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Fast ascent rate during the 2017-2018 Plinian eruption of Ambae (Aoba) volcano; a petrological investigation.

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ABSTRACT

In September 2017, after more than a hundred years of quiescence, Ambae (Aoba), Vanuatu's largest volcano, entered a new phase of eruptive activity, triggering the evacuation of the island's 11,000 inhabitants resulting in the largest volcanic disaster in the country's history. Three subsequent eruptive phases in November 2017, March 2018 and July 2018 expelled some of the largest tropospheric and stratospheric SO₂ clouds observed in the last decade. Here, we investigate the mechanisms and dynamics of this eruption. We use major elements, trace elements, and volatiles in olivine and clinopyroxene hosted melt inclusions, embayments, crystals and matrix glasses together with clinopyroxene geobarometry and

olivine, plagioclase and clinopyroxene geothermometry to reconstruct the physical and chemical evolution of the magma as it ascends to the surface. Volatile elements in melt inclusions and geobarometry data suggest that the magma originated from depth of ~ 14 km before residing at shallow (~ 0.5 to 3 km) levels. Magma ascent to the surface was likely facilitated by shallow phreatic eruptions that opened a pathway for magma to ascend. Succeeding eruptive phases are characterised by increasingly primitive compositions with evidence of small amounts of mixing having taken place. Mg-Fe exchange diffusion modelling yields olivine residence times in the magma chamber ranging from a few days to a year prior to eruption. Diffusion modelling of volatiles along embayments (melt channels) from the first two phases of activity and microlite number density suggest rapid magma ascent in the range of 15 to 270 km/h, 4 to 75 m/s (decompression rates of 0.1 to ~ 2 MPa/s) corresponding to a short travel time between the top of the shallow reservoir and the surface of less than two minutes.

INTRODUCTION

Ambae (also known as Aoba; $167^{\circ}50'E$, $15^{\circ}23'S$), is an island volcano located in the central part of the Vanuatu archipelago. The volcano summit is located 3900 m above the sea floor and 1496 m above sea level (Fig. 1). The emerged part of the edifice is 38×16 km and host a pre- 2017 eruption population of about $11,000$ inhabitants. Although part of a particularly active volcanic arc, Ambae's magmatic system has been very little studied compared to its volcanic neighbour Ambrym (e.g., Firth et al. 2016; Allard et al. 2016; Sheehan and Barclay 2016). Yet the volcano's magmatic activity together with the presence of large acid lakes at the volcano's summit, poses a particularly high risk to the island's population due to the potential for highly explosive phreato-magmatic eruptions and especially lahars. An understanding of Ambae's magmatic system is an essential first step towards a better

comprehension of its past and forecasting of its future activity. In this contribution, we aim to draw such an understanding by studying deposits from the 2017-2018 eruption of Ambae, the largest in recorded history.

Ambae is located in the central part of the Vanuatu arc, created by the north-eastward subduction of the Australian plate underneath the Pacific plate (e.g., Daniel et al. 1989). This central part of the arc is characterised by the collision of the D'Entrecasteaux ridge, a 100 km wide Eocene–Oligocene island arc complex on the Australian subducting plate, with the fore-arc (e.g., Collot et al. 1985). Ambae is a shield volcano composed of a pile of primitive basalts (Gorton 1977; Eggins 1993). At the summit is a 5 x 4 km caldera partially filled by 3 lakes one of which was considered among the world's largest acid crater lakes (Voui lake) (Bani et al. 2009a). According to local knowledge, the last significant eruption occurred about 350 years ago, with lava flows reaching the sea near N'dui N'dui (Warden 1970). Volcanic activity at Ambae in the last hundred years has been limited to fumarolic manifestation (Wiart 1995; Bani et al. 2009b). However since the early 1990s Ambae resumed eruptive activity within lake Voui, first with a phreatic eruption in 1995, followed by a Surtseyan eruption in 2005 that formed a 500 m width and 50 m high islet in the centre of the lake (Nemeth et al. 2006). This latter eruption has disrupted the equilibrium within lake Voui leading to a spectacular colour change from blue to red in 2006 (Bani et al. 2009b).

On 6 September 2017, a new eruptive period started with series of strong explosions (heard from neighbouring islands) that generated ash-laden plumes into the atmosphere. These first eruptions were likely phreatic, judging by the absence of juvenile material in the ejected products (Bani et al., in prep.), but injected large quantities (38,000 tons) of SO₂ into the atmosphere (OMI data, Bani et al., in prep.). On 22 September 2017, lava reached the surface

76 producing fire fountains, Strombolian eruptions and lava flows. In early October 2017 the
77 surficial activity had subsided, the period from 22 September to 6 October 2017 is referred to
78 here as phase 1 of activity. On 21 October 2017, a second phase of high-intensity eruptive
79 activity started. Intense ash fall and acid rains were reported on coastal villages. 39,000 tons
80 of SO₂ were injected on the first two days of activity alone (OMPS data, Bani et al., in prep.).
81 By the end of November, the eruption intensity had subsided with ash and gas impacts
82 limited only to the top of the island. The period from 21 October to 7 December 2017 is
83 referred to here as phase 2 of activity. In mid-March 2018 a third intense eruptive phase
84 started producing ash falls and acid rains with even more destructive impacts on the
85 population and environment than during the prior phases. A night glow, presumably from fire
86 fountains, could be seen from Maewo and Pantacost islands, 50 km away from the summit.
87 Gas jetting was sustained for more than 20 days and clearly audible in a 15 km radius. The
88 eruption intensity decreased around 15 April after significant release of volcanic gases to the
89 atmosphere (123,000 tons released on April 6, 2018 alone; OMI data, Bani et al., in prep.).
90 The period from mid-March to mid-April 2018 is referred to here as phase 3 of activity.
91 Three months later, in mid-July 2018, the eruption intensity increased once again to its most
92 violent and destructive eruptive period. Ash fall covered the entire island, momentarily
93 plunging coastal villages in total darkness, causing houses to collapse. By early August 2018
94 the eruption intensity had waned. The period of July 2018 is referred to here as phase 4 of
95 activity. This July 2018 episode released an estimated 400,000 tons of sulfur dioxide into the
96 upper troposphere and stratosphere at an altitude of 16.9 km (NASA Earth Observatory,
97 2019). All together an estimated 3.2 Tg of SO₂ was released during the 2017-2018 Ambae
98 eruption placing it among the top 6 largest eruption of the last 40 years in terms of SO₂
99 loading (Bani et al., in prep.). A detailed narrative of the eruption and its impacts in terms of
100 ash fall, lahar and gas emissions will be presented elsewhere (Bani et al., in prep.).

101

102 In this contribution, we perform a petrological investigation of the magma expelled during
103 the four phases of activity. We present major element, trace element, and volatile content in
104 olivine and clinopyroxene hosted melt inclusions, embayments, matrix glasses and bulk rock
105 together with clinopyroxene geobarometry, olivine, plagioclase, and clinopyroxene
106 geothermometry and major element and volatile diffusion modelling. We find that the Ambae
107 plumbing system extends from 0.5 to 14 km depth with most of the magma involved in the
108 2017-2018 eruption sourced from a shallow reservoir at 0.5 to 3 km depth. Phreatic eruptions
109 in early September 2017 likely served as trigger to or facilitated the subsequent magmatic
110 activity but was itself likely triggered by magmatic gases. Olivine residence times in the
111 magma chamber(s) range from a few days to a year prior to eruption. We constrain the ascent
112 rate of the magma from the shallow reservoir to the surface at 4 to 75 m/s. This fast ascent
113 rate suggests that large amounts of pre-eruptive volatiles must have been present in the
114 reservoir, consistent with satellite observations of large SO₂ emissions and of eruptive clouds
115 reaching the lower stratosphere.

116

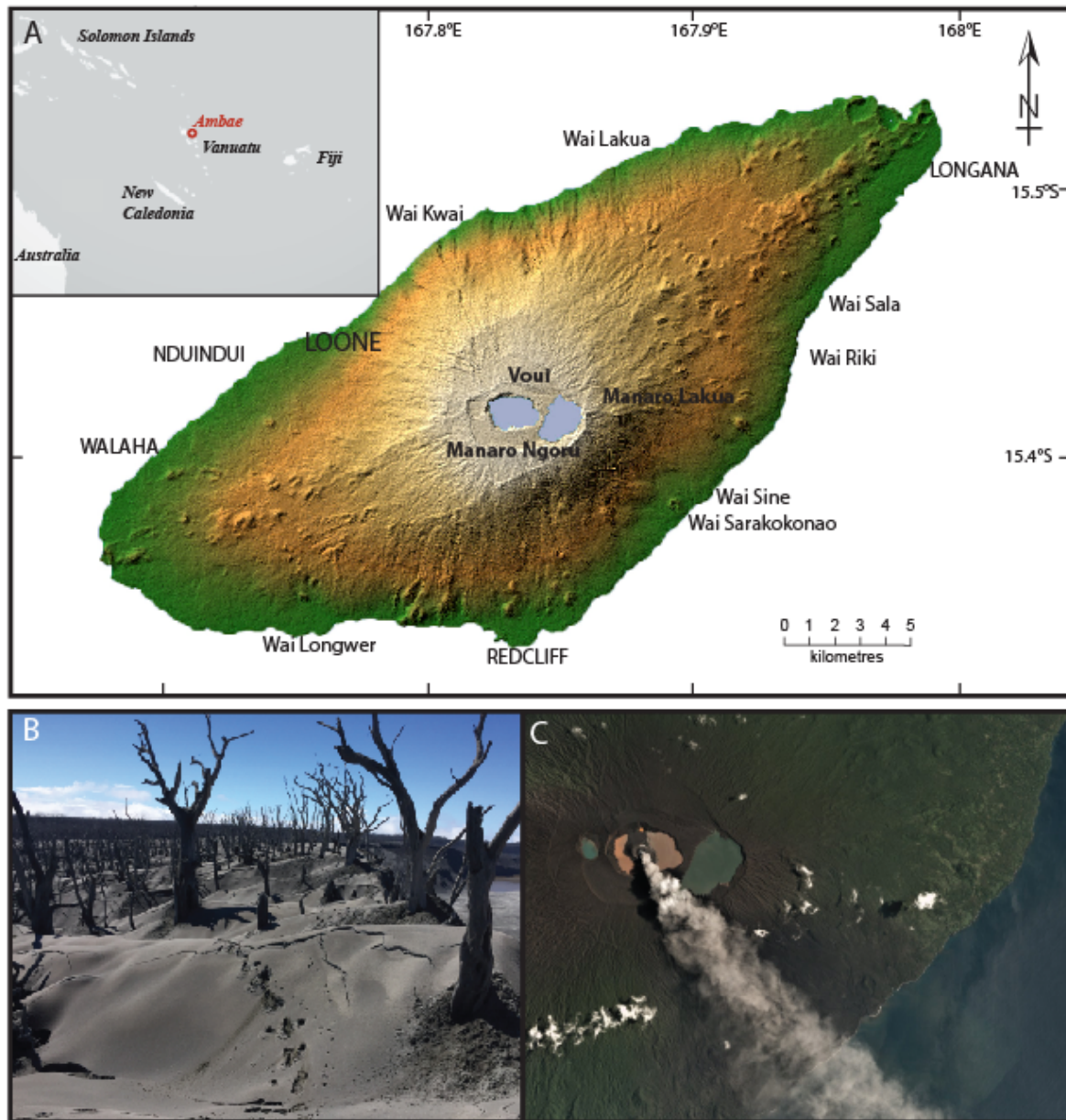


Figure 1: *A. Topographic hill shaded Shuttle Radar Topography Mission (SRTM) image of Ambae island (credit NASA). Location of all coastal settlements are marked together with the location of Voui lake. Upper left inset shows the location of Ambae/Aoba island within the south Pacific. B. Ash deposit on the north-western part of Voui crater (image take on 21 May 2018). C. Planetscope Ortho tile satellite image of the summit area taken 10 March 2018 showing the strong degassing at that time (image credit: ESA)*

METHOD

Samples

Tephra samples from phase one (September 2017), two (November 2017) and three (March 2018) were collected in the summit area within a few days to weeks of deposition. Ash samples from phase four (July 2018) were collected 10 km west of the vent during deposition. Samples from phase 1, 2 and 3 consisted of lapilli-size scoria with fragments typically about 1 cm in diameter. Scoriae from eruptive phase 1, 2 and 3 and ash from eruptive phase 4 were embedded in epoxy resin and polished to investigate their texture and the composition of crystal phases and interstitial glasses. For eruptive phase 1, 2 and 3, part of the same samples was crushed and sieved, and melt-inclusion bearing, euhedral and unbroken olivine and pyroxene crystals were handpicked under a binocular microscope (in the 0.5- and 2-mm grain size fraction). Melt inclusions and embayments were exposed by polishing single crystals individually. Olivine crystals were oriented for polishing in order to expose a maximum length of the embayments, from the end to the mouth of the embayment. Polishing was performed without the use of diamond as to entirely avoid possible C contamination in subsequent ion probe analysis, instead silicon carbide mats for coarse polishing and corundum mats for final 3 μm , 1 μm and $\frac{1}{4}$ μm polishing were used. Single crystals were subsequently mounted in indium. Melt inclusions are entirely glassy, often contain a contraction bubble and are typically about 50 to 200 μm in diameter (Fig. 2B).

Analyses of major elements

Bulk tephra samples were powdered, mixed with LiBO_2 , and melted in an induction oven at 1050 $^{\circ}\text{C}$ for 4.5 min using graphite crucibles. Resulting glass beads were then dissolved in a solution of deionized water and nitric acid and finally diluted by a factor of 2000. The

solutions were analysed by ICP-AES at Laboratoire Magmas et Volcans in Clermont-Ferrand.

Individual melt inclusions, embayments, matrix glasses and crystals were analysed with a Cameca SX100 electron microprobe at the Laboratoire Magmas et Volcans in Clermont-Ferrand. For crystalline phases, beam conditions of 15 kV and 15 nA were employed, while for glass analyses, the beam current was 6-8 nA and the beam defocussed to 10 μ m. Sodium was analysed first as to limit the effects of Na loss. The instrument was calibrated on natural and synthetic mineral standards and glasses : wollastonite (Si, Ca), MnTiO₃ (Ti, Mn), Al₂O₃ (Al in minerals), standard glass VG2 (Al in glasses, see Óladóttir et al. 2008), Cr₂O₃ (Cr), fayalite (Fe), forsterite (Mg), NiO (Ni), albite (Na), orthoclase (K) and apatite (P). Counting times were 10s minimum. Olivine core to rim transects were acquired by making measurements every 10 μ m from the crystal centre to the rim along a line perpendicular to the crystal face. In order to produce accurate analyses for minor elements, acceleration voltage was increased to 20 kV, beam current to 100 nA, and counting times to 20 s (Fe), 50 s (Mn, Ni), and 100 s (Al, P).

Ion Probe analyses of volatile elements

Volatile (H₂O, CO₂, Cl, F, S) content in melt inclusions, embayments, and matrix glasses were determined using a Cameca IMS 1280 ion microprobe at CRPG-CNRS-Nancy. Analytical conditions were similar to other volatile studies (e.g., Hauri et al. 2002; Bouvier et al. 2008; Shimizu et al. 2009; Rose-Koga et al. 2014; Moussallam et al. 2015). We used a Cs⁺ primary beam with a current of 1 nA and an electron gun to compensate for charge build-up at the sample surface. A 3-minute pre-sputter with a 30 \times 30 μ m square raster was applied, then analyses were performed on the 15 to 20 μ m spot in the center of the rastered-clean area by a mechanical aperture placed at the secondary ion image plane. The mass resolving \approx 7000

(with the contrast aperture at 400 μm , the energy aperture at 40 eV, the entrance slit at 52 μm and the exit slit at 173 μm) meant that complete discrimination of mass interferences was achieved ($^{34}\text{S}^1\text{H}$ on ^{35}Cl , ^{17}O on $^{16}\text{O}^1\text{H}$, $^{29}\text{Si}^1\text{H}$ on ^{30}Si and $^{31}\text{P}^1\text{H}$ on ^{32}S). We collected signals for ^{12}C (8 seconds), ^{17}O (3 seconds), $^{16}\text{O}^1\text{H}$ (6 seconds), ^{18}O (3 seconds), ^{19}F (4 seconds), ^{27}Al (3 seconds), ^{30}Si (3 seconds), ^{32}S (4 seconds) and ^{35}Cl (6 seconds; counting times in parenthesis), with 2 seconds waiting time after each switch of the magnet. This cycle was repeated 10 times during one analysis. One measurement lasted 12 min per spot.

The concentrations were determined using calibration curves (see supplementary materials) obtained, at the beginning and at the end of each session, by measuring a set of natural basaltic glasses during the same session and under the same analytical conditions (KL2G, Jochum et al. 2006); KE12, Mosbah et al. 1991; VG2, Jarosewich et al. 1980; experimental glasses N72, M35, M40 and M48, Shishkina et al. 2010 and MacQuarie glasses 47963 and 25603, Kamenetsky et al. 2000) with a large range of concentrations of volatile elements overlapping those of our samples. Maximum errors, based on reproducibility over 10 cycles of analyses, were less than 15% for CO_2 , 3% for Cl, 4% for S, and 5% for F and H_2O .

LA-ICP-MS analyses of trace elements

Trace element measurements on melt inclusions, embayments and glasses were carried out using a laser ablation system (193 nm Excimer Resonetics M-50E) associated with an inductively coupled plasma mass spectrometer (Agilent 7500 cs LA-ICPMS at the Laboratoire Magmas et Volcans, Clermont-Ferrand). Analysis were performed following classical procedures outlined in previous studies (e.g., Le Voyer et al. 2010; Rose-Koga et al. 2012). We used a pulse energy of about 3 mJ, a spot diameter between 15 and 33 μm and a laser pulse frequency of 2-3Hz, depending in the inclusion size to keep a fluence at sample surface of about 4 J/cm^{-2} . The background was measured for 30 to 40 seconds before ablation

and analysis time was approximately 100s. Data reduction was performed using the GLITTER software (www.es.mq.edu.au/GEMOC). This technique uses CaO (measured by EMP) as an internal standard. Reproducibility and accuracy of the analysis were constrained by systematic analysis of 2 standards BCR2-G and 612 at the beginning, in the middle and at the end of the session. Typical errors on the samples (1σ error of mean) is less than 10 % for all trace elements, except for Lu (<20%) and B (<40%).

Volatile diffusion modelling

Concentration profiles recorded in the embayments were fitted by a diffusion model similar to that of Ferguson et al. (2016). While the detail of the model and test results are explained in the **Supplementary**, a brief description is given here. An embayment profile was measured in a tube-like pocket of melt found within an olivine crystal. Because of its shape and incompatibility of volatile elements, such as H, C, S, Cl, and F, in olivine, diffusive exchange of elements was considered only possible at the mouth of the embayment. This justifies the use of diffusion solution in 1-D geometry. Naturally, the concentrations of volatile elements are systematically lower at the mouth of embayment. Having these geometrical constraints, we have developed a model with the following initial and boundary conditions for Fick's Second Law:

$$\frac{dC}{dt} = D(C_W) \frac{\partial^2 C}{\partial x^2}, \quad (1)$$

$$C = c_{initial}, x > 0, t = 0; \quad C = c_{initial}, x = \infty, t > 0,$$

$$C = f(t), x = 0,$$

where $D(C_W)$ is the diffusion coefficient of water in melt as a function of total dissolved water concentration; C , the concentration of diffusing species; t , time; and x , distance. Concentrations of volatile elements were set at a constant value $c_{initial}$, in the embayment, determined by the plateau concentration value inside the embayment. Concentration at the

interface was set by a function with respect to time, which determines H₂O-CO₂ solubility in magma along the linear isothermal decompression. In a magma ascending for eruption, and for fast diffusing species such as H₂O, a closed system degassing approximates the volatile element segregation better than an open system, as the latter requires complete segregation of magma from gas at every instant of degassing. We used SolEx (Witham et al. 2012) to determine the closed system, equilibrium degassing path for H₂O, CO₂, S, and Cl, and the $f(t)$ was determined from the SolEx result with a constant decompression rate (dP/dt). The diffusion equation was solved by Crank-Nicolson finite difference method for the 1-D geometry.

This diffusion calculation had therefore four parameters: (1) diffusion coefficient, (2) initial concentration, (3) concentration of a volatile element at $x=0$ which changes with time (*i.e.* decompression), (4) the mass of equilibrium bubble/fluid present at the moment of the magma ascent, and (5) the decompression rate. The first three parameters were independently determined with reasonable confidence from the experimental data, concentration profile, and magma temperature. The parameter (4) was unknowable, and it influenced the degassing path significantly (see the **Supplementary**). Here, we have tested its influence by varying from 0 to 1 wt % of pre-ascent volatile mass. By setting four parameters, the decompression rate was determined by minimizing the least-square residuals. While the method of Ferguson et al. (2016) minimizes X simultaneously for three volatile species (H₂O, CO₂, and S), it was not possible in our case. Diffusion profile of each volatile species were hence solved separately. R-based script, together with an example profile is given in **Appendix B**.

Olivine diffusion modelling

Chemical gradients (Fe–Mg) in olivines were modelled in one dimension with the DIPRA software program (Girona and Costa 2013). DIPRA is a finite difference code that solves Fick's second law of diffusion in order to model the one-dimensional diffusion of Fe – Mg (as forsterite, $Fo = Mg/(Mg + Fe)$, in mole fraction), Mn, Ni, and Ca in olivine. Diffusion anisotropy was accounted for by determining the olivine crystallographic axes using EBSD and the angles between the electron microprobe traverse and the three crystallographic axes. Temperature and pressure calculations are detailed below, oxygen fugacity was fixed at the Nickel-Nickel-Oxide buffer in all calculations. Initial conditions for sample Ph1Ol2E_P1 and Ph1Ol2E_P2 used a step function while all other calculations were performed using a homogeneous profile (see Fig. S6). The use of a step function for Ph1Ol2E_P1 and Ph1Ol2E_P2 was justified by the high and non-random residual that would result from using a homogeneous profile. Note that the effect of water on the Fe-Mg interdiffusion (e.g., Wang et al. 2004; Hier-Majumder et al. 2005) is not taken into consideration in the DIPRA calculations.

Post-entrapment crystallisation modelling

Corrections of major and trace element composition in olivine-hosted melt inclusions for post entrapment crystallisations (PEC) was performed. It consisted of dissolving increments of equilibrium olivine into the melt inclusion liquid until the Fe-Mg K_d reaches the equilibrium value. The equilibrium value for the Mg/Fe exchange coefficient between olivine and liquid, K_d , was determined by a model accounting for the influence H_2O and alkalies (Toplis 2005 with $T = 1145\text{ }^{\circ}C$ and $P = 1000\text{ bars}$), and the predicted $K_d = 0.25 \pm 0.01$. It should be noted that this value of K_d lower than the common value of 0.3 reflects that the Ambae melts were rich in Na_2O and K_2O . To test the effect of oxygen fugacity, Fe^{3+}/Fe_{tot} ratio was varied from 0.14 to 0.32 for the PEC calculation, reflecting the range observed for subduction zone

magmas, from FMQ-1 to FMQ+2; (e.g., Kress and Carmichael 1991; Bénard et al. 2018). With $\text{Fe}^{3+}/\text{Fe}_{\text{tot}} = 0.14$, we added successive increment of olivine back into the melt recalculating an observed Kd after each step and the process stopped when observed Kd = predicted Kd (0.25 ± 0.01 here). By doing this, the mean mass of olivine added was only of 1% (with a standard deviation of 0.2%, $n=57$) suggesting that very little PEC took place. At higher $\text{Fe}^{3+}/\text{Fe}_{\text{tot}} = 0.32$ (FMQ+2) or $\text{Fe}^{3+}/\text{Fe}_{\text{tot}} = 0.23$ (FMQ+1), the redox-corrected observed Kd is already too high compared to the “Toplis-predicted” Kd so that no olivine can be added to the melt. In the following, the corrected value reported are calculated for $\text{Fe}^{3+}/\text{Fe}_{\text{tot}} = 0.14$.

Estimation of microlite number density

Toramaru et al. (2008) developed a model to link microlite number density (N) in pyroclastic rocks to the rate of water exsolution which in turn can be linked to the rate of magma ascent. Microlite crystallization can happen as a consequence of cooling or decompression-induced degassing. In the case of decompression-induced degassing, a numerical simulation performed by Toramaru (2008) shows that the microlite number density is directly linked to the increase in liquidus temperature caused by degassing:

$$\frac{\partial T_L}{\partial t} = \left(\frac{N}{a}\right)^{2/3} \quad (2)$$

With the constant a mostly a function of melt chemistry and water concentration:

$$a = 3 \cdot 10^{15 + 0.345(C_{\text{SiO}_2} - 50) - 0.65C_{\text{H}_2\text{O}}} \quad (3)$$

where major element concentrations are in weight %.

The relationship between the increase in liquidus temperature T_L and the degassing rate is given by the derivative of the liquidus depression curve:

$$\frac{\partial T_L}{\partial t} = \frac{\partial T_L}{\partial C_{\text{H}_2\text{O}}} \cdot \frac{\partial C_{\text{H}_2\text{O}}}{\partial t} \quad (4)$$

296 The liquidus depression curve has been determined experimentally for olivine ($\frac{\partial T_L}{\partial C_{H_2O}} =$
 297 $40.4 - 5.94C_{H_2O} + 0.2283C_{H_2O}^2$; Médard and Grove 2008), and plagioclase ($\frac{\partial T_L}{\partial C_{H_2O}} =$
 298 $54.66e^{-0.29C_{H_2O}}$; Almeev et al. 2012) in basaltic melts. In the absence of an experimental
 299 calibration for clinopyroxene, we calculated ($\frac{\partial T_L}{\partial C_{H_2O}} \sim 26$) for the bulk scoria of phase 1 using
 300 the pMelts software (Ghiorso et al. 2002).
 301 The degassing can in turn be converted into a decompression rate using the derivative of the
 302 water solubility curve:

$$303 \quad \frac{\partial P_{H_2O}}{\partial t} = \frac{\partial P_{H_2O}}{\partial C_{H_2O}} \cdot \frac{\partial C_{H_2O}}{\partial t} \quad (5)$$

304 For our trachybasaltic bulk composition, we fit a quadratic function to the solubility data
 305 produced by the model of Moore et al. (1998) which gives $\frac{\partial P_{H_2O}}{\partial C_{H_2O}} = 182.3C_{H_2O}$
 306 In a steady state, the decompression rate at the microlite nucleation depth is linked to the
 307 ascent velocity (Toramaru et al. 2008):

$$308 \quad V_N = \frac{1}{\rho g} \cdot \frac{\partial P_{H_2O}}{\partial t} \quad (6)$$

309 where ρ is the magma density at the microlite nucleation depth.

310

311 RESULTS

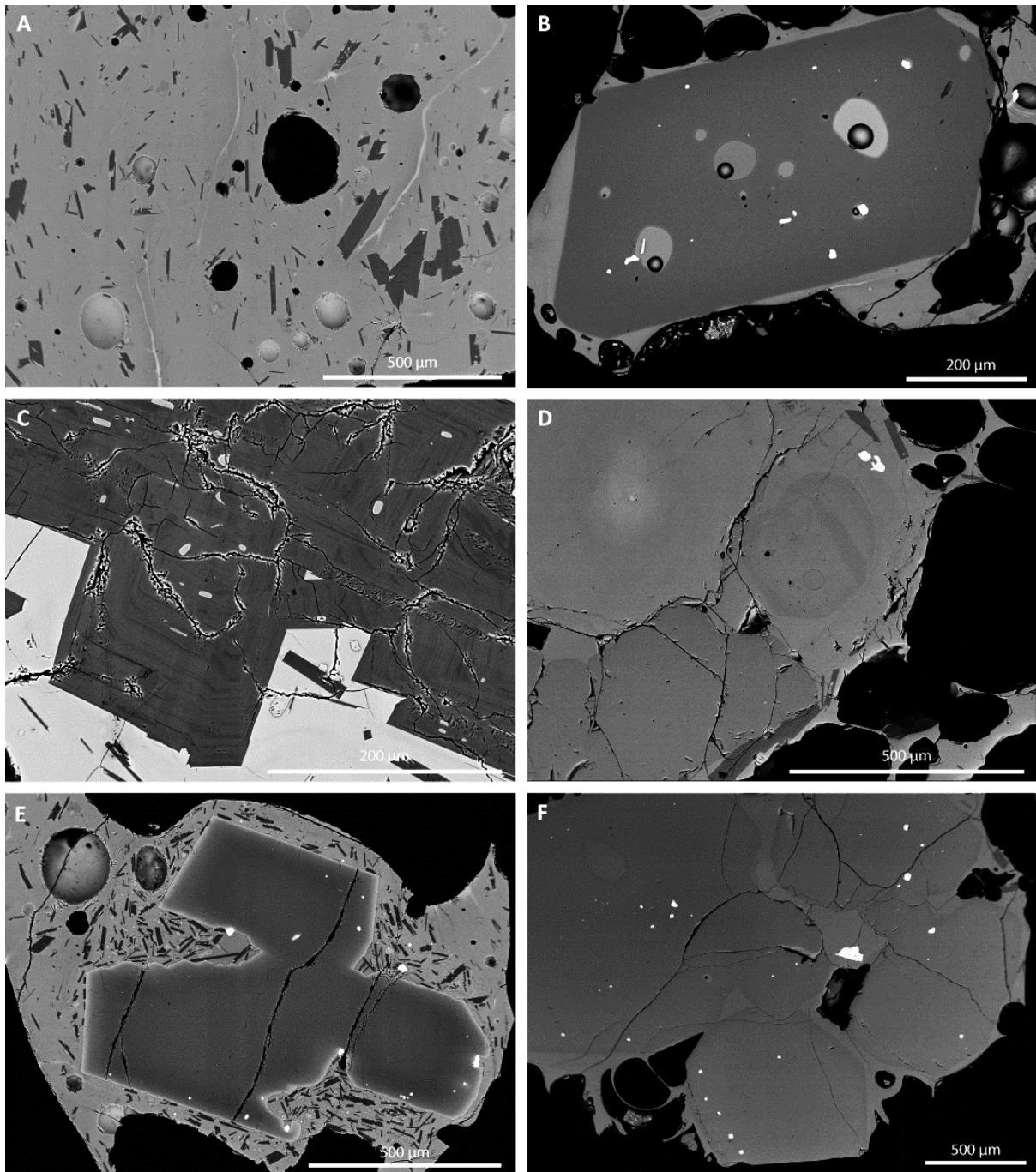
312 Mineralogy

313 The investigated scoriae are glassy, and typically have low crystallinity (Fig. 2A), with rare
 314 phenocrysts of plagioclase, olivine, and clinopyroxene, and microlites of the same phases.
 315 Small euhedral spinel microlites are also present, often associated with olivine phenocrysts
 316 (Fig. 2B). Most olivine phenocrysts have compositions ranging from Fo₈₅ to Fo₇₉, as expected
 317 for equilibrium with the bulk rock compositions (Table S1, S2, S3, S4, Fig. 3 and 4). Four
 318 types of zoning patterns can be identified in olivine phenocrysts (Fig. 3, 4B): (1) no zoning

319 (homogeneous crystals, Fig 3D, 4B); (2) normal zoning (Fig 3A, 3B, , 3C, 3F, 3G 4B); (3)
 320 reverse zoning (Fig 2B, 2F, 3H, 3I); and (4) complex zoning (Fig 2E). Homogeneous crystals
 321 are dominant in phase 1, and a few are still present in phase 2. Crystals with weak normal
 322 zoning (less than 1.5 Fo unit) are only present in phase 1 and again in phase 4. Reverse
 323 zoning is typical for most crystals from phase 2 and 3. Crystals with the stronger reverse
 324 zoning also show rounded edges indicative of dissolution. Reverse zoning can be interpreted
 325 as reflecting either magma mixing (Ruth et al. 2018) or reheating of a magma reservoir.
 326 Complex zoned phenocrysts have a weak internal reverse zoning very similar to the one
 327 observed in the previous type, followed by a strong normal zoning. These complex zoned
 328 phenocrysts are associated with a high microlite density, and present mostly in phase 1,
 329 although one crystal has been found in phase 2 and one in phase 3. These complex zoned
 330 phenocrysts are very similar to the olivine phenocrysts described from the April 2007
 331 eruption of Piton de la Fournaise (Reunion Island) where they have been interpreted as
 332 products of lateral magma transfer in the plumbing system (Albert et al. 2019). As for the
 333 Reunion samples, normal zoning patterns around melt inclusions are identical to normal
 334 zoning patterns near the crystal rims, indicating that the zoning is not due to magma mixing
 335 (Albert et al. 2019).

336 A significant fraction of the olivine phenocrysts (Fig. 4A) have core compositions more
 337 magnesian than Fo₈₄ and are thus not in equilibrium with any of the measured bulk rock
 338 compositions (Fig. 4). We interpret those crystals as antecrysts, formed from a batch of more
 339 primitive magma. A few small olivine microlites have compositions ranging from Fo₈₂ to
 340 Fo₇₄ (Table S3). Clinopyroxene phenocrysts are normally zoned, with Mg# ranging from 85
 341 to 78 (Table S5, calculated assuming all Fe as FeO). Oscillatory zoning is occasionally
 342 observed (Fig. 2D). Clinopyroxene microlites are small and difficult to analyse but have Mg#
 343 in the range 80-72. Plagioclase phenocrysts are euhedral, normally zoned and commonly

show oscillatory zoning (Fig.2C). Their composition varies from An₈₁ in the core to An₇₀ near the rims (Table S6). Euhedral plagioclase microlites have compositions varying from An₈₀ to An₅₈. Spinel compositions are highly variable: spinel included in olivine phenocrysts are chromites (41 wt% Cr₂O₃), whereas the rare spinel microlites tend to be richer in iron (58 wt% FeO+Fe₂O₃ on average, up to 76 wt%).



350 **Figure 2:** *A. BSE microphotograph of the typical glassy texture of the pyroclasts (phase 1),*
 351 *with plagioclase (micro-)phenocrysts and microlites and a few olivine and clinopyroxene*
 352 *microlites. B. BSE microphotograph of a normally zoned olivine phenocryst with large glassy*
 353 *melt inclusions associated with spinel crystals and retraction bubbles (phase 2). C. Enhanced*
 354 *BSE microphotograph of oscillatory zoning in a plagioclase phenocryst that contain small*
 355 *melt-inclusions (phase 1). D. BSE microphotograph of clinopyroxene phenocrysts with*
 356 *complex oscillatory zoning and melt inclusions (phase 2). E. BSE microphotograph of an*
 357 *olivine phenocryst showing complex zoning: slight inverse zoning around the core, followed*
 358 *by clean normal zoning on the edges (phase 1). The phenocryst also contains spinel*
 359 *inclusions and is surrounded by an increased microlite density. F. Enhanced BSE*
 360 *microphotograph of a group of olivine phenocrysts with a slight reverse zoning around the*
 361 *edges (phase 2). The phenocrysts contain spinel and melt inclusions.*

362

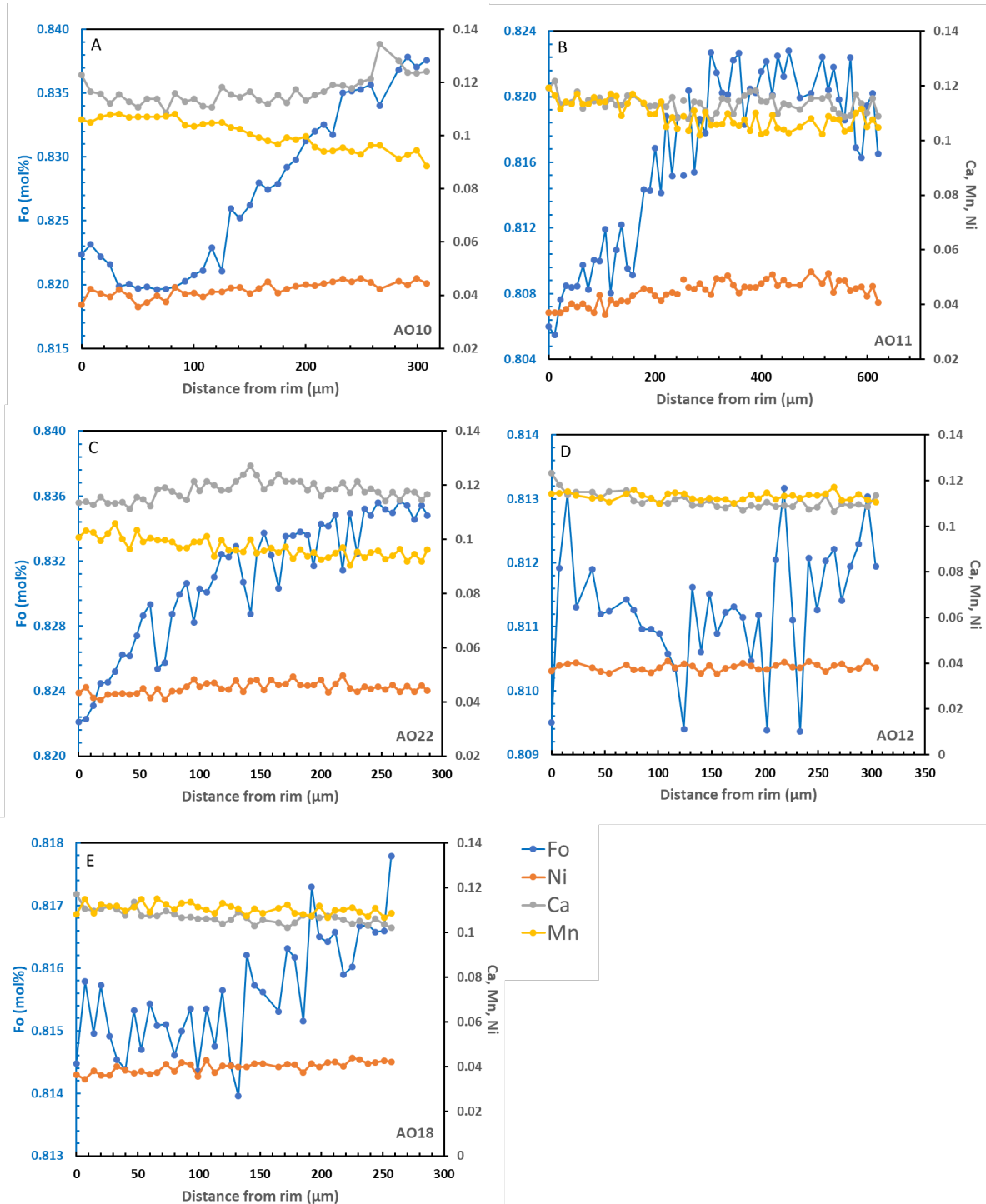


Figure 3: Rim to core electron microprobe %Fo (mol%) ($= 100 \times \text{Mg} / [\text{Fe} + \text{Mg}]$), Ca, Mn, and Ni traverses in olivine crystals from phase 1 (A to E) and phase 3 (F to J). AO10, AO11, AO22 and AO18 show predominantly normal zoning with AO10 showing an alternation from normal to reverse zoning close to the rim. AO12 shows no resolvable zoning pattern. Ol5ph3, Ol2ph3 and Ol1ph3 show normal zoning while Ol7ph3 and Ol3ph3 shows reverse zoning.

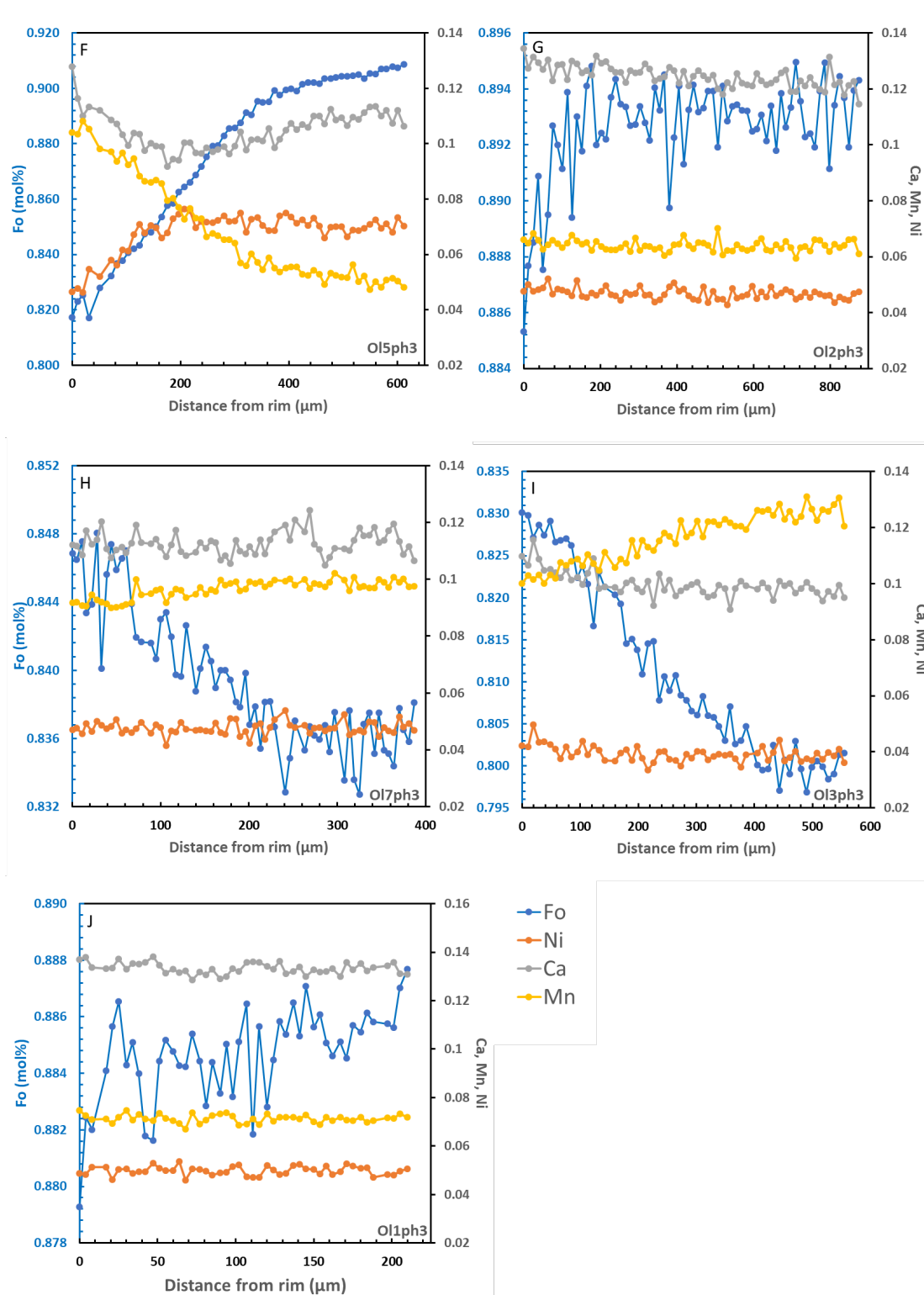


Figure 3con: Rim to core electron microprobe %Fo (mol%) ($= 100 \times \text{Mg} / [\text{Fe} + \text{Mg}]$), Ca, Mn, and Ni traverses in olivine crystals from phase 1 (A to E) and phase 3 (F to J). AO10, AO11, AO22 and AO18 show predominantly normal zoning with AO10 showing an alternation from normal to reverse zoning close to the rim. AO12 shows no resolvable zoning pattern. Ol5ph3, Ol2ph3 and Ol1ph3 show normal zoning while Ol7ph3 and Ol3ph3 shows reverse zoning.

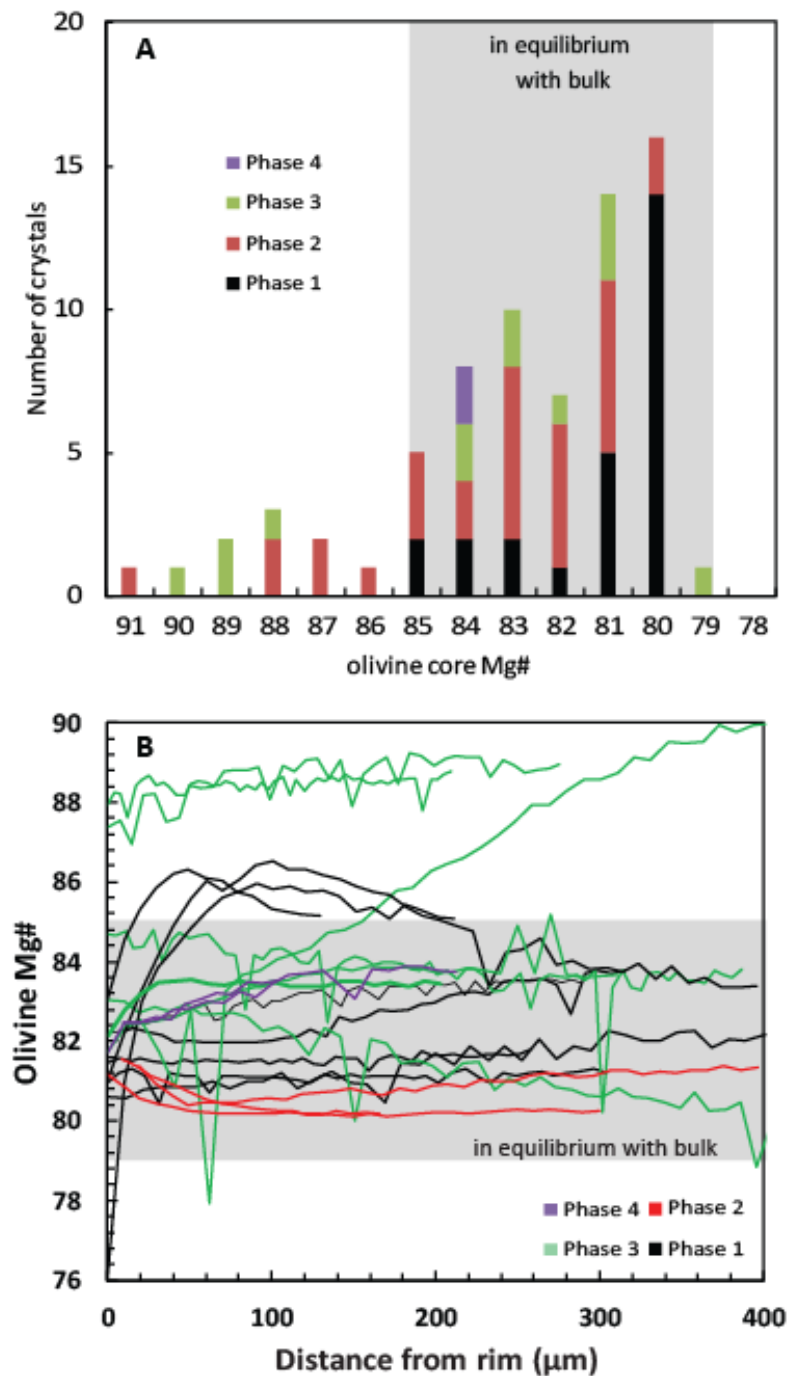


Figure 4: A. Frequency diagram of the Fo content in olivine cores from phase one, two and three. The range of Fo composition in equilibrium with bulk is calculated using measured bulk rock compositions and a K_d of 0.3. Note that a significant fraction of phase 2 and 3 olivines have core composition more magnesian-rich than expected for equilibrium with the bulk. B. Fo profiles of olivine phenocrysts from phases 1, 2, 3 and 4.

Major elements

The PEC corrected composition of melt inclusions, embayments and matrix glasses in major and trace element is given in [Tables S1 and S7](#). Uncorrected compositions are given in [Tables S2 and S8](#).

Samples for the 2017-2018 eruption are the most differentiated volcanic rocks on the island, with the exception of one trachyandesite analysis reported by Eggins (1993). The bulk rock compositions ([Table S1](#)) plot in the trachybasalt and basaltic trachyandesite fields of the TAS diagram ([Fig. 5](#)), as an extension of previously published analyses for Ambae island (Warden 1970; Gorton 1977; Eggins 1993). Previously reported whole rocks (and melt inclusion) compositions from older (undated) deposits on Ambae are more primitive, mostly tholeiitic and alkali basalts. The younger rocks of Ambae belong to the low-Ti suite of Eggins (1993). This suite starts with voluminous magnesian olivine- and clinopyroxene-phyric basalts (“ankaramites”) that form the largest part of the island. More differentiated plagioclase-phyric basalts are confined to stream courses and represent the youngest eruptive products of the volcano, including lava flows from the latest significant eruption of the volcano that produced the N’dui N’dui flow about 350 years ago (Warden 1970). Products of the 2017-2018 eruptive phase are slightly more differentiated than these plagioclase-phyric basalts, and likely represent further evolution along a fractional crystallization trend ([Fig. 5](#)). In total alkali-silica space, most inclusions plot as alkali basalt, trachy-basalt or basaltic trachyandesite while matrix glasses are basaltic trachy-andesite in composition. In details, melt inclusions, matrix glasses and bulk rock compositions from phase 1 tend to be slightly more evolved than the ones of phase 2, 3 and 4.

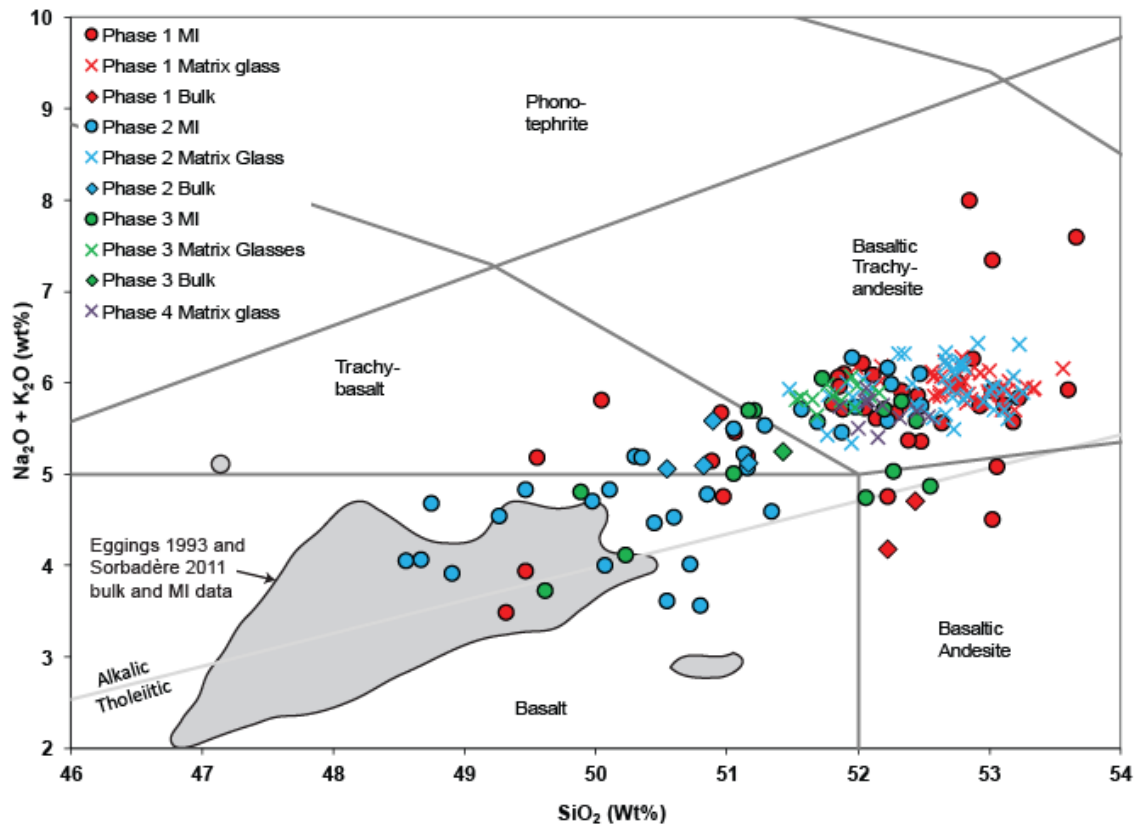


Figure 5: Total alkalis versus silica. Melt inclusions, matrix glasses and bulk rock compositions (all normalized) are reported for the first three phases of eruptive activity between 2017 and 2018 and matrix glass compositions are reported for phase 4. Bulk rock and melt inclusion composition data from older (undated) deposits are from Eggings (1993) and Sorbadere et al. (2011).

Melt inclusions have Mg# ranging from 31.7 to 59.6 (assuming all iron is FeO). Host olivine crystals have compositions ranging from Fo₇₉ to Fo₉₁. As expected for near equilibrium melt inclusion-host pairs, there is a clear correlation between the olivine and melt inclusions compositions (Fig. 6A). Interestingly, olivine erupted during the first phase of activity (September 2017) tend to have lower Forsterite content (Fo < 84%) and to contain inclusions with low Mg# (Mg# < 50) while olivine erupted during the subsequent two phases of activity (November 2017 and March 2018) have a larger range of Forsterite content (Fo up to 91%) and contain inclusions with a larger range of Mg# (Mg# up to 60). In addition, melt inclusions from the first phase tend to be richer in Si, Al, Fe, Mn and K, and poorer in Ca and Mg compared to melt inclusions from the second and third phases (Fig. 6). This is also seen

with the interstitial glasses which show a small increase in MgO, Mg# and CaO/Al₂O₃, and a small decrease in Na₂O and K₂O between phases 1, 2, 3 and 4.

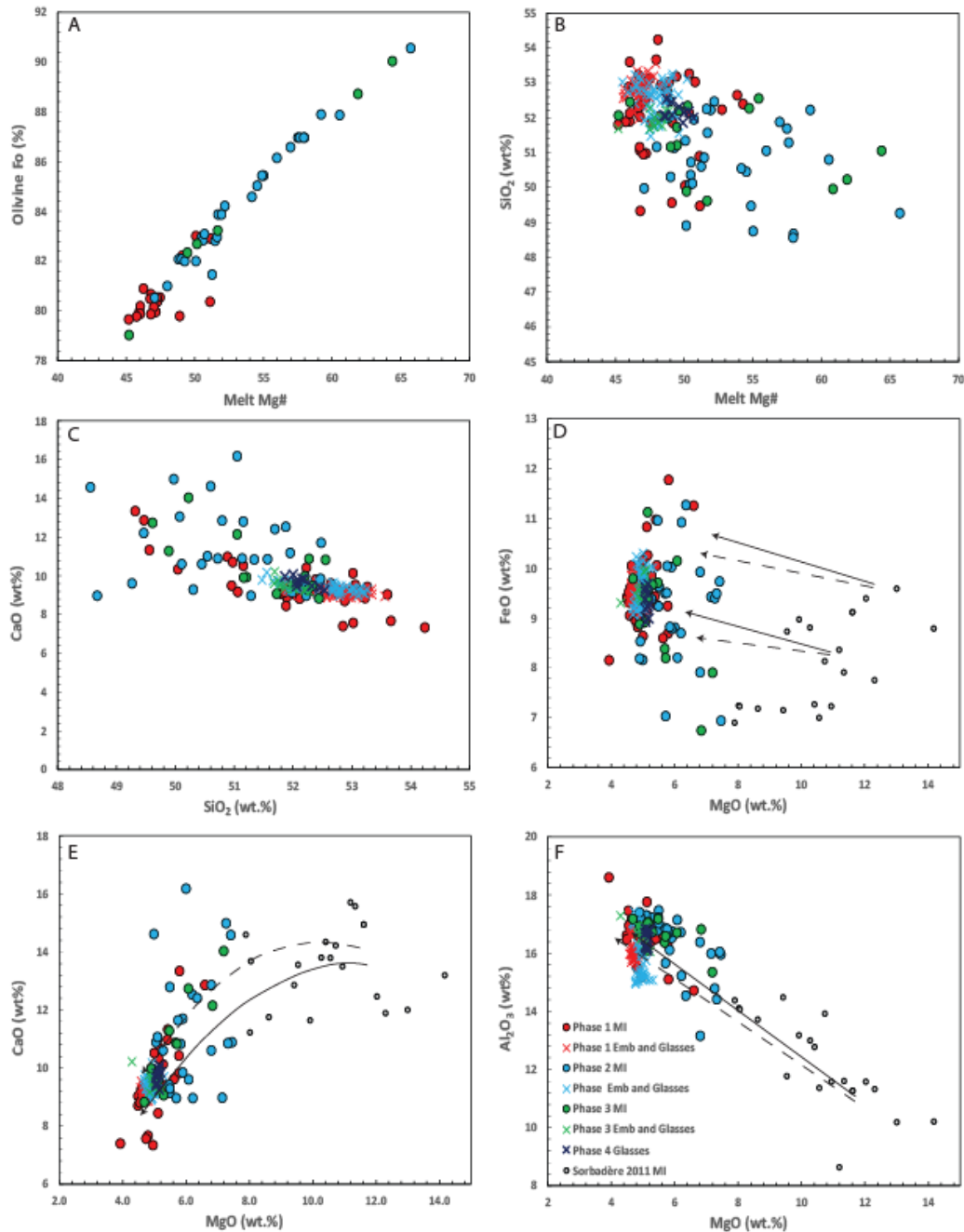
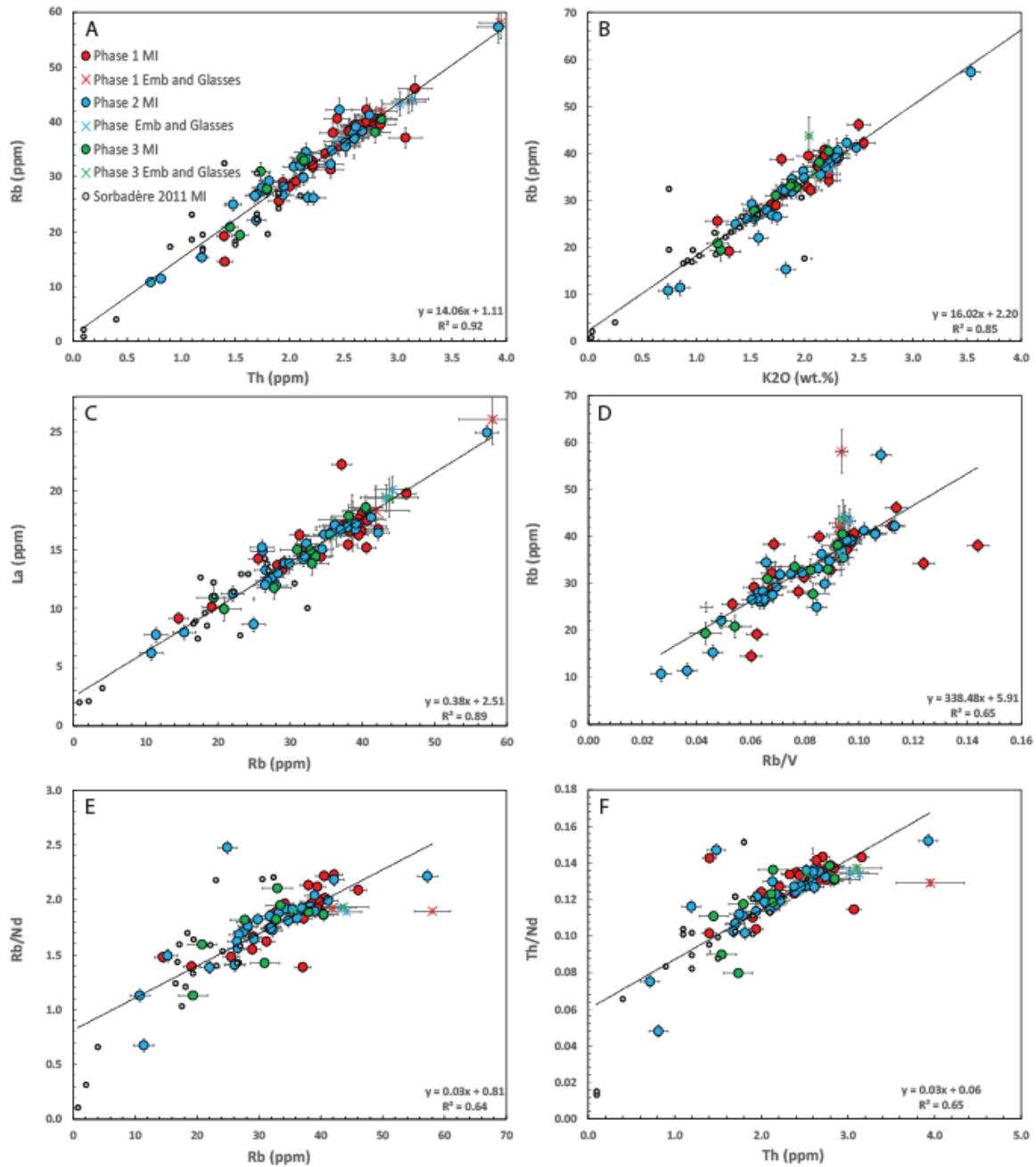


Figure 6: Major element compositions of olivine and pyroxene-hosted melt inclusions (olivine-hosted inclusions are PEC-corrected). **A.** Olivine Forsterite content vs melt inclusion Mg#. **B.** SiO₂ vs melt Mg#. **C.** CaO vs SiO₂. **D.** FeO_T vs MgO. **E.** CaO vs MgO. **F.** Al₂O₃ vs MgO. More primitive melt inclusions data from Ambae are also shown, together with MELTS-calculated liquid lines of descent at 200 MPa (dotted arrow) and 500 MPa (solid arrow) (Sorbadere et al. 2011).

435 **Trace elements**

436 Trace and rare earth elements concentrations in melt inclusions and matrix glasses show no
437 systematic differences between the first three eruptive phases (Fig. S4 and 7). The rare earth
438 element distribution, normalised to chondrite (Sun and McDonough 1989), is typical of arc
439 magmas, being depleted in heavy rare earth elements. The positive correlations, passing close
440 to the origin, between incompatible element pairs (Fig. 7A and 7B) show that melt inclusion
441 compositions of the three eruption phases plot on a single trend of magma differentiation,
442 suggesting they belong to a co-genetic magmatic series.

443



444

445 **Figure 7:** Trace element variation diagrams from melt inclusions and matrix glasses from the
 446 2017-2018 eruption (this study; note that olivine-hosted MI are PEC-corrected) and from
 447 older deposits (Sorbadere et al. 2011). **A.** Plot of Rb versus Th. **B.** Plot of Rb versus K₂O. **C.**
 448 Plot of La versus Rb. **D.** Plot of Rb versus Rb/V. **E.** Plot of Rb/Nd versus Rb. **F.** Plot of Th/Nd
 449 versus Th. Linear regression lines through all data are presented on each plot. The data from
 450 **A** and **B** define linear trends that passes through the origin as expected for fractional
 451 crystallization while data from **C** define a linear correlation that does not pass through the
 452 origin, potentially indicative of mixing between two components. Data from **D** is not well
 453 explained by a linear relationship ($R^2=0.65$) as expected for melts related by either mixing or
 454 fractional crystallisation. Data from **E** and **F** are not well explained by linear relationships
 455 ($R^2=0.64$ and 0.65 respectively), potentially indicating mixing between two components. On
 456 diagram **E** and **F** fractional crystallisation should result in no variations in Rb/Nd and Th/Nd
 457 as a function of Rb and Th concentrations respectively.

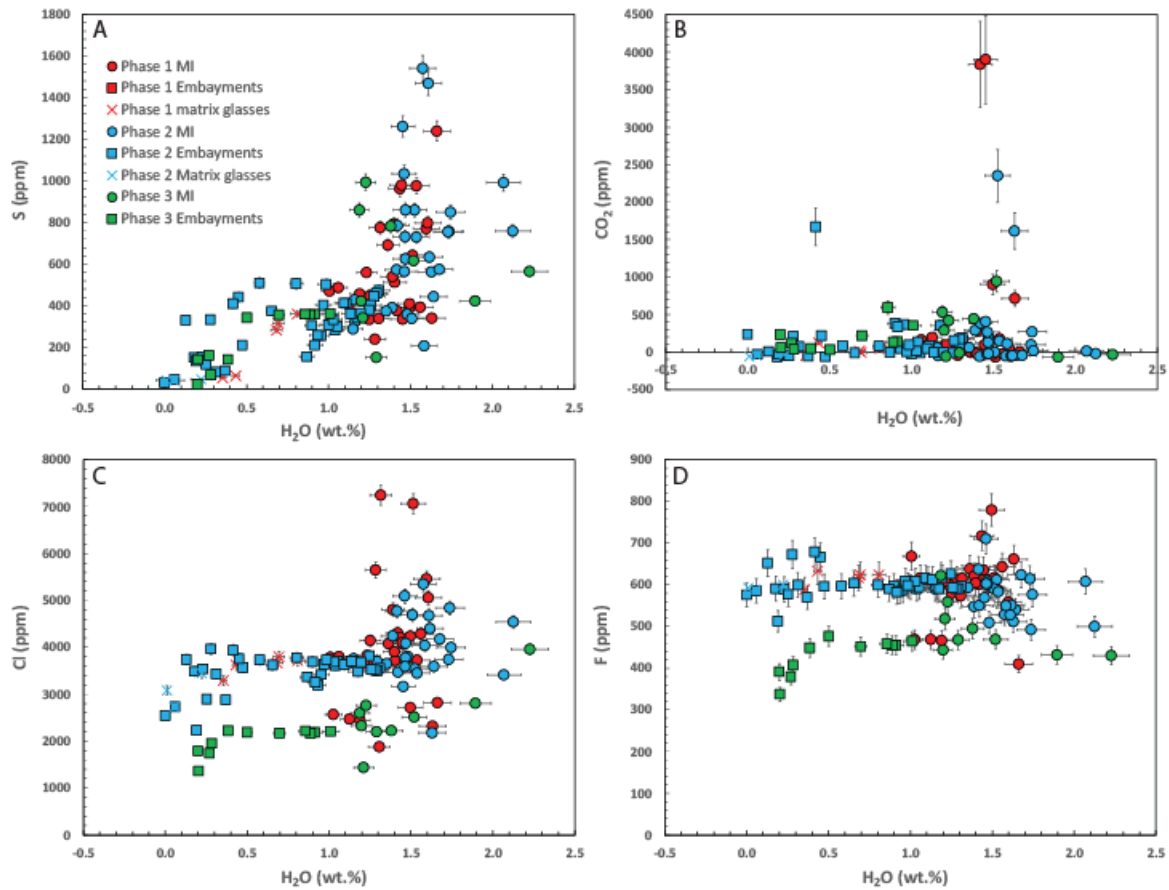
Volatiles

In melt inclusions and matrix glasses, chlorine ranges from 1435 to 7243 ppm, fluorine from 410 to 779 ppm, water from 0 to 2.2 wt.%, CO₂ from 0 to 3899 ppm and sulfur from 30 to 1541 ppm. Both H₂O and CO₂ in melt inclusions have the potential to be altered during post-entrapment processes such as diffusive H⁺ exchanges between the MI and the melt through the olivine (e.g., Massare et al. 2002; Danyushevsky et al. 2002; Chen et al. 2011; Gaetani et al. 2012; Bucholz et al. 2013; Portnyagin et al. 2019) or CO₂ diffusion between the MI and its shrinkage bubble (e.g., Anderson and Brown, 1993; Steele-Macinnis *et al.*, 2011; Wallace *et al.*, 2015). While there is no evidence for either process in the investigated sample set, neither can be discarded. A conservative approach is hence to consider the measured H₂O and CO₂ values as minima. Melt inclusions from phase one tend to be richer in chlorine and fluorine and poorer in water than melt inclusions from phase two and three, with phase three being the poorest in chlorine and fluorine. No systematic differences are observed in the CO₂ and sulfur content of the melt inclusions between the three phases.

Neither chlorine, fluorine nor CO₂ correlates with Nb despite being theoretically similarly incompatible (Fig. S5A, B and D). Likewise, water does not correlate with Ce despite being theoretically similarly incompatible (Fig. S5C). Finally, the sulfur content does not correlate with FeO content suggesting that sulphide saturation did not take place during the magmatic evolution recorded by melt inclusions. Together all these observations suggest that the volatile contents of melt inclusions and matrix glasses are not determined by differentiation and must be more reflective of degassing processes even though no clear open- nor closed-system degassing trend is shown by the data. Little correlations are observed between volatile element excepted for water and sulfur (Fig. 8). The majority of the sixty-eight melt inclusions

482 analysed have low CO₂ content, between detection limit (<50) and 500 ppm, while only five
 483 inclusions have CO₂ content above 1000 ppm.

484



485

486 **Figure 8:** Volatile elements abundance in melt inclusions, embayments and matrix glasses
 487 from Ambae.

488

Geothermobarometry and volatile saturation pressure

The presence of euhedral microlites of plagioclase, clinopyroxene, and olivine indicates that interstitial glasses are saturated with all three phases. Interstitial glasses are in equilibrium with olivine microlite rims ($\text{Fe/Mg } K_d = 0.27\text{--}0.29$; Roeder and Emslie 1970), clinopyroxene microlite rims ($\text{Fe/Mg } K_d = 0.19\text{--}0.33$; Putirka 2008), and plagioclase microlite rims ($\text{Ab/An } K_d = 0.17\text{--}0.32$, Putirka 2008). Pre-eruptive temperatures were thus calculated using a variety of mineral-saturation models (Table 1). At least 15 glass compositions were analysed for each phase; for a given phase, all analyses are within analytical error of each other, so only the average value was used for thermobarometry calculations. All these estimates are within error of each other and are consistent with a pre-eruptive temperature of $1145 \pm 15^\circ\text{C}$, with possibly a small increase of 8°C between phase 1 and phase 4. These calculations were performed without taking into account the influence of water on mineral-saturation temperatures. Water concentrations in matrix glasses vary between 0.01 and 0.81 wt% (Table S9). Addition of a maximum of 0.81 wt% H_2O results in a 30°C decrease (Médard and Grove 2008), i.e. a minimum pre-eruptive temperature of $1115 \pm 15^\circ\text{C}$.

Since interstitial glasses are saturated with olivine, plagioclase and clinopyroxene, it is possible to estimate the equilibration pressure from the pressure-dependence of the olivine-plagioclase-clinopyroxene cotectic (Yang et al. 1996). The calibration of Yang et al. (1996) as implemented by Kelley and Barton (2008) gives pressure estimates of $50 \pm 110\text{ MPa}$, indicating that the latest stages of crystallization occurred in the upper 6 km of the crust. Given that the bulk rock compositions are also saturated with olivine, plagioclase and clinopyroxene, it is possible to estimate a pressure where the melt equivalent to the bulk rock was extracted using the same method. The average pressure at which fractional crystallization from a more mafic magma produces the bulk rock composition is $133 \pm 110\text{ MPa}$. The model

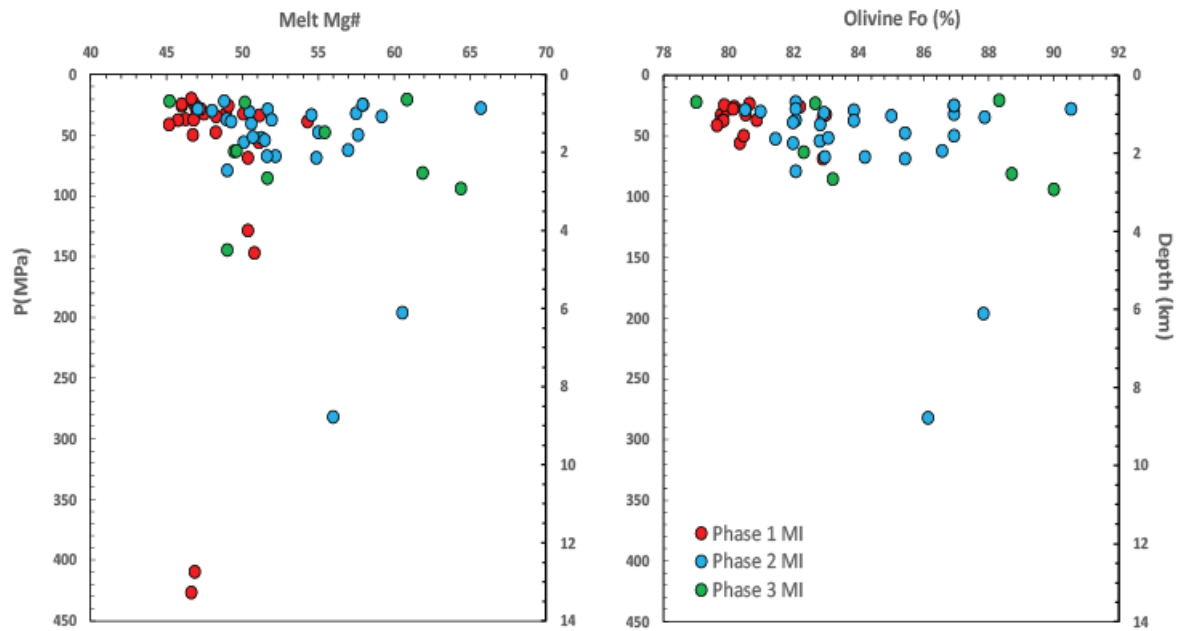
of Kelley and Barton (2008) has been developed for water-poor Icelandic basalts. Addition of water stabilizes clinopyroxene, resulting in pressure estimates that are too high (Husen et al. 2016). If interstitial glasses contained water when they last equilibrated with the phenocrysts, the equilibration pressure would be lower, i.e. within error of the surface.

Crystallization pressures can also be estimated using the pressure-sensitive incorporation of the jadeite component in clinopyroxene (e.g., Putirka, 2008). Barometers have been calibrated either purely from the jadeite component of the clinopyroxene, or from jadeite partitioning between clinopyroxene and melt. This equation is very sensitive to temperature and water concentration in melts, but less sensitive to the exact melt composition. We simultaneously solved for pressure and temperature using equation (30) and (33) of Putirka (2008) to minimize temperature effects. We initially paired microlite and phenocryst rims compositions with matrix glass compositions, phenocryst core compositions with bulk rock compositions, and melt inclusion compositions with their host phenocrysts, and only kept clinopyroxene analyses for which $\text{Fe/Mg } K_d = 0.28 \pm 0.08$ (Putirka 2008). Phenocrysts cores are, however, rarely in equilibrium with the bulk composition, and more commonly in near-equilibrium with the matrix glass, so we paired phenocryst core compositions with matrix glasses. This approximation could produce a ~ 30 MPa error on pressure, much smaller than the accuracy of the barometer. Water concentration in glasses was assumed to be 1.5 wt%, the average water content in melt inclusions, except for melt inclusions in which water concentrations have been analysed. Equilibrium between 11 melt inclusions and their host clinopyroxenes gives pressures between 43 and 326 MPa, with an average pressure of 184 ± 75 MPa at an average temperature of 1089 ± 11 °C. The temperature is lower than the temperature estimated from glass thermometry, since it is calculated with a higher water content. Identical pressures and temperatures were obtained from 26 analyses of phenocryst

cores (170 ± 86 MPa at 1091 ± 13 °C) and 28 analyses of phenocryst and microlite rims (188 ± 73 MPa at 1093 ± 12 °C), with a global average of all pressure calculations of 181 ± 78 MPa. Individual pressures vary between 13 and 328 MPa. Interestingly, the use of a clinopyroxene-only barometer (equation 32b of Putirka et al. 2008) produces similar results, with an average of 155 ± 65 MPa. Uncertainties due to barometer calibrations are quite large, with a standard error of estimate of 160 MPa for equation (30) of Putirka (2008). The final pressure estimates of 180 ± 160 MPa indicates that crystallization happened between 1 and 12 km, assuming a 2900 kg.m^{-3} density for the basaltic crust. These values are consistent with estimates from melt inclusion volatile elements concentrations, and within error of estimates based on cotectic melt compositions using bulk rock compositions. Pressure estimated from cotectic melt compositions of interstitial glasses are lower and register the latest stages of crystallization, whereas mineral compositions can register earlier, possibly deeper, stages.

Melt inclusions entrapment pressures were calculated using the model of Iacono-Marziano et al., (2012) for $\text{H}_2\text{O}-\text{CO}_2$ saturation pressure. They yield entrapment pressures between 20 and 427 MPa (Fig. 9), consistent with pressures of 43 to 326 MPa estimated from clinopyroxene/melt barometry. Given the above-reported error on the volatile content determination and the error on the model, the results can conservatively be taken to be accurate at $\pm 20\%$. Yet, as discussed previously, H_2O and CO_2 values are to be considered as minimum values due to possible diffusion of both species out of the MI. The entrapment pressures are therefore also to be taken as minimum pressure estimates. The two deepest inclusions are hosted in pyroxenes and their entrapment pressure estimates from volatile saturation (410 and 427 MPa) is fairly consistent, although higher, than their host pyroxene calculated crystallisation pressure (277 and 302 ± 160 MPa). Most other pyroxene-hosted inclusions record volatile saturation pressure lower than their host pyroxene calculated

564 crystallisation pressure yet within the standard error. Saturation pressures are also consistent
 565 with the calculated cotectic pressures, and point to a shallow magma storage system
 566 somewhere between 0.5 and 3 km depth for most of the crystals.

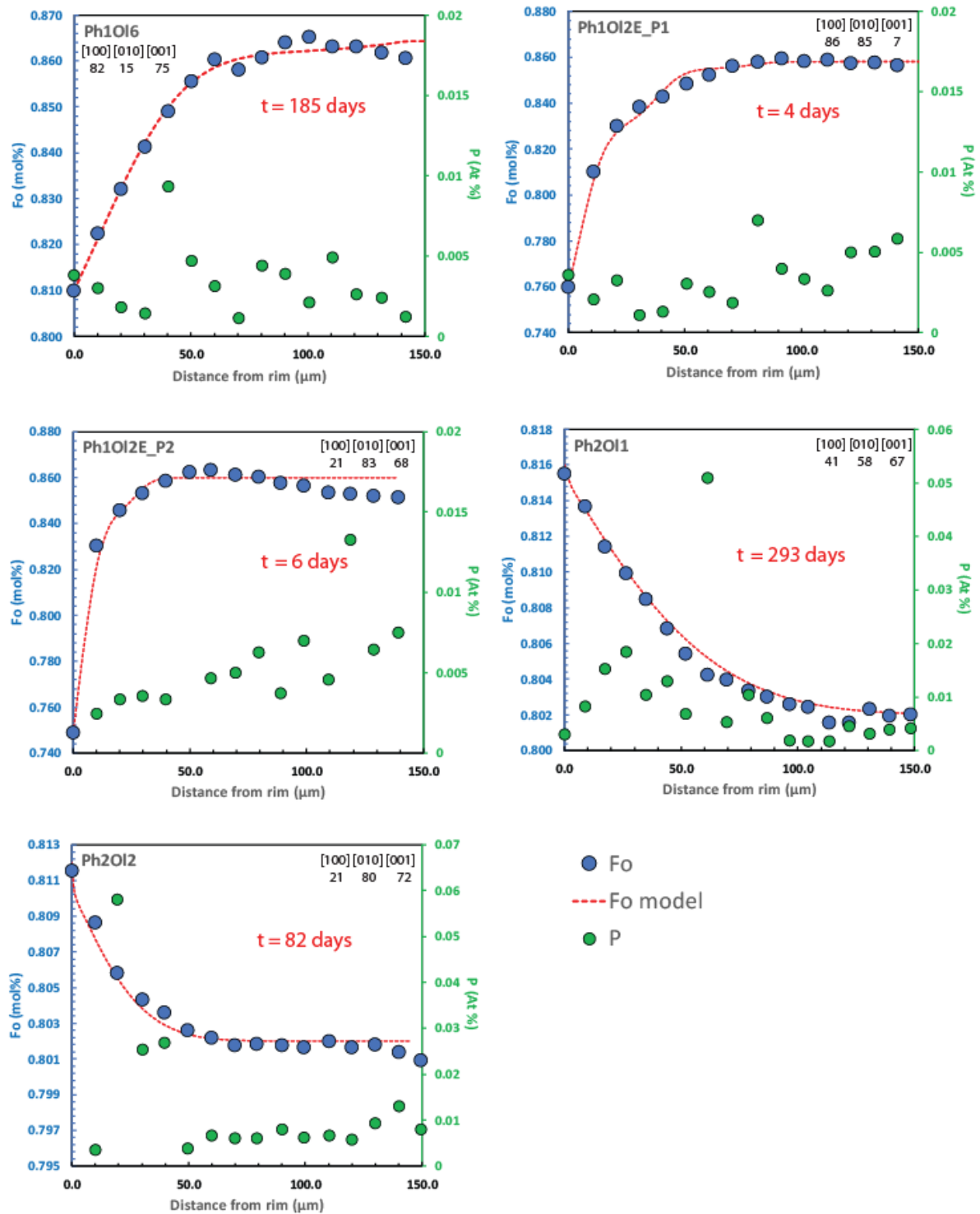


567

568 **Figure 9:** Melt inclusions entrapment pressure calculated from H_2O - CO_2 saturation
 569 pressure, compared to melt Mg# and host olivine forsterite content. Note that some of the MI
 570 are hosted in pyroxenes, hence not plotted on the right-hand side diagram.
 571

Residence time

We modelled the %Fo concentrations of five profiles obtained from two olivine from phase 1 and two olivine from phase 2 (Fig. 10). The diffusion model parameters and results are provided in Table 2. For normally zoned phase 1 olivine we determined diffusion timescales ranging from 4 to 185 days. For reversely zoned olivine from phase 2 we determined diffusion timescales ranging from 82 to 293 days (keeping in mind that phase 2 occurred about two months after phase 1). While our dataset is limited, it does already highlight a variety of diffusion timescales recorded by olivine phenocryst/antecryst. These results suggest that the shallow magma chamber has been periodically recharged over the days to year preceding the eruption. The latest recorded recharge event potentially occurred only a few days prior to the eruption while the oldest recorded event occurred about a year before the eruption.



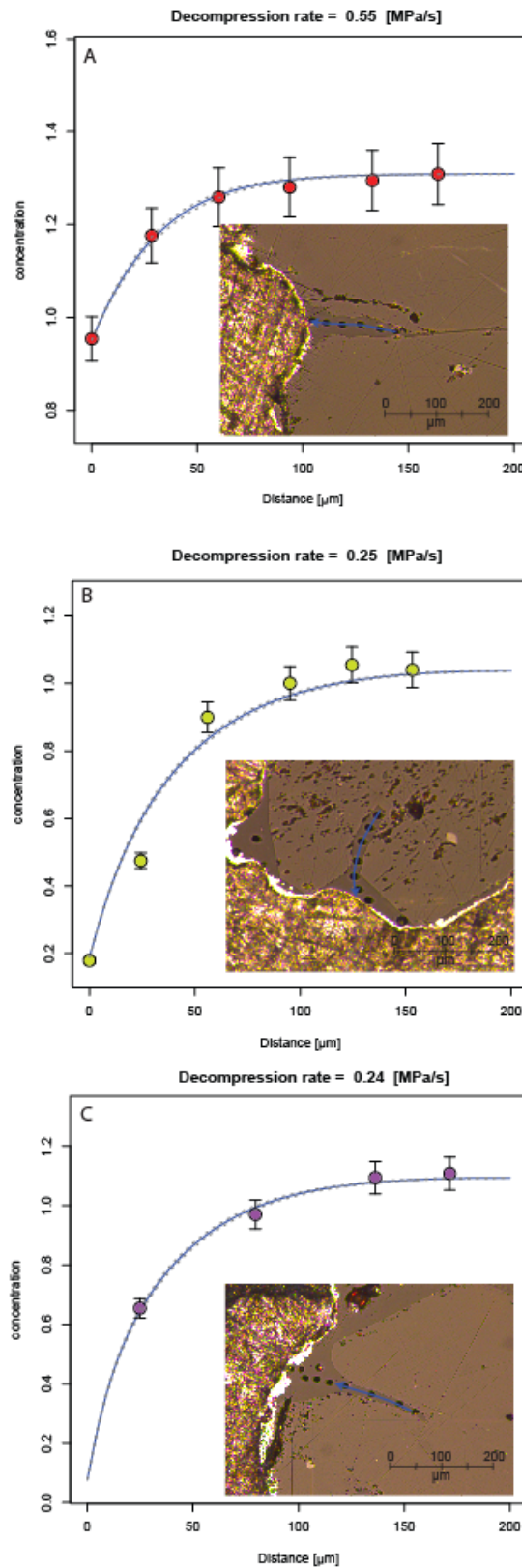
584

585 **Figure 10:** Results of diffusion modelling compared to %Fo concentrations profiles
586 (measured by EMPA) in two olivine crystals from phase 1 and two olivine crystals from
587 phase 2 of the 2017-2018 Ambae eruptions. Modelling was performed using the DIPRA
588 software (Girona and Costa 2013). Phosphorus profiles are shown and do not correlate with
589 %Fo content as expected for Fe-Mg zoning resulting from diffusion although the error on
590 phosphorus measurement ($3\sigma = \pm 0.002$) could partly mask any correlation. Initial conditions
591 are shown in Fig. S6.
592

Ascent rates

Volatile diffusion modelling was performed to match the measured H₂O concentration profiles (obtained by SIMS) along three melt embayments from phase two of the 2017-2018 eruption (note that the olivine used here are different from the ones used for Fe-Mg diffusion modelling in the previous section). Best fit solutions of the diffusion models are shown in **Fig. 11** with initial conditions and model parameters given in **Table 3**. These results give decompression rates of 0.24 ± 0.04 , 0.25 ± 0.06 and 0.55 ± 0.20 MPa/s which corresponds to ascent rates in the order of 33-75 km/h, assuming a unique melt/rock density of 2665 kg.m^{-3} for the trachybasaltic magma (Lange and Carmichael 1990). Concentrations of H₂O at the edge of the profiles are 0.06, 0.18, and 0.95 wt% indicating the degassing process can be quenched at about 4 MPa.

Diffusion modelling was also performed to reproduce S and Cl concentration profiles (obtained by SIMS). However, the measured profiles showed S and Cl concentrations at the edge of the embayments and in the surrounding matrix glass, that are much higher than expected for equilibrium at 0.1 MPa (i.e., ~50 to 200 ppm S and > 3400 ppm Cl remaining). This suggests that decompression rates were too high for S and Cl to maintain equilibrium during degassing, resulting in oversaturation of these species in the melt, as shown experimentally for CO₂ (Pichavant et al. 2013).



613

614 **Figure 11:** Results of diffusion modelling compared to H_2O concentration (measured by
 615 SIMS) in three melt embayments from phase two of the 2017-2018 Ambae eruptions. Insets
 616 show microphotograph of each embayment with blue arrows marking the location of the
 617 SIMS analyses.

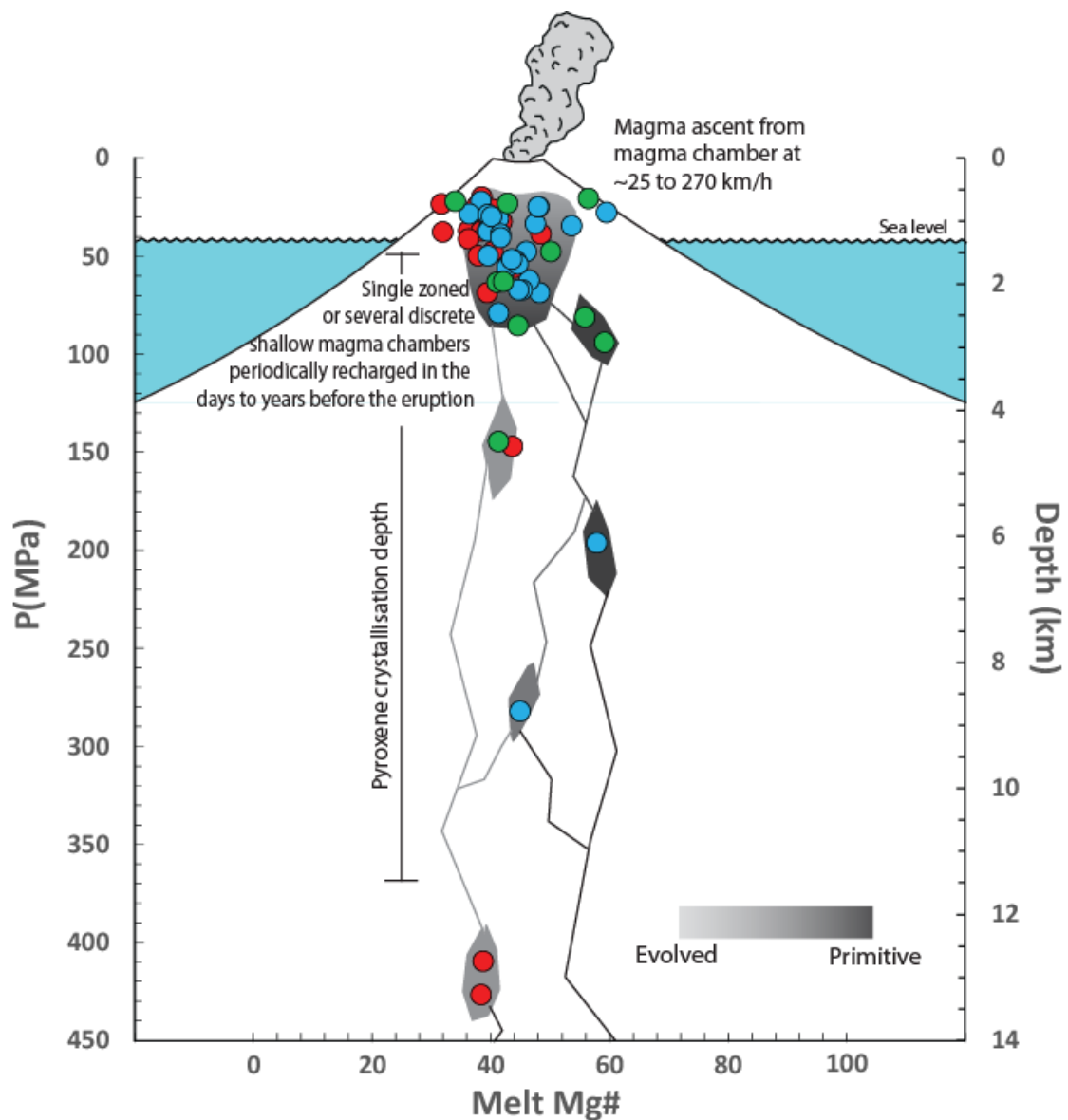
Ascent rates can also be estimated from the microlite density (Toramaru et al. 2008). The number and nature of microlites is extremely similar between samples of phases 1, 2 and 3 (Fig. S7). A higher number of microlites is observed in ashes from phase 4, but this is mainly due to the abundance of very small microlites that are not observed in other phases. These microlites might have crystallized during eruption. 2D microlite number densities and phase proportions were estimated using BSE images. This allowed us to compute average microlite sizes using the method of Hammer et al. (1999). To calculate 3D microlite number densities, we then divided the 2D microlite number densities by the average microlite size. For samples from phase 1, average microlite number densities averaged on seven BSE images are $3.8 \cdot 10^{13} \text{ m}^{-3}$ for plagioclase, $9.7 \cdot 10^{12} \text{ m}^{-3}$ for clinopyroxene, and $1.8 \cdot 10^{12} \text{ m}^{-3}$ for olivine. Given the small number of crystals (2 to 10 per image), the olivine MND has a very large uncertainty, and will not be taken into account. Using a maximum volatile content at microlite nucleation of 2.23 wt% H_2O (highest measurement in melt inclusions) results in decompression rates of $1.6 \cdot 10^5 \text{ Pa.s}^{-1}$ and $1.1 \cdot 10^5 \text{ Pa.s}^{-1}$ calculated from plagioclase and clinopyroxene, respectively. The density of the trachybasaltic magma is 2665 kg.m^{-3} (Lange and Carmichael 1990), producing ascent rates of $4\text{-}6 \text{ m.s}^{-1}$, or $15\text{-}23 \text{ km.h}^{-1}$, consistent with values determined from volatile diffusion.

DISCUSSION

The Ambae plumbing system

Clinopyroxene crystallisation pressures and melt inclusions entrapment pressures both suggest that the magma involved in the 2017-2018 Ambae eruption had been crystallising pyroxene and olivine at depths of 0.5 to 14 km minimum. Most melt inclusions, however, record low entrapment pressures corresponding to depth of 0.5 to 3 km. This suggest that most of the erupted magma originated from a shallow magma chamber located at sea level

644 (Fig. 12). There is no relationship between the melt inclusions entrapment pressures and
 645 markers of magmatic differentiation (Fig. 9), suggesting that the plumbing system is not
 646 arranged in a succession of progressively shallower and more evolved magma chambers.
 647 Instead, multiple chambers of variously differentiated magma must occupy similar depth
 648 levels, following horizontally separated pathways (Fig. 12).
 649



650

651 **Figure 12:** Schematic representation of the Ambae plumbing system on which melt inclusions
 652 entrapment pressure and Mg# are overlaid. Magmas at different stages of differentiation
 653 migrate through the crust and mix via episodic recharge events in a shallow magma chamber
 654 located at sea level. During the eruption, evolved magma is emptied preferentially towards

the beginning while more primitive magma is erupted during later phases. Note that there is no relationship between Mg# and horizontal coordinates of the melts.

The 2017-2018 emptying of Ambae's plumbing system

The 2017-2018 eruption started with phreatic activity and the creation of a small pyroclastic cone at the lake centre several days before the first appearance of phase one lava at the surface (Bani et al., in prep). A phreatic or phreatomagmatic start to a magmatic eruption is a commonly observed occurrence at volcanoes with previously-closed conduits, examples include the 1955 Bezymianny (Gorshkov 1959), 1980 Mt St Helens (Lipman and Mullineaux 1981), 1990 Kelut (Bourdier et al. 1997), 1990–1995 Unzen (Nakada et al. 1999) and 2017-ongoing Nevados de Chillan (Moussallam et al. 2018) eruptions. Even if they do not necessarily lead to edifice destruction (e.g., 1888 Bandai eruption, Yamamoto et al. 1999), phreatic events can vaporize and expel water in shallow aquifers, unloading the upper crust hence changing the stress field around, and the magmatic pressure within, a magma reservoir (Pinel and Jaupart 2005). As such, phreatic events can act as trigger or facilitate magmatic eruptions. Given the delay observed between the phreatic eruptions (6 September 2017) and magma arrival at the surface (22 September 2017), it is conceivable that the phreatic eruptions played a role in favouring the magmatic ones, although it is likely that the phreatic eruptions themselves were driven by magmatic gases given the large quantities of SO₂ they released (Bani et al., in prep.).

Bulk rocks, matrix glasses and melt inclusions from phase one are compositionally more evolved than those of phase two and three (Fig. 5 and 6). Melt inclusions from phase two and three have higher Mg#, MgO, CaO content, lower Si and Al content and tend to be hosted in more forsteritic olivine (Fig. 6). An overlap in composition exist however between the three phases such that phase two and three, whilst dominated by more primitive melt inclusions, do

also contain inclusions that have similar compositions as the ones from phase one. Normal zoning of olivine and pyroxene phenocrysts is most commonly observed in the early and late phases of the eruption, yet reverse zoning is common also shown by a number of crystals, particularly in phase 2 and 3 (Fig. 2, 3 and 10). The observed compositional continuity (as opposed to a bimodal distribution) shown by the melt inclusions, from primitive to evolved end members (Fig. 5,6) suggest a simple magmatic evolution with all erupted melts related by simple liquid line of descent from the same primitive parental magma (potentially similar to the 8 to 14 wt.% MgO melts described by Sorbadère et al. 2011) (Fig. 6). Trace element variation diagrams from melt inclusions and matrix glasses (Fig. 7) show a strong linear correlation between Rb, Th and K₂O (highly incompatible elements) that points towards the origin as expected for melts related by fractional crystallisation. This again suggests that all melt inclusion compositions (including from older lavas) are consistent with simple fractional crystallisation evolution from a single parental magma.

Some trace element systematic do hint at mixing processes having taken place. A strong linear correlation is evident between La and Rb (Fig. 7C) but a regression through the data does not pass through the origin potentially indicative of mixing between two components (e.g., Schiano et al. 2010). Plots of Rb/Nd versus Rb and Th/Nd vs Th (Fig. 7E and F) define trends that are not well described by linear relationships suggesting again mixing between two components. On these diagrams, simple fractional crystallisation would result in no variations in Rb/Nd and Th/Nd as a function of Rb and Th concentrations respectively (e.g., Schiano et al. 2010). In addition, the fact that numerous olivine crystals exhibit reverse zoning and the bi-modal distribution observed in olivine forsterite content with a significant proportion of phase 2 and 3 olivine recording core Fo content higher in magnesium than

expected for equilibrium with the bulk rock (Fig. 4), both suggest that some of the olivine are antecrystic and that some amount of mixing has taken place.

Diffusion timescales from olivine zoning show multiple recharge events occurring a few days to a year prior to the 2017-2018 eruption (exact timescale depending on initial conditions). It is possible therefore that the latest recharge event played a role in triggering (via magmatic gas release) the initial phreatic and subsequent magmatic activity as the timing (a few days before phase 1 eruption) is consistent with the onset of phreatic and seismic activity.

Based on all these observations we suggest that:

1. The Ambae plumbing system was periodically recharged by co-genetic magmatic batches in the days to year prior to eruption.
2. The initial phreatic activity, propelled by magmatic gases, likely opened a pathway or modified the local stress field allowing pressurized magma to travel to the surface.
3. The eruption first tapped into the most evolved magma chamber(s) erupting phase one magma. As the eruption progresses from phase one to four, the most evolved chamber(s) got progressively exhausted opening the way to more primitive (and yet not necessarily deeper) magma from other chambers.
4. During or prior to the eruption some small amount of mixing between co-genetic magmas from different chambers occurred.

Magma ascent rate

Modelling of volatile diffusion profiles in melt embayments suggest magma decompression rates in the order of 0.24 to 0.55 MPa/s, corresponding to ascent rates of 9 to 21 m/s (33 to 75 km/h) while calculated decompression rates based on microlite number densities are in the

order 0.11 to 0.16 MPa/s corresponding to ascent rates of 4 to 6 m/s (15 to 23 km/h). Considering that the two methods are entirely independent their broad agreement is worth highlighting and brings confidence in the determined values. Given a shallow magma chamber at 0.5 to 3 km depth, it must hence have taken 24 to 122 seconds for the magma to travel from the top of the magma chamber to the surface. The decompression and ascent rates we calculate are comparable to other volcanoes using the same embayment volatile diffusion modelling method. Liu et al., (2007) investigated the 26.5 ka phreatomagmatic Oruanui eruption of Taupo (New Zealand) and found a decompression rate of 0.001 to 0.007 MPa/s. Myers et al., (2018) investigated the 25.4 ka Oruanui eruption of Taupo, 0.767 Ma Bishop Tuff and 2.08 Ma Huckleberry Tuff and found decompression rates in the order of 10^{-3} to 10^{-1} MPa/s. Humphreys et al., (2008) found decompression rates of 0.9-1.6 MPa/s for the 1980 Plinian eruption of Mount St Helens, (Washington, USA). Lloyd et al., (2014) found decompression rate of 0.3 to 0.5 MPa/s for the basaltic andesite sub-Plinian eruption of Volcán de Fuego (Guatemala) while Ferguson et al., (2016) found decompression rates between 0.05 and 0.45 MPa/s during fire fountain to sub-Plinian basaltic eruptions at Kīlauea volcano (Hawaii, USA). **Figure 13** shows a compilation of currently available magmatic decompression rate estimated using the embayment volatile diffusion method and compares it to eruption magnitude. For basaltic eruptions a trend seems to emerge of increasing decompression rate (i.e. ascent rate) with increasing eruption magnitude. The trend however does not extend to more silicic and higher magnitude eruptions with the largest, ultraplinian, eruptions yielding comparatively slow decompression rates.

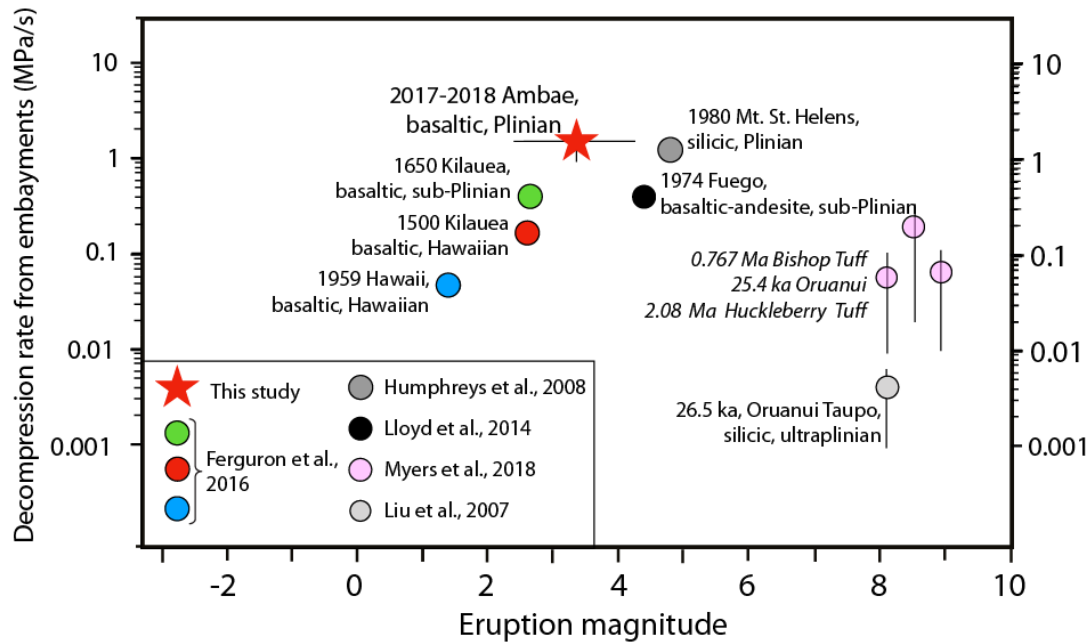


Figure 13: Compilation of decompression rate estimates using the embayment volatile diffusion method and comparison with eruption magnitude. For basaltic eruptions a trend of increasing decompression rate (i.e. ascent rate) with increasing eruption magnitude can be observed. The trend however does not extend to more silicic and higher magnitude eruptions.

Calculated magma ascent rate is strongly influenced by a parameter that is difficult to assess: the amount of pre-existing fluid/bubble before the onset of the final eruptive ascent (Ferguson et al., 2016). This is because the curve of the H_2O depletion with respect to pressure varies towards deeper degassing with more pre-existing fluid (see Fig. S1). While Ferguson et al. (2016) used H_2O - CO_2 -S profiles to simultaneously constrain this parameter, it was not possible for the case here as CO_2 , S and Cl show significant disequilibrium (super-saturation compared to H_2O). The values reported here are minimum ascent rate with pre-existing fluid mass of 0 %. Supposing a value of fluid mass of 1 %, the ascent rate increases by approximately a factor of 2.5. As the volatile element abundances in the melt inclusions are somewhat higher than the highest value of the embayments, an estimated pre-existing fluid mass of 0.5 to 1.5 % could be close to reality. A faster magma ascent rate is furthermore consistent with the strong disequilibrium recorded in slower diffusing elements (CO_2 , S, Cl). The magma ascent rate can hence be estimated to be in the order of 1-2 MPa/s.

Considering pre-eruptive volatiles in our calculations, magma ascent rate during the 2017-2018 Ambae eruption therefore appears as one of the fastest among other eruptions where the same methodology has been applied (Liu et al. 2007; Humphreys et al. 2008; Lloyd et al. 2014; Ferguson et al. 2016). This fast ascent was likely propelled by pre-eruptive volatiles, consistent with observations that the eruption reached the lower stratosphere (NASA Earth Observatory 2019) and released extremely large amounts of gases into the atmosphere (2.08 Tg of SO₂ in the first nine months, equivalent to a third of the atmospheric loading from the 1982 El Chichon eruption) while expelling relatively minor amounts of solid materials during the same period (Bani et al., in prep).

CONCLUSIONS

The 2017-2018 eruption of Ambae (Aoba) volcano was the largest recorded event in Vanuatu since the 1929 eruption of Ambrym. It expelled large quantities of gases to the atmosphere and prompted local authorities to evacuate the 11,000 inhabitants of the islands making it the largest volcanic disaster in the country's recorded history. In this contribution we used major, trace and volatile elements in bulk rocks, matrix glasses, melt inclusions, embayments and minerals to shed light on the magmatic processes operating during this eruption. We found that (i) The Ambae plumbing system was periodically recharged by magmatic batches in the year prior to eruption with the latest recharge event occurring only a few days before eruption. (ii) the eruption was likely facilitated by an initial phase of phreatic activity, itself triggered by magmatic gases potentially associated with the latest magmatic recharge event. (iii) The magma released during the eruption started crystallising at 14 to 5 km depth before residing in a shallow magma chamber, located at ~0.5 to 3 km depth. (iv) From the shallow magma chamber to the surface, minimum magma decompression rate ranges from 0.11 to

799 0.55 MPa/s, with a possible actual rate of 1-2 MPa/s, corresponding to average ascent rates in
800 the order of 15 to 270 km/h and a travel time between the top of the shallow reservoir and the
801 surface of 9 seconds to 2 minutes. Such a fast ascent rate, together with the high eruptive
802 plume heights (reaching the lower stratosphere) and the large quantities of SO₂ released,
803 suggest that a large amount of pre-eruptive volatiles must have been present in the reservoir.

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