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# Explosive behavior of intermediate magmas: the example of Cotopaxi volcano (Ecuador)

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## **Keypoints**

- Five eruptions occurred at Cotopaxi characterized by a small variation in composition but spanning a wide range of intensity were studied.
- ESPs, textural data and conduit modeling were combined to investigate how magma rheology and ascent dynamics influence eruptive behavior.
- Variabilities in crystal content and magma composition result in feedbacks among crystallization, melt viscosity and volatile exsolution.

24 **Abstract**

25 The variability in intensity and style shown by explosive volcanism has been traditionally  
26 explained by a complex interplay among melt composition and pre-eruptive volatile content,  
27 which modulate magma ascent and conduit dynamics. However, magmas having similar  
28 compositions may be characterized by subtle textural changes affecting magma rheology and  
29 eventually explosive dynamics. Here we study five eruptions occurred at Cotopaxi volcano  
30 (Ecuador) in the last 2000 years characterized by a small variation in magma composition but  
31 spanning a wide range of intensity to investigate how these parameters control variations in  
32 eruptive intensity. We combined eruption source parameters (ESPs), obtained from the application  
33 of recent models to all the available field data, with new textural data and state-of-the-art conduit  
34 dynamics modeling. We found that, despite having variable microlite content and texture, the  
35 effect of microlite on magma rheology is partly counterbalanced by variable phenocryst  
36 abundance, resulting in a comparable total crystal content. The combination of modeling results  
37 with textural data and ESPs suggests that subtle variability in crystal content and magma  
38 composition may be accompanied by strong feedback effects among crystallization, changes in  
39 melt/magma viscosity and volatile exsolution, with microlite crystallization resulting in a rapid  
40 change of magma rheology and modifications in the explosive dynamics. By combining ESPs  
41 with quantitative textural data (i.e. melt normalized vesicle number density) and conduit  
42 modelling, we also show how general observed correlations between composition and texture of  
43 juvenile products with eruption intensity are not evident when applied to eruptions characterized  
44 by a small compositional range.

45

46 **1. Introduction**

47 Explosive eruptions are among the most spectacular and destructive phenomena on Earth. During  
48 magma rise to the surface, variable interconnected processes of degassing, bubble nucleation and  
49 growth, as well as degassing-induced crystallization, force the magma to rapidly change its  
50 rheology and eventually fragment. Once fragmentation occurs, the mixture of gas and pyroclasts is  
51 ejected at high velocity in the atmosphere forming convective volcanic plumes and/or feeding  
52 lateral pyroclastic density currents (PDCs). Recent examples of volcanic eruptions of different  
53 magnitude, such as those of Tungurahua (Ecuador, 1999), Chaitén (Chile, 2008), Eyjafjallajökull  
54 (Iceland, 2010), Cordón Caulle (Chile, 2011), Kelud and Sinabung (Indonesia, 2013-14), and  
55 Fuego (Guatemala, 2018), clearly demonstrate the variable aspects of explosive volcanism, often  
56 resulting in devastating consequences for ecosystems and population living nearby active  
57 volcanoes (e.g. Craig et al., 2016; Elissondo et al., 2016; Few et al., 2017; Martin et al., 2009;  
58 Mazzocchi et al., 2010; McCausland et al., 2019).

59 The large variability in intensity and style of explosive eruptions is primarily controlled by magma  
60 ascent dynamics, which modulate the Eruption Source Parameters (ESPs; e.g., exit velocity, mass  
61 eruption rate, column height), and is in turn influenced by the melt composition and pre-eruptive  
62 volatile content (Jaupart, 1996; Huppert, 2000; Gonnermann and Manga, 2007; Cashman, 2004).  
63 Basaltic explosive activity typically takes the form of low-intensity Hawaiian and Strombolian  
64 eruptions (Houghton and Gonnermann, 2008), but also sporadic, transient, higher intensity phases  
65 have been observed or described (e.g. Williams, 1983; Walker et al., 1984; McPhie et al., 1990;  
66 Rosi et al., 2006; Höskuldsson et al., 2007; Scollo et al., 2007; Coltelli et al., 1998; Costantini et  
67 al., 2009; Perez et al., 2009). Increasing magma evolution coupled with higher volatile contents is  
68 commonly associated with higher intensity activity which takes the form of sub-Plinian to Plinian  
69 eruptions, depending on their steadiness (Cioni et al., 2015). Particularly, Plinian and sub-Plinian

70 eruptions from andesitic eruptive centers consist of multiphase eruptive episodes with shifting  
71 styles, producing complex pyroclastic successions, which have been described, among others, at  
72 Colima and Nevado de Toluca (Mexico), Somma-Vesuvius (Italy), and Taranaki and Ruapehu  
73 (New Zealand) volcanoes (e.g. Arce et al., 2003, 2005; Cioni et al., 2008; Saucedo et al., 2010;  
74 Pardo et al., 2012; Macías et al., 2017; Torres-Orozco et al., 2018).

75 Important progress has been made in understanding magma ascent and eruption dynamics during  
76 the last decades (Sparks, 1986; Vergnolle and Jaupart, 1986; Carey and Sigurdsson, 1989; Klug  
77 and Cashman, 1996; Melnik et al., 2005; Gonnermann and Manga, 2007), but several key  
78 questions still arise from the observation of the large variability in the eruptive style and explosive  
79 dynamics of magmas with comparable characteristics (i.e. bulk chemistry and volatile and crystal  
80 content). Such variability in eruptive style, which can even take the form of effusive-explosive  
81 transitions also in the absence of important changes in magma composition, has been related to  
82 many complex sub-surface processes such as increase in magma vesicularity, decompression-  
83 induced crystallization (predominantly microlites) and related viscosity increase, progressive loss  
84 of the exsolved fluid phase, magma shearing in the conduit, sintering, and viscous dissipation at  
85 conduit walls (Stevenson et al., 1996; Manga et al., 1998; Hammer et al., 2000; Rust et al., 2003;  
86 Polacci et al., 2001; Schipper et al., 2013; Tuffen et al., 2013; Cassidy et al., 2018; Wadsworth et  
87 al., 2020). How these processes interact with each other and to what extent they affect eruptive  
88 dynamics is however still an open question, particularly crucial for magmas having compositions  
89 spanning from basalts to andesites, for which subtle textural or compositional changes of the melt  
90 phase, often occurring syn-eruptively, may have primary effects on rheology and explosive  
91 dynamics (Mader et al., 2013; Lindoo et al., 2017; Arzilli et al. 2019a).

92 The last 2000 years of explosive activity of Cotopaxi, an andesitic central volcano situated in the  
93 Eastern Cordillera of the Ecuadorian Andes, offer the unique opportunity to investigate pyroclastic  
94 products of moderate- to high-intensity eruptions resulting from the ascent of andesitic magma  
95 with a restricted silica range (Barberi et al., 1995; Costantini, 2010; Pistolesi et al., 2011). Despite  
96 the similar composition (from andesite to dacite; 56 to 62 wt. % SiO<sub>2</sub>), well-documented eruptions  
97 of this volcano present a large variability in ESPs, spanning more than one order of magnitude in  
98 mass eruption rate (MER). In this study, we selected five eruptions to discuss the factors  
99 controlling eruption dynamics in magmas of intermediate composition. ESPs of the considered  
100 Cotopaxi eruptions have been updated as part of this investigation based on the most recent  
101 available models, in order to have a comprehensive and homogenous characterization of the  
102 eruption dynamics associated with the different selected events. We thus investigated how  
103 composition, microtextural characteristics, volatile disequilibrium during magma ascent,  
104 degassing efficiency, and conduit dynamics may affect such explosive behavior.

105

## 106 **2. Geological background and targeted eruptions**

107 The volcanic activity in Ecuador is distributed both trench-ward and behind the volcanic arc and is  
108 related to the subduction of the oceanic Nazca plate carrying the aseismic Carnegie Ridge,  
109 produced by the passage of the plate over the Galapagos hotspot. The surface volcanism results in  
110 a broad (up to 110 km) volcanic arc (Bourdon et al., 2003) and consists of three different volcanic  
111 chains (i.e., the Western and the Eastern Cordilleras, and the Andean foothill).

112 Cotopaxi volcano, an ice-capped, 5897 m-high perfectly symmetrical cone, lies within the Inter-  
113 Andean valley, a structural depression within the two cordilleras. Alexander von Humboldt,  
114 explorer, geographer and naturalist, was the first European who tried to reach Cotopaxi's summit

115 during the five years spent in South America, from 1799 to 1804. He reached an altitude of 4500  
116 m, and only in 1872 the crater top was reached by the German geologist Wilhelm Reiss.

117 The volcano started its formation ~560 ka ago with the construction of an ancient stratovolcano  
118 (Paleocotopaxi), whose activity was characterized by large explosive events and deposition of  
119 rhyolitic tephra deposits (e.g., Barberi et al., 1995). After a period of potential rest, the volcanic  
120 activity resumed 100–150 ky ago (Barberi et al., 1995; Hall, 1977; Hall and Mothes, 2008) and  
121 was interrupted by a large flank failure ~4500 yr BP (Barberi et al., 1995; Hall, 1977; Hall and  
122 Mothes, 2008; Smyth and Clapperton, 1986). The scar is presently completely filled by younger  
123 products and a hummocky topography is the only remaining indication of this giant debris  
124 avalanche.

125 While historical chronicles concerning Cotopaxi activity are available starting from the time of the  
126 Spanish conquest, geological descriptions of Cotopaxi activity date back to the eighteenth century  
127 in a series of scientific monographs and works by La Condamine (1751), von Humboldt (1837–  
128 1838), Reiss (1874), Sodiro (1877), Stübel (1897), Whympfer (1892), Wolf (1878, 1904), Reiss  
129 and Stübel (1869–1902), Hradecka et al. (1974), Miller et al. (1978), Hall (1987), Hall and von  
130 Hillebrandt (1988), and Mothes (1992). More recently, Barberi et al. (1995), Hall and Mothes  
131 (2008) and Pistolesi et al. (2011) investigated and simplified the scheme of the last andesitic  
132 eruptive cycle after the ~4500 yr BP flank failure, during which multiple scoria and pumice falls,  
133 lava and pyroclastic flows contributed to the formation of the present edifice (Table S1).

134 Pistolesi et al. (2011) conducted a detailed stratigraphic study related to eruptive products post-XII  
135 century which, combining field data with historical chronicles and radiocarbon ages, and  
136 highlighting the presence of 21 continuous tephra units. All these units, organized in plane-parallel  
137 sequences and separated by erosive surfaces, were characterized both physically and

138 compositionally. The chronostratigraphic scheme covers a time window bracketed between the  
139 emplacement of the Quilotoa co-ignimbrite ash, a regional marker dated at AD 1150 (Di Muro et  
140 al., 2008; Mothes and Hall, 2008), and the last important Cotopaxi explosive event occurred in  
141 1877, before the recent reactivation of 2015 (Gaunt et al., 2016). Barberi et al. (1995) studied the  
142 main eruptions related to a longer period of activity, starting from the sector collapse episode at  
143 ~4500 yr BP, and identified the deposits of at least 15 older eruptions (from Plinian to sub-Plinian)  
144 below the tephra Layers detailed in Pistolesi et al. (2011), including therefore eruptions over a  
145 period of time between about 2000 and 800 years ago. Hall and Mothes (2008) furtherly detailed  
146 the stratigraphic framework proposed by Barberi et al. (1995), also recognizing local deposits of  
147 lava flows and other minor pyroclastic deposits between the main eruptions.

148 Based on the chrono-stratigraphy and composition of the products of the recent period, Pistolesi et  
149 al. (2011) suggested that during the past eight centuries the volcanic activity was not regularly  
150 spaced over time. Between the thirteenth and eighteenth centuries, the activity was in fact  
151 characterized by isolated Plinian and sub-Plinian episodes followed by long phases of substantial  
152 degassing and suggestive of a period of low magma input rate. In the eighteenth century, the  
153 system was fueled by a significant input of volatile-rich, mafic magma, resulting in high intensity  
154 Plinian eruptions (1742-44 and 1766-68) followed by several short-lived events defining clusters  
155 of eruptions, and in a significant increase in the average eruptive rate. By revising the first-order  
156 approximation proposed by Barberi et al. (1995), who obtained an average recurrence time of 117  
157 years dividing the time lapse of 2000 yr by the number of tephra beds counted in the same  
158 interval, Pistolesi et al. (2011) proposed that Cotopaxi was characterized by an uneven magma  
159 input rate also in the last 2000 years of activity, with periods of very frequent, low- to high-

160 intensity activity, and scattered, mid- to high-intensity eruptions separated by longer repose  
161 intervals.

162 The five targeted eruptions of this study cover the last ~1000 years of activity and include Plinian  
163 and sub-Plinian events including a limited range of magma compositions. Starting from the oldest  
164 studied event, we selected the eruptions of Layers 5, 3, 2 and 1 (according to the nomenclature in  
165 Barberi et al., 1995) and the last event occurred in 1877 (layer P<sub>E</sub> of Pistolesi et al., 2011).

166 Layer 5 is a fallout deposit of black scoriaceous lapilli bearing abundant lithic fragments of gray  
167 lava. Slightly younger than 1180 yr BP, Layer 5 is related to a Plinian event of andesitic  
168 composition (57.9 wt. % SiO<sub>2</sub>). Layer 3 represents the largest Plinian pumice fallout deposit of the  
169 last 2000 years of Cotopaxi activity (Barberi et al., 1995). Dated at 820±80 yr BP, in agreement  
170 with the presence of the Quilotoa ash immediately above, the deposit is represented by a well-  
171 sorted, symmetrically-graded bed of pumice. In the upper part, a characteristic 10 cm-thick bed  
172 richer in dark lithics and lava chips is ubiquitous. Slightly more evolved than Layer 5 (62.3 wt. %  
173 SiO<sub>2</sub>), Layer 3 is the result of a Plinian eruption. Layers 1 and 2 (M<sub>B</sub> and M<sub>T</sub> in Hall and Mothes,  
174 2008 and Pistolesi et al., 2011) form a pair of black and white tephra layers traceable around the  
175 volcano resulting from Plinian activity. Interlayered stream sediments and debris-flow deposits at  
176 valley sites were scatteredly observed between these layers, as well as sporadic small pockets of  
177 organic material at the base of Layer 1. These are suggestive that the two beds were probably  
178 emplaced in a close time interval, even though they have a slightly different dispersal. Multiple  
179 tongues of loose, coarse-grained tephra deposits mainly composed of dark, cauliflower scoria  
180 bombs, which are several meters wide and up to 3 m thick and locally welded, are often found  
181 around the cone, and have been interpreted as scoria flows derived from boiling over activity  
182 during this Plinian eruption (Layer 1). Layers 2 and 1 were related to the events occurred in AD

183 1742–1744 and AD 1766–1768, respectively, and were associated with catastrophic lahars  
184 resulting from the rapid melting of the summit glacier. Both are andesitic in composition, with  
185 Layer 2 slightly more evolved. The last high-intensity, sub-Plinian eruption at Cotopaxi occurred  
186 in 1877 (Layer P<sub>E</sub> in Pistolesi et al., 2011), fed by andesitic magma as well (58.8 wt. % SiO<sub>2</sub>).  
187 Stratigraphic position and dispersal are well in agreement with detailed historical descriptions  
188 given by Wolf (1878) and Sodiro (1877) for this eruption. Scoria flows generation that  
189 accompanied the eruption was responsible for ice melting and consequent generation of  
190 destructive lahars, also well reported in the contemporary chronicles.

191

### 192 **3. Methods**

193 In order to study the five selected Cotopaxi eruptions, we adopted a strategy based on the  
194 integration of newly estimated ESPs (e.g., total erupted volume, column height, MER),  
195 geochemical and textural information, and conduit modeling. All the products selected for the  
196 textural studies and the physical volcanology data were collected during several field surveys  
197 carried out between 2005 and 2008. ESPs were re-calculated based on new available models for  
198 data treatment which were not available at the time of previous elaborations (e.g. Barberi et al.,  
199 1995; Pistolesi et al., 2011). New textural data of the erupted products (density, bubble size  
200 distribution and crystal content) are also included, while geochemical information is derived from  
201 the literature. All these data sources were used to constrain the input parameters of conduit  
202 simulations in order to investigate how differences in rheology, gas exsolution and crystallization  
203 may affect fragmentation and eruption dynamics. In this section, we summarize the different  
204 strategies and assumptions used to calculate the ESPs, measure textural data, and develop  
205 numerical simulations of conduit dynamics.

206

### 207 **3.1 Eruption Source Parameters**

208 In order to provide a comprehensive and homogeneous characterization of the five selected  
209 Cotopaxi eruptions the most recent models for the determination of ESPs have been applied and  
210 their results compared (Table 1).

211 First, the models of Pyle (1989), Fierstein and Nathenson (1992), and Bonadonna and Houghton  
212 (2005) had already been applied by Pistolesi et al. (2011) and Biass and Bonadonna (2011) for the  
213 determination of the tephra deposit volume. These models are based, respectively, on the  
214 integration of the Exponential applied to 1 segment (for Eruption 1877, Layers 1, 2 and 3),  
215 Exponential applied to 2 segments (for Layer 5), and Power-Law best fit of tephra-deposit  
216 thickness versus distance from the vent expressed as the square root of the area enclosed by the  
217 associated isopach contours (Fig. 1). The Power-Law limits of integration were set at 100 km  
218 (Eruption 1877, Layer 1 and Layer 2; Pistolesi et al., 2011), 100-500 km (Layer 3; Biass and  
219 Bonadonna, 2011), and 150 km (Layer 5; Biass et al., 2019). Here we have also applied the more  
220 recent model of Bonadonna and Costa (2012) that integrates the tephra-deposit volume based on  
221 the Weibull best fit (Table 1).

222 Second, the methods of Carey and Sparks (1986) and Rossi et al. (2019) have been used for the  
223 determination of plume height from the distribution of the largest lithics around the volcano (Fig.  
224 1). In particular, the model of Rossi et al. (2019) builds on the model of Carey and Sparks (1986),  
225 commonly used for the determination of plume height, by implementing additional key aspects of  
226 plume dynamics, cloud spreading and particle sedimentation, such as the effect of wind advection  
227 on the buoyant plume. In fact, plumes that are bent by the action of wind have the potential to  
228 sediment a given clast size further from the vent than vertical plumes characterized by the same

229 height. Depending on wind and eruptive conditions, the model of Rossi et al. (2019) might,  
230 therefore, return lower values of plume height with respect to the model of Carey and Sparks  
231 (1986) that only considers wind advection of particles settling in the atmosphere. The results of  
232 the models of Carey and Sparks (1986) and Rossi et al. (2019) were averaged over all lithic  
233 contours associated with the average of the 3 axis of the 5 largest clasts (i.e. 3.2 cm and 1.6 cm for  
234 Eruption 1877; 6.4 cm, 3.2 cm and 1.6 cm for Layer 1; 3.2 cm, 1.6 cm and 0.8 cm for Layer 2; 3.2  
235 cm, 1.6 cm and 0.8 cm for Layer 3; 6.4 cm, 3.2 cm, 1.6 cm and 0.8 cm for Layer 5; Biass and  
236 Bonadonna, 2011). The determination of the plume height with the model of Carey and Sparks  
237 (1986) was carried out using the Matlab script of Biass et al. (2015) based on the crosswind and  
238 downwind ranges taken from the isopleth maps of Pistolesi et al. (2011) (Eruption 1877 and  
239 Layers 1 and 2) and of Biass and Bonadonna (2011) (Layers 3 and 5).

240 Finally, the strategies of Wilson and Walker (1987), Mastin et al. (2009) and Degruyter and  
241 Bonadonna (2012) have been used to determine the MER. In particular, the model of Degruyter  
242 and Bonadonna (2012) also accounts for wind advection of the buoyant plume, typically resulting  
243 in higher values of MER with respect to the models of Wilson and Walker (1987) and Mastin et  
244 al. (2009) in case of strong wind speeds. The equation of Wilson and Walker (1987) for the  
245 determination of MER was applied considering an empirical normalization constant of 0.295 as  
246 suggested by Pistolesi et al. (2011) and Biass and Bonadonna (2011) for silica-poor magmas. The  
247 application of the theoretical equation of Degruyter and Bonadonna (2012) requires values of  
248 magmatic temperature taken as 1223 K for Layer 3 (andesite) and 1273 K for Eruption 1877 and  
249 Layers 1, 2 and 5 (basaltic andesite). Additionally, the tropopause height was fixed at 17 km  
250 above sea level and wind is averaged across plume height considering the maximum value as  
251 derived using the model of Rossi et al. (2019) and a linear decay profile to sea level as suggested

252 by Carey and Sparks (1986). It is also important to note that the vent height is taken at 5.9 km,  
253 which is especially relevant to convert to height above the vent the results derived from both the  
254 models of Carey and Sparks (1986) and Rossi et al. (2019), which are in turn expressed in height  
255 above the sampling level, averaged to be about 3 km above sea level. Such a high vent is also  
256 associated with a low atmospheric density and temperature ( $0.7028 \text{ kg/m}^3$  and  $249.7 \text{ K}$ ,  
257 respectively), which largely impact the calculation of MER with the equation of Degruyter and  
258 Bonadonna (2012).

259 Finally, the minimum duration of the eruptions was calculated based on the combination between  
260 the peak value of MER and the Erupted Mass obtained from the average volumes of the deposits,  
261 computed considering a deposit density of  $700 \text{ kg/m}^3$  for Eruption 1877 and Layers 2 and 3, and  
262  $950 \text{ kg/m}^3$  for Layers 1 and 5 (Biass and Bonadonna, 2011).

263

### 264 **3.2 Textural data**

265 Density of 100 randomly picked clasts (pumices or scoriae) larger than 16 mm for each layer was  
266 measured in air and water after sealing using the method of Houghton and Wilson (1989). Density  
267 data were then converted into vesicularity values using the bulk density of the samples, measured  
268 on the sample powders with a helium pycnometer. Data were then plotted as histograms and used  
269 for selecting representative clasts for vesicles and crystal image analysis.

270 Vesicles in pyroclastic products are the result of the complex overlapping of degassing,  
271 nucleation, expansion and eventually coalescence processes in magmas (Sparks, 1978; Cashman  
272 et al., 1994) and for this reason parameters such as size, spatial arrangements, shape as well as  
273 vesicle size distribution (VSD) and number density can be used to make inferences on the physical

274 processes that influence the magma ascent dynamics along the volcanic conduit (Toramaru, 2006;  
275 Cashman et al., 1994; Shea et al., 2010).

276 VSD and vesicle number density data were estimated using 1x photo scans and Scanning Electron  
277 Microscope Back-Scattered Electrons (SEM-BSE) images of polished sections of 1 to 3 clasts of  
278 pumice and scoria from the fallout deposits of the five targeted eruptions, representative of the  
279 average, lowest and highest densities measured in these pyroclastic products. SEM-BSE images  
280 were collected at different magnifications (25×, 100×, 250× and 500×) using electron microscopy  
281 facilities in labs of the University of Geneva and MEMA center (University of Florence). Images  
282 were then analyzed using the software ImageJ (Schneider et al., 2012) to obtain the complete size  
283 distribution of the entire vesicle population. The different minerals (both phenocrysts and  
284 microlites) and vesicles recognized on the images were segmented and saved on different binary  
285 images, and their shape, size and area measured. Vesicles showing clear signs of coalescence were  
286 manually de-coalesced using the vestiges of thin septa still present along the vesicle margins, in  
287 order to obtain the original dimensional parameters of each vesicle. Data obtained from images  
288 with different magnifications were renormalized to the total area investigated in the 1× scans. 2D  
289 data from image analysis were finally reconverted to 3D using the stereological model proposed  
290 by Sahagian and Proussevitch (1998), which allows to derive  $N_V$  (number of vesicles per unit  
291 volume) from  $N_A$  (number of vesicles per unit area) and VSD in the assumption of spherical  
292 vesicles. In particular, the VSD data were obtained using a reference area (in  $\text{mm}^2$ ), calculated  
293 excluding the area of vesicles touching the edge of the image as well as the area occupied by  
294 phenocrysts. Conversely, the total investigated area was used to estimate the total vesicularity and  
295 vesicle number density.

296 The final vesicle volume fraction for each size class  $V_{fi}$  was calculated using the values of  $N_V$  for  
297 each size class ( $N_{Vi}$ ) and of the equivalent volume of each size class  $V_i$ :

$$298 \quad V_{fi} = N_{Vi} \cdot V_i \quad (1)$$

299 Given the assumption of spherical vesicle shape, the derived values for vesicle volume fractions  
300 ( $V_{fi}$ ) are generally less reliable than those obtained directly from density measurements. For this  
301 reason,  $V_{fi}$  was normalized to the total clast vesicularity as derived from the density  
302 measurements, so obtaining the adjusted volume fractions  $V_{fc}$  for each sample. By analogy, also  
303 the  $N_V$  values were normalized ( $N_V^m$ ) to the melt volume ( $1 - \phi$ ), where  $\phi$  represents the vesicle  
304 fraction derived from the density measurements (Proussevitch et al., 2007).

305 Phenocryst and microlite (plagioclase, clinopyroxene, orthopyroxene, oxide) contents were also  
306 measured by image analysis on selected SEM backscattered electron images. The measured data  
307 were referred to the image area excluding bubbles. Given that thin sections may be only partially  
308 representative of the largest crystals of the mineral assemblage, total crystal content was also  
309 estimated solving linear least-squares mass-balance equations with a dedicated spreadsheet, using  
310 whole-rock, residual groundmass glass and mineral phases compositions from the variable  
311 literature (Barberi et al., 1995; Pistolesi et al., 2011; Saalfeld et al., 2019; Table 2).

312

### 313 **3.3 Conduit dynamics**

314 The five selected eruptions, considering their similar compositions but different MERs and crystal  
315 contents, represent useful case studies to address the conditions that control the intensity of  
316 andesitic (s. l.) explosive eruptions. Thus, we investigated the conduit dynamics of the five  
317 targeted Cotopaxi eruptions (i.e., Layers 5, 3, 2, 1 and Eruption 1877) by adopting the 1D steady-  
318 state conduit model MAMMA (de' Michieli Vitturi and Aravena, 2021). The full system of

319 equations can be found in de' Michieli Vitturi and Aravena (2021). This two-phase numerical  
320 model accounts for the most important processes that magmas experience during ascent in the  
321 conduit, such as changes in rheology, gas exsolution, crystallization, outgassing, and magma  
322 fragmentation. The model includes a series of relaxation parameters for controlling the rate of  
323 crystallization ( $\tau^{(c)}$ ), gas exsolution ( $\tau^{(d)}$ ), and the pressure difference between the two phases  
324 ( $\tau^{(p)}$ ). Here we assumed pressure equilibrium between the phases (i.e.,  $\tau^{(p)} \ll 1$  s), while both  
325 equilibrium and disequilibrium conditions were tested for gas exsolution (in particular, from  
326  $\tau^{(d)} = 10^{-4}$  s for equilibrium conditions to  $\tau^{(d)} = 10$  s for disequilibrium conditions). Please  
327 note that larger values of  $\tau^{(d)}$  tend to reproduce effusive eruptions and in other cases they delay  
328 strongly the crystallization process, which is not in agreement with textural data of some of the  
329 targeted events. Crystallization relaxation time was assumed to be large ( $\tau^{(c)} = 1000$  s) in  
330 eruptions characterized by small volume fractions (< 5 vol. %) of microlites (i.e. Layer 2, Layer 3  
331 and Eruption 1877), and of the order of 10 s for eruptions that present a significant volume  
332 fraction (~30 vol. %) of microlites (i.e. Layer 1 and Layer 5; Table 2). These settings allowed us  
333 to reproduce in the numerical simulations such high contents of microlites. The reasons that can  
334 result in a delayed crystallization process (or not) are discussed below. On the other hand,  
335 modelling magma fragmentation is challenging because it can be generated by different processes  
336 in nature (e.g., Dingwell, 1996; Papale, 1999; Spieler et al., 2004; Mueller et al., 2008;  
337 Gonnermann and Manga, 2013; Cashman and Scheu, 2015; Arzilli et al., 2019b; Taddeucci et al.,  
338 2021). Brittle fragmentation in silicic magmas is expected to be triggered by high strain rates  
339 derived from the acceleration of the ascending magma or by bubble overpressure due to gas  
340 exsolution in presence of restricted bubble expansion. Instead, in low-viscosity magmas, brittle  
341 fragmentation is expected to be strongly influenced by syn-eruptive crystallization processes

342 (Moitra et al., 2018; Arzilli et al., 2019b). Different criteria have been implemented to describe the  
343 fragmentation position in conduit models (Gonnermann and Manga, 2013; Cashman and Scheu,  
344 2015), including the use of a critical volume fraction of bubbles (e.g. Sparks, 1978), stress-based  
345 (e.g., Papale, 1999) and strain-based criteria (e.g., Zhang, 1999). In particular, in this work magma  
346 fragmentation is assumed to occur when a critical volume fraction of bubbles is reached, which is  
347 considered to be equal to the vesicularity measured in the volcanic products. In this way,  
348 simulations are not associated with a specific fragmentation mechanism and are simply  
349 constrained considering textural data (Table 2). In any case, note that Degruyter et al. (2012)  
350 showed that a critical gas volume fraction has similar consequences as a critical strain rate or  
351 overpressure in a one-dimensional conduit model.

352 As a first approximation, we also assumed isothermal conditions and cylindrical conduits, even  
353 though MAMMA allows to model depth-dependent conduit geometries (e.g. Aravena et al.,  
354 2018a; 2018b) and non-isothermal conditions (e.g. La Spina et al., 2015). Based on literature-  
355 derived information and new data of ESPs and textural information (see Section 4), we selected a  
356 set of constitutive equations to define and parametrize magma rheology, crystallization,  
357 outgassing, water exsolution and the equations of state for each studied eruption (Tables 3, S2 and  
358 captions therein). Because some eruption conditions are poorly constrained (e.g., water content,  
359 initial temperature), we considered ranges of values (Table 3; Martel et al., 2018). Inlet pressure  
360 was assumed to be equal to the lithostatic pressure, considering a conduit length of 8 km and a  
361 lithostatic pressure gradient of 25.5 MPa/km (i.e.  $\rho$  of 2600 kg/m<sup>3</sup>). Because MER is an  
362 independently estimated parameter (basing on field data) for each one of the studied eruptions (see  
363 Table 1), for each set of input conditions, we iterated on conduit radius up to reach the conduit

364 geometry conditions consistent with the MER values (Table 1). In this way, MER represents an  
365 input parameter of our numerical simulations, while conduit radius represents an output variable.  
366 The outputs of MAMMA are the profiles along the conduit of key parameters such as velocity,  
367 pressure, density, crystals content and volume fraction of gas. By performing large sets of  
368 numerical simulations, we study the dependence of some model outputs on the described, uncertain  
369 input parameters (i.e. water content, temperature and equilibrium degree of gas exsolution). The  
370 measured contents of phenocrysts and microlites of each targeted eruption were adopted to  
371 recognize the input conditions (i.e. temperature, water content, equilibrium degree) that allow  
372 simulating their characteristics. When possible, we use this information to constrain key parameters  
373 of each targeted eruption, such as fragmentation depth and conduit dimensions, among others. Note  
374 that, because of computational limitations, a set of uncertain input parameters were considered fixed  
375 in our numerical simulations for a given eruption (e.g., critical volume fraction of bubbles, conduit  
376 length, inlet pressure). In fact, the results presented in this paper are based on ~15,000 conduit  
377 simulations (as a consequence of the variable input parameters and the iterative procedure to  
378 calculate the conduit radius) and thus the inclusion of additional variable inputs would have  
379 produced computational capacity problems. In any case, varying these parameters tends to show a  
380 much smaller effect on numerical results than modeled variations in water content, temperature and  
381 relaxation parameters of crystallization and gas exsolution. An additional significant simplification  
382 of our model is related to the effect of crystals in magma viscosity (Costa et al., 2007). Despite the  
383 well-documented influence of crystal size distribution and crystal shape in magma rheology  
384 (Cimarelli et al., 2011; Del Gaudio, 2014; Moitra and Gonnermann, 2015), the lack of generalized  
385 formulations able to consider the main characteristics of a system composed of phenocrysts and,  
386 eventually microlites with different shapes, hinders the consideration of more complex assumptions.

387 We remark that, despite the well-known limitations of numerical modeling for constraining  
388 numerically eruption conditions in presence of a series of uncertain parameters as those cited above,  
389 these results provide useful information for comparison purposes and to discuss general issues  
390 regarding high-intensity explosive eruptions driven by mildly evolved magmas.

391

## 392 **4. Results**

### 393 **4.1 Eruptive Source Parameters**

394 The five selected Cotopaxi eruptions are associated with a wide range of ESPs (Table 1). In  
395 particular, the volume of the studied tephra deposits ranges between  $2.3 \times 10^7 \text{ m}^3$  (Eruption 1877)  
396 and  $6.0 \times 10^8 \text{ m}^3$  (Layer 3) when using the method of Pyle (1989) based on the integration of the  
397 Exponential fit (Pistolesi et al., 2011; Biass and Bonadonna, 2011; Fig. 1 and Table 1). Given the  
398 limited number of isopach curves that result in a single exponential segment in most cases, the  
399 integration of the Weibull fit provides similar results to the integration of the Exponential fit.  
400 Nonetheless, the integration of the Power Law fit (Bonadonna and Houghton, 2005) results in  
401 volumes about 3 times larger, comprised between  $4.9 \times 10^7 \text{ m}^3$  (Eruption 1877) and  $1.5 \times 10^9 \text{ m}^3$   
402 (Layer 3) (Pistolesi et al., 2011; Biass and Bonadonna, 2011; Biass et al., 2019; Table 1). In  
403 particular, the volumes of the tephra deposits of Layers 3 and 5 were also confirmed by the  
404 inversion of field data using the TEPHRA2 model (i.e.  $2.4 \times 10^9 \text{ m}^3$  for Layer 3 and  $5.0 \times 10^8 \text{ m}^3$  for  
405 Layer 3; Biass and Bonadonna, 2011). Average values of tephra-deposit volume for each eruption  
406 (excluding the inversion results) correspond to  $3.2 \pm 1.2 \times 10^7 \text{ m}^3$  (Eruption 1877),  $2.3 \pm 1.1 \times 10^8 \text{ m}^3$   
407 (Layer 1),  $3.7 \pm 3.5 \times 10^8 \text{ m}^3$  (Layer 2),  $8.8 \pm 4.4 \times 10^8 \text{ m}^3$  (Layer 3), and  $3.0 \pm 0.6 \times 10^8 \text{ m}^3$  (Layer 5).  
408 As expected, values of maximum plume height derived from the model of Rossi et al. (2019) are  
409 between 25-48% lower than those derived from the model of Carey and Sparks (1986) with the

410 exception of the Eruption 1877, for which the difference is only 8%. Such a large difference is  
411 related to the high wind speeds associated with the eruptions of Layers 1, 2, 3 and 5. It is worth  
412 noting that the wind value reported in Table 1 corresponds to the maximum value at the  
413 tropopause, while the associated values of wind speed averaged along the plume height are 2.1  
414 m/s, 7.2 m/s, 6.1 m/s, 6.7 m/s and 6.3 m/s for Eruption 1877, Layer 1, Layer 2, Layer 3 and Layer  
415 5, respectively.

416 Finally, the peak of MER, determined with the theoretical equation of Degruyter and Bonadonna  
417 (2012) and computed using the values of maximum plume height and wind speed derived from the  
418 model of Rossi et al. (2019), results in values between  $9.1 \times 10^6$  kg/s (Eruption 1877) and  $5.9 \times 10^7$   
419 kg/s (Layer 1; Table 1). These MER values are generally higher than those derived with the  
420 models of Wilson and Walker (1987) and Mastin et al. (2009) for the same sets of plume heights  
421 (with the exception of Eruption 1877), with discrepancies between 22% and 62% and between  
422 23% and 36%, respectively. Interestingly, the model of Rossi et al. (2019), that takes into account  
423 the effect of the wind shear on plume rise, results in Layer 1 being associated with the highest  
424 plume (20.3 km above vent; a.v.), the strongest wind at tropopause (18.7 m/s) and the highest  
425 MER with the three strategies considered in this work ( $2.2$ - $5.9 \times 10^7$  kg/s). In contrast, the  
426 application of the model of Carey and Sparks (1986) results in Layer 3 being associated with the  
427 highest plume (26.8 km a.v.) and the strongest wind (18.3 m/s at tropopause), so giving the  
428 highest MER with the three strategies considered here ( $0.7$ - $1.9 \times 10^8$  kg/s).

429 The minimum duration of each eruption was derived by dividing the Erupted Mass (estimated  
430 from the average value of volume reported above and the associated deposit density) by the peak  
431 MER (calculated with the equation of Degruyter and Bonadonna, 2012) and resulted in 0.5, 1.0,  
432 3.8, 4.8 and 2.1 hours for Eruption 1877, Layer 1, Layer 2, Layer 3 and Layer 5, respectively.

433

#### 434 **4.2 Textural data of the juvenile fraction**

435 The studied products are moderately porphyritic with a total maximum crystal content between  
436 16% (Layer 3) and 40% (Layer 1; Table 2). This range of values derives from a mass balance  
437 estimate considering the major elements compositions of the whole rock, glass analyses and the  
438 mineralogical assemblage. Mineralogy of juvenile clasts is characterized by a rather uniform  
439 phenocryst assemblage that includes plagioclase (max. 15%), clinopyroxene (max. 2%),  
440 orthopyroxene (max. 3%) and Fe-Ti oxides (max. 1%) as fundamental minerals. Microlite  
441 abundance is variable. In particular, the events with the lowest SiO<sub>2</sub> contents are characterized by  
442 a microlite content up to about 30 vol. % (Layers 1 and 5) while the events with the highest SiO<sub>2</sub>  
443 contents have a glassy groundmass (Eruption 1877, Layers 2 and 3). In general, microlites have  
444 skeletal features, suggestive of a rapid growth under large undercooling (e.g. Shea and Hammer,  
445 2013).

446 The average clast density distribution associated with the five analyzed layers varies between 0.62  
447 g cm<sup>-3</sup> and 1.24 g cm<sup>-3</sup> (55-77% density-derived clast vesicularity; Fig. 2; Table 2), with the  
448 lowest values associated with the two most evolved eruptions (Layers 2 and 3). Overall, clast  
449 density has a large, unimodal distribution (0.4-2 g cm<sup>-3</sup>), with Layer 1 and Eruption 1877 showing  
450 a narrower distribution around intermediate (~1 g cm<sup>-3</sup>) values. Density distribution of Layer 5 is  
451 bimodal, showing the widest variability and the highest measured density values within the  
452 studied volcanic products (Fig. 2).

453 Vesicle size has a polydisperse distribution, with the smallest vesicles around 8 μm and the largest  
454 around 6 mm. The smallest bubbles have a spherical to elongate shape, and the largest  
455 deformation of this bubble population is mainly visible in samples from Layer 2 and Layer 3 (Fig.

456 3). Clear evidence of bubble collapse only characterizes the smallest bubbles of the pumices from  
457 Eruption 1877. In the samples of all the eruptions, the largest vesicles are always irregular,  
458 probably due to processes of bubble coalescence and/or collapse. This is particularly evident in the  
459 densest products of the studied samples (Eruption 1877 and Layer 5). Overall, the Vesicle Volume  
460 Distributions (VVDs) are polymodal, with two to three distinct modes, except for products of  
461 Eruption 1877, characterized by an asymmetric, unimodal distribution, with a sharply truncated  
462 coarse tail (Fig. 3). Contrary to the highly dispersed bubble size distributions of the products from  
463 Layers 1, 2 and 5, the products of Eruption 1877 and Layer 3 are characterized by a lower  
464 dispersion of bubble size, with a marked peak around  $\sim 0.5\text{-}0.9$  and  $\sim 0.3\text{-}0.6$  mm, respectively  
465 (Fig. 3). The maximum bubble size of the products of Eruption 1877 is  $\sim 1.1$  mm, significantly  
466 lower than that observed in the other eruptions. On the other hand, the bubble size volume  
467 distribution in the products of Layer 3 shows a secondary mode between 2 and 6 mm (Fig. 3). The  
468 diagrams of VSD (ln population density vs. size) for all the eruptions always show curved,  
469 concave upward distributions, that in some cases (Eruption 1877 and Layers 2 and 3) can be  
470 described by the combination of three main linear segments (Fig. S1)

471 In general, values of melt normalized vesicle number density ( $N_V^m$ ) of Layers 1, 2, and 5 are  
472 similar, ranging from  $1.06 \times 10^8$  to  $4.80 \times 10^8$   $\text{cm}^{-3}$ , whereas the  $N_V^m$  of the Eruption 1877 is one  
473 order of magnitude lower ( $1.47 \times 10^7$   $\text{cm}^{-3}$ ). The products of Layer 3 are instead characterized by  
474 the highest values of  $N_V^m$  ( $1.37 \times 10^9$   $\text{cm}^{-3}$ ; Fig. 3 and Table 2).

475

### 476 **4.3 Conduit dynamics**

477 To describe numerical results, we will focus on three of the selected eruptions: Eruption 1877,  
478 Layer 1 and Layer 3, which include volcanic products both with glassy groundmass (Eruption

479 1877 and Layer 3) and with ~30 vol. % of microlites (Layer 1), and also present the minimum  
480 (Eruption 1877) and maximum (Layer 1) MER and the minimum (Layer 1) and maximum (Layer  
481 3) silica content of the studied events (Table 1). The detailed modeling results associated with the  
482 other targeted events are included in the Supplementary Material.

483

#### 484 **4.3.1 Eruption 1877**

485 Figure 4 presents representative examples of the results derived from specific conduit simulations,  
486 selected in order to include the cases of simulations with scarce crystallization (upper panels) and  
487 with significant microlite content (lower panels). In the upper panels, we show the profiles along  
488 the conduit of key physical parameters (mixture pressure, microlite volume fraction, viscosity, gas  
489 volume fraction, dissolved water mass fraction and mixture velocity) for two simulations  
490 performed for the Eruption 1877. In these simulations, two different values for the gas exsolution  
491 relaxation parameter were used. For each simulation, the conduit radius was calculated in order to  
492 reproduce the MER of the Eruption 1877, and, in both cases, it is around 13 m. Magma ascent  
493 occurs with sparse microlite crystallization, which results in a magma characterized by a viscosity  
494 close to that of the melt near the fragmentation level. Considering the specific input conditions for  
495 these simulations (water content of 3.1 wt. % and temperature of 950°C), numerical results  
496 suggest that gas exsolution occurs from depths of about 4.5 km and fragmentation occurs at ~2.0  
497 km depth. However, these and other results such as viscosity and exit pressure are strongly  
498 controlled by unconstrained input parameters (i.e. water content and temperature), and thus the  
499 analysis of a larger set of simulations is needed to describe properly the eruption dynamics. Under  
500 these specific circumstances, the adoption of contrasting values of the relaxation parameter for gas  
501 exsolution does not have a strong influence on the modeled dynamics of magma ascent (Fig. 4),

502 due to the long characteristic time of magma ascent compared to the adopted values of  $\tau^{(d)}$  (note  
503 that the use of larger values of  $\tau^{(d)}$  result in effusive events).

504 Figure 5 presents a summary of the results of a set of simulations considering fixed conditions for  
505 the relaxation parameters associated with crystallization and gas exsolution (in particular,  $\tau^{(c)} =$   
506  $1000\text{ s}$  and  $\tau^{(d)} = 10^{-4}\text{ s}$ ). The high value adopted for  $\tau^{(c)}$  is here justified by the absence of an  
507 important microlite crystallization in the melt. Each panel presents a color scale of different output  
508 parameters of our simulations (fragmentation depth, conduit radius, exit pressure, exit velocity,  
509 magma viscosity at the fragmentation level and melt viscosity at the fragmentation level) as a  
510 function of initial temperature and water content. Additionally, two superposed contour maps  
511 indicate the initial volume fraction of phenocrysts (continuous contour black lines) and volume  
512 fraction of microlites (dashed contour black lines). Phenocryst volume fraction decreases when  
513 water content and temperature increase, with values between 0 vol. % and >35 vol. % for the  
514 adopted variation range of input parameters. Measured values of magma crystallinity in terms of  
515 phenocryst and microlite content (Table 2) are then used to partially constrain the field of expected  
516 variability of initial temperature, water content, and of all the other output parameters represented  
517 in Figure 5. Because the observed content of phenocrysts is about 35 vol. %, we remark that not  
518 all the combinations of water content and input temperature are compatible with the results of  
519 textural analyses of pyroclastic products of this eruption. This consideration, adopting a tolerance  
520 range of 5 vol. %, is then used to define plausible ranges of eruption conditions. Under the  
521 simulation conditions assumed for Eruption 1877 (i.e. high value of  $\tau^{(c)}$ ), the numerical results are  
522 associated with a very low (about 3 vol. %) microlite content, in agreement with the measured  
523 data.

524 We also observe that narrow conduits are sufficient to produce the MER of this eruption, with  
525 values between ~12 m and ~15 m. Magma fragmentation depth ranges from ~1.5 km to ~2.2 km  
526 (Fig. 5). On the other hand, numerical simulations allow constraining other eruption parameters  
527 such as exit pressure (~10 Atm to ~20 Atm, choked conditions) and exit velocity (~100 m/s to  
528 ~120 m/s). Magma (i.e. melt + crystals + gas) and melt viscosity at the fragmentation level range  
529 between  $6.0 \times 10^4$  Pa s and  $2.2 \times 10^5$  Pa s and between  $6.3 \times 10^4$  Pa s and  $1.4 \times 10^5$  Pa s, respectively.  
530 It is important to note that, in addition to the influence of crystals (Costa, 2005) and bubbles  
531 (Costa et al., 2007), the resulting magma viscosity at the fragmentation level is controlled by the  
532 coupled effect of magma composition, temperature, and gas volume fraction at fragmentation (i.e.  
533 the measured vesicularity of volcanic products).

534 Only slight changes occur in numerical results when other conditions are considered for the  
535 relaxation parameter of gas exsolution (Fig. S2). Based on our numerical results (i.e. Figs. 5 and  
536 S2), we defined a set of likely eruption conditions for Eruption 1877 that are summarized in Table  
537 4. Some parameters can be well constrained, such as conduit dimensions and the order of  
538 magnitude of magma viscosity at the fragmentation level, while other outputs (i.e. fragmentation  
539 depth) are characterized by a significant uncertainty.

540

#### 541 **4.3.2 Layer 1**

542 The bottom panels of Figure 4 display the profiles along the conduit of key physical parameters  
543 associated with two specific simulations performed for Layer 1. The differences in the results of  
544 simulations performed with variable values of gas exsolution relaxation parameter are here  
545 significant; conduit radius, computed through an iterative procedure with the aim of simulating the  
546 estimated MER of Layer 1, results in values of ~15.5 m (for  $\tau^{(d)} = 10^{-4}$  s) and ~17 m (for

547  $\tau^{(d)} = 10 \text{ s}$ ). In contrast to the upper panels of Figure 4, these simulations were performed in  
548 order to accomplish a significant crystallization of microlites as observed in the natural samples of  
549 this eruption (i.e. using a lower value for  $\tau^{(c)}$ ), which results in a larger difference between the  
550 magma viscosity and melt viscosity at the fragmentation level.

551 Figure 6 displays the results associated with a set of simulations performed for Layer 1, for which  
552 we considered fixed conditions for the relaxation parameters related to crystallization ( $\tau^{(c)} =$   
553  $10 \text{ s}$ ) and gas exsolution ( $\tau^{(d)} = 10^{-4} \text{ s}$ ). Phenocryst volume fraction ranges between 0 vol. %  
554 and >40 vol. % for the modeled range of water content and temperature. Again, we highlight that  
555 not all the combinations of water content and input temperature are compatible with the  
556 phenocrysts content measured for this eruption (about 20 vol. %). The low value adopted for  $\tau^{(c)}$   
557 allows for significant microlite crystallization, reaching concentrations of the order of 30 vol. %,  
558 in agreement with the texture observed in the associated volcanic products. In this case as well,  
559 these considerations were used to define plausible ranges of eruption conditions, which are less  
560 constrained (especially in terms of water content) respect to what observed for the Eruption 1877  
561 simulations. It is worth highlighting that larger values of  $\tau^{(c)}$  and  $\tau^{(d)}$  are not able to model the  
562 observed concentration of microlites and thus their results are not included here.

563 The higher MER associated with Layer 1 reflects into larger conduits, with a maximum range for  
564 radius between  $\sim 14 \text{ m}$  and  $\sim 19 \text{ m}$ . Fragmentation depths compatible with phenocrysts and  
565 microlites contents vary between  $\sim 0.3 \text{ km}$  and  $\sim 1.0 \text{ km}$ , while exit pressure and exit velocity  
566 ranges between  $\sim 60 \text{ Atm}$  and  $\sim 120 \text{ Atm}$  and from  $\sim 130 \text{ m/s}$  to  $\sim 190 \text{ m/s}$ , respectively. Finally,  
567 magma and melt viscosity at the fragmentation level range between  $9.1 \times 10^3 \text{ Pa s}$  and  $8.7 \times 10^4 \text{ Pa s}$   
568 and between  $6.9 \times 10^3 \text{ Pa s}$  and  $1.9 \times 10^4 \text{ Pa s}$ , respectively. The differences between magma and  
569 melt viscosity derive in this case from the dominant effect of crystals (Costa, 2005).

570 Figure S3 presents the equivalent results associated with disequilibrium conditions of gas  
571 exsolution. Based on these results (i.e., Figs. 6 and S3), a set of likely eruption conditions for  
572 Layer 1 was defined, presented in Table 4. Please note that differences between the results  
573 summarized in Table 4 and the description presented in the previous paragraph for Figure 6 are  
574 small.

575

### 576 **4.3.3 Layer 3**

577 Layer 3 represents an end-member for the five selected eruptions, showing the most evolved  
578 magma composition coupled with a very low crystal content. In Figure 7, we present the results of  
579 a set of simulations performed for this eruption, considering  $\tau^{(c)} = 1000 \text{ s}$  and  $\tau^{(d)} = 10^{-4} \text{ s}$ .  
580 The volume fraction of phenocrysts varies between 0 vol. % and ~30 vol. % for the adopted values  
581 of water content and temperature. Also in this case, only a restricted portion of the panels  
582 presented in Figure 7 is compatible with the content of phenocrysts measured in the products of  
583 Layer 3 (about 10 vol. %) allowing to define plausible ranges for eruption conditions. On the other  
584 hand, the value adopted for  $\tau^{(c)}$  is manifested in sparse crystallization, compatible with the  
585 characteristics of this eruption. In this case, the very low content of microlites in all the  
586 simulations reflects in a less robust constrain for eruption conditions, at least in terms of initial  
587 water content and temperature.

588 The results presented in Figure 7 suggest a conduit radius between ~13 m and ~18 m for Layer 3.  
589 Fragmentation depths compatible with textural data vary between ~0.1 km and ~1.2 km, exit  
590 pressure ranges from ~40 Atm to ~90 Atm, and exit velocity varies between ~130 m/s and ~190  
591 m/s. On the other hand, magma and melt viscosity at the level of fragmentation vary between  
592  $6.5 \times 10^3 \text{ Pa s}$  and  $6.6 \times 10^4 \text{ Pa s}$  and between  $5.6 \times 10^4 \text{ Pa s}$  and  $3.8 \times 10^5 \text{ Pa s}$ , respectively.

593 The equivalent results derived from disequilibrium conditions of gas exsolution are displayed in  
594 Figure S4. From these results (i.e., Figs. 7 and S6), we defined likely eruption conditions for Layer  
595 3 (Table 4). Numerical results associated with Layers 2 and 5 are presented in Figures S4, S5, S7,  
596 and S8 and a summary of the derived constraints for key eruption parameters is included in Table  
597 4.

598

## 599 **5. Discussion**

### 600 **5.1 Conduit modeling and eruption dynamics**

601 Given the significant range of ESPs for the recent activity at Cotopaxi volcano, and based on the  
602 limited range of magma compositions involved in this activity, we emphasize the importance of  
603 having accurate parameters (both textural and physical) in order to address and discuss how  
604 magma rheology may affect eruption dynamics. To accomplish this, we revised all the available  
605 field data in order to retrieve ESPs based on the most recent models, which were then combined  
606 with detailed textural analyses of the juvenile products and with state-of-the-art conduit dynamics  
607 modeling.

608 Results summarized in Table 4 indicate only slight differences for the radius of the conduit  
609 associated with all eruptions, suggesting that other factors controlled the variability of MER of the  
610 studied eruptions (Fig. 8a). Instead, there is a good correlation among MER and dynamic features  
611 of the erupting mixture such as exit pressure and exit velocity, with the eruptions characterized by  
612 the highest values of MER (i.e. Layers 1 and 5) having also the highest contents of microlites and  
613 the highest exit pressures (50 – 120 Atm) and velocities (130 – 190 ms<sup>-1</sup>). Conversely, the  
614 eruptions with products characterized by a glassy groundmass (i.e. Eruption 1877 and Layer 2)  
615 have exit pressures lower than 40 Atm coupled with a low exit velocity (90 – 130 ms<sup>-1</sup>; Fig. 8b, c).

616 Layer 3, which is significantly richer in  $\text{SiO}_2$  than the other targeted eruptions, presents  
617 intermediate results for the two variables. Interestingly, the event characterized by the lowest  
618 values of exit pressure and velocity (i.e. Eruption 1877) is also associated with boiling-over  
619 activity resulting in the emplacement of widespread scoria flow deposits during the eruption  
620 (Barberi et al., 1995; Pistolesi et al., 2011). The presence of a crater (well visible at Cotopaxi)  
621 could have modulated the decompression of the jet: according to the numerical results of Woods  
622 and Bower (1995), jets with a low overpressure can rapidly decompress well below the external  
623 pressure and re-equilibrate through a shock wave inside the crater. This recompression results in a  
624 drastic change of the mixture velocity and hence in a low fountaining of the column feeding the  
625 boiling over activity.

626 Although fragmentation depth could not be unequivocally constrained, significant differences  
627 have been recognized among the studied eruptions, showing a general decrease with increasing  
628 MER; in this context, those eruptions with a higher groundmass crystallinity (Layers 1 and 5) also  
629 show the shallowest fragmentation depth (Fig. 8d).

630 An unclear relationship is shown between MER and magma (i.e. melt + crystals + bubbles)  
631 viscosity at the fragmentation depth (Fig. 8e), with the events with the highest MER also having  
632 the lowest magma viscosity at this level. Similar, high values of MER characterize both eruptions  
633 with the lowest  $\text{SiO}_2$  contents (Layers 1 and 5) and Layer 3, which shows the highest  $\text{SiO}_2$   
634 content. Interestingly, even though it is well known that silica content is positively correlated with  
635 melt viscosity (e.g. Giordano et al., 2008), the influence of microlites, more abundant in the  
636 eruptions characterized by the lowest contents of silica (and highest MER), results in an apparent  
637 lower value of magma viscosity at the fragmentation level (Fig. 8f). This counter-intuitive result,  
638 which is also manifested in an apparent lack of correlation between MER and silica content,

639 highlights the critical importance of crystallization in controlling the eruption dynamics of  
640 intermediate magmas.

641

## 642 **5.2 Texture, composition and eruptive source parameters**

643 Size distributions of vesicles from the different eruptions show variable characteristics. Data from  
644 Layer 1 and Layer 5 suggest similar conditions of vesiculation, with a quite large size range  
645 covered by the vesicles, a negligible amount of vesicle coalescence and clearly concave upward  
646 VSD plots, suggesting continuous/accelerating bubble nucleation and growth (Shea et al., 2010;  
647 Fig. S1). This was possibly accomplished by a relatively low magma ascent velocity (averaged  
648 along the conduit), which allowed also the occurrence of an important, degassing-induced,  
649 microlite crystallization. The high exit pressure suggested for these two eruptions by the modeling  
650 results (Table 4) is in general agreement with this interpretation. Pumice clasts from Layer 2 and  
651 Layer 3 present clear multimodal VVDs (Fig. 3), suggestive of a complex vesiculation history.  
652 However, while the vesicle size distribution of Layer 3 can be interpreted as related to the  
653 presence of important coalescence effects on a distribution characterized by a curved, concave  
654 upward trend, the complex VVD diagram of Layer 2 (Fig. 3), associated with a VSD curve  
655 characterized by three different segments (Fig. S1), reveals the occurrence of few, distinct  
656 episodes of vesiculation during ascent. The ESPs and dynamical parameters estimated for Layer 3  
657 (high MER, high exit pressure and velocity, shallow fragmentation depth) are all indicative of a  
658 large eruption characterized by a rapid, continuous and accelerating process of magma ascent,  
659 decompression and degassing, well recorded by the general features of vesicle shape and size.  
660 Pumices from the Eruption 1877 show the largest variability in terms of vesicle distribution; these  
661 are in fact characterized by important differences in shape and volume density of bubbles, and by

662 the presence, in most of the samples, of collapsed vesicles. The shape of the VSD and CVD (Figs.  
663 3, S1) diagrams suggest the presence of few, distinct episodes of vesicle nucleation and growth.  
664 The small conduit radius related to this eruption (Table 4), inducing important lateral gradients in  
665 the ascending magma column, could explain the large variability observed in the texture of the  
666 vesicular material. The partial outgassing of the magma (suggested by the presence of collapsed  
667 vesicles) was however not sufficient to force an important microlite crystallization (low  
668 undercooling?).

669 A positive correlation between  $N_V^m$  and whole-rock  $\text{SiO}_2$  content of the studied samples exists for  
670 the products of Layers 1, 2, 3 and 5 (Fig. 9a). An important discrepancy from this linear trend  
671 occurs for the Eruption 1877 products, which show a lower value of  $N_V^m$  respect to that observed  
672 for similar compositions. However, the texture of these products reveals clear evidence of open-  
673 degassing behavior, which could have largely reduced the vesicle number density. The variation of  
674 the general texture of the analyzed products with the increase of  $\text{SiO}_2$  content suggests that the  
675 more evolved compositions are characterized by smaller and more deformed (although still with  
676 convex shapes) bubbles, with bubble walls thinner than the other products (Fig. 3). In general, we  
677 observe a variation of the texture of Cotopaxi products with increasing  $\text{SiO}_2$  content, even though  
678 in a limited range (Table S3). However, it is worth noting that these products are characterized by  
679 significant differences in glass matrix composition due to the differences in crystal content, with a  
680 transition from andesitic (Layers 1 and 5) to dacitic (Eruption 1877, Layer 2 and Layer 3), where  
681 the greatest changes in bubble texture seem to be recorded. Despite this rough correlation among  
682 silica content and  $N_V^m$ , no clear relationship exists among  $N_V^m$  and MER (Fig. 9b), possibly  
683 related to the fact that for 3 out of 5 eruptions the MER variability spans a range of values that lie  
684 within the associated errors. A clear negative correlation also exists between  $N_V^m$  and magma

685 viscosity (Fig. 9e, f), suggesting that bubble nucleation (+ resorption) may be hindered at high  
686 viscosity. This is particularly evident for Eruption 1877, for which the lowest  $N_V^m$  is clearly  
687 associated with vesicle resorption and collapse. Unexpectedly, we observe a positive correlation  
688 between  $N_V^m$  and total erupted volume (Fig. 9c). The significance of this clear correlation is  
689 somewhat obscure, and it should be possibly linked with the complex interactions between the  
690 dynamics of reservoir emptying and the associated pressure changes, that could in turn reflect  
691 onto the dynamics of bubble growth. The absence of a clear correlation between parameters like  
692 MER, silica content or melt viscosity and vesicle number density suggests that the relation  
693 evidenced by Toramaru (2006) among some of these parameters should be accurately  
694 reconsidered, particularly when looking at restricted silica variability.

695

### 696 **5.3 Controls of magma features on eruptive parameters in eruptions of intermediate magma** 697 **compositions**

698 In order to explain how eruption dynamics (e.g. MER) may be controlled by pre-eruptive  
699 parameters and magma rheology, we have combined here textural data, ESPs and dynamic  
700 parameters derived from numerical modeling. Unfortunately, we have no direct data for pre-  
701 eruptive volatile (water) content of the magma, although a rough indirect estimation can be  
702 derived from fitting the results of numerical modeling (Figs. 5, 6, 7 and S2 to S8) with some  
703 observables. The aim of this discussion is to show any possible relation between the observed or  
704 inferred magma features (composition, crystal content,  $N_V$ , water content, viscosity) and measured  
705 or modeled dynamical parameters. Eruptions characterized by the most mafic compositions also  
706 correspond to the most crystallized samples (Layers 1 and 5); modeling results suggest that these  
707 eruptions were possibly associated with volatile-rich magmas ( $H_2O > 3.5$  wt. %) and a relatively

708 low melt viscosity (Figs. 6, S3, S7, S8). All these conditions possibly favored eruptive dynamics  
709 dominated by the highest MER and the highest exit pressures and velocities.

710 High volatile contents (Figs. 7 and S6) also characterized the SiO<sub>2</sub>-richer magma of Layer 3,  
711 erupted at high MER but without developing a diffuse groundmass crystallization. The high melt  
712 viscosity and the rapid magma ascent could have forced in this case an important delay in syn-  
713 eruptive vesiculation, forced to occur at high overpressure over a short length below the  
714 fragmentation level and so resulting in a high N<sub>V</sub> value and a poorly dispersed (except for the  
715 effects of coalescence) vesicle size distribution (Fig. 3).

716 A glassy groundmass also characterizes the products of Layer 2 and Eruption 1877, having an  
717 intermediate composition and a low volatile content. These eruptions have the lowest MER  
718 between the five targeted events, and clearly differentiate respect to the two crystal-rich eruptions  
719 discussed above. Despite their similarity in terms of some of the most important dynamical or  
720 compositional parameters, these two events clearly differ for the dynamics of magma degassing as  
721 recorded in the textural features of their products (Fig. 3).

722 As a whole, two end members can be observed for texture of the melt, one having a very low (0-  
723 2%) microlite content (Eruption 1877, Layers 2 and 3), and one with high (30%) microlite content  
724 (Layers 1 and 5), with no group having intermediate characteristics. The microlite texture is  
725 however partly counterbalanced by phenocryst abundance, resulting in total crystal content of 16  
726 to 40 wt. %. Despite a range in matrix glass composition (andesitic to dacitic) due to variable  
727 microlite content, overall crystallinity is thus partly counterbalanced by phenocryst abundance,  
728 resulting in a more subtle variability among the studied eruptions. This, coupled to a small range  
729 of bulk-rock composition, may suggest strong feedbacks effects among crystallization, changes in  
730 melt/magma viscosity and volatile exsolution. Once started, the process of microlite crystallization

731 may result in a rapid change in magma rheology, further enhancing gas exsolution and microlite  
732 formation, with the latter becoming rapidly dominant during magma ascent.

733 It has been suggested that magma characteristics (i.e. composition and texture) scale with eruption  
734 intensity (i.e. MER). In particular, although the relationship between the evolution of the volatile  
735 fraction within the melt and the explosivity of the eruption is not totally understood, bubble  
736 number density vs. SiO<sub>2</sub> (Toramaru, 2006) and MER vs. N<sub>v</sub><sup>m</sup> (Alfano et al., 2012; Houghton et al.,  
737 2010) were used to suggest a positive correlation between intensity and tephra texture. On one  
738 hand, this general correlation cannot be generalized to all cases, and it is limited to those cases in  
739 which vesiculation occurs under near-equilibrium condition, as shown by Rust and Cashman  
740 (2011). In fact, while at a broader scale the trend may appear evident, different behaviors can be  
741 extrapolated when considering the diverse compositional sub-groups (e.g. basaltic, phonolitic,  
742 andesitic-rhyolitic; Alfano et al., 2011), with further complications related to microlite abundance  
743 (Moitra et al., 2013). Our data further confirm that, when considering a small range of magma  
744 composition, the observed variability in MER is coupled to a wide range of vesicle number  
745 densities which barely define a clear trend. This apparent scatter suggests that, at comparable  
746 degree of magma evolution (in our case 59.5 < whole rock SiO<sub>2</sub> wt.% < 64.9), other factors (e.g.  
747 magma ascent velocity, degree of undercooling and microlite nucleation and growth) are at play in  
748 governing conduit dynamics, with bubble nucleation and evolution resulting in complex size  
749 distributions and largely variable N<sub>v</sub><sup>m</sup> values. For the studied eruptions, we envisage that the  
750 combination of variable (0-2% to 30%) microlite contents with different phenocryst abundances  
751 (total crystal content range of 16 - 40 wt.%) may eventually control eruption dynamics which,  
752 despite a limited variability of bulk magma composition, resulted in a wide range of ESPs, as  
753 retrieved from deposit studies. Also, modeling results suggest this variability, with microlite-rich

754 mafic eruptions (L1, L5) showing high MDR, exit pressure, velocity and water content despite a  
755 low melt viscosity, and more evolved cases (L2 and 1877) having lower water content and  
756 dynamical parameters coupled to a glassy groundmass and a higher viscosity. As a matter of fact,  
757 the presented data demonstrate the existence of fully non-linear, complex inter-relationships  
758 between the different dynamical and textural parameters, warning about the use of simple (or  
759 simplistic) relations for deriving general laws. The example of Cotopaxi is particularly relevant in  
760 showing that this complexity can characterize a very limited range of magma compositions which  
761 however resulted in eruptive events having ESPs spanning two orders of magnitude. It has  
762 recently shown that for rhyolitic compositions and under homogeneous nucleation conditions, a  
763 general trend predicted by the bubble number density decompression rate meter of Toramaru  
764 (2006) still holds for  $N_V > 10^7 \text{ cm}^{-3}$  (Hajimirza et al., 2019), while the relation cannot be applied  
765 when  $N_V$  is lower for diffusion to affect saturation. Decompression-independent bubble number  
766 densities in decompression experiments, albeit in hydrated phonolitic melts, has been shown by  
767 Allabar and Nowak (2018). The little or no dependency of number density on eruption intensity  
768 (or decompression rate) that we observe here may be ascribed to the large textural variability (e.g.  
769 microlite content) and volatile saturation conditions, which may result in a complex interplay  
770 between diffusion and decompression nearly unrelated to the slight compositional variability.

771

## 772 **6. Conclusions**

773 Explosive behavior of volcanoes results from the complex interplay among initial magma  
774 properties and ascent dynamics along the conduit. This interplay in turn modulates continuous  
775 changes in extensive and intensive parameters before final magma fragmentation. By examining  
776 five eruptions occurred at Cotopaxi volcano, we had the opportunity of exploring how events

777 characterized by a small variation in magma composition resulted in explosive events spanning a  
778 wide range of intensity, from sub-Plinian to Plinian. To accomplish this, we combined both  
779 textural and physical parameters in order to address and discuss how magma rheology may affect  
780 eruption dynamics. All the available field data were revised in order to retrieve ESPs based on the  
781 most recent models, which were then combined with detailed textural analyses of the juvenile  
782 products and with state-of-the-art conduit dynamics modeling. We found that the five selected  
783 eruptions can be grouped in two end members in relation to texture of the products, one having a  
784 very low (0-2%) and one high (30%) microlite content. Nonetheless, the effect of microlite  
785 content is partly counterbalanced by phenocryst abundance, resulting in total crystal content of 16  
786 to 40 wt. %. The combination of conduit modeling results with textural data and ESPs suggests  
787 that subtle variability in crystal content and magma composition may be accompanied by strong  
788 feedbacks effects among crystallization, changes in melt/magma viscosity and volatile exsolution,  
789 with microlite crystallization resulting in a rapid change in magma rheology and eventually in  
790 different explosive dynamics.

791 Finally, we also emphasize that the general observed correlation among magma characteristics  
792 (i.e. composition and texture) with eruption intensity (i.e. MER) may be problematic when applied  
793 to eruptions characterized by small compositional variability. While at a broader scale the trend  
794 may be evident, particularly considering compositional sub-groups (e.g. basaltic, phonolitic,  
795 andesitic-rhyolitic), we show that, when dealing with subtle variations in magma composition, the  
796 observed variability in MER is coupled to a variety of textural characteristics (i.e. number  
797 densities) such that the definition of a clear trend results problematic. Data scattering may be  
798 related to a complex series of parameters at play (magma ascent velocity, degree of undercooling

799 and microlite nucleation and growth) resulting in very different explosive dynamics despite the  
800 limited compositional range.

801

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813 et al. (2015) available at [https://github.com/e5k/CareySparks86\\_Matlab](https://github.com/e5k/CareySparks86_Matlab) and at  
814 <https://vhub.org/resources/3922>.

815

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823

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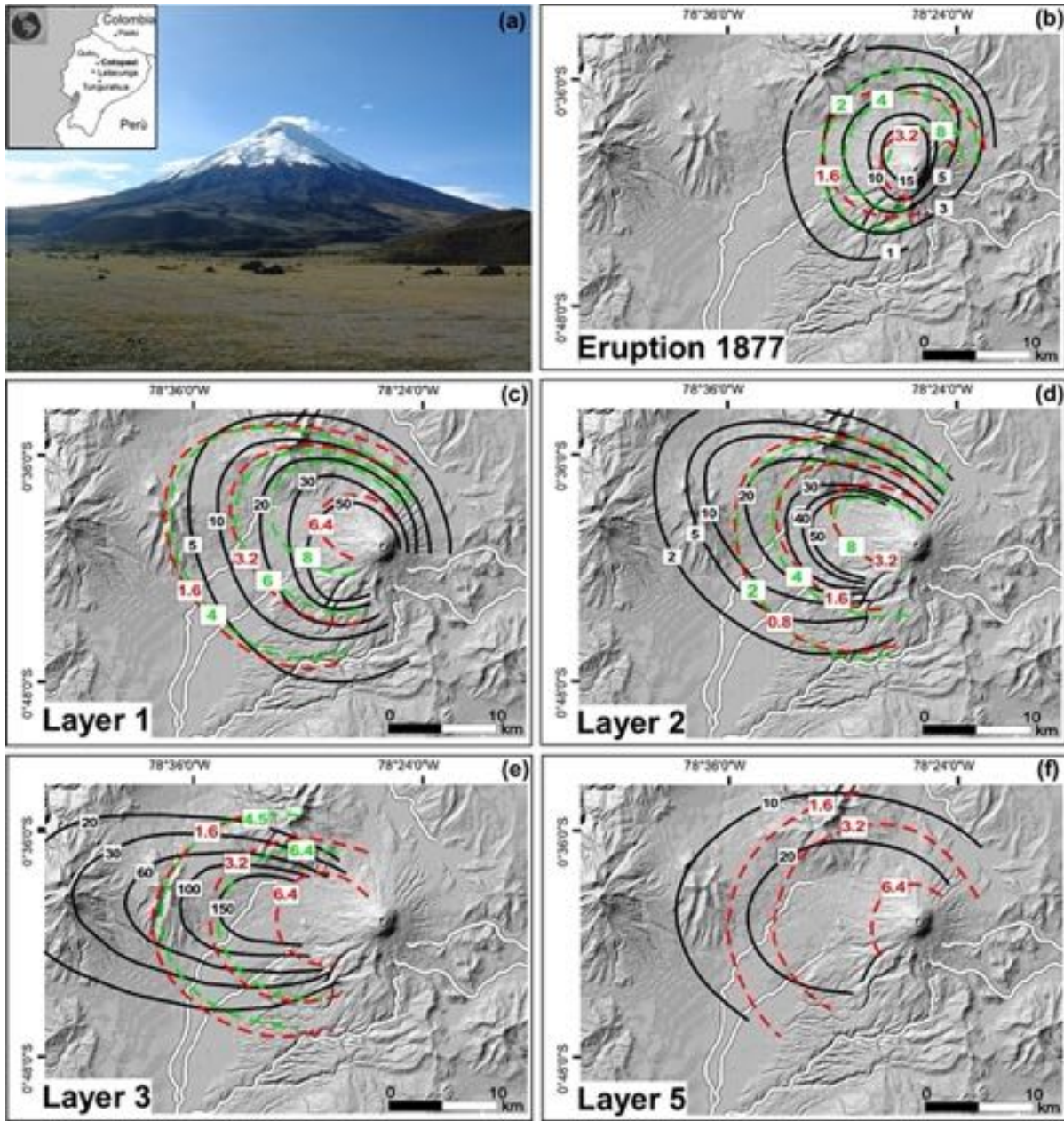
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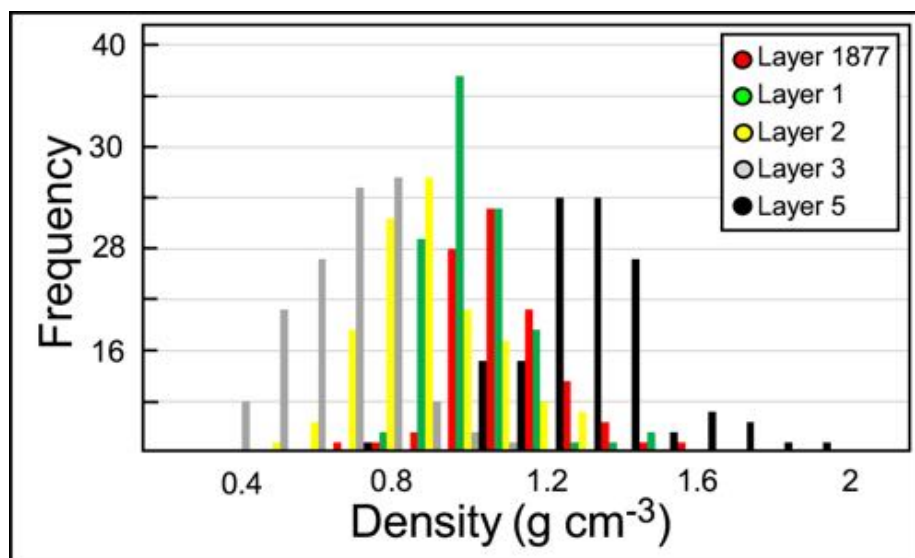
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**Figure 1.** (a) Cotopaxi volcano as seen from the north (localization in the inset). (b-f) Isopach (black lines) contours and isopleth (lithic: red dashed lines; pumice: green dashed lines) contours for the five targeted eruptions of Cotopaxi volcano, whose relief is presented in shaded maps. Data presented here are derived from Barberi et al. (1995), Pistolesi et al. (2011) and Biass and Bonadonna (2011).

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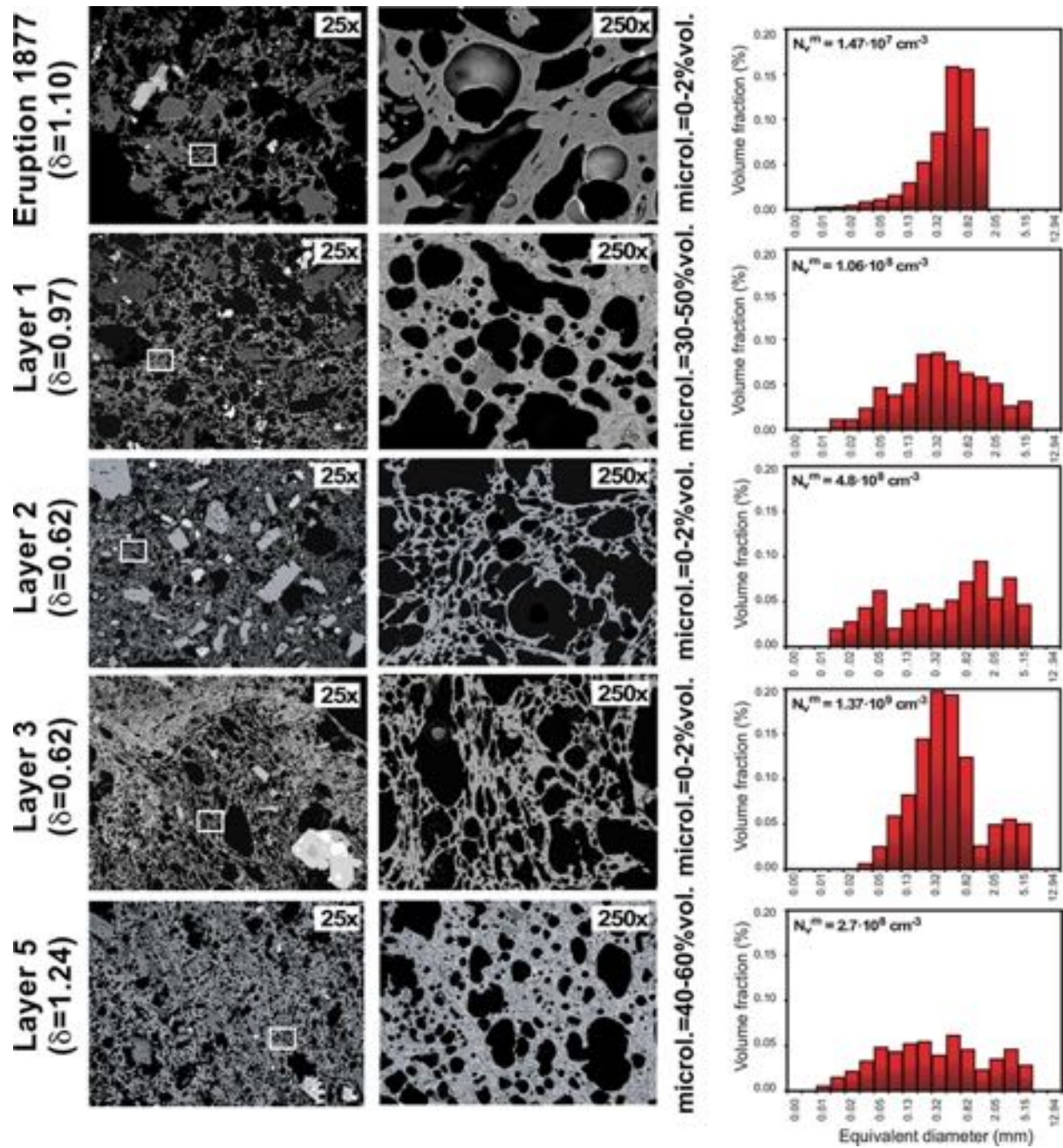
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**Figure 2.** Frequency distributions of bulk density values (gr cm<sup>-3</sup>) of juvenile fragments for the five selected eruptions (Eruption 1877, and Layers 1, 2, 3 and 5).



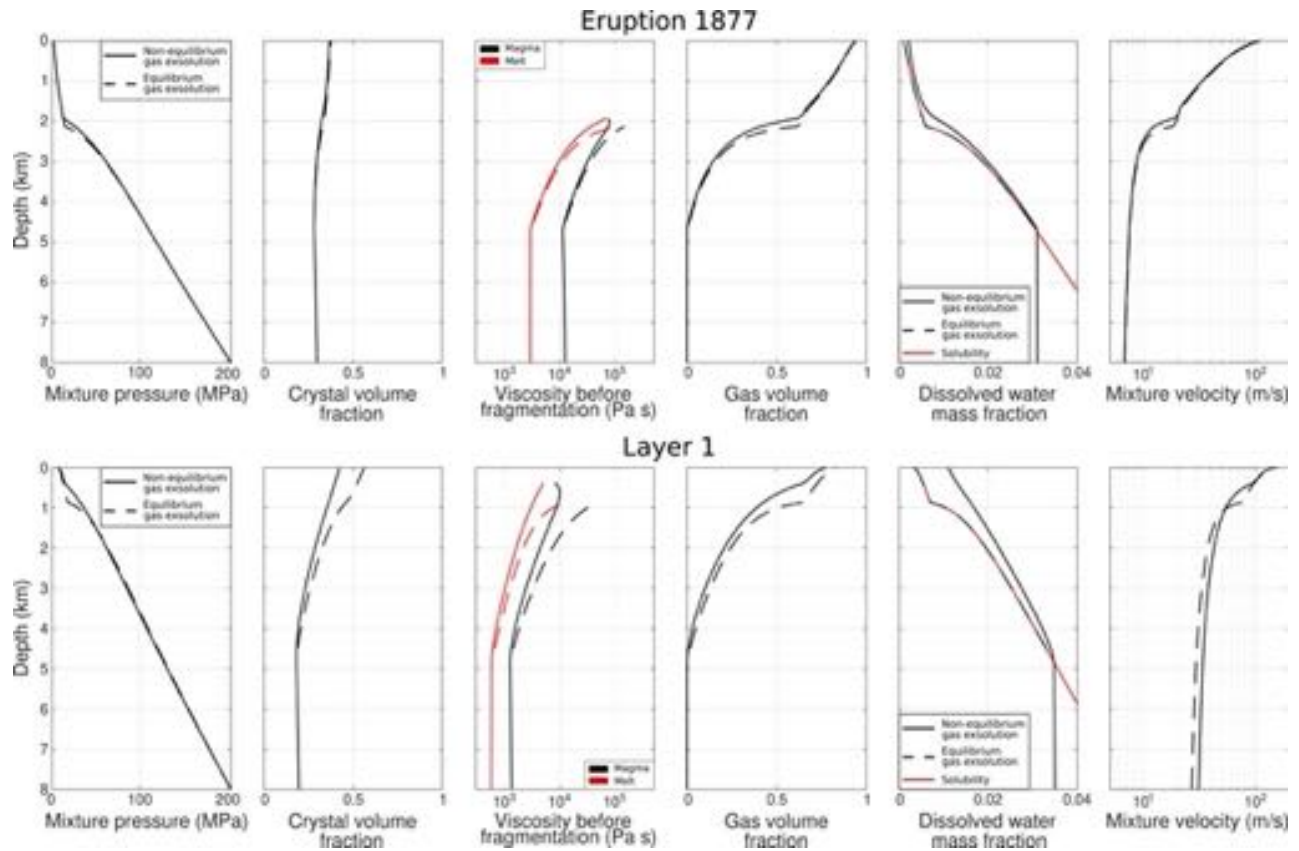
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1175 **Figure 3.** Backscattered Scanning Electron Microscope images of bubble textures and bubble

1176 size distribution from Cotopaxi's Eruption 1877, Layer 1, Layer 2, Layer 3 and Layer 5.

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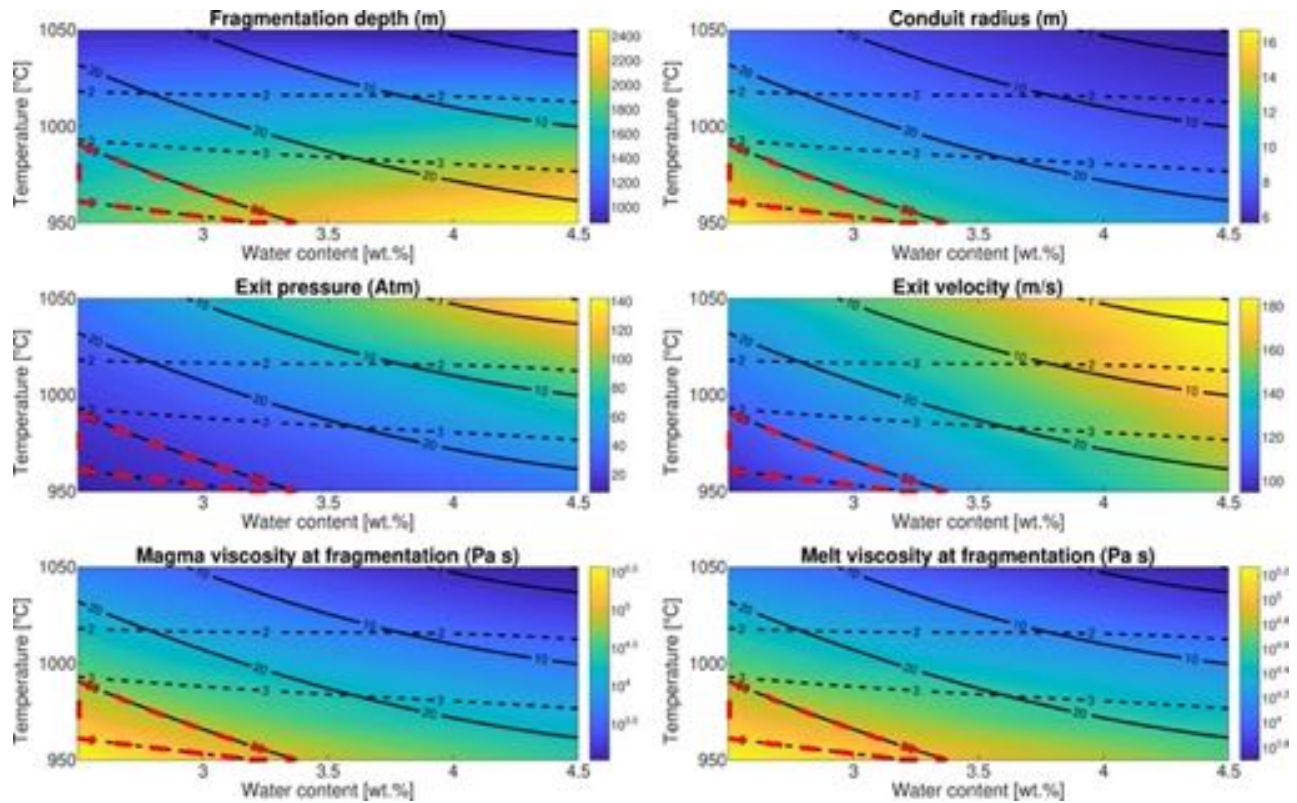


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 1180 **Figure 4.** Profiles along the conduit of key physical parameters associated with four specific  
 1181 simulations. Upper panels: Eruption 1877 (equilibrium and non-equilibrium conditions for gas  
 1182 exsolution; water content: 3.1 wt. %; temperature: 950°C). Lower panels: Layer 1 (equilibrium and  
 1183 non-equilibrium conditions for gas exsolution; water content: 3.5 wt. %; temperature: 1000°C).  
 1184 Conduit radius was defined in order to reproduce the computed MER of each eruption (Table 1).  
 1185 Other input parameters are presented in Table 3.

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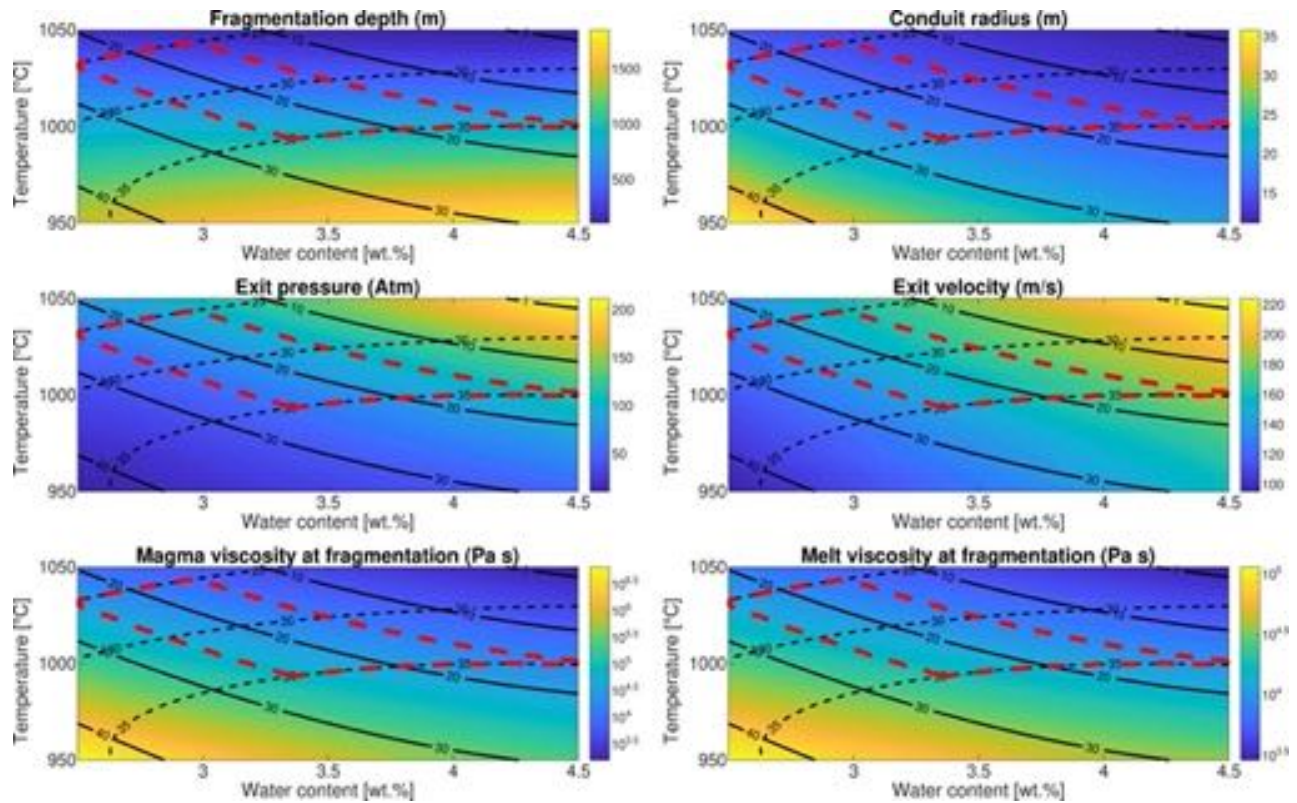
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1190 **Figure 5.** Summary of numerical results associated with Eruption 1877 considering fixed  
 1191 conditions for crystallization and gas exsolution relaxation parameters ( $\tau^{(c)} = 1000$  s and  $\tau^{(d)} =$   
 1192  $10^{-4}$  s). Other inputs are presented in Table 3. Each panel presents a color scale of different output  
 1193 parameters (fragmentation depth, conduit radius, exit pressure, exit velocity, magma viscosity at  
 1194 fragmentation and melt viscosity at fragmentation) as a function of initial temperature and water  
 1195 content. Two superposed contour maps indicate the volume fraction of phenocrysts (continuous  
 1196 lines) and the volume fraction of microlites (dashed lines). The area enclosed by red dashed lines is  
 1197 compatible with our estimates of the volume fraction of microlites and phenocrysts for this  
 1198 eruption, considering a range of tolerance of 5 vol. %.

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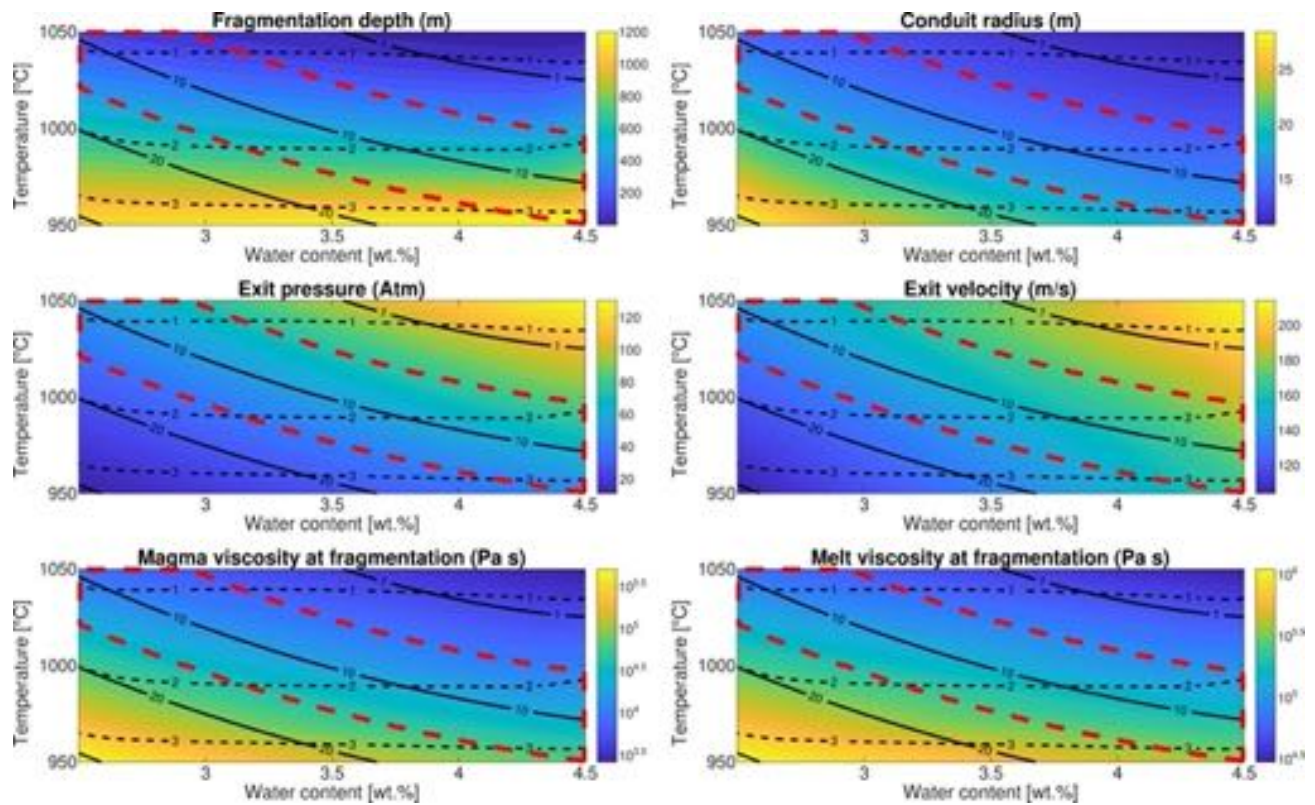


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**Figure 6.** Summary of numerical results associated with Layer 1 considering fixed conditions for crystallization and gas exsolution relaxation parameters ( $\tau^{(c)} = 10 \text{ s}$  and  $\tau^{(d)} = 10^{-4} \text{ s}$ ). Other inputs are presented in Table 3. Each panel presents a color scale of different output parameters (fragmentation depth, conduit radius, exit pressure, exit velocity, magma viscosity at fragmentation and melt viscosity at fragmentation) as a function of initial temperature and water content. Two superposed contour maps indicate the volume fraction of phenocrysts (continuous lines) and the volume fraction of microlites (dashed lines). The area enclosed by red dashed lines is compatible with our estimates of the volume fraction of microlites and phenocrysts for this eruption, considering a range of tolerance of 5 vol. %.

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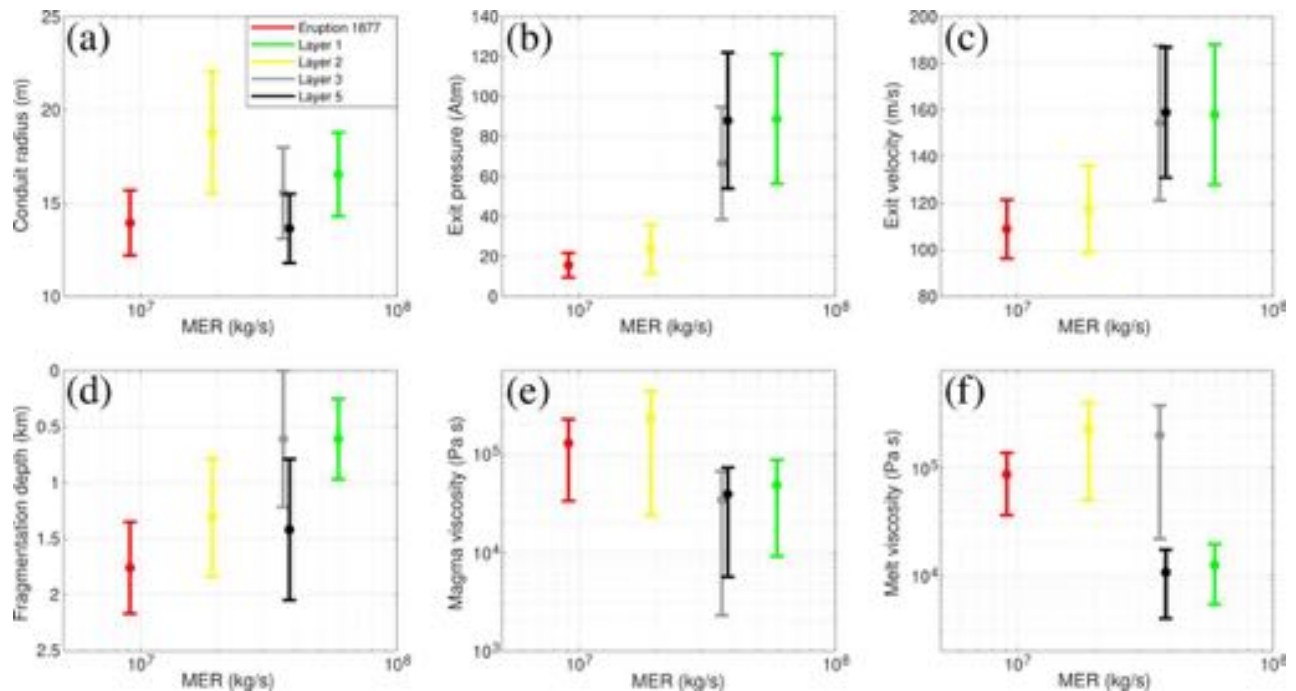


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1212 **Figure 7.** Summary of numerical results associated with Layer 3 considering fixed conditions for  
 1213 crystallization and gas exsolution relaxation parameters ( $\tau^{(c)} = 1000 \text{ s}$  and  $\tau^{(d)} = 10^{-4} \text{ s}$ ). Other  
 1214 inputs are presented in Table 3. Each panel presents a color scale of different output parameters  
 1215 (fragmentation depth, conduit radius, exit pressure, exit velocity, magma viscosity at fragmentation  
 1216 and melt viscosity at fragmentation) as a function of initial temperature and water content. Two  
 1217 superposed contour maps indicate the volume fraction of phenocrysts (continuous lines) and the  
 1218 volume fraction of microlites (dashed lines). The area enclosed by red dashed lines is compatible  
 1219 with our estimates of the volume fraction of microlites and phenocrysts for this eruption,  
 1220 considering a range of tolerance of 5 vol. %.

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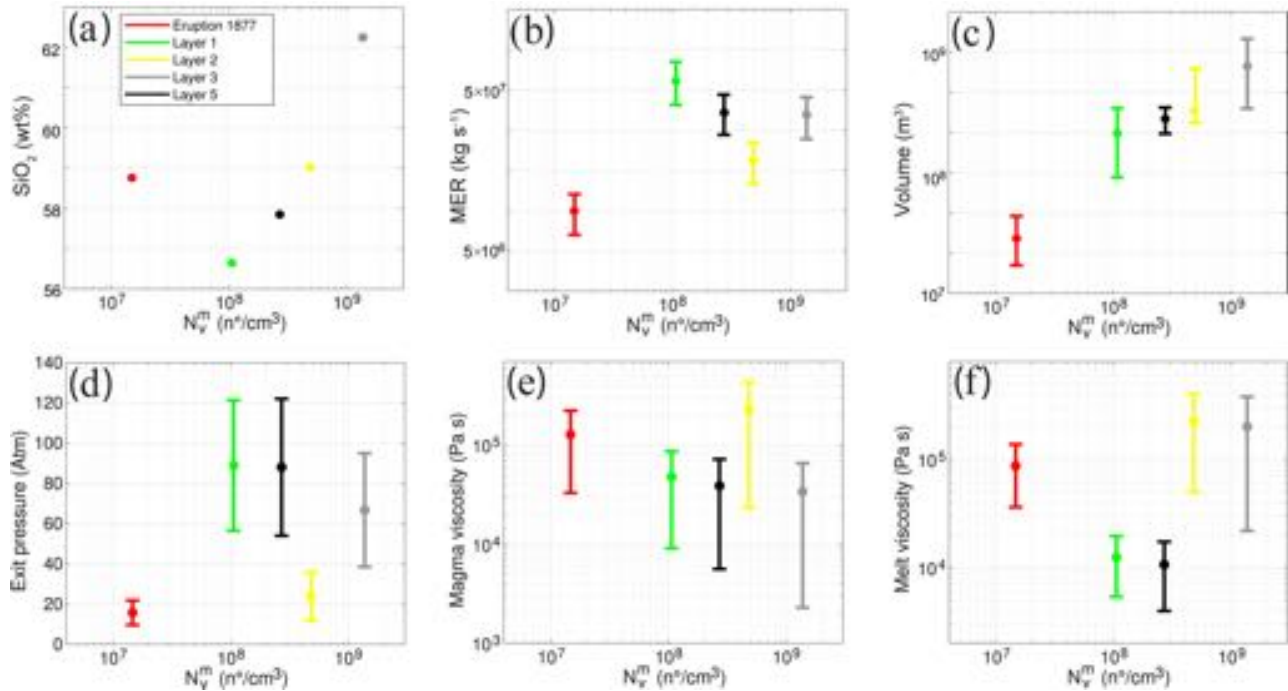
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1224 **Figure 8.** Relationships between MER (calculated from the inversion based on Rossi et al. (2019)  
 1225 and Degruyter and Bonadonna (2012) models; Table 1) of the five selected Cotopaxi eruptions and  
 1226 other key eruption parameters (conduit radius, exit pressure, exit velocity, fragmentation depth,  
 1227 magma viscosity and melt viscosity) derived from conduit modeling. The variability ranges of these  
 1228 parameters were defined considering exclusively the numerical simulations whose results are  
 1229 compatible with textural data (microlites and phenocrysts contents) derived from the studied  
 1230 volcanic products. In other words, we considered the areas enclosed by the red dashed lines in  
 1231 Figures 5-7 and S2-S8.

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1234 **Figure 9.** Relationships between Number of vesicles per unit volume ( $N_v^m$ ,  $\text{cm}^{-3}$ ) of the five  
 1235 selected Cotopaxi eruptions and other key eruption parameters (silica content, MER, volume, exit  
 1236 pressure, magma viscosity and melt viscosity) derived from conduit modeling. MER and volume  
 1237 are calculated from the inversion based on Rossi et al. (2019) and Degruyter and Bonadonna (2012)  
 1238 models, and from the average of Exponential, Power Law and Weibull fits, respectively (see Table  
 1239 1). The variability ranges of the parameters presented in the lower panels were defined considering  
 1240 exclusively the numerical simulations whose results are compatible with textural data (microlites  
 1241 and phenocrysts contents) derived from the studied volcanic products. In other words, we  
 1242 considered the areas enclosed by the red dashed lines in Figures 5-7 and S2-S8.

1243

1244 **Table 1.** Eruptive Source Parameters for the five selected eruptions.

Erup./ Layer	Volume (m <sup>3</sup> )				Average plume height above vent (km)		Average wind at tropopause (m/s)		Average Mass Eruption Rate – D&B12 (kg/s)		Average Mass Eruption Rate – Ma09 (kg/s)		Average Mass Eruption Rate – W&W87 (kg/s)		Duration (hours)
	<i>Exp.</i>	<i>PL</i>	<i>W</i>	<i>Inv.</i>	<i>C&amp;S86</i>	<i>RBD 19</i>	<i>C&amp;S86</i>	<i>RBD19</i>	<i>C&amp;S86</i>	<i>RBD19</i>	<i>C&amp;S86</i>	<i>RBD19</i>	<i>C&amp;S86</i>	<i>RBD19</i>	
<b>1877</b>	2.3×10 <sup>7a</sup>	4.9×10 <sup>7a</sup>	2.5×10 <sup>7</sup>	-	16.5 <sup>a*</sup>	15.2	6.4 <sup>a*</sup>	4.5	1.5×10 <sup>7</sup>	9.1×10 <sup>6</sup>	1.6×10 <sup>7</sup>	1.1×10 <sup>7</sup>	9.8×10 <sup>6</sup>	7.1×10 <sup>6</sup>	0.5
<b>1</b>	1.4×10 <sup>8a</sup>	4.0×10 <sup>8a</sup>	1.7×10 <sup>8</sup>	-	25.4 <sup>a*</sup>	20.3	17.5 <sup>a*</sup>	18.7	1.4×10 <sup>8</sup>	5.9×10 <sup>7</sup>	9.5×10 <sup>7</sup>	3.8×10 <sup>7</sup>	5.5×10 <sup>7</sup>	2.2×10 <sup>7</sup>	1.0
<b>2</b>	1.3×10 <sup>8a</sup>	8.6×10 <sup>8a</sup>	1.2×10 <sup>8</sup>	-	21.4 <sup>a*</sup>	16.1	15.5 <sup>a*</sup>	13.3	6.7×10 <sup>7</sup>	1.9×10 <sup>7</sup>	4.7×10 <sup>7</sup>	1.4×10 <sup>7</sup>	2.8×10 <sup>7</sup>	8.9×10 <sup>6</sup>	3.8
<b>3</b>	6.0×10 <sup>8b</sup>	1.5×10 <sup>9b</sup>	5.5×10 <sup>8</sup>	2.4×10 <sup>9b</sup>	26.8 <sup>b*</sup>	18.1	18.3 <sup>b*</sup>	16.0	1.9×10 <sup>8</sup>	3.6×10 <sup>7</sup>	1.2×10 <sup>8</sup>	2.3×10 <sup>7</sup>	6.8×10 <sup>7</sup>	1.4×10 <sup>7</sup>	4.8
<b>5</b>	2.9×10 <sup>8c</sup>	3.8×10 <sup>8c</sup>	2.3×10 <sup>8c</sup>	5.0×10 <sup>8b</sup>	25.6 <sup>b*</sup>	18.7	17.0 <sup>b*</sup>	15.3	1.5×10 <sup>8</sup>	3.8×10 <sup>7</sup>	9.8×10 <sup>7</sup>	2.7×10 <sup>7</sup>	5.7×10 <sup>7</sup>	1.6×10 <sup>7</sup>	2.1

1245 Exp.: Exponential strategy; PL; Power-Law strategy; W: Weibull strategy; Inv.: Inversion Strategy; C&S86: Carey and Sparks (1986); RBD19:  
1246 Rossi et al. (2019); D&B12: Degruyter and Bonadonna (2012); Ma09: Mastin et al. (2009); W&W87: Wilson and Walker (1987).  
1247 All values presented in this table are calculated as part of this work with the exception of: <sup>a</sup>, from Pistoiesi et al. (2011); <sup>b</sup>, from Biass and  
1248 Bonadonna (2011); <sup>c</sup>, from Biass et al. (2019). The values presented with the symbol “\*” indicate that the CW and DW ranges have been  
1249 calculated from the corresponding paper, but the plume height was calculated with the Matlab script of Biass et al. (2015).  
1250 Italic values in the exponential column for the calculation of volume indicates values derived with the method of Fierstein and Nathenson (1992)  
1251 for two segments. All the other values are calculated with the method of Pyle (1989) for one exponential segment.  
1252 Heights and winds at the tropopause calculated with C&S86 and RBD19 are averaged over all lithic contours associated with the average of the 3  
1253 axes of the 5 largest clasts (3.2 cm and 1.6 cm for Eruption 1877; 6.4 cm, 3.2 cm and 1.6 cm for Layer 1; 3.2 cm, 1.6 cm and 0.8 cm for Layer 2;  
1254 3.2 cm, 1.6 cm and 0.8 cm for Layer 3; 6.4 cm, 3.2 cm, 1.6 cm and 0.8 cm for Layer 5).  
1255 Height above vent = Height with C&S86 or RBD19 – (Vent height – Sampling height), where Vent height = 5.9 km and Sampling height = 3.0  
1256 km.

1257 Parameters used for the calculation of MER with D&B12 equation are: magmatic temperature (1223 K for Layer 3 and 1273 K for Eruption 1877,  
1258 Layer 1, Layer 2 and Layer 5), tropopause height (17 km), while wind is averaged across plume height.

1259 **Table 2.** Main physical parameters of tephra samples estimated from textural analyses. Whole-rock  
 1260 and glass matrix data are from Barberi et al. (1995) and from Pistolesi et al. (2011). Mass balance  
 1261 calculations were made with a dedicated spreadsheet using whole rock and groundmass glass  
 1262 analyses from Pistolesi et al. (2011) and mineral data from Saalfeld et al. (2019).

Parameters	Eruption 1877	Layer 1	Layer 2	Layer 3	Layer 5
Average clast density (g/cm <sup>3</sup> )	1.10	0.98	0.84	0.62	1.24
Vesicularity derived from density (%)	61.4	65.1	69.7	77.0	54.9
Vesicularity derived from image analysis (%)	38.3	41.4	53.8	57.9	36.6
Microlites content derived from image analysis (vol. %)	0	30	0	0	30
Phenocrysts reanalysed (vol. %)	35	20	30	10	20
Total crystal content derived from mass balance (wt. %)	33	40	35	16	28
Whole-rock SiO <sub>2</sub> (wt. %, normalized)	58.8	56.7	59.1	62.3	57.9
Glass matrix SiO <sub>2</sub> (wt. %)	62.6	59.5	64.4	64.9	60.4
N <sub>A</sub> (n°/cm <sup>2</sup> )	8.80×10 <sup>3</sup>	5.13×10 <sup>4</sup>	2.37×10 <sup>5</sup>	3.22×10 <sup>5</sup>	1.24×10 <sup>5</sup>
N <sub>V</sub> (n°/cm <sup>3</sup> )	5.56×10 <sup>6</sup>	3.70×10 <sup>7</sup>	1.49×10 <sup>8</sup>	3.15×10 <sup>8</sup>	1.16×10 <sup>8</sup>
N <sub>V</sub> <sup>m</sup> (n°/cm <sup>3</sup> )	1.47×10 <sup>7</sup>	1.06×10 <sup>8</sup>	4.80×10 <sup>8</sup>	1.37×10 <sup>9</sup>	2.70×10 <sup>8</sup>
Variation range of vesicles diameter (mm)	0.008 – 1.1	0.008 – 6	0.010 – 6	0.008 – 6	0.008 – 6

**Table 3.** Constitutive equations and main input parameters used in numerical simulations.

Constitutive Equation	Eruption 1877	Layer 1	Layer 2	Layer 3	Layer 5
Melt viscosity <sup>1</sup>	Giordano et al. (2008)	Giordano et al. (2008)	Giordano et al. (2008)	Giordano et al. (2008)	Giordano et al. (2008)
Effect of bubbles <sup>2</sup>	Costa et al. (2007)	Costa et al. (2007)	Costa et al. (2007)	Costa et al. (2007)	Costa et al. (2007)
Effect of crystals	Costa (2005)	Costa (2005)	Costa (2005)	Costa (2005)	Costa (2005)
Crystallization <sup>3</sup>	alphaMELTS calibration	alphaMELTS calibration	alphaMELTS calibration	alphaMELTS calibration	alphaMELTS calibration
Water solubility <sup>4</sup>	Henry's law	Henry's law	Henry's law	Henry's law	Henry's law
Outgassing <sup>5</sup>	Forchheimer's law	Forchheimer's law	Forchheimer's law	Forchheimer's law	Forchheimer's law
Equations of state (exsolved gas)	Ideal gas	Ideal gas	Ideal gas	Ideal gas	Ideal gas
Equations of state (melt, crystals and pyroclasts) <sup>6</sup>	Mie-Gruneisen equations	Mie-Gruneisen equations	Mie-Gruneisen equations	Mie-Gruneisen equations	Mie-Gruneisen equations
Input parameters	Eruption 1877	Layer 1	Layer 2	Layer 3	Layer 5
Inlet pressure (MPa)	203.8	203.8	203.8	203.8	203.8
Conduit length (km)	8.0	8.0	8.0	8.0	8.0
Magma water content (wt. %)	2.5 – 4.5	2.5 – 4.5	2.5 – 4.5	2.5 – 4.5	2.5 – 4.5
Temperature (°C)	950 – 1050	950 - 1050	950 - 1050	950 - 1050	950 - 1050
Relaxation time for crystallization (s)	1000	10	1000	1000	10
Relaxation time for gas exsolution (s)	10 <sup>-4</sup> - 10	10 <sup>-4</sup> - 10	10 <sup>-4</sup> - 10	10 <sup>-4</sup> - 10	10 <sup>-4</sup> - 10
Mass discharge rate (kg/s)	9.1×10 <sup>6</sup>	5.9×10 <sup>7</sup>	1.9×10 <sup>7</sup>	3.6×10 <sup>7</sup>	3.8×10 <sup>7</sup>
Bubble number density (m <sup>-3</sup> )	1.47×10 <sup>13</sup>	1.06×10 <sup>14</sup>	4.80×10 <sup>14</sup>	1.37×10 <sup>15</sup>	2.70×10 <sup>14</sup>
Gas volume fraction for fragmentation <sup>7</sup>	61.4%	65.1%	69.7%	77.0%	54.9%

<sup>1</sup>Computed using literature-derived data of glass composition for each eruption (Costantini, 2010).<sup>2</sup>Costa et al. (2007) adopted a generalisation of Llewellyn and Manga (2005).<sup>3</sup>For each eruption, a set of alphaMELTS (Smith and Asimow, 2005) simulations was performed, considering literature-derived data to define the magma composition (bulk-rock values; Costantini, 2010) and variable values for water content ( $w$ ; 0.0 wt. % - 4.5 wt. %), pressure ( $p$ ; 1 – 4000 bar) and temperature ( $T$ ; 950°C – 1050°C). Equilibrium crystallinity ( $\beta_{eq}$ ) was fitted considering the following relationship:
$$\beta_{eq} = \max(0, \min(1, a_{T^2} \cdot T^2 + a_T \cdot T + a_{p^2} \cdot p^2 + a_p \cdot p + a_{w^2} \cdot w^2 + a_w \cdot w + a_{pw} \cdot p \cdot w + a_{Tw} \cdot T \cdot w + a_{Tp} \cdot T \cdot p + a_0)),$$

where  $T$ ,  $p$  and  $w$  are expressed in K, bar and mass fraction, respectively. Fit coefficients for each eruption are presented in Table S2.

<sup>4</sup>Fit derived from water solubility data on andesitic melts (Botcharnikov et al., 2015).<sup>5</sup>Following Degruyter et al. (2012).<sup>6</sup>Le Métayer et al. (2005).

<sup>7</sup>Assumed equal to the vesicularity estimates derived from density measurements (Table 2). This is because they are generally more reliable than estimates obtained directly from image analysis because of the uncertainty associated with the use of stereological models.

1265

**Table 4.** Main results derived from conduit numerical modeling.

<b>Parameter</b>	<b>Eruption 1877</b>	<b>Layer 1</b>	<b>Layer 2</b>	<b>Layer 3</b>	<b>Layer 5</b>
Conduit radius (m)	12 – 16	14 – 19	16 – 22	13 – 18	12 – 16
Fragmentation depth (km)	1.4 – 2.2	0.3 – 1.0	0.8 – 1.8	<1.2	0.8 – 2.1
Exit pressure (Atm)	10 – 20	60 – 120	10 – 40	40 – 100	50 – 120
Exit velocity (m/s)	100 – 120	130 – 190	100 – 140	120 – 190	130 – 190
Magma viscosity at fragmentation level (Pa s)	$3.3 \times 10^4 - 2.2 \times 10^5$	$9.1 \times 10^3 - 8.7 \times 10^4$	$2.3 \times 10^4 - 4.4 \times 10^5$	$2.3 \times 10^3 - 6.6 \times 10^4$	$5.6 \times 10^3 - 7.2 \times 10^4$
Melt viscosity at fragmentation level (Pa s)	$3.6 \times 10^4 - 1.4 \times 10^5$	$5.4 \times 10^3 - 1.9 \times 10^4$	$5.0 \times 10^4 - 4.1 \times 10^5$	$2.2 \times 10^4 - 3.8 \times 10^5$	$4.0 \times 10^3 - 1.7 \times 10^4$

1266

1267

Figure 1.

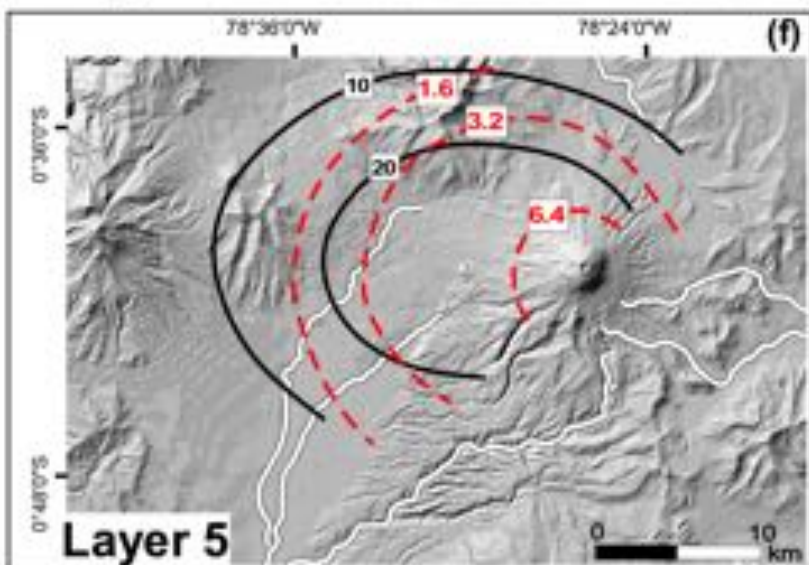
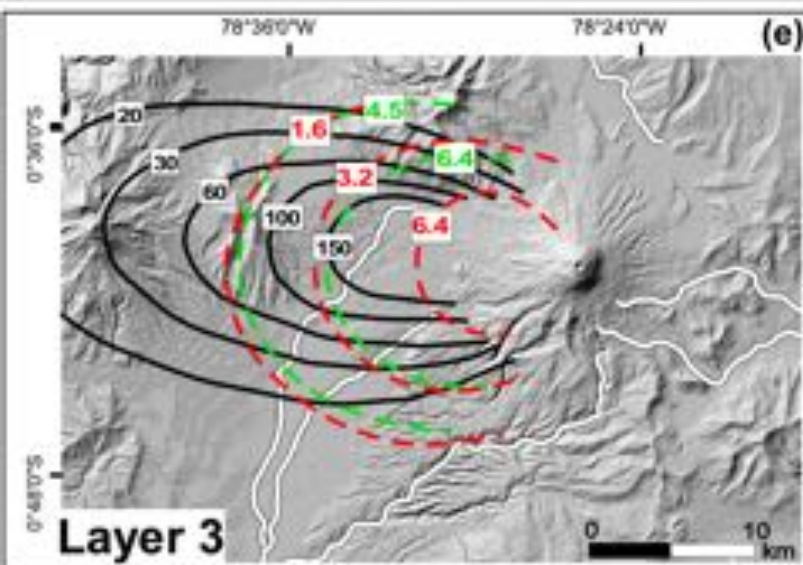
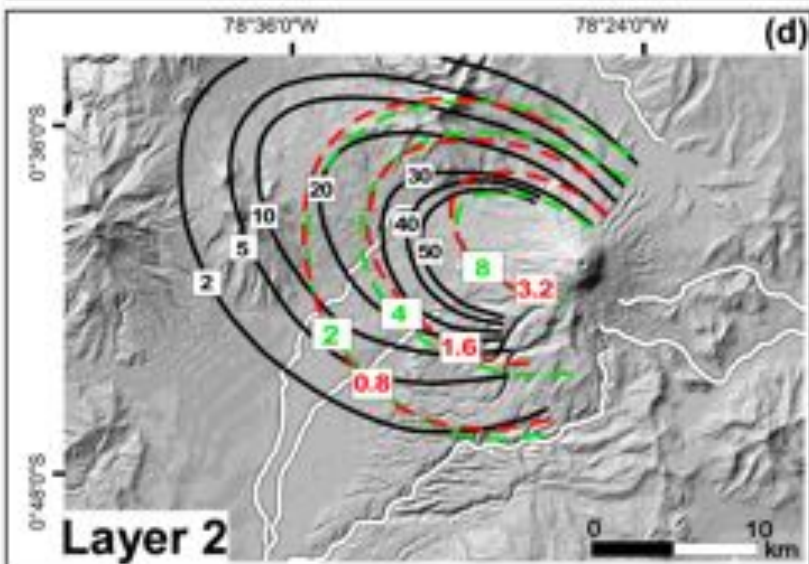
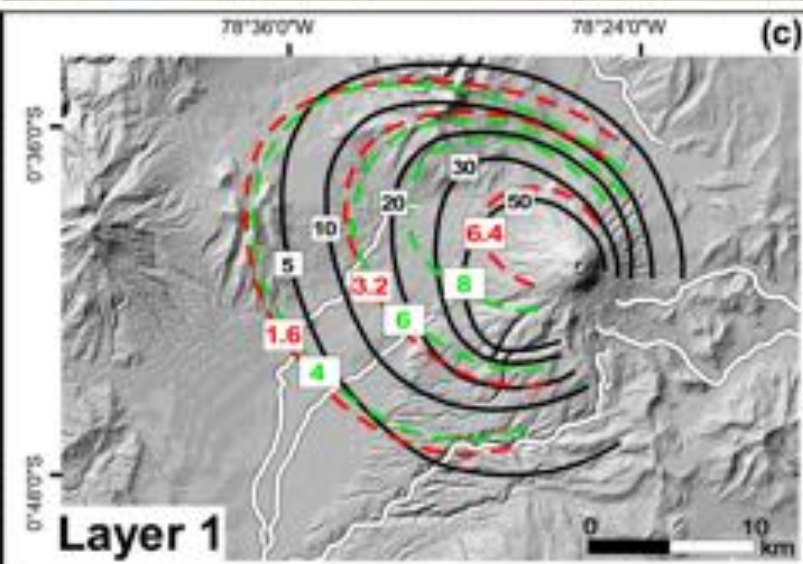
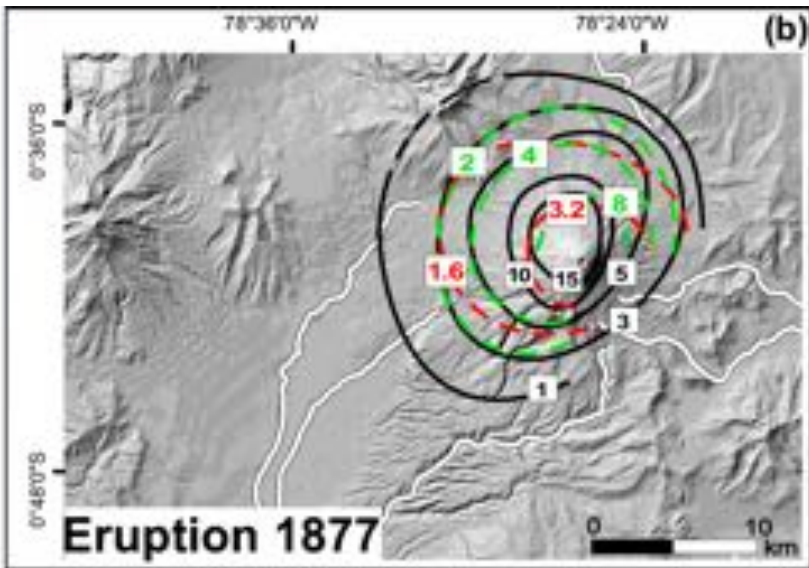
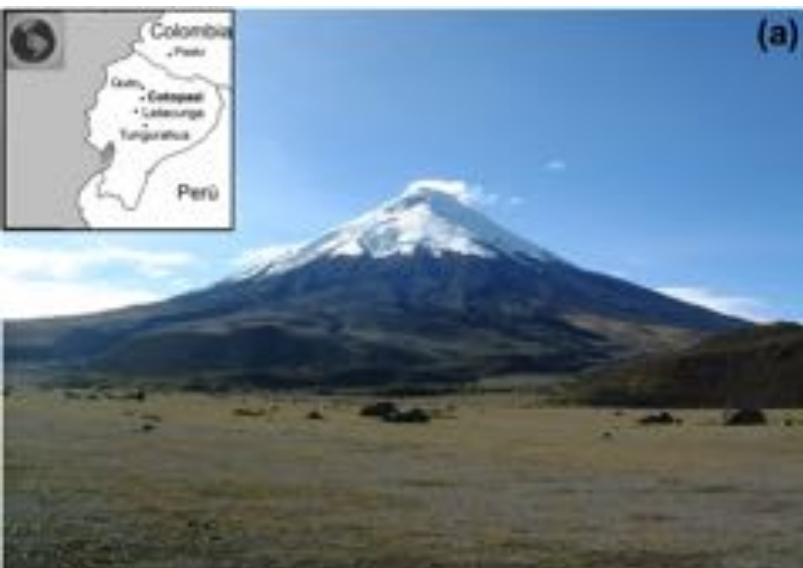


Figure 2.

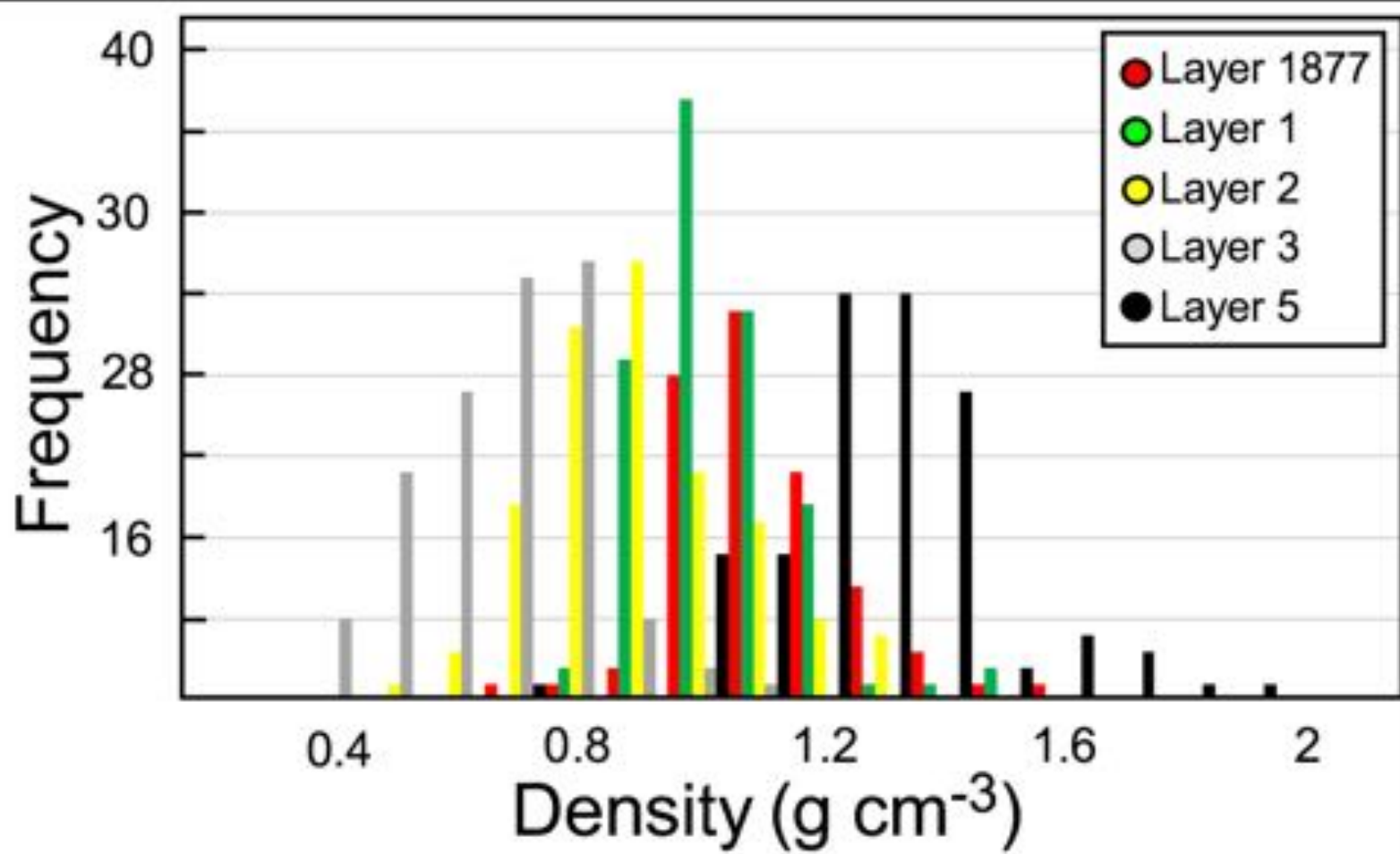
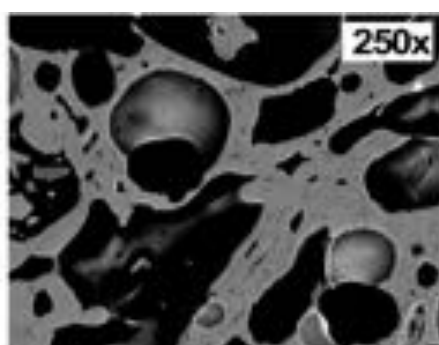
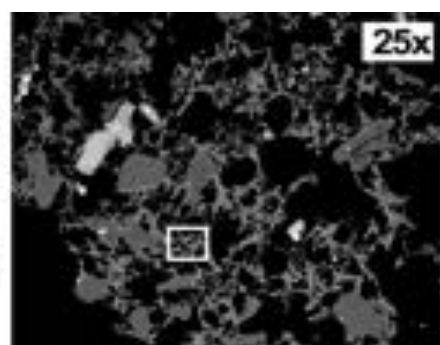
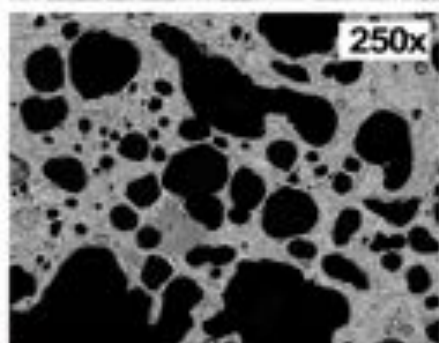
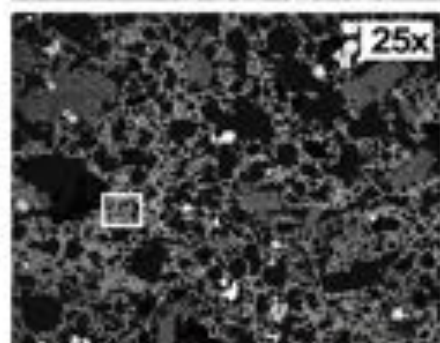


Figure 3.

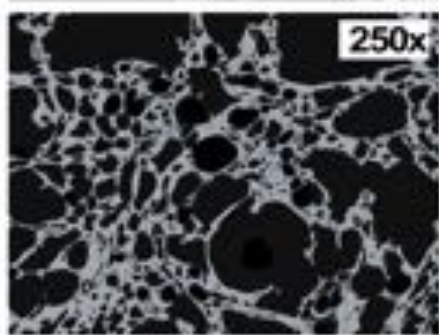
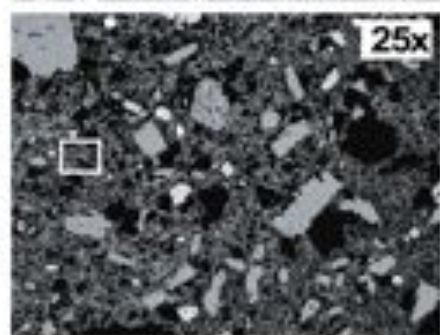
Eruption 1877  
( $\delta=1.10$ )



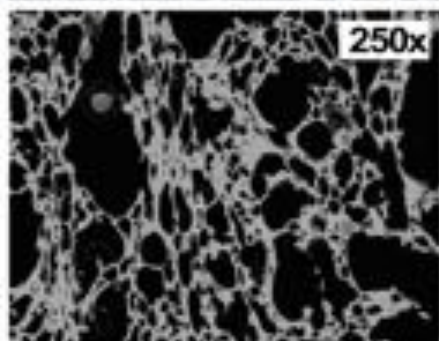
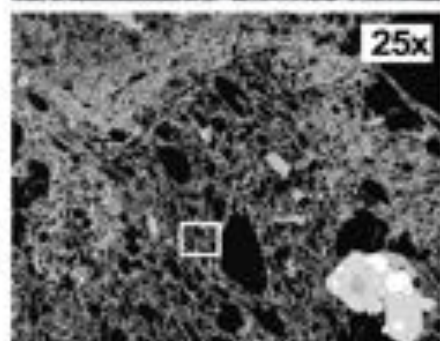
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( $\delta=0.97$ )



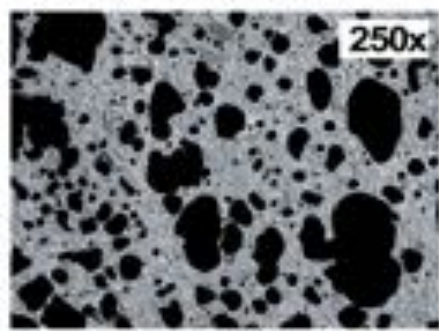
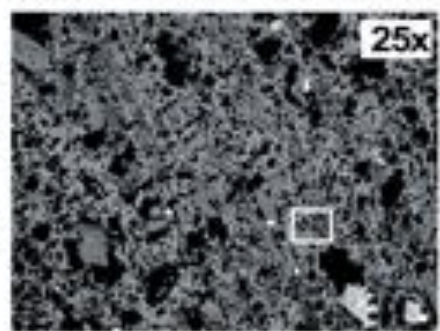
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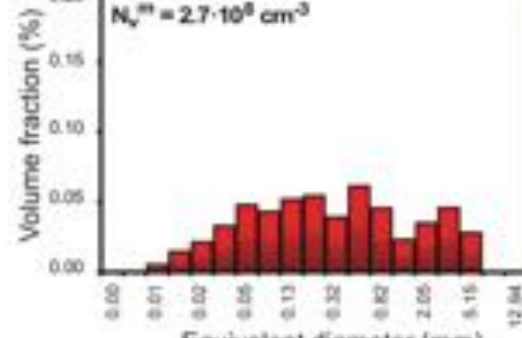
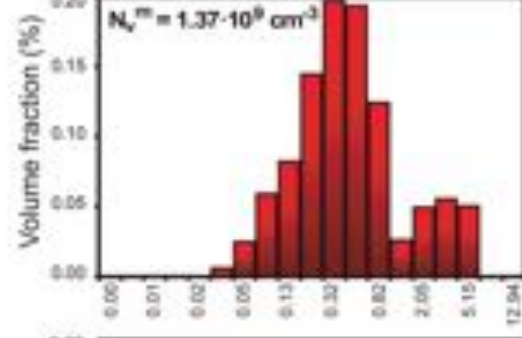
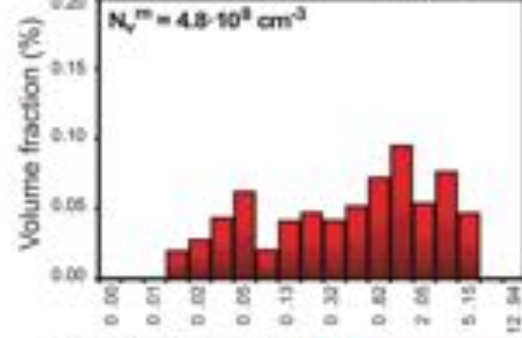
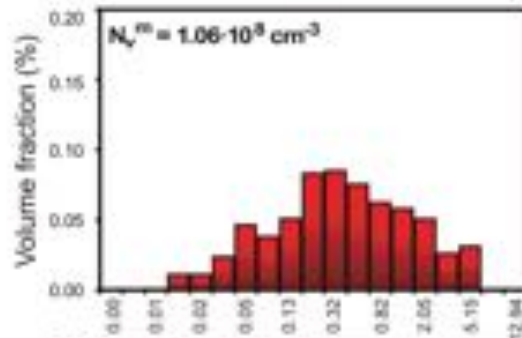
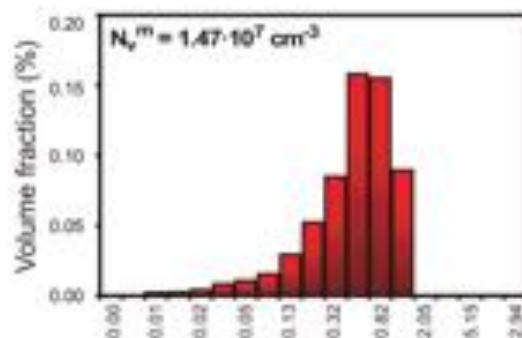
Layer 3  
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Layer 5  
( $\delta=1.24$ )

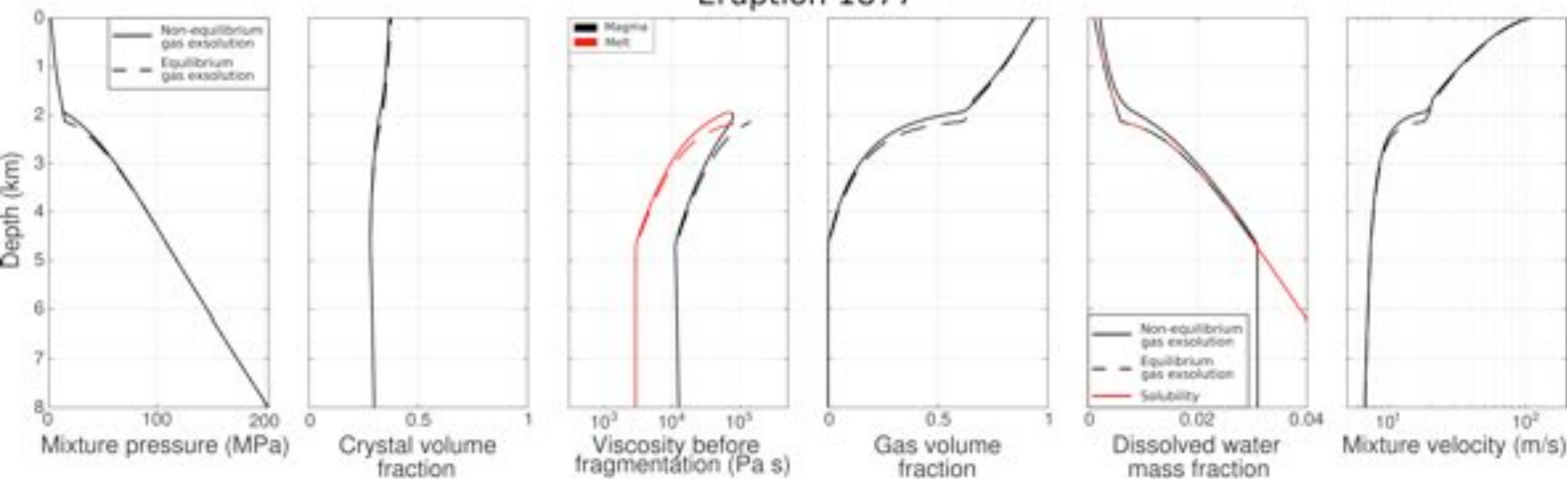


microl.=0-2%vol. microl.=30-50%vol. microl.=0-2%vol. microl.=0-2%vol. microl.=40-60%vol.



**Figure 4.**

## Eruption 1877



## Layer 1

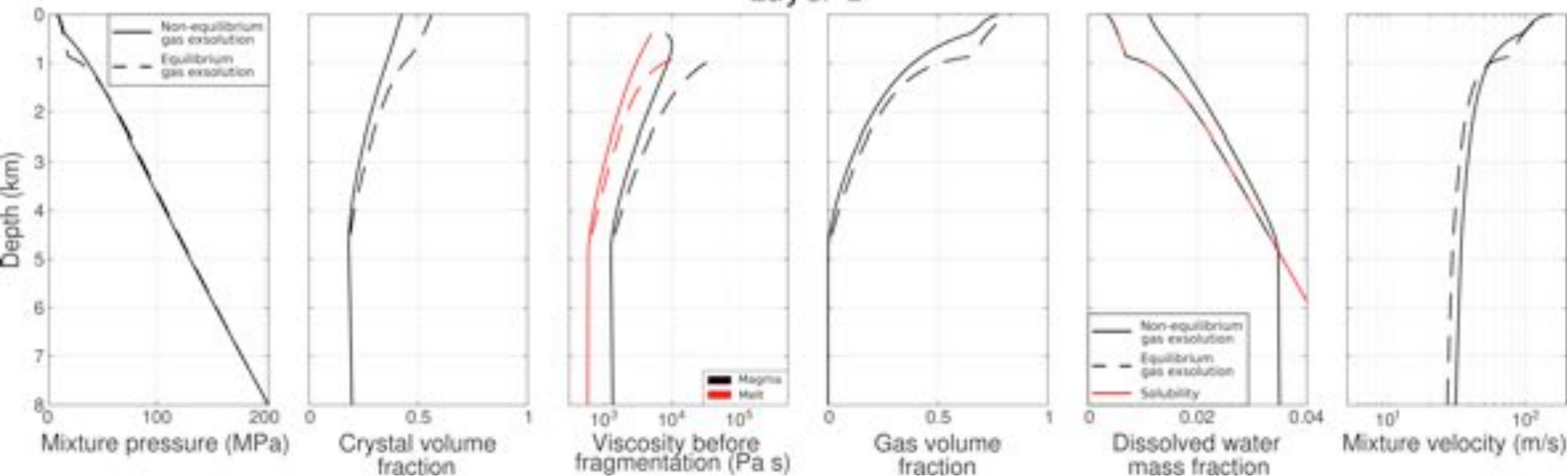


Figure 5.

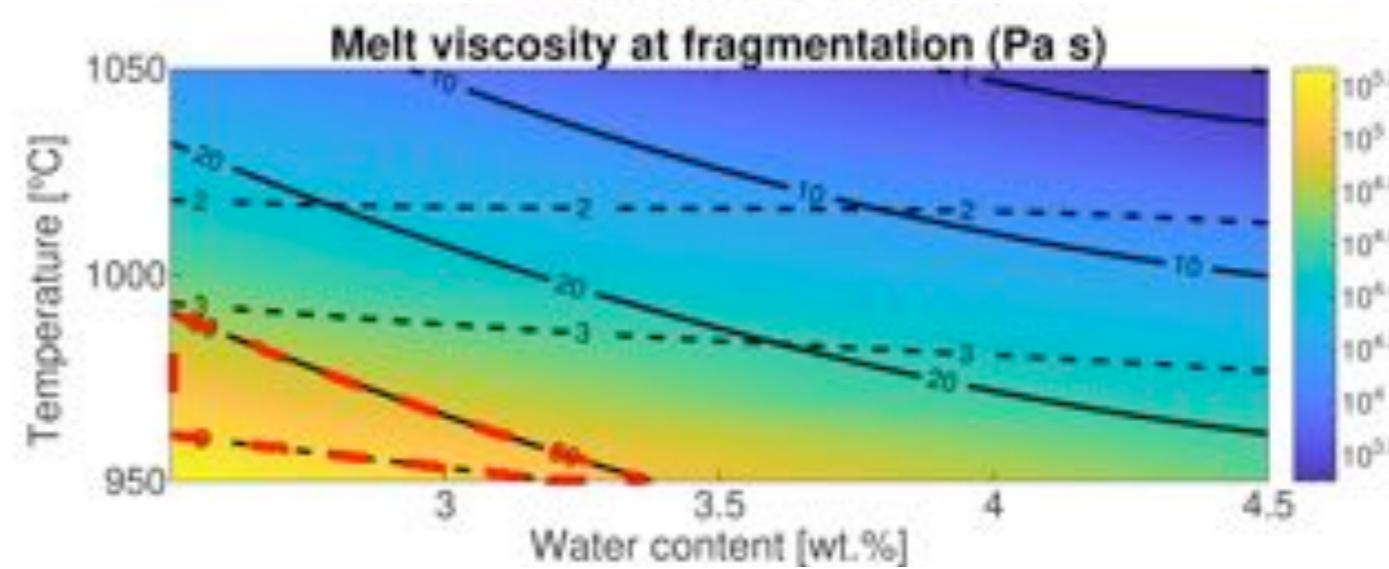
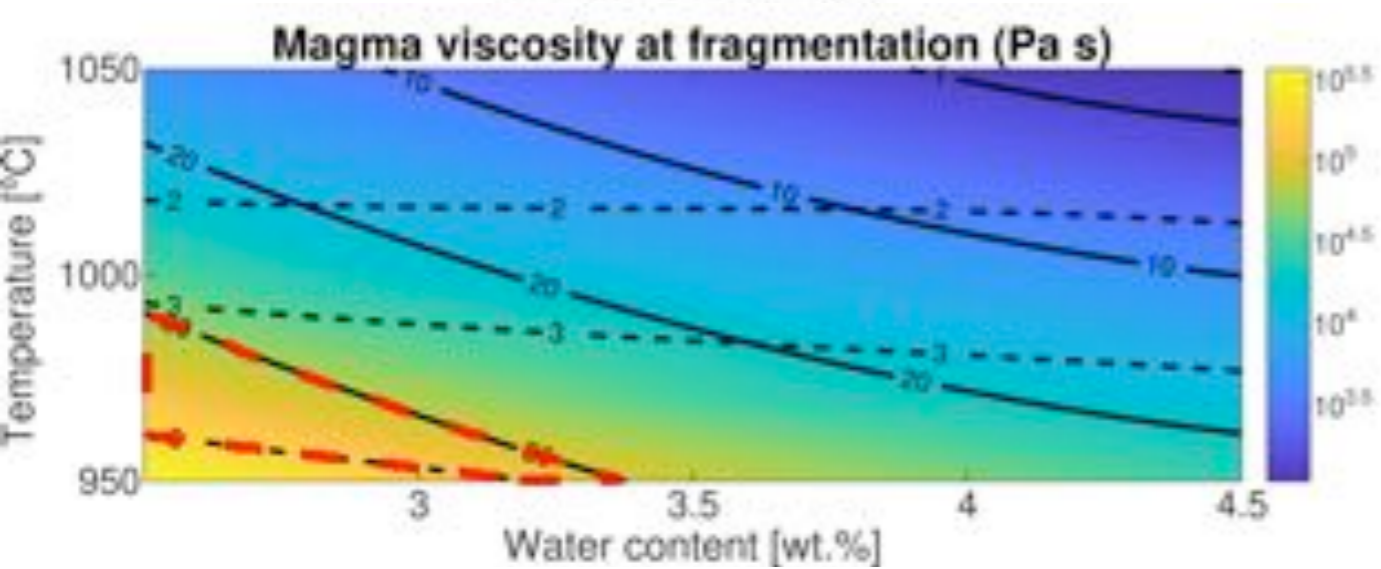
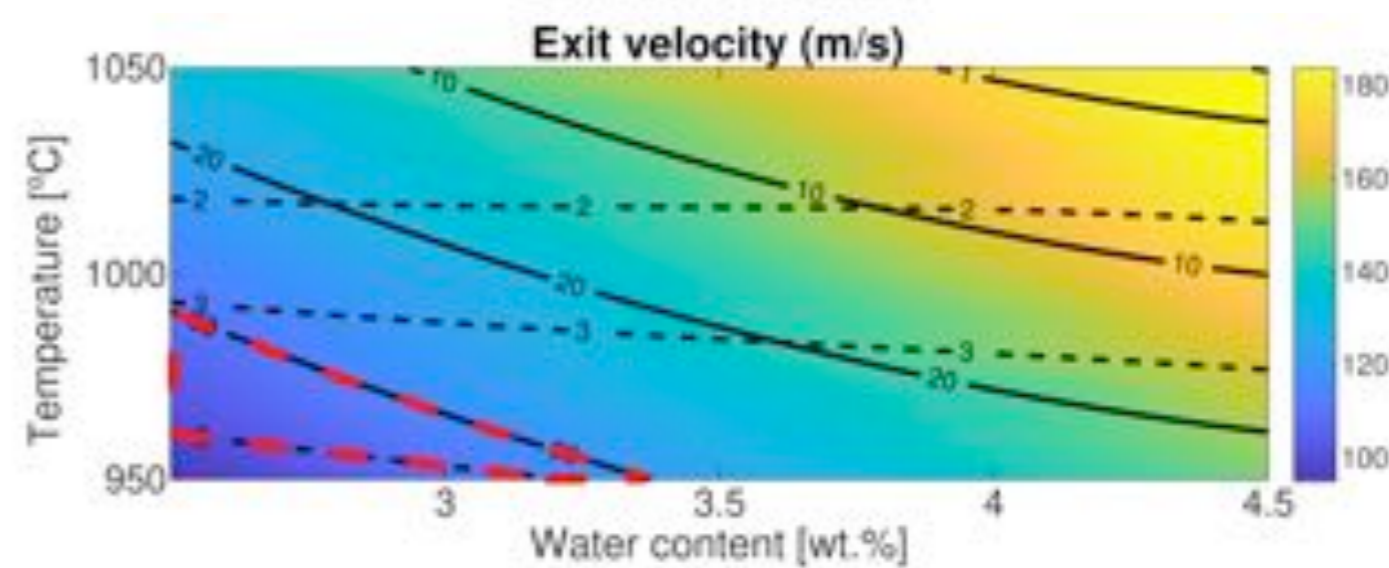
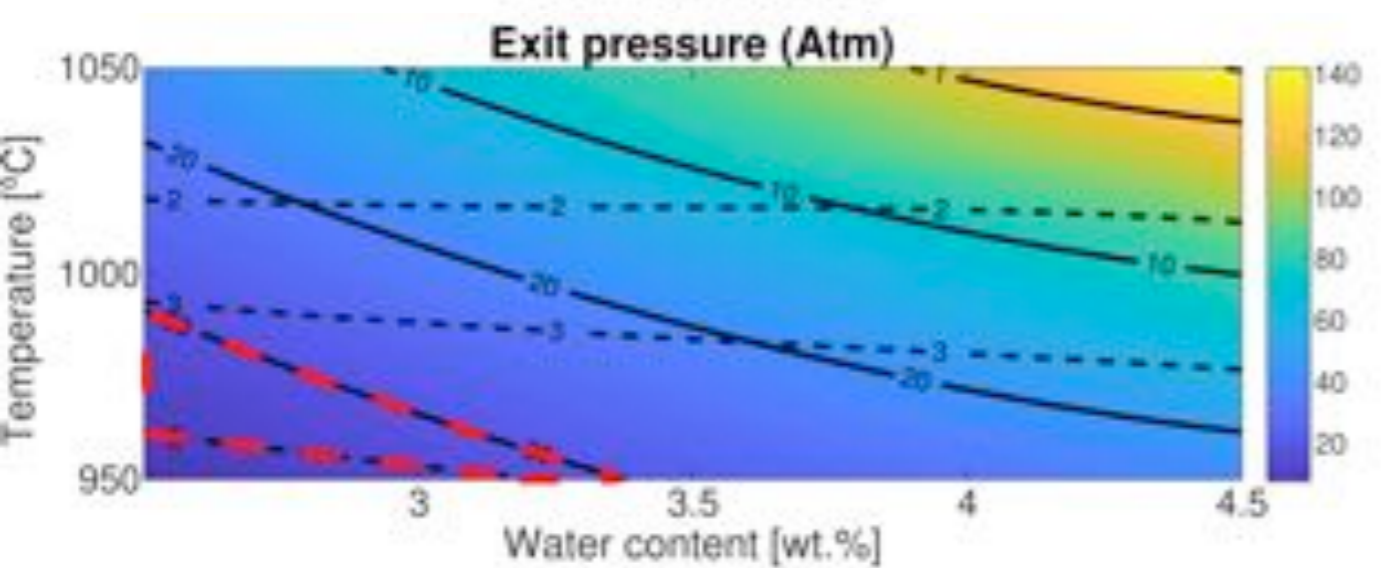
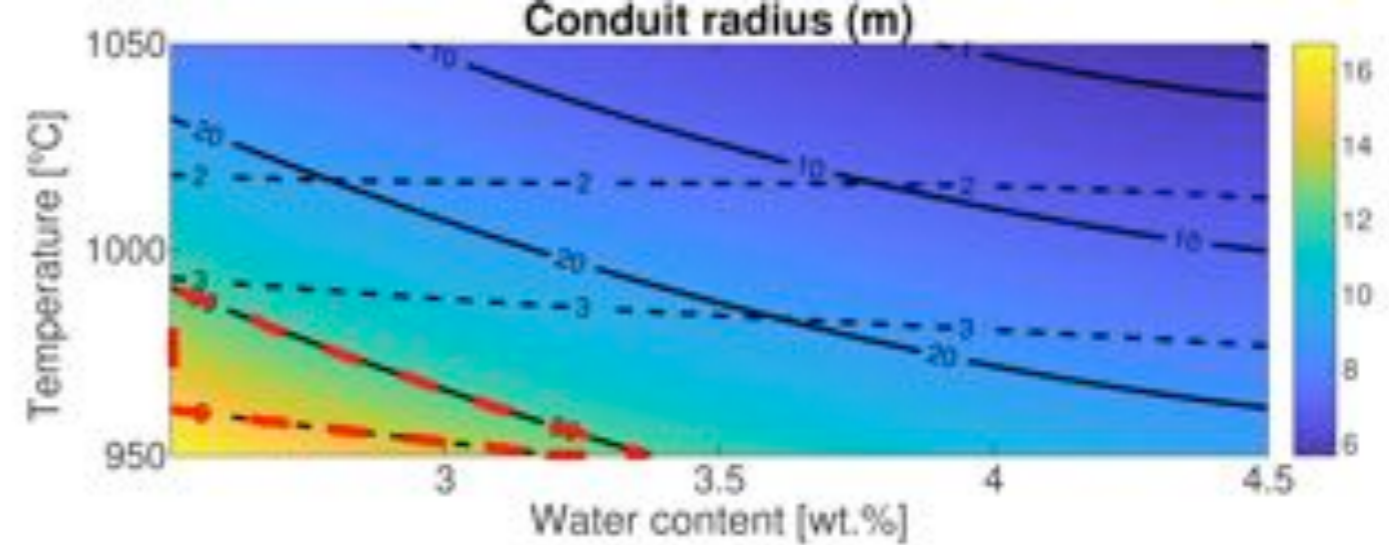
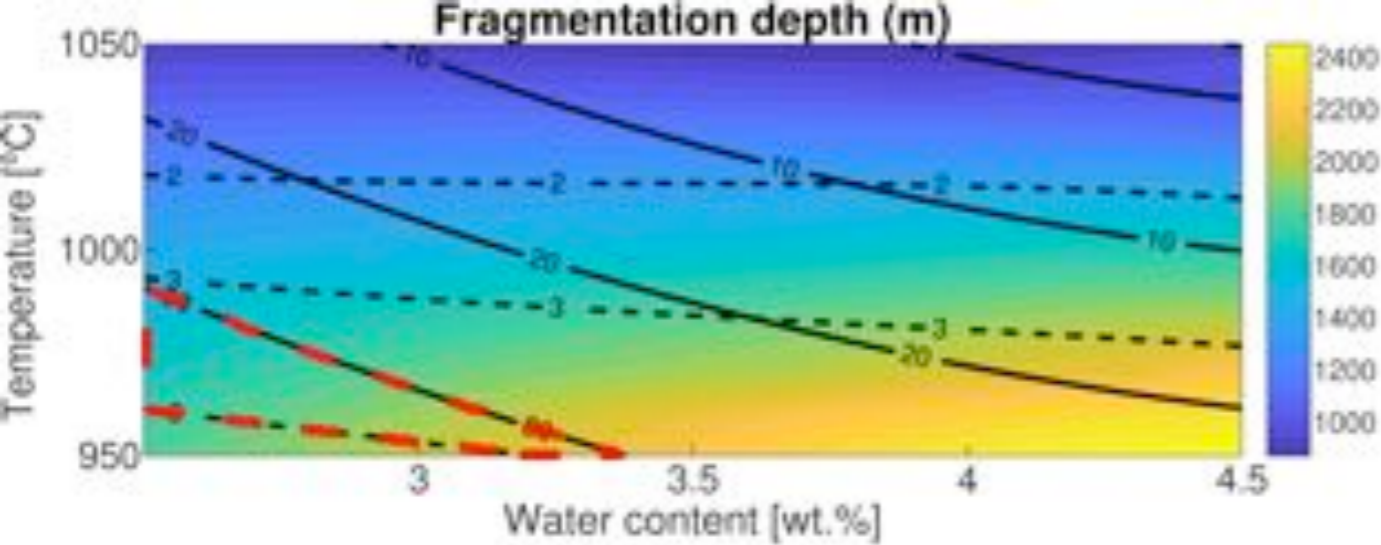
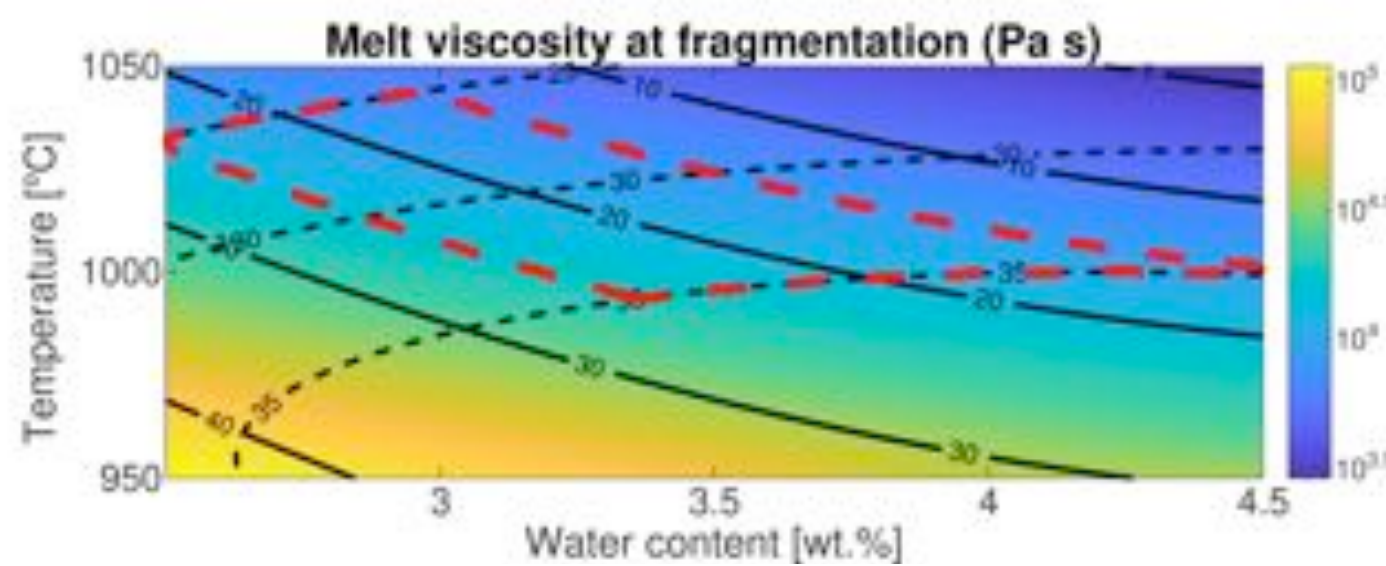
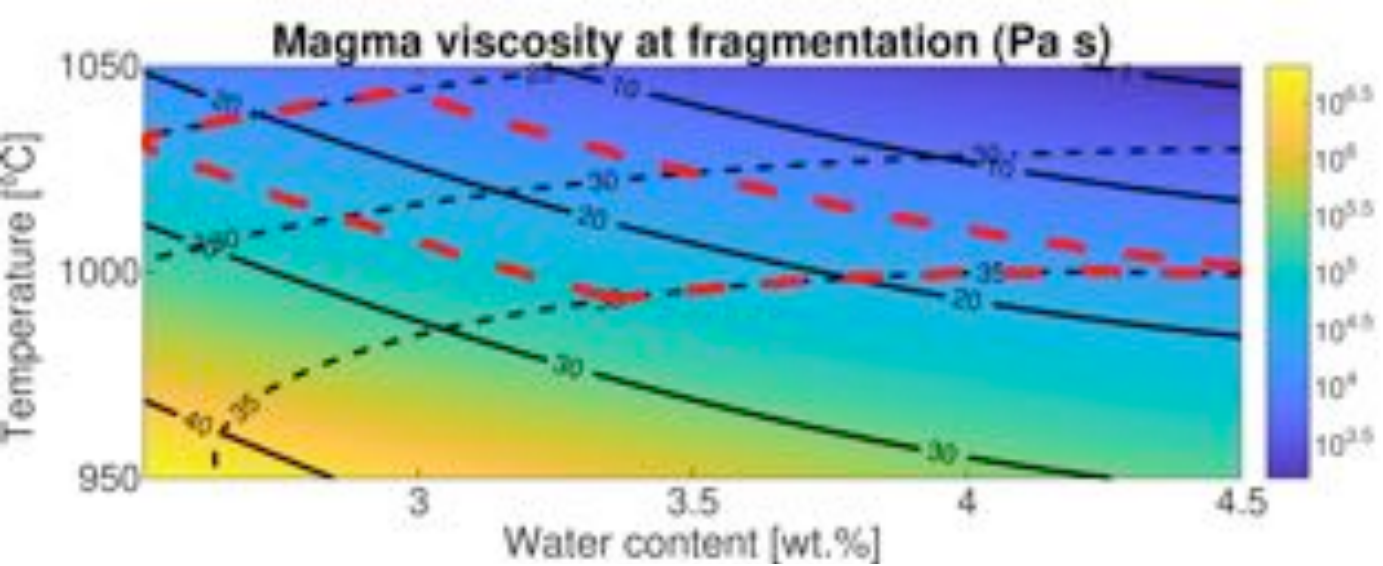
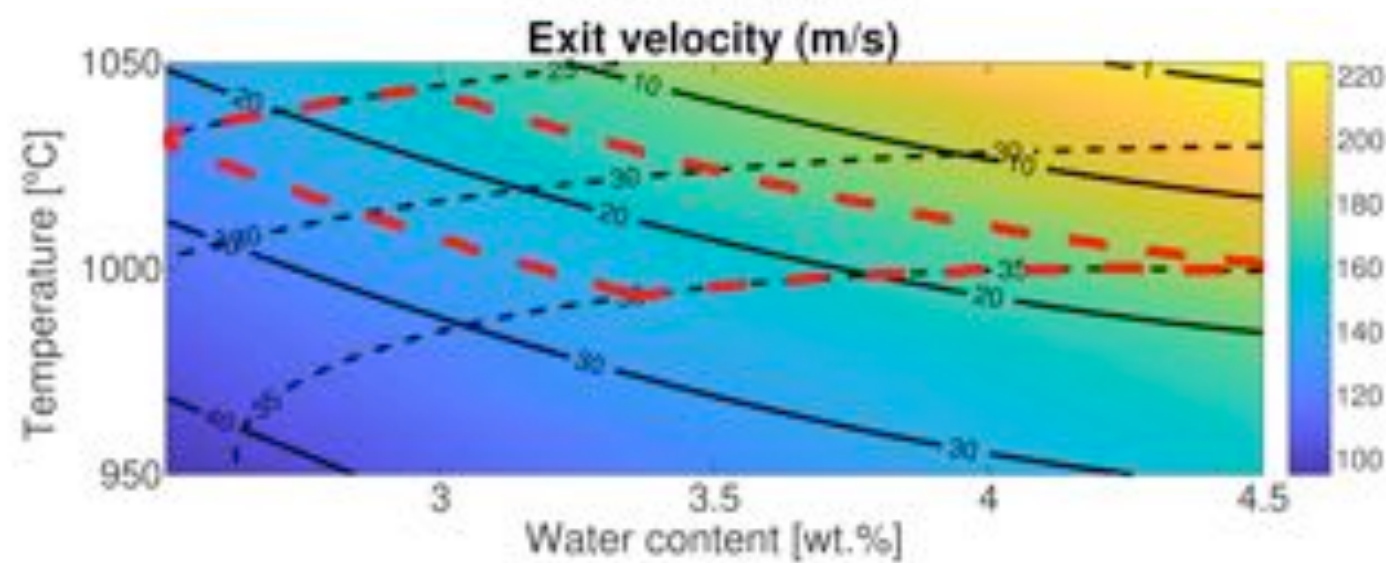
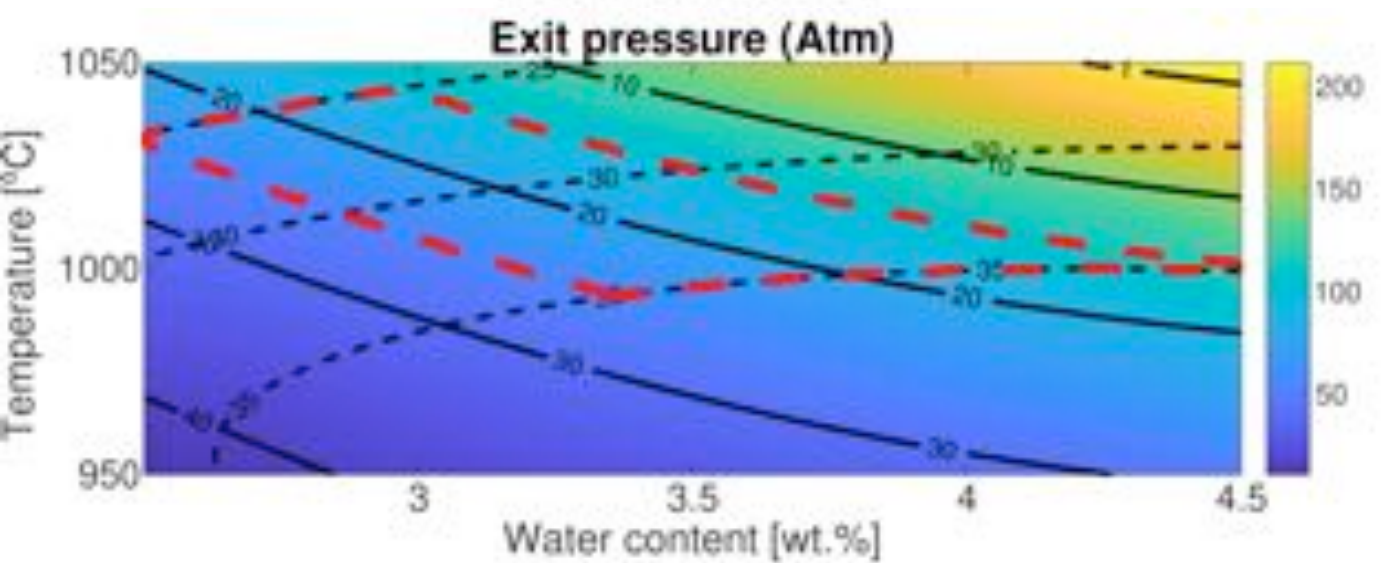
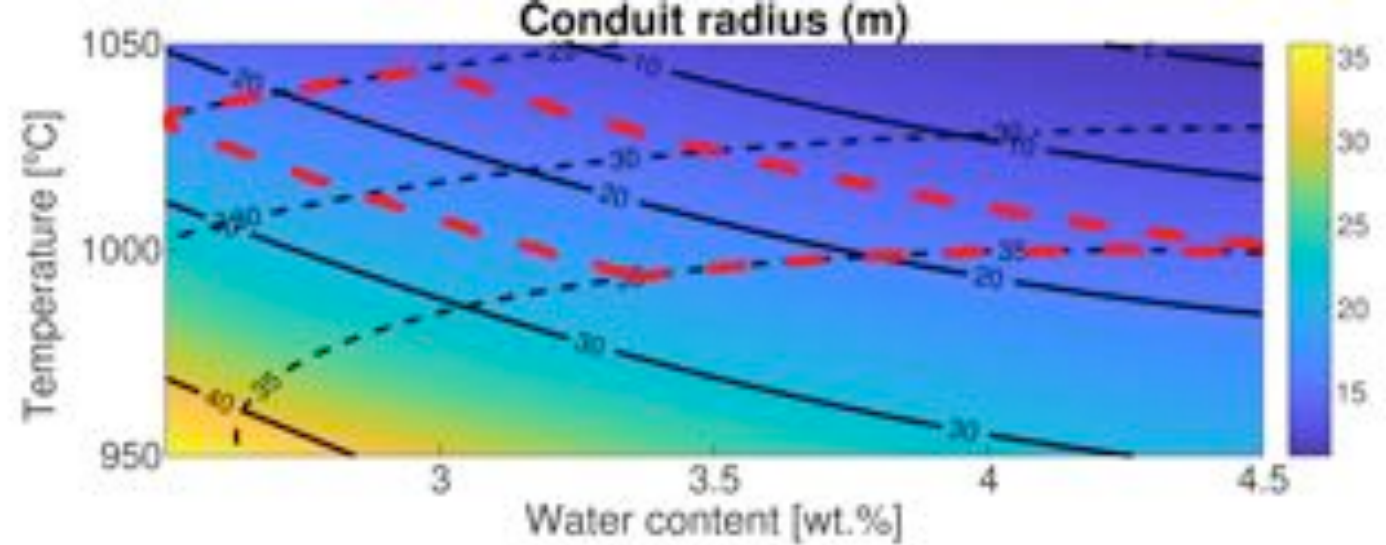
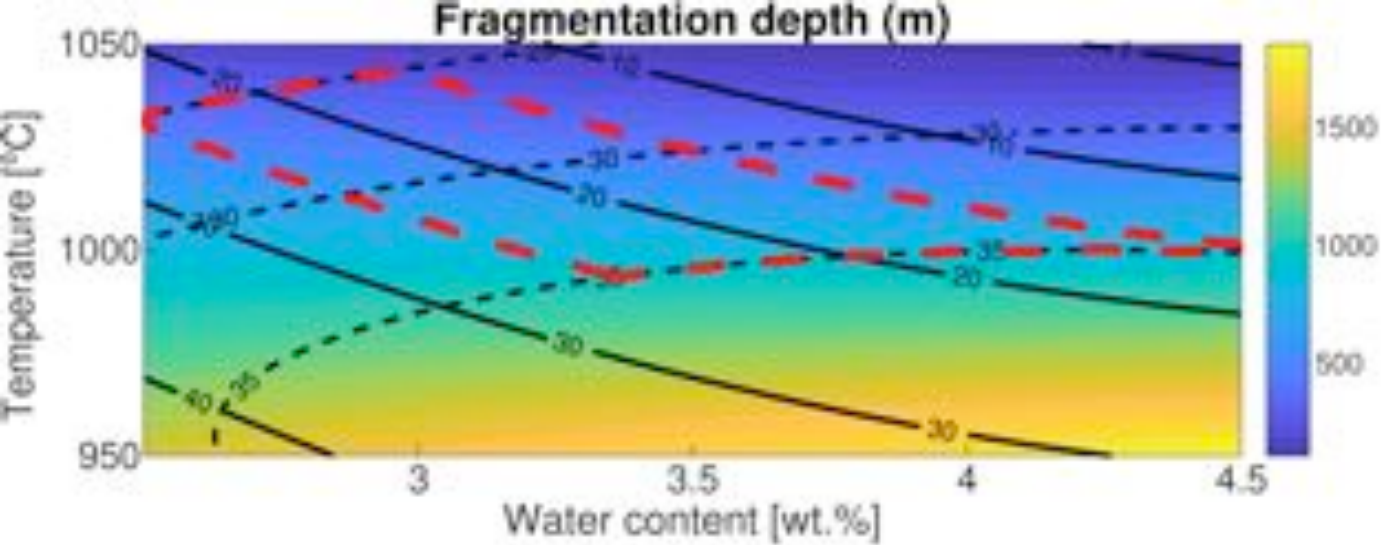


Figure 6.



**Figure 7.**

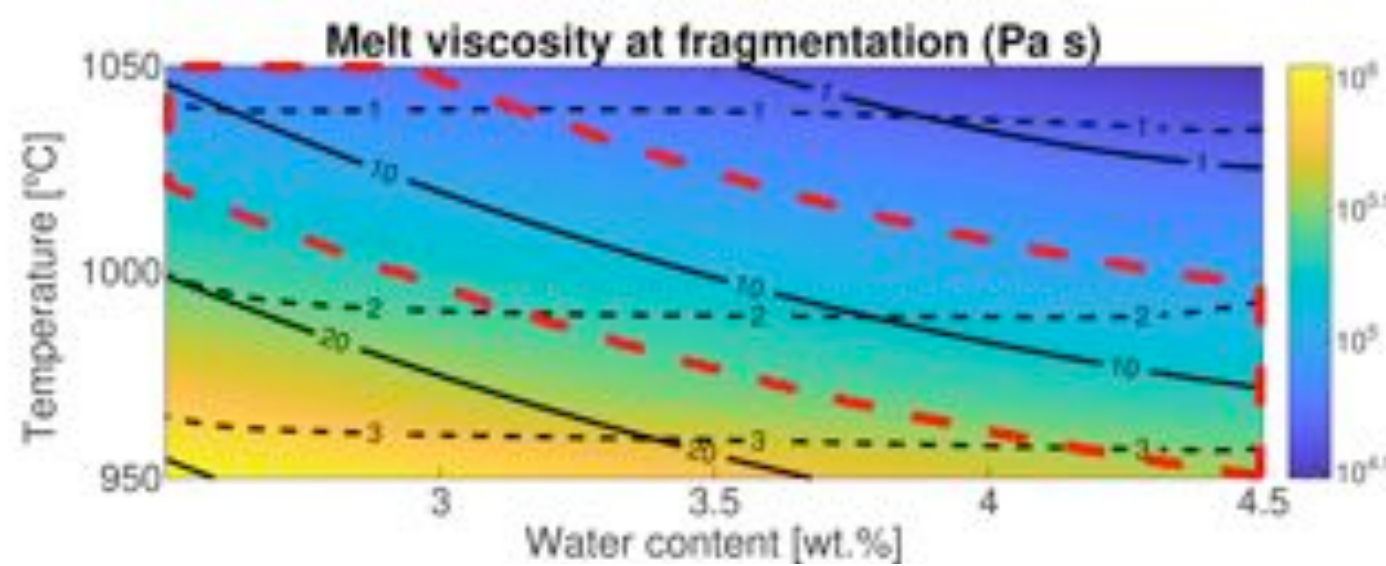
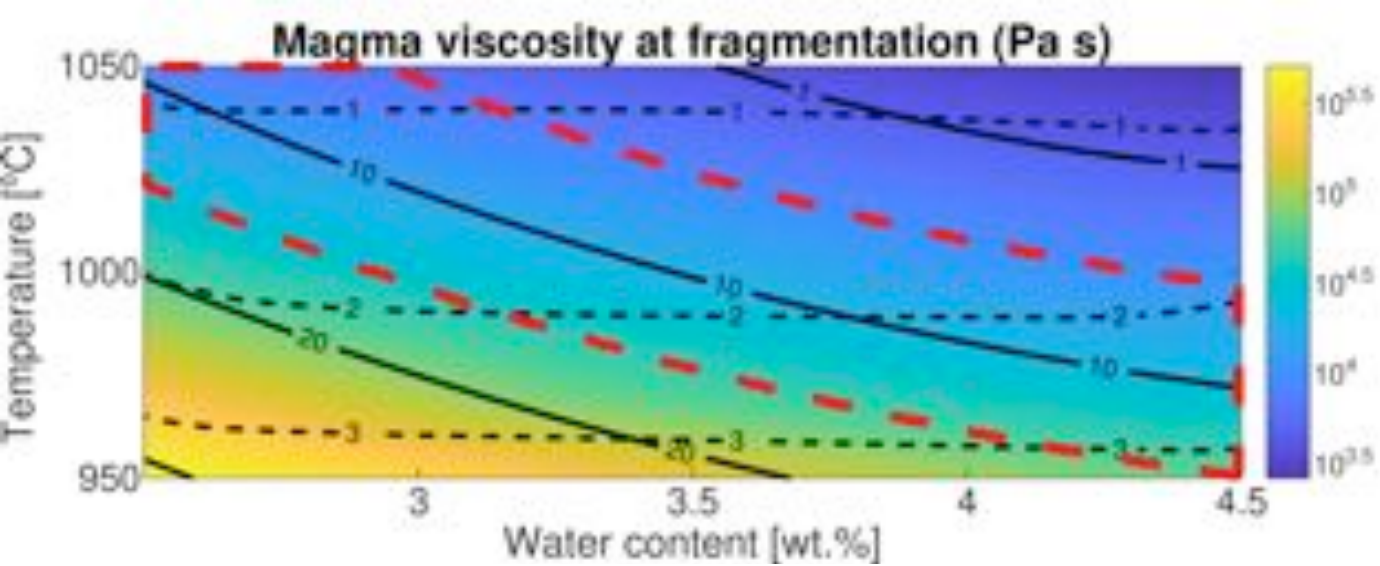
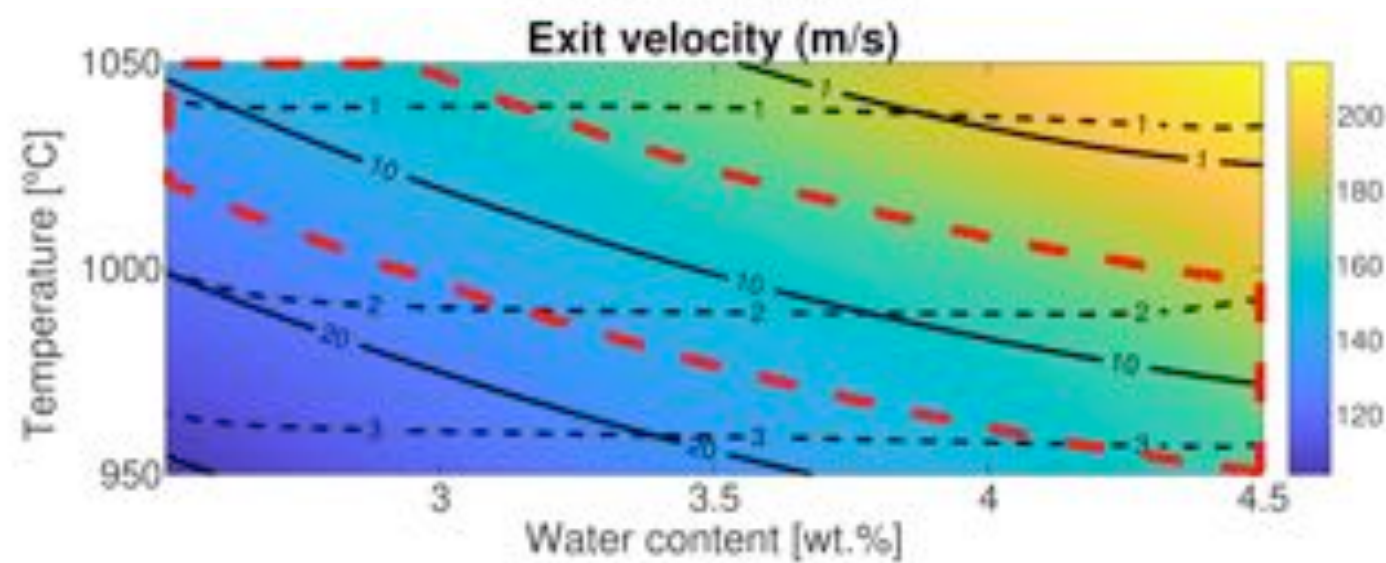
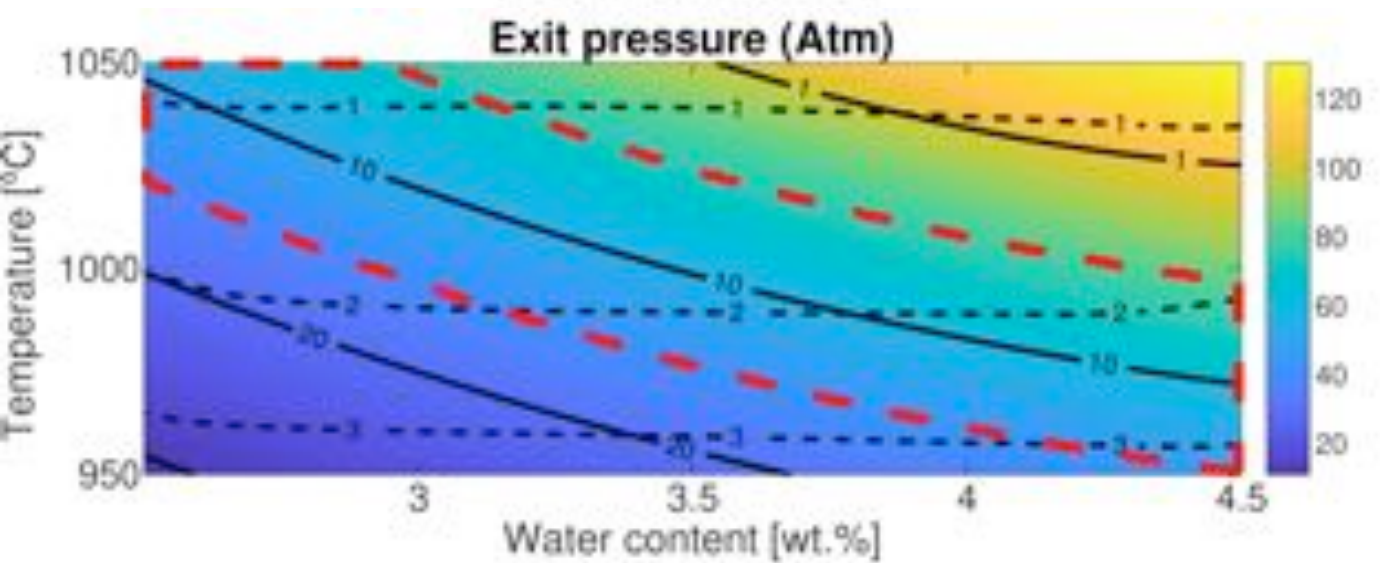
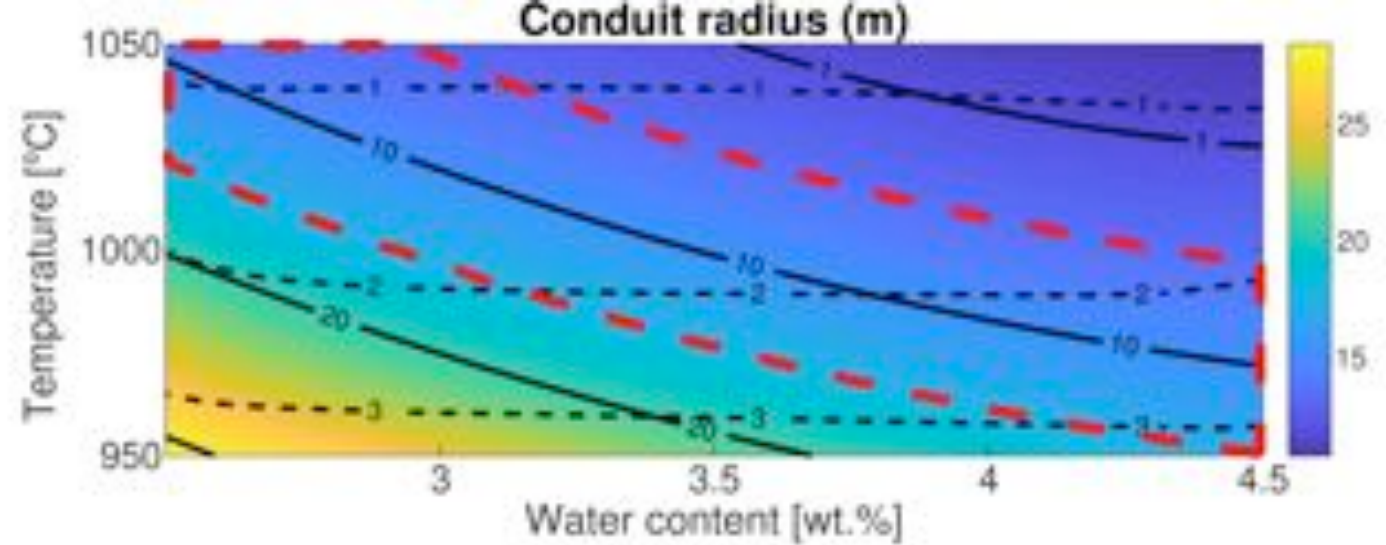
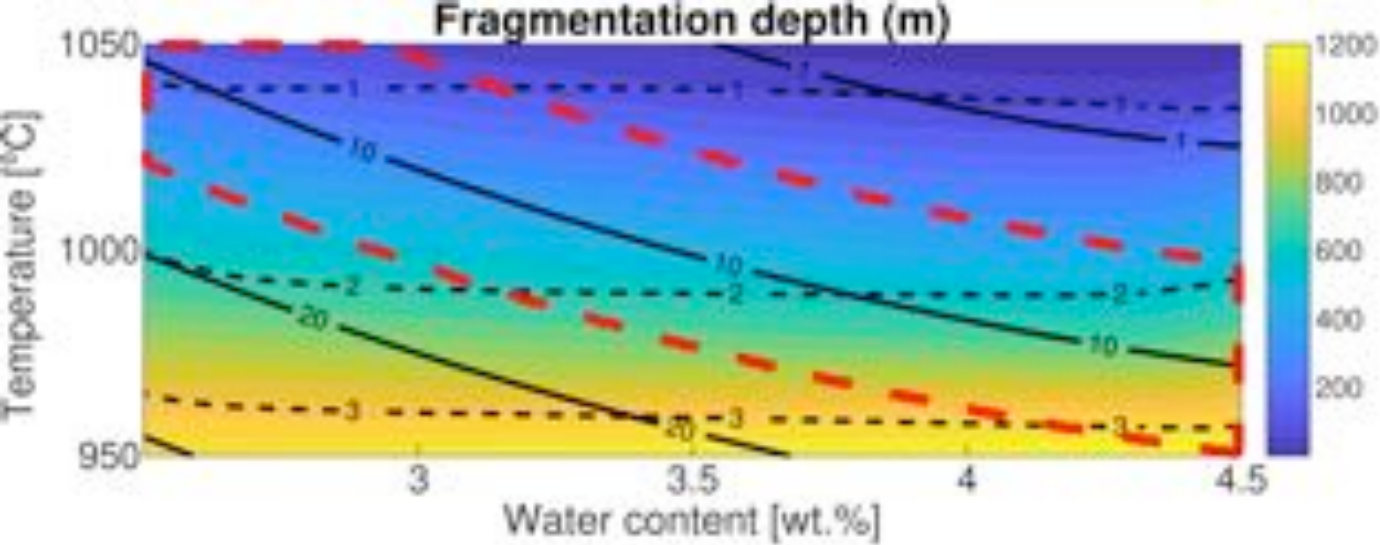


Figure 8.

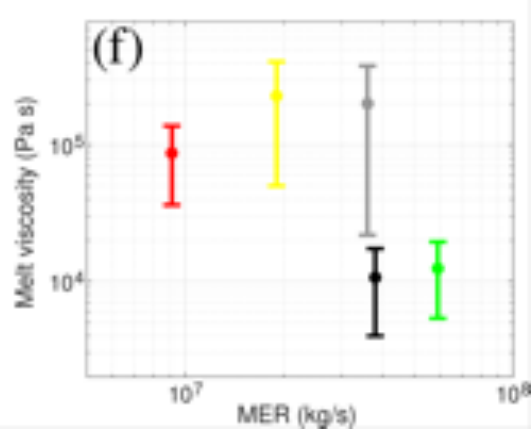
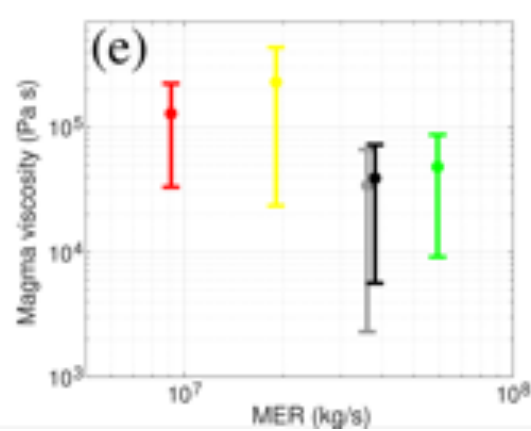
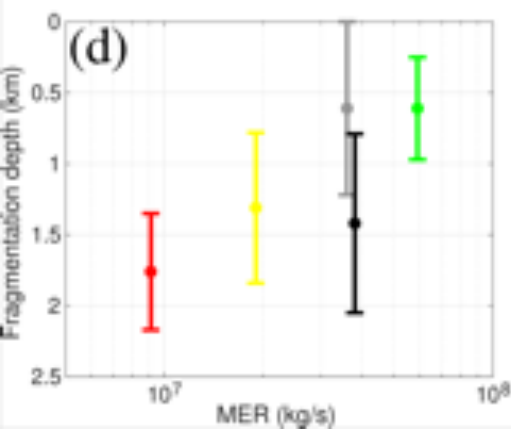
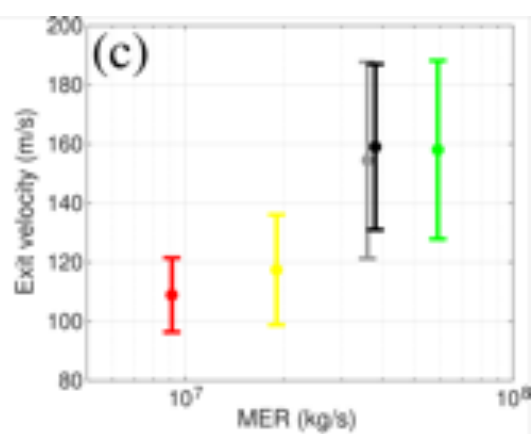
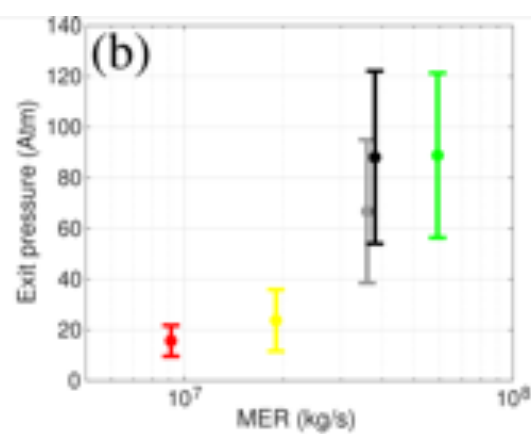
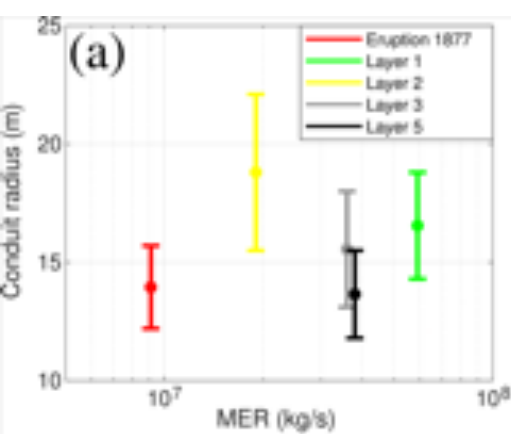
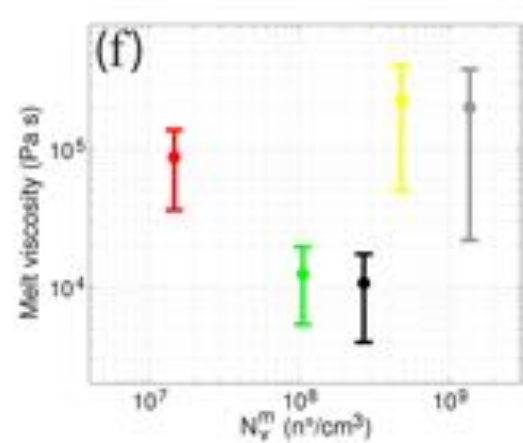
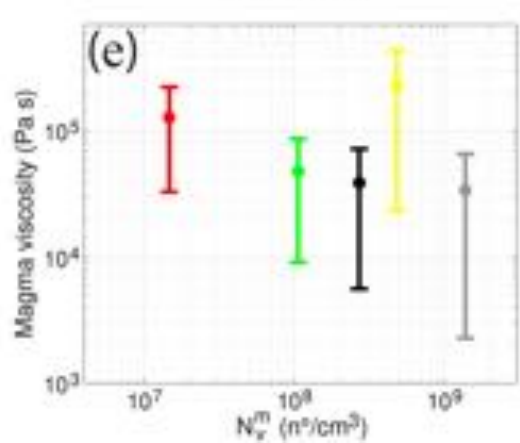
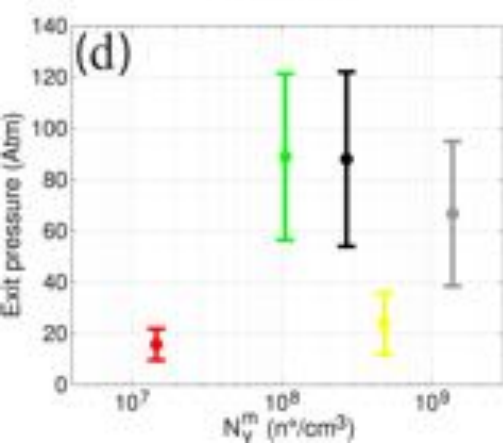
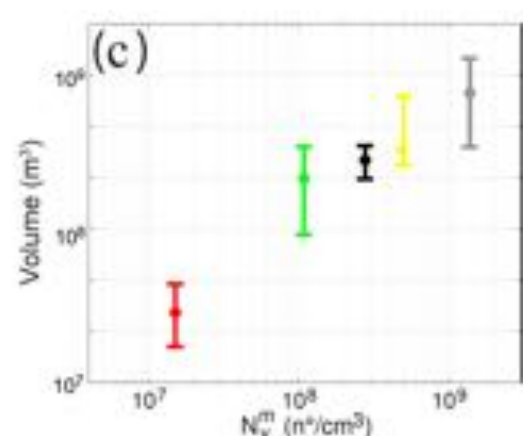
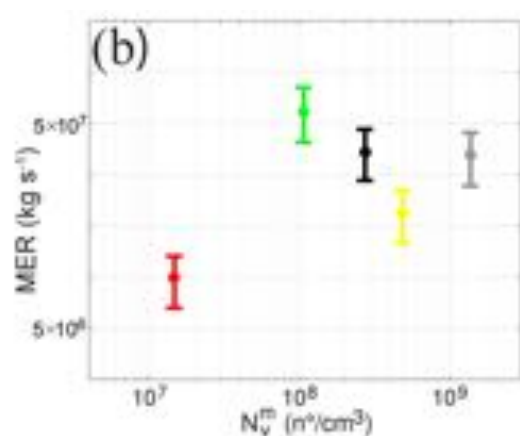
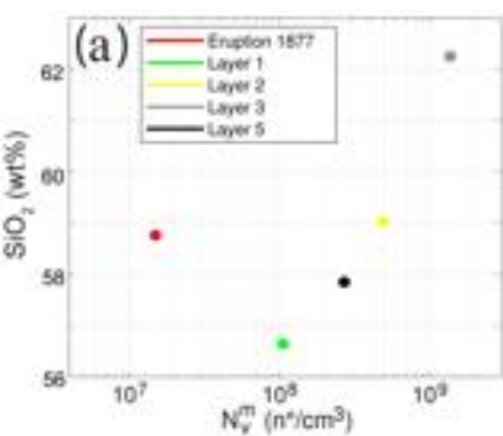


Figure 9.



1 **Table 1.** Eruptive Source Parameters for the five selected eruptions.

Erup./ Layer	Volume (m <sup>3</sup> )				Average plume height above vent (km)		Average wind at tropopause (m/s)		Average Mass Eruption Rate – D&B12 (kg/s)		Average Mass Eruption Rate – Ma09 (kg/s)		Average Mass Eruption Rate – W&W87 (kg/s)		Duration (hours)
	<i>Exp.</i>	<i>PL</i>	<i>W</i>	<i>Inv.</i>	<i>C&amp;S86</i>	<i>RBD 19</i>	<i>C&amp;S86</i>	<i>RBD19</i>	<i>C&amp;S86</i>	<i>RBD19</i>	<i>C&amp;S86</i>	<i>RBD19</i>	<i>C&amp;S86</i>	<i>RBD19</i>	
<b>1877</b>	2.3×10 <sup>7</sup> <sup>a</sup>	4.9×10 <sup>7</sup> <sup>a</sup>	2.5×10 <sup>7</sup>	-	16.5 <sup>a*</sup>	15.2	6.4 <sup>a*</sup>	4.5	1.5×10 <sup>7</sup>	9.1×10 <sup>6</sup>	1.6×10 <sup>7</sup>	1.1×10 <sup>7</sup>	9.8×10 <sup>6</sup>	7.1×10 <sup>6</sup>	0.5
<b>1</b>	1.4×10 <sup>8</sup> <sup>a</sup>	4.0×10 <sup>8</sup> <sup>a</sup>	1.7×10 <sup>8</sup>	-	25.4 <sup>a*</sup>	20.3	17.5 <sup>a*</sup>	18.7	1.4×10 <sup>8</sup>	5.9×10 <sup>7</sup>	9.5×10 <sup>7</sup>	3.8×10 <sup>7</sup>	5.5×10 <sup>7</sup>	2.2×10 <sup>7</sup>	1.0
<b>2</b>	1.3×10 <sup>8</sup> <sup>a</sup>	8.6×10 <sup>8</sup> <sup>a</sup>	1.2×10 <sup>8</sup>	-	21.4 <sup>a*</sup>	16.1	15.5 <sup>a*</sup>	13.3	6.7×10 <sup>7</sup>	1.9×10 <sup>7</sup>	4.7×10 <sup>7</sup>	1.4×10 <sup>7</sup>	2.8×10 <sup>7</sup>	8.9×10 <sup>6</sup>	3.8
<b>3</b>	6.0×10 <sup>8</sup> <sup>b</sup>	1.5×10 <sup>9</sup> <sup>b</sup>	5.5×10 <sup>8</sup>	2.4×10 <sup>9</sup> <sup>b</sup>	26.8 <sup>b*</sup>	18.1	18.3 <sup>b*</sup>	16.0	1.9×10 <sup>8</sup>	3.6×10 <sup>7</sup>	1.2×10 <sup>8</sup>	2.3×10 <sup>7</sup>	6.8×10 <sup>7</sup>	1.4×10 <sup>7</sup>	4.8
<b>5</b>	2.9×10 <sup>8</sup> <sup>c</sup>	3.8×10 <sup>8</sup> <sup>c</sup>	2.3×10 <sup>8</sup> <sup>c</sup>	5.0×10 <sup>8</sup> <sup>b</sup>	25.6 <sup>b*</sup>	18.7	17.0 <sup>b*</sup>	15.3	1.5×10 <sup>8</sup>	3.8×10 <sup>7</sup>	9.8×10 <sup>7</sup>	2.7×10 <sup>7</sup>	5.7×10 <sup>7</sup>	1.6×10 <sup>7</sup>	2.1

2 *Exp.*: Exponential strategy; *PL*: Power-Law strategy; *W*: Weibull strategy; *Inv.*: Inversion Strategy; *C&S86*: Carey and Sparks (1986); *RBD19*:  
3 Rossi et al. (2019); *D&B12*: Degruyter and Bonadonna (2012); *Ma09*: Mastin et al. (2009); *W&W87*: Wilson and Walker (1987).  
4 All values presented in this table are calculated as part of this work with the exception of: <sup>a</sup>, from Pistolesi et al. (2011); <sup>b</sup>, from Biass and  
5 Bonadonna (2011); <sup>c</sup>, from Biass et al. (2019). The values presented with the symbol “\*” indicate that the CW and DW ranges have been  
6 calculated from the corresponding paper, but the plume height was calculated with the Matlab script of Biass et al. (2015).  
7 *Italic* values in the exponential column for the calculation of volume indicates values derived with the method of Fierstein and Nathenson (1992)  
8 for two segments. All the other values are calculated with the method of Pyle (1989) for one exponential segment.  
9 Heights and winds at the tropopause calculated with *C&S86* and *RBD19* are averaged over all lithic contours associated with the average of the 3  
10 axes of the 5 largest clasts (3.2 cm and 1.6 cm for Eruption 1877; 6.4 cm, 3.2 cm and 1.6 cm for Layer 1; 3.2 cm, 1.6 cm and 0.8 cm for Layer 2;  
11 3.2 cm, 1.6 cm and 0.8 cm for Layer 3; 6.4 cm, 3.2 cm, 1.6 cm and 0.8 cm for Layer 5).

- 12 Height above vent = Height with C&S86 or RBD19 – (Vent height – Sampling height), where Vent height = 5.9 km and Sampling height = 3.0  
13 km.
- 14 Parameters used for the calculation of MER with D&B12 equation are: magmatic temperature (1223 K for Layer 3 and 1273 K for Eruption 1877,  
15 Layer 1, Layer 2 and Layer 5), tropopause height (17 km), while wind is averaged across plume height.



**Table 2.** Main physical parameters of tephra samples estimated from textural analyses. Whole-rock and glass matrix data are from Barberi et al. (1995) and from Pistolesi et al. (2011). Mass balance calculations were made with a dedicated spreadsheet using whole rock and groundmass glass analyses from Pistolesi et al. (2011) and mineral data from Saalfeld et al. (2019).

Parameters	Eruption 1877	Layer 1	Layer 2	Layer 3	Layer 5
Average clast density (g/cm <sup>3</sup> )	1.10	0.98	0.84	0.62	1.24
Vesicularity derived from density (%)	61.4	65.1	69.7	77.0	54.9
Vesicularity derived from image analysis (%)	38.3	41.4	53.8	57.9	36.6
Microlites content derived from image analysis (vol. %)	0	30	0	0	30
Phenocrysts reanalysed (vol. %)	35	20	30	10	20
Total crystal content derived from mass balance (wt. %)	33	40	35	16	28
Whole-rock SiO <sub>2</sub> (wt. %, normalized)	58.8	56.7	59.1	62.3	57.9
Glass matrix SiO <sub>2</sub> (wt. %)	62.6	59.5	64.4	64.9	60.4
N <sub>A</sub> (n°/cm <sup>2</sup> )	8.80×10 <sup>3</sup>	5.13×10 <sup>4</sup>	2.37×10 <sup>5</sup>	3.22×10 <sup>5</sup>	1.24×10 <sup>5</sup>
N <sub>V</sub> (n°/cm <sup>3</sup> )	5.56×10 <sup>6</sup>	3.70×10 <sup>7</sup>	1.49×10 <sup>8</sup>	3.15×10 <sup>8</sup>	1.16×10 <sup>8</sup>
N <sub>V</sub> <sup>m</sup> (n°/cm <sup>3</sup> )	1.47×10 <sup>7</sup>	1.06×10 <sup>8</sup>	4.80×10 <sup>8</sup>	1.37×10 <sup>9</sup>	2.70×10 <sup>8</sup>
Variation range of vesicles diameter (mm)	0.008 – 1.1	0.008 – 6	0.010 – 6	0.008 – 6	0.008 – 6

**Table 3.** Constitutive equations and main input parameters used in numerical simulations.

Constitutive Equation	Eruption 1877	Layer 1	Layer 2	Layer 3	Layer 5
Melt viscosity <sup>1</sup>	Giordano et al. (2008)	Giordano et al. (2008)	Giordano et al. (2008)	Giordano et al. (2008)	Giordano et al. (2008)
Effect of bubbles <sup>2</sup>	Costa et al. (2007)	Costa et al. (2007)	Costa et al. (2007)	Costa et al. (2007)	Costa et al. (2007)
Effect of crystals	Costa (2005)	Costa (2005)	Costa (2005)	Costa (2005)	Costa (2005)
Crystallization <sup>3</sup>	alphaMELTS calibration	alphaMELTS calibration	alphaMELTS calibration	alphaMELTS calibration	alphaMELTS calibration
Water solubility <sup>4</sup>	Henry's law	Henry's law	Henry's law	Henry's law	Henry's law
Outgassing <sup>5</sup>	Forchheimer's law	Forchheimer's law	Forchheimer's law	Forchheimer's law	Forchheimer's law
Equations of state (exsolved gas)	Ideal gas	Ideal gas	Ideal gas	Ideal gas	Ideal gas
Equations of state (melt, crystals and pyroclasts) <sup>6</sup>	Mie-Gruneisen equations	Mie-Gruneisen equations	Mie-Gruneisen equations	Mie-Gruneisen equations	Mie-Gruneisen equations
Input parameters	Eruption 1877	Layer 1	Layer 2	Layer 3	Layer 5
Inlet pressure (MPa)	203.8	203.8	203.8	203.8	203.8
Conduit length (km)	8.0	8.0	8.0	8.0	8.0
Magma water content (wt. %)	2.5 – 4.5	2.5 – 4.5	2.5 – 4.5	2.5 – 4.5	2.5 – 4.5
Temperature (°C)	950 – 1050	950 - 1050	950 - 1050	950 - 1050	950 - 1050
Relaxation time for crystallization (s)	1000	10	1000	1000	10
Relaxation time for gas exsolution (s)	10 <sup>-4</sup> - 10	10 <sup>-4</sup> - 10	10 <sup>-4</sup> - 10	10 <sup>-4</sup> - 10	10 <sup>-4</sup> - 10
Mass discharge rate (kg/s)	9.1×10 <sup>6</sup>	5.9×10 <sup>7</sup>	1.9×10 <sup>7</sup>	3.6×10 <sup>7</sup>	3.8×10 <sup>7</sup>
Bubble number density (m <sup>-3</sup> )	1.47×10 <sup>13</sup>	1.06×10 <sup>14</sup>	4.80×10 <sup>14</sup>	1.37×10 <sup>15</sup>	2.70×10 <sup>14</sup>
Gas volume fraction for fragmentation <sup>7</sup>	61.4%	65.1%	69.7%	77.0%	54.9%

<sup>1</sup>Computed using literature-derived data of glass composition for each eruption (Costantini, 2010).<sup>2</sup>Costa et al. (2007) adopted a generalization of Llewellyn and Manga (2005).<sup>3</sup>For each eruption, a set of alphaMELTS (Smith and Asimow, 2005) simulations was performed, considering literature-derived data to define the magma composition (bulk-rock values; Costantini, 2010) and variable values for water content ( $w$ ; 0.0 wt. % - 4.5 wt. %), pressure ( $p$ ; 1 – 4000 bar) and temperature ( $T$ ; 950°C – 1050°C). Equilibrium crystallinity ( $\beta_{eq}$ ) was fitted considering the following relationship:
$$\beta_{eq} = \max(0, \min(1, a_{T^2} \cdot T^2 + a_T \cdot T + a_{p^2} \cdot p^2 + a_p \cdot p + a_{w^2} \cdot w^2 + a_w \cdot w + a_{pw} \cdot p \cdot w + a_{Tw} \cdot T \cdot w + a_{Tp} \cdot T \cdot p + a_0)),$$

where  $T$ ,  $p$  and  $w$  are expressed in K, bar and mass fraction, respectively. Fit coefficients for each eruption are presented in Table S2.

<sup>4</sup>Fit derived from water solubility data on andesitic melts (Botcharnikov et al., 2015).<sup>5</sup>Following Degruyter et al. (2012).<sup>6</sup>Le Métayer et al. (2005).<sup>7</sup>Assumed equal to the vesicularity estimates derived from density measurements (Table 2). This is because they are generally more reliable than estimates obtained directly from image analysis because of the uncertainty associated with the use of stereological models.

**Table 4.** Main results derived from conduit numerical modeling.

<b>Parameter</b>	<b>Eruption 1877</b>	<b>Layer 1</b>	<b>Layer 2</b>	<b>Layer 3</b>	<b>Layer 5</b>
Conduit radius (m)	12 – 16	14 – 19	16 – 22	13 – 18	12 – 16
Fragmentation depth (km)	1.4 – 2.2	0.3 – 1.0	0.8 – 1.8	<1.2	0.8 – 2.1
Exit pressure (Atm)	10 – 20	60 – 120	10 – 40	40 – 100	50 – 120
Exit velocity (m/s)	100 – 120	130 – 190	100 – 140	120 – 190	130 – 190
Magma viscosity at fragmentation level (Pa s)	$3.3 \times 10^4 - 2.2 \times 10^5$	$9.1 \times 10^3 - 8.7 \times 10^4$	$2.3 \times 10^4 - 4.4 \times 10^5$	$2.3 \times 10^3 - 6.6 \times 10^4$	$5.6 \times 10^3 - 7.2 \times 10^4$
Melt viscosity at fragmentation level (Pa s)	$3.6 \times 10^4 - 1.4 \times 10^5$	$5.4 \times 10^3 - 1.9 \times 10^4$	$5.0 \times 10^4 - 4.1 \times 10^5$	$2.2 \times 10^4 - 3.8 \times 10^5$	$4.0 \times 10^3 - 1.7 \times 10^4$