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1 ***Can destabilization rims of hydrous minerals be used to constrain magma***  
2 ***ascent kinetics at lava dome volcanoes?***

3

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6

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10

11

12 **Keywords:**

13 biotite destabilization; trachyte lava; plumbing system; degassing; Chaîne des Puys; Sarcoui

14

15

16

17 ***Abstract***

18 Time constraints on igneous processes related to eruption triggering, e.g., magma mixing or  
19 ascent in the conduit, are needed in any risk mitigation attempt. In this context, magma  
20 ascent rate and kinetics are key parameters as they may correspond to the response time  
21 available to civil protection during volcanic unrest. Several tools available to quantify such  
22 durations include diffusion chronometry, isotopic geochemistry, and hydrous mineral  
23 destabilization related to magma degassing during ascent through the conduit. Here I discuss  
24 the possible limitations of the widely used hydrous mineral destabilization chronometry  
25 applied to minerals within large lava bodies that cool relatively slowly (over days or weeks)  
26 such as thick lava flows and lava domes. Based on the type case of the Sarcoui dome (Chaîne  
27 des Puys, France) and its associated phreatomagmatic outbreak deposits, I suggest that  
28 hydrous mineral destabilization rims may, in some cases, develop at the surface during dome  
29 emplacement. From this perspective, preeruptive timescales calculated based on  
30 disequilibrium kinetics will be greatly underestimated, leading to a serious issue in the  
31 reconstruction of the eruption dynamics and its possible applications to emergency  
32 management for future eruptions. More generally, hydrous mineral destabilization  
33 chronometry should be used with great caution. Nevertheless, it remains a choice tool to  
34 quantify magma ascent rates for eruptions during which magmas quench upon arrival at the  
35 surface (e.g., Plinian, Vulcanian, or phreatomagmatic eruptions), with pumiceous textures  
36 being a good indicator of quenching. In the case of lava dome emplacement, I suggest that  
37 the minerals embedded in pumiceous clasts emitted during explosive phases are more  
38 reliable candidates for chronometry studies than crystals within the dome itself as those  
39 clasts might represent the fresh magma that triggered the explosion.

## 40 ◦ **Introduction**

41           Understanding the mechanisms and quantifying the time scales and kinetics  
42 associated with volcanic unrest, eruption triggering, and magma ascent to the surface  
43 remain some of the holy grails of modern volcanology and igneous petrology. This has  
44 particular societal impact in the case of silica-rich magmas associated with the most  
45 explosive and destructive eruptions, for which risk mitigation requires such time constraints.  
46 In addition to the information provided by seismic and geophysical monitoring (e.g., [Endo et al. 1981](#);  
47 [Gudmundsson et al. 2014](#); [Sigmundsson et al. 2014](#)), igneous rocks record the  
48 various stages of their evolution, offering the potential for time constraints (e.g., [Spiegelman  
& Elliott 1993](#); [Rutherford 2008](#); [Druitt et al. 2012](#)). The main chronometric tools available  
49 are diffusion chronometry (e.g., [Demouchy et al., 2006](#); [Boudon et al. 2015](#); [Shea et al. 2015](#);  
50 [Ferguson et al., 2016](#); [Costa et al., 2020](#)), U-Th disequilibrium series ([Girard et al. 2017](#)), and  
51 mineral destabilization chronometry ([Rutherford & Hill 1993](#); [Faure et al., 2001](#); [Browne &  
52 Gardner, 2006](#)).

54           Since the pioneering study of [Rutherford & Hill \(1993\)](#), who quantified the ascent rate  
55 of Mount St. Helens dacite magmas in both dome forming eruptions, and large explosive  
56 Plinian to Subplinian eruptions, hundreds of studies have used their experimentally  
57 calibrated approach. Their method relies on the advancement of the destabilization reaction  
58 that affects hydrous minerals (e.g., pyroxene-oxide-plagioclase rims around amphibole)  
59 during magma degassing. Although initially calibrated for amphibole, similar destabilization  
60 reactions are reported for other hydrous minerals like biotite ([Venezky & Rutherford, 1999](#);  
61 [Underwood et al., 2012](#)). This reaction is usually considered to occur during magma ascent  
62 as volatile solubility decreases dramatically upon decompression. Alternatively, it has been  
63 proposed and demonstrated experimentally that the reaction may occur when a more

64 primitive hotter magma intrudes a differentiated and partially cooled magma chamber (e.g.,  
65 [Feeley & Sharp, 1996](#); [Plechov et al. 2008](#); [De Angelis et al., 2015](#)). Decompression-related  
66 dehydration drives hydrous minerals (e.g., amphibole or mica) to unstable conditions  
67 ([Rutherford 2008](#)), forming a reaction rim whose width is directly correlated to the duration  
68 between the onset of mineral destabilization and the final quenching of the lava. It is  
69 generally assumed that magmas are volatile saturated at mid-crustal or shallow magma  
70 chamber depths and that degassing starts as soon as magma begins its ascent (as an  
71 example, [Métrich & Wallace, 2008](#)). Thus, the depth of the magma chamber and the  
72 duration of degassing obtained from destabilization rims can be converted into an ascent or  
73 decompression rate. Nevertheless, volatile saturation of magmas at magma chamber depth  
74 should be demonstrated for each case study, as volatile-undersaturated melts would start to  
75 degas at shallower levels in the conduit, leading to over-estimated ascent rates. More  
76 generally any process with the potential to destabilize hydrous minerals should be carefully  
77 explored before proposing an ascent rate of magmas. As an example, during the cooling of  
78 large lava effusions or extrusions at the surface (e.g., thick lava flows or lava domes), the  
79 temperature of the lava may remain largely above the solidus temperature for several days  
80 or weeks (as an example, [Dzurisin et al., 1990](#)), enhancing the destabilization of hydrous  
81 minerals and prolonging the development of the destabilization rim ([Garcia & Jacobson,](#)  
82 [1979](#); [Devine et al., 1998](#)); such a process would lead to under-estimated ascent rates.

83         Based on the type case of the Sarcoui dome (Chaîne des Puys, France) destabilized  
84 biotite crystals, I show here that in some cases the destabilization rims may develop after  
85 magma ascent and during lava dome emplacement. I then explore more generally the utility  
86 of hydrous mineral destabilization rims in estimating the ascent rates of explosive and  
87 effusive or extrusive eruptions.

88

89 ◦ **Geological context**

90           The Chaîne des Puys (CdP) is situated in the French Massif Central to the west of the  
91 Oligocene Limagne rift and borders the city of Clermont-Ferrand, a metropolitan area with  
92 ~480,000 inhabitants (Figure 1). The CdP is a 30-km-long north-south alignment of around a  
93 hundred monogenic cones, domes, and maars emplaced between ~100 and  $6.7 \pm 0.1$  ka  
94 (Figure 1; Juvigne 1992; Boivin et al. 2009). French Massif Central magmatism is related to  
95 asthenospheric upwelling (Hoernle et al. 1995; Granet et al. 1995) and melting of a strongly  
96 metasomatized HIMU-type mantle (Lustrino & Wilson 2007; Hamelin et al. 2009; France et  
97 al. 2015; Gu et al. 2016, 2018; the HIMU isotopic signature likely documenting the influence  
98 of a recycled component in the mantle source, Zindler & Hart, 1986). Volatile- and silica-rich  
99 magmas, like the trachyte lavas forming the Sarcoui edifice studied herein ( $\sim 12.5 \pm 0.5$  ka;  
100 Miallier et al. 2017), formed through protracted differentiation, and ~5%–10% crustal  
101 contamination (Villemant et al. 1980; Condomines et al. 1982; Hamelin et al. 2009; Martel et  
102 al. 2013; France et al. 2016). Recently, it has been proposed that the CdP plumbing system  
103 may still contain some residual melt, although any potential eruption would require an  
104 unlikely mush rejuvenation (France et al. 2016) as described for large silicic eruptions  
105 (Bachmann & Bergantz, 2003; Hubert et al., 2010).

106           Sarcoui volcano is a lava dome of ca. 1 km in diameter and of ca. 150 m high; it is  
107 composed of massive trachyte lava containing biotite and alkali feldspar phenocrysts (whole-  
108 rock compositions available in Martel et al. 2013; Table 1). The lava dome has suffered  
109 several destabilization events and is bordered by several block-and-ash deposits (BAD) that  
110 are locally buried below massive trachyte lavas, highlighting that destabilization occurred  
111 during dome growth (Miallier et al. 2017).

112

113 ◦ **Methods**

114 Field work was conducted to reconstruct the stratigraphy, field relations between the  
115 various facies, and sample the key lithologies. Major element mineral compositions and  
116 backscattered electron images (BSE) were determined using a CAMECA SX100 electron  
117 probe (SCMEM; Université de Lorraine, France) equipped with five spectrometers and the  
118 “Peak sight” operating system. Data were obtained using a 12 kV acceleration potential, a  
119 static (fixed) beam, K $\alpha$  emissions from all elements, and a matrix correction based on  
120 [Pouchou & Pichoir \(1991\)](#). Crystals were analyzed with a beam current of 10 nA, a focused  
121 beam, and counting times of 20 s on background and 40 s on peak.

122

123 ◦ **Results: Stable and destabilized biotite at Sarcoui volcano**

124 ***Field evidence of an initial explosive phase followed by lava dome emplacement***

125 At various outcrops located in the two northeast quarries, pyroclastic deposits are  
126 recognized below the dome deposits ([Figure 2](#)), and attest of an initial explosive phase for  
127 which fragmentation could have been phreatomagmatic. Samples for the present study  
128 were collected in two old quarries (45°48'54.4"N, 2°59'15.9"E, and 45°48'58.9"N,  
129 2°59'09.9"E; [Table S1](#)). Both outcrops contain dark and roundish, relatively dense (in  
130 comparison to the massive dome trachyte lava) trachyte lapilli and blocks with a quenched  
131 aspect that can be classified as cauliflower bombs ([Figures 2a, b](#)). Also present are angular  
132 accidental blocks of basalt and gneiss (full thin sections presented in [Figures 2c and d](#),  
133 respectively) embedded in an ash matrix. Most of the trachyte lapilli and blocks display a  
134 banded texture with vesicle-poor and -rich bands that seems to preserve conduit-related  
135 shearing ([Shea et al. 2012; France et al. 2016](#)). The characteristics of the pyroclastic deposits

136 related to this initial explosive phase, and more specifically the presence of cauliflower  
137 bombs, and of angular accidental blocks of basalt and gneiss embedded in an ash matrix  
138 suggest that the fragmentation could have been phreatomagmatic. Although the origin of  
139 fragmentation should be considered with caution ([White & Valentine, 2016](#)), based on those  
140 evidences, and relying on previous volcanological studies in the Chaîne des Puys area (e.g.,  
141 [De Goer et al., 1992](#); [Miallier et al., 2004](#); [Boivin et al., 2009](#)), I will hereafter name this  
142 pyroclastic unit "phreatomagmatic deposits". This phreatomagmatic deposit may  
143 correspond to units SU2 and SU3 from [Miallier et al. \(2017\)](#), although they did not identify  
144 the accidental angular and fragmented material. Similar to their observations, no paleosol or  
145 erosion features have been observed between the phreatomagmatic deposits and the  
146 overlying dome lavas or pyroclastic density currents related BADs ([Figure 2](#)).  
147 Thermoluminescence dating of the phreatomagmatic deposits, and  $^{238}\text{U}$ - $^{230}\text{Th}$  dating of  
148 dome samples give similar ages ( $12.6 \pm 3$  ka, and  $12.8 \pm 1.4$  ka, respectively,  $2\sigma$  errors;  
149 [Miallier et al. 2004](#)). Field evidence and geochronology therefore support a model in which  
150 the phreatomagmatic stage represents the initial breakout that was directly followed by  
151 dome emplacement, in agreement with early detailed field studies ([De Goër et al. 1991](#)).

152

### 153 ***Petrography and phase compositions***

154 The similar mineral assemblages and the perfect match of mineral compositions of  
155 the dome and initial phreatomagmatic products presented below, highlight that both  
156 eruptive phases were fed by the same trachyte magma. Both trachyte magmas display  
157 porphyritic textures with 5 to 10% phenocrysts of plagioclase ( $2 \times 0.6$  mm) and biotite ( $\sim 1.5$   
158 mm) with no difference observed between the phenocryst fraction of dome and  
159 phreatomagmatic samples (point counting estimates). [Martel et al. \(2013\)](#) showed that this



160 mineral assemblage in the Sarcoui magma had an equilibrium temperature of  $\sim 800^{\circ}\text{C}$ .  
161 Plagioclase is zoned, and biotite displays skeletal morphologies (Figure 3). The matrix is  
162 composed of microlites of plagioclase with potassic feldspar rims, biotite, and iron oxide.  
163 Microphenocrysts ( $\sim 100\ \mu\text{m}$ ) of apatite and zircon are recognized throughout the matrix and  
164 included in the biotite and plagioclase phenocrysts. Crystallites, the smallest microlites, are  
165 also similar in composition and assemblage (potassic feldspar, biotite, and iron oxide)  
166 between the dome and phreatomagmatic samples, but are  $\sim 2$  times larger in dome samples.  
167 Phenocrysts and microlites are oriented in the same direction in all samples; in the  
168 cauliflower bombs, this highlights that the larger microlites had already crystallized when the  
169 trachyte magma was migrating through the conduit. Crystallites are only locally oriented. In  
170 dome samples, all biotite crystals are rimmed by dark, oxide-rich, in average  $20\ \mu\text{m}$  large  
171 destabilization coronas that contain potassic feldspar, Ti-magnetite, plagioclase, and rare  
172 ilmenite (Figures 3, 4). Only rare Mg-rich clinopyroxene is present suggesting that in a  
173 similar way as for amphibole destabilization, the reaction is not equilibrated, and thus have  
174 likely proceeded with melt present (Rutherford & Hill, 1993). Several examples of reaction  
175 rims are presented in Figure 4 along with an attempt of mass balance based on compositions  
176 presented in Table 1. The estimated reaction is  $0.5\ \text{biotite} + 0.5\ \text{melt} \Rightarrow 0.63\ \text{K-feldspar} +$   
177  $0.18\ \text{titanomagnetite} + 0.12\ \text{plagioclase} + 0.04\ \text{clinopyroxene} + 0.03\ \text{ilmenite} + \text{melt}_2$  (with  
178  $\text{melt}_2$  enriched in both Mg and K, and depleted in Fe). Similar to the pioneering work of  
179 Rutherford & Hill (1993), the mass balance residuals are relatively high, suggesting that the  
180 reaction proceeded with melt present before and after an increment of reaction. In contrast,  
181 no destabilization features are observed in phreatomagmatic samples (Figure 3). The  
182 matrices of phreatomagmatic samples contain micro-vesicle-rich and -poor bands,  
183 consistent with the low densities of the bombs ( $1.5\text{-}2\ \text{g/cm}^3$ ; France et al. 2016; Miallier et al.

184 2017), whereas dome sample matrices contain widespread vesicles, consistent with the  
185 lower densities locally observed in the dome lavas ( $\sim 1.3 \text{ g/cm}^3$ ; France et al. 2016; Miallier et  
186 al. 2017).

187 Mineral compositions are very similar in the dome and phreatomagmatic samples  
188 (Figure 5). Biotite contains 5.5 wt%  $\text{TiO}_2$ , 14 wt%  $\text{Al}_2\text{O}_3$ , and 12.5 wt% MgO with no  
189 compositional difference between the massive dome trachyte lava and the dense cauliflower  
190 bombs (Table 1). Feldspar compositions (Figure 5; Table 1) vary from andesine ( $\text{An}_{31}\text{Ab}_{64}\text{Or}_5$ )  
191 to alkali feldspar ( $\text{An}_1\text{Ab}_{49}\text{Or}_{50}$ ). Phenocryst cores are more Ca-rich (andesine to oligoclase)  
192 than microlite and crystallite cores (oligoclase to anorthoclase). The rims of all feldspar  
193 grains display more evolved compositions that range from  $\text{Or}_{33}$  to  $\text{Or}_{50}$ .

194 Oxygen isotopes measured in zircon grains record crustal contamination at magma  
195 chamber depth, but display no variation between the dome and phreatomagmatic samples  
196 (France et al. 2016).

197

## 198 ◦ **Discussion**

### 199 ***Development of biotite destabilization rims at Sarcoui lava dome***

200 The various petrographic and geochemical features described herein and in the  
201 literature show that the same magma, which experienced similar magma chamber and  
202 conduit processes, is responsible for the dome and phreatomagmatic eruptions at Sarcoui.  
203 Those various results highlight that the lava dome was emplaced immediately after the  
204 phreatomagmatic outbreak. To summarize, these main features are: 1) the lack of a paleosol  
205 between the phreatomagmatic deposits and dome ones (dome lavas or BAD; Figure 2); 2)  
206 the similar ages between the two facies; 3) the similar mineral cargo charges, and thus a  
207 similar influence on magma viscosity; and 4) the identical phase compositions of phenocrysts

208 that document magma chamber processes, microlites that seem to document conduit  
209 processes, and crystallites that document the very last stages of crystallization at or near the  
210 surface (Figure 5). The strong similarity in crystal proportion and mineral composition down  
211 to the smaller crystallites implies that each of the various generations of crystals crystallized  
212 in similar conditions and from similar melts during the dome and phreatomagmatic stages.  
213 Although the density of dome trachyte lavas varies slightly due to their micro-vesicularity  
214 that developed late during dome emplacement, both dome and phreatomagmatic samples  
215 are relatively dense and incipiently to poorly vesicular according to Houghton & Wilson  
216 (1989) classification (e.g., no pumice texture observed). In the case of the initial  
217 phreatomagmatic stage, this implies that most of the eruption's explosive energy came from  
218 water-magma interaction, and not from 'dry' explosive magma fragmentation. Therefore,  
219 the only difference between the phreatomagmatic and dome-extrusion phases was the  
220 duration of shallow crystallization (or rate of cooling), which was long (slow) for dome lavas  
221 and short (quenched) for phreatomagmatic samples, consistent with the biotite-  
222 destabilization rims and larger crystallites observed in the dome samples. These different  
223 emplacement modes are related to an external parameter: the presence (phreatomagmatic  
224 outbreak stage) or lack (dome stage) of a water table or abundant infiltrated basement fluids  
225 that could trigger phreatomagmatic fragmentation when the magma:water ratio is optimal  
226 (White & Valentine, 2016). Consequently, there is no indication of any difference between  
227 the phreatomagmatic and dome stages prior to the water-magma interaction, and similar  
228 ascent conditions are therefore likely. In that case, most, if not all, of the biotite  
229 destabilization rims are related to destabilization at depths shallower than the melt-water  
230 interactions of the phreatomagmatic stage, and thus very close to the surface or during  
231 dome emplacement. Similar conclusions were reached for some destabilization rims of

232 amphibole at Soufriere Hill, and in the Cascades ([Garcia & Jacobson, 1979](#); [Devine et al.,](#)  
233 [1998](#)). This also implies that a relatively quick conduit ascent should be considered for both  
234 the phreatomagmatic and dome stages with striking implications for risk mitigation.  
235 However, the ascent rate cannot yet be quantified, as only amphibole destabilization  
236 experiments are available; none have been reported for biotite. Once the kinetics of biotite  
237 destabilization upon magma ascent, or at surface conditions will be quantified with future  
238 experiments, the biotite destabilization rims in dome samples may be used to obtain  
239 minimum ascent rates, and an overall cooling rate of the dome lava. As a comparison, a lack  
240 of destabilization rim for amphibole crystals would document an ascent of magmas from  
241 reservoir to surface in less than 4 days, when a 20  $\mu\text{m}$  destabilization rim would develop in a  
242 lava dome at surface conditions in tens of days ([Rutherford & Hill, 1993](#)). According to  
243 [Underwood et al. \(2012\)](#) biotite destabilization is likely to develop faster than amphibole  
244 destabilization due to the higher diffusivity of H in biotite, suggesting that those duration  
245 estimates would be maximum values. Those authors also documented that biotite and  
246 amphibole are more extensively reacted in the core of the Lassen Peak dome (US) where  
247 cooling rates are lower than in its external parts, suggesting that the development of  
248 destabilization rims at surface conditions is not restricted to Sarcoui dome, and should be  
249 seriously considered in any study.

250

### 251 ***General implications and conclusions***

252 The results described herein support a model in which magma ascent conditions in  
253 the conduit were similar for the phreatomagmatic (prior to water-magma interaction) and  
254 dome-extrusion phases at Sarcoui. It is therefore likely that the biotite destabilization rims  
255 observed exclusively in dome samples developed during dome emplacement, and should not

256 be used to quantify trachyte magma ascent rates in the conduit. If those were mistakenly  
257 used for purposes of ascent kinetics estimates, the latter would be greatly underestimated,  
258 implicating serious issues in the development of risk mitigation plans. My results suggest  
259 that the approach of [Rutherford & Hill \(1993\)](#) should be applied with caution when  
260 considering samples that may have cooled slowly (days to weeks) at the surface, e.g., thick  
261 lava flows, domes, and lava lakes. Samples that quenched upon arrival at the surface, i.e.,  
262 juvenile ejecta from explosive eruptions (scoria, pumice, phreatomagmatic bombs),  
263 nevertheless remain perfectly suitable to apply this method. Dome samples are not always  
264 unsuited to this method, but a detailed petrographic study is required to prove that  
265 destabilization rims formed during magma ascent through the conduit and not during dome  
266 emplacement at the surface. To apply [Rutherford & Hill's \(1993\)](#) method to domes, it seems  
267 that the best samples are those that are strongly vesiculated and associated with dilute  
268 pyroclastic density currents (PDCs; e.g., explosive dome eruptions), as they represent the  
269 quenching of fresh magma injected within the dome and the triggering of the explosive  
270 eruption ([Boudon et al. 2015](#)). This was for example the case for strongly vesiculated  
271 pyroclasts from BAD related to PDC from the 1991-1995 Unzen eruption ([Holtz et al., 2005](#);  
272 [Browne et al., 2006](#)). These destabilization rims can confidently be attributed to magma  
273 mixing or destabilization upon ascent, and may be used as chronometers. One striking  
274 difference between Unzen volcano biotite destabilization rims, and the ones studied herein  
275 is the presence of amphibole as a destabilization product at Unzen ([Browne et al., 2006](#)). As  
276 initially suggested by [Feeley & Sharp \(1996\)](#), the presence of amphibole may thus represent  
277 an indicator or rims formed upon ascent by opposition to rims formed at surface during  
278 dome cooling like in the present study where such a hydrous mineral is lacking. This will  
279 nevertheless have to be confirmed by future experimental constraints. Less-vesiculated and

280 rather degassed magmas associated with PDCs may rather represent the slowly cooled and  
281 degassed rigid dome carapace that fragments during the explosive eruption ([Boudon et al.](#)  
282 [2015](#)), and should not be used.

283

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418

419 **Figure captions:**

420 Figure 1: Simplified maps of the study area modified after [France et al. \(2016\)](#). The location  
421 of Chaîne des Puys in France is shown, and the location and geological map of Sarcoui  
422 volcano is presented (from [Boivin et al. 2009](#)). Red arrows and asterisks document the block  
423 and ash deposit location and flow directions. Numbers in rectangles refers to the  
424 determined ages.

425

426 Figure 2: Characteristics of the phreatomagmatic outcrops from the NE quarries of Sarcoui  
427 dome ([Figure 1](#)). The reddish ash and cauliflower bombs characteristic of a phreatomagmatic  
428 deposit (a-b) is overlaid by a block-and-ash deposit (BAD) characteristic of the Sarcoui dome  
429 pyroclastic density currents (PDC). Note in a) the lack of any paleosol in between  
430 phreatomagmatic deposit and the Sarcoui dome related BAD. c) and d) present full thin  
431 section images of accidental clasts that are part of the phreatomagmatic deposit. c) presents  
432 a basalt clast thin section under plane-polarized light; d) presents a gneiss clast under both  
433 plane- and cross-polarized light.

434

435 Figure 3: Microphotographs of trachyte lava from the Sarcoui phreatomagmatic stage  
436 (without biotite reaction rim; left microphotographs), and the Sarcoui dome (with a  
437 noticeable reaction rim around biotite; right microphotographs). Top pictures are under  
438 plane-polarized light, and bottom pictures under cross-polarized light. Bt: biotite, Pl:  
439 plagioclase. Backscattered electron images of the biotite grain from the lava dome sample  
440 are presented in [Figure 4b-c](#).

441

442 Figure 4: Characteristics of the biotite destabilization rims at Sarcoui lava dome. a-k panels  
443 are backscattered electron images, with h-i images obtained by modifying contrast and  
444 colors to better highlight phase relations. An attempt of reaction mass balance is presented  
445 ( $0.5 \text{ biotite} + 0.5 \text{ melt} \Rightarrow 0.63 \text{ K-feldspar} + 0.18 \text{ titanomagnetite} + 0.12 \text{ plagioclase} + 0.04$   
446  $\text{clinopyroxene} + 0.03 \text{ ilmenite} + \text{melt}_2$ ), it highlights that melt is present before and after the  
447 reaction, and that  $\text{melt}_2$  is enriched in both Mg and K, and depleted in Fe. Representative  
448 compositions of the related phases are presented in [Table 1](#). The high Mg# of the Cpx  
449 produced by the reaction is consistent with  $\text{melt}_2$  being enriched in Mg. Panels b-c document  
450 the lava dome biotite grain presented in [Figure 3](#).

451

452 Figure 5: Feldspar compositions in the studied dome and phreatomagmatic bomb (phreato)  
453 samples for Sarcoui dome. Left panel compare all data from the dome samples to all data  
454 from the phreatomagmatic bombs. Right panel compare the composition of phenocrysts,  
455 microlites, and crystallites.

456

457 Figure 6: Evolutionary model for Sarcoui volcano. Magma feeding both the phreatomagmatic  
458 and dome stages followed the same evolution and ascent conditions up to the shallow depth  
459 at which magma interacted with water during the phreatomagmatic stage. The optimal  
460 water:magma ratio triggering phreatomagmatic fragmentation is here schematized by the  
461 representation of a water-rich porous layer named "water table"; see [White & Valentine](#)  
462 [\(2016\)](#) for a discussion on this optimal ratio. The magma was quenched during

463 phreatomagmatism, whereas it slowly cooled during lava dome emplacement, resulting in  
464 extended magma degassing and associated hydrous mineral (biotite) destabilization.

465

466 Table 1: Representative analysis of feldspar, biotite, and of the minerals forming the  
467 breakdown reaction rims of biotite from Sarcoui samples. phreato: phreatomagmatic  
468 bombs; FdK potassic feldspar; ilm: ilmenite; TiMgt: titanomagnetite; Cpx: clinopyroxene;  
469 melt (800SAR2): melt used in the mass balance calculation as a reactant from [Martel et al.](#)  
470 [\(2013\)](#) experiments at 800°C. The buck-rock composition provided by [Martel et al. \(2013\)](#) is  
471 also presented.

472

473 Table S1: Sample list, type, and coordinates.

474

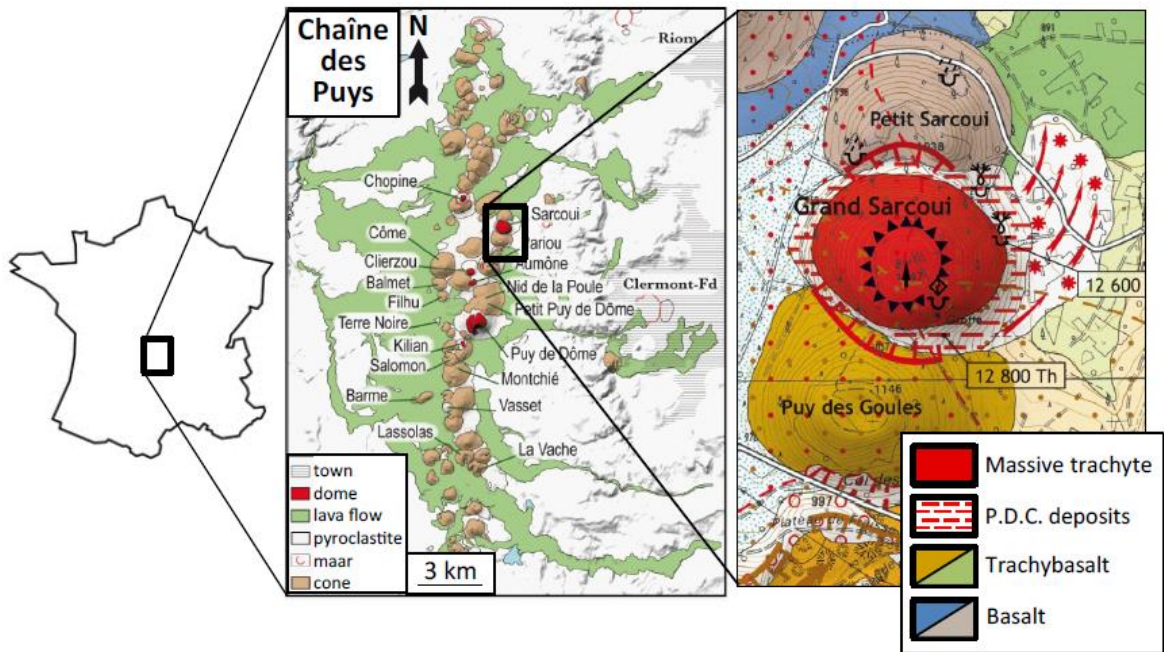


Figure 1

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476



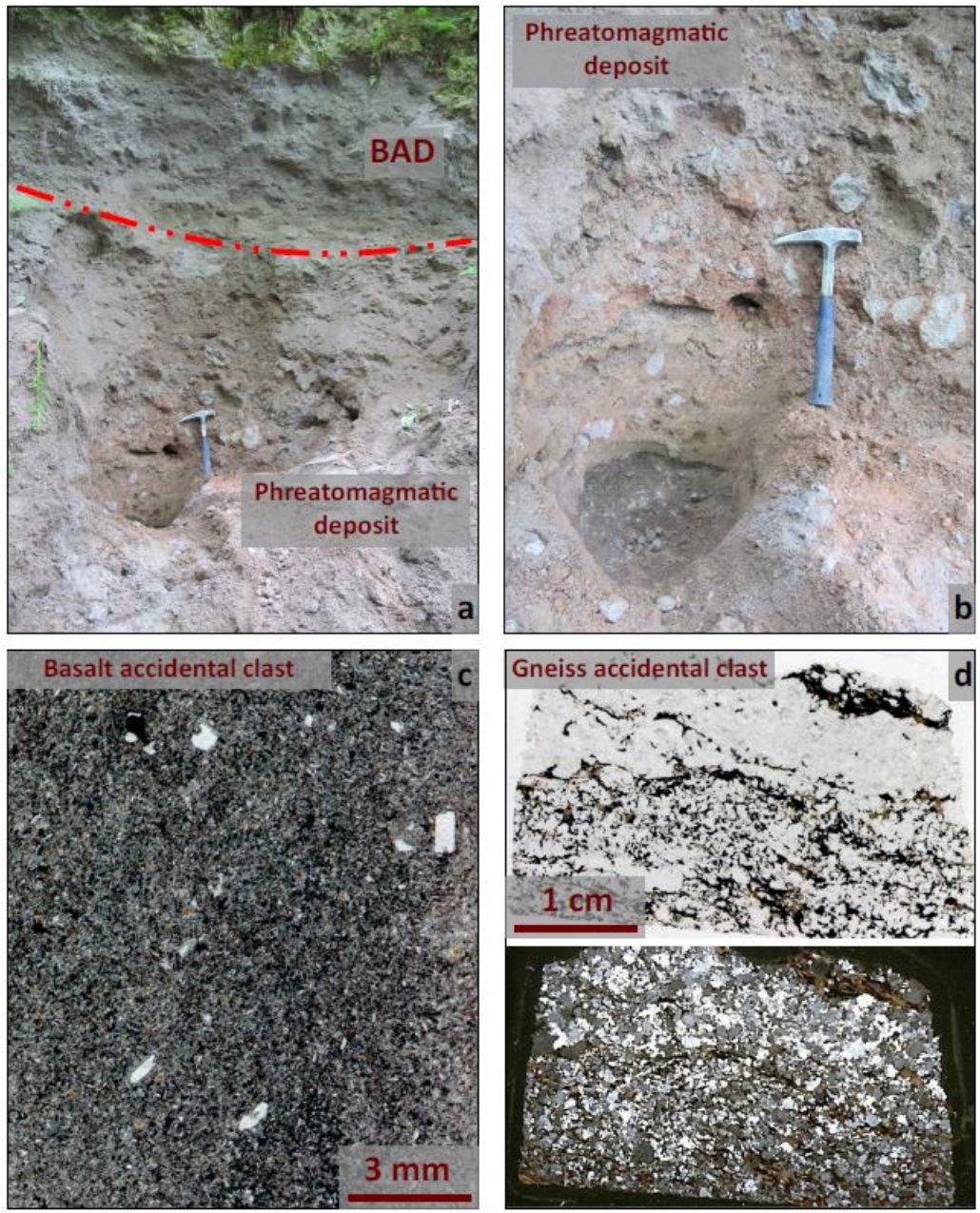
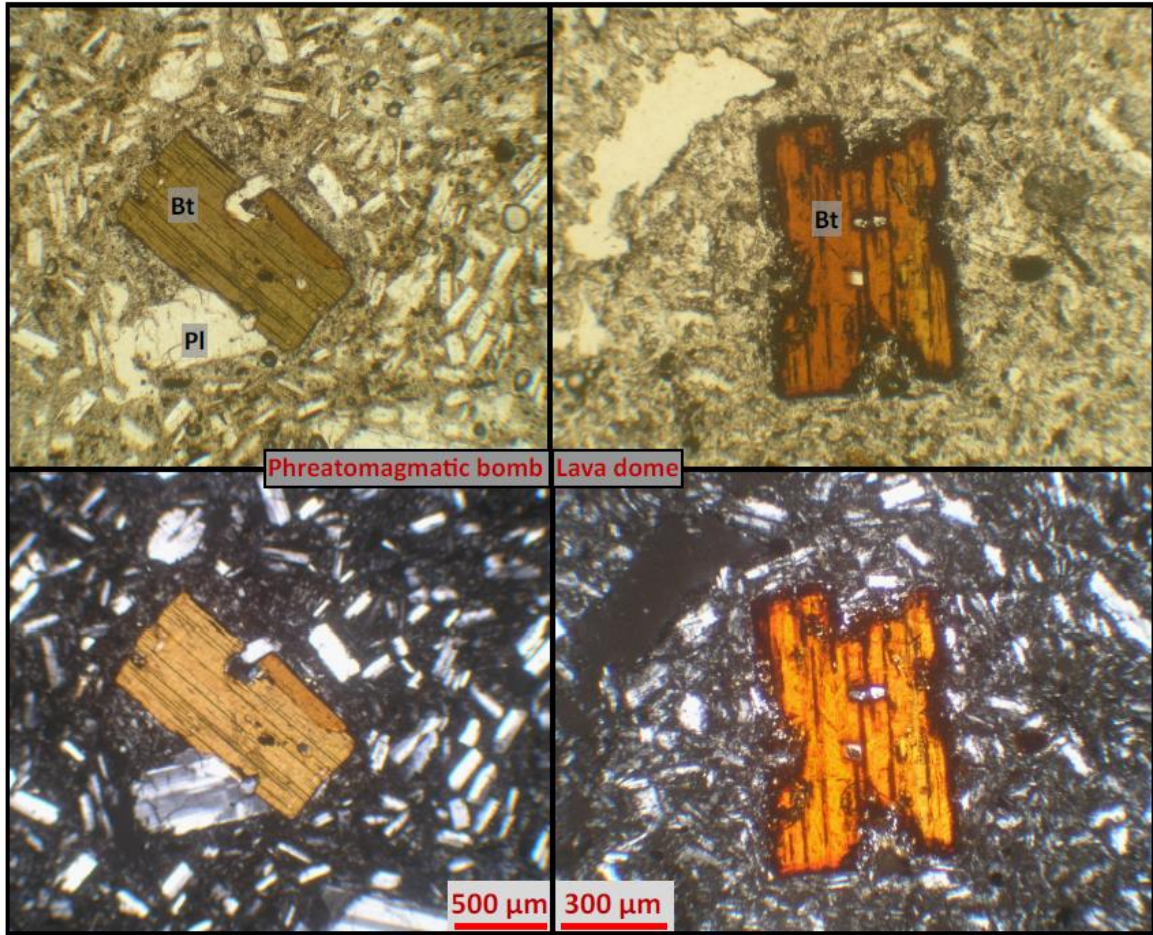


Figure 2

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478



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Figure 3

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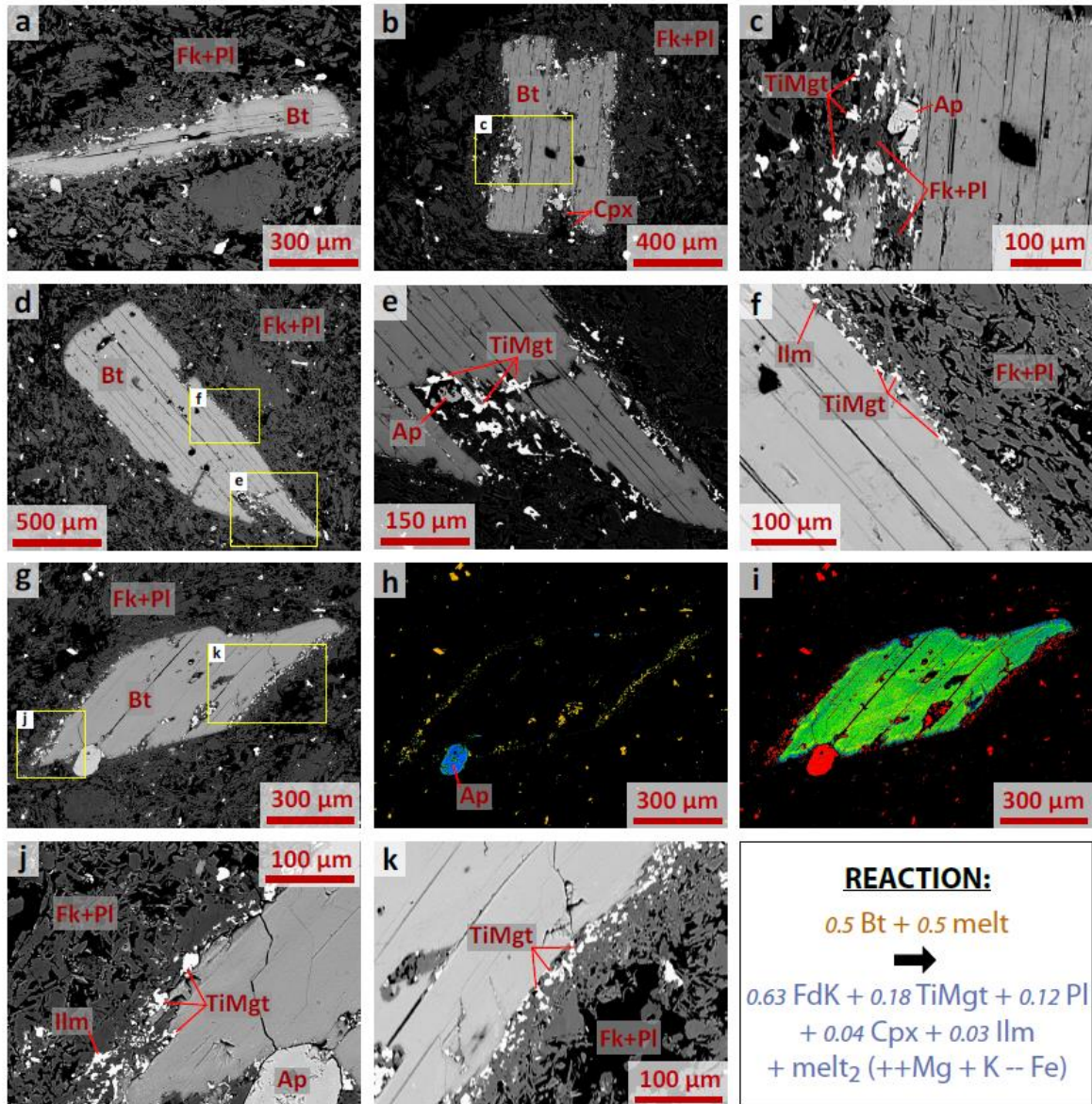


Figure 4

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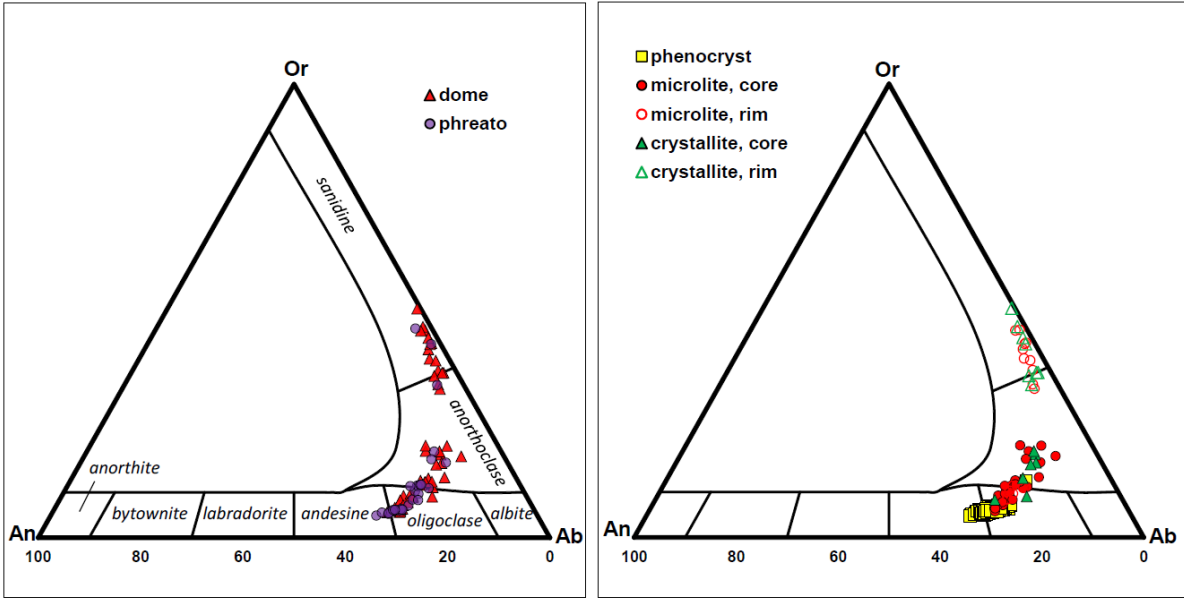


Figure 5

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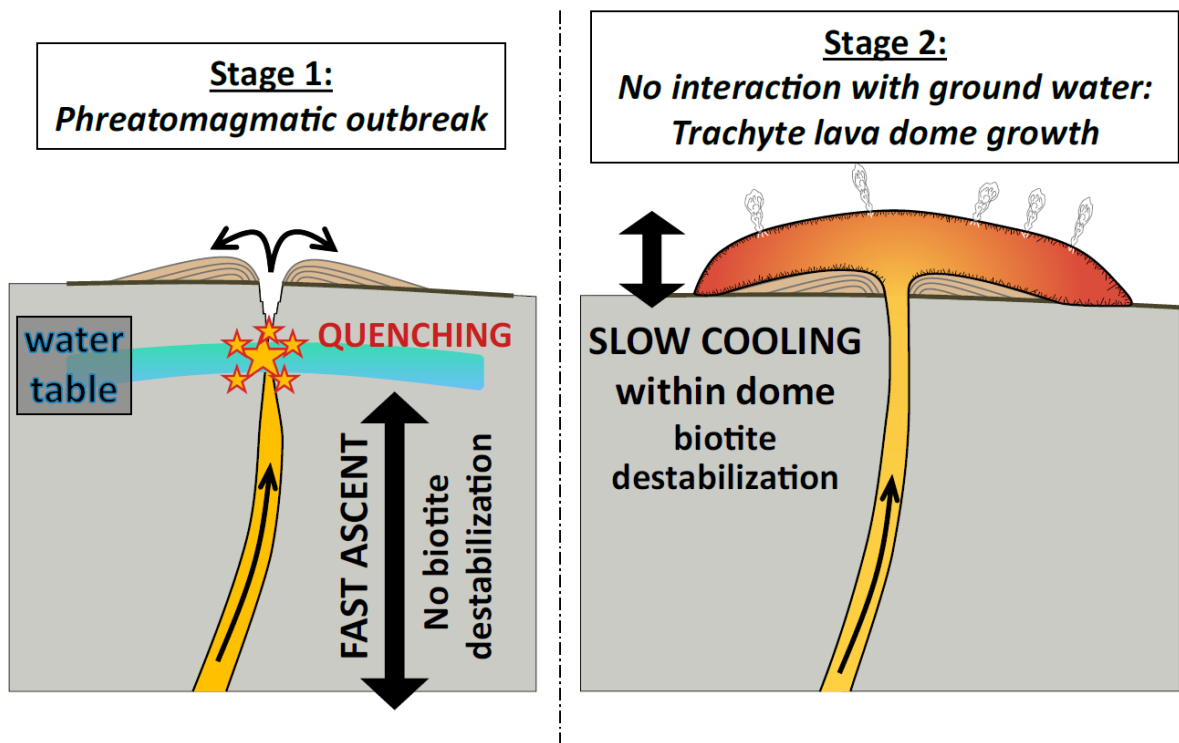


Figure 6

485

486

487 Table 1:

			<i>n</i>	SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	TiO <sub>2</sub>	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	MnO	MgO	FeO	Cl	Total	#Mg	An	Ab	Or		
Feldspar	Dome	phenocryst core	8	62.3	22.9	-	5.0	7.7	1.3	-	0.0	0.3	-	99.4		24.5	67.9	7.6		
		+/-		0.6	0.6		0.8	0.2	0.4		0.0	0.1				3.6	1.5	2.2		
		microlite core	16	63.3	22.0	-	4.0	7.8	2.0	-	0.0	0.3	-	99.4		19.6	69.0	11.4		
		+/-		1.4	1.0		1.2	0.3	0.8		0.0	0.1				5.7	2.1	4.6		
		crystallite core	7	64.3	21.4	-	3.3	7.9	2.4	-	0.0	0.4	-	99.7		16.0	69.7	14.3		
		+/-		1.4	0.8		0.9	0.3	0.7		0.0	0.1				4.4	1.6	4.0		
	phreato	microlite rim		10	66.3	18.7	-	0.6	6.4	6.9	-	0.0	0.4	-	99.4		3.0	56.9	40.0	
			+/-		0.5	0.5		0.2	0.3	0.8		0.0	0.1				1.2	3.2	4.3	
		crystallite rim		6	67.1	18.4	-	0.5	6.3	7.1	-	0.0	0.6	-	100.0		2.3	56.1	41.6	
			+/-		0.3	0.5		0.3	0.6	0.9		0.0	0.1				1.3	4.7	5.8	
		phenocryst core		14	61.5	23.4	-	5.7	7.6	1.0	-	0.0	0.3	-	99.5		27.7	66.5	5.8	
			+/-		0.4	0.4		0.4	0.2	0.1		0.0	0.1				1.7	1.3	0.4	
			microlite core		9	63.5	21.6	-	3.8	7.8	2.2	-	0.0	0.3	-	99.2		18.4	68.9	12.7
				+/-		0.9	0.8		0.8	0.1	0.7		0.0	0.1				4.0	1.2	3.8
			microlite rim		7	63.8	21.6	-	3.4	7.5	2.9	-	0.0	0.3	-	99.5		16.5	66.2	17.3
				+/-		1.2	1.3		1.4	0.8	2.1		0.0	0.1				6.8	6.5	12.8
crystallite rim		2	66.8	18.6	-	0.7	6.4	6.4	-	0.0	0.6	-	99.4		3.5	58.4	38.2			
	+/-		0.1	0.2		0.3	0.0	1.0		0.0	0.1				1.7	2.9	4.5			
Biotite	dome		14	35.8	14.3	5.41	0.0	0.76	8.8	0.59	12.5	17.2	0.17	95.6	56.4					
		+/-		0.5	0.6	0.12	0.0	0.06	0.2	0.15	0.2	0.5	0.10	0.5	0.9					
	phreato	14	35.7	14.0	5.37	0.0	0.77	8.7	0.53	12.6	17.6	0.24	95.6	56.1						
	+/-		0.4	0.3	0.29	0.0	0.05	0.1	0.10	0.2	1.2	0.42	0.8	1.8						
Representative compositions of biotite's rim minerals	FdK		64.9	18.5	0.3	0.6	5.6	6.9	0.1	0.0	1.5		98.4							
	ilm		0.2	1.0	33.5	0.1	0.0	0.1	1.7	3.5	48.6		88.8							
	TiMgt		0.1	1.3	11.2	0.0	0.1	0.1	3.6	4.9	68.5		89.7							
	Cpx		54.9	0.7	0.3	19.4	1.2	0.1	2.9	17.6	3.8		100.8	89.2						
	melt (800SAR2)		66.6	17.3	0.4	1.6	6.8	4.7	0.2	0.3	2.2		100.0							
Bulk-rock (from Martel et al. 2013)				65.5	17.0	0.4	1.7	6.9	4.7	0.2	0.5	2.7	0.04	99.7						

488

489