



HAL
open science

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

Lyderic France

► **To cite this version:**

Lyderic France. Can destabilization rims of hydrous minerals be used to constrain magma ascent kinetics at lava dome volcanoes?. *Bulletin of Volcanology*, 2020, 82 (10), pp.66. 10.1007/s00445-020-01405-4 . hal-02950420

HAL Id: hal-02950420

<https://hal.science/hal-02950420>

Submitted on 27 Sep 2020

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

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

3

4 Bulletin of Volcanology (2020) 82:66

5 <https://doi.org/10.1007/s00445-020-01405-4>

6

7 Lydéric France

8 Université de Lorraine, CNRS, CRPG, F-54000 Nancy, France

9 *Corresponding author: lyderic.france@univ-lorraine.fr

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
49 & Elliott 1993](#); [Rutherford 2008](#); [Druitt et al. 2012](#)). The main chronometric tools available
50 are diffusion chronometry (e.g., [Demouchy et al., 2006](#); [Boudon et al. 2015](#); [Shea et al. 2015](#);
51 [Ferguson et al., 2016](#); [Costa et al., 2020](#)), U-Th disequilibrium series ([Girard et al. 2017](#)), and
52 mineral destabilization chronometry ([Rutherford & Hill 1993](#); [Faure et al., 2001](#); [Browne &
53 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 ± 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 α 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}U - ^{230}Th dating of
148 dome samples give similar ages (12.6 ± 3 ka, and 12.8 ± 1.4 ka, respectively, 2σ 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 (2×0.6 mm) and biotite (~ 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 ~ 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 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% TiO_2 , 14 wt% Al_2O_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 Or_{33} to 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 μm 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 of 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

284 ***Acknowledgments:***

285 Mickael Demacon is thanked for numerous discussions related to the various processes
286 involved in Sarcoui trachyte magma genesis and evolution during his undergraduate project.
287 François Faure is thanked for assistance during field sampling. Georges Boudon is thanked
288 for discussions related to estimating the kinetics of igneous processes in silica-rich systems. I
289 thank the Editor, two anonymous reviewers, and Pavel Plechov for constructive comment
290 that improved the manuscript. This is CRPG contribution n°XXXX.

291 **References:**

- 292 Bachmann O, Bergantz GW (2003) Rejuvenation of the Fish Canyon magma body: A window into the
293 evolution of large-volume silicic magma systems. *Geology*, 31, 789–792.
- 294 Boivin P, Besson JC, Briot D, Camus G, Goer De Herve A, Gourgaud A, Labazuy P, Langlois E,
295 Larouzière FD, Livet M, Mergoil J, Miallier D, Morel JM, Vernet G, Vincent PM, (2009)
296 *Volcanologie de la Chaîne des Puys, Massif Central Français. Parc Naturel Régional des Volcans*
297 *d'Auvergne Clermont-Ferrand*, 196 p, 1/25 000 map.
- 298 Boudon G, Balcone-Boissard H, Villemant B, Morgan DJ (2015) What factors control superficial lava
299 dome explosivity? *Sci Rep* 5: 14551. doi: 10.1038/srep14551
- 300 Browne BL, Gardner JE (2006) The influence of magma ascent path on the texture, mineralogy, and
301 formation of Hornblende reaction rims. *Earth Planet. Sci. Lett.* 246: 161–176.
- 302 Browne BL, Eichelberger JC, Patino LC, Vogel TA, Dehn J, Uto K, Hoshizumi H (2006) Generation of
303 porphyritic and equigranular mafic enclaves during magma recharge events at Unzen volcano,
304 Japan. *J Petrol* 47-2: 301-328. doi:10.1093/petrology/egi076
- 305 Condomines M, Morand P, Camus G, Duthou L (1982) Chronological and geochemical study of lavas
306 from the Chaîne des Puys, Massif Central, France: evidence for crustal contamination. *Contrib*
307 *Mineral Petrol* 81: 296–303.
- 308 Costa F, Shea T, Ubide T (2020) Diffusion chronometry and the timescales of magmatic processes.
309 *Nature Reviews-Earth & Environment* 1: 201-214. doi:10.1038/s43017-020-0038-x
- 310 De Angelis SH, Larsen J, Coombs M, Dunn A, Hayden L (2015) Amphibole reaction rims as a record of
311 pre-eruptive magmatic heating: An experimental approach. *Earth and Planetary Science*
312 *Letters* 426: 235-245

313 De Goër A, Camus G, Gourgaud A, Kieffer G, Mergoil J, Vincent P (1991) *Volcanologie de la Chaîne des*
314 *Puys. Parc Naturel Régional des Volcans d’Auvergne Edition, 128pp.*

315 Demouchy S, Jacobsen SD, Gaillard F, Stern CR (2006) Rapid magma ascent recorded by water
316 diffusion profiles in mantle olivine. *Geology* 34-6: 429-432. doi:10.1130/G22386.1

317 Devine JD, Rutherford MJ, Gardner JE (1998) Petrologic determination of ascent rates for the 1995-
318 1997 Soufriere Hills volcano andesitic magma. *Geophysical Research Letters* 25-19: 3673-3676

319 Druitt TH, Costa F, Deloule E, Dungan M, Scaillet B (2012) Decadal to monthly timescales of magma
320 transfer and reservoir growth at a caldera volcano. *Nature* 482: 77-80.
321 doi:10.1038/nature10706

322 Dzurisin D, Denlinger RP, Rosenbaum JG (1990) Cooling rate and
322 thermal structure determined from progressive magnetization of the Dacite Dome at Mount
323 St. Helens, Washington. *J. Geophys. Res.* 95, 2763. doi:10.1029/JB095iB03p02763

324 Endo ET, Malone SD, Noson LL, Weaver CJ (1981) Locations, magnitudes and statistics of the March
325 20--May 18 earthquake sequence. *USGS Prof. Pap.* 1250: 93-108.

326 Faure F, Trolliard G, Montel JM, Nicollet C (2001) Nano-petrographic investigation of a mafic xenolith
327 (maar de Beaunit, Massif Central, France). *Eur. J. Mineral.* 13: 27-40.

328 Feeley TC, Sharp ZD (1996) Chemical and hydrogen isotope evidence for in situ dehydrogenation of
329 biotite in silicic magma chambers. *Geology* 24-11: 1021-1024.

330 Fergusson DJ, Gonnermann HM, Ruprecht P, Plank T, Hauri EH, Houghton BF, Swanson DA (2016)
331 Magma decompression rates during explosive eruptions of Kilauea volcano, Hawaii, recorded
332 by melt embayments. *Bull Volcanol* 78: 71. doi:10.1007/s00445-016-1064-x

333 France L, Chazot G, Kornprobst J, Dallai L, Vannucci R, Grégoire M, Bertrand H, Boivin P (2015) Mantle
334 refertilization and magmatism in old orogenic regions: the role of late-orogenic pyroxenites.
335 *Lithos* 232: 49–75. <http://dx.doi.org/10.1016/j.lithos.2015.05.017>

336 France L, Demacon M, Gurenko A, Briot D (2016) Oxygen isotopes reveal crustal contamination and a
337 large, still partially molten magma chamber in Chaîne des Puys (French Massif Central). *Lithos*
338 260: 328–338.

339 Garcia MO, Jacobson SS (1979) Crystal clots, amphibole fractionation and the evolution of calc-
340 alkaline magmas. *Contrib Mineral Petrol* 69: 319-327

341 Girard G, Reagan MK, Sims KWW, Thornber CR, Waters CL, Phillips EH (2017) ^{238}U - ^{230}Th - ^{226}Ra - ^{210}Pb -
342 ^{210}Po Disequilibria Constraints on Magma Generation, Ascent, and Degassing during the
343 Ongoing Eruption of Kīlauea. *J Petrol* 58-6: 1199–1226. doi:10.1093/petrology/egx051

344 Gudmundsson A, Lecoeur N, Mohajeri N, Thordarson T (2014) Dike emplacement at Bardarbunga,
345 Iceland, induces unusual stress changes, caldera deformation, and earthquakes. *Bull Volcanol*
346 76: 869.

347 Granet M, Wilson M, Achauer U (1995) Imaging a plume beneath the French Massif Central. *Earth*
348 *Planet Sci Lett* 136: 281–296.

349 Gu X, Deloule E, France L, Ingrin J (2016) Multi-stage metasomatism revealed by trace element and Li
350 isotope distributions in minerals of peridotite xenoliths from Allègre volcano (French Massif
351 Central). *Lithos* 264: 158–174.

352 Gu X, Ingrin J, Deloule E, France L, Xia Q (2018) Metasomatism in the sub-continental lithospheric
353 mantle beneath the south French Massif Central: Constraints from trace elements, Li and H in
354 peridotite minerals. *Chem Geol* 478: 2-17. doi:10.1016/j.chemgeo.2017.08.006

355 Hamelin C, Seitz HM, Barrat JA, Dosso L, Maury RC, Chaussidon M (2009) A low $\delta^7\text{Li}$ lower crustal
356 component: evidence from an alkalic intraplate volcanic series (Chaîne des Puys, French Massif
357 Central). *Chem Geol* 266: 205–217.

358 Hoernle K, Zhang YS, Graham D (1995) Seismic and geochemical evidence for large scale mantle
359 upwelling beneath the eastern Atlantic and western and central Europe. *Nature* 374: 34–39.

360 Holtz F, Sato H, Lewis J, Behrens H, Nakada S (2005) Experimental petrology of the 1991–1995 Unzen
361 dacite, Japan. Part 1: Phase relations, phase composition, and pre-eruptive conditions. *Journal*
362 *of Petrology* 46, 319–337.

363 Houghton BF, Wilson CJN (1989) A vesicularity index for pyroclastic deposits. *Bull Volcanol* 51: 451-
364 462

365 Huber C, Bachmann O, Manga M (2010) Two competing effects of volatiles on heat transfer in
366 crystal-rich magmas: Thermal insulation vs defrosting. *J Petrol* 51: 847–867

367 Juvigne E (1992) Studies on the age of two volcanic lacustrine craters of Auvergne (France). *Comptes*
368 *Rendus Acad Sci Série 2-314*: 401–404.

369 Lustrino M, Wilson M (2007) The Circum-Mediterranean Anorogenic Cenozoic Igneous Province.
370 *Earth Sci Rev* 81: 1–65.

371 Martel C, Champallier R, Prouteau G, Pichavant M, Arbaret L, Balcone-Boissard H, Boudon G, Boivin
372 P, Bourdier JL, Scaillet B (2013) Trachyte phase relations and implication for magma storage
373 conditions in the Chaîne des Puys (French Massif Central). *J Petrol* 54-6: 1071–1107.
374 <http://dx.doi.org/10.1093/petrology/egt006>.

375 Métrich N, Wallace P (2008) Volatile abundances in basaltic magmas and their degassing paths
376 tracked by melt inclusions. In: Putirka K, Tepley F (eds) *Minerals, Inclusions and Volcanic*
377 *Processes*. Mineralogical Society of America, *Reviews in Mineralogy and Geochemistry* 69: 363-
378 402.

379 Miallier D, Condomines M, Pilleyre T, Sanzelle S, Guittet J (2004) Concordant thermoluminescence
380 and ^{238}U – ^{230}Th ages for a trachytic dome (Grand Sarcoui) from the Chaîne des Puys (French
381 Massif Central). *Quat Sci Rev* 23-5/6: 709–715.

382 Miallier D, Pilleyre T, Boivin P, Labazuy P, Gailler LS, Rico J (2017) Grand Sarcoui volcano (Chaîne des
383 Puys, Massif Central, France), a case study for monogenetic trachytic lava domes. *J Volcanol*
384 *Geotherm Res* 345: 125–141. doi: 10.1016/j.jvolgeores.2017.07.015

385 Plechov PY, Tsai AE, Shcherbakov VD, Dirksen OV (2008) Opacitization conditions of hornblende in
386 Bezymyanni volcano andesites (March 30, 1956 Eruption). *Petrology* 16-1: 21-37.

387 Pouchou JL, Pichoir F (1991) Quantitative analysis of homogeneous or stratified microvolumes
388 applying the model “PAP”. In: Heinrich KFJ, Newbury DE (eds) *Electron Probe Quantification*,
389 Plenum, New York, pp 31–75.

390 Rutherford MJ (2008) Magma ascent rates. *Rev Mineral Geochem* 69: 241–271.

391 Rutherford MJ, Hill PM (1993) Magma ascent rates from amphibole breakdown: Experiments and the
392 1980-1986 Mount St. Helens eruptions. *J Geophys Res* 98: 19667-19685.

393 Shea T, Gurioli L, Houghton BF (2012) Transitions between fall phases and pyroclastic density
394 currents during the AD 79 eruption at Vesuvius: building a transient conduit model from the
395 textural and volatile record. *Bull Volcanol* 74: 2363–2381.

396 Shea T, Costa F, Krimer D, Hammer JE (2015) Accuracy of timescales retrieved from diffusion
397 modeling in olivine: A 3D perspective. *Am Mineral* 100: 2026-2042.

398 Sigmundsson F, Hooper A, Hreinsdóttir S, Vogfjörd KS, Ófeigsson BG, Heimisson ER, Dumont S, Parks
399 M, Spaans K, Gudmundsson GB, Drouin V, Árnadóttir T, Jónsdóttir K, Gudmundsson MT,
400 Högnadóttir T, Fridriksdóttir HM, Hensch M, Einarsson P, Magnússon E, Samsonov S,
401 Brandsdóttir B, White RS, Ágústsdóttir T, Greenfield T, Green RG, Hjartardóttir AR, Pedersen R,
402 Bennett RA, Geirsson H, La Femina PC, Björnsson H, Pálsson F, Sturkell E, Bean CJ, Möllhoff M,
403 Braiden AK, Eibl EPS (2014) Segmented lateral dyke growth in a rifting event at Bárðarbunga
404 volcanic system, Iceland. *Nature* 517: 191-195. doi:10.1038/nature14111

405 Spiegelman M, Elliott T (1993) Consequences of melt transport for uranium series disequilibrium in
406 young lavas. *Earth Planet Sci Lett* 118-1/4: 1-20

407 Underwood SJ, Feeley TC, Clyne MA (2012) Hydrogen isotope investigation of amphibole and biotite
408 phenocrysts in silicic magmas erupted at Lassen volcanic center, California. *Journal of*
409 *Volcanology and Geothermal Research* 227-228: 32-49

410 Venezky DY, Rutherford MJ (1999) Petrology and Fe–Ti oxide reequilibration of the 1991 Mount
411 Unzen mixed magma. *Journal of Volcanology and Geothermal Research* 89: 213–230

412 Villemant B, Joron JL, Jaffrezic H, Treuil M, Maury RC, Brousse R (1980) Cristallisation fractionnée
413 d'un magma basaltique alcalin: la série de la Chaîne des Puys (Massif Central, France). II.
414 *Géochimie. Bull Mineral* 103: 267–286.

415 White JDL, Valentine GA (2016) Magmatic versus phreatomagmatic fragmentation: Absence of
416 evidence is not evidence of absence. *Geosphere* 12-5: 1478–1488. doi:10.1130/GES01337.1

417 Zindler A, Hart, SR (1986) Chemical geodynamics. *Annu. Rev. Earth Planet. Sci.*

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 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 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

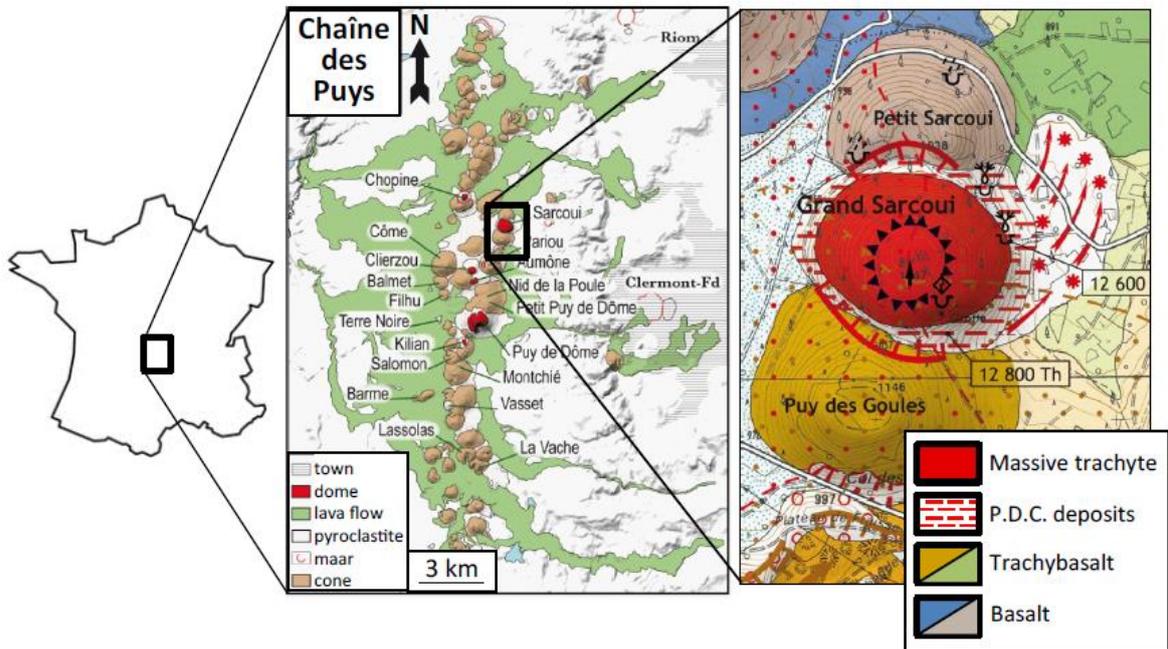


Figure 1

475

476

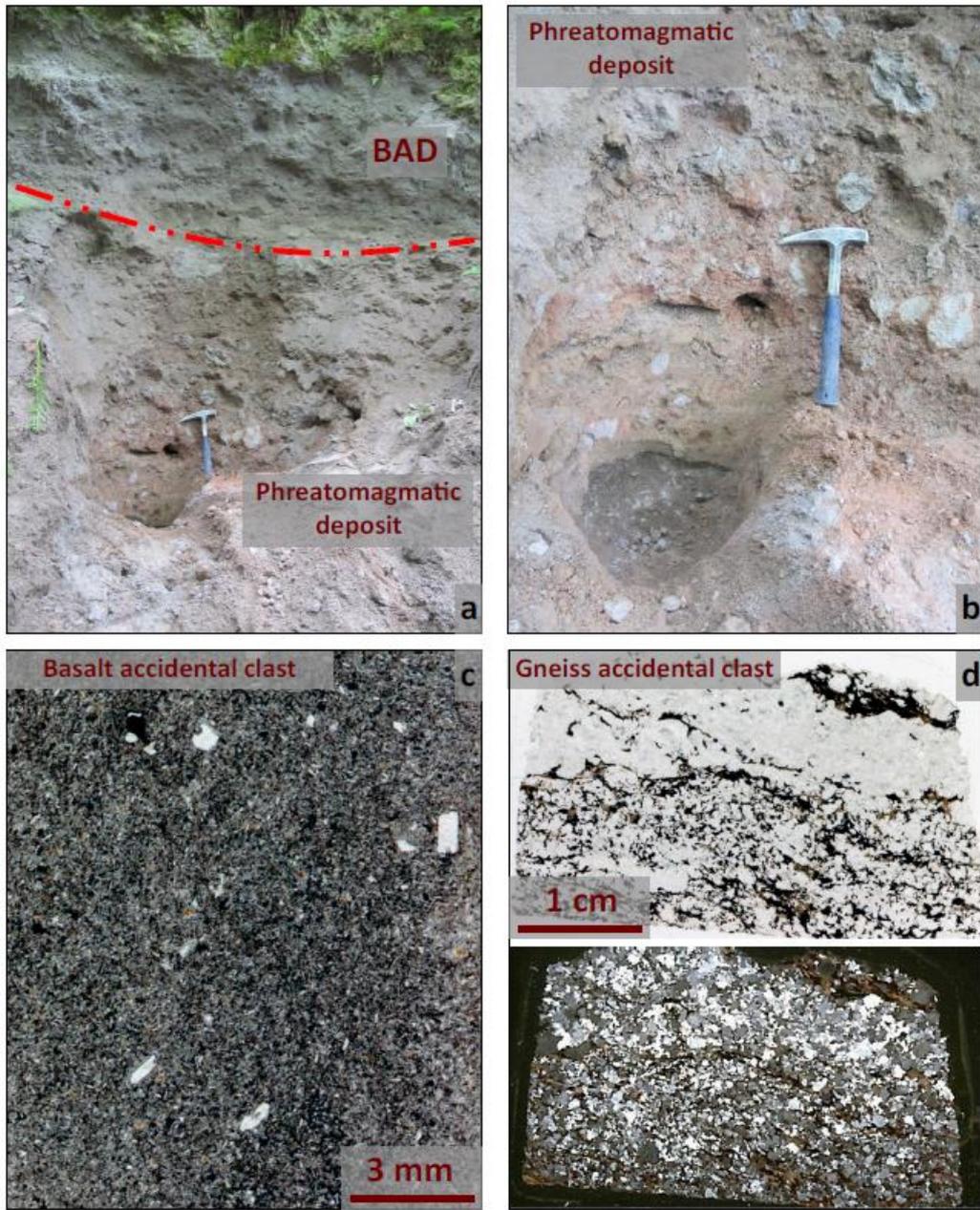
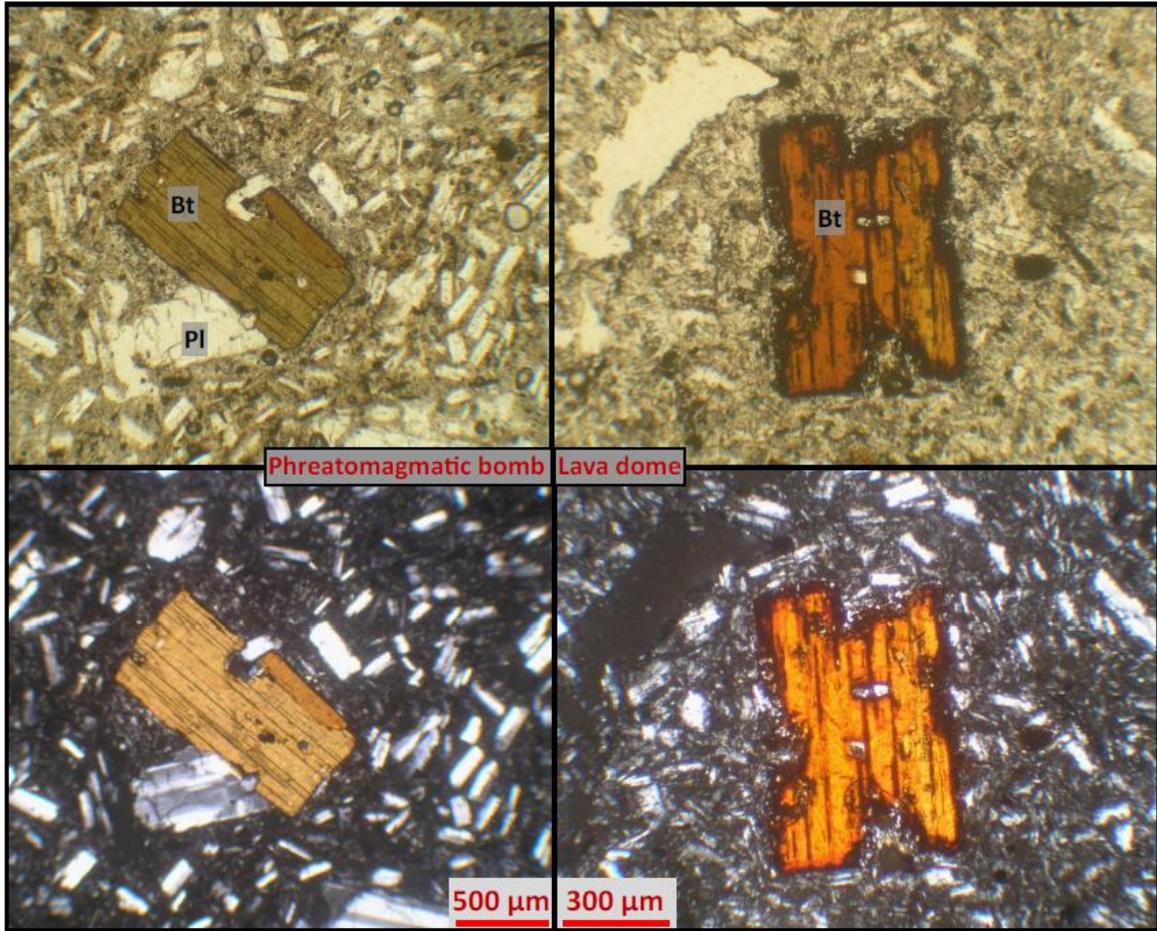


Figure 2

477

478



479

Figure 3

480

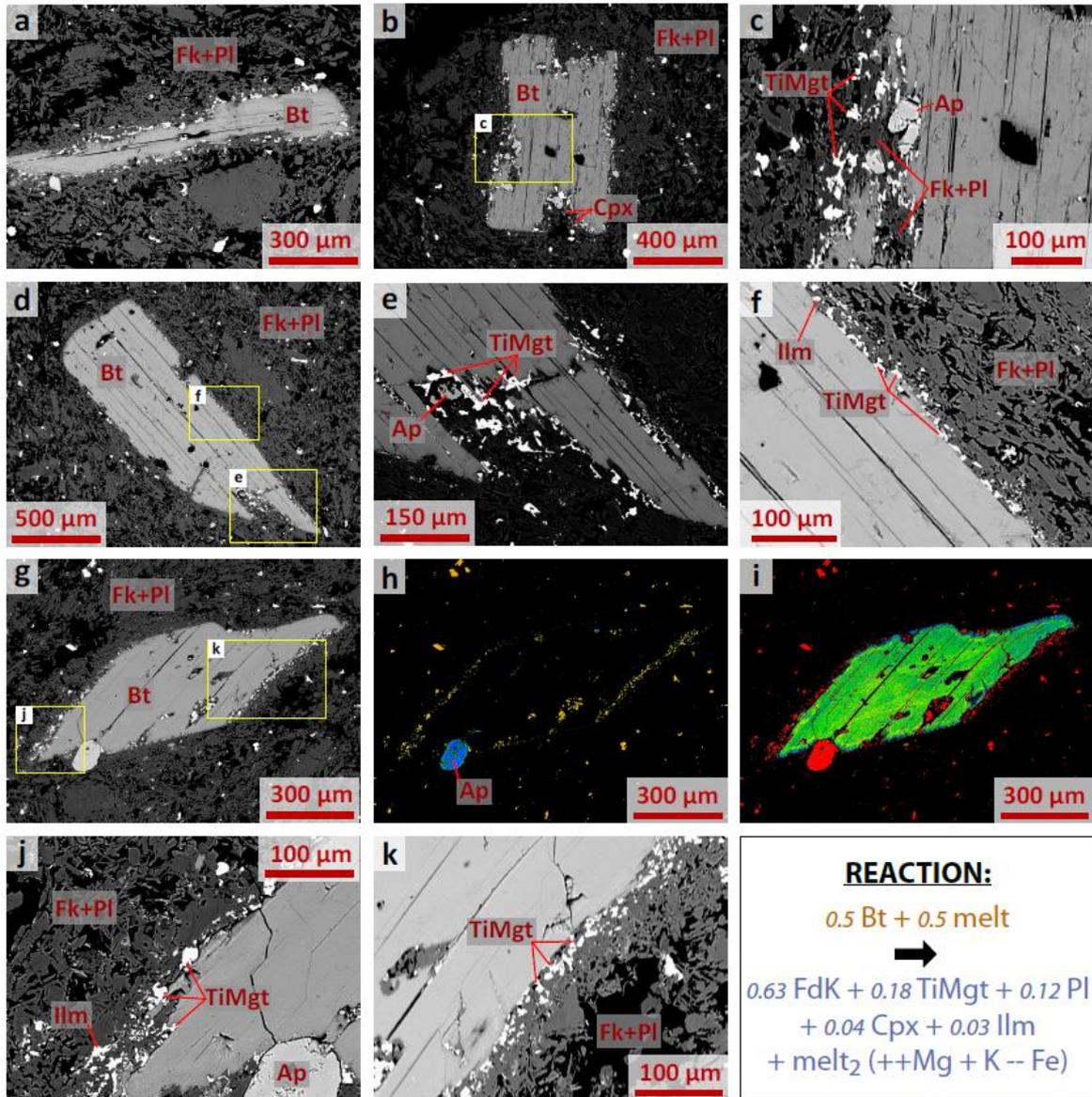


Figure 4

481

482

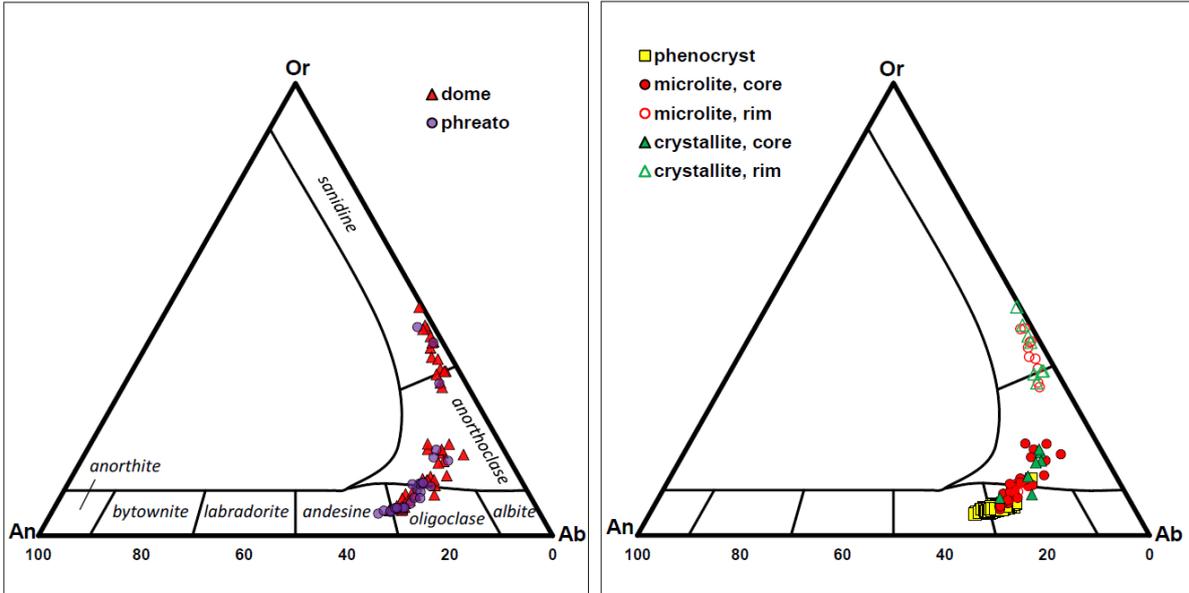


Figure 5

483

484

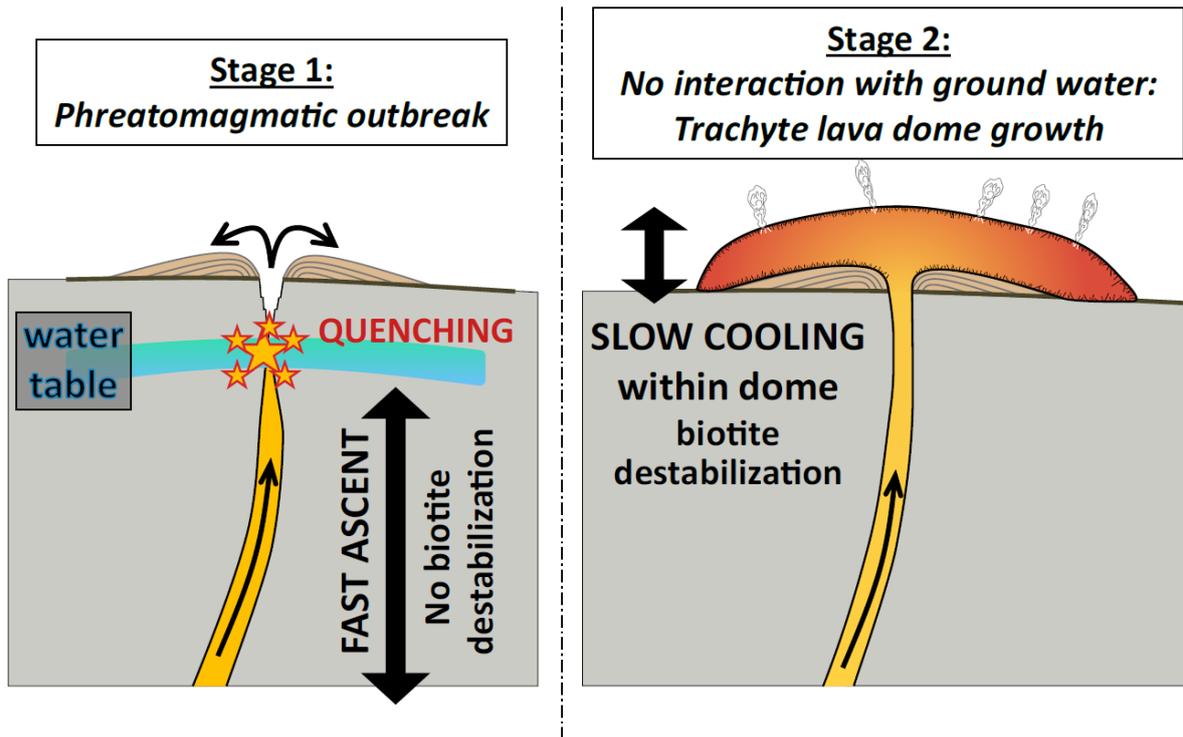


Figure 6

485

486

487 Table 1:

		<i>n</i>	SiO ₂	Al ₂ O ₃	TiO ₂	CaO	Na ₂ O	K ₂ 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