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1 How to turn off a lava lake? A petrological investigation of the
2 2018 intra-caldera and submarine eruptions of Ambrym
3 volcano.

4
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22
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24 eruption; geo-speedometer

25

26 **ABSTRACT**

27 In December 2018, an unusually large intra- and extra-caldera eruption took place at Ambrym
28 volcano (Vanuatu). The eruption drained the volcano's five active lava lakes and terminated,
29 at least momentarily, the surface activity that had been ongoing for decades to hundreds of
30 years, sustaining the largest recorded persistent degassing on the planet. Here, we investigate
31 the mechanisms and dynamics of this major eruption. We use major elements and volatiles in
32 olivine and clinopyroxene hosted melt inclusions, embayments, crystals and matrix glasses
33 together with clinopyroxene geobarometry as well as olivine and clinopyroxene
34 geothermometry and diffusion modelling in crystals and embayments to reconstruct the
35 chronology and timing of the subsurface processes that accompanied the eruption. We find that
36 the eruption began with the meeting, mingling and limited chemical mixing of mostly two
37 magma bodies occupying similar vertical but different horizontal locations in the crust, one
38 corresponding to the main plumbing system at Ambrym that fed the lava lakes and the other
39 corresponding to an older, previously cut-off and more chemically evolved branch of the
40 plumbing system. Within the primitive magma, two texturally distinct components – one
41 microlite-rich and one microlite-poor – can further be identified. The 2018 eruption hence
42 provides a detailed image of Ambrym's complex plumbing system. Our diffusion timescales
43 and geobarometric estimates coincide closely with geophysical observations. They point to a
44 reconnection of the evolved magmatic branch with the main system occurring less than 10 h
45 prior to the intra-caldera eruption and a period of two days for the subsequent >30 km lateral
46 magma transport along a deeper dike prior to submarine eruption just off the SE coast of the
47 island with the more primitive magma reaching first followed by mingled magma containing
48 both compositions. Magma ascent rates is estimated at 95 ± 24 m/s in the last ~2.5 km of ascent
49 during the intra-caldera eruption and at 80 ± 6 m/s in the last ~4 km of ascent during the
50 submarine eruption. Comparison with other lava-lake-draining eruptions reveal striking

51 similarities both in terms of precursory activity, with lake level rising prior to eruption in all
52 cases, and in term of plumbing system organisation with the presence of peripheral magma
53 pockets, isolated from the main magmatic system but that can be mobilized and erupted when
54 met by dikes propagating laterally from the main system.

55

56 **INTRODUCTION**

57 Lava lakes are the emblem of persistent degassing volcanic activity. They are rare examples of
58 volcanic systems having reached a metastable equilibrium whereby gas and magma motions
59 result in conduit dynamics allowing for efficient gas release whilst maintaining molten magma
60 from the chamber to the surface (e.g., Tazieff 1994; Harris et al. 2005; Witham and Llewelin
61 2006; Harris 2008; Oppenheimer et al. 2009; Burgi et al. 2014; Moussallam et al. 2015a, 2016;
62 Allard et al. 2016b). As a result, volcanoes with lava lakes tend to rank amongst the major
63 emitters of volcanic gases to the atmosphere – at least for the past decade (Carn et al. 2017) –
64 and to maintain surficial activity that can remain nearly unchanged for decades to centuries. In
65 2018 two of the most iconic lava lake volcanoes, Kīlauea and Ambrym, both experienced major
66 eruptions whereby subsurface magma migration, associated with an episode of caldera
67 collapse, caused the rapid drainage of their lava lakes (Neal et al. 2019; Shreve et al. 2019).

68

69 In this study we investigate the mechanism behind the 2018 Ambrym eruption. We present
70 petrological observations of glasses, crystals and melt inclusions of both the initial intra-caldera
71 fire-fountain eruption and subsequent submarine eruption. We then compare these findings to
72 geophysical and surficial observation to derive a detailed understanding of the cause and timing
73 of the eruption as well as the architecture of Ambrym’s plumbing system.

74

75 **THE 2018 ERUPTION OF AMBRYM VOLCANO**

76 A detailed account of the temporal evolution of the eruption in terms of deformation, seismic,
77 thermal and SO₂ gas emissions from ground and satellite observations is given in Shreve et al.,
78 (2019). Here we summarise the main events focussing on the deposits and ground observations.
79 We divide the eruption in two parts, the intra-caldera and the submarine eruption.

80

81 **Intra-caldera eruption**

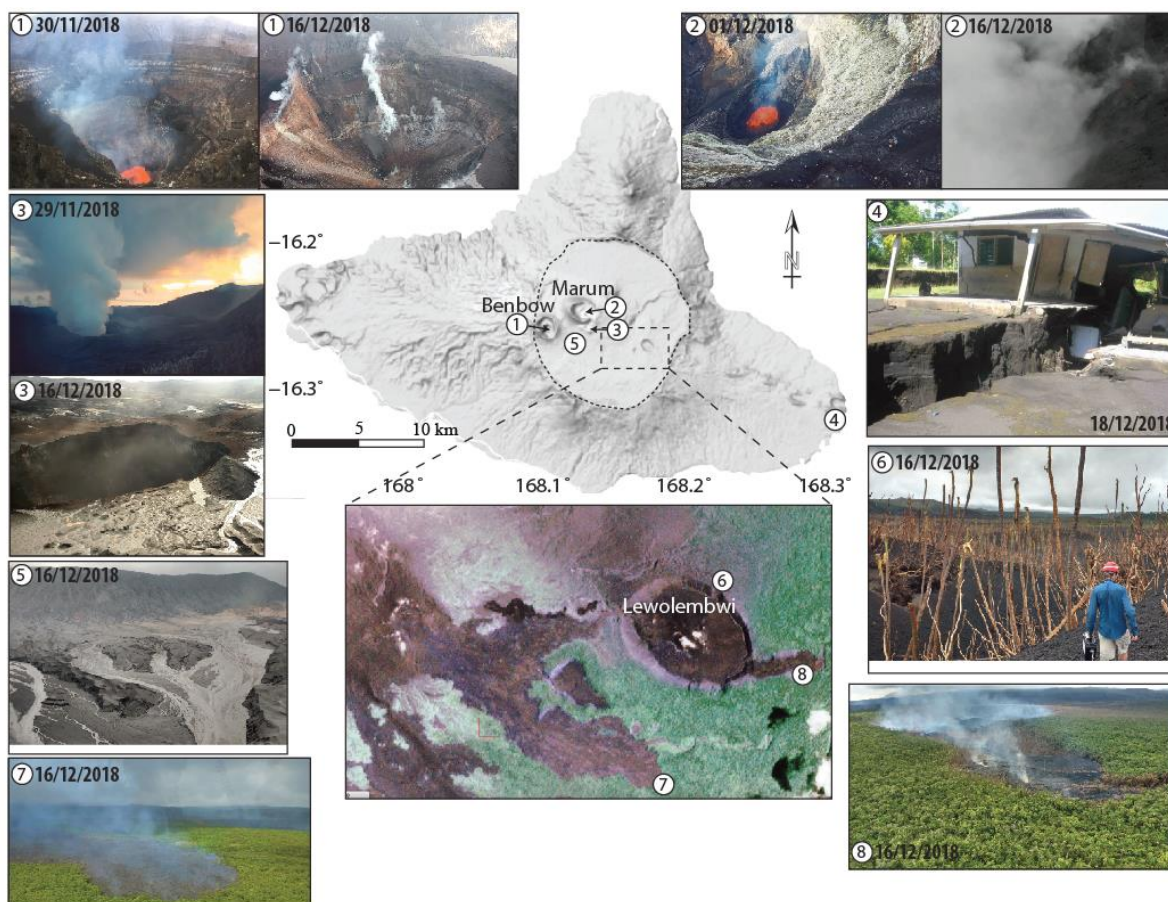
82 On 14 December 2018 (all date and time in UTC), the volcano-seismic crisis began with a few
83 events detected inside the caldera. Lava reached the surface around 23h20 UTC from two
84 fissures according to high time-resolution thermal observations from Advanced Himawari
85 Imager (AHI) aboard HIMAWARI 8 geostationary satellite (Shreve et al. 2019). Two fissures
86 opened, one trending N110° at 590 m a.s.l. and cutting through the Lewolembwi tuff ring (re-
87 activation of a pre-existing fracture) and the other trending N-S at 730 m a.s.l. and located near
88 the eruption site of the 2015 lava flows. The first fissure produced mainly scoria deposits
89 covering a 1 km radius (with minor pahoehoe lava flows close to the eruption site) while the
90 second fissure produced a blocky, 10⁶ m³ (Shreve et al. 2019) lava flow (Fig. 1, panel 6, 7 and
91 8). Both fissures likely resulted in fire fountain style of eruption (visually confirmed only at
92 the second fissure). As lava extrusion proceeded on the SE flank of the volcano during the 15
93 of December, lava lake activity at the summit ceased according to satellite thermal observations
94 (Shreve et al. 2019). Directly prior to the eruption, five lava lakes were being observed by the
95 authors in the summit region, two at Benbow and three at Marum (with lake level at Marum's
96 main lava lake noticeably increasing in the weeks prior to the eruption; Fig. S1) (Fig. 1, panel
97 1 and 2). The crater walls around all five lava lakes had partially collapsed inward by 16
98 December, one lava lake at Marum crater had been replaced by a water lake at that time. A
99 previously vigorously degassing vent located just south of Marum (called Maben Mbwelesu or

100 Niri Taten, source of the 1988–1989 basaltic lava flow) also ceased degassing (Fig. 1, panel 3).
101 According to HIMAWARI observations, syn-eruptive SO₂ degassing mainly took place on 15
102 and 16 December, ceasing early in the morning of 16 December (UT) (Shreve et al. 2019). On
103 15 December at 17:45 and 19:45 and on 16 December at 1:00 and 5:00 ash clouds were
104 observed from the Vanuatu Meteorology and Geohazards Department’s webcam. On 16
105 December at 3:30, lahars were observed on the flank of both Benbow and Marum (Fig. 1, panel
106 5). Yet no rainfall had been reported since the onset of the eruption, suggesting that the water
107 originated from within the edifices, possibly as a result of compaction during the period of
108 subsidence highlighted by InSAR observations (see next section). By the end of December 16,
109 surficial activity inside the caldera had ceased. Videos from a helicopter overflight in the
110 caldera on 16 December are provided in the supplementary.

111

112 **Dike propagation and submarine eruption**

113 At 20h21 on December 15 the seismic activity increased sharply marking the beginning of
114 magma propagation into the SE rift zone (Shreve et al., 2019). By 17 December at 12:00,
115 inversion of InSAR data revealed ~3 m of opening along a >30 km long dike, of 419 to
116 532×10^6 m³ total volume, dipping ~70° to the south, and extending from within the caldera
117 to beyond the eastern coast (Shreve et al., 2019). Around 16:00 on December 17 magma
118 migration stopped, likely marking the onset of a shallow submarine eruption just off the SE
119 coast of Ambrym, near the villages of Pamal and Ulei, confirmed in the following days as
120 basaltic pumice drifted to shore. This large magma migration episode in the subsurface was
121 accompanied by inflation in the SE part of the island (Fig 1, panel 4) but also by subsidence
122 (>2 m) of the caldera floor (Shreve et al., 2019).



123

124 **Fig. 1** Central panel: shaded hillside view of Ambrym Island (source CNES/Airbus), dotted
 125 contour outlines the ~12 km diameter caldera. Lower panel: True-colour Planet image taken
 126 on 31-02-2019 showing the extent of lava flow and scoria deposits (emplaced on 14-16
 127 December 2018). Numbers on the central and lower panels show the location of the
 128 photographs with corresponding numbers. Panel 1 shows one of two lava lakes at Benbow
 129 crater before and after/during the eruption. Panel 2 shows one of three lava lakes at Marum
 130 crater before and after/during the eruption. Crater walls around all five original lava lakes
 131 partially collapsed, one lava lake was replaced by a water lake by December 16. Panel 3 shows
 132 the Maben Mbwelesu or Niri Taten vent before and after/during the eruption. Panel 4 shows
 133 the surface deformation and resulting fracturing affecting the coastal village of Pamal, located
 134 close to the site of inferred submarine eruption starting on 17 December. Panel 5, view of
 135 Marum with lahar coming down the flanks of the edifice from all directions whilst no rainfall
 136 had occurred, suggesting that groundwater, expelled due to compaction was generating the
 137 mud flows. Panel 6, view from the flank of Lewolembwi crater with palm trees blasted and
 138 partially buried by scoria. Panel 7, view of the lava flow with active front marked by burning
 139 vegetation. Panel 8, view of the fissure running through the Lewolembwi crater
 140

141 **METHODOLOGY**

142 **Samples**

143 Tephra samples from the intra-caldera eruption were collected on 16 December 2018, North of
144 the Lewolembwi crater (at 16°16'3.65"S; 168°10'13.61"E, sample name AMB2018_FF_S1 and
145 at 16°15'42.42"S; 168°10'17.41"E, sample name AMB2018_FF_S2). Tephra from both
146 samples were lapilli-size but sample AMB2018_FF_S2, being located further from the fissure
147 was composed of smaller (typically < 1cm diameter) lapilli. An ash sample was collected south
148 of the Marum and Benbow craters (at 16°16'4.29"S; 168° 7'3.24"E, sample name
149 AMB2018_FF_S3). Tephra samples from the submarine eruption were collected on 18
150 December 2018 (sample name AMB2018_SUB_S1) and 04 February 2019 (sample name
151 AMB2018_SUB_PAMAL_1) from the beach near the coastal village of Pamal. In both cases
152 the pumices were floating and deposited onshore. The exact timing of the deposition of each
153 wave of scoria is unknown. Both samples consist of lapilli size (typically < 1cm diameter),
154 highly vesicular fragments. Details on sample processing are provided in the **supplementary**.

155

156 **Analyses of major, trace and volatile elements**

157 Major element compositions of bulk tephra were analysed on a HORIBA-Jobin-Yvon
158 ULTIMA C ICP-AES at Laboratoire Magmas et Volcans in Clermont-Ferrand using the
159 procedure described in Moussallam et al. (2019). Trace element compositions of melt
160 inclusions and glasses were analysed using a laser ablation system associated with an
161 inductively coupled plasma mass spectrometer of the Laboratoire Magmas et Volcans,
162 Clermont-Ferrand (193 nm Excimer Resonetics M-50E laser with an Agilent 7500 ICP-MS).
163 Volatile (H₂O, CO₂, Cl, F, S) content in melt inclusions, embayments, and matrix glasses were
164 determined using a Cameca IMS 1280 ion microprobe at CRPG-CNRS-Nancy. Analytical
165 conditions were similar to other volatile studies (e.g., Hauri et al. 2002; Bouvier et al. 2008;

166 Shimizu et al. 2009; Rose-Koga et al. 2014, 2020.; Moussallam et al. 2015b, 2019). Details of
167 all three methods are provided in the **supplementary**.

168

169 **Volatile and olivine diffusion modelling**

170 Concentration profiles recorded in the embayments were fitted by a diffusion model similar to
171 the one of Ferguson et al. (2016), building on the model presented in Moussallam et al. (2019).

172 Chemical gradients (Fe–Mg, Mn and Ca) in olivine crystals were modelled in one dimension
173 using the DIPRA software (Girona and Costa 2013). Details of both methods are provided in
174 the **supplementary**.

175

176 **Assessment of post-entrapment crystallisation.**

177 The amount of post-entrapment crystallisation (PEC) for olivine-hosted inclusions was
178 estimated using the Petrolog3 software (Danyushevsky and Plechov 2011). Calculations were
179 performed using the olivine-melt model of Danyushevsky, (2001), the density model of Lange
180 and Carmichael, (1990), the model for melt oxidation of Kress and Carmichael, (1988) and the
181 model of Toplis, (2005) for the compositional dependence of the olivine-liquid Fe-Mg
182 exchange coefficient (Kd). Calculations were performed assuming a system buffered at the
183 nickel nickel-oxide (NNO) equilibrium. Note that calculations in Petrolog3 are performed
184 under anhydrous conditions at 1 atm. As the inclusions showed no sign of iron loss (**Fig. 3D**),
185 measured FeO_t concentration were taken as final concentration. The resulting PEC estimates
186 range from -10 to 5% with an average of -3% and standard deviation of $\pm 3\%$. Performing the
187 same calculations at the quartz-fayalite-magnetite (QFM) buffer yield similar results with a
188 range from -11 to 3%, an average of -4% and standard deviation of $\pm 3\%$. Given that most
189 (77%) olivine-hosted inclusions are modelled to have undergone no or negative amounts of
190 PEC we consider that it can safely be assumed that most inclusions have retained their

191 entrapment composition and not been affected by any significant amounts of PEC. In addition,
192 the fact that olivine-hosted melt inclusions, pyroxene-hosted melt inclusions and matrix glasses
193 all record the same range in compositions argues against any significant post-entrapment
194 crystallisation effect on the melt inclusion compositions.

195

196 **RESULTS**

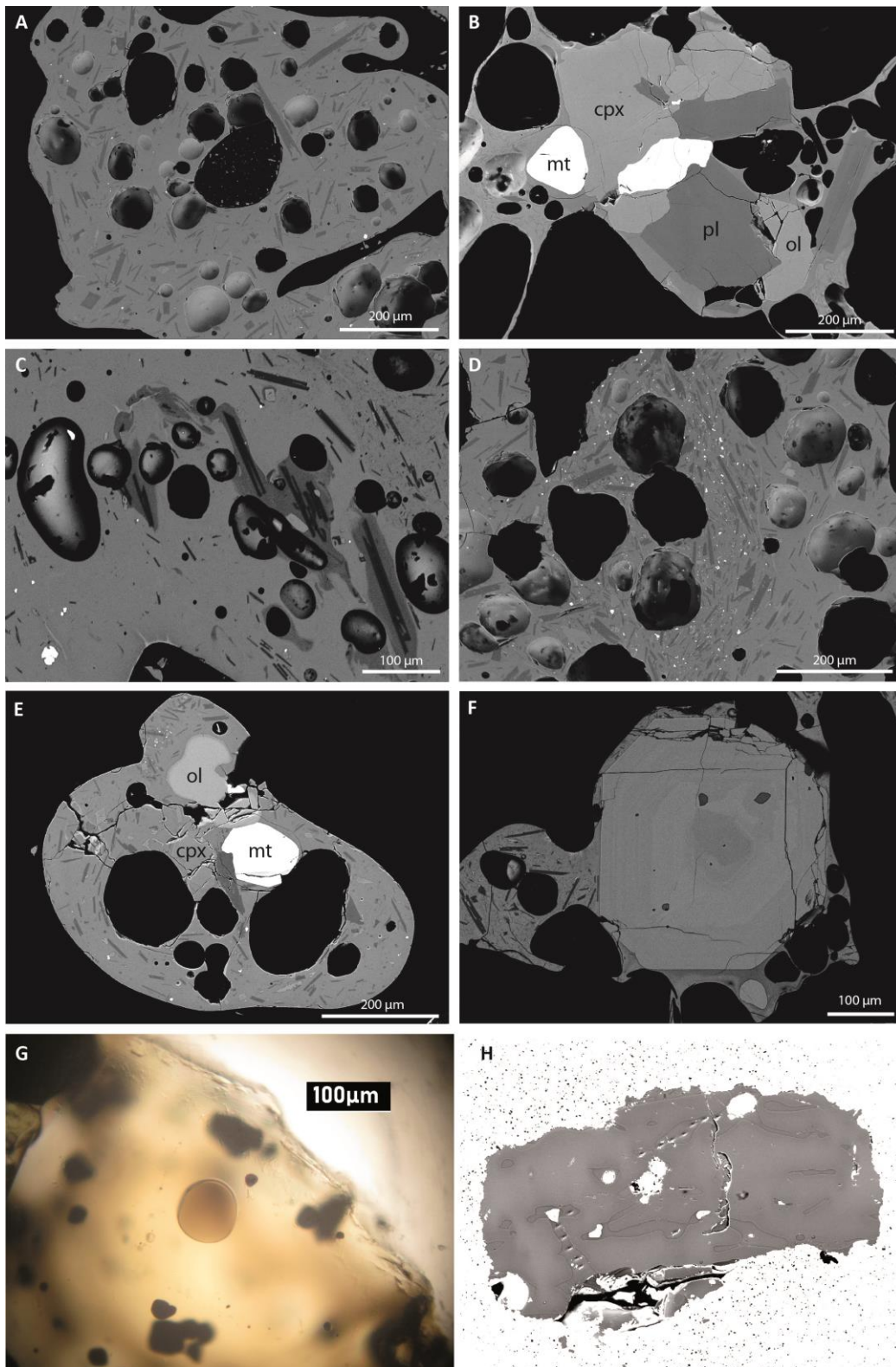
197 **Mineralogy and texture**

198 The bulk of the eruptive products is composed of glassy, crystal-poor, mafic scoriae containing
199 numerous round vesicles (Fig. 2A). There is no significant difference between the subaerial
200 and submarine scoriae. Microlites ($< 200 \mu\text{m}$) are predominantly plagioclase, with small
201 amounts of olivine, clinopyroxene, and magnetite. Larger crystals are rare, and include olivine
202 and plagioclase phenocrysts, sometimes associated with magnetite. Glomerocrysts composed
203 of olivine, clinopyroxene, plagioclase, and magnetite (Fig. 2B) are relatively frequent, and
204 often coated by a darker glass, with a more silicic composition than the dominant mafic glass.
205 Evidence of mingling at the microscopic scale between a mafic glass and a silicic glass has
206 been observed in all eruptive products (Fig. 2C), although the silicic component is always
207 subordinate, and no macroscopic sample with a dominant silicic composition has been found.
208 Apart from the glomerocrysts, the silicic glass often contains plagioclase microlites. There is
209 also evidence of mingling between the phenocryst-poor mafic melt, and a third component with
210 a similar mafic composition but a higher crystallinity (Fig. 2D), indicating that at least three
211 different magmas were erupted at the same time.

212

213 Olivine phenocrysts show a bimodal distribution (Fig. 3A; Table S2; S3 and S4), with a main
214 population around Fo_{74} in equilibrium with the mafic magma (Fe/Mg olivine K_d of $0.30 \pm$
215 0.04), and a second population around Fo_{66} in equilibrium with the silicic magma (Fe/Mg

216 olivine K_d of 0.30 ± 0.04). Most low Mg# and high Mg# olivines (37 / 45) are usually
217 homogeneous. However, when included in the mafic magma, low-Mg# olivines show rounded
218 shapes indicative of dissolution (Fig. 2E), and a small rim with a higher Mg# (four crystals,
219 hereafter called type 1 reverse zoning). Another four olivine crystals (two from each
220 population) show a more pronounced, large scale reverse zoning (type 2 reverse zoning, see
221 timescales section below). Olivine phenocrysts also often contain glassy melt inclusions that
222 preserve the composition of their parent magma (Fig. 2G). A few high-Mg# olivine crystals
223 contain large numbers of inclusions and embayments (Fig. 2H), a texture possibly acquired
224 through initial skeletal growth or as a result of dissolution. Clinopyroxene crystals mostly
225 belong to the glomerocrysts originating from the silicic magma and have an average Mg# of
226 72 (Fig. 3B). Some of those clinopyroxene crystals show evidence of multiple growth stages,
227 dissolution, and oscillatory zoning (Fig. 2F), indicating a protracted history for the cooling and
228 crystallization of the silicic magma. Composition of plagioclase phenocrysts vary from An₈₅ to
229 An₄₈.
230



231

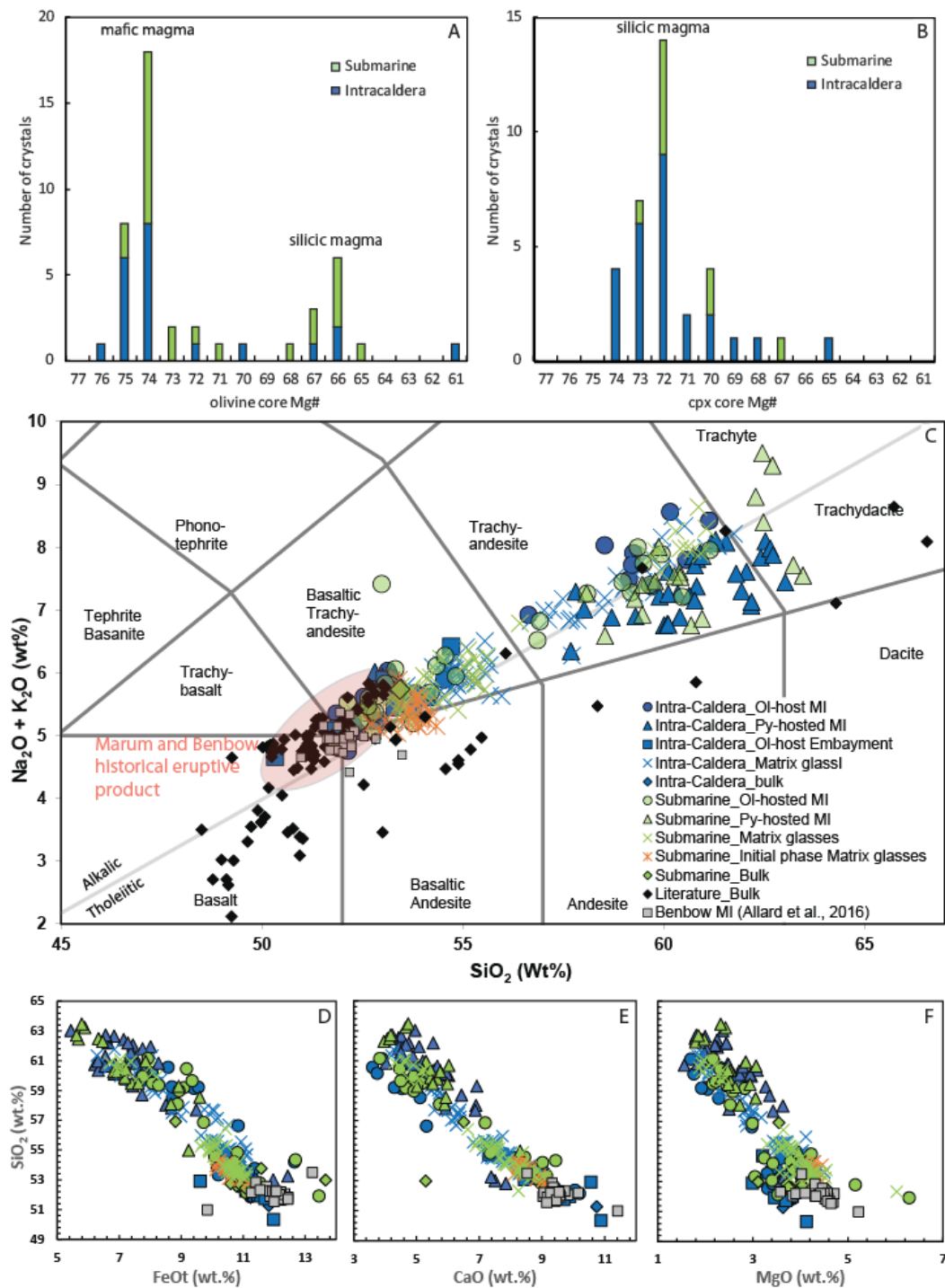
232 **Fig. 2 A-F.** BSE microphotographs of typical textures for subaerial and submarine eruptive
 233 products: **A.** Typical texture of the dominant, crystal-poor mafic magma, with microlites of
 234 plagioclase and minor olivine, clinopyroxene and magnetite. **B.** Plagioclase-clinopyroxene-
 235 olivine-magnetite glomerocryst surrounded by a small layer of microlite-free silicic melt
 236 (darker grey) within a more microlite-rich mafic melt. **C.** Mingling between a crystal-poor
 237 mafic melt (lighter grey, dominant) and a silicic melt (darker grey with darker plagioclase

238 *microlites, less abundant). The upper right side of the picture shows a mafic melt with higher*
 239 *crystallinity. D. Mingling between a crystal-poor and a crystal-rich mafic melts. Both melts*
 240 *contain the same microlite assemblage of plagioclase, olivine, clinopyroxene and magnetite,*
 241 *with less magnetite in the crystal-poor melt. E. Another glomerocryst from the silicic melt,*
 242 *now partially dissolved in the mafic melt, with a thin magnesian overgrowth rim around the*
 243 *olivine crystal. F. Clinopyroxene megacryst, surrounded by a small amount of silicic glass*
 244 *(dark grey, bottom of picture), in turn surrounded by mafic, microlite-rich glass (left). Evidence*
 245 *of multiple growth stages, dissolution, and oscillatory zoning indicates a protracted history for*
 246 *the silicic magma. Dark glassy melt inclusions in the megacryst core have a trachydacitic*
 247 *composition. G. Transmitted light optical microphotograph of an olivine crystal with a glassy*
 248 *melt inclusion. H. BSE microphotograph of an olivine crystal with numerous glassy melt*
 249 *inclusions and embayments (black spots are from SIMS analyses)*
 250

251 **Major elements**

252 The major element composition of melt inclusions, embayments and matrix glasses is given in
 253 **Tables S1** and reported in **Fig. 3** together with bulk rock, matrix glass and melt inclusion
 254 analyses from the literature. The composition of melt inclusions and matrix glasses from the
 255 2018 Ambrym eruption is bimodal with modes centred around ~53 and ~60 wt.% SiO₂ and
 256 compositions ranging from 50 to 63 wt.% SiO₂. Melt inclusions from both the intra-caldera
 257 and submarine eruptions capture the full range of compositions. Matrix glasses from the onset
 258 of submarine eruption however (sample collected on 18 December 2018, labelled initial phase
 259 on subsequent figures) shows very restricted basaltic trachy-andesite composition with no
 260 trachy-andesite to trachy-dacite component as opposed to matrix glasses from subsequent
 261 submarine eruptive activity (sample collected on 04 February 2019) which show both
 262 components and very limited intermediate compositions (i.e. limited chemical mixing). Most
 263 pyroxene-hosted melt inclusions (yet not all) are of the more evolved composition while
 264 olivine-hosted melt inclusions are of both components with very limited intermediate
 265 compositions. This is in agreement with the presence of two populations of olivine phenocrysts
 266 and only one population of clinopyroxene phenocrysts (**Fig. 3 A, B**). Bulk compositions of the
 267 intra-caldera and submarine eruptions are both basaltic trachy-andesite. Melt inclusions have
 268 Mg# ranging from 24 to 60 (assuming all iron is FeO). Host olivine crystals have compositions

269 ranging from Fo₆₅ to Fo₇₆ with no systematic difference between melt inclusions and olivine
270 from intra-caldera and submarine eruptions. The range of observed composition from the 2018
271 eruption is covered by bulk rock analyses of older deposits. Evolved compositions are rare at
272 Ambrym, the one reported here are similar to the high-potassium (HK) andesite to dacite series
273 described by Picard et al. (1994) previously found only around the Lewolembwi crater. The
274 more primitive compositions reported here on the other hand are typical of the recent historic
275 trachy-basaltic to basaltic trachy-andesite activity, very close in composition to that reported
276 for bulk, matrix glass and melt inclusions in recent deposits of lava expelled from Marum and
277 Benbow craters (Firth et al. 2016; Allard et al. 2016a; Sheehan and Barclay 2016) (Fig. 3). The
278 observed range in melt inclusion compositions can be mostly explained in term of mixing
279 between two end members, a more trachy-basaltic (at ~51 wt.% SiO₂) and a more trachy-dacitic
280 (at ~63 wt.% SiO₂) component (estimated end member compositions given in Table S1).
281



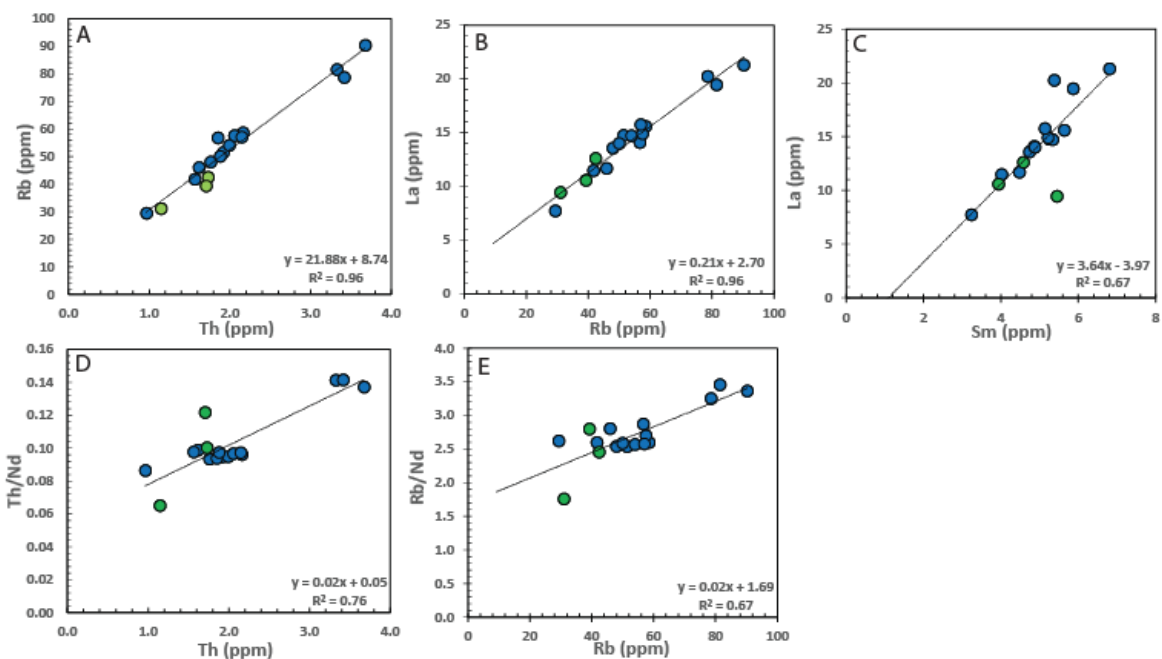
282

283 **Fig. 3 A.** Frequency diagram of the Mg# of olivine cores from the intra-caldera and submarine
 284 eruptions. **B.** Frequency diagram of the Mg# of pyroxene cores from the intra-caldera and
 285 submarine eruptions. **C.** Total alkalis vs silica. Melt inclusions, matrix glasses, and bulk rock
 286 compositions (all normalized to 100% on a volatile-free basis) for the 2018 intra-caldera (blue
 287 symbols) and submarine (green and orange symbols) eruptions. Bulk rock literature data are
 288 from locations covering the entire island (Gorton 1977; Picard et al. 1994; Firth et al. 2016;
 289 Allard et al. 2016a; Sheehan and Barclay 2016). Faded pink ellipse shows the typical
 290 composition of bulk lava, matrix glasses and melt inclusions from 1913 to 2014 eruptive
 291 products from Marum and Benbow craters (Firth et al. 2016; Allard et al. 2016a; Sheehan and
 292 Barclay 2016). **D, E and F.** SiO₂ vs FeO_{tot}, CaO and MgO diagrams

293

294 **Trace elements**

295 Trace and rare earth elements concentrations in melt inclusions and matrix glasses are reported
 296 in **Table S5** and show no systematic differences between the intra caldera and submarine
 297 eruptions albeit limited data. Highly incompatible elements define positive linear correlation
 298 offset from the origin (**Fig. 4A-C**) suggesting mixing between two components or evolution
 299 through partial melting (fractional crystallization is expected to result in linear correlation
 300 passing through the origin, e.g., Schiano et al. 2010). Plots of incompatible trace element ratios
 301 versus abundance (**Fig. 4D and E**) show a positive correlation also consistent with either mixing
 302 or partial melting (as fractional crystallization would result in no variations of the ratio with
 303 concentration, e.g., Schiano et al. 2010).



304

305 **Fig. 4** Trace-element variation diagrams from melt inclusions and matrix glasses from the
 306 intra-caldera (blue) and submarine (green) eruption. **A.** Plot of Rb vs Th. **B.** Plot of La vs
 307 Rb. **C.** Plot of La vs Sm. **D.** Plot of Th/Nd vs Th. **E.** Plot of Rb/Nd vs Rb. Linear regression
 308 lines through all data are presented on each plot
 309

310 Volatiles

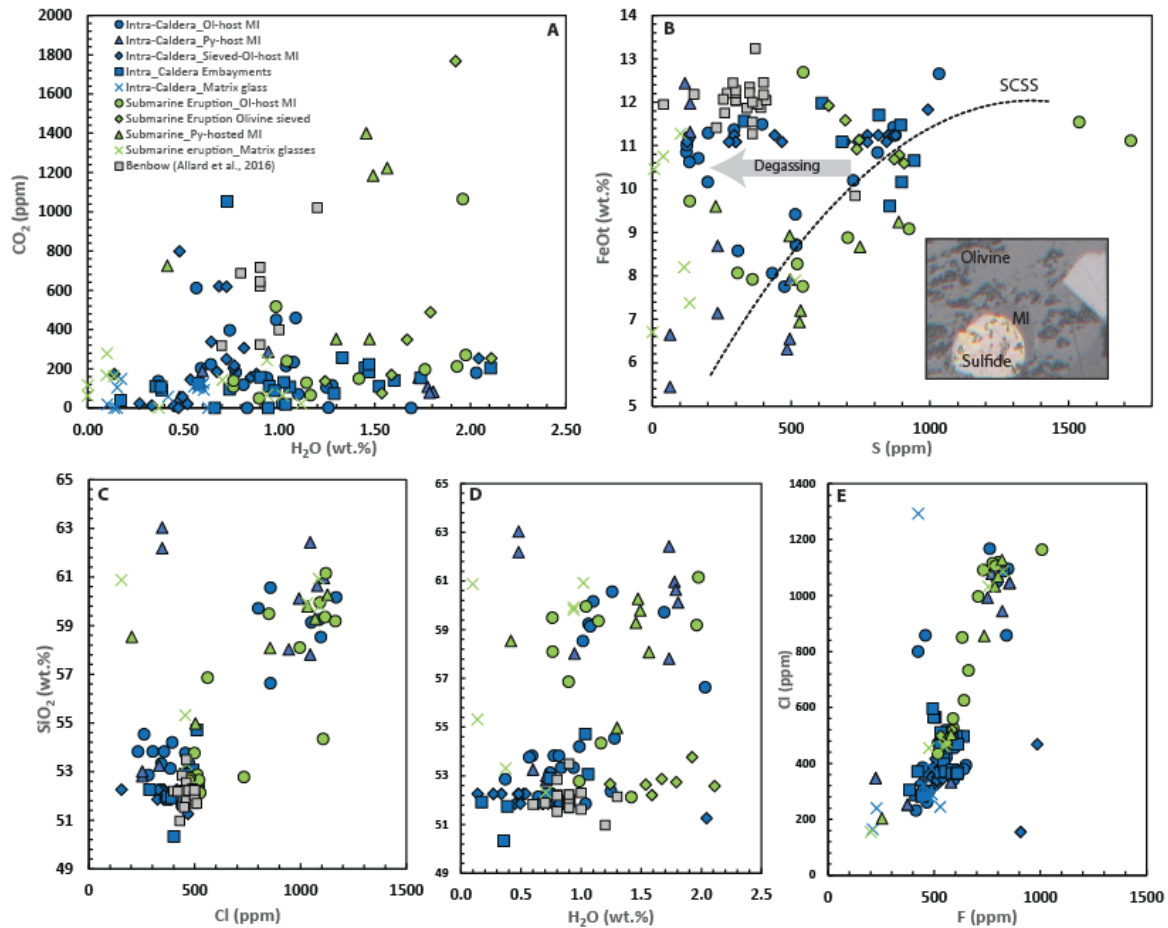
311 Volatile contents from melt inclusions, embayments and matrix glasses from both the intra-
312 caldera and submarine eruptions are given in [Table S6](#). In melt inclusions and matrix glasses,
313 chlorine ranges from below detection to 1200 ppm, fluorine from 200 to 1000 ppm, water from
314 below detection to 2.1 wt.%, CO₂ from below detection to 1800 ppm and sulfur from below
315 detection to 1700 ppm. H₂O and CO₂ values in melt inclusions are regarded as minimum values
316 from the time of entrapment given that H⁺ diffusion out of the melt inclusion (e.g., Chen et al.
317 2011) and CO₂ diffusion in shrinkage bubbles (e.g., Anderson and Brown 1993) cannot be
318 discarded.

319

320 CO₂ vs H₂O abundance in melt inclusion does not follow a single degassing trend but might
321 instead reflect the combination of several distinct degassing paths. Sulfur abundance is partly
322 controlled by degassing as evidenced by the systematic difference between melt inclusions and
323 matrix glasses. The presence of sulfides in some melt inclusions ([Fig. 5](#)) indicates that the melt
324 must have been at saturation with a liquid sulfide phase prior to ascent and degassing. The large
325 size of the sulfides (~20 μm, [Fig. 5](#) and MI photographs in appendix) precludes their formation
326 from secondary processes. This is further indicated by the observed negative correlation
327 between the maximum sulfur content of melt inclusions and their FeO content following the
328 expected trend of sulfur content at sulfide saturation ([Fig. 5](#)). Both chlorine and fluorine
329 contents are correlated with melt composition, the basaltic melts end member having around
330 450 ppm Cl and 550 ppm F and the dacitic melt end member having around 1100 ppm Cl and
331 800 ppm F. The original water content of the more mafic and more silicic melts mostly
332 overlaps, although the mafic component has melt inclusions with low (<0.75 wt.%) water
333 content that are largely absent from the more silicic component indicating more extensive
334 shallow crystallisation. Mafic melt inclusions from the submarine eruption tend to record

335 higher water and CO₂ content than mafic melt inclusions from the intra-caldera eruption. There
 336 are no correlations between the melt inclusion CO₂ content and their major element
 337 composition.

338



339

340 **Fig. 5** Volatile elements abundance in melt inclusions, embayments and matrix glasses from
 341 Ambrym. Inset: reflected-light microphotograph showing a ~20 μ m sulfide in an olivine-hosted
 342 melt inclusion (olivine AMB A12). Dotted curve on panel B show calculations of the sulfur
 343 content at sulfide saturation (SCSS) calculated at 1200°C, 100 MPa using the model of Smythe
 344 et al. (2017)

345

346

347 Geothermobarometry and volatile saturation pressure

348 Melt inclusions entrapment pressures were calculated using the model of Iacono-Marziano et

349 al., (2012) for H₂O-CO₂ saturation pressure. They yield entrapment pressures between 5 and

350 280 MPa (Fig. 6). Given the above-reported error on the volatile content determination and the

351 error on the model, the results can conservatively be taken to be precise at $\pm 20\%$. Yet, as

352 discussed previously, H₂O and CO₂ values are to be considered as minimum values due to
353 possible diffusion of both species. The entrapment pressures are therefore also to be taken as
354 minimum pressure estimates. Another reason to assume these pressure estimates are minimum
355 is that the model of Iacono-Marziano et al., (2012) that we use here has been shown, at least
356 for alkali basalt compositions to overestimate the amount of water that can be dissolved in
357 the melt (see Fig. 13 in Shishkina et al. 2014).

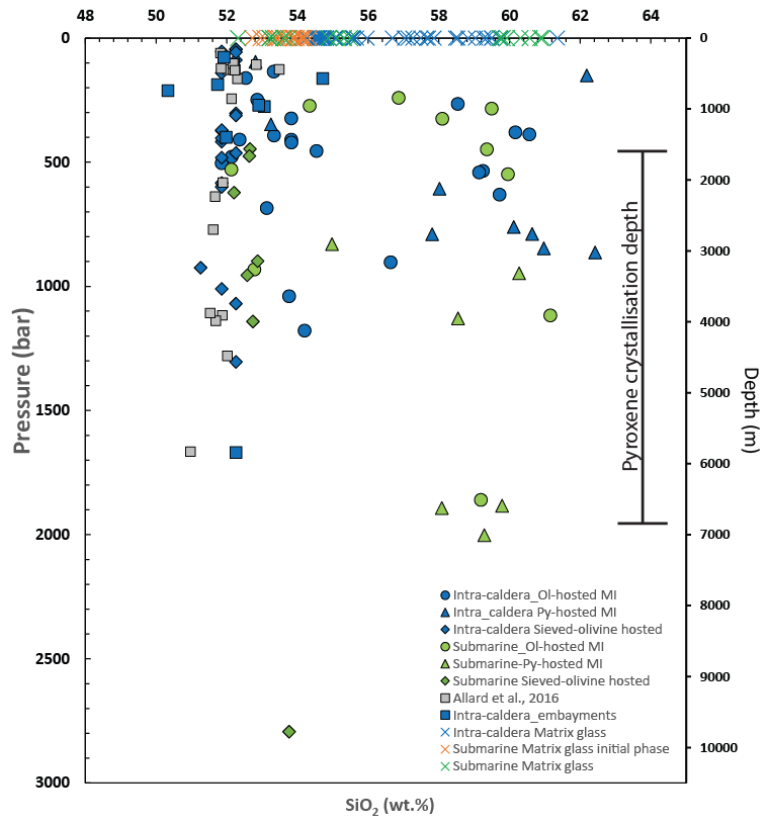
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359 Shown in Fig.6 is the absence of systematic difference between the entrapment pressures of
360 the more basaltic and the more dacitic melt inclusions. Both magmas appear to have
361 crystallized over similar pressure ranges. Similarly, we see no strong systematic differences
362 between the entrapment pressure of melt inclusions from crystals erupted during the intra-
363 caldera and those erupted during the submarine eruption, although we note that the five most
364 deeply entrapped (180 to 280 MPa) inclusions are all from crystals erupted in the submarine
365 eruption suggesting contribution from a deeper magma that was not present in the intra-caldera
366 eruption. Similarly, all the most shallowly entrapped (< 24 MPa) melt inclusions are from
367 crystals erupted during the intra-caldera eruption suggesting that the shallowest (top 750 m)
368 portion of the magmatic system contributed only to the intra-caldera eruption and might have
369 been emptied prior to the submarine eruption.

370 The volatile content of matrix glasses erupted during the submarine eruption should
371 theoretically reflect the hydrostatic pressure under which they were erupted. Volatile contents
372 indicate pressures of last equilibration in the range of 2.5 to 50 MPa. We attribute this large
373 range to the rapid ascent (see ascent rates section) that would have impeded equilibrium
374 degassing (Pichavant et al. 2013) and to the potentially explosive nature of the submarine
375 eruption preventing re-equilibration at extrusion pressure. Taking the lowest recorded

376 equilibrium pressure as the maximum emplacement depth suggests that the submarine eruption
 377 took place less than 250m underwater.

378



379

380 **Fig. 6** Melt inclusions entrapment pressure calculated from H_2O-CO_2 saturation pressure
 381 using the model of Iacono-Marziano et al., (2012) for all data shown including the data from
 382 Allard et al., 2016, compared to melt SiO_2 content. Note that all matrix glasses are arbitrarily
 383 plotted at 1 bar
 384

385 Since all glasses are saturated with olivine, pre-eruptive temperatures were calculated using the
 386 olivine-melt thermometers of Ford et al. (1983) and Beattie et al. (1993). Both thermometers
 387 give near-identical results for the mafic glasses (Ford: 1125 °C; Beattie: 1112 °C; average of
 388 1120 °C). The silicic melt was slightly colder, with estimated temperatures between 1074
 389 (Beattie) and 1112 °C (Ford), for an average temperature of 1100 °C. Taking into account a
 390 1.2 wt% water content in the magma (average of all analysed melt inclusions) results in lower

391 estimated pre-eruptive temperatures of 1078 °C for the mafic magma and 1054 °C for the silicic
392 magma, using the model of Médard and Grove (2008).

393 Previous work on clinopyroxene-based thermobarometry for the 2005-2007 eruptive products
394 of the then-active lava lakes were published by Sheehan and Barclay (2016). Using three
395 different thermometer/barometer combinations (Putirka et al. 2003; eq. (32a) and (32d) of
396 Putirka et al. 2008; eq. (30) and (33) of Putirka et al. 2008), they report a tight range of average
397 storage pressures for Mg#74-76 clinopyroxenes between 390 and 485 MPa. Work on a larger
398 range of eruptive products (from 1913 to 2014, clinopyroxenes Mg#61-74) by Firth et al.
399 (2016) using eq. (30) and (33) of Putirka produces a similar pressure range of 433 ± 86 MPa
400 (excluding one of their datapoint at 1000 MPa). Using the same eq. (30) and (33) of Putirka et
401 al. (2008) on 14 Mg#70-74 clinopyroxenes from the 2018 eruptive products associated with
402 melt inclusions returns average crystallisation pressures of 145 ± 59 MPa at temperatures of
403 1042 ± 25 °C, in excellent agreement with the pre-eruptive temperature for the silicic magma
404 estimated from olivine-melt thermometry. There is no overlap between our dataset and
405 previous datasets (Sheehan and Barclay 2016; Firth et al. 2016), suggesting that two different
406 storage levels were sampled by the 2018 eruption and previous eruptions. Since most
407 clinopyroxene crystals (if not all) come from the silicic magma, we use an average melt
408 composition of the silicic melt inclusions, and average water content of 1.2 wt% to apply the
409 same combination of thermometer and barometer to a larger dataset of 81 clinopyroxene
410 phenocryst analyses. The resulting pre-eruptive pressure is 121 ± 75 MPa at a temperature of
411 1034 ± 8 °C, identical to the values obtained with the more restricted dataset. Although using
412 the same barometer, no value came within error of the pressures determined by Firth et al.
413 (2016) and Sheehan and Barclay (2016), confirming that the 2018 eruption sampled a shallower
414 storage system than previous eruptions. Assuming a crustal density of 2900 kg.m^{-3} , those
415 crystallization pressure translates into crystallization depths of 4.3 ± 2.6 km (or 5.1 ± 2.1 km

416 for the restricted dataset of inclusion-bearing clinopyroxene phenocrysts), in excellent
417 agreement with the deeper part of the storage system imaged through saturation pressure of
418 melt inclusions (see also Allard et al. 2016a).

419

420 **Residence time**

421 The eight olivine crystals that showed significant (greater than 2 %) differences in their
422 forsterite content were investigated through detailed (1 to 10 μm step) compositional EMP
423 profiles. All 8 crystals record reverse zoning with core compositions around Fo₆₅ to Fo₇₀ and
424 rim compositions around Fo₇₀ to Fo₇₆. These reversely zoned crystals are rare because the more
425 silicic magma is volumetrically much smaller than the primitive one and contains much fewer
426 olivine (Fig. 3). As a result, the vast majority of olivine crystals in the mixed deposit are from
427 the mafic component and are in equilibrium with the carrier melt, which is dominantly of mafic
428 composition, hence showing no significant zoning. The chances of finding olivine crystals from
429 the silicic component that have been in contact with the mafic component and started re-
430 equilibrating diffusively are small and only 8 such reversely zoned olivine crystals were found.
431 Two very different diffusion profiles are recorded by these crystals; type 1 profiles (4 crystals)
432 show very sharp compositional variations occurring over a narrow 5 to 20 μm distance to the
433 crystal edge, while the other 4 (type 2 profiles) show very smooth core to rim compositional
434 variations occurring over distances greater than 60 μm from the crystal edge. Diffusion profiles
435 for these 4 crystals could not be modelled accurately as a clear “plateau” in terms of
436 composition was not constrained. Their profiles however are consistent with strict minimum
437 diffusion timescales on the order of years. The four crystals that displayed sharp compositional
438 profiles were modelled using the DIPRA software (Girona and Costa 2013) for diffusion
439 timescale using the variations in Fe-Mg, Mn, Ni and Ca content. Results together with the full
440 list of parameters used for the modelling are given in supplementary Table S7. Modelled Fe-

441 Mg, Mn and Ni diffusion profiles match the natural data extremely well and yield similar time
442 estimates (Fig. 7) confirming that these natural profiles are the result of diffusion and not
443 growth processes. Modelled Ca diffusion profiles tend to have much higher discrepancy, carry
444 large uncertainties, and do not always yield diffusion timescales consistent with those obtained
445 from the Fe-Mg, Mn and Ni diffusion profiles. Any modelled diffusion profile yielding $\geq 15\%$
446 discrepancy with the data was discarded.

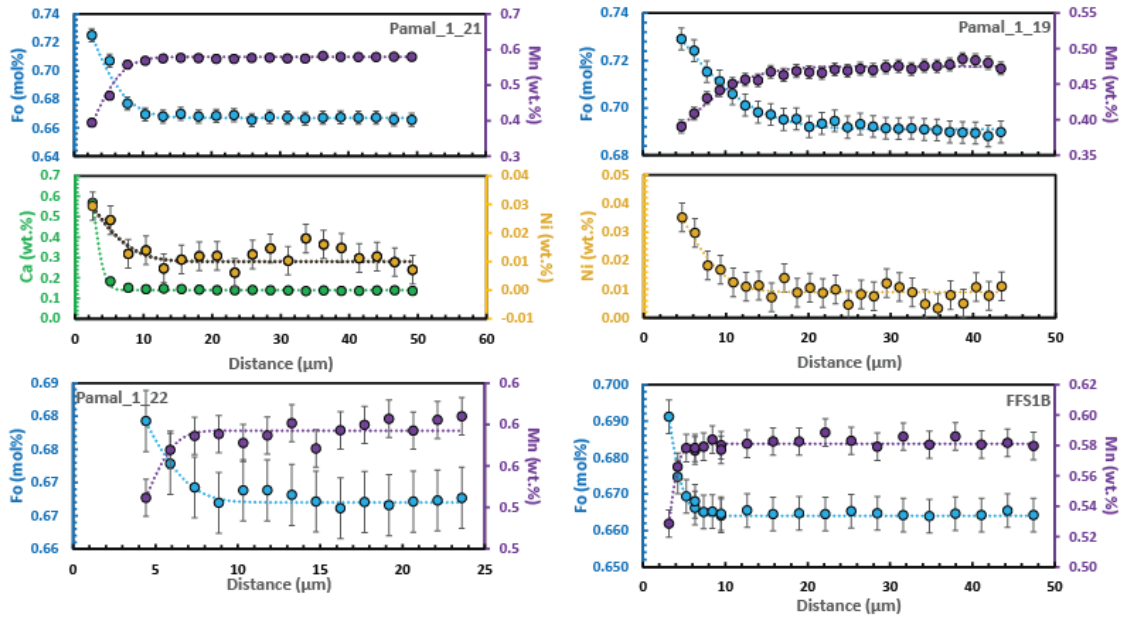
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448 Of the four olivine crystals for which diffusion timescale was investigated, one is from the
449 intra-caldera eruption while the other three are from the submarine eruption. The crystal from
450 the intra-caldera eruption record compositional profiles consistent with a diffusion timescale
451 of less than a day. Olivine crystals from the submarine eruption record compositional profiles
452 consistent with diffusion timescales of about two days indicating that mixing of these olivine
453 crystals (originating from the silicic magma) into the more primitive magma took place one to
454 a few days prior to eruption (Fig. 8).

455

456 The presence of four olivine phenocrysts with diffusion profiles indicative of long-term
457 diffusion, as well as frequent oscillatory zoning within the clinopyroxene phenocrysts indicate
458 the possibility of periodic interaction between the silicic and mafic magmatic systems, with at
459 least one interaction occurring more than a year prior to the 2018 eruption.

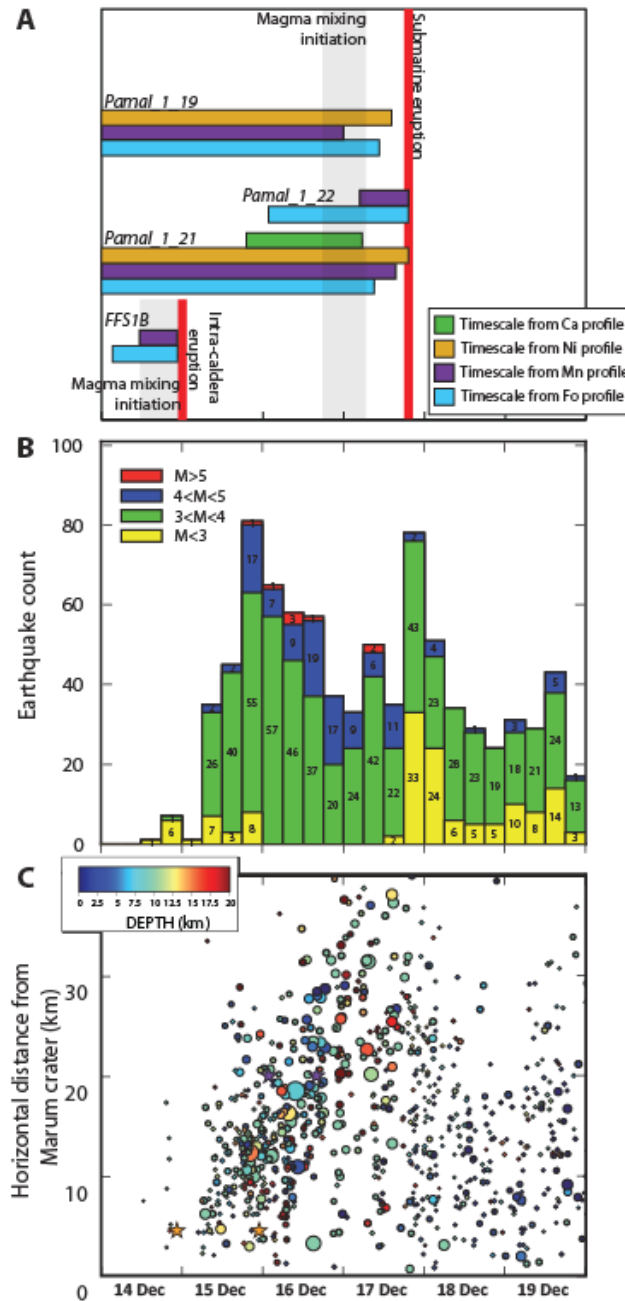
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461

462 **Fig. 7** Results of diffusion modelling (dotted lines) compared to Fo, Mn, Ni and Ca
 463 concentrations profiles (measured by EMPA, error bars are 2σ) in four olivine crystals from
 464 the intra-caldera and submarine eruptions. Modelling was performed using the DIPRA
 465 software (Girona and Costa 2013)

466



467
 468 **Fig. 8 A.** Summary of diffusion timescales obtained from modelling Fo, Mn, Ni and Ca diffusion
 469 profile. Each coloured bar shows the range of possible diffusion timescale taking into account
 470 unknown crystal orientation (modelling along the a and c axes) and the error on each diffusion
 471 model (derived from 2σ error on EMPA and a 25°C error on the estimated temperature). Note
 472 that calculated diffusion timescale, obtained in unit of time, are presented here in term of date
 473 assuming that diffusion stopped at the onset of eruption. The time window where all diffusion
 474 timescale overlap is taken to represent the time at which diffusion started which we interpret
 475 as the initiation of magma mixing (time at which olivine crystals became in contact with a more
 476 primitive melt and started developing reverse zoning). **B.** Evolution of seismicity at Ambrym in
 477 the period 14 to 20 of December (activity prior to 14 December is minimal), adapted from
 478 Shreve et al., (2019). **C.** Evolution of seismicity as a function of distance from Marum, with
 479 earthquakes coloured by depth, adapted from Shreve et al., (2019)
 480

481 **Ascent rates**

482 Volatile diffusion modelling was performed to match the measured H₂O and S concentration
483 profiles (obtained by SIMS) along three melt embayments from the intra-caldera and one from
484 the submarine eruption. All embayments are from melt compositions close to the basaltic end
485 member (i.e. main Marum and Benbow magma). Best fit solutions of the diffusion models are
486 shown in Fig. 9 with initial conditions, model parameters and results given in Table S8. For
487 each embayment we report the results of two type of models: one considering diffusion of the
488 specie of interest only, the other considering diffusion of both species and minimizing error on
489 the fit for both species simultaneously. In all cases, best fit of S profiles only, always result in
490 lower decompression rate estimates (from 0.01 to 0.6 MPa/s) than best fit of H₂O profiles (from
491 0.39 to 2.7 MPa/s). Embayment PG11 and AF2 show H₂O and S profiles that can be fitted well
492 simultaneously (Fig. 9), yielding decompression rates of 2.3 and 2.7 MPa/s respectively. These
493 two embayments are also the ones recording decompression from highest initial pressure (115
494 and 68 MPa respectively). Embayments AD5 and AE38 show significant discrepancy between
495 modelled and measured profiles when trying to fit both species simultaneously. These two
496 embayments record decompressions from lower pressures (34 and 37 MPa respectively). A
497 probable explanation is that PG11 and AF2 record direct magma ascent to the surface (from ~4
498 and ~2.5 km depth respectively) while AD5 and AE38 might have experienced a more complex
499 decompression path, only partially equilibrating at ~1 km depth prior to ascent or ascending in
500 discrete steps. We consider that our best estimates of magma ascent rates during the intra-
501 caldera and submarine eruptions are hence given by the simultaneous modelling of H₂O and S
502 diffusion profiles in embayments AF2 and PG11 respectively, giving ascent rates of 340 and
503 290 km/h (95 and 80 m/s) respectively. This corresponds to magma travel times of about 30
504 seconds from ~2.5 km depth to the surface during the intra-caldera eruption and of 1 minute
505 from ~4 km depth to the surface during the submarine eruption. These ascent rates are

506 extremely high. They are, by far the fastest estimates based on the volatile diffusion in
507 embayment method to date (see compilation in Moussallam et al. 2019) and are comparable to
508 estimates based on bubble number density for the andesitic 1997 Soufriere Hills eruption (12-
509 260 m/s; Giachetti et al. 2010) and basaltic-andesite 1986 Izu-Oshima (60-160 m/s; Toramaru
510 2006) eruptions.

511

512 An estimated $13 \times 10^6 \text{ m}^3$ of basaltic magma was degassed during the intra-caldera eruption in
513 about a day (Shreve et al. 2019) making it a magnitude 3.6 eruption with a log10 mass eruption
514 rate (MER) of 5.6. The amount of material erupted during the submarine eruption is unknown.
515 If we consider the volume of magma in the lateral dike (419 to $532 \times 10^6 \text{ m}^3$, Shreve et al.
516 2019) as representative for an eruption period of two days the submarine eruption would be of
517 magnitude 5.2 with a log10 MER of 7 (but note that the base assumption here is not verifiable
518 and these values should not be quoted). The point is that decompression rates of 2.7 and 2.3
519 MPa/s obtained for the intra-caldera and submarine eruptions respectively are much higher
520 than those obtained for eruptions of comparable magnitude using the same technique (see Fig.
521 13 in Moussallam et al. 2019). Whether these fast ascent rates are exceptional for basaltic
522 magmas will only become apparent as more eruptions are studied but to date they are by far
523 the highest.

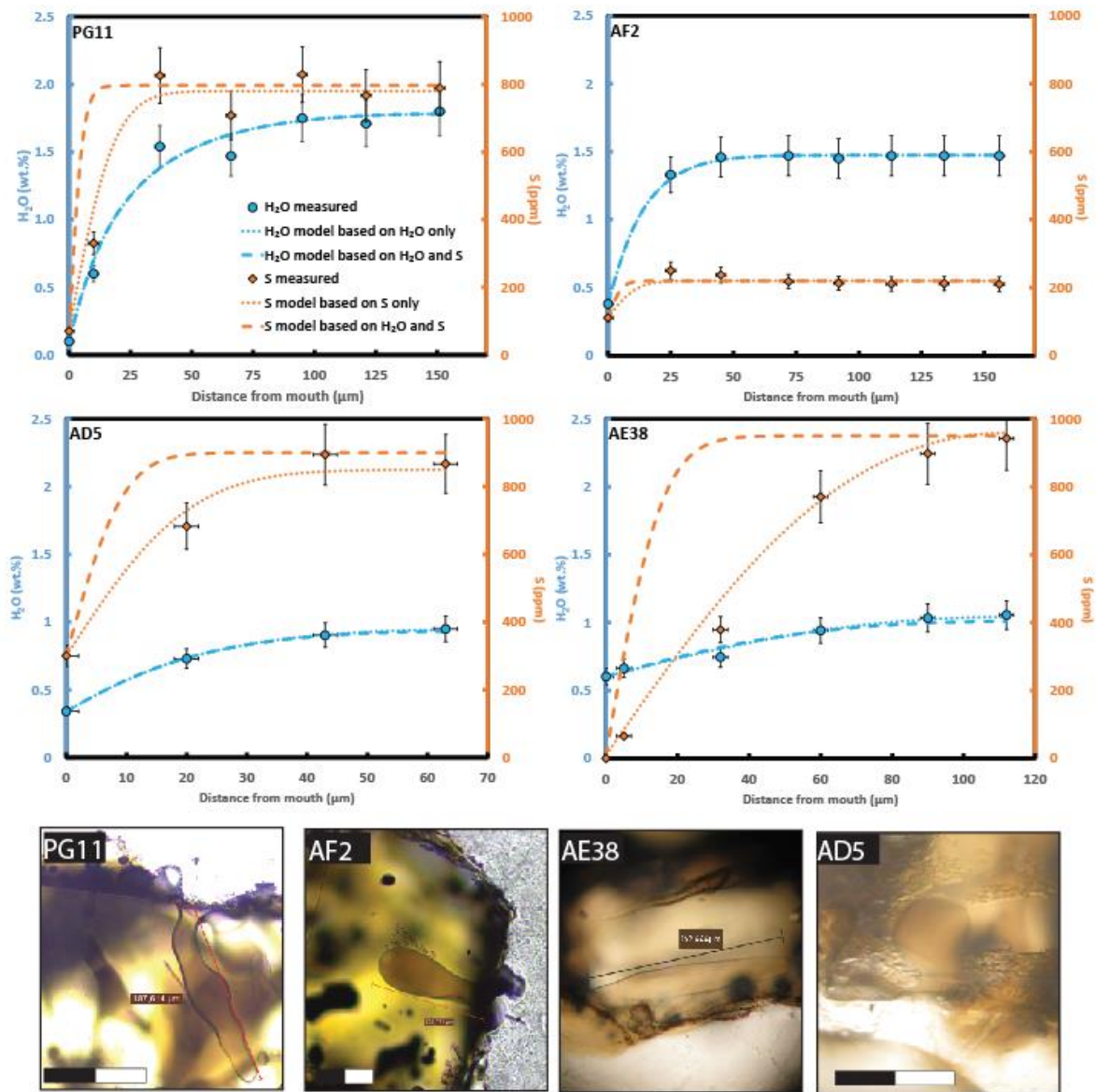
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525 From a simple mass conservation argument, the mass eruption rate and ascent velocity can be
526 used to estimate the conduit radius using:

$$527 \quad M = \pi r^2 \rho u \quad (1)$$

528 where M is mass eruption rate (in kg/s), r is the conduit radius (in m), ρ is the magma density
529 in ($\text{kg} \cdot \text{m}^{-3}$) and u is the velocity in ($\text{m} \cdot \text{s}^{-1}$). This would yield a conduit radius of about 1m for

530 the intra-caldera and 4m for the submarine eruption but we note that such an approach is over-
 531 simplistic as the ascent rate is not a constant and should be much higher at the surface than the
 532 average value calculated from the embayment technique (which assumes a constant
 533 decompression rate) and the eruption (at least the intra-caldera one) occurred along fissures
 534 and not a circular vent.
 535



536

537 **Fig. 9** Results of diffusion modelling compared to concentrations (measured by SIMS) for H_2O
 538 and S in three melt embayments from the intra-caldera and one (PG11) from the submarine
 539 eruption of Ambrym. Best fit results are shown for two type of models, one (dotted lines)
 540 considering diffusion of the specie of interest only, the other (dashed lines) considering

541 *diffusion of both species hence minimizing error on the fit for both species simultaneously.*
542 *Lower row show microphotograph of each embayment, scale bar represents 100 μm length*
543

544 **DISCUSSION**

545 **The Ambrym plumbing system**

546 The crystals and melt inclusions erupted during the 2018 eruption paint a new picture of the
547 plumbing system feeding Ambrym volcano. It is clear from this study and previous work on
548 recent deposits (Firth et al. 2016; Allard et al. 2016a; Sheehan and Barclay 2016) that the main
549 magmatic system underneath Marum and Benbow is of basaltic to basaltic-trachy-andesite
550 composition. Our results, together with previous melt inclusion studies (Allard et al. 2016a) all
551 point to this main magmatic system being vertically extensive, with no clear depth horizon of
552 magma repose/accumulation (Fig. 10). Taking into account the clinopyroxene barometry
553 results of Firth et al. (2016) and Sheehan and Barclay (2016), this vertical magma system might
554 extend to at least 14 km depth, although only the shallowest part of this system was involved
555 in the 2018 eruption. Evidence for this shallow plumbing system reaching up to the surface is
556 provided by the very shallow entrapment pressures of a significant number of melt inclusions,
557 as well as the presence of degassed magma batches, present as a microlite-rich mingled
558 component in the scoriae. The presence of degassed magma is typical of open conduit
559 magmatic systems (e.g., Lautze and Houghton 2005; Gurioli et al. 2014), and evidence of
560 magma convection in the conduits below lava lakes (e.g., Kazahaya et al. 1994; Moussallam et
561 al. 2015b), as was probably the case prior to the 2018 eruption.

562

563 The additional information carried in the 2018 eruption deposits is the clear identification of a
564 second magmatic branch, of trachy-andesitic to trachy-dacitic composition. This branch also
565 appears vertically extensive (Fig. 10). We hypothesise that it must be located underneath the
566 Lewolembwi crater as magma of the same composition make up the Tuff ring forming this

567 crater and a lava flow directly north of it that was emplaced in 1986 (Robin et al. 1991; Picard
568 et al. 1994), as already proposed by Picard et al. (1994). This magmatic batch possibly extends
569 down to 7 km, but not as deep as the main mafic magmatic branch.

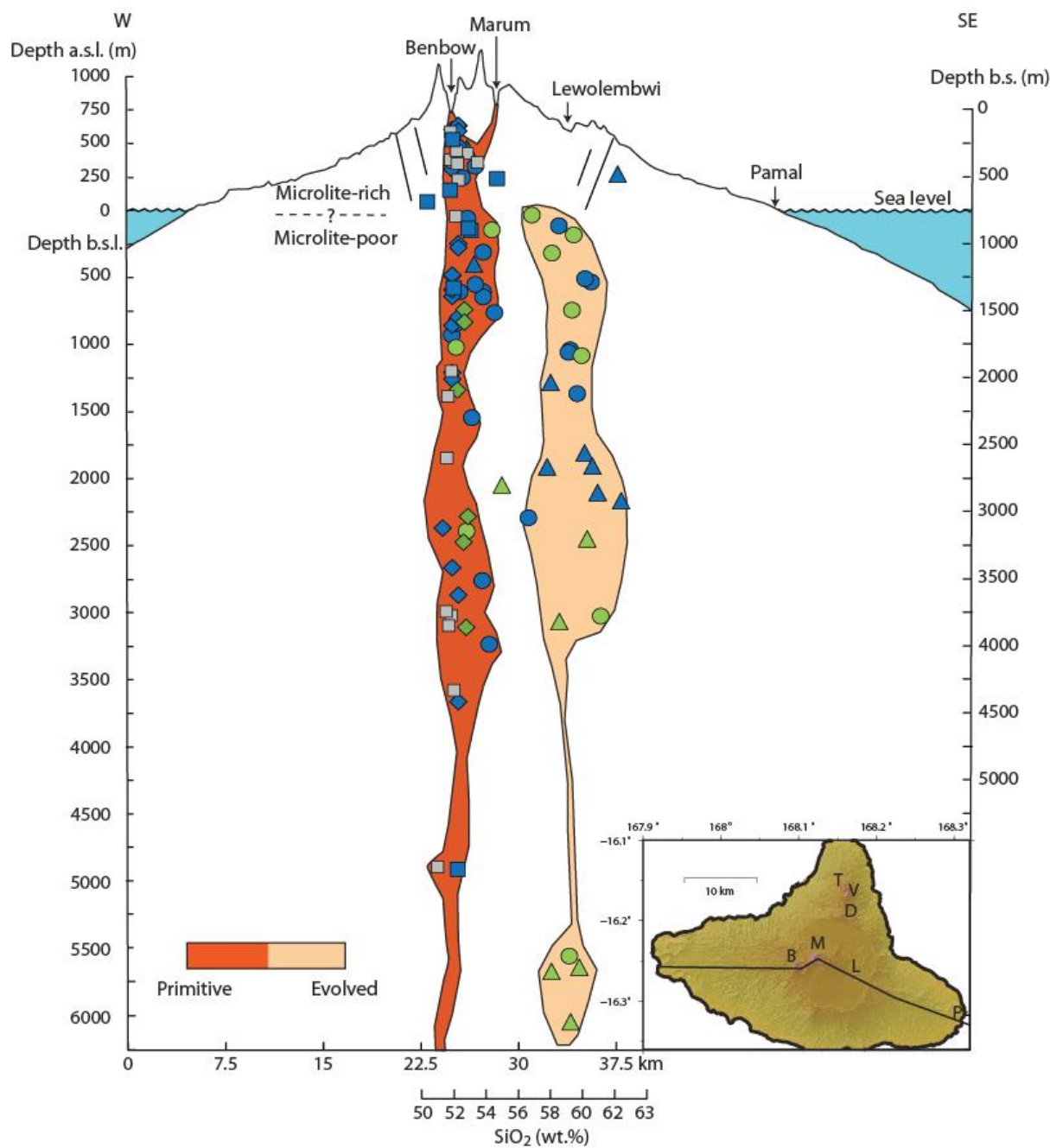
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571 The primitive and evolved magmatic branches are clearly chemically related (Fig.3 and Fig.4).
572 We hypothesise that the evolved, trachy-andesitic to trachy-dacitic magma was previously
573 linked at depth to the same parental source as the more primitive trachy-basaltic magma but
574 that this eastern branch was cut-out from the main magmatic channel and started to differentiate
575 in isolation, generating the Lewolembwi and 1986 deposits and being remobilized during the
576 2018 eruption (see next section).

577

578 Interactions between the main (trachy-basaltic) and peripheral (trachy-andesitic to trachy-
579 dacitic) magma chambers must have also occurred prior to the 2018 eruption as evidenced by
580 the presence of diffusion profiles in some olivines that record timescales in the order of years
581 at the very minimum (i.e. could be decades or hundreds of years the profiles were not complete
582 enough to calculate diffusion timescales). Trace element variations also suggest that chemical
583 mixing might have played a role in the resulting compositions and the presence of frequent
584 oscillatory zoning within the clinopyroxene phenocrysts might be further evidence of episodic
585 mafic input into the more evolved magmatic reservoir. How much episodic chemical mixing
586 has occurred between these two components over the years is unknown. Also unknown is how
587 much such interactions have contributed to past eruptions at Ambrym.

588



589

590 **Fig. 10** Schematic cross-section representing the Ambrym plumbing system on which melt
 591 inclusions entrapment pressure and SiO₂ content are superimposed (with no implied relation
 592 between SiO₂ content and horizontal coordinates of the melt). Inset DEM shows the location
 593 of the in-land part of the transect used to draw the elevation profile. The more mafic (dark
 594 orange) and more silicic (light beige) magma chambers are drawn conceptually, their vertical
 595 extent is constrained by MI entrapment pressures and pyroxene barometry, but their horizontal
 596 extent is unconstrained. We hypothesise that the more mafic magma is located underneath the
 597 Benbow and Marum crater as this is the composition that has feed all known eruptive activity
 598 at these craters while we envision that the more silicic magma is located underneath the
 599 Lewolembwi crater as magma of the same composition as our evolved end-member, make up
 600 the Tuff ring forming this crater and the 1986 flow directly north of it (Robin et al. 1991;
 601 Picard et al. 1994). The upper, degassed part of the mafic system is composed of microlite-rich

602 *magma, while the deeper part of the system is microlite-poor, the exact depth level separating*
603 *the two is unconstrained*

604

605 **Triggers and timing of the 2018 Ambrym eruption**

606 Evidence of magma mingling during the 2018 Ambrym eruption are extremely clear, both in
607 the intra-caldera and in the submarine eruptions. Magma mingling can be seen at the micron to
608 millimetre scale in both deposits (Fig. 2). Chemical mixing is evidenced mostly in the
609 composition of the matrix glasses marginally bridging the trachy-basaltic and trachy-andesitic
610 to trachy-dacitic end member compositions (keeping in mind that these could partially also be
611 reflecting mingling at a very small scale) while the melt inclusion compositions mostly record
612 these two end members but show a clear compositional gap in between at SiO₂ contents
613 between ~55 and ~57 wt.% (Fig. 3). These evidence all suggest, to a first order, that encounter
614 of the two magmas occurred for a relatively short time prior to eruption given that mingling
615 textures are preserved and chemical mixing between the two end members is incomplete. This
616 is confirmed quantitatively by the modelling of diffusion profiles in olivine crystals from the
617 evolved components brought in contact with the more mafic component at calculated
618 timescales of hours to a few days prior to eruption. While it is clear that the encounter of these
619 two magmas played an important part in this eruption the real “trigger” is the event that brought
620 the main branch of the plumbing system (trachy-basalt) in contact with the previously isolated
621 and more differentiated eastern branch (trachy-andesite to trachy-dacite). In the following
622 discussion we combine our petrological findings with visual and geophysical observations to
623 constrain the nature and timing of the processes that occurred during the 2018 eruption.

624

625 1. An over pressurizing magmatic system

626 In the two weeks directly preceding the eruption the height of the main lava lake at Marum
627 crater increased rapidly with a rise in the lake level of ~60 m between 30 November and 14

628 December 2018 (Fig. S1). This increase in lake level likely indicates increased pressurization
629 of the magmatic system around that time.

630

631 2. First diking event, magma mixing, lava lake drainage and intra-caldera eruption.

632 Fe-Mg and Mn diffusion modelling from an olivine crystal erupted between 14 December at
633 23:20 and 15 December at 00:01 UTC at Lewolembwi crater records diffusion timescales <
634 10h, indicating than magma mixing was initiated within 10h of eruption (i.e. in the afternoon
635 of the 14 December 2018). This should be taken as an order of magnitude estimate being based
636 on a single diffusion profile. Geodetic modelling by Shreve et al., (2019) based on inversion
637 of InSAR interferogram for images covering the November 3 to December 15, 00:24 interval
638 identified the emplacement of a shallow dike with 2 m maximum opening (Fig.11). While the
639 exact timing of this dike emplacement cannot be resolved by InSAR, we suggest that this
640 dyking event likely put in contact the two previously disconnected branches of the Ambrym
641 plumbing system with primitive magma from the main branch intersecting evolved magma
642 from the eastern branch and rising together to erupt at the surface. The intersection of the two
643 magma bodies probably occurred at shallow, 1 to 2.5 km depth (according to dike geodetic
644 modelling), draining the top 1 km portion of the main magmatic branch (as evidenced by the
645 abundance of magma from the top 1km in the intra-caldera eruption and lack of any such
646 magma in the subsequent submarine eruption) (Fig. 11), causing the disappearance of the five
647 lava lakes and partial collapse of the Marum and Benbow craters. Magma ascent from ~2.5 km
648 depth to the surface took less than 1 minute for magma ascending directly but was slower for
649 magma experiencing more complex history such as descending and laterally migrating prior to
650 eruption.

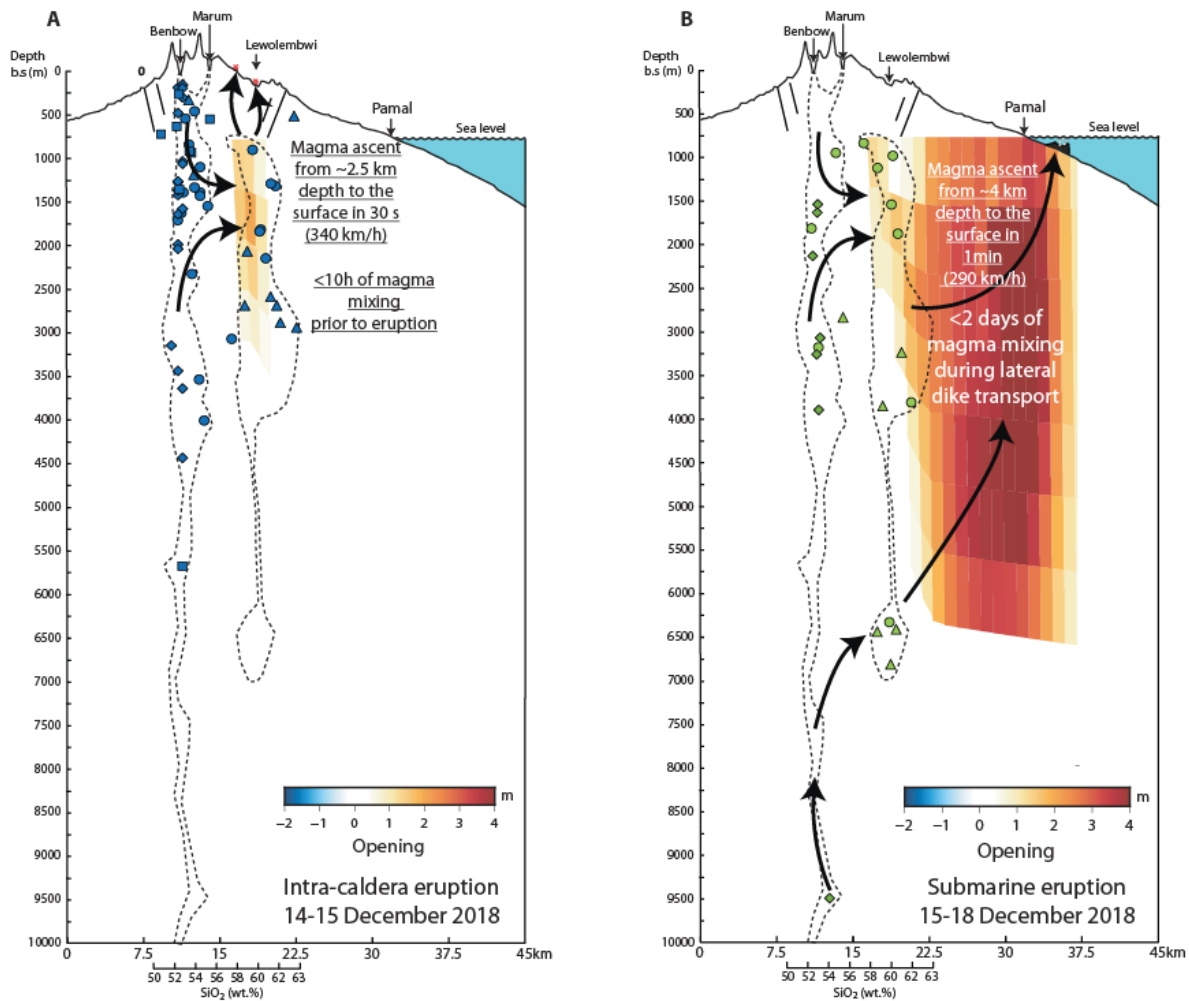
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652 3. Second diking event, magma migration and submarine eruption.

653 Fe-Mg, Mn, Ni and Ca diffusion modelling from three olivine crystals erupted during the
654 submarine eruption are all consistent with diffusion timescales <2 days. The estimated start of
655 the submarine eruption is at 16:00 on 17 December, based on abrupt decrease of seismic
656 moment release (Shreve et al., 2019). Magma mixing was therefore initiated sometime between
657 the 15 December evening and the 17 December (Fig. 7). This corresponds to a period of intense
658 seismic activity and surface deformation starting with a Mw 5.6 strike-slip earthquake on 15
659 December at 20h21 and interpreted from geodetic modelling as the lateral, eastward
660 propagation of a dike with ~ 3 m of opening along >30 km distance (Fig. 11) (Shreve et al.,
661 based on inversion of InSAR interferogram for images covering the 15 to 18 December period).
662 Magma mixing and dike propagation are therefore synchronous, and it seems likely that both
663 magma type became entrained in the same dike where mixing occurred. The tip of the dike
664 however was likely composed solely of the “main branch”, more primitive magma as the very
665 first scoria that reached the island shore showed no indication of mixing and are all of basaltic
666 trachy-andesite to basaltic-andesite composition (Fig. 3). One caveat to consider however is
667 that while the start date of the submarine eruption is well estimated, we do not have an exact
668 date or time on the moment at which our scoria samples were erupted. The eruption might have
669 continued for up to two months (Shreve et al., 2019). Our diffusion timescales might hence not
670 be directly comparable to seismic events preceding the eruption initiation, but they still indicate
671 timescales of magma mixing with <2 days of mixing prior to eruption. While lateral magma
672 transport might have taken up to two days, vertical ascent from ~ 4 km to the submarine eruption
673 site took just a minute, ascending at an average speed of ~ 80 km/h.

674

675



676

677 **Fig. 11** Schematic cross-section of Ambrym volcano comparing the entrapment pressure of
 678 melt inclusions erupted during (A) the intra-caldera eruption and (B) the submarine eruption
 679 with the depth of diking events in the period (A) November 3 to December 15 at 00:24 and (B)
 680 December 15 to 18 (geodetic models from Shreve et al., 2019). Dotted lines represent the
 681 hypothesized locations of the primitive and evolved magmatic branches (see Fig. 10 caption
 682 and text for details). Black arrows show the inferred directions of magma migration. In (A) the
 683 upper ~1km of the main magmatic branch underneath Benbow and Marum is drained, mixes
 684 with the eastern, evolved magmatic branch and erupted together at Lewolembwi (the
 685 composition of the lava erupted synchronously between Lewolembwi and Marum is unknown).
 686 In (B) eastward magma migration might originate, or at least has a contribution from deeper
 687 levels. Timescales of magma mixing and ascent rate from diffusion modelling of reverse
 688 compositional zoning in olivine and volatile in melt embayments are reported for each eruption
 689

690

691 Comparison with other lava-lake-draining rift eruptions

692 If we look in detail at the last three eruptions that terminated or partially drained lava lakes at
693 basaltic volcanoes along a rift axis (2017-2020 Erta 'Ale, 2018 Kīlauea and 2018 Ambrym),
694 clear patterns seem to emerge. First, they all seem to be preceded by high levels of their lava
695 lake. At Kīlauea, the 2018 eruption was preceded by a high level of the lava lake at the summit
696 crater and a rising lava pond at Pu'u 'Ō'ō (Neal et al. 2019). At Erta 'Ale the lake level had
697 been rising (yet not constantly) since 2000 (Barnie et al. 2016), overflowing its bank directly
698 prior to the 2017 fissure eruption (Global Volcanism Program, 2017). As seen previously, at
699 Ambrym the lake level rose rapidly in the weeks prior to eruption (Fig. S1). Another, older
700 example is the 1977 lake-draining fissure eruption of Nyiragongo Volcano, which was
701 preceded by a rise in the lake level of 200 m from 1959 to 1976 (Tazieff 1977). An inescapable
702 conclusion seems to be that pressure build up in the magmatic system (as tracked by rising lava
703 lake level), always precedes this type of eruption.

704

705 In all these examples, lake drainage (or subsidence) is then synchronous with magma migration
706 and lateral fissure eruptions (Tazieff 1977; Moore et al. 2019; Neal et al. 2019; Shreve et al.
707 2019). This phenomenon was (to our knowledge) first documented at Kīlauea volcano in 1924
708 where drainage of the hundred-year-long lava lake accompanied an intrusion in the eastern rift
709 zone (Jaggard and Finch 1924) and has since been documented extensively at Kīlauea (see
710 review by Patrick et al. 2019). Yet another example is the 1913 eruption of Ambrym which
711 had a remarkably similar pattern to the 2018 eruption. In 1913, the eruption began with a fissure
712 eruption inside the caldera followed by magma migration along the rift axis (to the west) and
713 magma eruption near the shoreline (Németh and Cronin 2011). Although it is not known how
714 the lava lake level responded at that time, Németh and Cronin (2008)'s investigation of

715 Marum's crater walls revealed that episodes of lava lake pounding and subsequent drainage to
716 feed flank eruptions and excavation of the crater by associated phreatic to phreatomagmatic
717 eruptions has been a common occurrence throughout the volcano's history. Lake drainage
718 through dike migration and flank eruption is hence probably as recurrent a phenomenon at
719 Ambrym as it is at Kīlauea.

720

721 A peculiarity of the 2018 Ambrym eruption is the extrusion of evolved lava together with lava
722 of the same composition as the lava lakes. As seen in the previous section, it appears that a dike
723 propagating from the main magmatic branch intersected and remobilised evolved magma from
724 a peripheral magma chamber. Interestingly the same phenomenon seems to have occur during
725 the 2018 Kīlauea eruption during which an evolved (andesitic) peripheral magmatic pocket
726 was intersected by the main dike and remobilised during the eruption (although in the Kīlauea
727 case this occurred significantly further down-rift from the main system, Gansecki et al. 2019).
728 The presence of such isolated, differentiated and still eruptible magmatic branches at the
729 periphery of main magmatic systems might therefore be a more common occurrence than
730 previously realised.

731 CONCLUSIONS

732 The 2018 Ambrym eruption, whilst volumetrically mostly occurring underground, was
733 spectacular. It drained five active lava lakes, caused partial collapse of their crater accompanied
734 with large, possibly phreatic, explosions, a 2 m subsidence of the 12 km diameter caldera,
735 magma migration for 30 km horizontal distance and eruption of mingled and slightly
736 chemically mixed magma at intra-caldera fissure eruptions and just off the SE coast as a
737 shallow submarine eruption. Much like previous ones, this eruption highlighted the fact that
738 whilst most population centres on Ambrym island are located at large distance from the main
739 vents, those located along the rift axis will continue to be at risk and impacted by future eruptive
740 events as magma migrates quickly to the shoreline. In this contribution we presented the results
741 of major and volatile element analyses in bulk rocks, matrix glasses, melt inclusions,
742 embayments, and minerals to shed light on the nature and timing of magmatic processes
743 operating during this eruption. The main conclusions we draw from our findings are (1) The
744 eruption began with the meeting, mingling and limited chemical mixing of trachy-basaltic
745 magma from the main magmatic system with more evolved trachy-andesitic to trachy-dacitic
746 magma from an older peripheral and previously cut-off branch. (2) In detail, the trachy-basaltic
747 magma is itself composed of two mingled component, one microlite-poor and one microlite-
748 rich, the second possibly reflecting degassing processes due to convection at shallow depth
749 underneath the lava lakes. (3) The primitive and evolved magmatic branches interact
750 periodically, with at least one interaction occurring more than a year prior to the 2018 eruption.
751 (4) Magma mixing took place less than 10h prior to intra-caldera eruption and for about 2 days
752 during magma transport in a >30 km dike from the centre of the island to the coast prior to
753 submarine eruption. Magma ascent from 2.5 and 4 km depth to the surface took place at rates
754 in the order of 95 ± 24 to 80 ± 6 m/s. (5) Comparison with other lava-lake-draining eruptions
755 reveals that lake level rise – indicating pressurisation of the magmatic system – always

756 precedes this type of eruption, highlighting the usefulness of this parameter for future
 757 monitoring. Furthermore, the presence of peripheral, more evolved magma pockets, cut-off
 758 from the main magmatic system but still mobilizable and eruptible could be a more common
 759 occurrence than previously realised.

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