



**HAL**  
open science

## Evolution of magma decompression and discharge during a Plinian event (Late Bronze-Age eruption, Santorini) from multiple eruption-intensity proxies

Madison Myers, Timothy H. Druitt, Federica Schiavi, Lucia Gurioli, Taya Flaherty

### ► To cite this version:

Madison Myers, Timothy H. Druitt, Federica Schiavi, Lucia Gurioli, Taya Flaherty. Evolution of magma decompression and discharge during a Plinian event (Late Bronze-Age eruption, Santorini) from multiple eruption-intensity proxies. *Bulletin of Volcanology*, 2021, 83 (3), 10.1007/s00445-021-01438-3 . hal-03269773

**HAL Id: hal-03269773**

**<https://uca.hal.science/hal-03269773v1>**

Submitted on 22 Nov 2021

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

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

[Click here to view linked References](#)

1

2

3 **Evolution of magma ascent and discharge during a Plinian event**  
4 **(Late Bronze-Age eruption, Santorini) from multiple eruption-**  
5 **intensity proxies**

6

7 Madison L. Myers<sup>a,b</sup>, Timothy H. Druitt<sup>a</sup>, Federica Schiavi<sup>a</sup>, Lucia Gurioli<sup>a</sup>, and Taya Flaherty<sup>a</sup>

8

9 <sup>a</sup> Université Clermont Auvergne, CNRS, IRD, OPGC, Laboratoire Magmas et Volcans, F-63000 Clermont-  
10 Ferrand, France11 <sup>b</sup> Montana State University, Department of Earth Sciences, Bozeman MT 59717-1272

12

13

14

15

16

17

18

19

20

21 \*Corresponding author. Email address: madison.myers@montana.edu

## 22 **Abstract**

23           We have coupled three independent methods to investigate the time-evolution of eruptive  
24 intensity during the sub-Plinian and Plinian phases of the 3600 y BP Late Bronze Age eruption of  
25 Santorini Volcano: (1) mass eruption rate based on new lithic isopleth maps for multiple layers of  
26 the fall deposit, (2) magma decompression rate calculated from vesicle number densities, and (3)  
27 magma decompression rate calculated from H<sub>2</sub>O gradients in melt reentrants, with methods 2 and  
28 3 measured on the same suite of pyroclasts. Mass eruption rate increased by two orders of  
29 magnitude, reaching  $210 \times 10^6 \text{ kg s}^{-1}$  at the peak of the Plinian phase (plume height  $28.4 \pm 1.0$   
30 km); it then declined in the final stage of fallout emplacement following the first generation of  
31 pyroclastic surges. Decompression rates from melt reentrants ( $0.008$  to  $0.25 \text{ MPa s}^{-1}$ ) are two to  
32 three orders of magnitude lower than those from vesicle number densities, assuming heterogeneous  
33 vesicle nucleation ( $2$  to  $19 \text{ MPa s}^{-1}$ ). Melt reentrants probably record slow decompression in the  
34 deep feeder conduit, whereas vesicles record much higher rates of decompression in the shallow  
35 conduit due to the steep, non-linear pressure gradients associated with magma vesiculation and  
36 fragmentation. Upwardly converging flow from a dike-like, deep conduit to a more cylindrical,  
37 shallow conduit may also have played a role in causing upwardly accelerating flow. Variations in  
38 deep decompression rate recorded by melt reentrants are decoupled from mass eruption rate,  
39 whereas those recorded by vesicles lie in between. Taken with the transition from unsteady to  
40 steady Plinian eruption conditions, this may reflect the existence of transient flow conditions in  
41 the conduit system due to progressing widening and lengthening of a deep feeder dike as Plinian  
42 eruption progressed. As the mass eruption rate rose to its peak value, the fragmentation level fell  
43 in the conduit due to increasing rates of magma strain and decompression.

44

45 **Keywords:** Plinian eruption, Santorini, decompression rate, conduit development, vesicle number  
46 density, melt reentrant

47

## 48 **Introduction**

49 Caldera-forming explosive eruptions are amongst the most devastating natural events on  
50 Earth (**Mason et al. 2004; Miller and Wark 2008**). As more than 500 million people live within  
51 the maximum exposure range of a volcano (**Doocy et. al 2013**), refining techniques to measure  
52 how quickly magma moves to the surface, and how magma flow evolves over the course of an  
53 eruption, is crucial for improved hazards monitoring and risk mitigation. Determining the  
54 processes that control transitions in eruption style, which commonly take place during eruptions  
55 due to shifts in external and internal parameters, is also important (**Hildreth and Drake 1992;**  
56 **Woods and Koyaguchi 1994; Castro and Gardner 2008; Degruyter et al. 2012**). Few silicic  
57 caldera-forming eruptions have ever been recorded by modern scientific instrumentation, so  
58 understanding of the processes leading up to, and during, such eruptions relies heavily on studies  
59 of their erupted products.

60 Three main clast-based techniques are available for estimating magma flow rate from  
61 measurements on pyroclastic fall deposits. One uses clast-isopleth data to estimate eruption plume  
62 height through application of plume model inversions (**Carey and Sparks 1986; Bonadonna and**  
63 **Costa 2013**). By splitting the fall deposit into multiple layers, this technique can yield variations  
64 of mass eruption rate with time (**Sulpizio et al. 2010; Houghton et al. 2014**). Decompression rates  
65 for particular volcanic episodes can also be determined using individual pyroclasts. For instance,  
66 the number density of vesicles generated by ascent-driven nucleation is a function of magma  
67 decompression rate (e.g., **Klug and Cashman 1996; Mourtada-Bonnefoi and Laporte 2004;**

68 **Massol and Koyaguchi 2005; Toramaru 2006; Cluzel et al. 2008; Shea et al. 2010; Giachetti**  
69 **et al. 2010; Hamada et al. 2010; Martel and Iacono-Marziano 2015; Hajimirza et al. 2019;**  
70 **Nishiwaki and Toramaru 2019**), although several processes occur in the conduit (e.g.  
71 coalescence, heterogenous nucleation, etc.) that may complicate interpretation. Another approach  
72 is to model gradients of dissolved H<sub>2</sub>O and (if present in detectable quantities) CO<sub>2</sub> preserved in  
73 crystal-hosted melt reentrants to estimate the decompression rate experienced by the crystal during  
74 ascent to the surface and prior to eruption quench (**Liu et al. 2007; Myers et al. 2016, 2018**).  
75 Although several studies have investigated the evolution of mass eruption rate with vesicle number  
76 density (and calculated decompression rate) through a volcanic sequence, no study has directly  
77 compared decompression rates determined using melt reentrants with those using vesicle number  
78 density from the same pyroclasts.

79 In this study we compare and contrast all three independent methods during the opening sub-  
80 Plinian and Plinian fallout phases of the Late Bronze-Age (LBA) eruption of Santorini (**Bond and**  
81 **Sparks 1976; Taddeucci and Wohletz 2001; Druitt 2014**). We integrate the results with pre-  
82 existing interpretations for how eruptive activity proceeded during these eruptive phases. A major  
83 conclusion of our work is that decompression rates recorded by the deep and shallow conduit  
84 systems are decoupled, and we propose a possible explanation in the case of the LBA eruption.

85

## 86 **Geologic Background**

87 The Santorini volcanic field is located on the Aegean volcanic arc, within a NE-SW  
88 trending rift zone (Fig. 1a). It is the most active volcanic center in the eastern Mediterranean region,  
89 having had at least twelve Plinian eruptions over the last 360 ky, interspersed with episodes of lava  
90 emission and smaller explosive events (**Druitt et al. 1999**). The-present-day (LBA) caldera is up

91 to 400 m deep and is cut by a prominent NE-SW lineament called the Kameni Line, which is  
92 parallel to the rift zone (Fig. 1b). Vents of historical volcanic activity are aligned along the Kameni  
93 Line, as were epicenters of the seismic unrest of 2011-12 (**Kaviris et al. 2015**), highlighting that  
94 it is a major line of weakness in the caldera.

95 The LBA eruption (also commonly called the Minoan eruption) was the most recent Plinian  
96 event at Santorini, occurring ~3600 years ago. It discharged 48-86 km<sup>3</sup> of magma and comminuted  
97 rock debris, and collapsed a large part of the modern-day caldera. It began with sub-Plinian  
98 precursory explosions, the plumes from which were largely dispersed to the south (phase 0)  
99 (**Heiken and McCoy 1990; Cioni et al. 2000**). The eruption then evolved through four main  
100 phases (Fig. 1c): a Plinian plume (phase 1), pyroclastic surges (phase 2), low-temperature,  
101 pyroclastic flows (phase 3), and voluminous hot, fluidized pyroclastic flows (phase 4) (**Bond and**  
102 **Sparks 1976; Heiken and McCoy 1984; Sparks and Wilson 1990; Taddeucci and Wohletz**  
103 **2001; Druitt 2014**).

104 The vent for phase 1 has been extrapolated to a location on the Kameni Line; the plume  
105 was dispersed to the east to south-east and grew with time, producing a reversely graded pumice  
106 fall deposit preserved up to 6 m thick (**Bond and Sparks 1976; Heiken and McCoy 1984**). The  
107 resulting fall deposit has a volume of at least 1-2 km<sup>3</sup> dense-rock equivalent, and has been  
108 subdivided into four layers, P1a - P1d (**Druitt 2014**). Layer P1a is crudely stratified and records  
109 an initial, unsteady Plinian plume; layer P1b is non-stratified and records a steadier plume. Layer  
110 P1c is a thin pyroclastic surge bed (commonly referred to in previous publications as the  
111 'phreatomagmatic break' or 'flow break'); it is overlain by layer P1d, the last fall layer prior to the  
112 transition to eruptive phase 2 (Figs. 1c, 2c,d).

113           In this paper we focus solely on the fall deposits of phases 0 and 1. Juvenile pyroclasts in  
114 these phases are white rhyodacite pumices, microphenocryst-rich andesitic pumices, and (in the  
115 upper portion P1b) cauliform andesitic enclaves. Although the rhyodacite pumices account for >  
116 99% of all LBA magma erupted, microphenocryst-rich pumices can comprise 40-90% of the  
117 juvenile material in P0 and the lower levels of P1 (see Table 1 from **Druitt 2014**). The influx of  
118 cauliform andesitic enclaves occurs part way up layer P1b, with their size increasing in conjunction  
119 with those of the rhyodacitic pumices and lithic clasts. The microphenocryst-rich pumices and  
120 cauliform andesitic enclaves are inferred to represent a separate magma intrusion that was  
121 intersected during the tapping of the main rhyodacitic reservoir (**Druitt 2014**). In this study we  
122 focus on the rhyodacitic pumices, and the cauliform enclaves and microphenocryst-rich pumices  
123 are not considered. Previous work of **Wilson and Houghton (1990)** noted only a slight decrease  
124 in rhyodacitic pumice density ( $\sim 0.6$  to  $\sim 0.5$  kg/m<sup>3</sup>), and overall consistent vesicularity (73-82%),  
125 upwards through P1.

126           Lithic components in the LBA fallout units have been interpreted to record changing levels  
127 of fragmentation with time (**Druitt 2014**). Lithics in P0 and at the base of P1 are dominated by  
128 lavas, inferred to be shallowly derived. About mid-way through P1b the proportion of altered lava  
129 lithics increases, and hydrothermally iron-stained blocks of holocrystalline to glass-bearing  
130 granitoids appear in abundance (Figs. 1c, 2b). These then increase in size and abundance to the top  
131 of P1b, where they account for a third of the lithic assemblage. Following the first pyroclastic  
132 surge (P1c), the final fall layer (P1d) is again dominated by fresh lavas. Making the assumption  
133 that free lithics are derived from or above the fragmentation level (e.g. **Barberi et al. 1989**), the  
134 data indicate that the magma fragmentation level was shallow during phase P1a, then descended  
135 with time, reaching the top of melt-bearing plutonic lithologies towards the end of phase P1b.

136 Phase equilibria experiments and melt inclusion volatile contents have constrained the pre-  
137 eruptive storage conditions of the ~70 wt. % SiO<sub>2</sub> rhyodacite to 850-880 °C, 100-160 MPa, 4-5  
138 wt.% H<sub>2</sub>O and <200 ppm CO<sub>2</sub> (Cottrell et al. 1999; Cadoux et al. 2014; Druitt et al. 2016;  
139 Flaherty et al. 2018).

140

## 141 **Methods**

142 We divided the fall deposit into six layers, expanding upon the original four: P0, P1a1,  
143 P1a2, P1b1, P1b2, and P1d (Figs. 1c, Fig. 2). Layer P1a1 is defined by two reversely zoned units  
144 with the transition to P1a2 marked by a coarsening of grain size (Fig. 2c, d). A thin parting of  
145 coarse ash separates layers P1a2 and P1b1, indicating a transient lowering of the eruption plume.  
146 The boundary between P1b1 and P1b2 is identified by the first appearance of cauliflower andesitic  
147 enclaves. The six layers could be traced between multiple sites around the caldera rim. Maximum  
148 lithic size and fall layer thickness measurements were made at 43 localities (see [Supplementary](#)  
149 [Table 1](#)) to determine the evolution of plume height, wind direction, and mass eruption rate during  
150 phases 0 and 1, following the methods of Carey and Sparks (1986). For each site, layer thickness  
151 was measured at several local sections, and an average was used on the isopach map (Fig. 3).  
152 Following the recommendations of Biass and Bonadonna (2011), three dimensions of the five  
153 largest lithic clasts from a 1 m<sup>2</sup> section were measured for each fall layer, and the mean of these  
154 five averaged diameters was used to create a lithic isopleth map ([Supplementary Table 1](#)).  
155 Thickness and lithic measurements for P0 were presented previously by (Cioni et al. 2000). We  
156 use their estimate for phase 0 plume height and convert it to mass eruption rates in the same way  
157 as for our phase 1 data (see Table 1).



158           Single pumice clasts were collected from within each of the six fall layers, with three  
159 samples collected from the thick P1b (lower, middle, upper), resulting in eight samples for this  
160 study – one in P0 and seven in P1 (Supplementary Table 2). Given the restricted range of pumice  
161 vesicularity and densities (**Wilson and Houghton 1990**), no particular clast density class was  
162 targeted. All samples from P1 were analyzed for melt reentrants and vesicle number density from  
163 single pumice clasts; P0 pumices are significantly smaller (< 5 cm) than those of P1 used in this  
164 study (6-13 cm: Supplementary Table 2), requiring multiple clasts to be crushed to obtain enough  
165 crystals to find melt reentrants. The pumices were scrubbed to remove any adhering ash matrix,  
166 and to ensure that microphenocryst-rich pumices were avoided, as they may have had a different  
167 decompression history. Effort was made to avoid pumices with strongly elongated vesicles  
168 (**Taddeucci and Wohletz 2001**), which would affect estimation of vesicle number density. For  
169 vesicle analysis, we chose to focus on the rim of each pumice clast, as rims are less likely to have  
170 been affected by post-fragmentation vesicle growth and coalescence (**Thomas et al. 1994; Shea**  
171 **et al. 2010**). Vesicle Number Densities from thick sections of pumice rims, were imaged using a  
172 Scanning Electron Microscope at image magnifications from  $\times 25$  up to  $\times 1000$ , and then quantified  
173 by analyzing six  $\times 500$  and two  $\times 1000$  images per pumice clast in FOAMS (**Shea et al. 2010**). The  
174 choice of analyzing these two high magnifications serves to focus our counting on the smallest  
175 vesicles, which dominate vesicle number density (**Houghton et al. 2010**). In our calculations we  
176 assumed a constant vesicularity of 75%, based upon the relatively homogenous vesicularity data  
177 of **Wilson and Houghton (1990)**. Full processing procedures and discussion of error analyses can  
178 be found in the Supplementary Material.

179           The remaining portions of each pumice clast were crushed using a mortar and pestle, sieved  
180 to 250- and 500- $\mu\text{m}$  size fractions, and picked for plagioclase, clinopyroxene and orthopyroxene.

181 All selected crystals were coated in vesiculated glass (Supplementary Figure 1). The crystal  
182 separates were then submerged in isopropyl alcohol to search for melt reentrants. This method  
183 worked well for plagioclase, but for the darker pyroxenes it was commonly necessary to mount  
184 the crystals in Crystalbond, slightly polish into their interiors, and use an optical microscope to  
185 search for melt reentrants. Plagioclase was found to contain the majority of melt reentrants  
186 (perhaps due to the challenge of observing them in the pyroxenes), mostly situated on crystal faces.  
187 Due to the requirements of the 1D diffusion code (see below for more information), care was taken  
188 to select only those rare melt reentrants with a simple morphology (i.e. no bottle neck, internal  
189 vesicles or minerals, multiple tubes, etc.), which would complicate diffusion assumptions  
190 (Supplementary Figure 2). We also selected reentrants with a single vesicle at their mouths to  
191 ensure efficient diffusive exchange between the reentrant and the degassing, external melt (after  
192 **Lloyd et al. 2014**). Using this criterion, only a few melt reentrants were found per 50-100 crystals,  
193 although this was quite variable between layers. Sixteen melt reentrants in total were studied (Fig.  
194 4): fourteen in plagioclase, one in orthopyroxene and one in clinopyroxene. All melt reentrants  
195 were measured along their lengths (ranging from 80 to 320  $\mu\text{m}$ ) for  $\text{H}_2\text{O}$  and  $\text{CO}_2$  concentrations  
196 using a Thermo-Nicolet Nexus 670 Fourier transform infrared spectrometer (FTIR) interfaced with  
197 a Continuum IR microscope at the University of Oregon using a computer-controlled stage. Of  
198 these sixteen melt reentrants, five were found to have flat profiles and were not able to be fit using  
199 the diffusion model; these reentrants were not further evaluated. Following FTIR analysis at the  
200 University of Oregon (FTIR-UO), eight of the eleven melt reentrants were re-analyzed (3 were  
201 lost in this process) at the University Clermont Auvergne by FTIR (FTIR-UCA) and Raman  
202 spectroscopy (Raman-UCA). This allowed us to evaluate the dependence of modeled

203 decompression rates on the measured profiles. All methods for these three separate instruments,  
204 and detailed transect information, can be found in the Supplementary Material.

205 To model the measured H<sub>2</sub>O diffusion profiles, we applied the 1-D constant decompression  
206 model of **Myers et al. (2018)**. This model, based on that of **Liu et al. (2007)**, allows for comparison  
207 of the measured and simulated profiles for various decompression rates. The boundary condition  
208 at the contact between the host melt and the mouth of the melt reentrant is based on the melt H<sub>2</sub>O  
209 and CO<sub>2</sub> solubility at a given pressure, updated at each decompression step, and is assumed to be  
210 in equilibrium with the external melt outside the crystal (**Liu et al. 2007; Myers et al. 2016, 2018**).

211

## 212 **Results**

### 213 *Plume heights and mass eruption rates*

214 Isopach and lithic isopleth maps for phase P1 are shown on Fig. 3. The broadening of  
215 isopleths upwards in the fall deposit (Fig. 3b-f) records an increase in plume height with time  
216 (particularly between P1b1 and P1b2), prior to a decrease in P1d. Wind direction recorded by the  
217 isopleths changes from SSE-wards in P1a1 and P1a2, to E-wards in P1b1, to NE-wards in P1b2.  
218 Layer P1d appears to have little wind influence on its dispersion. Wind directions recorded by  
219 isopach maps of the same levels agree with those for the isopleths for P1a, but disagree for P1b,  
220 where isopach wind directions are less rotated than those indicated by the isopleths (Fig. 3a vs. g).  
221 This could be due to the fact that each fallout layer is integrating an overall increase in eruption  
222 intensity throughout its thickness, so that the upper portion of the layer is probably where the  
223 largest clasts reside. Essentially, where the isopleths are recording the wind direction of the upper  
224 portion of any given layer, the isopach thickness is averaging the wind direction of that entire layer.

225 Plume heights for each layer are estimated using the **Carey and Sparks (1986)** inversion  
226 model (Table 1 and Fig. 5). As the height estimate is based solely on the crosswind range (half  
227 width of the depositional envelope), uncertainties presented for plume height are based on the  
228 ‘wobble-room’ available in the crosswind range isopleth placement (Fig. 3). Conversion of isopleth  
229 data to plume height used the 3.2 cm contour, the only one available for all five layers, although  
230 this is loosely constrained for P1a1 and P1a2 (Fig. 3). Plume height estimates were also made for  
231 the few other isopleths available, however agreement is typically poor between conversions  
232 (Supplementary Figures 3 and 4). Although it is feasible to take an average plume height based on  
233 several isopleths, because this can only be completed for certain layers, with quality of isopleths  
234 not all being equal, we have chosen to focus our interpretations on the 3.2 cm isopleth. Taking  $8.5$   
235  $\pm 1.5$  km as the estimated plume height for P0 (**Cioni et al. 2000**), the 3.2 cm isopleths suggest an  
236 initial increase in plume height, with the greatest height of  $28.4 \pm 1.0$  km reached in P1b2,  
237 decreasing to  $21.7 \pm 1.1$  km in P1d, following initial pyroclastic surge (P1c) production (Fig. 5,  
238 Supplemental Figure 4).

239 Plume heights were converted to mass eruption rate using the methods of **Wilson and**  
240 **Walker (1987)** and **Mastin et al. (2009)**, yielding values of  $1.7 \times 10^6$  to  $2.1 \times 10^8$  kg s<sup>-1</sup> and  $1.0 \times$   
241  $10^6$  to  $1.5 \times 10^8$  kg s<sup>-1</sup>, respectively (Table 1). Error bars associated with the Wilson and Walker  
242 (1987) estimates are propagated based on those determined for each plume height (Table 1).  
243 Although the conversion based on the **Wilson and Walker (1987)** formula is consistently higher,  
244 it falls within error of the **Mastin et al. (2009)** value (Table 1). Importantly, the relative range of  
245 two orders of magnitude represented by P0 and P1 is consistent between the two methods. The  
246 lowest mass eruption rate is associated with P0, and the highest with P1b2 (Fig. 5).  
247

248 *Decompression rates from pumice Vesicle Number Densities*

249       Visually there is a noticeable change in the sizes of vesicles in the rims of pumice clasts up  
250 through P0 and P1, with smaller (average diameter 4-5  $\mu\text{m}$ ) vesicles found at the top of P1a and  
251 the top of P1b (Fig. 6, Table 2). The largest vesicles, with average diameters of 9-13  $\mu\text{m}$ , are  
252 observed in P0 (largest), at the base of P1a, at the base of P1b, and in P1d. Pumice-rim vesicle  
253 number density values span 1.5 orders of magnitude, from  $2.50 \times 10^6$  to  $6.56 \times 10^7 \text{ mm}^{-3}$ , falling  
254 within the range typical for explosive rhyolitic eruptions (**Giachetti et al. 2010; Houghton et al.**  
255 **2010**). The lowest vesicle number density is from P0, and the highest is from P1b2.

256       Vesicle number densities were converted to decompression rates using equation (2) from  
257 **Toramaru (2006)**. This requires estimates of the interfacial tension between  $\text{H}_2\text{O}$  and melt, the  
258 starting saturation pressure and temperature, and the melt  $\text{H}_2\text{O}$  diffusivity, where the largest  
259 uncertainty comes from the choice of interfacial tension (**Shea 2017**). Using a temperature  
260 (850  $^\circ\text{C}$ ), pressure (165 MPa) and  $\text{H}_2\text{O}$  content (5.2 wt.%) appropriate for the LBA magma (**Druitt**  
261 **et al. 2016; Flaherty et al. 2018**), the diffusivity of  $\text{H}_2\text{O}$  is found to be  $1.1 \times 10^{-11} \text{ m}^2 \text{ s}^{-1}$  (equation  
262 27 from **Zhang et al. 2007**). The only remaining unknown is the interfacial tension, which relies  
263 on whether vesicle nucleation was homogenous or heterogeneous. We estimate decompression  
264 rates using both nucleation mechanisms, where the interfacial tension is taken to be  $0.025 \text{ N m}^{-1}$   
265 and  $0.12 \text{ N m}^{-1}$  for heterogeneous and homogeneous nucleation, respectively (**Shea 2017**).  
266 Resulting decompression rates recorded by the vesicle number densities are 2-19  $\text{MPa s}^{-1}$   
267 (heterogeneous) or 50-450  $\text{MPa s}^{-1}$  (homogeneous). In each case, the lowest rate is from P0 and  
268 the highest is from P1b2 (Fig. 5).

269

270 *Melt Reentrant Volatile Profiles*

271           Of a total of eleven measured H<sub>2</sub>O profiles in melt reentrants, nine display gradual  
272 decreases in H<sub>2</sub>O concentration as the transect approaches the crystal rim and the remaining two  
273 preserve slightly flatter profiles (Supplementary Figure 5). Some variabilities in H<sub>2</sub>O  
274 concentrations and gradients are observed between the three instruments used (FTIR-UO, FTIR-  
275 UCA, Raman-UCA), but the shape of each diffusion profile is usually similar (Fig. 7,  
276 Supplementary Table 4). Most melt reentrants measured by FTIR spectroscopy contain interior  
277 H<sub>2</sub>O concentrations (H<sub>2</sub>O = 2.0-5.0 wt.%) lower than those found in isolated melt inclusions in the  
278 same phenocryst phases (H<sub>2</sub>O = 4.0-5.4 wt.%, Table 2 of **Druitt et al. 2016**); however, some H<sub>2</sub>O  
279 concentrations determined by Raman spectroscopy (H<sub>2</sub>O = 3.2-5.9 wt.%) are higher than those  
280 found by FTIR. The absence of detectable CO<sub>2</sub> in the melt reentrants is not surprising given the  
281 low concentrations of CO<sub>2</sub> in isolated melt inclusions (<200 ppm; **Druitt et al. 2016**) and  
282 compared to other silicic systems (**Myers et al. 2018**).

283           All 1-D decompression models requires an estimate of the starting pressure, initial  
284 dissolved H<sub>2</sub>O concentration, temperature, and exsolved gas content. Models were run assuming  
285 a pre-eruptive temperature of 850 °C, a constant decompression rate and isothermal conditions. A  
286 starting pressure (165 MPa) and H<sub>2</sub>O concentration (5.2 wt.%) were determined based on the lack  
287 of measurable CO<sub>2</sub>, but relatively high H<sub>2</sub>O, measured in the interiors of most melt reentrants. This  
288 pressure is broadly consistent with that estimated from melt inclusions (100-160 MPa), with H<sub>2</sub>O  
289 concentrations representing upper end values measured from melt inclusions, but with CO<sub>2</sub> at the  
290 lower end of the dataset (**Druitt et al. 2016**). To objectively choose the best-fit profile, we used  
291 an iterative grid-search function to optimize fitting of the measured profiles (**Myers et al. 2018**).  
292 Unlike in **Myers et al. (2018)**, we cycled through a range of decompression rates and  
293 fragmentation pressures (pressure where diffusion ceases in the model), rather than initial gas

294 content. This allowed us to test whether the selected best-fit fragmentation pressure shifts with  
295 time during eruption. For low-CO<sub>2</sub> systems, the selected decompression rate is less sensitive to  
296 starting gas content, so we assumed no initial exsolved gas phase.

297 It was found that for those melt reentrants where concentration profiles were measured by  
298 all three methods (eight of eleven), modeled decompression rates were fairly consistent (Fig. 5;  
299 Supplementary Figure 5). Good model fits for all profiles could be achieved using our pre-ascent  
300 storage pressure estimate of 165 MPa. Hereon, we focus on the decompression results based on  
301 the University of Oregon FTIR, where all eleven melt reentrants were measured. The resulting  
302 decompression rates range between 0.008 and 0.25 MPa s<sup>-1</sup>, but without any systematic variation  
303 with stratigraphic height (Fig. 5). For those layers where multiple melt reentrants were measured  
304 and modeled (RDP18, RDP13 and RDP5) agreement between the different decompression rate  
305 estimates is good, especially compared to the two orders of magnitude represented by the entire  
306 dataset (Fig. 5).

307

## 308 **Discussion**

### 309 *Time variations in plume height and mass eruption rate*

310 Our isopleth data indicate that the plume rose from P0 to P1b, before diminishing in height  
311 after production of the pyroclastic surges of P1c (Fig. 5). Previous estimates placed the maximum  
312 plume height for phase 1 at  $36 \pm 5$  km, with a maximum eruption rate of  $1.4 - 4.2 \times 10^8$  kg s<sup>-1</sup>  
313 (**Sigurdsson et al. 1990, Sparks and Wilson 1990**, based on the data of **Bond and Sparks 1976**).  
314 Our estimate of maximum plume height is lower than this (layer P1b2: plume height  $28.4 \pm 1.0$   
315 km, mass eruption rate  $2.1 \pm 0.6 \times 10^8$  kg s<sup>-1</sup> based on **Wilson and Walker 1987**), probably because

316 we restricted sampling to a 1 m<sup>2</sup> area, whereas **Bond and Sparks (1976)** sampled over the entire  
317 outcrop (R.S.J. Sparks, written communication).

318

### 319 *Comparison of the decompression rate estimates*

320 Calculated magma decompression rates based on vesicle number density differ by more  
321 than an order of magnitude, depending on the nucleation assumption (2-19 MPa s<sup>-1</sup> - heterogeneous  
322 or 50-450 MPa s<sup>-1</sup> - homogeneous; Fig. 5). Homogenous nucleation is commonly assumed to  
323 dominate in more evolved melts (**Mangan and Sisson 2000**) since silicate minerals are thought to  
324 be poor nucleation sites for vesicles (**Hurwitz and Navon 1994**). Much of the requirement for  
325 heterogenous nucleation is based on the presence of Fe-Ti oxides, long established as the best  
326 nucleation site for vesicles (**Hurwitz and Navon 1994**). However, a review of vesicle size  
327 distributions and magma decompression rates by **Shea (2017)** demonstrates that the assumption of  
328 homogenous nucleation, and the resulting interfacial tension value, produces decompression  
329 estimates that are inconsistent with other nucleation-based rate-meters, and often requires  
330 overpressures greater than the inferred depth of storage. **Shea (2017)** argues that heterogenous  
331 nucleation is likely in all magmas (basaltic through rhyolitic), perhaps facilitated by oxide  
332 nanolites. The following lines of evidence suggest that heterogenous nucleation occurred during  
333 ascent of the LBA magma: (1) H<sub>2</sub>O gradients measured in reentrants can be modeled by constant  
334 decompression from the storage region to the fragmentation level, requiring that diffusion to a  
335 degassing external melt was occurring early on in magma ascent (and hence that the external melt  
336 was growing vesicles); (2) apparent nucleation of vesicles on Fe-Ti oxide grains contained within  
337 melt reentrants (Supplementary Figure 2 – not evaluated for H<sub>2</sub>O profiles/decompression rate).



338 Even assuming heterogeneous nucleation, the decompression rates from melt reentrants  
339 ( $0.008$  to  $0.25 \text{ MPa s}^{-1}$ , average  $0.06 \text{ MPa s}^{-1}$ ) are 2-3 orders of magnitude lower than those based  
340 on vesicle number density ( $2$ - $19 \text{ MPa s}^{-1}$ ), although they each span about an order of magnitude.  
341 Previous workers have also noted discrepancies between the rates retrieved from melt reentrants  
342 and those from vesicles (Shea 2017; Cassidy et al. 2018). The offset can be understood by  
343 considering the kinetics of the two processes. While vesicles nucleate on timescales of seconds in  
344 response to rapid magma decompression (e.g., Toramaru 2006; Hajimirza et al. 2019;  
345 Nishiwaki and Toramaru 2019), the ability of a  $\text{H}_2\text{O}$  gradient in a melt reentrant to react to  
346 external decompression is limited by the rate of  $\text{H}_2\text{O}$  diffusion in melt. Melt reentrants will cease  
347 recording diffusive  $\text{H}_2\text{O}$  loss if the magma decompression rate becomes too fast. Liu et al. (2007)  
348 estimate this threshold to be  $\sim 0.25 \text{ MPa s}^{-1}$  in silicic systems, which indeed corresponds to the  
349 uppermost decompression rates reported here and by Myers et al. (2018). While melt reentrants  
350 are unable to record decompression rates much higher than this value, vesicle number densities  
351 can record much faster rates (Shea 2017). The offset in decompression rates estimated by the two  
352 methods suggests that the decompression rate of the magma increased during ascent, from  $<0.25$   
353 (recorded by melt reentrants) to  $\gg 0.25 \text{ MPa s}^{-1}$  (recorded by vesicles). We envisage two possible  
354 explanations, distinguishing between deeper and shallower levels of the feeder conduit.

355

356 1. *Nonlinear pressure gradient during ascent to the fragmentation level.* As the magma  
357 ascends the feeder conduit, water diffuses from the melt to vesicles and the magma  
358 vesiculates. The mixture of gas and melt is then accelerated to the surface and undergoes  
359 fragmentation. During this process the magma can experience rapidly accelerating  
360 decompression due to: (1) increasing gas fraction as the magma degasses and the gas phase

361 expands (i.e., mass continuity); (2) large ( $\gg$  lithostatic) nonlinear pressure gradients  
362 immediately beneath the fragmentation level due to the marked increase in melt viscosity  
363 as the dissolved water content decreases (**Cashman and Scheu 2015; Gonnerman, 2015**)  
364 (as a silicic melt with 5 wt.% water degasses to 1 wt.% water, its viscosity increases by 2-  
365 3 orders of magnitude (**Giordano et al. 2008; Romine and Whittington 2015**); (3) rapid  
366 decompression associated with magma fragmentation. The high rates of magma  
367 decompression preceding, and accompanying, fragmentation may in some cases trigger a  
368 second vesicle nucleation event (**Massol and Koyaguchi 2005; Toramaru 2006;**  
369 **Hamada et al. 2010; Hajimirza et al. 2019**).

370

371 2. *Upwardly converging magma flow*. Accelerating magma flow may also result from  
372 changes in conduit geometry with depth beneath the volcano, for example a feeder dike at  
373 depth focusing upwards into a narrow, shallow conduit. Syn-eruptive magma flow through  
374 a deep, dike-shaped conduit transitioning into a shallow cylindrical conduit has been called  
375 upon at a number of volcanoes, such as Montserrat in the Antilles (**Costa et al. 2007**) and  
376 Somma–Vesuvius in Italy (**Massaro et al. 2018**). The location of the vents of LBA phases  
377 P0 and P1 on the NE-SW-trending Kameni Line (Fig. 1b), a long-lived line of weakness in  
378 Santorini caldera, suggests that it is possible that the LBA magma left its upper crustal  
379 reservoir in a dike, then converged into a more cylindrical conduit at shallow levels. This  
380 is also inferred to have occurred during historical activity at Santorini (**Pyle and Elliott**  
381 **2006**, Fig. 8). The upward focusing of magma flow from a 2 km-long dike (the deep  
382 conduit) into a 20 m-long segment of that dike (the shallow conduit) could have increased  
383 the ascent rate of the LBA magma by two orders of magnitude, providing another possible

384 explanation of the offset between melt reentrants and vesicle number density. In this  
385 mechanism, the LBA melt reentrants would record the initial slow stage of magma ascent  
386 through the deeper feeder dike, and vesicle number density the much faster stage of focused  
387 magma flow through the shallow conduit.

388

389 In summary, we infer that the 2-3 orders of difference between the decompression rates  
390 recorded by melt reentrants and vesicles can be attributed to the non-linear pressure gradient  
391 associated with the fragmentation level, possibly coupled with a downward flaring of the feeder  
392 conduit. Since melt reentrants only record slower decompression conditions ( $<0.25 \text{ MPa s}^{-1}$ ), they  
393 likely record flow during deeper magma ascent, whereas vesicle number densities are weighted  
394 towards the rapid, nonlinear decompression of faster, shallow ascent associated with fragmentation  
395 (Fig. 8; **Massol and Koyaguchi 2005; Toramaru 2006; Hamada et al. 2010; Hajimirza et al.**  
396 **2019**).

397

#### 398 *Integration of the different parameter sets*

399 We now compare time variations of the different eruption intensity proxies (Fig. 5). Mass  
400 eruption rate is lowest in P0 and rises steadily to a maximum near the top of P1b, before decreasing  
401 in P1d. Decompression rate from vesicle number density mimics mass eruption rate by increasing  
402 from P0 to P1a2, being highest at the top of P1b2, then decreasing in P1d. However, it differs from  
403 mass eruption rate in that it shows a small drop from P1a2 to P1b1. Finally, decompression rate  
404 from melt reentrants increases from P0 to P1a1, then decreases markedly into P1a2 and P1b1  
405 before rising again. Time-variations of the three parameters suggests: (1) magma decompression  
406 rate in the deep conduit (recorded by melt reentrants) is decoupled from mass eruption rate at the

407 surface; (2) vesicles record rates intermediate between that in the deep conduit and the mass  
408 eruption rate; and (3) changes in decompression rate in the shallow conduit appear to lag behind  
409 those in the deep conduit (for example, the drop in decompression rate from P1a1 to P1a2 in the  
410 deep conduit occurs in vesicle number density from P1a2 to P1b1; Fig. 5).

411         The observed decoupling between the deep and shallow levels of the conduit may be  
412 understood if the deep conduit was a dike, as discussed earlier (Fig. 8). In eruptions fed by dikes  
413 that transition into shallow cylindrical conduits, significant decoupling of the two levels of the  
414 feeder system are possible (Costa et al. 2007; Massaro et al. 2018). The dike can act as a  
415 magmatic capacitor, storing magma (and pressure) by elastic deformation of the dike walls before  
416 conveying it to the fragmentation level in the shallow cylindrical conduit. The shallow conduit,  
417 generated by vent erosion, in turn modulates the mass eruption rate. Abrupt changes in deep  
418 conduit conditions can cause transient flow states, which, due to the high viscosities of silicic melts,  
419 can take several hours to propagate up into the shallow conduit (de' Michieli Vitturi et al. 2010).  
420 Such effects may explain why time-variations in our shallow (vesicle number density-derived)  
421 decompression rate appear to be intermediate between the deep (melt reentrant-derived)  
422 decompression rate and the mass eruption rate.

423         It is unclear what could have caused magma ascent rate in the feeder dike to drop by over  
424 an order of magnitude during P1a1 and P1b1. Assuming that the data are representative of the  
425 magma as a whole, one possibility is that the melt reentrants are recording short-term fluctuations  
426 in deep magma decompression rate not captured by the mass eruption rate data. While mass  
427 eruption data are layer-averaged, melt reentrants are derived from individual magma parcels  
428 (pyroclasts) and may record processes taking place on shorter timescales. The bedded nature of  
429 P1a shows that eruptive conditions early on in phase 1 were unsteady, and that quasi-steady

430 conditions were not attained until P1b. The occurrence of a coarse-ash parting at the P1a-P1b  
431 boundary (Figs. 1c, 2c,d) suggests that the mass eruption rate declined greatly at this point in the  
432 eruption, as supported by the decrease in decompression rate preserved by melt reentrants and  
433 VND, although not shown by our isopleth data. This suggests that melt reentrants and vesicles,  
434 due to their ability to respond rapidly to changing decompression conditions (although on subtly  
435 different timeframes), may record changes in eruption history that might not be recognized as  
436 significant time breaks in the field. The low decompression rates recorded by melt reentrants in  
437 P1a2 and at the base of P1b may therefore be inherited from early unsteadiness in deep magma  
438 flow as the conduit geometry evolved in response to changes in magma pressure and crustal  
439 stresses. Early transient flow and eruption unsteadiness might have been due to widening and  
440 lengthening of the deep feeder dyke as the eruption gained pace, as also invoked for the Pomici di  
441 Avellino eruption of Vesuvius (**Massaro et al. 2018**).

442

#### 443 *Behavior of the fragmentation level*

444 Magma fragmentation rate is controlled by a number of factors, including magma  
445 composition, volatile content, rheology, decompression rate, and strain rate (**Cashman and Scheu**  
446 **2015; Gonnerman, 2015, Cassidy et al. 2018** and references therein). The parameter variations  
447 in Figure 5 offer a possible explanation for the fall in fragmentation level during P1b, as was  
448 inferred from lithic data (**Druitt 2014**). This idea is supported by high mouth pressures preserved  
449 by melt reentrants at the P1b2 level (80-90 MPa; Supplementary Figure 6). Vertical migration of  
450 the fragmentation level will depend on the competition between the fragmentation rate and the rate  
451 of supply of magma from depth. Neither magma composition, dissolved volatile content, nor  
452 rheology changed significantly during the eruption (**Flaherty et al. 2018**). Thus, the main process

453 causing the fragmentation level to fall throughout P1b was likely the increasing magma flow rate  
454 in the shallow conduit. This could have increased the fragmentation rate by increasing the rates of  
455 strain, decompression and development of gas overpressure in vesicles (**Gonnerman 2015**).  
456 Indeed, the number of free crystals (relative to magma crystal content) increase upwards in fall  
457 unit P1b, which when coupled with analysis of particle populations, is consistent with increasing  
458 fragmentation rate during Plinian eruption prior to phase P1c (**Taddeucci and Wohletz 2001**).  
459 However, the apparent decline in magma supply from the deep feeder dyke early on during P1b  
460 (recorded by melt reentrants) may have also played a role by starving the fragmentation level of  
461 ascending magma. Once the fragmentation level had dropped significantly in P1b, access of  
462 seawater to the conduit, and/or the onset of plume instabilities, caused pyroclastic surge production  
463 (P1c). This was then followed by re-establishment of shallow fragmentation (inferred by lithics)  
464 and Plinian phase P1d.

465

#### 466 *Rapid Exploitation of a Pre-existing Conduit System*

467 Previous work on the opening phases of large rhyolitic eruptions (Huckleberry Ridge,  
468 Oruanui and Bishop caldera-forming eruptions) found that melt reentrant diffusion profiles  
469 recorded at least a two-stage decompression history: (1) an initial stage of slow decompression  
470 prior to (2) a final, faster ascent associated with the explosive eruption (**Myers et al. 2018**). This  
471 explanation was required because measured diffusion profiles in crystals from the opening phases  
472 were best recreated using starting volatile concentrations lower than the pre-eruptive magma  
473 storage concentrations recorded by isolated melt inclusions. However, as the eruptions progressed  
474 (most notably the Bishop Tuff), the modeled starting conditions started to approach the storage  
475 concentrations, probably due to a transition from an initial, sluggish ascent to a fully developed

476 feeder system. These observations, however, are not applicable to the LBA eruption, where all  
477 melt reentrant profiles can be reproduced using pre-eruptive storage depth as the initial condition,  
478 even in phase 0 (albeit assuming low CO<sub>2</sub>). This could suggest that either feeder development was  
479 relatively rapid in the LBA system, or that the ascending magma exploited a pre-existing zone of  
480 weakness.

481         Relevant to this idea is that alongside the eruption of the main rhyodacitic LBA magma,  
482 phases 0 and 1 were accompanied by a chemically and mineralogically distinct magma (preserved  
483 as microphenocryst-rich pumices) that has been interpreted previously to represent an intrusion  
484 already present beneath the Plinian vent prior to the LBA eruption (see **Druitt 2014** for full data  
485 and interpretations); indeed this other magma accounts for up to 40 % of pumice discharged during  
486 phase 0. It is therefore likely that the LBA magma exploited an already existing intrusion to reach  
487 the surface, pushing out some of its contents. Furthermore, seismic tomography has imaged a  
488 vertical cylinder of low-density rock beneath Santorini caldera extended down to 3 km, which may  
489 also suggest the existence of a long-lived structural pathway that could have been exploited by the  
490 ascending LBA magma (**Hoof et al. 2019**).

491

## 492 **Conclusions**

493 We have compared and contrasted three eruption intensity proxies for the sub-Plinian and Plinian  
494 opening phases of the Late Bronze-Age eruption of Santorini: plume height and mass eruption rate  
495 from lithic isopleths, magma decompression rate from pumice vesicle number densities, and  
496 magma decompression rate from H<sub>2</sub>O diffusion gradients in crystal-hosted melt reentrants. The  
497 two decompression-rates determinations were carried out on the same suite of pyroclasts. The aim

498 was to obtain insight into what each technique records and detail the processes of conduit magma  
499 flow early on in a caldera-forming eruption. The main conclusions are as follows.

500 1. Plume heights during fallout accumulation increased from  $8.5 \pm 1.5$  km to  $28.4 \pm 1$  km,  
501 then decreased to  $21.7 \pm 1.1$  km following a transient phase of pyroclastic surge  
502 emplacement. Mass eruption rate during the fallout phases is estimated to have reached  
503  $210 \times 10^6$  kg s<sup>-1</sup>.

504 2. Decompression rates estimated from vesicle number densities are 2-3 orders of magnitude  
505 higher than those from melt reentrant H<sub>2</sub>O diffusion gradients (0.008 to 0.25 MPa s<sup>-1</sup>) if  
506 heterogeneous vesicle nucleation is assumed (2-19 MPa s<sup>-1</sup>), and 4 orders of magnitude  
507 higher if homogeneous nucleation is assumed (50-450 MPa s<sup>-1</sup>). Whereas melt reentrants  
508 record slow (<0.25 MPa s<sup>-1</sup>) magma decompression during flow deep in the conduit, vesicle  
509 number densities record much higher rates of decompression in the shallow conduit,  
510 probably due to the steep, non-linear pressure gradients associated with magma  
511 fragmentation. Convergent flow from a dike-shaped deep conduit to a more cylindrical  
512 shallow conduit may also in part explain the high rates of magma decompression recorded  
513 by vesicles. The two methods therefore provide information on syn-eruptive magma  
514 decompression rates at different levels in the conduit.

515 3. The time-variation of magma decompression rate in the deep conduit (recorded by melt  
516 reentrants) is decoupled from mass eruption rate at the surface, whereas that in the shallow  
517 conduit (recorded by vesicles) lies in between. Changes in shallow decompression rate  
518 appear to lag behind those in deep decompression rate. A decrease in decompression rate  
519 preserved by melt reentrants during the transition from unsteady to steady flow may have



520 resulted from transient flow phenomena in a deep feeder dike, possibly caused by an event  
521 of dike widening and/or lengthening.

522 4. Peak eruptive conditions during the Plinian phase were preceded by a fall in the  
523 fragmentation level, perhaps due to the increasing rates of shallow magma decompression  
524 and strain causing the rate of magma fragmentation to exceed that of magma supply from  
525 deeper in the conduit.

526 5. Melt reentrants and vesicles, due to their ability to respond rapidly to changing  
527 decompression conditions, may record changes in eruption history that might not be  
528 recognized as significant time breaks in the field.

529 6. The data support a previously published interpretation that the LBA magma made its way  
530 to the surface through a pre-existing zone of weakness, likely associated with an intrusion  
531 related to an earlier phase of magma movement.

532

### 533 **Acknowledgements**

534 The authors would like to thank Bruce Houghton for conversations surrounding isopleth/isopach  
535 data, and helpful reviews by the editor, one anonymous reviewer and Alessandro Vona that led to  
536 an improved manuscript. This work was funded through a Clermont Auvergne University  
537 Postdoctoral grant to Druitt and an NSF grant EAR-1922513 to Myers. We thank Christophe  
538 Constantin for thin sections, Jean-Marc Hénot for assistance with the SEM, and Erica Duncan,  
539 Emma Kerins, Megan Saalfeld, and Crystal Christensen for processing SEM images. This is  
540 Laboratory of Excellence ClerVolc Contribution Number 448.

541

### 542 **References**

- 543 Barberi F, Cioni R, Rosi M, et al (1989) Magmatic and phreatomagmatic phases in explosive  
544 eruptions of Vesuvius as deduced by grain-size and component analysis of the pyroclastic  
545 deposits. *J Volcanol Geotherm Res* 38:287–307. [https://doi.org/10.1016/0377-](https://doi.org/10.1016/0377-0273(89)90044-9)  
546 [0273\(89\)90044-9](https://doi.org/10.1016/0377-0273(89)90044-9)
- 547 Biass S, Bonadonna C (2011) A quantitative uncertainty assessment of eruptive parameters  
548 derived from tephra deposits: the example of two large eruptions of Cotopaxi volcano,  
549 Ecuador. *Bull Volcanol* 73:73–90. <https://doi.org/10.1007/s00445-010-0404-5>
- 550 Bonadonna C, Costa A (2013) Plume height, volume, and classification of explosive volcanic  
551 eruptions based on the Weibull function. *Bull Volcanol* 75:742.  
552 <https://doi.org/10.1007/s00445-013-0742-1>
- 553 Bond A, Sparks RSJ (1976) The Minoan eruption of Santorini, Greece. *J Geol Soc* 132:1–16.  
554 <https://doi.org/10.1144/gsjgs.132.1.0001>
- 555 Cadoux A, Scaillet B, Druitt TH, Deloule E (2014) Magma Storage Conditions of Large Plinian  
556 Eruptions of Santorini Volcano (Greece). *J Petrol* 55:1129–1171.  
557 <https://doi.org/10.1093/petrology/egu021>
- 558 Carey S, Sparks RSJ (1986) Quantitative models of the fallout and dispersal of tephra from  
559 volcanic eruption columns. *Bull Volcanol* 48:109–125.  
560 <https://doi.org/10.1007/BF01046546>
- 561 Cashman KV, Scheu B (2015) Magmatic fragmentation. *The encyclopedia of volcanoes* (pp.  
562 459-471). Academic Press.
- 563 Cassidy M, Manga M, Cashman K, Bachmann O (2018) Controls on explosive-effusive volcanic  
564 eruption styles. *Nat Commun* 9:2839. <https://doi.org/10.1038/s41467-018-05293-3>
- 565 Castro JM, Gardner JE (2008) Did magma ascent rate control the explosive-effusive transition at  
566 the Inyo volcanic chain, California. *Geology* 36:279. <https://doi.org/10.1130/G24453A.1>
- 567 Cioni R, Gurioli L, Sbrana A, Vougioukalakis G (2000) Precursory phenomena and destructive  
568 events related to the Late Bronze Age Minoan (Thera, Greece) and AD 79 (Vesuvius,  
569 Italy) Plinian eruptions; inferences from the stratigraphy in the archaeological areas. *Geol*  
570 *Soc Lond Spec Publ* 171:123–141. <https://doi.org/10.1144/GSL.SP.2000.171.01.11>
- 571 Cluzel N, Laporte D, Provost A, Kannewischer I (2008) Kinetics of heterogeneous bubble  
572 nucleation in rhyolitic melts: implications for the number density of bubbles in volcanic  
573 conduits and for pumice textures. *Contrib Mineral Petrol* 156:745–763.  
574 <https://doi.org/10.1007/s00410-008-0313-1>
- 575 Costa A, Melnik O, Sparks RSJ, Voight B (2007) Control of magma flow in dykes on cyclic lava  
576 dome extrusion. *Geophys Res Lett* 34:L02303. <https://doi.org/10.1029/2006GL027466>

- 577 Cottrell E, Gardner JE, Rutherford MJ (1999) Petrologic and experimental evidence for the  
578 movement and heating of the pre-eruptive Minoan rhyodacite (Santorini, Greece).  
579 *Contrib Mineral Petrol* 135:315–331. <https://doi.org/10.1007/s004100050514>
- 580 de' Michieli Vitturi M, Clarke AB, Neri A, Voight B (2010) Transient effects of magma ascent  
581 dynamics along a geometrically variable dome-feeding conduit. *Earth Planet Sci Lett*  
582 295:541–553. <https://doi.org/10.1016/j.epsl.2010.04.029>
- 583 Degruyter W, Bachmann O, Burgisser A, Manga M (2012) The effects of outgassing on the  
584 transition between effusive and explosive silicic eruptions. *Earth Planet Sci Lett* 349–  
585 350:161–170. <https://doi.org/10.1016/j.epsl.2012.06.056>
- 586 Doocy S, Daniels A, Dooling S, Gorokhovich Y. The Human Impact of Volcanoes: a Historical  
587 Review of Events 1900-2009 and Systematic Literature Review. *PLOS Curr Disasters*  
588 2013 Apr 16 . Edition 1.:  
589 <https://doi.org/10.1371/currents.dis.841859091a706efebf8a30f4ed7a1901>.
- 590 Druitt TH (2014) New insights into the initiation and venting of the Bronze-Age eruption of  
591 Santorini (Greece), from component analysis. *Bull Volcanol* 76:794.  
592 <https://doi.org/10.1007/s00445-014-0794-x>
- 593 Druitt TH, Edwards L, Mellors RM, et al (1999) Santorini Volcano. *Geol Soc Mem* 19:
- 594 Druitt TH, Mercier M, Florentin L, et al (2016) Magma Storage and Extraction Associated with  
595 Plinian and Interplinian Activity at Santorini Caldera (Greece). *J Petrol* 57:461–494.  
596 <https://doi.org/10.1093/petrology/egw015>
- 597 Flaherty T, Druitt TH, Tuffen H, et al (2018) Multiple timescale constraints for high-flux magma  
598 chamber assembly prior to the Late Bronze Age eruption of Santorini (Greece). *Contrib*  
599 *Mineral Petrol* 173:75. <https://doi.org/10.1007/s00410-018-1490-1>
- 600 Giachetti T, Druitt TH, Burgisser A, et al (2010) Bubble nucleation, growth and coalescence  
601 during the 1997 Vulcanian explosions of Soufrière Hills Volcano, Montserrat. *J Volcanol*  
602 *Geotherm Res* 193:215–231. <https://doi.org/10.1016/j.jvolgeores.2010.04.001>
- 603 Giordano D, Russell JK, Dingwell DB (2008) Viscosity of magmatic liquids: A model. *Earth*  
604 *Planet Sci Lett* 271:123–134. <https://doi.org/10.1016/j.epsl.2008.03.038>
- 605 Gonnermann HM, Manga M (2003) Explosive volcanism may not be an inevitable consequence  
606 of magma fragmentation. *Nature* 426:432–435. <https://doi.org/10.1038/nature02138>
- 607 Hajimirza S, Gonnermann HM, Gardner JE, Giachetti T (2019) Predicting Homogeneous Bubble  
608 Nucleation in Rhyolite. *J Geophys Res Solid Earth* 124:2395–2416.  
609 <https://doi.org/10.1029/2018JB015891>
- 610 Hamada M, Laporte D, Cluzel N, et al (2010) Simulating bubble number density of rhyolitic  
611 pumices from Plinian eruptions: constraints from fast decompression experiments. *Bull*  
612 *Volcanol* 72:735–746. <https://doi.org/10.1007/s00445-010-0353-z>

- 613 Heiken G, McCoy F (1990). Precursory activity to the Minoan eruption, Thera, Greece. In:  
614 Hardy DA (ed.) Thera and the Aegean World III, vol 2. Thera Foundation, London, pp  
615 13–18.
- 616 Heiken G, McCoy F, Jr (1984) Caldera development during the Minoan eruption, Thira,  
617 Cyclades, Greece. *J Geophys Res* 89: 8441-8462
- 618 Hildreth W, Drake R (1992) Volcan Quizapu, Chilean Andes. *Bull Volcanol* 54:93–125
- 619 Hooft EEE, Heath BA, Toomey DR, et al (2019) Seismic imaging of Santorini: Subsurface  
620 constraints on caldera collapse and present-day magma recharge. *Earth Planet Sci Lett*  
621 514:48–61. <https://doi.org/10.1016/j.epsl.2019.02.033>
- 622 Houghton BF, Carey RJ, Cashman KV, Wilson CJ, Hobden BJ, Hammer JE (2010) Diverse  
623 patterns of ascent, degassing, and eruption of rhyolite magma during the 1.8 ka Taupo  
624 eruption, New Zealand: evidence from clast vesicularity. *J Volcanol Geotherm Res*  
625 195:31-47.
- 626 Houghton BF, Carey RJ, Rosenberg MD (2014) The 1800a Taupo eruption: “III wind” blows the  
627 ultraplinian type event down to Plinian. *Geology* 42:459–461.  
628 <https://doi.org/10.1130/G35400.1>
- 629 Hurwitz S, Navon O (1994) Bubble nucleation in rhyolitic melts: Experiments at high pressure,  
630 temperature, and water content. *Earth Planet Sci Lett* 122:267–280.  
631 [https://doi.org/10.1016/0012-821X\(94\)90001-9](https://doi.org/10.1016/0012-821X(94)90001-9)
- 632 Kaviris G, Papadimitriou P, Kravvariti Ph, et al (2015) A detailed seismic anisotropy study  
633 during the 2011–2012 unrest period in the Santorini Volcanic Complex. *Phys Earth*  
634 *Planet Inter* 238:51–88. <https://doi.org/10.1016/j.pepi.2014.11.002>
- 635 Klug C, Cashman KV (1996) Permeability development in vesiculating magmas: implications  
636 for fragmentation. *Bull Volcanol* 58:87–100. <https://doi.org/10.1007/s004450050128>
- 637 Liu Y, Anderson AT, Wilson CJN (2007) Melt pockets in phenocrysts and decompression rates  
638 of silicic magmas before fragmentation. *J Geophys Res* 112:B06204.  
639 <https://doi.org/10.1029/2006JB004500>
- 640 Lloyd AS, Ruprecht P, Hauri EH, et al (2014) NanoSIMS results from olivine-hosted melt  
641 embayments: Magma ascent rate during explosive basaltic eruptions. *J Volcanol*  
642 *Geotherm Res* 283:1–18. <https://doi.org/10.1016/j.jvolgeores.2014.06.002>
- 643 Mangan M, Sisson T (2000) Delayed, disequilibrium degassing in rhyolite magma:  
644 decompression experiments and implications for explosive volcanism. *Earth Planet Sci*  
645 *Lett* 183:441–455. [https://doi.org/10.1016/S0012-821X\(00\)00299-5](https://doi.org/10.1016/S0012-821X(00)00299-5)
- 646 Martel C, Iacono-Marziano G (2015) Timescales of bubble coalescence, outgassing, and foam  
647 collapse in decompressed rhyolitic melts. *Earth Planet Sci Lett* 412:173–185.  
648 <https://doi.org/10.1016/j.epsl.2014.12.010>

- 649 Mason BG, Pyle DM, Oppenheimer C (2004) The size and frequency of the largest explosive  
650 eruptions on Earth. *Bull Volcanol* 66:735–748. [https://doi.org/10.1007/s00445-004-0355-](https://doi.org/10.1007/s00445-004-0355-9)  
651 9
- 652 Massaro S, Costa A, Sulpizio R (2018) Evolution of the magma feeding system during a Plinian  
653 eruption: The case of Pomici di Avellino eruption of Somma–Vesuvius, Italy. *Earth*  
654 *Planet Sci Lett* 482:545–555. <https://doi.org/10.1016/j.epsl.2017.11.030>
- 655 Massol H, Koyaguchi T (2005) The effect of magma flow on nucleation of gas bubbles in a  
656 volcanic conduit. *J Volcanol Geotherm Res* 143:69–88.  
657 <https://doi.org/10.1016/j.jvolgeores.2004.09.011>
- 658 Mastin LG, Guffanti M, Servranckx R, et al (2009) A multidisciplinary effort to assign realistic  
659 source parameters to models of volcanic ash-cloud transport and dispersion during  
660 eruptions. *J Volcanol Geotherm Res* 186:10–21.  
661 <https://doi.org/10.1016/j.jvolgeores.2009.01.008>
- 662 Miller CF, Wark DA (2008) Supervolcanoes and their explosive supereruption. *Elements* 4:11–  
663 15. <https://doi.org/10.2113/GSELEMENTS.4.1.11>
- 664 Mourtada-Bonnefoi CC, Laporte D (2004) Kinetics of bubble nucleation in a rhyolitic melt: an  
665 experimental study of the effect of ascent rate. *Earth Planet Sci Lett* 218:521–537.  
666 [https://doi.org/10.1016/S0012-821X\(03\)00684-8](https://doi.org/10.1016/S0012-821X(03)00684-8)
- 667 Myers ML, Wallace PJ, Wilson CJN, et al (2016) Prolonged ascent and episodic venting of  
668 discrete magma batches at the onset of the Huckleberry Ridge supereruption,  
669 Yellowstone. *Earth Planet Sci Lett* 451:285–297.  
670 <https://doi.org/10.1016/j.epsl.2016.07.023>
- 671 Myers ML, Wallace PJ, Wilson CJN, et al (2018) Ascent rates of rhyolitic magma at the onset of  
672 three caldera-forming eruptions. *Am Mineral* 103:952–965. [https://doi.org/10.2138/am-](https://doi.org/10.2138/am-2018-6225)  
673 2018-6225
- 674 Nishiwaki M, Toramaru A (2019) Inclusion of Viscosity Into Classical Homogeneous  
675 Nucleation Theory for Water Bubbles in Silicate Melts: Reexamination of Bubble  
676 Number Density in Ascending Magmas. *J Geophys Res Solid Earth* 124:8250–8266.  
677 <https://doi.org/10.1029/2019JB017796>
- 678 Pyle DM, Elliott JR (2006) Quantitative morphology, recent evolution, and future activity of the  
679 Kameni Islands volcano, Santorini, Greece. *Geosphere* 2:253.  
680 <https://doi.org/10.1130/GES00028.1>
- 681 Romine WL, Whittington AG (2015) A simple model for the viscosity of rhyolites as a function  
682 of temperature, pressure and water content. *Geochim Cosmochim Acta* 170:281–300.  
683 <https://doi.org/10.1016/j.gca.2015.08.009>
- 684 Shea T (2017) Bubble nucleation in magmas: A dominantly heterogeneous process? *J Volcanol*  
685 *Geotherm Res* 343:155–170. <https://doi.org/10.1016/j.jvolgeores.2017.06.025>

- 686 Shea T, Houghton BF, Gurioli L, et al (2010) Textural studies of vesicles in volcanic rocks: An  
687 integrated methodology. *J Volcanol Geotherm Res* 190:271–289.  
688 <https://doi.org/10.1016/j.jvolgeores.2009.12.003>
- 689 Sigurdsson H, Carey S, Alexandri G, Vougioukalakis G, & 10 others (2006) Marine  
690 investigations of Greece’s Santorini volcanic field. *EOS, Trans Amer Geophys Union* 87:  
691 337, 342
- 692 Sparks RSJ, Wilson CJN (1990) The Minoan deposits: a review of their characteristics and  
693 interpretation In: Hardy DA (ed.) *Thera and the Aegean World III*, vol 2. Thera  
694 Foundation, London, pp 89-99
- 695 Sulpizio R, Cioni R, Di Vito MA, et al (2010) The Pomici di Avellino eruption of Somma-  
696 Vesuvius (3.9 ka bp). Part I: stratigraphy, compositional variability and eruptive  
697 dynamics. *Bull Volcanol* 72:539–558. <https://doi.org/10.1007/s00445-009-0339-x>
- 698 Taddeucci J, Wohletz KH (2001) Temporal evolution of the Minoan eruption (Santorini,  
699 Greece), as recorded by its Plinian fall deposit and interlayered ash flow beds. *J Volcanol*  
700 *Geotherm Res* 19
- 701 Thomas N, Jaupart C, Vergnolle S (1994) On the vesicularity of pumice. *Journal of*  
702 *Geophysical Research: Solid Earth*, 99(B8), pp.15633-15644.
- 703 Toramaru A (2006) BND (bubble number density) decompression rate meter for explosive  
704 volcanic eruptions. *J Volcanol Geotherm Res* 154:303–316.  
705 <https://doi.org/10.1016/j.jvolgeores.2006.03.027>
- 706 Wilson L, Walker GPL (1987) Explosive volcanic eruptions - VI. Ejecta dispersal in plinian  
707 eruptions: the control of eruption conditions and atmospheric properties. *Geophys J Int*  
708 89:657–679. <https://doi.org/10.1111/j.1365-246X.1987.tb05186.x>
- 709 Wilson, CJN, Houghton, BF (1990) Eruptive mechanisms in the Minoan eruption: evidence from  
710 pumice vesicularity. *Thera and the Aegean world III*, 2, 122-128.
- 711 Woods AW, Koyaguchi T (1994) Transitions between explosive and effusive eruptions of silicic  
712 magmas. *Nature* 370:641–644. <https://doi.org/10.1038/370641a0>
- 713 Zhang Y, Xu Z, Zhu M, Wang H (2007) Silicate melt properties and volcanic eruptions:  
714 SILICATE MELT PROPERTIES. *Rev Geophys* 45:.  
715 <https://doi.org/10.1029/2006RG000216>

716

717

## 718 **Figure Captions**

719 **Fig. 1** Summary of geological features of Santorini and the Late Bronze Age (LBA) eruption after Flaherty  
720 et al. (2018) and references therein.

721 (a) Map of Santorini and its host rift zone, showing major normal faults (circle on the hanging wall),  
722 and the three main subsidence basins. The extent of submarine ignimbrite from the LBA eruption  
723 mapped from seismic profiles is shown in yellow. A location map is included as inset.  
724 (b) Map of Santorini caldera, showing the products of the LBA eruption (including the three main  
725 ignimbrite fans from eruptive phase 4), the outline of the shallow Cape Riva caldera that existed  
726 prior to the LBA eruption, the Kameni Line, and the approximate location of the Plinian vent.  
727 (c) Schematic log of the products of the LBA eruption, with the distributions of the juvenile  
728 components.

729 **Fig. 2** Field photos of the Plinian pumice fall deposit of phase 1 (P1) of the eruption, with a meter stick  
730 represented in each photo. (a) The pumice fall deposit overlying breccias from an earlier eruption,  
731 and itself overlain by the bedded tuffs of phase 2. (b) The top of P1 showing the abundance of  
732 hydrothermally stained granitoid clasts at this level. (c,d) The P1 pumice fall deposit in southern  
733 Santorini where layer P1b is underlain by layer P1a, with a fine ash separating the two. The thin  
734 lapilli fallout of P0 lies at the bases of these outcrops. GPS locations for each site can be found in  
735 Supplementary Table 1.

736 **Fig. 3** Thickness and maximum lithic size data from 43 sample locations. (a-f) isopleth data (mean diameter  
737 in cm of three axes of the five largest lithic fragments in 1 m<sup>2</sup> of each layer); (h-l) isopach data  
738 (cm). Figures (a) and (g) summarize the wind directions inferred from each technique. P0 wind  
739 direction taken from Cioni et al. (2000). The implied source vent for the eruption is shown by the  
740 black dot, matching that of Bonds and Sparks (1976).

741 **Fig. 4** Photomicrographs of melt reentrants exposed in plagioclase (a-c) and pyroxene (d) crystals. All melt  
742 reentrants clearly display a bubble at their mouth, a requirement to ensure that efficient diffusive  
743 exchange is occurring between the reentrant and surrounding melt. Simple morphologies that  
744 lacked strong bottle-necks were chosen in order to validate the 1D assumption.

745 **Fig. 5** Simplified stratigraphy of the deposits from phases 0 (P0) and 1 (P1) of the eruption. Eight pumice  
746 samples (names shown in boxes) were taken from six layers (labelled in bold), with the lithic

747 population percentages from Druitt (2014) shown as pie diagrams. Plume heights were calculated  
748 using the Carey and Sparks (1986) model from the isopleth data in Figure 3. Errors in plume height  
749 are based on uncertainties in isoline drawing. Plume heights were converted to mass eruption rate  
750 using the formula of Wilson and Walker (1987). Decompression rates ( $\text{MPa s}^{-1}$ ) are presented for  
751 melt reentrants (triangles), based on modeling of diffusion gradients preserved in plagioclase and  
752 pyroxene-hosted melt reentrants, and vesicle number densities (circles). Melt reentrant profiles  
753 were measured by FTIR spectroscopy at the University of Oregon (gray triangle), and by FTIR  
754 (open triangle) and Raman (black triangle) spectroscopy at the University of Clermont Auvergne.  
755 Vesicle number densities were converted to decompression rates using the formula of Toramaru  
756 (2006), based on the assumption of either heterogenous (gray circle) or homogenous (+ circle)  
757 vesicle nucleation. Error bars for melt-reentrant-based decompression rates are based on the  
758 goodness of fit for the modeled profiles.

759 **Fig. 6** Scanning electron microscope images, all taken at  $\times 500$ , of the eight pumice samples imaged,  
760 representing six layers of the pumice fall deposits; three pumice clasts are taken from layer P1b2.  
761 The glass is shown in white, and the vesicles in black on each image. Yellow scale in bottom right  
762 corner is  $100 \mu\text{m}$ .

763 **Fig. 7** Concentration profiles for two separate melt reentrants, RDP 18 and RDP 20, both from P1a,  
764 measured for  $\text{H}_2\text{O}$  profiles using three separate instruments. Each solid curve represents the best-  
765 fit decompression profile using the 1D diffusion model presented in Myers et al. (2018). UO:  
766 University of Oregon; UCA: University of Clermont Auvergne. For RDP 20, a fluorescent  
767 background characterizes the Raman spectra acquired at the extremities of this melt reentrant,  
768 resulting in slightly higher water contents estimated by Raman spectroscopy (see supplementary  
769 material for more analytical details).

770 **Fig. 8** Schematic showing a snapshot in time of two possible explanations why the decompression rates  
771 recorded by melt reentrants are at least 2-3 orders of magnitude lower than those recorded by



772 vesicle number densities. The figure shows two profiles through the LBA Plinian vent: (a) one  
773 perpendicular to the Kameni Line (profile A-B), and (b) the other parallel to it (profile C-D). The  
774 interpretation of each profile is as follows: (a) As magma ascends and degasses in the conduit, the  
775 viscosity (and hence pressure gradient) rises dramatically. Melt reentrants are thought to record  
776 the relatively low decompression rates ( $dP/dt$ ) deep in the conduit, and vesicle number densities  
777 (VND) to record the high decompression rates associated with fragmentation. (b) An additional  
778 effect may have been that the eruption was fed through a dike along the Kameni Line, which then  
779 focussed into a cylindrical conduit at shallow levels. Due to upwardly converging flow, the  
780 decompression rates in the dike would have been slower than those in the shallow conduit.

781

## 782 **Tables**

783 **Table 1:** Descriptions of each of the six fallout layers and parameters calculated from lithic  
784 isopleth maps.

785 **Table 2:** Vesicle size distribution parameters for each pumice clast analyzed.

786 **Table 3:** Information from individual reentrant profiles used to constrain decompression rates  
787 using the 1D code described in Myers et al. (2018). All H<sub>2</sub>O and CO<sub>2</sub> measurements are based on  
788 results from FTIR measurements at the University of Oregon (see Supplementary Table 6 for  
789 results from the other techniques). For the diffusion model, starting pressure ( $P_i=165$  MPa) and  
790 initial H<sub>2</sub>O content (5.2 wt.%) are based on the isolated melt inclusion data of Druitt et al.  
791 (2016), where final pressure ( $P_f$ ) is determined through the best-fit calculation. Conversion of  
792 decompression rate ( $dP/dt$ ) to ascent rate assumes a magmastatic pressure gradient, a magma  
793 density of 2600 kg m<sup>-3</sup> and a magma fragmentation depth at 1 km.

794

Figures 1-8

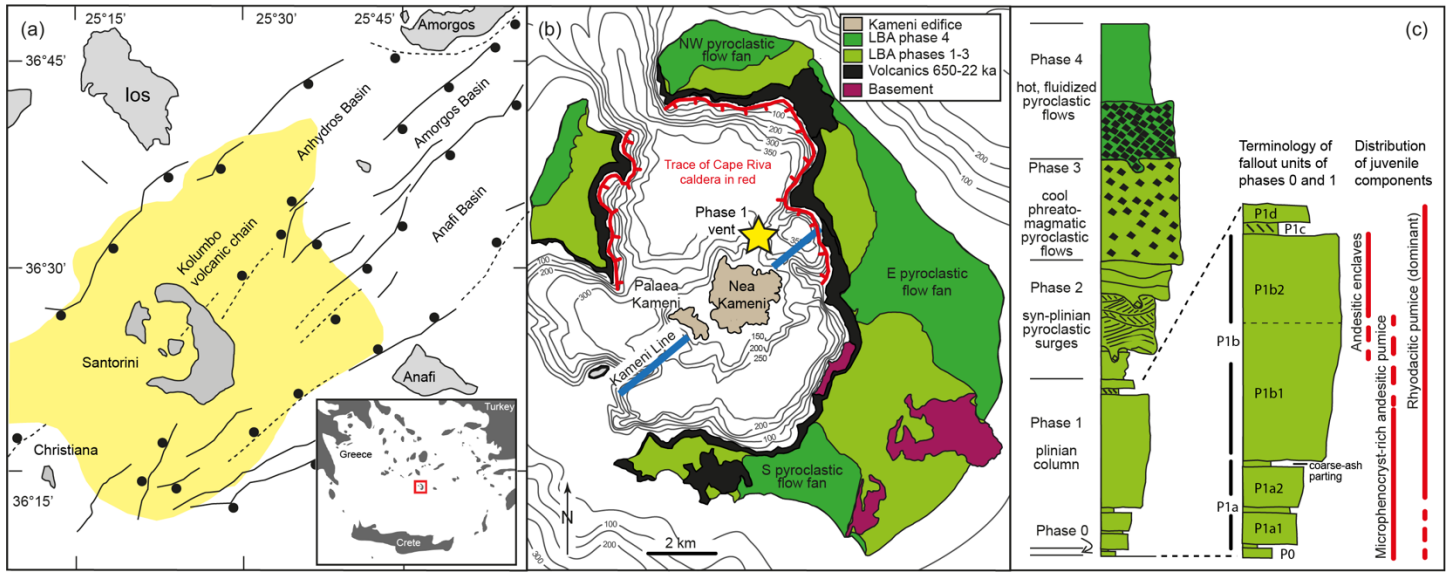
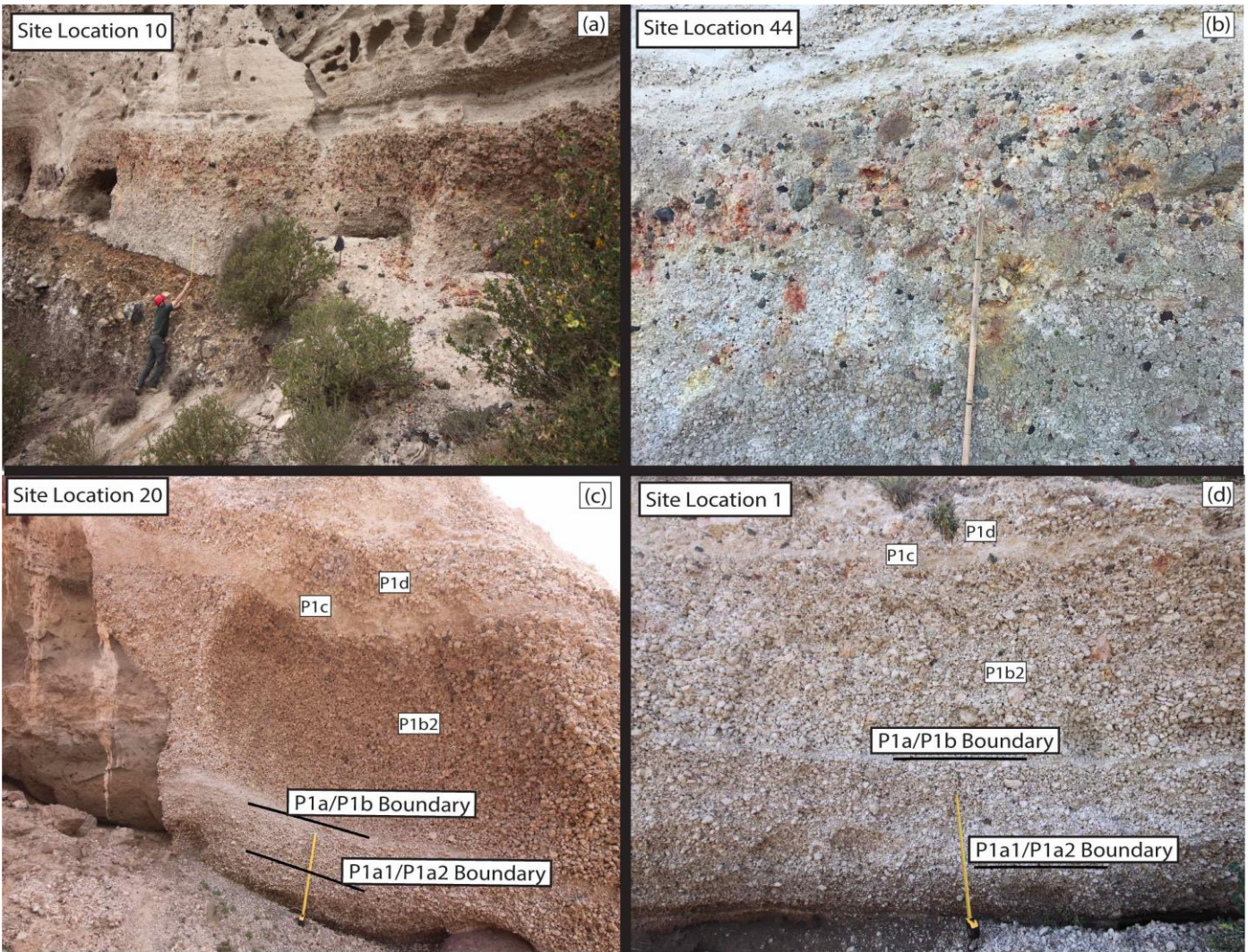
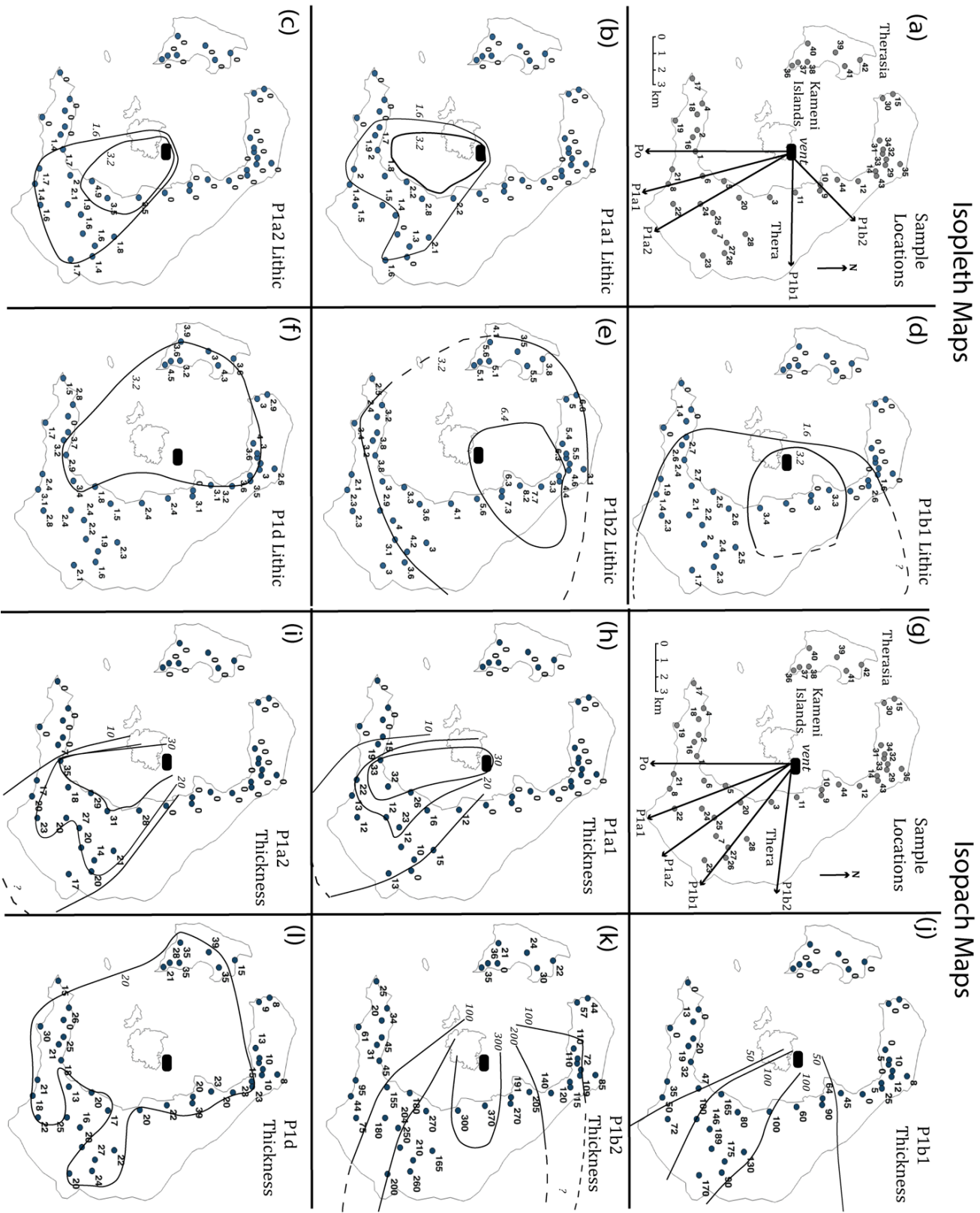


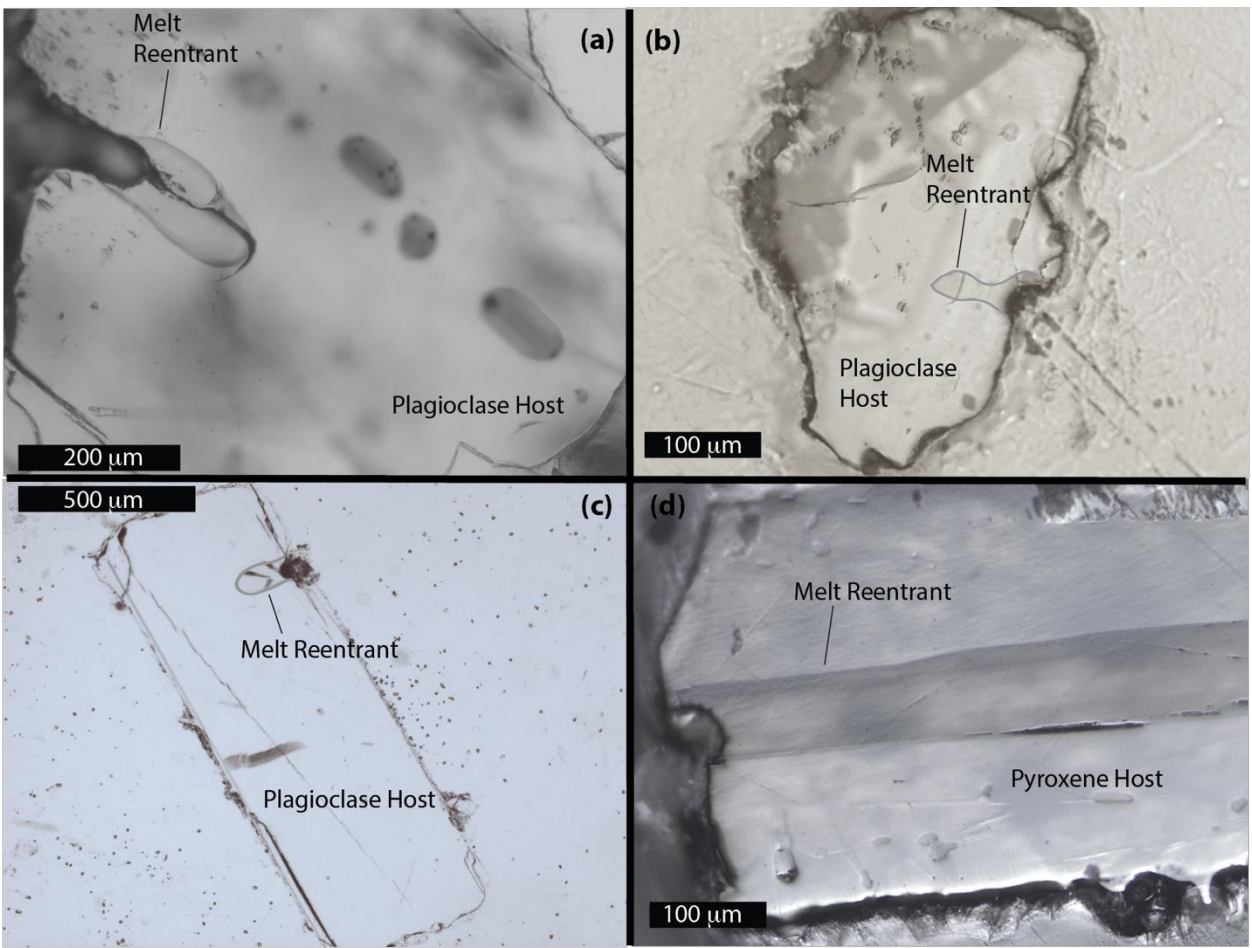
Fig. 1



**Fig. 2**



**Fig. 3**



**Fig. 4**

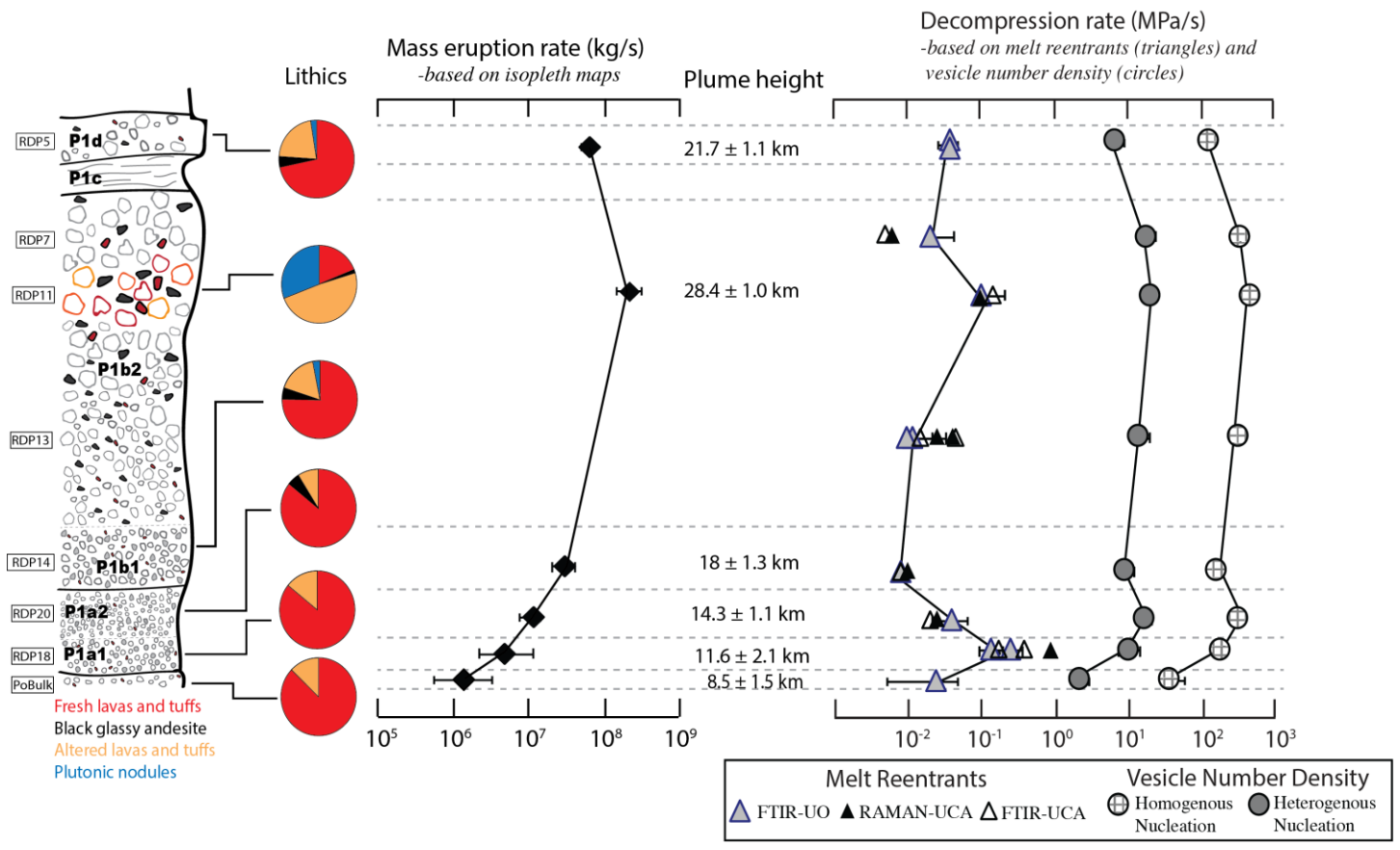
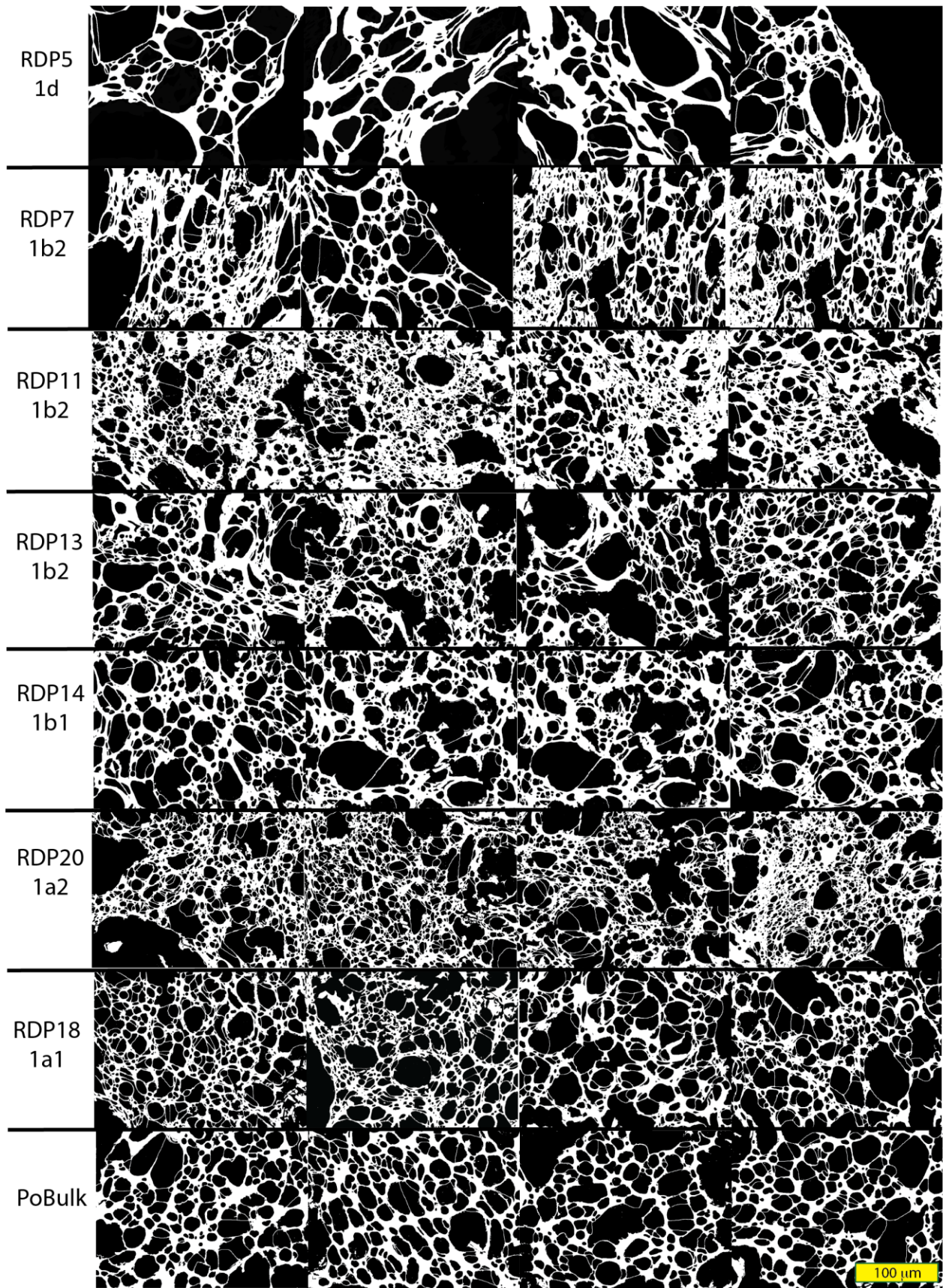
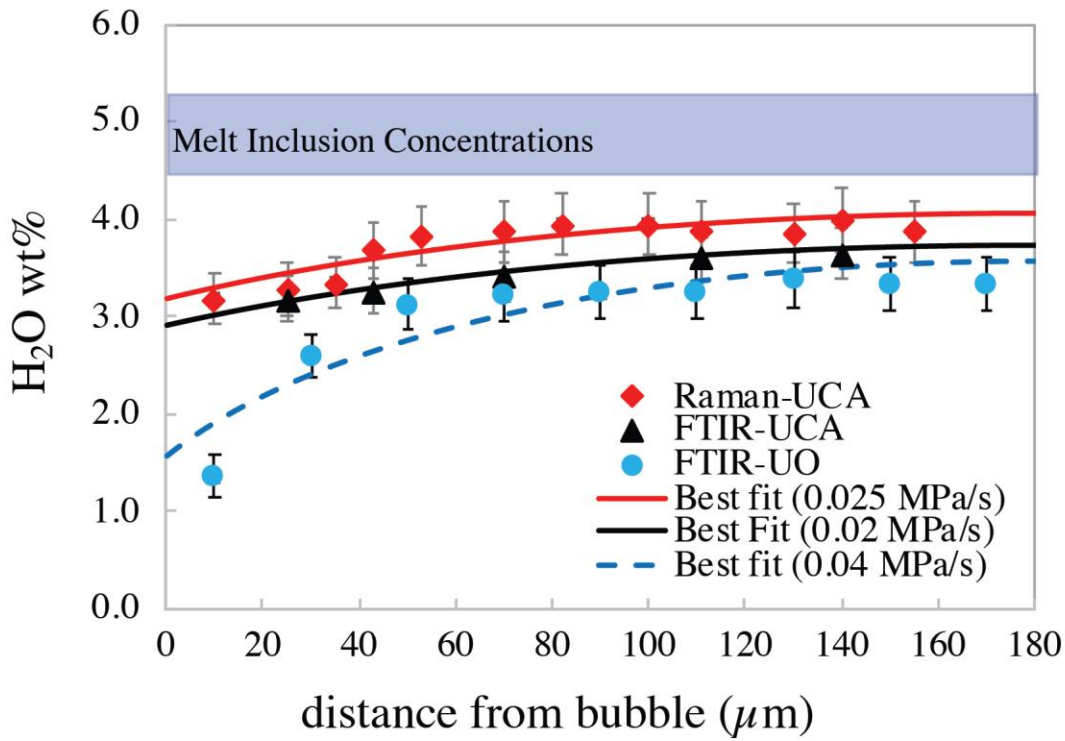
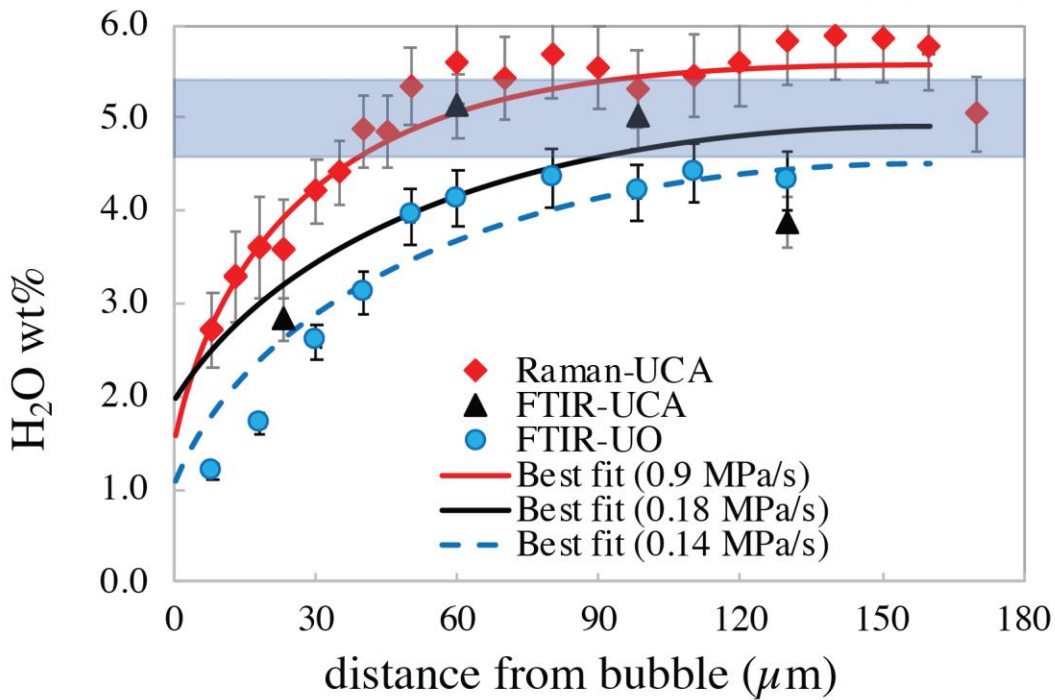


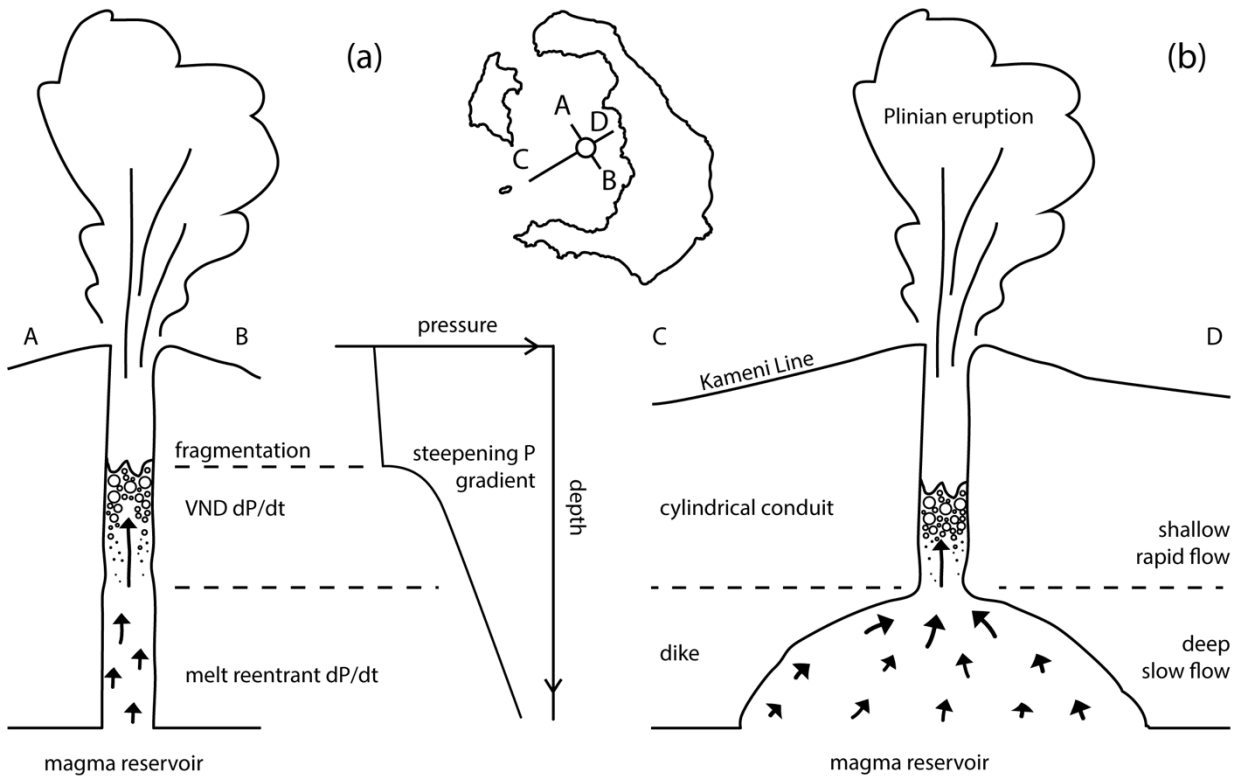
Fig. 5



**Fig. 6**

**RDP20- RE 1****RDP18- RE 2****Fig. 7**





**Fig. 8**

## Tables (1-3)

Table 1:

Layer	Characteristics	Plume Height (km)* <sup>1</sup>	Wind Direction	MER (kg/s) <sup>2</sup>	MER (kg/s) <sup>3</sup>
P0	Coarse ash fall with some microphenocryst-rich pumices. Beds A of Cioni et al. (2000)	$8.5 \pm 1.5$ <sup>4</sup>	S	$1.7 \pm 1.2 \times 10^6$	$1.0 \times 10^6$
P1a1	Lapilli pumice fall with abundant microphenocryst-rich pumices. Reversely graded. Gray coloring.	$11.6 \pm 2.1$	SSE	$5.8 \pm 3.9 \times 10^6$	$3.7 \times 10^6$
P1a2	Lapilli pumice fall with abundant microphenocryst-rich pumices. Two reversely zoned layers. Gray coloring.	$14.3 \pm 1.1$	SE	$1.3 \pm 0.4 \times 10^7$	$8.8 \times 10^6$
P1b1	Pumice fall deposit with whiter appearance. Microphenocryst-rich pumices scarce.	$18.0 \pm 1.3$	E	$3.4 \pm 1.1 \times 10^7$	$2.3 \times 10^7$
P1b2	Pumice fall deposit with white appearance. Microphenocryst-rich pumices scarce. P1b1/P1b2 boundary marked by incoming of rare cauliform andesitic scoria.	$28.4 \pm 1.0$	NE	$2.1 \pm 0.6 \times 10^8$	$1.5 \times 10^8$
P1d	Following pyroclastic surge deposit (P1c), final pumice fall layer before passage to P2.	$21.7 \pm 1.1$	No Wind ?	$7.2 \pm 1.3 \times 10^7$	$5.0 \times 10^7$

\*The plume height presented is determined using the 3.2 cm isopleth, assuming a lithic density of  $2500 \text{ kg/m}^3$ , and calculated using the plume inversion model of Carey and Sparks (1986), except for P0, where we used the value of Cioni et al. (2000). For each layer, the minimum and maximum plume height was determined based on uncertainties in the cross and downwind ranges; the average of this range is presented here along with the estimated uncertainty. Mass Eruption Rate (MER) was estimated using the methods of Wilson and Walker (1987) and Mastin et al. (2009), and the uncertainty estimated for the Wilson and Walker (1987) value. <sup>1</sup> Carey and Sparks (1986); <sup>2</sup> Wilson and Walker (1987); <sup>3</sup> Mastin et al. (2009); <sup>4</sup> Cioni et al. (2000).

**Table 2:**

Sample	Location	Fall Layer	Average Vesicle Diameter ( $\mu\text{m}$ ) <sup>a</sup>	$N_V$ ( $\text{mm}^{-3}$ )	$N_V$ <sup>b</sup> corrected ( $\text{mm}^{-3}$ )	$N_V$ <sup>c</sup> uncertainty ( $\text{mm}^{-3}$ )	$dP/dt$ <sup>c</sup> Heterog (MPa/s)	$dP/dt$ <sup>c</sup> Homog. (MPa/s)
P0	Location 1	P0	13	$6.24 \times 10^5$	$2.50 \times 10^6$	$7.58 \times 10^5$	2	50
RDP18	Location 1	P1a1	12	$6.43 \times 10^6$	$2.57 \times 10^7$	$1.26 \times 10^7$	11	240
RDP20	Location 1	P1a2	4	$1.45 \times 10^7$	$5.78 \times 10^7$	$9.04 \times 10^6$	17	420
RDP14	Location 11	P1b1/P1b2	9	$5.37 \times 10^6$	$2.15 \times 10^7$	$5.94 \times 10^6$	9	220
RDP13	Location 11	P1b2	7	$1.07 \times 10^7$	$4.28 \times 10^7$	$1.68 \times 10^7$	15	340
RDP11	Location 11	P1b2	5	$1.64 \times 10^7$	$6.56 \times 10^7$	$1.50 \times 10^7$	19	450
RDP7	Location 11	P1b2	8	$1.56 \times 10^7$	$6.26 \times 10^7$	$1.77 \times 10^7$	19	440
RDP5	Location 11	P1d	9	$3.42 \times 10^6$	$1.37 \times 10^7$	$4.16 \times 10^6$	7	160

<sup>a</sup> Vesicle diameter is an average of the two axes.

<sup>b</sup> Vesicle Number Density ( $N_V$ ) is corrected based on a vesicularity of 75%, taken from Wilson and Houghton (1990).

<sup>c</sup>  $N_V$  uncertainty is based on the 1 sigma SD of the  $N_V$  determined by analyzing each 500x image individually.

<sup>d</sup> Magma decompression rates ( $dp/dt$ ) were calculated from  $N_V$  using the equations from Toramaru (2006), using assumptions of both heterogenous and homogenous nucleation.

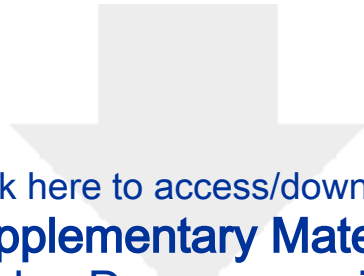
**Table 3:**

Melt Reentrant Name	Fall Layer	Length of Melt Reentrant ( $\mu\text{m}$ )	H <sub>2</sub> O Interior (wt.%)	CO <sub>2</sub> Interior (ppm)	Interior Pressure (MPa)	H <sub>2</sub> O Mouth (wt.%)	CO <sub>2</sub> Mouth (ppm)	Pressure Mouth (MPa)	dP/dt Best fit (MPa/s)	dP/dt uncertainty (MPa/s)	Ascent Rate (m/s)	Best Fit Fragmentation Pressure (MPa)
Po-1	Po	220	3.8	0	110	2.8	0	65	0.025	0.02	1.17	50
RDP18-1	P1a1	90	4.3	0	119	2.9	0	61	0.25	0.1	11	40
RDP18-2	P1a1	160	4.3	0	119	1.5	0	18	0.14	0.05	4.9	10
RDP20-1	P1a2	180	3.4	0	80	1.4	0	16	0.04	0.02	1.5	20
RDP14-4	P1b1	320	3.2	0	72	2.31	0	40	0.008	0.001	0.3	30
RDP13-1	P1b2	80	2.9	0	60	2.6	0	50	0.012	0.02	0.56	50
RDP13-5	P1b2	80	1.9	0	28	1.67	0	23	0.01	0.01	0.37	20
RDP11-6	P1b2	110	4.2	0	115	3.3	0	76	0.1	0.1	6.3	80
RDP7-5	P1b2	140	4.3	0	119	3.8	0	97	0.02	0.02	1.43	90
RDP5-1	P1d	170	3.8	0	97	2.8	0	57	0.035	0.01	1.79	60
RDP5-2	P1d	90	2.4	0	43	1.8	0	26	0.035	0.01	1.3	20



Click here to access/download  
**Supplementary Material**  
SUPPLEMENT LAB-2020\_Dec.docx





[Click here to access/download](#)

**Supplementary Material**

Supplemental Tables Decompression DEC 2020.xlsx

