1141 For Sn, the strong enrichment observed in granites (B2: 300-800 ppm; B1: 800-1600
1142 ppm, Raimbault et al., 1995) has no counterpart in experimental melts (Fig. 9; 11). In the
1143 ~NNO+3 charges, Sn melt concentrations are controlled by cassiterite crystallization, Sn
1144 partitioning toward Li-Mi and Hm and Sn loss to the capsule. In Beauvoir granites, Sn is
1145 carried mainly by magmatic cassiterite implying that Beauvoir melts became cassiterite-
1146 saturated at some point during their evolution. Sn whole rock concentrations and their
1147 variations within and between units mainly reflect the proportion of cassiterite crystals which,
1148 at cassiterite saturation, is controlled by the Sn content of the initial melt (a high Sn melt will
1149 produce a higher amount of cassiterite than a low Sn melt when brought at cassiterite
1150 saturation). Experimental data have shown that cassiterite solubility mainly depends on fO_2 at
1151 a given temperature and melt A/CNK, the influence of F being negligible (Bhalla et al., 2005
1152 and references therein). Cassiterite solubilities are of the order of 100 ppm at NNO+2-3 but
1153 increase rapidly to the several 1000 ppm range or more under reducing conditions (Bhalla et
1154 al., 2004 and references therein). Therefore, magmatic oxidation is a very efficient
1155 mechanism to promote cassiterite saturation (Linnen et al., 1995; 1996; Pichavant et al.,

1156 2016). At Beauvoir, several lines of evidence for a late progressive oxidation of the intrusion
1157 have been presented (Cuney et al., 1992 and see the redox section above). This promoted
1158 cassiterite saturation in melts enriched in Sn by crystallization-differentiation at the intrusion
1159 level (Michaud and Pichavant, 2020). Sn contents of initial melts are unknown but can be
1160 estimated to be of a few hundred ppm for B2 or more for B1, considering the respective
1161 whole rock concentrations. Some Sn was extracted and transported by postmagmatic
1162 hydrothermal fluids as suggested by the Sn depletion in the upper part of the cupola (Cuney et
1163 al., 1992; Raimbault et al., 1995). The final Sn distribution in the intrusion was probably
1164 controlled by the re-mobilization by late postmagmatic fluids of Sn from cassiterite and Sn-
1165 bearing micas.

1166

1167 **Broader implications for granitic magmatism**

1168 One important result from this study is the confirmation of the very low magmatic
1169 temperatures of the Beauvoir granites (see above). This demonstrates that natural granites
1170 solidi are in fact quite variable. Many (Burnham and Nekvasil, 1986; London et al., 1989;
1171 Xiong et al., 2002; Lukkari and Holtz, 2007; Aksyuk et al., 2016; this study) are much colder
1172 than the H₂O-saturated haplogranite solidus (Tuttle and Bowen, 1958). It is important to
1173 consider the latter as a theoretical reference rather than as a universal benchmark for the
1174 solidus of granitic rocks (see Ackerson et al., 2018).

1175 Two-feldspar thermometry and more recent geothermobarometric methods such as Ti-
1176 in-quartz generally have yielded temperatures between 400 and 600°C for the crystallization
1177 of granitic rocks (Moazzen and Droop, 2005; Glazner and Johnson, 2013; Ackerson et al.,
1178 2018; London et al., 2020). Despite the particularly low solidus temperatures for Beauvoir,
1179 albite and K-feldspar record equilibration temperatures that are still ~150°C colder (see
1180 above). These subsolidus equilibration temperatures have led to speculations on mechanisms
1181 of final consolidation of granitic bodies (Ackerson et al., 2018). By far the most likely
1182 explanation is that such low crystallization temperatures indicate fluid-assisted mineral re-
1183 equilibration during the post-magmatic evolution, the interpretation proposed above for
1184 Beauvoir and other granites (Moazzen and Droop, 2005; Lukkari and Holtz, 2007). For
1185 example, feldspars experience major chemical, textural and structural transformations during
1186 the cooling of granitic bodies (Parsons, 1978; Smith and Brown, 1988). These reactions are
1187 triggered by fluid percolation and fluid-rock interaction as demonstrated by fluid inclusions
1188 (Kontak et al., 1984) and stable isotope studies (Taylor and Forester, 1971; Fouillac and
1189 Rossi, 1991). Both quartz and feldspars in the GPF drill core show a systematic decrease in

1190 $\delta^{18}\text{O}$ from base to top, interpreted to indicate interaction with meteoric fluids at high
1191 temperature (Fouillac and Rossi, 1991; Cuney et al., 1992). Thus, despite its exceptionally
1192 low solidus temperature which makes the post-magmatic “window” narrower, the Beauvoir
1193 granite reminds us of the importance of re-equilibration processes involving hydrothermal
1194 fluids in the final solidification of granitic plutons.

1195

1196 CONCLUSIONS

1197 Crystallization experiments performed on two Beauvoir granite samples (B1 and B2) simulate
1198 the magmatic evolution and chemical fractionation in strongly peraluminous, F- and P-rich
1199 rare metal magmas. The main conclusions of the study are the following:

1200 - Beauvoir melts crystallize a magmatic assemblage which includes quartz, albitic
1201 plagioclase, K-feldspar, Li-mica, apatite, cassiterite and columbo-tantalite. This assemblage,
1202 joined by topaz and amblygonite in the B1 experiments, closely reproduces the granite
1203 mineralogy. Fe-Ti oxides are present for $f\text{O}_2$ between $\sim\text{NNO}+3$ and $\text{NNO}+1$.

1204 - The crystallization experiments confirm the very low solidus temperatures of the
1205 granites. In the 100-300 MPa range, B1 and B2 have liquidus temperatures lower than the
1206 H_2O -saturated haplogranite solidus, an exceptional behavior characteristic of flux-rich
1207 granitic melts.

1208 - Upon progressive crystallization, residual melts become more F- and P-rich. Melt
1209 major element compositions evolve at constant Qz contents and become more peraluminous.
1210 F-rich topaz and amblygonite crystallize as sub-liquidus phases from these evolved melts.

1211 - Comparison between natural and experimental mineral compositions suggest that
1212 feldspars in the granite have been re-equilibrated during the post-magmatic evolution. Both
1213 natural and experimental micas follow similar trends from Li phengites to zinnwaldites and
1214 polyolithionites. Compositions closely similar to the natural Beauvoir micas crystallize in the
1215 B1 experiments.

1216 - The $f\text{O}_2$ ranged from $< \text{NNO}+1$ to $> \text{NNO}-1.4$ during the early magmatic stage, later
1217 evolving to highly oxidizing conditions at the magmatic-hydrothermal transition. F, Li melt
1218 concentrations in B1 experiments reproduce or are slightly lower than the concentrations in
1219 the natural magma. Those in B2 underestimate most natural B2 concentrations since the
1220 starting sample has geochemical attributes of B'2 compositions. Glass major elements suggest
1221 that our two starting compositions are not comagmatic. The major distinction between the B
1222 and B' geochemical series at Beauvoir is confirmed. The intrusion was made from at least two
1223 independent magma batches.

1224 - The low liquidus temperatures, narrow saturation temperature intervals and grouped
1225 residual melts imply that the Beauvoir granites represent crystallized liquids. Beauvoir melts
1226 started to crystallize at depths (≥ 200 MPa) significantly higher than their emplacement level
1227 (80 MPa). Our results stress the importance of deep-seated processes (either partial melting or
1228 differentiation) as a way to generate the specific major element compositions of Beauvoir
1229 melts.

1230 - Upon magmatic crystallization, the melt becomes enriched in Li, Be, B, Rb, Cs, W,
1231 U, Nb, Ta and depleted in Sr and Pb. Fractionation trends observed in Beauvoir granites are
1232 similar to those in experiments suggesting that crystallization at the emplacement level
1233 controls chemical fractionation for most elements.

1234 - Sn is compatible in micas and hematite under highly oxidizing conditions. This,
1235 together with cassiterite crystallization and Sn loss, prevents Sn enrichment in residual melts.
1236 Cassiterite saturation in Beauvoir magmas was promoted by late magmatic oxidation of the
1237 intrusion.

1238 - Most of the mineralogical and geochemical characteristics that make the Beauvoir
1239 granite distinctive result from magmatic rather than hydrothermal processes. However,
1240 feldspar and mica compositions and fractionation patterns for some elements (Mn) confirm
1241 that post-magmatic fluids were involved in the final stages of crystallization of the intrusion.

1242

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1253

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1257

1258 **SUPPLEMENTARY DATA**

1259 Supplementary data for this paper are available at Journal of Petrology online.

1260

1261 **REFERENCES**

- 1262 Ackerson, M.A., Mysen, B.O., Tailby, N.D. & Watson, E.B. (2018). Low-temperature
1263 crystallization of granites and the implications for crustal magmatism. *Nature*,
1264 doi.org/10.1038/s41586-018-0264-2.
- 1265 Aïssa, M., Weisbrod, A. & Marignac, C. (1987). Caractéristiques chimiques et
1266 thermodynamiques des circulations hydrothermales du site d'Echassières. *Géologie de la*
1267 *France* **2-3**, 335-350.
- 1268 Aksyuk, A.M., Konyshev, A.A., Korzhinskayaa, V.S. & Shapovalova, Yu.B. (2016).
1269 Experimental study of physical and chemical melting conditions of rare-metal granites at
1270 the Voznesenka ore cluster, Primorye Region. *Doklady Earth Sciences* **470**, 921–923.
- 1271 Albarède, F. (1995). *Introduction to geochemical modelling*. Cambridge University Press, 543
1272 p.
- 1273 Aubert, G. (1969). *Les coupoles granitiques de Montebros et d'Echassières (Massif central*
1274 *français) et la génèse de leurs minéralisations en étain, lithium, tungstène et béryllium*.
1275 Mémoire BRGM 46, 345 p.
- 1276 Audétat, A., Garbe-Schönberg, D., Kronz, A., Pettke, T., Rusk, B., Donovan, J.J. & Lowers,
1277 H.A. (2014). Characterisation of a natural quartz crystal as a reference material for
1278 microanalytical determination of Ti, Al, Li, Fe, Mn, Ga and Ge. *Geostandards and*
1279 *Geoanalytical Research* **39**, 171-184.
- 1280 Augier, R., Choulet, F., Faure, M. & Turrillot, P. (2015). A turning-point in the evolution of
1281 the Variscan orogen: the ca. 325 Ma regional partial-melting event of the coastal South
1282 Armorican domain (South Brittany and Vendée, France). *Bulletin de la Société*
1283 *Géologique de France* **186**, 63-91.
- 1284 Barbey, P., Villaros, A., Marignac, C. & Montel, J.-M. (2015). Multiphase melting, magma
1285 emplacement and P-T-time path in late-collisional context: the Velay example (Massif
1286 Central, France). *Bulletin de la Société Géologique de France* **186**, 93-116.
- 1287 Bénard, F., Moutou, P. & Pichavant, M. (1985). Phase relations of tourmaline leucogranites
1288 and the significance of tourmaline in silicic magmas. *Journal of Geology* **93**, 271-291.
- 1289 Bhalla, P., Holtz, F., Linnen, R.L. & Behrens, H. (2005). Solubility of cassiterite in evolved
1290 granitic melts: effect of T, fO₂, and additional volatiles. *Lithos* **80**, 387-400.

- 1291 Breiter, K. (2012). Nearly contemporaneous evolution of the A- and S-type fractionated
1292 granites in the Krušné hory/Erzgebirge Mts., Central Europe. *Lithos* **151**, 105–121.
- 1293 Breiter, K., Fryda, J., Seltmann, R. & Thomas, R. (1997). Mineralogical evidence for two
1294 magmatic stages in the evolution of an extremely fractionated P-rich rare-metal granite:
1295 the Podlesí stock, Krušné Hory, Czech Republic. *Journal of Petrology* **38**, 1723-1739.
- 1296 Breiter et al. (2014). Trace element composition of quartz from different types of pegmatites:
1297 A case study from the Moldanubian Zone of the Bohemian Massif (Czech Republic).
1298 *Mineralogical Magazine* **78**, 703-722.
- 1299 Burnham, C.W. & Nekvasil, H. (1986). Equilibrium properties of granite pegmatite magmas.
1300 *American Mineralogist* **71**, 239-263.
- 1301 Cerny, P., Blevin, P.L., Cuney, M. & London, D. (2005a). Granite-related ore deposits.
1302 *Economic Geology 100th Anniversary Volume*, 337-370.
- 1303 Cerny, P., Masau, M., Goad, B.E. & Ferreira, K. (2005b). The Greer Lake leucogranite,
1304 Manitoba, and the origin of lepidolite-subtype granitic pegmatites. *Lithos* **80**, 305-321.
- 1305 Charoy, B., Chaussidon, M., Le Carlier de Veslud, C. & Duthou, J.-L. (2003). Evidence of Sr
1306 mobility in and around the albite–lepidolite–topaz granite of Beauvoir (France): an in-situ
1307 ion and electron probe study of secondary Sr-rich phosphates. *Contributions to*
1308 *Mineralogy and Petrology* **145**, 673-690.
- 1309 Christiansen, E.H., Sheridan, M.F. & Burt, D.M. (1986). The geology and geochemistry of
1310 Cenozoic topaz rhyolite from the western United States. *Geological Society of America*
1311 *Special Paper* **205**, 1–82.
- 1312 Costi, H.T., Dall’Agnol, R., Pichavant, M. & Rämö, O.T. (2009). The peralkaline tin-
1313 mineralized Madeira cryolite albite-rich granite of Pitinga, Amazonian craton, Brasil:
1314 petrography, mineralogy and crystallization processes. *Canadian Mineralogist* **47**, 1301-
1315 1327.
- 1316 Cuney, M. & Autran, A. (1987). Objectifs généraux du projet GPF Echassières n°1 et
1317 résultats essentiels acquis par le forage de 900 m sur le granite albitique à topaze-
1318 lépidolite de Beauvoir. *Géologie de la France* **2-3**, 7-24.
- 1319 Cuney, M. & Marignac, C. (2015). Rare metal, W and U deposits and related granites from
1320 the NE French Massif Central. Field Guide for the 13th Biennial SGA Meeting, 155 p.
- 1321 Cuney, M., Marignac, C. & Weisbrod, A. (1992). The Beauvoir topaz-lepidolite albite granite
1322 (Massif Central, France): The disseminated magmatic Sn-Li-Ta-Nb-Be mineralization.
1323 *Economic Geology* **87**, 1766-1794.

- 1324 Cuney, M. & Barbey, P. (2014). Uranium, rare metals, and granulite-facies metamorphism.
1325 *Geoscience Frontiers* **5**, 729-745.
- 1326 Debon, F. & Le Fort, P. (1983). A chemical-mineralogical classification of common plutonic
1327 rocks and associations. *Transactions of the Royal Society of Edinburgh, Earth Sciences*
1328 **73**, 135-149.
- 1329 Eugster, H.P. (1957). Heterogeneous reactions involving oxidation and reduction at high
1330 pressures and temperatures. *Journal of Chemical Physics* **26**, 1760–1761.
- 1331 Faure, M., Lardeaux, J.-M. & Ledru, P. (2009). A review of the pre-Permian geology of the
1332 Variscan French Massif Central. *Comptes Rendus Geosciences* **341**, 202–213.
- 1333 Fouillac, A. M. & Rossi, P. (1991). Near-solidus ¹⁸O depletion in a Ta-Nb-bearing albite
1334 granite: the Beauvoir Granite, France. *Economic Geology* **86**, 1704-1720.
- 1335 French, B.M., Jezek, P.A. & Appleman, D.E. (1978). Virgilite: a new lithium aluminum
1336 silicate mineral from the Macusani glass, Peru. *American Mineralogist* **63**, 461-465.
- 1337 Gardien, V., Lardeaux, J.-M., Ledru, P., Allemand, P. & Guillot, S. (1997). Metamorphism
1338 during late-orogenic extension: insights from the French Variscan belt. *Bulletin de la*
1339 *Société Géologique de France*, **168**, 271-286.
- 1340 Gebelin, A., Roger, F. & Brunel, M. (2009). Syntectonic crustal melting and high-grade
1341 metamorphism in a transpressional regime, Variscan Massif Central, France.
1342 *Tectonophysics* **477**, 229–243
- 1343 Glazner, A.F. & Johnson, B.R. (2013). Late crystallization of K-feldspar and the paradox of
1344 megacrystic granites. *Contributions to Mineralogy and Petrology* **166**, 777-799.
- 1345 Hanson, G.N. (1978). The application of trace elements to the petrogenesis of igneous rocks
1346 of granitic composition. *Earth and Planetary Science Letters* **38**, 26–43.
- 1347 Harlaux, M., Mercadier, J., Bonzi, W.M.-E., Kremer, V., Marignac, C. & Cuney, M. (2017).
1348 Geochemical signature of magmatic-hydrothermal fluids exsolved from the Beauvoir
1349 rare-metal granite (Massif Central, France): Insights from LA-ICPMS analysis of primary
1350 fluid inclusions. *Geofluids* 1925817, doi.org/10.1155/2017/1925817.
- 1351 Harlaux, M., Romer, R.L., Mercadier, J., Morlot, C., Marignac, C. & Cuney, M. (2018). 40
1352 Ma of hydrothermal W mineralization during the Variscan orogenic evolution of the
1353 French Massif Central revealed by U-Pb dating of wolframite. *Mineralium Deposita* **53**,
1354 21-51.
- 1355 Holtz, F., Johannes, W. & Pichavant, M. (1992a). Effect of excess aluminium on phase
1356 relations in the system Qz-Ab-Or: experimental investigation at 2 kbar and reduced H₂O
1357 activity. *European Journal of Mineralogy* **4**, 137-152.

- 1358 Holtz, F., Johannes, W. & Pichavant, M. (1992a). Peraluminous granitoids: the effect of
1359 alumina on melt composition and coexisting minerals. *Transactions of the Royal Society*
1360 *of Edinburgh: Earth Sciences* **83**, 409-416.
- 1361 Holtz, F., Dingwell, D.B. & Behrens, H. (1993). Effects of F, B₂O₃ and P₂O₅ on the
1362 solubility of water in haplogranite melts compared to natural silicate melts. *Contributions*
1363 *to Mineralogy and Petrology* **113**, 492-501.
- 1364 Huang, F., Scaillet, B., Wang, R., Erdmann, S., Chen, Y., Faure, M., Liu, H., Xie, L., Wang,
1365 B. & Zhu, J. (2021). Experimental constraints on intensive crystallization parameters and
1366 fractionation in A-type granites: A case study on the Qitianling pluton, South China.
1367 *Journal of Geophysical Research Solid Earth* **124**, doi.org/10.1029/2019JB017490.
- 1368 Hulsbosch, N. (2019). Nb-Ta-Sn-W distribution in granite-related ore systems: fractionation
1369 mechanisms and examples from the Karagwe-Ankole belt of Central Africa. In *Ore*
1370 *Deposits, Origin, Exploitation, and Exploitation*, AGU Geophysical Monograph Series,
1371 75-107.
- 1372 Icenhower, J. & London, D. (1995). An experimental study of element partitioning among
1373 biotite, muscovite and coexisting peraluminous silicic melt at 200 MPa (H₂O). *American*
1374 *Mineralogist* **80**, 1229–1251.
- 1375 Icenhower, J.P. & London, D. (1997). Partitioning of fluorine and chlorine between biotite
1376 and granitic melt: experimental calibration at 200 MPa H₂O. *Contributions to*
1377 *Mineralogy and Petrology* **127**, 17–29.
- 1378 Johan, V., Monier, G. & Rossi, P. (1988). *Etude pétrographique systématique des granites du*
1379 *sondage GPF1 de Beauvoir*. Doc BRGM 124, 91 p.
- 1380 Kontak, D.J, Clark, A.H. & Farrar E. (1984). The influences of fluid and rock compositions,
1381 and tectono-thermal processes on Al-Si distribution in alkali feldspars in granitoid rocks,
1382 S.E. Peru. *Bulletin de Minéralogie* **107**, 387-400.
- 1383 Kosakevitch, A. (1976). *Evolution de la minéralisation en Li, Ta et Nb dans la coupole*
1384 *granitique de Beauvoir (massif d’Echassières, Allier)*. Document BRGM, 205 p.
- 1385 Kovalenko, V.I. (1979). *Experimental studies on the conditions of formation of rare metal-,*
1386 *Li-, F-rich granites*. Nauka, Moscow, 150 p.
- 1387 Kovalenko, V.I., Tsareva, G.M. & Cuney, M. (1998). Major-, trace elements and water in the
1388 magma of the Beauvoir rare metal granite, France (melt inclusion data). *Doklady*
1389 *Akademii Nauk* **358**, 667–671.

- 1390 Lahlafi, M. (1997). *Rôle des micas dans la concentration des éléments légers (Li, Be et F)*
1391 *dans les granites crustaux: étude expérimentale et cristallogénétique*. Thèse de Doctorat,
1392 Université d'Orléans, 362 p.
- 1393 Lardeaux, J.-M. (2014). Deciphering orogeny: a metamorphic perspective. Examples from
1394 European Alpine and Variscan belts. Part II: Variscan metamorphism in the French
1395 Massif Central - a review. *Bulletin de la Société Géologique de France* **185**, 281–310.
- 1396 La Roche, H. de, Stussi, J.-M. & Chauris, L. (1980). Les granites à deux micas hercyniens
1397 français. Essai de cartographies et de corrélations géochimiques appuyées sur une banque
1398 de données. Implications pétrologiques et métallogéniques. *Sciences de la Terre, Nancy*
1399 **XXIV**, 5-121.
- 1400 Linnen, R.L. (1998). The solubility of Nb-Ta-Zr-Hf-W in granitic melts with Li and Li + F:
1401 constraints for mineralization in rare metal granites and pegmatites. *Economic Geology*
1402 **93**, 1913-1925.
- 1403 Linnen, R.L. & Keppler, H. (1997). Columbite solubility in granitic melts: consequences for
1404 the enrichment and fractionation of Nb and Ta in the Earth's crust. *Contributions to*
1405 *Mineralogy and Petrology* **128**, 213-227.
- 1406 Linnen, R.L. & Cuney, M. (2005). Granite-related rare-element deposits and experimental
1407 constraints on Ta-Nb-W-Sn-Zr-Hf mineralization. In: Linnen RL and Samson IM (eds)
1408 *Rare-element geochemistry and mineral deposits*, Geological Association of Canada, **17**,
1409 45-68.
- 1410 Linnen, R.L., Pichavant, M., Holtz, F. & Burgess, S. (1995). The effect of fO₂ on the
1411 solubility, diffusion and speciation of tin in haplogranitic melt at 850°C and 2 kbar.
1412 *Geochimica et Cosmochimica Acta* **59**, 1579-1588.
- 1413 Linnen, R.L., Pichavant, M. & Holtz, F. (1996). The combined effect of fO₂ and melt
1414 composition on SnO₂ solubility and tin diffusivity in haplogranitic melts. *Geochimica et*
1415 *Cosmochimica Acta*, **60**, 4965-4976.
- 1416 London, D. (1992). Phosphorus in S-type magmas: the P₂O₅ content of feldspars from
1417 peraluminous granites, pegmatites, and rhyolites. *American Mineralogist* **77**, 126-145.
- 1418 London, D. (2008). *Pegmatites*. Canadian Mineralogist Special Publication **10**, 368 p.
- 1419 London, D., Morgan, G.B.VI & Hervig, R.L. (1989). Vapor-undersaturated experiments with
1420 Macusani glass+ H₂O at 200 MPa, and the internal differentiation of granitic pegmatites.
1421 *Contributions to Mineralogy and Petrology* **102**, 1-17.

- 1422 London, D., Morgan, G.B., VI, Babb, H.A. & Loomis, J.L. (1993). Behavior and effects of
1423 phosphorus in the system $\text{Na}_2\text{O}-\text{K}_2\text{O}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{P}_2\text{O}_5-\text{H}_2\text{O}$ at 200 MPa (H_2O).
1424 *Contributions to Mineralogy and Petrology* **113**, 450–465.
- 1425 London, D., Wolf, M.B., Morgan, G.VI & Gallego Garrido, M. (1999). Experimental silicate–
1426 phosphate equilibria in peraluminous granitic magmas, with a case study of the
1427 Alburquerque batholith at Tres Arroyos, Badajoz, Spain. *Journal of Petrology* **40**, 215–
1428 240.
- 1429 London, D., Morgan, G.VI & Wolf, M.B. (2001). Amblygonite-montebrazite solid solutions
1430 as monitors of fluorine in evolved granitic and pegmatitic melts. *American Mineralogist*
1431 **86**, 225–233.
- 1432 London, D., Hunt, L.E., Schwing, C.R. & Guttery, B.M. (2020). Feldspar thermometry in
1433 pegmatites: truth and consequences. *Contributions to Mineralogy and Petrology* **175**, 8,
1434 doi-org.insu.bib.cnrs.fr/10.1007/s00410-019-1617-z.
- 1435 Lukkari, S. & Holtz, F. (2007). Phase relations of a F-enriched peraluminous granite: an
1436 experimental study of the Kymi topaz granite stock, southern Finland. *Contributions to*
1437 *Mineralogy and Petrology* **153**, 273–288.
- 1438 Luth, W.C. (1976). *Granitic rocks*. In: Bailey, D.K., MacDonald, R. (Eds.), *The Evolution of*
1439 *Crystalline Rocks*. Academic Press, London, 335–417.
- 1440 Manning, D.A.C. (1981). The effect of fluorine on liquidus phase relationships in the system
1441 $\text{Qz}-\text{Ab}-\text{Or}$ with excess water at 1 kbar. *Contributions to Mineralogy and Petrology* **76**,
1442 206–215.
- 1443 Marcoux, E., Barré, B., Pichavant, M. & Poujol, M. (2021). Âge et genèse de la coupole
1444 granitique à métaux rares (Sn, Li, Nb-Ta, W) de Montebraz (Creuse, Massif central
1445 français). *BSGF – Earth Science Bulletin* **192**, 16.
- 1446 Marignac, C., Cuney, M., Kesraoui, M. & Bouabssa, L. (2009). Differentiation-
1447 dedifferentiation processes in the rare metal granite (RMG) lineages. Proceedings of the
1448 10th biennial SGA meeting, Townsville, EGRU, 991–993.
- 1449 Martin, J.S. (1983). An experimental study of the effect of lithium on the granite system.
1450 *Proceedings of the Ussher Society* **5**, 417–420.
- 1451 Michaud, J.A.-S. & Pichavant, M. (2020). Magmatic fractionation and the magmatic-
1452 hydrothermal transition in rare metal granites: evidence from Argemela (Central
1453 Portugal). *Geochimica et Cosmochimica Acta* **289**, 130–157.

- 1454 Michaud, J.A.-S., Pichavant, M. & Villaros, A. (2021). Rare elements enrichment in crustal
1455 peraluminous magmas: insights from partial melting experiments. *Contributions to*
1456 *Mineralogy and Petrology* **176**, 96, doi.org/10.1007/s00410-021-01855-9.
- 1457 Moazzen, M. & Droop, G.T.R. (2005). Application of mineral thermometers and barometers
1458 to granitoid igneous rocks: the Etive Complex, W Scotland. *Mineralogy and Petrology*
1459 **83**, 27-53.
- 1460 Monier, G. & Tegye, M. (1985). *Etude pétrographique préliminaire du sondage "GPF*
1461 *Echassières I"*. Doc. BRGM n° 100, 41-93.
- 1462 Monier, G., Charoy, B., Cuney, M., Ohnenstetter, D. & Robert, J.-L. (1987). Evolution
1463 spatiale et temporelle de la composition des micas du granite albitique à topaze-lépidolite
1464 de Beauvoir. *Géologie de la France* **2-3**, 179-188.
- 1465 Monnier, L., Salvi, S., Melleton, J., Bailly, L., Béziat, D., de Parseval, P., Gouy, S. & Lach, P.
1466 (2019). Multiple Generations of wolframite mineralization in the Echassières district
1467 (Massif Central, France). *Minerals* **9**, 637, doi:10.3390/min9100637.
- 1468 Monnier, L., Salvi, S., Jourdan, V., Salla, S., Bailly, L., Melleton, J. & Béziat, D. (2020).
1469 Contrasting fluid behavior during two styles of greisen alteration leading to distinct
1470 wolframite mineralizations: The Echassières district (Massif Central, France). *Ore*
1471 *Geology Reviews* **124**, 103648.
- 1472 Muller, A., Romer, R. & Pedersen, R.-B. (2017). The Sveconorwegian pegmatite province -
1473 thousands of pegmatites without parental granites. *Canadian Mineralogist* **55**, 283–315.
- 1474 Nabelek, P.I., Russ-Nabelek, C. & Denison, J.R. (1992a). The generation and crystallization
1475 conditions of the Proterozoic Harney Peak Leucogranite, Black Hills, South Dakota,
1476 USA: Petrologic and geochemical constraints. *Contributions to Mineralogy and*
1477 *Petrology* **110**, 173-191.
- 1478 Nabelek, P.I., Russ-Nabelek, C. & Haeussler, G.T. (1992b). Stable isotope evidence for the
1479 petrogenesis and fluid evolution in the Proterozoic Harney Peak leucogranite, Black
1480 Hills, South Dakota. *Geochimica et Cosmochimica Acta* **56**, 403-417.
- 1481 Nekvasil, H. & Burnham, C.W. (1987). The calculated individual effects of pressure and
1482 water content on phase equilibria in the granite system. In *Magmatic Processes:*
1483 *Physicochemical Principles*. The Geochemical Society, Special Publication **1**, 433–445.
- 1484 Neukampf, J., Ellis, B., Magna, T., Laurent, O. & Bachmann, O. (2019). Partitioning and
1485 isotopic fractionation of lithium in mineral phases of hot, dry rhyolites: the case of the
1486 Mesa Falls Tuff, Yellowstone. *Chemical Geology* **506**, 175–186.
- 1487 Parsons, I. (1978). Feldspars and fluids in cooling plutons. *Mineralogical Magazine* **42**, 1-17.

- 1488 Pearce, N.J.G., Perkins, W.T., Westgate, J.A., Gorton, M.P., Jackson, S.E., Neal, C.R.,
1489 Chenery, S.P. (1997). A compilation of new and published major and trace element data
1490 for NIST SRM 610 and NIST SRM 612 glass reference materials. *Geostandard*
1491 *Newsletter* **21**, 115–144.
- 1492 Pichavant, M. (1987). Effects of B and H₂O on liquidus phase relations in the haplogranite
1493 system at 1 kbar. *American Mineralogist* **72**, 1056–1070.
- 1494 Pichavant, M., Boher, M., Stenger, J.-F., Aïssa, M. & Charoy, B. (1987a). Relations de phases
1495 des granites de Beauvoir entre 1 et 3 kbar en conditions de saturation en H₂O. *Géologie*
1496 *de la France* **2–3**, 77–86.
- 1497 Pichavant, M., Valencia Herrera, J., Boulmier, S., Briquieu, L., Joron, J.-L., Juteau, M.,
1498 Michard, A., Sheppard, S.M.F., Treuil, M. & Vernet, M. (1987b). The Macusani glasses,
1499 SE Peru: evidence of chemical fractionation in peraluminous magmas. In *Magmatic*
1500 *Processes: Physicochemical Principles*. The Geochemical Society, Special Publication **1**,
1501 359–373.
- 1502 Pichavant, M., Villaros, A., Deveaud, S., Scaillet, B. & Lahlafi, M. (2016). The influence of
1503 redox state on mica crystallization in leucogranitic and pegmatitic liquids. *Canadian*
1504 *Mineralogist* **54**, 559-581.
- 1505 Piwinski, A.J. & Martin, R.F. (1970). An experimental study of equilibrium with granitic
1506 rocks at 10 kbar. *Contributions to Mineralogy and Petrology* **29**, 1-10.
- 1507 Pollard P. J. (2021). The Yichun Ta-Sn-Li deposit, South China: Evidence for extreme
1508 chemical fractionation in F-Li-P-Rich magma. *Economic Geology* **116**, 453-469.
- 1509 Raimbault, L. (1998). Composition of complex lepidolite-type granitic pegmatites and of
1510 constituent columbite-tantalite, Chédeville, Massif Central, France. *Canadian*
1511 *Mineralogist* **36**, 563–583.
- 1512 Raimbault, L. & Burnol, L. (1998). The Richemont rhyolite dyke, Massif Central, France: a
1513 subvolcanic equivalent of rare metal granites. *Canadian Mineralogist* **36**, 265-282.
- 1514 Raimbault, L., Cuney, M., Azencott, C., Duthou, J.-L. & Joron, J.-L. (1995). Geochemical
1515 evidence for a multistage magmatic genesis of Ta-Sn-Li mineralization in the granite at
1516 Beauvoir, French Massif Central. *Economic Geology* **90**, 548-576.
- 1517 Roda-Robles, E., Pesquera, A., Gil-Crespo, P.P., Vieira, R., Lima, A., Garate-Olave, I.,
1518 Martins, T. & Torres-Ruiz, J. (2016). Geology and mineralogy of Li mineralization in the
1519 Central Iberian Zone (Spain and Portugal). *Mineralogical Magazine* **80**, 103-126.
- 1520 Roda-Robles, E., Villaseca, C., Pesquera, A., Gil-Crespo, P.P., Vieira, R., Lima, A. & Garate-
1521 Olave, I. (2018). Petrogenetic relationships between Variscan granitoids and Li-(F-P)-

- 1522 rich aplite-pegmatites in the Central Iberian Zone: geological and geochemical
1523 constraints and implications for other regions from the European Variscides. *Ore*
1524 *Geology Reviews* **95**, 408-430.
- 1525 Romer, R.L. & Pichavant, M. (2021). Rare metal granites and pegmatites. In: Alderton, D.,
1526 Elias, S.A. (eds.), *Encyclopedia of Geology*, 2nd edition., **5**, 840-846.
- 1527 Rossi, P., Autran, A., Azencott, C., Burnol, L., Cuney, M., Johan, V. & Raimbault, L. (1987).
1528 Logs pétrographiques et géochimiques du granite de Beauvoir dans le sondage GPF
1529 “Echassières I”: minéralogie et géochimie comparée. *Géologie de la France* **2-3**, 111–
1530 135.
- 1531 Scaillet, B., Pichavant, M. & Roux, J. (1995). Experimental crystallization of leucogranite
1532 magmas. *Journal of Petrology* **36**, 663–705.
- 1533 Schmidt, B., Scaillet, B. & Holtz, F. (1995). Accurate control of fH(2) in cold-seal pressure
1534 vessels with the Shaw membrane technique. *European Journal of Mineralogy* **7**, 893–
1535 903.
- 1536 Simpson, D.R. (1977). Aluminum phosphate variants of feldspar. *American Mineralogist* **62**,
1537 351-355.
- 1538 Sirbescu, M.-L., Schmidt, C., Veksler, I.V., Whittington, A.G. & Wilke, M. (2017).
1539 Experimental crystallization of undercooled felsic liquids: generation of pegmatitic
1540 texture. *Journal of Petrology* **58**, 539-568.
- 1541 Smith, J.V. & Brown, W.L. (1988). *Feldspar minerals. Volume 1 Crystal Structures,*
1542 *Physical, Chemical, and Microtextural Properties*. Springer, 828 p.
- 1543 Stewart, D.B. (1978). Petrogenesis of lithium-rich pegmatites. *American Mineralogist* **63**,
1544 970-980.
- 1545 Soufi, M. (2021). Origin and physical-chemical control of topaz crystallization in felsic
1546 igneous rocks: Contrasted effect of temperature on its OH–F substitution. *Earth Science*
1547 *Reviews* **213**, 103467.
- 1548 Taylor, H.P.Jr. & Forester, R.W. (1971). Low-O18 igneous rocks from the intrusive
1549 complexes of Skye, Mull, and Ardnamurchan, Western Scotland. *Journal of Petrology*
1550 **12**, 465-497.
- 1551 Tischendorf, G., Gottesmann, B., Förster, H.-J. & Trumbull, R.B. (1997). On Li-bearing
1552 micas: estimating Li from electron microprobe analyses and an improved diagram for
1553 graphical representation. *Mineralogical Magazine* **61**, 809–834.

- 1554 Tuttle, O.F., Bowen, N.L. (1958). Origin of granite in the light of experimental studies in the
 1555 system NaAlSi₃O₈-KAlSi₃O₈-SiO₂-H₂O. *Geological Society of America Memoir* 74,
 1556 153 p.
- 1557 Xiong, X.L., Zhao, Z.H., Zhu, J.C., Rao, B. & Lai, M.Y. (1998). Phase equilibria in the
 1558 granite-H₂O-HF system and effect of fluorine on granitic melt structure. *Chinese Journal*
 1559 *of Geochemistry* **17**, 114-122.
- 1560 Xiong, X.L., Zhao, Z.H., Zhu, J.C. & Rao, B. (1999). Phase relations in albite granite-H₂O-
 1561 HF system and their petrogenetic applications. *Geochemical Journal* **33**, 199-214.
- 1562 Xiong, X.L., Rao, B., Chena, F.R., Zhu, J.C. & Zhao, Z.H. (2002). Crystallization and melting
 1563 experiments of a fluorine-rich leucogranite from the Xianghualing Pluton, South China,
 1564 at 150 MPa and H₂O-saturated conditions. *Journal of Asian Earth Sciences* **21**, 175-188.
- 1565 Van Lichtervelde, M., Holtz, F. & Melcher, F. (2018). The effect of disequilibrium
 1566 crystallization on Nb-Ta fractionation in pegmatites: Constraints from crystallization
 1567 experiments of tantalite-tapiolite. *American Mineralogist* **103**, 1401-1416.
- 1568 Villaros, A., Pichavant, M. & Deveaud, S. (2019). Mica-liquid trace elements partitioning and
 1569 the granite-pegmatite connection: The St-Sylvestre complex (Western French Massif
 1570 Central). *Chemical Geology* **528**, 119265.
- 1571 Weidner, J.R. & Martin, R.F. (1987). Phase equilibria of a fluorine-rich leucogranite from the
 1572 St. Austell pluton, Cornwall. *Geochimica et Cosmochimica Acta* **51**, 1591-1597.
- 1573 Wen, S. & Nekvasil, H. (1994). SOLVALC: an interactive graphics program package for
 1574 calculating the ternary feldspar solvus and for two feldspar geothermometry. *Computers*
 1575 *and Geosciences* **20**, 1025e1040.
- 1576 Zhu, J.C., Li, R.K., Li, F.C., Xiong, X.L., Zhou, F.Y. & Huang, X.L. (2001). Topaz albite
 1577 granites and rare-metal mineralization in the Limu district, Guangxi provine, southeast
 1578 China. *Mineralium Deposita* **36**, 393-405.

1579

1580 **FIGURE CAPTIONS**

1581 **Fig. 1.** SEM images of major phases and textures in crystallization experiments. (a) euhedral
 1582 quartz (Qz) and plagioclase (Pl) in glass (Gl), charge B2N-18. (b) large homogenous
 1583 plagioclase crystals (Pl) in glass (Gl), charge B1N-1. (c) large heterogeneous feldspar crystals
 1584 in glass (Gl), charge B2N-3. The potassic and sodic zones are denoted by K and Na
 1585 respectively. Notice the progressive and geometrically complex transitions between the two
 1586 zones and the small quartz (Qz) crystals at the extremities of the feldspars. (d) euhedral
 1587 homogeneous K-feldspar crystals (Af) coexisting with homogeneous plagioclase (Pl) in

1588 composite crystals embedded in glass (Gl), charge B1N-22. (e) Li-mica crystals (Li-Mi) in
1589 glass (Gl), charge B2N-25. Apatite (Ap) appears in white at the bottom of the image. (f)
1590 Euhedral topaz crystals (Tpz) coexisting with quartz (Qz), K-feldspar (Af), plagioclase (Pl),
1591 Li-mica (small white grey interstitial phases and aggregates) and glass (Gl) in charge B1-21.
1592 (g) Several crystals of Li-Al phosphates (LiP) of the amblygonite-montebrasite series
1593 dispersed in charge B1-21. Notice also the presence of topaz (Tpz). (h) Crystal of hematite
1594 (Hm) in glass (Gl), charge B2N-19. See Tables 2 and 3 for experimental conditions.

1595

1596 **Fig. 2.** Summary P-T phase diagram for the B1 Beauvoir composition under H₂O-saturated
1597 conditions and for a fO_2 between NNO+3 and NNO+1. The solid symbols are experimental
1598 data points from Table 2. Solid curves represent phase equilibria constrained from
1599 crystallization experiments (this study) and dashed curves from melting experiments
1600 (Pichavant et al., 1987a). Solidus curves appear in black and Qz, Pl, Af and Li-Mi saturation
1601 curves in red, blue, green and yellow respectively. Phases are stable on the left of their
1602 respective saturation curve. The P-T location of minimum liquidus temperature points in the
1603 H₂O-saturated haplogranite system (“haplogranite minimum”, Tuttle and Bowen, 1958) is
1604 shown for comparison (grey solid curve). See text for the construction of the diagram and for
1605 interpretations.

1606

1607 **Fig. 3.** Summary P-T phase diagram for the B2 Beauvoir composition under H₂O-saturated
1608 conditions and for a fO_2 between NNO+3 and NNO-1. The solid symbols are experimental
1609 data points from Table 3. Solid curves represent phase equilibria constrained from
1610 crystallization experiments (this study) and dashed curves from melting experiments
1611 (Pichavant et al., 1987a). Solidus curves appear in black and Qz, Pl, Af and Li-Mi saturation
1612 curves in red, blue, green and yellow respectively. Phases are stable on the left of their
1613 respective saturation curve. The P-T location of minimum liquidus temperature points in the
1614 H₂O-saturated haplogranite system (“haplogranite minimum”, Tuttle and Bowen, 1958) is
1615 shown for comparison (grey solid curve). See text for the construction of the diagram and for
1616 interpretations.

1617

1618 **Fig. 4.** Changes in Beauvoir experimental melt compositions during progressive
1619 crystallization and decreasing melt fraction (expressed in wt%). Symbols are in red for B1
1620 and in blue for B2. Melt fractions are melt (L) proportions in experimental charges (solid
1621 symbols) calculated by mass balance (Table 2 and 3). For starting melts (open symbols), melt

1622 fractions are 100%. (a) F concentrations (wt%) vs. melt fraction; (b) P₂O₅ concentrations
1623 (wt%) vs. melt fraction; (c) normative Corundum (%) vs. melt fraction; (d) normative
1624 Anorthite (%) vs. melt fraction. In (a), numbers in grey refer to the charge numbering in
1625 Tables 2 and 3. Data for experimental melts are from Table 4 and for starting melts from
1626 Table 1. Note that the F and P₂O₅ concentrations are not normalized to 100% anhydrous.
1627 Error bars are shown.

1628
1629 **Fig. 5.** Qz, Ab, Or compositions of Beauvoir starting materials (stars) and experimental
1630 glasses (open circles). Symbols are in red for the B1 and in blue for the B2 composition.
1631 Normative compositions (data in Tables 1 and 4) are projected on the Qz-Ab-Or space.
1632 Numbers in grey refer to the charge numbering in Tables 2 and 3. Note that, in panel B1,
1633 compositions of glass B1O-1 and of starting materials overlap and, in panel B2, compositions
1634 of glasses B2N-19, B2N-20, B2N-25, B2N-26 and of starting materials overlap. For
1635 reference, the positions of H₂O-saturated minimum liquidus temperature points in the
1636 haplogranite system at 100, 200 and 300 MPa (Tuttle and Bowen, 1958) are shown as black
1637 squares. The positions of H₂O-saturated minimum liquidus temperature points in the
1638 haplogranite system at 100 MPa with 1 and 2 wt% added F (Manning, 1981) are shown as
1639 black circles.

1640
1641 **Fig. 6.** Compositions of feldspars in the crystallization experiments. All data from Table S2
1642 are plotted. The B1 symbols and tie-lines are in red and the B2 in blue. Pl: plagioclase; Af: K-
1643 feldspar; I: heterogeneous feldspar. Notice the heterogeneous feldspars plotting intermediate
1644 between Pl and Af. Two-feldspar equilibration temperatures calculated using the
1645 SOLVCALC software of Wen and Nekvasil (1994) for 4 representative Pl and Af pairs are
1646 shown together with experimental pressures on their respective tie-lines. For charge B2N-24,
1647 the calculated temperature is 578°C (vs. 580°C, Table 3) and, for charge B1N-21, 567°C (vs.
1648 580°C, Table 2). For charge B2N-20, two plagioclase compositions have been selected for the
1649 calculations, the < 2 mol% An difference between the two being representative of the
1650 chemical dispersion within charges. The two Pl-Af pairs yield temperatures of 608 and 588°C
1651 (vs. 620°C, Table 3). Results of other two-feldspar calculations are summarized in text.
1652 Average compositions of feldspars in B1 and B2 starting rocks (data from Pichavant et al.,
1653 1987a) are shown for comparison.

1654

1655 **Fig. 7.** Compositions of (a) experimental and (b) natural micas plotted in the $\text{Fe}_t + \text{Mn} + \text{Ti} -$
 1656 Al^{VI} (apfu) vs. $\text{Mg} - \text{Li}$ (apfu) diagram of Tischendorf et al. (1997). Abbreviated names of
 1657 end-member compositions appear in grey (nomenclature after Tischendorf et al., 1997). Data
 1658 for experimental micas include electron microprobe (EMPA) and SEM analyses (Table 7 and
 1659 S3). For the SEM analyses, F concentrations are corrected by a constant factor ($\times 2.3$). The
 1660 data for the natural Beauvoir micas (Monier et al., 1987) include electron microprobe
 1661 (EMPA) and wet chemical analyses (CHEM). The data for the natural Argemela micas are
 1662 averages of EMPA analyses from Michaud and Pichavant (2020). Except for the wet chemical
 1663 analyses of Beauvoir micas, the Li_2O contents of micas analyzed by EMPA and SEM are all
 1664 calculated from their F concentrations, using the [di 1] equation of Tischendorf et al. (1997).
 1665 In (a), 21, 22, 23 refer to the charge number in Table 2. The two arrows represent the
 1666 evolutionary trends (1) and (2) discussed in text.

1667
 1668 **Fig. 8.** (a) $\text{Mg} - \text{Li}$ (apfu) and (b) $\text{Fe}_t + \text{Mn} + \text{Ti} - \text{Al}^{\text{VI}}$ (apfu) of experimental micas as a
 1669 function of the F content of the coexisting melt. Data for experimental micas include electron
 1670 microprobe (EMPA) and SEM analyses (Table 7 and S3). For the SEM analyses, F
 1671 concentrations are corrected by a constant factor ($\times 2.3$). Li_2O concentrations of micas
 1672 analyzed by EMPA and SEM are all calculated from their F contents, using the [di 1]
 1673 equation of Tischendorf et al. (1997). F concentrations in experimental melts are raw data
 1674 from Table 4 (i.e., not normalized to 100%) with error bar shown. 21, 22, 23, 25 refer to the
 1675 charge number in Table 2 (i.e., 21-23 refers to the B1N-21 to -23 and 25 to the B2N-25
 1676 charge). The two arrows represent the evolutionary trends (1) and (2) discussed in text.

1677
 1678 **Fig. 9.** Evolution of trace element concentrations during progressive crystallization of the B2
 1679 Beauvoir melt. Concentrations of trace elements (Li, Be, B, Mn, Rb, Cs, W, U, Nb, Ta, Sn,
 1680 Sr) in selected B2 glasses (C), referenced to those in the initial melt (C_0 , represented by the
 1681 crystal-free charge B2N-26), are plotted as $\log(C/C_0)$ vs. the melt fraction (f). Glass
 1682 concentrations are from Table 9 and melt fractions from Table 3. Dashed curves are Rayleigh
 1683 fractionation trends calculated (see Hanson, 1978) for bulk mineral/melt partition coefficients
 1684 (D) of 0.01, 0.5 and 3. The horizontal dotted line corresponds to $C = C_0$. See text.

1685
 1686 **Fig. 10.** F concentration in experimental micas as a function of the F content of the coexisting
 1687 melt. Data for experimental micas include electron microprobe (EMPA) and SEM analyses
 1688 (Table 7 and S3). F concentrations from SEM data are corrected by a constant factor ($\times 2.3$). F

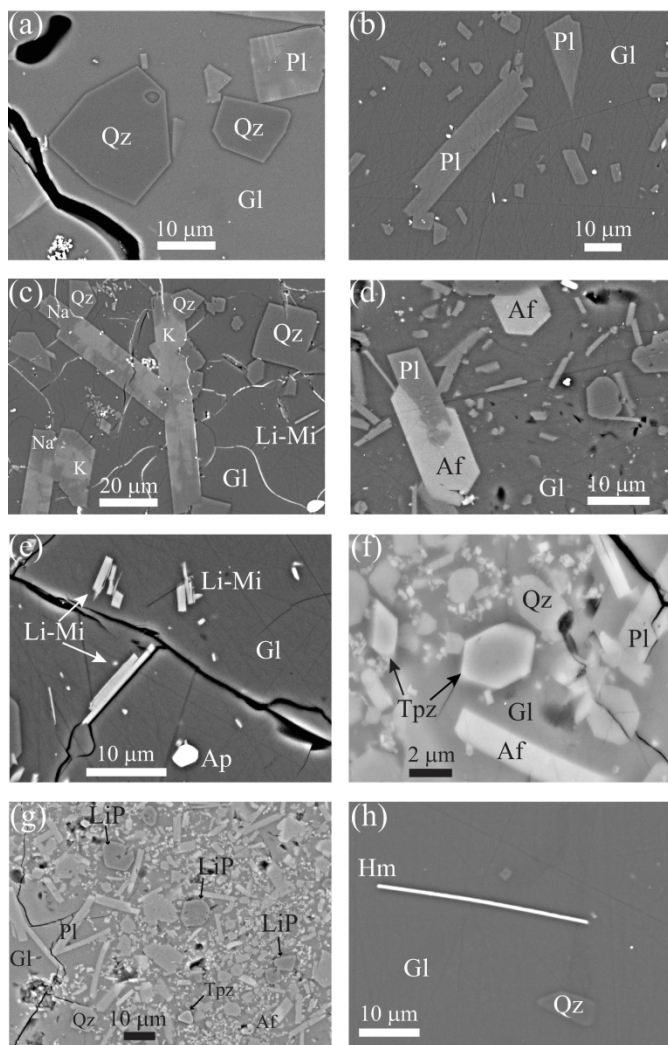
1689 concentrations in experimental melts are raw data from Table 4 (i.e., not normalized to 100%)
 1690 with error bar shown. 21, 22, 23, 25 refer to the charge number in Table 2 (i.e., 21-23 refers to
 1691 the B1N-21 to -23 and 25 to the B2N-25 charge).

1692

1693 **Fig. 11.** Comparison between major element fractionation in Beauvoir granites (left column)
 1694 and in experimental glasses from this study (right column). Top panels (a) and (b): major
 1695 element compositions plotted in the A-B diagram of Debon and Le Fort (1983). Bottom
 1696 panels (c) and (d): major element compositions plotted in the Q*3-B*3-F*3-M*3
 1697 representation of La Roche et al. (1980), see also Cuney and Marignac (2015) and inset for
 1698 the significance of the Q*3, B*3, F*3, M*3 apexes. Concentrations are expressed in
 1699 millications, i.e., as 1000 x atomic concentrations (at.). Abbreviations: Qz: quartz, Ann:
 1700 annite, Phl: phlogopite, Pl: plagioclase, Af: K-feldspar, Mu: muscovite. In (d), experimental
 1701 mica compositions from charges B1N23 and B2N-25 are also plotted (respectively Li-Mi B1
 1702 and Li-Mi B2). The Beauvoir whole rock data are exclusively from Raimbault et al. (1995)
 1703 and the experimental data are from this paper (averages of starting compositions in Table 1
 1704 and selected glass compositions from Table 4). Notice the specific Q*3-rich compositions of
 1705 glasses B1N-17 and B2N-3 (respectively labelled 17 and 3) from 100 MPa crystallization
 1706 charges.

1707

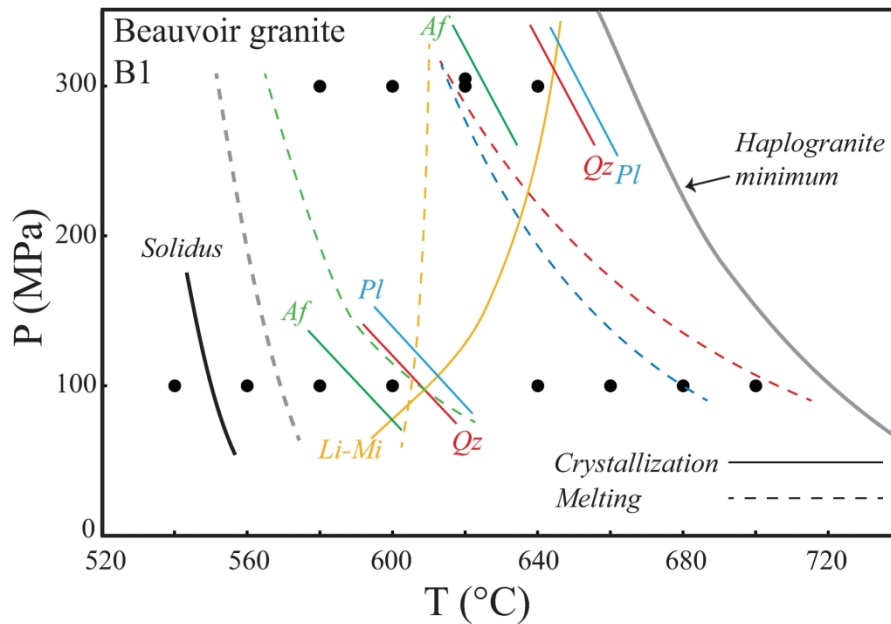
1708 **Fig. 12.** Comparison between trace element (Sn, Ta, Be, Nb, Li, P, U, F, Rb, W, Ti, Mn, Zn,
 1709 Pb, Sr, Cs) fractionation in experiments from this study and in Beauvoir granites. Horizontal
 1710 axis: $\log(C/C_0)$ where C is the trace element concentration in the most fractionated glass B2-
 1711 24 and C_0 the trace element concentration in the parental glass B2-26 (Table 9). Vertical axis:
 1712 $\log(C/C_0)$ where C is the trace element concentration in B1 granite (ESP 270, Raimbault et
 1713 al., 1995) and C_0 the trace element concentration in glass B2-26 (Table 9) assumed
 1714 representative of the B2 granite. Experimental and natural enrichment and depletion
 1715 directions are indicated. The location of data points relative to the 1:1 dotted line indicates
 1716 how close elements are from an evolution identical in experiments and granites, whether an
 1717 enrichment or a depletion. Color coding reflects the four groups of elements distinguished,
 1718 those enriched both in the experiments and in the granites (red circles), those depleted both in
 1719 the experiments and in the granites (black circles), those whose behavior should be controlled
 1720 by accessory phase solubilities (blue squares) and those very much fractionated neither in the
 1721 experiments nor in the granite (grey diamonds).



Pichavant, Fig.1

Fig. 1. SEM images of major phases and textures in crystallization experiments. (a) euhedral quartz (Qz) and plagioclase (Pl) in glass (Gl), charge B2N-18. (b) large homogeneous plagioclase crystals (Pl) in glass (Gl), charge B1N-1. (c) large heterogeneous feldspar crystals in glass (Gl), charge B2N-3. The potassic and sodic zones are denoted by K and Na respectively. Notice the progressive and geometrically complex transitions between the two zones and the small quartz (Qz) crystals at the extremities of the feldspars. (d) euhedral homogeneous K-feldspar crystals (Af) coexisting with homogeneous plagioclase (Pl) in composite crystals embedded in glass (Gl), charge B1N-22. (e) Li-mica crystals (Li-Mi) in glass (Gl), charge B2N-25. Apatite (Ap) appears in white at the bottom of the image. (f) Euhedral topaz crystals (Tpz) coexisting with quartz (Qz), K-feldspar (Af), plagioclase (Pl), Li-mica (small white grey interstitial phases and aggregates) and glass (Gl) in charge B1-21. (g) Several crystals of Li-Al phosphates (LiP) of the amblygonite-montebbrasite series dispersed in charge B1-21. Notice also the presence of topaz (Tpz). (h) Crystal of hematite (Hm) in glass (Gl), charge B2N-19. See Tables 2 and 3 for experimental conditions.

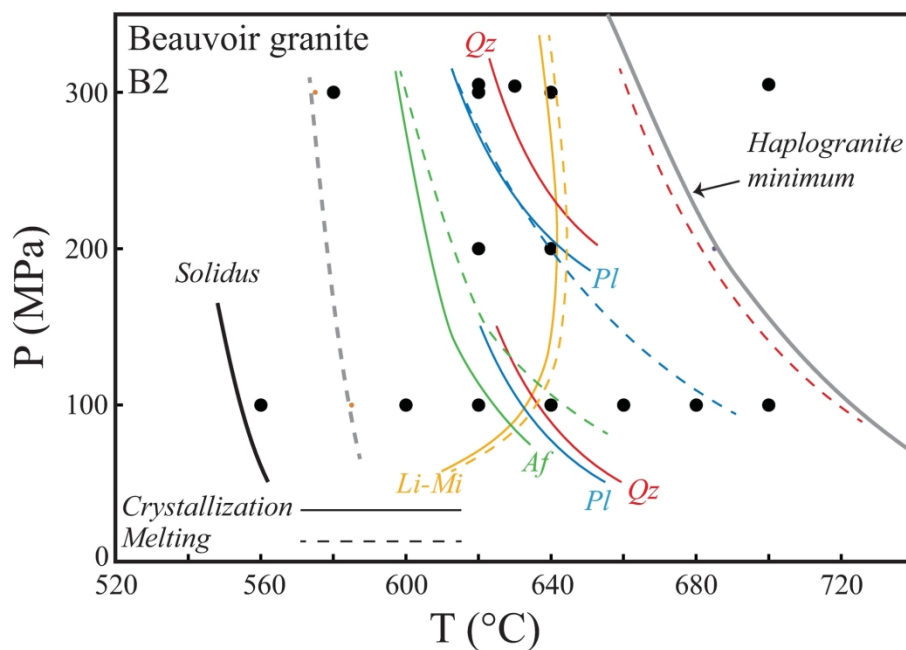
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Pichavant, Fig.2

Fig. 2. Summary P-T phase diagram for the B1 Beauvoir composition under H₂O-saturated conditions and for a fO₂ between NNO+3 and NNO+1. The solid symbols are experimental data points from Table 2. Solid curves represent phase equilibria constrained from crystallization experiments (this study) and dashed curves from melting experiments (Pichavant et al., 1987a). Solidus curves appear in black and Qz, Pl, Af and Li-Mi saturation curves in red, blue, green and yellow respectively. Phases are stable on the left of their respective saturation curve. The P-T location of minimum liquidus temperature points in the H₂O-saturated haplogranite system ("haplogranite minimum", Tuttle and Bowen, 1958) is shown for comparison (grey solid curve). See text for the construction of the diagram and for interpretations.

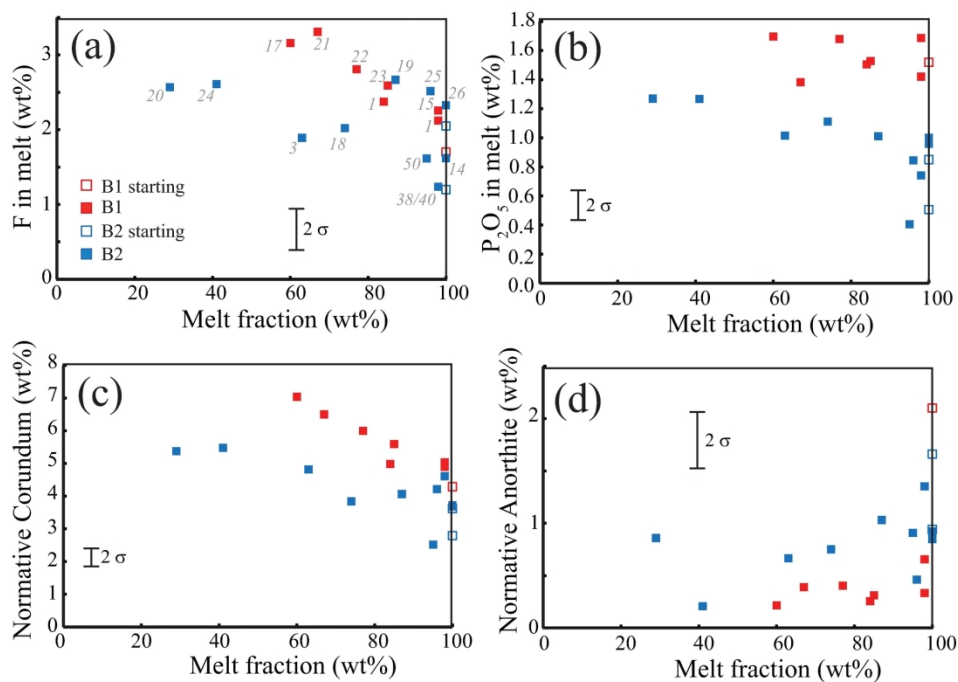
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Pichavant, Fig.3

Fig. 3. Summary P-T phase diagram for the B2 Beauvoir composition under H₂O-saturated conditions and for a fO₂ between NNO+3 and NNO-1. The solid symbols are experimental data points from Table 3. Solid curves represent phase equilibria constrained from crystallization experiments (this study) and dashed curves from melting experiments (Pichavant et al., 1987a). Solidus curves appear in black and Qz, Pl, Af and Li-Mi saturation curves in red, blue, green and yellow respectively. Phases are stable on the left of their respective saturation curve. The P-T location of minimum liquidus temperature points in the H₂O-saturated haplogranite system ("haplogranite minimum", Tuttle and Bowen, 1958) is shown for comparison (grey solid curve). See text for the construction of the diagram and for interpretations.

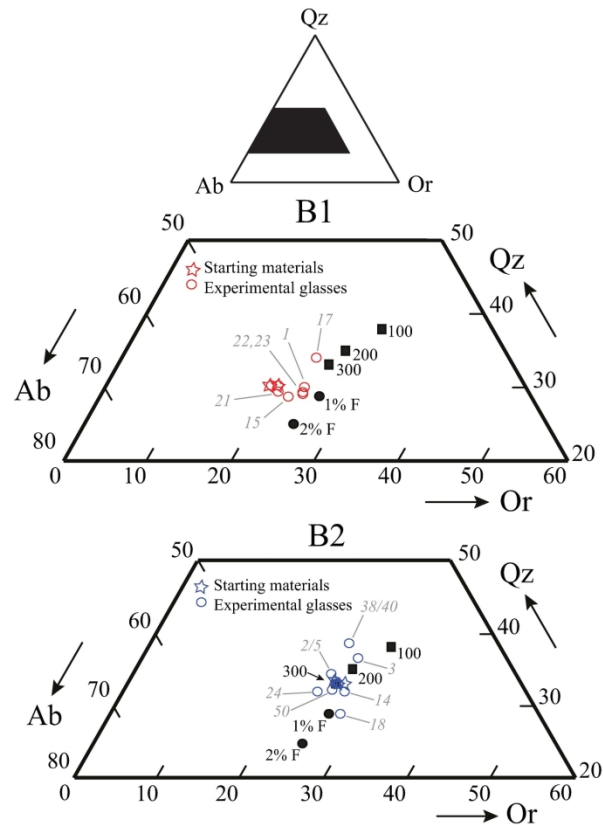
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Pichavant, Fig.4

Fig. 4. Changes in Beauvoir experimental melt compositions during progressive crystallization and decreasing melt fraction (expressed in wt%). Symbols are in red for B1 and in blue for B2. Melt fractions are melt (L) proportions in experimental charges (solid symbols) calculated by mass balance (Table 2 and 3). For starting melts (open symbols), melt fractions are 100%. (a) F concentrations (wt%) vs. melt fraction; (b) P₂O₅ concentrations (wt%) vs. melt fraction; (c) normative Corundum (%) vs. melt fraction; (d) normative Anorthite (%) vs. melt fraction. In (a), numbers in grey refer to the charge numbering in Tables 2 and 3. Data for experimental melts are from Table 4 and for starting melts from Table 1. Note that the F and P₂O₅ concentrations are not normalized to 100% anhydrous. Error bars are shown.

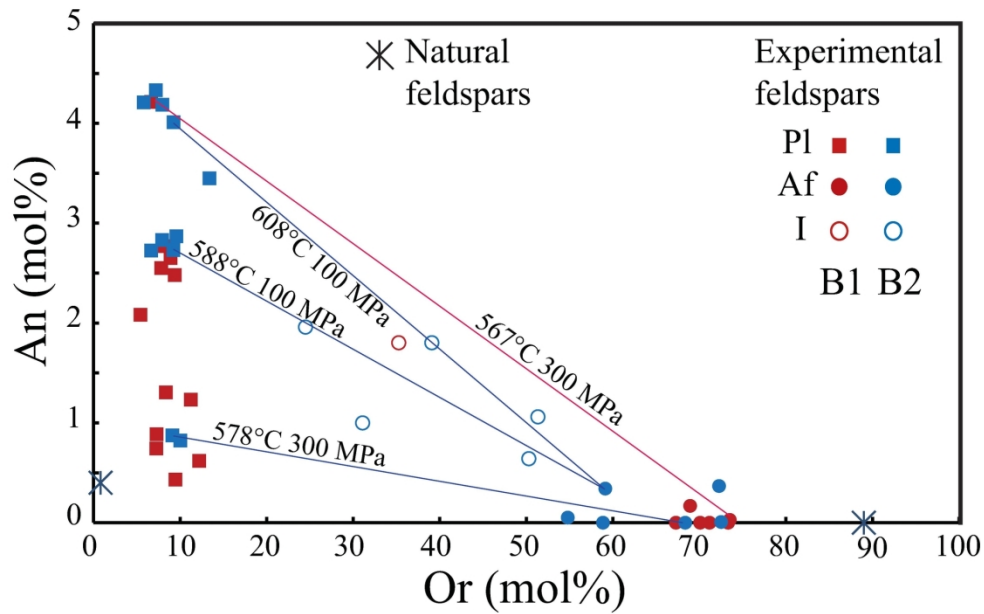
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Pichavant, Fig. 5

Fig. 5. Qz, Ab, Or compositions of Beauvoir starting materials (stars) and experimental glasses (open circles). Symbols are in red for the B1 and in blue for the B2 composition. Normative compositions (data in Tables 1 and 4) are projected on the Qz-Ab-Or space. Numbers in grey refer to the charge numbering in Tables 2 and 3. Note that, in panel B1, compositions of glass B1O-1 and of starting materials overlap and, in panel B2, compositions of glasses B2N-19, B2N-20, B2N-25, B2N-26 and of starting materials overlap. For reference, the positions of H₂O-saturated minimum liquidus temperature points in the haplogranite system at 100, 200 and 300 MPa (Tuttle and Bowen, 1958) are shown as black squares. The positions of H₂O-saturated minimum liquidus temperature points in the haplogranite system at 100 MPa with 1 and 2 wt% added F (Manning, 1981) are shown as black circles.

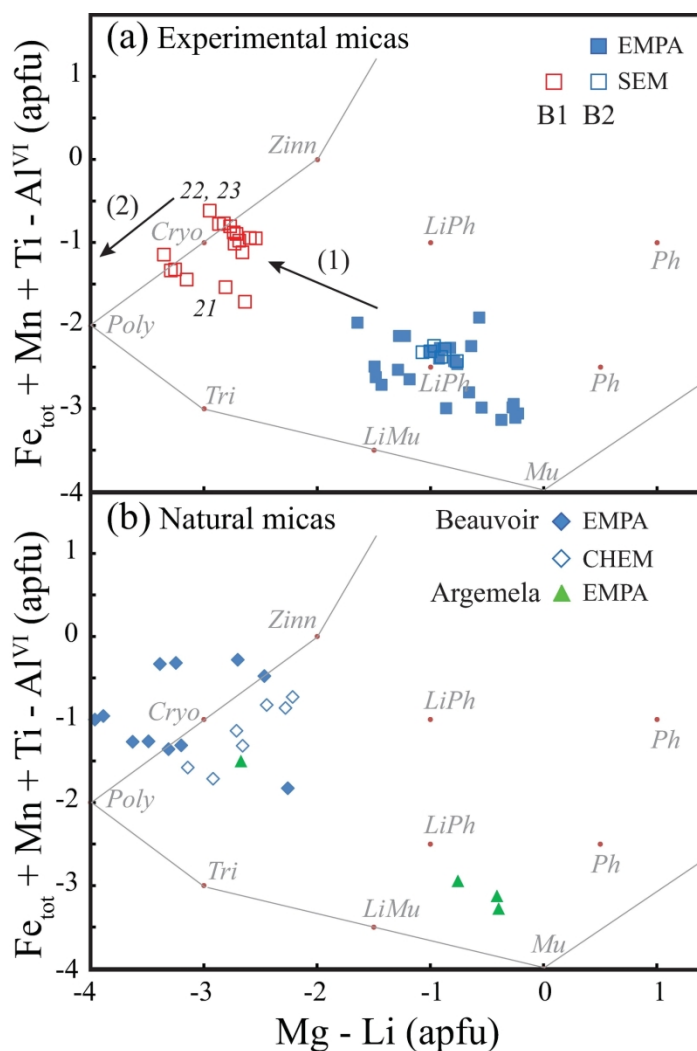
134x250mm (300 x 300 DPI)



Pichavant, Fig. 6

Fig. 6. Compositions of feldspars in the crystallization experiments. All data from Table S2 are plotted. The B1 symbols and tie-lines are in red and the B2 in blue. Pl: plagioclase; Af: K-feldspar; I: heterogeneous feldspar. Notice the heterogeneous feldspars plotting intermediate between Pl and Af. Two-feldspar equilibration temperatures calculated using the SOLVCALC software of Wen and Nekvasil (1994) for 4 representative Pl and Af pairs are shown together with experimental pressures on their respective tie-lines. For charge B2N-24, the calculated temperature is 578°C (vs. 580°C, Table 3) and, for charge B1N-21, 567°C (vs. 580°C, Table 2). For charge B2N-20, two plagioclase compositions have been selected for the calculations, the < 2 mol% An difference between the two being representative of the chemical dispersion within charges. The two Pl-Af pairs yield temperatures of 608 and 588°C (vs. 620°C, Table 3). Results of other two-feldspar calculations are summarized in text. Average compositions of feldspars in B1 and B2 starting rocks (data from Pichavant et al., 1987a) are shown for comparison.

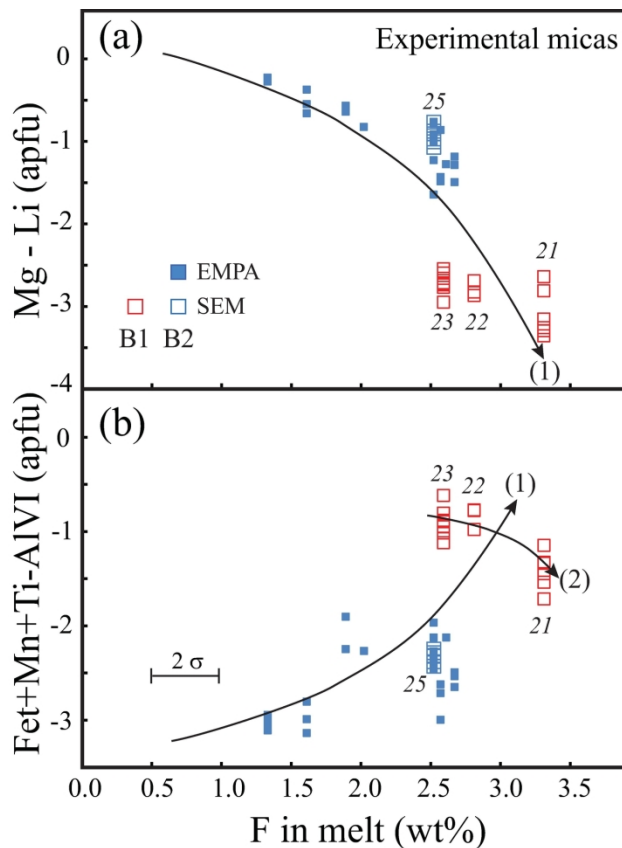
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Pichavant, Fig.7

Fig. 7. Compositions of (a) experimental and (b) natural micas plotted in the $Fe_{\text{tot}} + Mn + Ti - Al^{VI}$ (apfu) vs. $Mg - Li$ (apfu) diagram of Tischendorf et al. (1997). Abbreviated names of end-member compositions appear in grey (nomenclature after Tischendorf et al., 1997). Data for experimental micas include electron microprobe (EMPA) and SEM analyses (Table 7 and S3). For the SEM analyses, F concentrations are corrected by a constant factor ($\times 2.3$). The data for the natural Beauvoir micas (Monier et al., 1987) include electron microprobe (EMPA) and wet chemical analyses (CHEM). The data for the natural Argemela micas are averages of EMPA analyses from Michaud and Pichavant (2020). Except for the wet chemical analyses of Beauvoir micas, the Li₂O contents of micas analyzed by EMPA and SEM are all calculated from their F concentrations, using the [di 1] equation of Tischendorf et al. (1997). In (a), 21, 22, 23 refer to the charge number in Table 2. The two arrows represent the evolutionary trends (1) and (2) discussed in text.

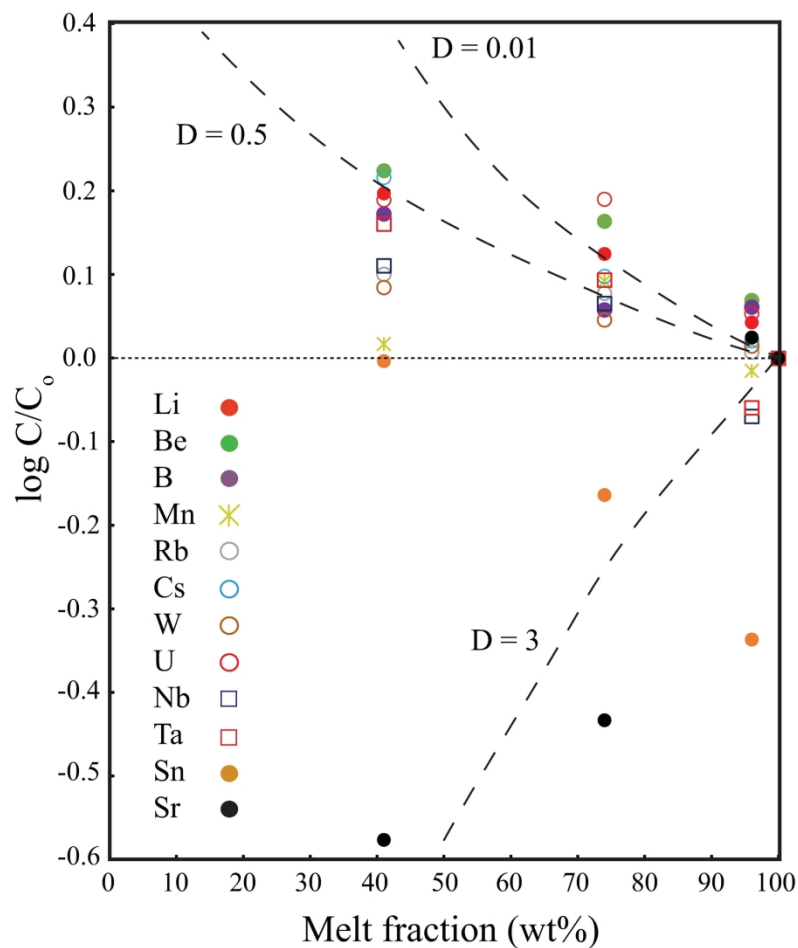
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Pichavant, Fig.8

Fig. 8. (a) Mg - Li (apfu) and (b) Fet + Mn + Ti - AlVI (apfu) of experimental micas as a function of the F content of the coexisting melt. Data for experimental micas include electron microprobe (EMPA) and SEM analyses (Table 7 and S3). For the SEM analyses, F concentrations are corrected by a constant factor ($\times 2.3$). Li₂O concentrations of micas analyzed by EMPA and SEM are all calculated from their F contents, using the [di 1] equation of Tischendorf et al. (1997). F concentrations in experimental melts are raw data from Table 4 (i.e., not normalized to 100%) with error bar shown. 21, 22, 23, 25 refer to the charge number in Table 2 (i.e., 21-23 refers to the B1N-21 to -23 and 25 to the B2N-25 charge). The two arrows represent the evolutionary trends (1) and (2) discussed in text.

154x249mm (300 x 300 DPI)

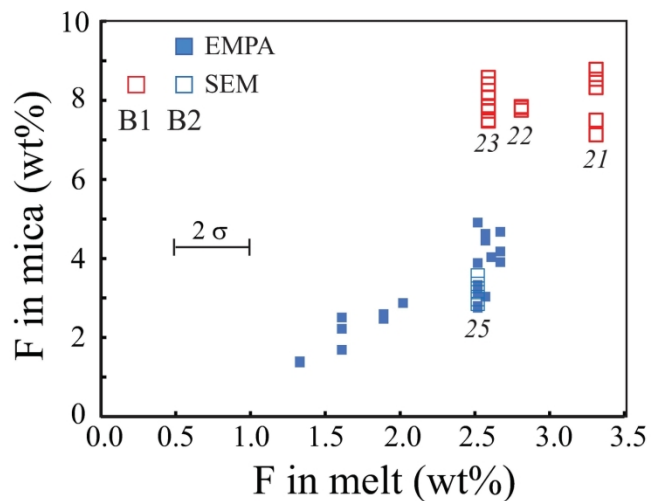


Pichavant, Fig.9

Fig. 9. Evolution of trace element concentrations during progressive crystallization of the B2 Beauvoir melt. Concentrations of trace elements (Li, Be, B, Mn, Rb, Cs, W, U, Nb, Ta, Sn, Sr) in selected B2 glasses (C), referenced to those in the initial melt (C_0 , represented by the crystal-free charge B2N-26), are plotted as $\log (C/C_0)$ vs. the melt fraction (f). Glass concentrations are from Table 9 and melt fractions from Table 3.

Dashed curves are Rayleigh fractionation trends calculated (see Hanson, 1978) for bulk mineral/melt partition coefficients (D) of 0.01, 0.5 and 3. The horizontal dotted line corresponds to $C = C_0$. See text.

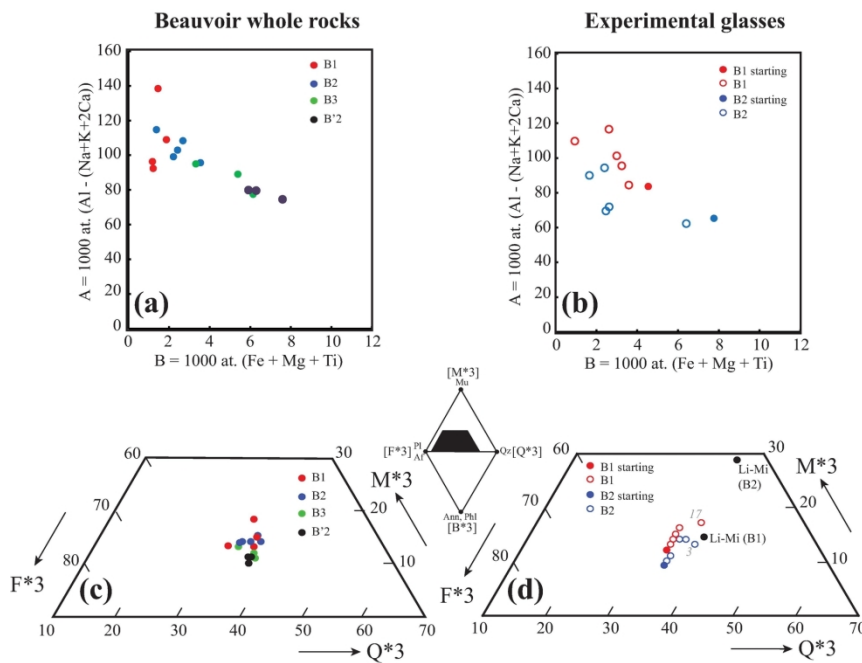
175x249mm (300 x 300 DPI)



Pichavant, Fig.10

Fig. 10. F concentration in experimental micas as a function of the F content of the coexisting melt. Data for experimental micas include electron microprobe (EMPA) and SEM analyses (Table 7 and S3). F concentrations from SEM data are corrected by a constant factor ($\times 2.3$). F concentrations in experimental melts are raw data from Table 4 (i.e., not normalized to 100%) with error bar shown. 21, 22, 23, 25 refer to the charge number in Table 2 (i.e., 21-23 refers to the B1N-21 to -23 and 25 to the B2N-25 charge).

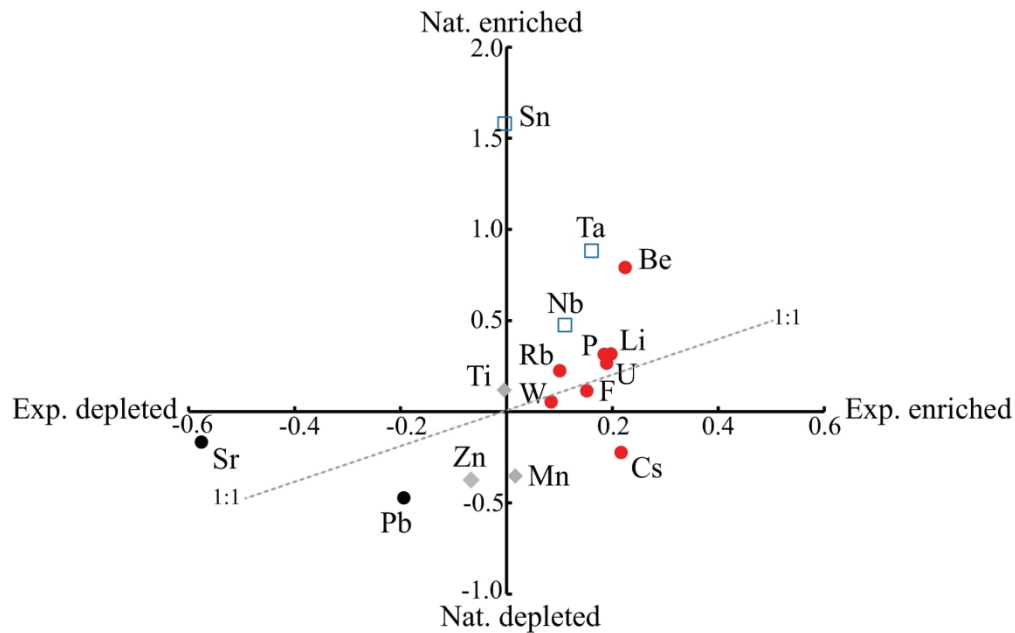
149x224mm (300 x 300 DPI)



Pichavant, Fig. 11

Fig. 11. Comparison between major element fractionation in Beauvoir granites (left column) and in experimental glasses from this study (right column). Top panels (a) and (b): major element compositions plotted in the A-B diagram of Debon and Le Fort (1983). Bottom panels (c) and (d): major element compositions plotted in the Q*3-B*3-F*3-M*3 representation of La Roche et al. (1980), see also Cuney and Maignac (2015) and inset for the significance of the Q*3, B*3, F*3, M*3 apices. Concentrations are expressed in millications, i.e., as 1000 x atomic concentrations (at.). Abbreviations: Qz: quartz, Ann: annite, Phl: phlogopite, Pl: plagioclase, Af: K-feldspar, Mu: muscovite. In (d), experimental mica compositions from charges B1N23 and B2N-25 are also plotted (respectively Li-Mi B1 and Li-Mi B2). The Beauvoir whole rock data are exclusively from Raimbault et al. (1995) and the experimental data are from this paper (averages of starting compositions in Table 1 and selected glass compositions from Table 4). Notice the specific Q*3-rich compositions of glasses B1N-17 and B2N-3 (respectively labelled 17 and 3) from 100 MPa crystallization charges.

171x221mm (300 x 300 DPI)



Pichavant, Fig.12

Fig. 12. Comparison between trace element (Sn, Ta, Be, Nb, Li, P, U, F, Rb, W, Ti, Mn, Zn, Pb, Sr, Cs) fractionation in experiments from this study and in Beauvoir granites. Horizontal axis: $\log(C/C_0)$ where C is the trace element concentration in the most fractionated glass B2-24 and C_0 the trace element concentration in the parental glass B2-26 (Table 9). Vertical axis: $\log(C/C_0)$ where C is the trace element concentration in B1 granite (ESP 270, Raimbault et al., 1995) and C_0 the trace element concentration in glass B2-26 (Table 9) assumed representative of the B2 granite. Experimental and natural enrichment and depletion directions are indicated. The location of data points relative to the 1:1 dotted line indicates how close elements are from an evolution identical in experiments and granites, whether an enrichment or a depletion. Color coding reflects the four groups of elements distinguished, those enriched both in the experiments and in the granites (red circles), those depleted both in the experiments and in the granites (black circles), those whose behavior should be controlled by accessory phase solubilities (blue squares) and those very much fractionated neither in the experiments nor in the granite (grey diamonds).

187x223mm (300 x 300 DPI)

Table 1. Compositions of starting materials

sample #	B1 WR	B1 glass 1997	B2 WR	B2 glass 1997	B2 glass 2017	
n		6		6	10	
		av*		av*	av	sd
SiO ₂	69.04	69.36	71.62	71.18	72.47	0.88
TiO ₂	0.03	0.03	0.02	0.01	0.02	0.03
Al ₂ O ₃	16.72	16.48	15.79	15.66	14.78	0.87
Fe ₂ O ₃	0.05	nd	0.04	nd	nd	nd
FeO	0.12	0.33	0.47	0.62	0.40	0.06
MnO	0.03	0.02	0.10	0.09	0.07	0.06
MgO	0.02	0.04	0.01	0.04	0.01	0.01
CaO	0.22	0.40	0.19	0.32	0.18	0.05
Na ₂ O	5.11	5.16	4.52	4.50	4.48	0.23
K ₂ O	3.17	2.97	3.84	3.89	4.07	0.16
P ₂ O ₅	1.50	1.52	0.95	0.85	0.51	0.15
SnO ₂	nd	nd	nd	nd	0.00	0.01
F	2.39	1.70	1.77	1.20	2.05	0.30
Li ₂ O	1.28	nd	0.71	nd	nd	nd
CO ₂	0.03	nd	0.03	nd	nd	nd
H ₂ O	0.40	nd	0.23	nd	nd	nd
Total	98.83	98.01	99.58	98.36	99.04	
O=F	1.01	0.72	0.75	0.51	0.86	
Total	97.82	97.29	98.83	97.86	98.18	
Qz	28.34	28.73	31.50	30.78	31.72	
Ab	45.87	46.29	39.86	39.88	39.51	
Or	19.87	18.59	23.65	24.06	25.04	
An	1.16	2.10	0.98	1.66	0.94	
Co	4.76	4.29	4.02	3.62	2.79	
Total	100.00	100.00	100.00	100.00	100.00	
A/CNK	1.37	1.33	1.32	1.28	1.22	

Granite whole-rock (WR) samples analyzed by wet chemistry (Pichavant et al., 1987a)

Glasses analyzed by electron microprobe. n: number of electron microprobe spots; av: average; sd: standard deviation

*data from Lahlafi (1997)

FeO = FeO total except for the WR data (FeO = FeO)

nd: not determined

Qz, Ab, Or, An, Co: normative quartz, albite, orthoclase, anorthite and corundum concentrations

A/CNK = Al₂O₃/(Na₂O+K₂O+CaO) molar

Table 2. Conditions and results of the B1 crystallization experiments.

Charge #	T °C	P MPa	duration h	phase assemblage	Σr^2 control	fO ₂	Δ NNO	Reference
<i>Nancy</i>								
B1N-16	540	100	2064	Qz, Pl, Af, Mu	nd	unbuffered	~+3	
B1N-11	560	100	1800	(L), Qz, Pl, Af, Mu	nd	unbuffered	~+3	(1)
B1N-17	580	100	2064	L(60), Qz(8), Pl(24), Af(6), Mu(1), Ap	0.03	unbuffered	~+3	
B1N-1	600	100	1800	L(84), Qz (2), Pl(13), Mu(1)	0.09	unbuffered	~+3	(1)
B1N-13	640	100	1104	L(100)	nd	unbuffered	~+3	(1)
B1N-9	660	100	1560	L(100)	nd	unbuffered	~+3	(1)
B1N-7	680	100	1560	L(100)	nd	unbuffered	~+3	(1)
B1N-2	700	100	672	L(100)	nd	unbuffered	~+3	(1)
B1N-21	580	300	2184	L(67), Qz(9), Pl(16), Af(6), Mu(1) Tpz, Li-P, Ap	0.00	unbuffered	~+3	
B1N-15	600	300	1104	L(98), Pl(2), Ap	0.11	unbuffered	~+3	(1)
B1N-22	620	300	2184	L(77), Qz(7), Pl(14), Af(2), (Mu), Ap	0.01	unbuffered	~+3	
B1N-23	640	300	2184	L(85), Qz(2), Pl(12), (Mu), Ap	0.06	unbuffered	~+3	
<i>Orléans</i>								
B1O-1	620	300	1008	L(98), Pl(2), Ap	0.02	intrinsic	+1	

L: silicate glass, Qz: quartz, Pl: plagioclase, Af: K-feldspar, Mu: muscovite, Tpz: topaz, Li-P: amblygonite, Ap: apatite.

Δ NNO: deviation in log fO₂ from the NNO oxygen buffer (Chou, 1987) calculated at the same P and T than the experiment.

Numbers in brackets are proportions of major phases (wt%) calculated by mass balance. Σr^2 = sum of residuals. (phase A): < 1 wt% phase A present.

(1) experiments partially described in Pichavant et al. (1987a).

Table 3. Conditions and results of the B2 crystallization experiments.

Charge #	T °C	P MPa	duration h	phase assemblage	Σr^2	fO ₂ control	Δ NNO	Reference
<i>Nancy</i>								
B2N-12	560	100	1800	(L), Qz, Pl, Af, Mu	nd	unbuffered	~+3	(1)
B2N-3	600	100	1080	L(63), Qz(9), Pl(15), Af(9), Mu(4)	0.00	unbuffered	~+3	(1)
B2N-20	620	100	2064	L(29), Qz(20), Pl(28), Af(20), Mu(4), Hm	0.14	unbuffered	~+3	
B2N-14	640	100	1104	L(100)	nd	unbuffered	~+3	(1)
B2N-10	660	100	1560	L, Hm	nd	unbuffered	~+3	(1)
B2N-8	680	100	1560	L, Hm	nd	unbuffered	~+3	(1)
B2N-4	700	100	672	L(100)	nd	unbuffered	~+3	(1)
B2N-18	620	200	2064	L(74), Qz(10), Pl(11), Mu(5)	0.00	unbuffered	~+3	
B2N-19	640	200	2064	L(87), Qz(5), Pl(5), Mu(3), Hm	0.08	unbuffered	~+3	
B2N-24	580	300	2184	L(41), Qz(19), Pl(21), Af(10), Mu(9)	0.02	unbuffered	~+3	
B2N-25	620	300	2184	L(96), Qz(2), Mu(2), Hm, Ap	0.07	unbuffered	~+3	
B2N-26	640	300	2184	L(100)	nd	unbuffered	~+3	
<i>Orléans</i>								
B2O-2/5*	620	300	1008-4032	L, Qz, Pl, (Af), Mu, Mt, Ap	nd	intrinsic	+1	(2)
B2O-6/7*	700	300	168-1008	L, Ap, Mt	nd	intrinsic	+1	(2)
B2O-38/40*	620	300	240-360	L(98), Bt(2), Ap	0.49	CCO buffer	-1.4	(2)
B2O-41	700	300	168	L, Ap	nd	CCO buffer	-1.4	(2)
B2O-50	630	304	1123	L(95), Mu(5), Ap	0.15	Ar-H ₂	-1	

L: silicate glass, Qz: quartz, Pl: plagioclase, Af: K-feldspar, Mu: muscovite, Bt: biotite, Ap: apatite, Hm, hematite, Mt: magnetite.

Δ NNO: deviation in log fO₂ from the NNO oxygen buffer (Chou, 1987) calculated at the same P and T than the experiment.

Numbers in brackets are proportions of major phases (wt%) calculated by mass balance. Σr^2 = sum of residuals. nd: not determined (mass balance calculation not performed). (phase A): < 1 wt% of phase A present.

*Multiple experiments were performed for different durations under identical P-T conditions and results are grouped (see text).

- (1) experiments partially described in Pichavant et al. (1987a).
- (2) experiments partially described in Lahlafi (1997) and Pichavant et al. (2016).

Table 4. Major element compositions of experimental glasses

sample #	B1		B1		B1		B1	
charge #	B1N-1		B1N-15		B1N-17		B1N-21	
n	5		5		6		6	
	av	sd	av	sd	av	sd	av	sd
SiO ₂	63.37	0.22	62.70	0.63	61.56	0.55	61.90	0.98
TiO ₂	0.01	0.01	0.00	0.01	0.02	0.02	0.01	0.01
Al ₂ O ₃	15.23	0.23	15.42	0.26	15.85	0.27	16.54	0.62
FeO	0.19	0.11	0.13	0.09	0.15	0.08	0.05	0.05
MnO	0.03	0.04	0.04	0.04	0.04	0.05	0.04	0.05
MgO	0.04	0.05	0.03	0.02	0.01	0.02	0.01	0.01
CaO	0.04	0.02	0.06	0.03	0.04	0.02	0.07	0.02
Na ₂ O	4.40	0.11	4.73	0.11	3.95	0.20	4.67	0.19
K ₂ O	3.34	0.03	3.07	0.09	3.08	0.15	2.91	0.19
P ₂ O ₅	1.51	0.16	1.42	0.14	1.70	0.18	1.38	0.09
SnO ₂	nd		nd		nd		nd	
F	2.37	0.10	2.26	0.05	3.16	0.09	3.31	0.32
Total	90.52		89.87		89.55		90.88	
O=F	1.00		0.95		1.33		1.39	
Total	89.52		88.91		88.22		89.49	
Qz	28.86		27.07		31.61		27.26	
Ab	43.07		46.58		39.59		45.89	
Or	22.84		21.13		21.54		19.96	
An	0.26		0.33		0.22		0.39	
Co	4.99		4.89		7.04		6.50	
Total	100.00		100.00		100.00		100.00	
A/CNK	1.39		1.37		1.60		1.51	

Glasses analyzed by electron microprobe. n: number of electron microprobe spots; av: average; sd: standard deviation. For B1, glasses from charges B1N-11, -13, -9, -7, -2 (Table 2) have not been analyzed and for B2, glasses from charges B2N-10, -8, -4 and B2O-6/7, -41 (Table 3)

*data from Lahfafi (1997) normalized to 100%

FeO = FeO total

nd: not determined

Qz, Ab, Or, An, Co: normative quartz, albite, orthoclase, anorthite and corundum concentrations

A/CNK = Al₂O₃/(Na₂O+K₂O+CaO) molar

B1		B1		B1		B2		B2	
B1N-22		B1N-23		B1O-1		B2N-3		B2N-14	
5		6		6		5		5	
av	sd	av	sd	av*	av	sd	av		
62.22	0.54	63.20	0.44	71.19	65.93	0.66	65.48	1.13	
0.01	0.02	0.01	0.01	0.02	0.02	0.02	0.01	0.02	
16.10	0.24	16.00	0.18	17.22	14.43	0.16	14.38	0.57	
0.15	0.04	0.19	0.12	0.03	0.16	0.08	0.19	0.06	
0.05	0.07	0.06	0.08	0.00	0.14	0.05	0.08	0.06	
0.03	0.04	0.02	0.02	0.01	0.01	0.01	0.01	0.01	
0.07	0.03	0.05	0.01	0.13	0.12	0.03	0.16	0.02	
4.43	0.12	4.52	0.11	5.20	3.72	0.20	4.10	0.18	
3.25	0.10	3.31	0.04	3.28	3.58	0.07	3.75	0.14	
1.68	0.28	1.53	0.08	1.69	1.02	0.13	0.96	0.07	
nd		nd		nd	nd		nd		
2.81	0.10	2.59	0.08	2.12	1.89	0.22	1.62	0.13	
90.81		91.48		100.90	90.99		90.75		
1.18		1.09		0.89	0.80		0.68		
89.63		90.39		100.00	90.20		90.07		
27.70		27.66		28.96	34.59		30.61		
43.58		43.95		45.38	35.83		39.50		
22.32		22.48		19.97	24.09		25.25		
0.40		0.31		0.66	0.67		0.92		
6.00		5.59		5.03	4.82		3.72		
100.00		100.00		100.00	100.00		100.00		
1.47		1.44		1.40	1.41		1.29		

B2 B2N-18		B2 B2N-19		B2 B2N-20		B2 B2N-24	
7		10		6		2	
av	sd	av	sd	av	sd	av	sd
65.02	0.23	65.09	0.38	62.90	0.56	64.50	0.46
0.02	0.02	0.03	0.05	0.03	0.02	0.02	0.02
15.05	0.09	14.41	0.13	15.01	0.13	15.63	0.24
0.30	0.13	0.12	0.03	0.08	0.05	0.14	0.13
0.07	0.03	0.07	0.07	0.12	0.07	0.07	0.08
0.01	0.01	0.02	0.02	0.01	0.01	0.01	0.01
0.13	0.03	0.18	0.02	0.15	0.03	0.04	0.02
4.31	0.11	4.03	0.17	3.96	0.20	4.38	0.25
3.99	0.04	3.61	0.14	3.36	0.23	3.27	0.12
1.11	0.10	1.01	0.17	1.27	0.14	1.27	0.13
nd		0.00		nd			
2.02	0.15	2.67	0.33	2.57	0.07	2.61	0.09
92.04		91.24		89.45		91.94	
0.85		1.12		1.08		1.10	
91.19		90.11		88.37		90.84	
27.55		31.45		31.25		30.07	
41.19		39.03		39.24		42.25	
26.66		24.43		23.28		22.00	
0.75		1.03		0.86		0.21	
3.84		4.06		5.38		5.48	
100.00		100.00		100.00		100.00	
1.29		1.33		1.44		1.44	

B2		B2		B2		B2		B2	
B2N-25		B2N-26		B2O-2/5		B2O-38/40		B2O-50	
2		10		8		12		6	
av	sd	av	sd	av*	sd	av*	sd	av	sd
64.80	0.56	64.52	0.39	73.48	0.02	73.97	0.04	67.79	0.66
0.01	0.02	0.03	0.05	0.02	0.02	0.04	0.04	0.02	0.03
14.46	0.23	13.94	0.21	15.76	0.16	15.79	0.51	13.72	0.21
0.16	0.03	0.42	0.05	0.16	0.09	0.51	0.09	0.36	0.05
0.08	0.06	0.09	0.07	0.11	0.07	0.09	0.03	0.07	0.07
0.01	0.01	0.01	0.01	0.07	0.11	0.03	0.27	0.01	0.01
0.08	0.05	0.15	0.02	0.11	0.27	0.27	0.16	0.16	0.04
4.13	0.14	3.99	0.15	4.57	4.57	4.07	4.38	4.38	0.20
3.55	0.10	3.63	0.17	3.77	3.77	3.78	3.66	3.66	0.12
0.85	0.39	1.00	0.12	1.19	1.19	0.74	0.41	0.41	0.10
0.01	0.01	0.01	0.01	nd	nd	nd	0.00	0.00	0.01
2.52	0.31	2.33	0.32	1.33	1.33	1.24	1.61	1.61	0.32
90.65		90.11		100.57		100.53		92.19	
1.06		0.98		0.56		0.52		0.68	
89.59		89.12		100.01		100.01		91.51	
31.02		31.43		32.96		36.01		31.15	
40.20		39.17		39.61		35.18		41.35	
24.10		24.87		22.82		22.85		24.07	
0.46		0.85		0.55		1.35		0.91	
4.22		3.68		4.05		4.61		2.52	
100.00		100.00		100.00		100.00		100.00	
1.34		1.30		1.34		1.40		1.20	

Table 5. Representative compositions of experimental quartz

sample #	B1		B1		B1		B2		B2	
charge #	B1N-17		B1N-21		B1N-22		B2N-3		B2N-18	
SiO ₂	94.59	94.12	98.67	98.49	99.78	98.52	99.15	99.89	99.25	
TiO ₂	0.00	0.00	0.04	0.00	0.00	0.00	0.00	0.00	0.03	
Al ₂ O ₃	2.88	3.01	0.43	0.08	0.18	0.20	0.08	0.23	0.33	
FeO	0.00	0.00	0.01	0.07	0.00	0.00	0.00	0.01	0.13	
MnO	0.07	0.06	0.08	0.06	0.00	0.00	0.13	0.00	0.00	
MgO	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	
CaO	0.00	0.00	0.00	0.00	0.02	0.01	0.03	0.01	0.00	
Na ₂ O	0.02	0.11	0.05	0.00	0.02	0.01	0.01	0.03	0.00	
K ₂ O	0.00	0.09	0.01	0.01	0.02	0.00	0.00	0.00	0.06	
P ₂ O ₅	0.09	0.14	0.03	0.00	0.00	0.00	0.05	0.09	0.05	
SnO ₂	nd	nd	nd	nd	nd	nd	nd	nd	nd	
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	
Total	97.66	97.53	99.31	98.71	100.01	98.74	99.44	100.25	99.87	
O=F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	
Total	97.66	97.53	99.31	98.71	100.01	98.74	99.44	100.25	99.87	

FeO = FeO total

nd: not determined

Table 6. Representative compositions of experimental feldspars

sample #	B1		B1		B1		B1		B1	
charge #	B1N-1		B1N-15		B1N-17		B1N-21		B1N-22	
feldspar	Pl	Pl*	Pl*	Pl	I	Af	Pl	Af	Pl	Af
SiO ₂	65.12	67.87	67.41	67.15	65.09	64.22	67.43	63.93	66.90	63.31
TiO ₂	0.02	nd	nd	0.02	0.00	0.00	0.00	0.00	0.00	0.00
Al ₂ O ₃	19.75	20.31	20.13	19.47	19.34	19.78	20.46	19.06	20.43	19.09
FeO	0.00	nd	nd	0.00	0.12	0.00	0.00	0.00	0.07	0.19
MnO	0.22	nd	nd	0.00	0.00	0.07	0.06	0.00	0.00	0.00
MgO	0.00	nd	nd	0.00	0.00	0.00	0.00	0.00	0.01	0.00
CaO	0.57	0.18	0.15	0.12	0.35	0.03	0.51	0.01	0.53	0.00
Na ₂ O	10.06	10.30	10.27	9.39	6.81	3.41	9.99	2.91	9.83	3.26
K ₂ O	1.40	1.24	1.23	2.00	5.80	11.55	1.62	12.28	1.50	11.62
P ₂ O ₅	1.36	nd	nd	1.24	1.25	0.96	0.66	1.02	1.15	0.82
F	0.00	nd	nd	0.00	0.25	0.01	0.04	0.10	0.09	0.00
Total	98.48	99.90	99.19	99.39	99.00	100.02	100.77	99.30	100.51	98.29
O=F	0.00	0.00	0.00	0.00	0.10	0.00	0.02	0.04	0.04	0.00
Total	98.48	99.90	99.19	99.39	98.89	100.02	100.75	99.26	100.47	98.29
Or	8.1	7.3	7.3	12.2	35.3	68.9	9.4	73.5	8.9	70.1
Ab	89.1	91.8	92.0	87.2	62.9	30.9	88.1	26.5	88.5	29.9
An	2.8	0.9	0.7	0.6	1.8	0.2	2.5	0.0	2.7	0.0

Pl: plagioclase; Af: K-feldspar, I: heterogeneous feldspar

*data from Pichavant et al. (1987a)

**data from Lahlafi (1997)

FeO = FeO total

nd: not determined

Ab, Or, An: molar proportions of albite, orthoclase, anorthite end-members

B1 B1N-23 PI	B1 B1O-1 PI**	B2 B2N-3 PI	PI*	Af	I	Af*	B2 B2N-18 PI	B2 B2N-19 PI	B2 B2N-20 PI	I
66.90	65.75	66.23	66.50	64.47	63.36	65.29	66.20	66.06	65.82	64.55
0.00	nd	0.00	nd	0.00	0.00	nd	0.00	0.03	0.00	0.01
20.43	20.04	20.72	20.48	19.05	19.49	19.48	20.31	20.57	20.94	20.15
0.07	nd	0.17	nd	0.28	0.00	nd	0.06	0.11	0.14	0.00
0.00	nd	0.13	nd	0.00	0.00	nd	0.00	0.15	0.00	0.03
0.01	nd	0.00	nd	0.01	0.03	nd	0.01	0.01	0.00	0.00
0.53	0.43	0.57	0.69	0.00	0.13	0.01	0.60	0.88	0.80	0.41
9.83	10.55	10.45	9.19	4.54	5.61	4.90	10.08	10.44	9.60	8.57
1.50	0.94	1.17	2.25	9.88	8.74	9.04	1.67	1.02	1.55	4.34
1.15	0.90	1.04	nd	1.01	0.67	nd	0.84	0.91	0.88	1.08
0.09	nd	0.00	nd	0.00	0.00	nd	0.00	0.00	0.05	0.00
100.51	98.61	100.48	99.11	99.23	98.03	98.72	99.77	100.18	99.78	99.14
0.04	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00
100.47	98.61	100.48	99.11	99.23	98.03	98.72	99.77	100.18	99.76	99.14
8.9	5.4	6.7	13.4	58.9	50.3	54.8	9.6	5.8	9.2	24.5
88.5	92.5	90.6	83.2	41.1	49.1	45.1	87.6	90.0	86.8	73.5
2.7	2.1	2.7	3.5	0.0	0.6	0.1	2.9	4.2	4.0	2.0

	B2		B2
	B2N-24		B2O-2/5
Af	Pl	Af	I**
63.78	66.77	66.28	63.45
0.00	0.01	nd	nd
19.04	20.16	19.91	18.49
0.00	0.00	nd	nd
0.00	0.00	nd	nd
0.00	0.01	nd	nd
0.07	0.17	0.2	0.07
4.52	10.25	7.52	2.89
10.03	1.75	5.23	11.60
0.92	0.75	0.9	0.44
0.06	0.00	nd	nd
98.43	99.88	100.04	96.94
0.02	0.00	0.00	0.00
98.40	99.88	100.04	96.94
59.1	10.0	31.1	72.3
40.5	89.2	67.9	27.4
0.3	0.8	1.0	0.4

Table 7. Representative compositions of experimental micas

sample #	B1	B1	B1	B2	B2	B2	B2	B2	B2	B2	B2
charge #	B1N-21	B1N-22	B1N-23	B2N-3	B2N-18	B2N-19	B2N-20	B2N-24	B2N-25	B2O-2/5*	B2O-50
SiO ₂	53.61	51.19	51.04	47.36	45.92	45.00	45.62	47.07	43.70	45.00	45.74
TiO ₂	nd	nd	nd	0.21	0.20	0.05	0.00	0.00	0.00	0.04	0.06
Al ₂ O ₃	21.50	19.74	20.01	29.56	31.88	34.25	34.90	30.25	32.82	33.25	33.66
FeO	2.18	5.34	5.40	7.43	5.12	2.87	1.81	4.57	4.92	3.49	2.28
MnO	nd	nd	nd	0.17	0.17	0.07	0.22	0.44	0.00	0.04	0.00
MgO	0.16	1.17	2.26	0.60	0.47	0.27	0.11	0.10	0.16	0.29	0.26
CaO	nd	nd	nd	0.02	0.04	0.02	0.00	0.03	0.05	0.00	0.00
Li ₂ O	6.75	6.04	6.80	1.31	1.75	2.62	3.00	2.50	1.79	0.61	1.13
Na ₂ O	2.03	1.30	1.43	0.83	0.70	1.02	1.26	0.87	0.85	0.85	0.95
K ₂ O	6.24	7.26	7.12	9.17	10.15	8.96	9.56	9.13	9.65	9.59	10.12
P ₂ O ₅	0.78	0.78	0.23	0.17	0.10	0.39	0.30	0.23	0.06	0.00	0.24
SnO ₂	nd	nd	nd	nd	nd	0.17	0.11	0.10	0.14	nd	0.03
F	8.53	7.84	8.58	2.47	2.87	4.18	4.62	4.03	3.14	1.40	2.22
Total	101.79	100.66	102.87	99.31	99.37	99.89	101.51	99.33	97.27	94.56	96.69
O=F	3.59	3.30	3.61	1.04	1.21	1.76	1.95	1.70	1.32	0.59	0.94
Total	98.19	97.36	99.26	98.27	98.16	98.13	99.57	97.63	95.95	93.98	95.76
Mg-Li	-3.2555	-2.8253	-2.9487	-0.5697	-0.8262	-1.2878	-1.4836	-1.2776	-0.9247	-0.2760	-0.5502
Fe _t +Mn+Ti-Al ^{VI}	-1.3250	-0.7700	-0.6168	-1.9034	-2.2680	-2.5341	-2.6212	-2.1248	-2.3983	-2.9863	-2.9897

Analyses by SEM for the B1 and by EMPA for the B2 charges

Li₂O estimated from the F concentration (Tischendorf et al., 1997)

*data from Lahlafi (1997)

FeO = FeO total

Table 8. Representative compositions of accessory phases

sample # charge #	B1			B2				
	B1N-21			B2N-25	B2N-19	B2N-25		
phase	topaz	topaz	amblygonite	amblygonite	apatite	hematite	hematite	hematite
SiO ₂	32.07	37.01	4.56	2.05	0.07	1.26	0.95	0.13
TiO ₂	0.03	0.00	0.19	0.11	0.08	0.14	0.33	2.42
Al ₂ O ₃	57.60	50.47	35.34	35.39	0.01	2.71	2.22	0.79
FeO	0.04	0.04	0.11	0.03	0.32	83.14	85.19	83.53
MnO	0.15	0.07	0.14	0.00	3.63	0.13	0.23	0.00
MgO	0.04	0.01	0.02	0.04	0.03	0.00	0.04	0.02
CaO	0.02	0.05	0.00	0.03	51.84	0.00	0.05	0.02
Na ₂ O	0.06	0.49	0.24	0.23	0.00	0.13	0.11	0.00
K ₂ O	0.08	0.30	0.39	0.05	0.00	0.12	0.09	0.00
P ₂ O ₅	0.98	0.78	49.57	51.69	38.15	0.11	0.00	0.00
SnO ₂	0.00	0.00	0.00	0.00	0.02	1.39	1.40	1.57
F	20.93	18.33	13.25	13.07	5.63	0.22	0.13	0.00
Total	112.00	107.55	103.81	102.70	99.76	89.35	90.74	88.48
O=F	8.81	7.72	5.58	5.50	2.37	0.09	0.06	0.00
Total	103.18	99.83	98.23	97.19	97.39	89.26	90.69	88.48
XF	0.99	0.89	0.95	0.94	1.55	Hm=0.997	0.993	0.950
XOH	0.01	0.11	0.05	0.06	-0.55	Ilm=0.003	0.007	0.050

Total Fe as FeO

XF = at. F/(F+OH) with F+OH = 2 for topaz and apatite and F+OH = 1 for amblygonite

XOH = 1-XF

Hm=hematite, Ilm=ilmenite

Table 9. Selected trace element concentrations

sample #	B2								B2	B2
charge #	B2N-26		B2N-25		B2N-18		B2N-24		B2N-25	B2N-25
phase	Glass		Glass		Glass		Glass		Quartz	Kd Qz/L
	av (ppm)	sd/av(%)	av (ppm)	sd/av(%)	av (ppm)	sd/av(%)	av (ppm)	sd/av(%)	ppm	
Li	2737.6	1.1	3021.1	3.8	3648.6	5.2	4310.3	10.8	81.3	0.027
Be	32.0	3.3	37.5	7.8	46.6	8.6	53.6	15.8	<0.8	
B	10.0	4.9	11.6	9.3	11.5	10.6	14.9	19.2	<1.9	
Na	26154.3	1.5	27739.9	1.7	27848.6	3.9	29257.3	6.9	<22.6	
Al	66231.2	2.0	70296.0	4.8	78139.7	7.4	74862.6	5.0	367.9	0.005
P	3057.8	2.9	3110.0	3.7	3547.7	4.7	3621.3	10.3	152.8	0.049
Ti	41.8	7.5	19.4	10.5	66.9	14.5	41.4	2.2	2.4	0.123
Mn	483.0	2.0	466.4	1.4	598.8	4.4	501.8	5.0	<3.9	
Zn	114.6	6.8	108.1	4.4	110.9	9.1	98.2	8.4	2.8	0.026
Rb	2024.2	0.6	2057.8	2.9	2420.9	4.5	2550.3	7.1	<1.0	
Sr	9.0	4.7	9.5	3.5	3.3	17.9	2.4	2.7	<0.1	
Zr	3.3	6.9	2.8	26.7	4.0	12.8	4.3	7.6	<0.0	
Nb	38.1	4.0	32.5	2.3	44.3	7.0	49.2	3.6	<0.0	
Sn	35.4	3.6	16.3	26.6	24.3	16.8	35.1	22.3	3.0	0.185
Cs	624.6	1.3	655.1	2.7	782.7	4.3	1027.6	10.1	2.0	0.003
Ba	0.8	14.9	0.7	32.7	0.4	60.7	1.3	70.7	0.2	
La	0.1	13.9	0.1	72.8	0.3	21.1	0.0	6.1	<0.0	
Ta	23.2	5.7	20.2	12.2	28.7	17.6	33.5	9.3	<0.0	
W	36.2	3.6	37.4	3.6	40.2	6.2	44.0	13.5	0.1	0.002
Pb	78.1	1.6	94.8	3.4	240.5	3.5	50.0	7.8	<0.3	
U	8.6	4.5	9.7	4.2	13.3	8.6	13.3	11.4	<0.0	
Nb/Ta	1.64		1.61		1.54		1.47			

sd : standard deviation ; av : average