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Evidence for deep gas loss in open volcanic systems

Marielle COLLOMBET ^{a*}, Alain BURGISSER ^a, Mathieu COLOMBIER ^b, Elizabeth GAUNT ^c

^a Univ. Grenoble Alpes, Univ. Savoie Mont Blanc, CNRS, IRD, IFSTTAR, ISTerre, 38000 Grenoble, France.

^b Department of Earth and Environmental Sciences, Ludwig-Maximilians-Universität München, Germany.

^c Instituto Geofísico, Escuela Politécnica Nacional, Quito, Ecuador.

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* Corresponding author: marielle.collombet@univ-smb.fr

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20 Abstract

21 Previous studies of Vulcanian eruptive products have shown that the respective volcanic conduits were filled
22 for the most part with low-porosity magma (i.e. < 10 vol%) prior to eruption. Comparison with the
23 theoretical porosity distribution expected from closed-system degassing suggests that gas loss must have
24 taken place at depth within the magmatic column (between 3 and 5 km). At such high pressures (between 70
25 and 110 MPa), however, porosities are low enough (< 20 vol%) to rule out traditional gas loss mechanisms.
26 We tested if channelling, an outgassing mechanism based on bubble connection due to high crystal content
27 (>40 vol%) proposed to occur in mushy magma reservoirs, could also happen in volcanic conduits. We
28 reanalysed phenocryst, microlite, and porosity data from recent eruptions of Merapi volcano, Indonesia,
29 Soufrière Hills volcano, Montserrat, and Tungurahua volcano, Ecuador. Overall, these magmas had crystal
30 contents high enough for outgassing to occur by channelling. Gases could be channelled out of the magma
31 columns at various levels during ascent to yield mostly gas-depleted magma columns prior to explosive
32 behaviour. Such outgassing by channelling thus has the capacity to influence eruptive style. Depending on
33 the phenocryst content, microlite growth during ascent can either foster, or impede gas escape by
34 channelling. Considering the pervasive occurrence of microlites and ensuing high crystal contents in
35 volcanic conduits (between 40 and 70 vol%), the high likelihood of channelling implies that other outgassing
36 mechanisms might not be as dominant as previously envisioned.

37

38 **Keywords:** Silicic volcanoes, outgassing, crystal content, channelling, porosity

39 Introduction

40 The eruption of viscous magma at the Earth's surface often gives rise to abrupt regime changes, such as the
41 transition from the gentle effusion of a lava dome to brief but powerful Vulcanian explosions. These eruptive
42 regimes are controlled by outgassing, a process by which the gas contained in bubbles formed during ascent
43 can escape from the magma. In the case of highly viscous magmas (viscosity $>10^6$ Pa.s), individual bubbles
44 are trapped within the melt and cannot migrate individually through the magmatic column, hindering the
45 bubbly flow outgassing mechanism that is generally at play for low viscosity magmas (Parmigiani et al.
46 2016; Cardoso and Woods, 1999). Gas loss within highly viscous magmas is mainly related to bubble

connectivity and to percolation, which controls the ability for the gas to travel through these bubble connections, making the magma permeable under specific physical conditions. Pavourable physical conditions for percolation to occur depend on the amount, size, and shape of the bubbles, deformation, which promotes bubble connectivity (e.g., Blower 2001; Rust and Cashman 2004; Burgisser et al. 2017; Kushnir et al. 2017), and the presence of crystals (Lindoo et al. 2017; deGraffenried et al. 2019; Degruyter et al. 2019; Colombier et al. 2020). In some cases, brittle failure and cracks can also appear (Shields et al. 2014). When quantified, permeabilities deduced from the combination of these parameters range between 10^{-15} and 10^{-12} m² (e.g., Klug and Cashman, 1996; Farquharson et al. 2015; Kushnir et al. 2016; Burgisser et al. 2017). When integrated into numerical models, they contribute to reproduce effusive regimes that are compatible with field observations like bubble density distribution described on an eroded conduit, or extrusion rate during dome-forming activity (e.g., Melnik and Sparks, 2002; Collombet, 2009; Degruyter et al. 2012; Cassidy et al. 2018).

However, these numerical models fail to reproduce the very low porosity values inferred to occur in the conduit prior to Vulcanian eruptions. Figure 1 shows the porosity distribution prior to the Vulcanian explosion on February 10, 2010 at Soufrière Hills volcano, Montserrat (Burgisser et al. 2019). Samples collected in the field provide porosity data down to approx. 70 MPa (~3 km depth) that are significantly below the theoretical porosities suggested for a closed system. The outgassing mechanisms, which are based on bubble connection relationships implemented in various numerical models, could explain low porosity values near the surface but fail to reproduce porosity values at depth (beyond a corresponding pressure of 10 MPa). This implies that another outgassing mechanism is at play and/or that a large amount of gas has been evacuated from the deepest parts of the magma column (most likely above 70 MPa) during and/or before the ascent from the reservoir.

We test the hypothesis that this mismatch is due to a missing physical mechanism that greatly enhances outgassing efficiency within the conduit, even at very low porosity values (<10 vol%). This mechanism, channelling, which involves a rigid crystal framework, was first described for the case of magma chambers by Parmigiani et al.(2017) and Degruyter et al. (2019). Channelling relies on the capacity of crystal-rich

magmas to build sustainable channels for gas percolation independently from magma movements. For crystal volume fractions between 40 and 70 vol%, this mechanism allows gas permeability to reach the order of 10^{-10} m², which is two to five orders of magnitude higher than permeability calculated for bubbles alone at an equivalent porosity (Klug and Cashman, 1996; Burgisser et al. 2017). Under such permeable conditions, it is much more plausible that magma succeeds to extensively degas and outgas, even at the high pressures and low gas volume fractions typical of the deepest parts of a volcanic conduit. In addition, the formation of such channel-like porous pathways reduces significantly the percolation threshold, which is the critical vesicularity at which system-spanning bubble connectivity and permeability occur (e.g., Lindoo et al. 2017; Colombier et al. 2020). As a consequence, provided that crystallinity is high enough, outgassing can occur before or early during ascent, resulting in a magma column severely depleted in gas.

We revisited three data sets of the porosity distribution within volcanic conduits filled by andesitic magmas. They are from a Vulcanian explosion on February 11, 2010 at Soufrière Hills volcano, Montserrat (Burgisser et al. 2019), an explosive event on October 26, 2010 at Merapi volcano, Indonesia (Drignon et al. 2016), and a large Vulcanian event that occurred on July 14, 2013 at Tungurahua volcano, Ecuador (Gaunt et al. 2020). We show that the magmas feeding these Vulcanian events were crystal-rich enough to promote highly efficient outgassing by channelling.

Geological Context

The on-going eruption of Soufrière Hills is dominated by lava dome effusion with intermittent collapse. The magma emitted is andesitic in bulk composition with a rhyolitic melt. The fifth phase of lava extrusion since the beginning of the eruption was marked by a succession of dome growth as lava lobes and spines followed by partial dome collapse (Wadge et al. 2014b; Cole et al. 2014). The February 11, 2010 explosion was preceded by two small Vulcanian events on February 5 and 8 (Stinton et al. 2014b; Cole et al. 2015). The February 11 event was mostly likely triggered by a dome collapse that started 1 hour and 27 minutes prior to the onset of the Vulcanian explosion. Burgisser et al. (2019) analyzed twenty-three samples of the resulting pumice-rich, pyroclastic density current deposits and pumice fallout deposits. Phenocrysts are mostly

102 plagioclases that are up to 5 mm across (Higgins and Roberge, 2003). Large plagioclases have irregular
103 shapes and smaller plagioclases are euhedral. Euhedral amphiboles up to 12 mm across are also present with
104 a low number density. Microlites are mostly plagioclases with a concave up distribution in the 20-1000 μm^2
105 range (Couch et al. 2003) and short/long axis ratios of ~ 0.5 (Murch and Cole, 2019).
106
107 Merapi volcano produced in 2010 the largest explosive eruption since 1872 (Surono et al. 2012). This
108 unusual event, which emitted basaltic andesites with a rhyolitic melt (Costa et al. 2013), occurred in several
109 stages between October 26 and November 23 (Komorowski et al. 2013; Jenkins et al. 2013). The first large
110 explosion lasted for 2h on October 26, producing an ash plume and pyroclastic density currents. This
111 explosion had a phreatomagmatic component and was interpreted as a laterally directed explosion from a
112 newly formed cryptodome. Drignon et al. (2016) analyzed forty-one samples from a pumice levee of the
113 valley-filling pyroclastic deposits produced by the October 26 event. Phenocrysts populations vary little
114 between the recent eruptions (van der Zwan et al. 2013). Crystals that are 1.6–4 mm across make up one
115 population with a size distribution that is approximately a straight line. The most common phenocrysts are
116 plagioclases and pyroxenes that both have subhedral morphologies with evidences of resorption (Erdmann et
117 al. 2016). Microlites are mostly plagioclases with a rectangular prism habit, areas of 10–15 μm^2 , and
118 short/long axis ratios ranging from 0.35 to 0.67 (Costa et al. 2013).
119
120 The latest activity (1999 – 2016) at Tungurahua volcano varied from small-scale gas and ash venting and
121 Strombolian outbursts to rare sub-Plinian events (Eycheenne et al. 2012; Hall et al. 2015; Battaglia et al.
122 2019). The magma emitted is a low silica andesite with a high silica andesitic melt. The last six years of this
123 eruptive phase was marked by two to three months of relative quiescence followed by Vulcanian events and
124 ending with several weeks of low-explosivity Strombolian activity and ash venting (Hidalgo et al. 2015). On
125 July 14, 2013, an unusually large Vulcanian event occurred at 06h46 (local time), producing a large ash
126 column and pyroclastic density currents that travelled up to 7.5 km away from the vent. Gaunt et al. (2020)
127 analyzed twenty-seven samples from three lithological groups (pyroclastic current deposits, fallout pumice,
128 and ballistic ejecta) present in the ejecta. Plagioclase is the most abundant phenocryst phase, followed by
129 pyroxene (Samaniego et al. 2011). Both exhibit mostly euhedral habits. Microlites are mostly plagioclase

130 with habits ranging from rectangular prisms to acicular crystals (Gaunt et al. 2020). Dense plug rocks have
131 mean microlite areas of 4.2–26 μm^2 with mostly tabular morphologies. Juvenile vesicular clasts of the
132 pyroclastic current deposits have mean microlite areas of 10–66 μm^2 with occasional swallow-tail and
133 hopper morphologies.

134

135 Methods

136 We modified data treatment of the recompression model of Burgisser et al. (2010) used by Drignon et al.
137 (2016), Burgisser et al. (2019), and Gaunt et al. (2020). We briefly outline the data acquisition and the data
138 treatment done in these three studies before describing our modification. The petrological and textural
139 analysis of the samples was carried out in the three studies according to the procedures outlined in Burgisser
140 et al. (2010) and Drignon et al. (2016). Polished sections were imaged by Scanning Electron Microscopy
141 (SEM) or by element mapping based on Energy Dispersive Spectroscopy (EDS). Images were taken at
142 several magnifications depending on the average size of the vesicles and crystals. Images were used to
143 quantify the amounts of vesicles, glass, phenocrysts, and microlites in the samples by manually tracing each
144 of the elements (SEM) or by automatic thresholding and by successive subtractions of Boolean combinations
145 of the binary images of the analyzed elements (EDS). The segmented images were then used to measure the
146 proportions of phenocrysts, microlites, and vesicles. In this study, we use “vesicles” and “vesicularity” to
147 characterize voids regardless of genesis. We use “bubbles” and “porosity” to characterize voids resulting
148 from the exsolution of magmatic volatiles.

149

150 Imaged vesicles were subdivided in four types (Fig. 2, Burgisser et al. 2010; Giachetti et al. 2010). The
151 first type was composed of large, deformed vesicles of equivalent diameter $>300 \mu\text{m}$ and circularity <0.2 .
152 The second type was composed of small, rounded vesicles of equivalent diameter $<40 \mu\text{m}$ and circularity
153 >0.65 . The third type was composed of large angular voids between crystal fragments and of irregular voids
154 delimited by crystal faces that are found in dense, diktytaxitic samples. The last type was composed of the
155 remaining vesicles, which are of intermediate sizes and often composed of coalesced bubbles. Type 1
156 vesicles were caused by degassing during magma ascent from the reservoir. The sudden decompression

157 linked to the explosion nucleated new, syn-explosive bubbles. The first syn-explosive bubbles are of Type 4
158 because they had enough time to grow and coalesce before quenching. The latest syn-explosive bubbles are
159 of Type 2 because they were quenched as small, isolated vesicles. Finally, Type 3 vesicles that are gaps
160 between crystal fragments result from crystal failure caused by the decompression accompanying conduit
161 evacuation. The remaining Type 3 vesicles found in diktytaxitic samples are likely due to reorganization of
162 voids under shear (Laumonier et al. 2011).

163
164 The recompression model of Burgisser et al. (2010) assumes that, prior to explosion, the piece of magma to be
165 ejected is at a pressure P_i and porosity, ϕ , with water weight fraction distributed as exsolved, x_i , and
166 dissolved, s_i . The clast is quenched at pressure P_q , porosity ϕ_q , and dissolved water content, s_q . Quench
167 porosity, ϕ_q , is the sum of volume fractions of the four vesicle types. Only Type 1 vesicles were present in the
168 conduit prior to the explosion. Type 2 vesicle nucleate during the explosion but were not affected by
169 outgassing. Type 3 and 4 vesicles both appeared during the explosion and could participate to outgassing
170 because they feature various degrees of interconnection. The weight fraction of the exsolved water content at
171 quench time is thus composed of an outgassed part, x_{out} , and a part remaining within the clast that is divided
172 into three weight fractions: x_{q1} (Type 1 vesicles), x_{d1} (Type 2 vesicles), and x_{q2} (Type 3 and 4 vesicles).
173 Similarly, quench pressures differ from pre-explosive pressures. The conversion from quench to pre-
174 explosive conditions (Fig. 2) was done by a mass balance linking the state of the magma just before the
175 explosive event (pre-explosive state characterized by P_i , ϕ , s_i and x_i) to that once the magma has quenched
176 (quench state characterized by P_q , ϕ_q , s_q , x_{q1} , x_{d1} , and x_{q2}).

177
178 Model input data are the magma temperature, the densities of melt and crystals at that temperature, ϕ_q ,
179 and, to obtain P_i , one of plagioclase microlite content, or glass water content. In Gaunt et al. (2020), P_i was
180 obtained by converting the glass-referenced volume proportions of plagioclase microlites of each sample into
181 pressure. In Drignon et al. (2016) and Burgisser et al. (2019), P_i was calculated by using the glass water
182 content determined by elemental analyzer and converting successively glass water content into dissolved
183 water content and into pressure with a solubility relationship. The original recompression model has four free
184 parameters (Burgisser et al. 2010). The first one, $O = x_{out}/(x_{q1} + x_{q2} + x_{d1} + x_{out})$, quantifies the gas fraction

185 outgassed during the explosion to propel the ejecta. This poorly constrained parameter, which has a modest
 186 effect on pre-explosive values (Burgisser et al. 2010), is assumed to lie between 10% and 70–82% with a
 187 reference value of 50%. The second parameter, H , estimates the maximum, near-instant pressure drop that
 188 clasts can sustain before breaking (Mueller et al. 2008) as a proxy for quench pressure. In other words, P_q is
 189 the smallest of P_i and $P_{atm} + H P_{Mueller}$, where P_{atm} is the atmospheric pressure, $P_{Mueller}$ is the value of the
 190 pressure drop, and H is a multiplicative factor that was assumed to lie between 0.5 and 2 with a typical value
 191 of 1. The upper limits of O and H were chosen so that the maximum total water content of every clast lies
 192 below that inferred in the reservoir and that all clasts undergo syn-explosive inflation so as to match textural
 193 observations. The upper values of O were 70, 76, and 82% for Tungurahua, Soufrière Hills, and Merapi,
 194 respectively. The two last parameters of the model quantify the proportions of vesicles: $E=(x_{dl}+x_{q2})/x_{q1}$ and
 195 $F=x_{dl}/x_{q1}$. As vesicle types occur in similar proportions in the products of a given explosion, Burgisser et al.
 196 (2010) and the authors of the three data set studied herein (Drignon et al. 2016; Burgisser et al. 2019; Gaunt
 197 et al. 2020) used a lumped approach; they averaged vesicle proportions in all samples of a given explosion.
 198 They then calculated E and F using the median proportions and standard deviations of each vesicle type.
 199
 200 In this study, we attempted to gain precision in the pre-explosive porosities by considering for each clast the
 201 respective proportions of the vesicle types. These proportions were published in the original studies. Using
 202 individual vesicle proportions means that E and F were no longer considered as free parameters with a
 203 median value and a standard deviation. Instead, they were considered as input parameters with an assumed
 204 negligible uncertainty, which reduced the number of degrees of freedom of the recompression procedure
 205 from four to two. In the Drignon et al. (2016) study of the October 26 event at Merapi, for instance, the
 206 vesicles of Types 1, 2, and 3–4 were present in respective median proportions of 28.4:2.5:70, which yields
 207 $E=2.7$ and $F=0.14$. We replaced these median proportions by the individual values of each sample (AME 10
 208 A1: 33.5:3.1:63.4, etc.) to obtain E and F values for each sample. Beyond reducing the number of sources of
 209 uncertainties, this modification evidenced a shortcoming of the vesicle size distribution analysis, which led
 210 us to discard data from Drignon et al. (2016) (Supplementary Text S1).
 211

212 The recompression model has two types of uncertainty. Analytical uncertainties result from error propagation
 213 of the uncertainties on quench vesicularities, glass water contents, and plagioclase microlite proportions.
 214 Analytical uncertainties for each sample were calculated with four sets of model outputs that used the
 215 reference values of H and O and the respective minimum and maximum values of ϕ , microlite content and/or
 216 glass water content. Analytical errors on pre-explosive porosity are small in absolute value (<1 vol%). Our
 217 new procedure with fewer free parameters does not significantly affect these absolute errors ($<15\%$).
 218 Similarly, errors on pre-explosive pressure are modified by $<10\%$ and typically by $<2\%$. These changes will
 219 not be discussed further as they are smaller than those stemming from model uncertainties. This second type
 220 of uncertainty results from propagation of the uncertainties on H and O . The reference values and bounds of
 221 these two parameters were combined to yield five sets of pre-explosive pressures and porosities. The set with
 222 the reference values was kept as the average set and the two sets with the largest and smallest porosity values
 223 at any pressure were kept as extrema to characterize model uncertainty. Only these three sets are reported in
 224 the Results section.

225
 226 In the case of a rigid crystal network with crystal volume fraction ϕ_x between 0.4 and 0.7, the minimum
 227 porosity required to make percolation possible (i.e. the percolation threshold), ϕ_{cr} (Fig. 2), follows a third-
 228 degree polynomial equation-:

$$229 \quad \phi_{cr} = a\phi_x^3 - b\phi_x^2 + c\phi_x + d \quad (1)$$

230 where the constants are $a=0.7495$, $b=0.4268$, $c=-0.1626$, and $d=0.1478$ for Degruyter et al. (2019) or $a=2.75$,
 231 $b=2.79$, $c=0.6345$, and $d=0.0997$ for Parmigiani et al. (2017). The difference between these two sets of
 232 constants is due to the initial conditions chosen for the bubbles. In the case of Parmigiani et al. (2017), the
 233 initial positions of the bubbles and their respective size was set randomly, whereas these parameters followed
 234 a spinodal distribution in the case of Degruyter et al. (2019). A spinodal distribution means that the size and
 235 the position of bubbles satisfy the principle of mass conservation and capillary stresses, which leads to more
 236 consistent and more accurate initial conditions. This higher level of physical verisimilitude led us to use the
 237 set of parameters of Degruyter et al. (2019).

238
 239 The vesicle-free crystal content, ϕ_{xvf} , is related to ϕ_x by (Fig. 2):

$$\phi_x = (1 - \phi)\phi_{xvf} \quad (2)$$

Drignon et al. (2016) reported bulk volume fractions of vesicles, phenocrysts and microlite, from which ϕ_{xvf} (Table 1) was calculated using Eq (2). Burgisser et al. (2019) reported ϕ_q and ϕ_{xvf} . We used their original unpublished quantifications of microlites to obtain vesicle-free volume fractions of phenocrysts. Gaunt et al. (2020) reported bulk volume fractions of vesicles, of plagioclase and pyroxene phenocrysts, and of plagioclase and pyroxene microlite. We used their original unpublished quantifications of oxide phenocrysts and microlites to calculate ϕ_{xvf} .

For a given vesicle-free crystal content, there is one critical porosity such that $\phi = \phi_{cr}$:

$$a(1 - \phi)^3 \phi_{xvf}^3 - b(1 - \phi)^2 \phi_{xvf}^2 + c(1 - \phi)\phi_{xvf} + d - \phi = 0 \quad (3)$$

For each ϕ_{xvf} within the 40–70% interval, we solve equation (3) to obtain a solution $\phi_{cr} = f(\phi_{xvf})$. The associated uncertainties, $d\phi_{xvf}$, were obtained by error propagation and mostly reflect natural variability. Uncertainties on the critical porosity, $d\phi_{cr}$, were calculated using the variation of the f function: $d\phi_{cr} = df(d\phi_{xvf})$.

Results

Considering the vesicle proportions of each sample instead of an average value does not change much (<17%) the final values of pre-explosive pressures (Fig. S1). When using microlites (Gaunt et al. 2020), no pressure change occurs because the power-law relationship linking microlite proportion to pressure does not depend on vesicle content. Pre-explosive porosities are more influenced by this reduction of free parameters. Samples located at pressures >10 MPa change by <10% from the original value. Shallower samples undergo porosity changes of up to 25% with one outlier at 40% (Fig. S2), but the general trend with pressure remains (Fig. S1). Changes in model uncertainties compared to the ones from the original studies are small but uneven between data points, varying by <15% from the original value except for a couple of samples with vanishing porosity. As a result, the overall respective envelopes of model uncertainties are similar for each volcano.

Two conditions need to be met for outgassing to occur by channelling (Parmigiani et al. 2017; Degruyter et al. 2019). The first condition is that total crystal content must be between 40 and 70 vol%. Provided the germane amount of crystals, the second condition is that a critical porosity is reached. Channelling allows gas escape at very low gas volume fraction with a percolation threshold below 10 vol%. We used the critical values given by the empirical relationship of Degruyter et al. (2019) because they correspond to more realistic initial conditions, but similar results are found when using the Parmigiani et al. (2017) values. To take the empirical nature of these fits into account, we also considered the maximum and minimum percolation values they yield within the validity bounds of crystal content (5 and 12 vol% porosity, respectively). Once channelling is established, outgassing rate is controlled by a balance between magma permeability, driving pressure, and the collapse rate of the medium. As a result, outgassing could evacuate more gas than the critical porosity. Such hysteresis has been proposed to occur at low (Rust and Cashman, 2004; Michaut et al. 2009; Gonnermann et al. 2017) and high crystal contents (Colombier et al. 2020), albeit not by channelling. In the absence of more specific evidence, we considered that non-vanishing porosities lower than the percolation threshold could result from channel collapse.

To test if channelling could explain low porosity distributions within volcanic conduits before explosion, we assessed to which extent these two conditions of crystallinity and porosity were met in three sets of data on andesitic products from different volcanoes. These datasets are composed of the pressures associated with the corresponding pre-eruptive porosities and of the amounts of phenocrysts (which are generally already present in the magmatic chamber) and microlites (which develop during magma ascent). It is important to consider these two populations of crystals since they could, depending on how their amounts change during ascent, prevent or enhance channelling.

Montserrat

The amounts of “phenocrysts” and “phenocrysts plus microlites” in the Soufrière Hills samples from 2010 are plotted in Fig. 3a. Microlite crystallization during ascent increased the total amount of crystals from 30 vol% to approximatively 55 vol% on average, thereby helping channelling to develop. Figure 3b shows pre-

explosive porosities of the samples, of closed-system degassing, and of the channelling threshold. All threshold porosities are within a narrow range smaller than 5 vol%. Considering that all samples have a porosity well below that of closed-system degassing, it means that most of the gas present in the conduit has been lost. Physical conditions propitious to channelling were thus probably met during ascent. However, it appears that most measured porosities are slightly below the threshold values, except two points among the shallowest samples and one point around 57 MPa that are slightly above threshold. This situation indicates that, if channelling occurred, a little more than the gas in excess of the threshold value has been lost, deflating the magma back below the critical porosity.

Merapi

Data from the October 26, 2010 Merapi eruption are particularly interesting because they provide information down to ~10 km (270 MPa), which is close to the pure water exsolution level for that eruption (Erdmann et al. 2016; Drignon et al. 2016). Estimating the exsolution level at Merapi is a complex task because 1) a significant amount of CO₂ coexisted with H₂O in the gas phase and 2) the magmas feeding the 2010 eruption were stored at multiple levels and mixed prior to their final ascent towards the surface (Costa et al. 2013; Nadeau et al. 2013; Preece et al. 2014; Erdmann et al. 2016; Widiyantoro et al. 2018). The mixing process started by magma assembly at depth and continued during ascent. This induced complexities in the crystal cargo that hinders straightforward interpretations of porosity and crystal contents (Supplementary Text S2). Here, we framed the likely conduit conditions according to the experimental results of Erdmann et al. (2016) by performing two closed-system, multicomponent decompressions. Following Drignon et al. (2016), we used the model D-Compress (Burgisser et al. 2015) to simulate the porosity evolution of rhyolitic melts undergoing isothermal, equilibrium decompression at NNO+1. The first decompression represents the hotter recharge magma at 975 °C. It starts from 250 MPa with 5 wt% melt H₂O content and 400 ppm melt CO₂ content, which yields a CO₂/H₂O molar ratio of 0.2. The second represents the colder resident magma at 925 °C. It starts from 175 MPa with 3 wt% melt H₂O content and 500 ppm melt CO₂ content, which yields a CO₂/H₂O molar ratio of 0.66. The resident and recharge magmas were mixed in a way that we expect to fall within these two end members. The initial porosity is mostly unconstrained, so we considered a small (<< 1 vol%) and a large (10 vol%) value for each magma.

320

321 Figure 4a shows the phenocrysts and total crystal content for Merapi conduit prior to the October 26
322 explosion. Unlike at Soufrière Hills, phenocrysts alone are already within the range for which channelling is
323 possible. The amount of crystals already present in reservoir(s) was thus sufficient to foster outgassing at
324 depth without magma ascent if sufficient gas was present. There was also enough time for the awakened
325 reservoir to degas during the weeks to months prior to eruption (Budi-Santoso et al. 2013). If microlites are
326 taken into account, however, the total crystal content clearly rises above 70 vol% in most of the magmatic
327 column. Thus, if outgassing occurred by channelling, this outgassing mechanism took place earlier or during
328 the first phase of microlite crystallization. Once 70 vol% of crystals was reached, channelling mostly likely
329 stopped.

330

331 Pre-eruptive porosities, percolation thresholds, and associated uncertainties are shown in Fig. 4b. Above ~70
332 MPa, samples suggest that the conduit was filled with nearly gas-free magma, well below percolation
333 threshold. This observation makes the two scenarios starting with high initial porosities unlikely, so they
334 were not given further consideration. Assuming that the porosity evolution was controlled by an initially gas-
335 poor recharge magma, no bubbles are expected at pressures higher than 250 MPa, which is consistent with
336 the very small values (<1 vol%) of the samples located around that pressure. At ~150 MPa, the gas available
337 by closed-system degassing matches that needed for channelling (black line on Fig. 4b) although measured
338 porosities are well below percolation. Assuming instead that the eruptive products sampled the resident
339 magma brings the pressure at which closed-system porosities are within the threshold range from 150 to 70
340 MPa (green line on Fig. 4b). This situation is compatible with the vanishing measured porosities at higher
341 pressure and the fact that porosities increase below 70 MPa while remaining below the threshold porosities.

342

343 The combined data of Figs. 4a and 4b suggest that deeper than 3 km (70 MPa), there was not enough gas
344 available in the reservoir to be outgassed despite the favourable phenocryst content for channelling. Between
345 3 km and the surface, however, gas contents stemming from closed-system degassing would have been high
346 enough for outgassing to occur. Shallower than 3 km, crystal contents vary from 40–50 vol% before
347 microlite crystallization, to more than 70 vol % post-crystallization. If the shallow magma column stagnated

348 while (or before) microlite crystallization occurred, channelling could have started before microlite
349 crystallization and continued until 70 vol% of crystals was reached. If the magma still contained substantial
350 gas volume fraction once this high limit of crystallinity reached, other mechanisms, such as cracks due to
351 brittle failure of the melt, should be evoked to explain further outgassing. In any case, the degassing trends
352 suggest that the analysed clasts sampled the colder resident magma. As in the case of Soufrière Hills, pre-
353 eruptive shallow samples display porosities below the critical threshold, possibly suggesting outgassing
354 accompanied by deflating.

355 Tungurahua

356 Storage conditions at Tungurahua were more H₂O-rich than at Merapi, with a molar fraction of H₂O in the
357 gas phase of 0.8 (Andújar et al. 2017). As a result, D-Compress runs with mixed H₂O-CO₂ fluids do not
358 differ much from pure H₂O ones, and closed-system decompressions with pure water were assumed instead
359 for simplicity. Porosities and crystal contents from Tungurahua give information from the surface down to ~7
360 km depth (170 MPa), which, like at Merapi, is deep enough to constrain the point at which gas loss starts.
361 Similarly to Soufrière Hills, phenocrysts in the Tungurahua magma cannot trigger channelling alone (Fig.
362 5a). The total amount of crystals, however, approaches the lower limit of 40% around 100 MPa, and exceeds
363 it for pressures lower than 50 MPa. This implies that microlites were required for channelling to occur
364 possibly around 100 MPa, and more certainly around 50 MPa. Near the surface, the upper limit of 70% is
365 overcome, stopping channelling and potentially fostering a shallow gas-rich area that appears clearly on the
366 porosity data with values spanning from 25% to 50 vol% (Fig. 5b). Closed-system porosities suggest that gas
367 was available for channelling at ~150 MPa, although the amount of crystals was not sufficient at this depth to
368 foster channelling (Fig. 5a). The first calculated pre-eruptive porosity approaching the channelling threshold
369 is located around 110 MPa and the first pre-eruptive porosity which effectively coincides with that threshold
370 appears at ~50 MPa (Fig. 5b). As a result, gas loss by channelling may have started between 110 and 50 MPa
371 and was no longer possible below 10 MPa.

372

373 Discussion

374 Our data show that, to first order, channelling is a process that is consistent with the low porosities inferred
375 in volcanic conduits prior to Vulcanian explosions. That channelling can occur during ascent due to an
376 increase in crystal content brings a partial answer to the long-standing issue of gas transfer across the entire
377 depth of magmatic systems (e.g., Oppenheimer et al. 2009) despite sometimes very high magma viscosity
378 and no obvious physical mechanism of transfer (Edmonds et al. 2010). Gas percolation in magmas has long
379 been studied in decompression experiments, which have notably shown that the presence of >20 vol%
380 crystals lowers percolation thresholds compared to crystal-poor melts (Lindoo et al. 2017; deGraffenried et
381 al. 2019). Such type of viscous percolation occurs when bubbles can expand spherically while pushing
382 crystals apart (Colombier et al. 2020).

383
384 The percolation associated with channelling, on the other hand, is fostered by buoyancy-driven deformation
385 of the bubbles in the constricted space between crystals and subsequent coalescence at moderate to low Bond
386 numbers. Additional processes can also form similar channel-like interconnected porous pathways at low
387 critical porosity in crystal-rich magmas. For instance, Colombier et al. (2020) showed that brittle-viscous
388 coalescence is favoured at melt viscosities higher than 10^6 Pa.s and relatively high gas overpressure. Because
389 of such alternatives, conditions adverse to channelling do not imply the absence of outgassing. Figure 7
390 shows a literature compilation of vesiculation experiments on crystal-free to crystal-rich magmas and magma
391 analogues (Gardner et al. 2000; Larsen and Gardner 2000; Mangan and Sisson 2000; Gardner 2007;
392 Takeuchi et al. 2009; Shea et al. 2010; Okumura et al. 2012; Martel and Iacono-Marziano 2015;
393 Oppenheimer et al. 2015; Preuss et al. 2016; Colombier et al. 2020). Data is tallied in crystal-poor (<10
394 vol%), moderately crystal-rich (10-40 vol%), and crystal-rich (>50 vol%) categories, as well as whether
395 samples underwent closed-system degassing, or bore connected bubbles networks able to outgas. Figure 7
396 suggests that experimental magma outgassing has only been explored in low-pressure, shallow conditions,
397 which contrasts with our deep-seated determination of channelling during magma ascent. This compilation
398 also confirms that high crystal contents promote outgassing at low percolation thresholds.

399
400 The conditions leading to channelling are based on several key assumptions (Parmigiani et al. 2017;
401 Degruyter et al. 2019), the validity of which must be assessed in the case of conduit flow. One is that

capillary forces dominates buoyancy forces so that the Bond number is 0.1-1. The Bond number depends on gravity, the density difference between melt and gas, the surface tension, and the radius of the gas bubbles. A reasonable estimate for the melt density of the three studied volcanoes is 2300 kg/m³. According to Lyakhovsky et al. (1996), surface tension can be taken as 0.05 N/m. The most variable quantity is bubble size because it depends highly on the pressure and bubble number density. A Bond number between 0.1 and 1 implies a bubble radius between 0.2 and 0.8 mm. At depth, bubbles that size and smaller were likely, and the assumption of dominating capillary forces holds. The possibility that bubbles adhere to crystal surfaces was also left aside. This is a reasonable assumption in conduit conditions because microlites are mostly composed of non-wetting phases like plagioclase in arc magmas. We note, however, that wetting phases like Fe-Mg silicates or Fe-Ti oxides are also present, which suggests that investigating how crystal adherence affects bubble channelling is a worthy pursuit.

Two other assumptions of Parmigiani et al. (2017) and Degruyter et al. (2019) are that the ambient pressure is hydrostatic and that the crystal framework remains static during the onset of channelling. The lower crystallinity limit of 40 vol% stems from the assumption that crystalline matrix resists frictional sliding, which corresponds to a random loose packing. Their synthetic crystal size and shape distributions aim to represent a scaled version of phenocrysts prevailing in storage conditions. Microlite crystallization during ascent yielded in our three studied cases linear to concave-up crystal size distributions with a significant proportion of highly elongated microlites. Random loose packing and frictional effects in such distributions occur at volume fraction lower than 40 vol%, possibility as low as 8 vol% (Saar et al. 2001; Guo et al. 2013). The bubble coalescence leading to channelling, however, is controlled by how much interstitial space occurs between crystals, which is inversely proportional to the cubic root of crystal volume fraction. It is thus unlikely that the onset of channelling occurs at crystallinities much lower than that proposed by Parmigiani et al. (2017), depending on crystal shapes and sizes. Similarly, the transition from channelling to bubble trapping could also occur at a different (yet probably close to) volume fraction than the 70 vol% determined numerically.

Bubble trapping assumes a static solid matrix. It is no longer static when either local overpressure occurs, or a far-field stress is applied to the medium, which are two likely processes in ascending magmas. Different regimes of matrix deformation by local gas overpressure exist. Analogue experiments (Holtzman et al. 2012; Oppenheimer et al. 2015) show that a regime of slow solid matrix rearrangement occurs at the random loose packing, when gas bubbles begin to deform. This change is insensitive to liquid viscosity in the range 10-1000 Pa s. Another regime of rapid deformation of the immersed granular medium by gas-induced overpressure has been described as “venting” (Varas et al. 2015) or “capillary fracturing” (Holtzman et al. 2012). It occurs near the random close packing (Oppenheimer et al. 2015), and it is characterized by gas migration in a fracture-like manner. These regimes have recently been proposed to play a role in Strombolian dynamics when gas proportions are large and melt viscosities low (<5 Pa s; Barth et al. 2019; Oppenheimer et al. 2020). It is unclear yet probable that outgassing by channelling leads to limited grain displacement in melts of much higher viscosity. The two values of crystal content framing channelling are thus reasonable estimates that will be refined when reproduced experimentally with the assumption of static solid matrix relaxed. As the same reasoning holds for the critical porosity (Eq. (3)), the shaded regions in Fig. 3b, 4b, and 5b indicate the two extreme values ever reached by numerical simulations of channelling (Degruyter et al. 2019). Considering only these extreme values does not change the conclusion that conditions for channelling were met in the three studied cases.

Once channelling is established, gas evacuation can proceed to the point of fracture healing if large-scale stresses are applied (Caricchi et al. 2011; Laumonier et al. 2011; Heap et al. 2015). Such gas expulsion is thus expected in a crystal-rich magma sheared along a volcanic conduit (e.g., Thomas et al. 2019). Provided that Eq. (3) is accurate, this is the most probable explanation of the porosities close to but lower than the critical porosity at Soufrière Hills and Merapi (Figs. 3b and 4b). We reiterate that channelling in deforming suspension remains to be quantified to add weight to our inferences.

At Tungurahua, conditions for channelling were met between 110 MPa and a gas-rich region located at 1-10 MPa. Outgassing, however, could have happened through other mechanisms such as viscous percolation in the main part of the conduit owing to the low melt viscosity ($10^{3.7}$ Pa s at the reservoir level, Chevrel et al.

2015), or brittle-viscous coalescence shallowly, where much higher melt viscosities are expected. Figure 6 shows closed-system degassing paths for three starting points: at 200 MPa (reservoir level, Andújar et al. 2017), at 108 MPa, and at 34 MPa. These intermediate pressures correspond to the samples closest to the points where total crystal contents reach and exceed the lower bound for channelling, respectively. The decompression starting at 108 MPa overestimates shallower porosities, whereas that starting at 34 MPa closely follows the trend of the most porous samples. That the measured shallow porosities can be explained by decompressing the residual porosity around 40 MPa suggests that gas loss by channelling was enabled by microlite crystallization between 100 and 40 MPa (1.6-2.9 km depth). The gas-rich region at 1-10 MPa (40-400 m depth) could be due to the ascent of material immediately below that was fast enough to minimize outgassing, or to gas accumulation. Accumulation implies that outgassing was inefficient close to the surface. This could be caused by a combination of hindered channelling at high crystal content, hindered brittle-viscous coalescence because of insufficient gas overpressure, and impermeable plug and conduit walls. Gas accumulation is supported by the near-surface observation that measured porosities decrease to very low values (<1 vol%), revealing the presence of a low permeability, rigid plug (Gaunt et al. 2020). Since the amount of crystals is very high in this plug, cracks or bubble connections (e.g., Heap et al. 2014) were not efficient enough to evacuate the residual gas, thereby allowing the gas-rich region to form.

Our results suggest that outgassing by channelling could have occurred down to (and probably deeper than) 2.7 km at Soufrière Hills, shallower than 3 km depth at Merapi, and between 1.5 and 3 km at Tungurahua. One implication of such widespread channelling depths is that, besides rapidly changing magma rheology to possibly trigger more explosive eruptions (Arzilli et al. 2019), microlite could also spur gas escape. Another implication is that outgassing by channelling could occur regardless of near-surface conditions such as the presence of a dome, a plug, or a permeable wallrock. Finally, the permeabilities reached by channelling are high for magmas in general, and several orders of magnitude higher than commonly admitted permeabilities at low porosity (e.g., Farquharson et al. 2015; Kushnir et al. 2016). Such high permeabilities would affect transitions in eruptive style (Cassidy et al. 2018), and the corresponding permeability relationships given by Parmigiani et al. (2017) and Degruyter et al. (2019) can readily be tested in conduit flow models.

484 Conclusions

485 We revisited three reconstructions of porosity and crystal distributions prior to Vulcanian events that
486 occurred respectively on February 10, 2010 at Soufrière Hills volcano, Montserrat (Burgisser et al. 2019), on
487 October 26, 2010 at Merapi volcano, Indonesia (Drignon et al. 2016) and on July 14, 2013 at Tungurahua
488 volcano, Ecuador (Gaunt et al. 2020). They all feature porosities too low to be explained by common shallow
489 processes such as connected bubble pathways or cracks. Our data show that these magmas have high enough
490 crystal contents for outgassing to occur by a process named channelling. When the amount of crystals in a
491 three-phase magma lies within the 40-70 vol% range, the interstitial space is narrow enough to force bubbles
492 to form channels in which the gas buoyantly moves upwards but not so constricted that channel connections
493 are ruptured (Parmigiani et al. 2017). Channelling creates gas permeability of the order of 10^{-10} m^2 , which is
494 two to five orders of magnitude higher than the other permeability processes at the equivalent bubble
495 content.

496
497 At Soufrière Hills, microlite crystallization during ascent yielded conditions propitious to channelling from
498 near surface down to a depth greater than 2.7 km. If confidence is given in the critical porosity
499 quantification, outgassing deflated the magma slightly below the critical porosity, which suggests that large-
500 scale stresses such as those related to conduit wall shearing were also at play in controlling the evolution of
501 porosity during ascent.

502
503 At Merapi, the phenocrysts in the reservoirs could have fostered outgassing at depth without magma ascent.
504 Closed-system degassing in response to ascent, however, only yielded enough gas for channelling at mid-
505 conduit depths. The part of the conduit where both gas contents were large enough and calculated porosities
506 fall within the percolation range is located around 3 km depth. In this region and in the shallower parts of the
507 conduit, outgassing by channelling could have taken place during (and possibly before) microlite
508 crystallization but before the 70% crystal content was reached. This suggests that the strong gas depletion in
509 the magma column could have resulted from weak exsolution at the beginning of the ascent followed by
510 channelling at depths shallower than 3 km. Of the two end-member magmas involved in the eruption that we

511 considered in our analysis, this degassing behaviour is best explained if the conduit was filled by the colder
512 resident magma.

513

514 At Tungurahua, outgassing by channelling was enabled by microlite crystallization at 1.6–2.9 km depth.
515 Near the surface, a low permeability plug trapped a gas-rich region between 40 and 400 m depth. Inefficient
516 outgassing between 1.6 km and the gas accumulation region was probably caused by a combination of
517 hindered channelling at high crystal content, hindered brittle-viscous coalescence because of insufficient gas
518 overpressure, and impermeable plug and conduit walls. This suggests that the triggering of the paroxysmal
519 explosion was controlled by near-surface gas trapping.

520

521 We presented lines of evidence for deep gas loss prior to sudden explosive emissions of three basaltic
522 andesitic to andesitic magmas. They suggest that channelling is a viable mechanism for gas to be removed
523 from the magma column at various levels during ascent, yielding mostly gas-depleted magma columns prior
524 to eruption. Such outgassing by channelling could thus influence eruptive style. Depending on the
525 phenocryst content, microlite growth during ascent can either foster, or impede gas escape by channelling.
526 Considering the pervasive occurrence of microlites and ensuing high crystal contents in volcanic conduits,
527 the high likelihood of outgassing by channelling implies that other outgassing mechanisms might not be as
528 dominant as previously envisioned.

529

530 Tables

531 **Table 1:** Pre-explosive volatile proportions, vesicle-free crystal proportions, and pressures. Values in
 532 parenthesis are relative uncertainties in % except for the pressure, where they indicate lower and upper
 533 bounds of uncertainties in MPa. Uncertainties on phenocryst content are sometimes very low (<1%), which
 534 reflects that few SEM images were imaged to characterize phenocrysts. Uncertainties on the total crystal
 535 fraction include those due to microlite, which are obtained from a larger number of images. They thus reflect
 536 more accurately the natural spatial variability.

	Pressure (MPa)	Porosity ^a (vol%)	Gas ^b (ppmw)	Phenocrysts (vol%)	Crystals (vol%)
Soufrière Hills					
AMO 210 A	40 (-9,+10)	1.9 (1)	0.6	34 (2)	54 (10)
AMO 210 C	51 (-25,+31)	2.3 (2)	0.92	34 (12)	55 (22)
AMO 210 D	14 (-3,+3)	1.9 (1)	0.2	31 (6)	39 (11)
AMO 210 E	54 (-19,+22)	3.5 (0)	1.54	48 (0.3)	69 (9)
AMO 210 F	33 (-8,+9)	1.6 (1)	0.41	50 (2)	66 (13)
AMO 210 G	42 (-13,+15)	1.2 (3)	0.39	40 (12)	63 (11)
AMO 210 H	12 (-5,+6)	8.5 (1)	0.89	39 (8)	65 (12)
AMO 210 L	46 (-13,+14)	1.4 (2)	0.52	59 (3)	77 (6)
AMO 210 Q	41 (-7,+7)	3.4 (1)	1.14	32 (1.3)	40 (14)
WP1.095A	24 (-5,+5)	4.3 (0)	0.86	24 (0.4)	52 (3)
WP1.095B	16 (-6,+7)	7.2 (0)	0.95	32 (0.2)	59 (9)
WP1.108A	31 (-5,+5)	2.8 (0)	0.72	32 (0.3)	52 (10)
WP1.108B	56 (-5,+5)	7.8 (1)	3.68	36 (1.1)	48 (3)
WP2.200A	36 (-8,+9)	2.8 (0)	0.81	42 (0.3)	63 (19)
WP2.200B	31 (-2,+2)	3 (0)	0.76	31 (0.7)	42 (3)
WP2.201A	35 (-8,+9)	4.6 (0)	1.33	45 (0.2)	59 (10)
WP2.201B	69 (-14,+15)	2.4 (0)	1.36	31 (0.2)	51 (8)
WP2.329A	7 (-1,+1)	23.6 (0)	1.62	11 (1.2)	27 (15)
WP2.329B	41 (-4,+4)	2.7 (1)	0.89	51 (2)	65 (3)
Merapi Oct. 26					
AME 10 A1	75 (-22,+25)	2.3 (4)	1.27	33 (7)	70 (11)
AME 10 A2	100 (-34,+38)	1.5 (4)	1.12	44 (8)	79 (8)
AME 10 A3	150 (-123,+174)	1.2 (3)	1.32	40 (3)	82 (11)
AME 10 A4	100 (-22,+24)	1.6 (5)	1.16	45 (8)	80 (8)
AME 10 A5	73 (-25,+29)	2.4 (3)	1.32	47 (5)	76 (6)
AME 10 A6	58 (-12,+13)	3.1 (2)	1.32	46 (3)	72 (5)
AME 10 B1	158 (-48,+53)	0.9 (4)	1.03	41 (9)	80 (10)
AME 10 B2	63 (-11,+12)	2.3 (3)	1.05	50 (5)	72 (8)
AME 10 B4	268 (-106,+118)	0.6 (4)	1.14	36 (6)	87 (6)

AME 10 B5	260 (-153,+184)	0.4 (5)	0.73	44 (13)	87 (12)
AME 10 C1	159 (-57,+64)	1 (1)	1.13	38 (1)	80 (14)
AME 10 C2	141 (-31,+33)	0.7 (2)	0.69	38 (5)	81 (6)
AME 10 C3	55 (-17,+20)	2.8 (2)	1.16	46 (3)	70 (10)
AME 10 C4	77 (-24,+27)	1.4 (3)	0.78	44 (7)	77 (7)
AME 10 D1	107 (-25,+27)	1.1 (3)	0.88	44 (4)	79 (9)
AME 10 D2	85 (-37,+43)	1.4 (5)	0.88	42 (11)	79 (11)
AME 10 D3	64 (-37,+47)	2 (8)	0.96	34 (20)	69 (27)
AME 10 D4	63 (-14,+15)	2.8 (5)	1.34	42 (12)	71 (10)
AME 10 E1	75 (-23,+26)	1.2 (2)	0.67	42 (6)	74 (11)
AME 10 E2	202 (-67,+74)	0.9 (4)	1.26	57 (4)	88 (4)
AME 10 E3	59 (-11,+12)	2 (3)	0.86	43 (9)	74 (8)
AME 10 E4	77 (-15,+16)	1.9 (5)	1.1	45 (10)	72 (8)
AME 10 E5	227 (-78,+87)	0.5 (4)	0.85	45 (12)	86 (11)
AME 10 F1	12 (-5,+5)	2.3 (20)	0.2	21 (23)	54 (17)
AME 10 G1	42 (-16,+18)	2.3 (2)	0.76	52 (5)	75 (7)
AME 10 G2	75 (-54,+75)	1.1 (2)	0.78	51 (6)	82 (11)
AME 10 G3	44 (-21,+25)	0.3 (15)	0.48	41 (9)	70 (12)
AME 10 GA	94 (-25,+28)	2.8 (1)	0.89	48 (10)	81 (9)
AME 10 GB	196 (-80,+92)	1.6 (1)	0.86	47 (8)	88 (8)
Tungurahua					
1	0.8 (-0.2,+0.4)	36.8 (3)	0.33	15 (21)	55 (12)
2	0.5 (-0.1,+0.1)	0 (8)	0	46 (10)	73 (9)
4	0.9 (-0.2,+0.3)	0.9 (11)	0.01	42 (14)	67 (13)
6	0.5 (-0.1,+0.2)	0.3 (30)	0.001	39 (36)	70 (32)
7.5	0.6 (-0.2,+0.3)	0.8 (9)	0.004	30 (12)	62 (12)
11	3 (-1,+1)	0 (1)	0	25 (1.4)	44 (5)
12	109 (-63,+91)	0.3 (1)	0.24	34 (0.8)	38 (5)
13	5 (-2,+9)	44.2 (1)	2.53	32 (2)	48 (14)
14	31 (-17,+68)	8.5 (4)	2.07	38 (4)	45 (6)
16	1.6 (-0.4,+0.5)	15.7 (4)	0.21	31 (3)	55 (5)
17	24 (-8,+16)	1.7 (2)	0.29	36 (6)	43 (6)
18	34 (-15,+41)	6.1 (1)	1.57	34 (2)	43 (6)
20	66 (-14,+20)	1.5 (2)	0.69	31 (3)	36 (4)
21	5 (-3,+8)	3.2 (2)	0.12	36 (3)	51 (10)
22	2 (-1,+5)	0 (4)	0	26 (3)	48 (15)
26	51 (-36,+149)	8 (3)	3.14	32 (2)	39 (9)
27	15 (-5,+10)	8.5 (3)	1.01	34 (5)	42 (8)
28	5 (-2,+9)	36.5 (3)	1.86	29 (6)	45 (12)
29	30 (-17,+82)	7.7 (1)	1.77	30 (0.8)	36 (6)
30	4 (-1,+2)	51.7 (1)	3.14	23 (4)	41 (8)
34	1.2 (-0.3,+0.4)	45 (3)	0.69	25 (2)	53 (6)
35	4 (-1,+2)	11.5 (3)	0.4	28 (3)	46 (5)
36	5 (-1,+2)	8.1 (1)	0.33	30 (2)	47 (7)
38	31 (-14,+38)	10.6 (1)	2.62	39 (2)	45 (4)

39	31 (-17,+67)	1.5 (2)	0.33	34 (5)	40 (7)
40	10 (-0.1,+0.1)	25.7 (2)	2.43	38 (2)	49 (2)
41	174 (-79,+26)	0.3 (2)	0.4	26 (3)	32 (4)

Figure captions

Fig.1: Porosity as a function of pre-explosive pressure for the February 10, 2010 Vulcanian explosion at Soufrière Hills. Data are from Burgisser et al. (2019) and the «equilibrium» curve corresponds to closed-system degassing. Other curves are from the open-system degassing conduit flow models of Clarke et al. (2007), Melnik and Sparks (2002), the MAMMA code (La Spina et al. 2017; Degruyter et al. 2012), and Collombet (2009)

Fig. 2: Schematic depiction of the textural elements considered. Clasts represent the magma in “Quench state”, which is composed of melt (blue), crystals (red), and vesicles (white). Vesicles are pre-explosive (Type 1), syn-explosive and isolated (Type 2), syn-explosive gaps between broken crystals (Type 3), and syn-explosive and coalesced (Type 4). Prior to ejection, the magma is in a “Pre-explosive state”. It is composed of Type 1 vesicles that occupy a smaller volume than in the quench state because they are stored a high pressure in the conduit. The “ascent from the reservoir” allows the magma to reach the pre-explosive state in the conduit. If the critical porosity (ϕ_{cr}) is reached during ascent, channelling could cause vesicles to form channels that enable outgassing. The far-right panel is a visual depiction of the definitions of porosity, ϕ , vesicle-free crystal volume fraction, ϕ_{xvf} , and bulk crystal volume fraction, ϕ_x

Fig. S2: Pre-explosive porosities and associated model uncertainties of this study (Table 1) and those of Burgisser et al. (2019) (A), Gaunt et al. (2020) (B), and Drignon et al. (2016) (C)

Fig.3: a) Amount of phenocrysts (red) and phenocrysts plus microliths (blue) relative to pressure for Soufrière Hills according to Burgisser et al. (2019). The shaded region covers the validity domain for the development of channelling according to Parmigiani et al. (2017). b) Porosity thresholds calculated from Degruyter et al. (2019) and associated uncertainties (blue) for Soufrière Hills. Samples with crystal contents outside the validity range were discarded. Red dots are observed porosities. The

associated uncertainties are smaller than the symbols. The green line gives theoretical porosities for a closed system. The shaded region covers the extreme porosity threshold values

Fig.4: a) Contents of phenocrysts (red) and phenocrysts plus microlite (blue) as a function of pressure for Merapi according to Drignon et al. (2016). The shaded region covers the validity domain for the development of channelling according to Parmigiani et al. (2017). b) Porosity thresholds calculated from Degruyter et al. (2019) and associated uncertainties (blue) for Merapi. Red dots are observed porosities and the associated uncertainties are smaller than the symbols. Green and black lines give theoretical porosities for a closed system with 5 wt% H₂O and 400 ppm CO₂ in the melt (resident magma) and with 3 wt% H₂O and 500 ppm CO₂ in the melt (recharge magma) respectively. The shaded region covers the extreme porosity threshold values

Fig.5: a) Contents of phenocrysts (red) and phenocrysts plus microlites (blue) as a function of pressure for Tungurahua according to Gaunt et al. (2020). The shaded region covers the validity domain for the development of channelling according to Parmigiani et al. (2017). b) Porosity thresholds calculated from Degruyter et al. (2019) and associated uncertainties (blue) for Tungurahua. Samples with crystal contents outside the validity range were discarded. Red dots are observed porosities and the associated uncertainties are smaller than the symbols. Green line gives the theoretical porosities for a closed system. The shaded region covers the extreme porosity threshold values

Fig.6: Comparison at Tungurahua between pre-eruptive porosities (red circles), and theoretical porosities for closed systems starting decompressing at the reservoir level of 200 MPa (green line), at 108 MPa (cyan line), and at 34 MPa (magenta line)

Fig.7: Porosity profile as a function of depth for vesiculation experiments on crystal-free to crystal-rich magmas and magma analogues. The coloured areas respectively overlap samples with connected bubbles networks capable of outgassing. All other samples featured closed system degassing. Data are from Gardner et al. (2000), Larsen and Gardner (2000), Mangan and Sisson (2000), Gardner (2007); Takeuchi et al. (2009), Shea et al. (2010), Okumura et al. (2012), Martel and Iacono-Marziano (2015), Oppenheimer et al. (2015), Preuss et al. (2016), and Colombier et al. (2020)

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Table

	Pressure (MPa)	Porosity ^a (vol%)	Gas ^b (ppmw)	Phenocrysts (vol%)	Crystals (vol%)
Soufrière Hills					
AMO 210 A	40 (-9,+10)	1.9 (1)	0.6	34 (2)	54 (10)
AMO 210 C	51 (-25,+31)	2.3 (2)	0.92	34 (12)	55 (22)
AMO 210 D	14 (-3,+3)	1.9 (1)	0.2	31 (6)	39 (11)
AMO 210 E	54 (-19,+22)	3.5 (0)	1.54	48 (0.3)	69 (9)
AMO 210 F	33 (-8,+9)	1.6 (1)	0.41	50 (2)	66 (13)
AMO 210 G	42 (-13,+15)	1.2 (3)	0.39	40 (12)	63 (11)
AMO 210 H	12 (-5,+6)	8.5 (1)	0.89	39 (8)	65 (12)
AMO 210 L	46 (-13,+14)	1.4 (2)	0.52	59 (3)	77 (6)
AMO 210 Q	41 (-7,+7)	3.4 (1)	1.14	32 (1.3)	40 (14)
WP1.095A	24 (-5,+5)	4.3 (0)	0.86	24 (0.4)	52 (3)
WP1.095B	16 (-6,+7)	7.2 (0)	0.95	32 (0.2)	59 (9)
WP1.108A	31 (-5,+5)	2.8 (0)	0.72	32 (0.3)	52 (10)
WP1.108B	56 (-5,+5)	7.8 (1)	3.68	36 (1.1)	48 (3)
WP2.200A	36 (-8,+9)	2.8 (0)	0.81	42 (0.3)	63 (19)
WP2.200B	31 (-2,+2)	3 (0)	0.76	31 (0.7)	42 (3)
WP2.201A	35 (-8,+9)	4.6 (0)	1.33	45 (0.2)	59 (10)
WP2.201B	69 (-14,+15)	2.4 (0)	1.36	31 (0.2)	51 (8)
WP2.329A	7 (-1,+1)	23.6 (0)	1.62	11 (1.2)	27 (15)
WP2.329B	41 (-4,+4)	2.7 (1)	0.89	51 (2)	65 (3)
Merapi Oct. 26					
AME 10 A1	75 (-22,+25)	2.3 (4)	1.27	33 (7)	70 (11)
AME 10 A2	100 (-34,+38)	1.5 (4)	1.12	44 (8)	79 (8)
AME 10 A3	150 (-123,+174)	1.2 (3)	1.32	40 (3)	82 (11)
AME 10 A4	100 (-22,+24)	1.6 (5)	1.16	45 (8)	80 (8)
AME 10 A5	73 (-25,+29)	2.4 (3)	1.32	47 (5)	76 (6)
AME 10 A6	58 (-12,+13)	3.1 (2)	1.32	46 (3)	72 (5)
AME 10 B1	158 (-48,+53)	0.9 (4)	1.03	41 (9)	80 (10)
AME 10 B2	63 (-11,+12)	2.3 (3)	1.05	50 (5)	72 (8)
AME 10 B4	268 (-106,+118)	0.6 (4)	1.14	36 (6)	87 (6)
AME 10 B5	260 (-153,+184)	0.4 (5)	0.73	44 (13)	87 (12)
AME 10 C1	159 (-57,+64)	1 (1)	1.13	38 (1)	80 (14)
AME 10 C2	141 (-31,+33)	0.7 (2)	0.69	38 (5)	81 (6)
AME 10 C3	55 (-17,+20)	2.8 (2)	1.16	46 (3)	70 (10)
AME 10 C4	77 (-24,+27)	1.4 (3)	0.78	44 (7)	77 (7)
AME 10 D1	107 (-25,+27)	1.1 (3)	0.88	44 (4)	79 (9)
AME 10 D2	85 (-37,+43)	1.4 (5)	0.88	42 (11)	79 (11)
AME 10 D3	64 (-37,+47)	2 (8)	0.96	34 (20)	69 (27)
AME 10 D4	63 (-14,+15)	2.8 (5)	1.34	42 (12)	71 (10)
AME 10 E1	75 (-23,+26)	1.2 (2)	0.67	42 (6)	74 (11)
AME 10 E2	202 (-67,+74)	0.9 (4)	1.26	57 (4)	88 (4)
AME 10 E3	59 (-11,+12)	2 (3)	0.86	43 (9)	74 (8)
AME 10 E4	77 (-15,+16)	1.9 (5)	1.1	45 (10)	72 (8)
AME 10 E5	227 (-78,+87)	0.5 (4)	0.85	45 (12)	86 (11)

AME 10 F1	12 (-5,+5)	2.3 (20)	0.2	21 (23)	54 (17)
AME 10 G1	42 (-16,+18)	2.3 (2)	0.76	52 (5)	75 (7)
AME 10 G2	75 (-54,+75)	1.1 (2)	0.78	51 (6)	82 (11)
AME 10 G3	44 (-21,+25)	0.3 (15)	0.48	41 (9)	70 (12)
AME 10 GA	94 (-25,+28)	2.8 (1)	0.89	48 (10)	81 (9)
AME 10 GB	196 (-80,+92)	1.6 (1)	0.86	47 (8)	88 (8)
Tungurahua					
1	0.8 (-0.2,+0.4)	36.8 (3)	0.33	15 (21)	55 (12)
2	0.5 (-0.1,+0.1)	0 (8)	0	46 (10)	73 (9)
4	0.9 (-0.2,+0.3)	0.9 (11)	0.01	42 (14)	67 (13)
6	0.5 (-0.1,+0.2)	0.3 (30)	0.001	39 (36)	70 (32)
7.5	0.6 (-0.2,+0.3)	0.8 (9)	0.004	30 (12)	62 (12)
11	3 (-1,+1)	0 (1)	0	25 (1.4)	44 (5)
12	109 (-63,+91)	0.3 (1)	0.24	34 (0.8)	38 (5)
13	5 (-2,+9)	44.2 (1)	2.53	32 (2)	48 (14)
14	31 (-17,+68)	8.5 (4)	2.07	38 (4)	45 (6)
16	1.6 (-0.4,+0.5)	15.7 (4)	0.21	31 (3)	55 (5)
17	24 (-8,+16)	1.7 (2)	0.29	36 (6)	43 (6)
18	34 (-15,+41)	6.1 (1)	1.57	34 (2)	43 (6)
20	66 (-14,+20)	1.5 (2)	0.69	31 (3)	36 (4)
21	5 (-3,+8)	3.2 (2)	0.12	36 (3)	51 (10)
22	2 (-1,+5)	0 (4)	0	26 (3)	48 (15)
26	51 (-36,+149)	8 (3)	3.14	32 (2)	39 (9)
27	15 (-5,+10)	8.5 (3)	1.01	34 (5)	42 (8)
28	5 (-2,+9)	36.5 (3)	1.86	29 (6)	45 (12)
29	30 (-17,+82)	7.7 (1)	1.77	30 (0.8)	36 (6)
30	4 (-1,+2)	51.7 (1)	3.14	23 (4)	41 (8)
34	1.2 (-0.3,+0.4)	45 (3)	0.69	25 (2)	53 (6)
35	4 (-1,+2)	11.5 (3)	0.4	28 (3)	46 (5)
36	5 (-1,+2)	8.1 (1)	0.33	30 (2)	47 (7)
38	31 (-14,+38)	10.6 (1)	2.62	39 (2)	45 (4)
39	31 (-17,+67)	1.5 (2)	0.33	34 (5)	40 (7)
40	10 (-0.1,+0.1)	25.7 (2)	2.43	38 (2)	49 (2)
41	174 (-79,+26)	0.3 (2)	0.4	26 (3)	32 (4)

^a Values at 0 are <0.1 vol% and errors at 0 are <1%.

^b Values at 0 are <0.001 ppm per weight.

Figure

Figure

Figure

Figure

Figure

Figure

Figure

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