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# Constraints from Phase Equilibrium Experiments on Pre-eruptive Storage Conditions in Mixed Magma Systems: a Case Study on Crystal-rich Basaltic Andesites from Mount Merapi, Indonesia

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## ABSTRACT

Mount Merapi is one of Indonesia's and the world's most hazardous volcanoes. Existing constraints on the location and the crystallization conditions of its pre-eruptive magma reservoir based on mineral and glass thermobarometry and geophysical surveys are inconclusive, yet of immediate importance for future hazard mitigation. Here, we use two series of phase equilibrium experiments to quantify the conditions of pre-eruptive magma storage and magma recharge in Merapi's upper-crustal reservoir: (1) to characterize crystallization at total equilibrium conditions; (2) to characterize crystallization at local equilibrium conditions between crystal rims and host melt. We demonstrate that this experimental approach can constrain pre-eruptive crystallization conditions and their variation in crystal-rich, mixed magma systems, such as that of Merapi, for which standard crystallization experiments and mineral thermobarometry fail. In agreement with geophysical estimates, we infer that Mount Merapi's pre-eruptive reservoir partially crystallizes at  $\geq 100\text{--}200 \pm 75$  MPa and thus at relatively shallow depths of  $c. \geq 4\text{--}5$  to  $\sim 9$  km. Magmas are stored at  $\geq 925\text{--}950 \pm 25^\circ\text{C}$  with a melt  $\text{H}_2\text{O}$  content of  $\sim 3\text{--}4$  wt % and a vapour phase with an  $X_{\text{H}_2\text{O}}$  [ $\text{H}_2\text{O}/(\text{H}_2\text{O} + \text{CO}_2)$ ] of  $\sim 0.5\text{--}0.6 \pm 0.1$ . Pre-eruptive recharge magmas, in contrast, have a temperature  $> 950$  to  $< 1000^\circ\text{C}$ , a higher melt  $\text{H}_2\text{O}$  content of  $\sim 4\text{--}5$  wt % and a vapour phase with a higher  $X_{\text{H}_2\text{O}}$  of  $\sim 0.8 \pm 0.1$ . We hypothesize that these variations in melt  $\text{H}_2\text{O}$  content and vapour  $X_{\text{H}_2\text{O}}$  between resident and pre-eruptive recharge magmas relate to variable degrees of open-system degassing of magma parcels en route into the reservoir rather than to variations in volatile fluxing of the stored magmas.

**Key words:**  $\text{H}_2\text{O}$ ;  $\text{CO}_2$ ; crystal-rich; experiment; magma mixing; Merapi; volatiles

## INTRODUCTION

Constraining pre-eruptive magma storage conditions is of immediate importance for assessing hazard potential and hazard types at active volcanoes (Woods, 1995; Sparks, 2003; Scaillet *et al.*, 2008; Ruprecht & Bachmann, 2012; Cashman & Sparks, 2013), yet this is

rarely achieved. Mount Merapi in Indonesia, one of the world's most hazardous volcanoes, is a prime example, for which pre-eruptive magma storage conditions are poorly constrained. The volcano rises above densely populated countryside and is located  $\sim 30$  km from the metropolis of Yogyakarta. On average, Merapi erupts

every few years. Its historical eruptions have been dominated by periods of lava dome growth and eventual gravitational dome collapse that typically produce VEI1 to VEI3 pyroclastic flows and surges (Voight *et al.*, 2000; Charbonnier & Gertisser, 2008; Gertisser *et al.*, 2011). These flows clearly affect the local population, but proficient hazard management has limited the number of fatalities. However, explosive VEI4 eruptions that affect a much larger area and population occasionally take place, as in 2010, and the rock record shows that such explosive events were more common in prehistoric times (Andreastuti *et al.*, 2000; Camus *et al.*, 2000; Gertisser *et al.*, 2011). For future hazard management it is thus imperative to (1) closely monitor the current pre-eruptive magma reservoir and (2) assess whether highly explosive eruptions will remain the exception, or whether they may become more common.

Geophysical surveys—including seismic, ground deformation, and magnetotelluric studies—have inferred shallow magma storage zones at depths of ~1.5–2.5 km and at >5 km to ~8.5 km below Merapi's summit; however, the shallowest zones may represent the presence of brines rather than magma storage zones (Ratdomopurbo & Poupinet, 1995, 2000; Beauducel & Cornet, 1999; Müller & Haak, 2004; Budi-Santoso *et al.*, 2013). Petrological studies and mineral barometry have suggested a wide range of possible crystallization pressures: (1) 0 to >600 MPa and predominantly ~100–400 MPa or ~400–500 MPa for clinopyroxene (Gertisser, 2001; Chadwick *et al.*, 2013; Preece *et al.*, 2014); (2) >300 to ~900 MPa and predominantly ~400–500 MPa for amphibole (Costa *et al.*, 2013; Nadeau *et al.*, 2013). Glass inclusion hygrobarmetry and the evolved compositions of plagioclase and clinopyroxene indicate limited shallow magma storage at ≤300 MPa (Costa *et al.*, 2013; Nadeau *et al.*, 2013; Preece *et al.*, 2014). In combination, these calculated pressures suggest crystallization throughout the crust and upper mantle, with major magma crystallization zones at depths of ~4–18 km. Estimates for crystallization temperatures in the magma storage reservoir range widely from ~950°C using amphibole thermometry (Costa *et al.*, 2013; Erdmann *et al.*, 2014) to ~1050°C using clinopyroxene thermometry (Preece *et al.*, 2014). The limited data available indicate pre-eruptive melt H<sub>2</sub>O contents between ~3.5 and 5.0 wt % (Borisova *et al.*, 2013; Preece *et al.*, 2014).

The aim of our study is to provide experimentally constrained estimates for the pre-eruptive crystallization conditions and processes in Merapi's current pre-eruptive magma storage reservoir and, in particular, constraints on the crystallization pressure and melt volatile content, which are key hazard factors. We anticipated that previous estimates of crystallization pressure and thus the location of the reservoir could be improved upon, given that they were inferred on the basis of amphibole and clinopyroxene mineral barometers with large uncertainties (Fig. 1).

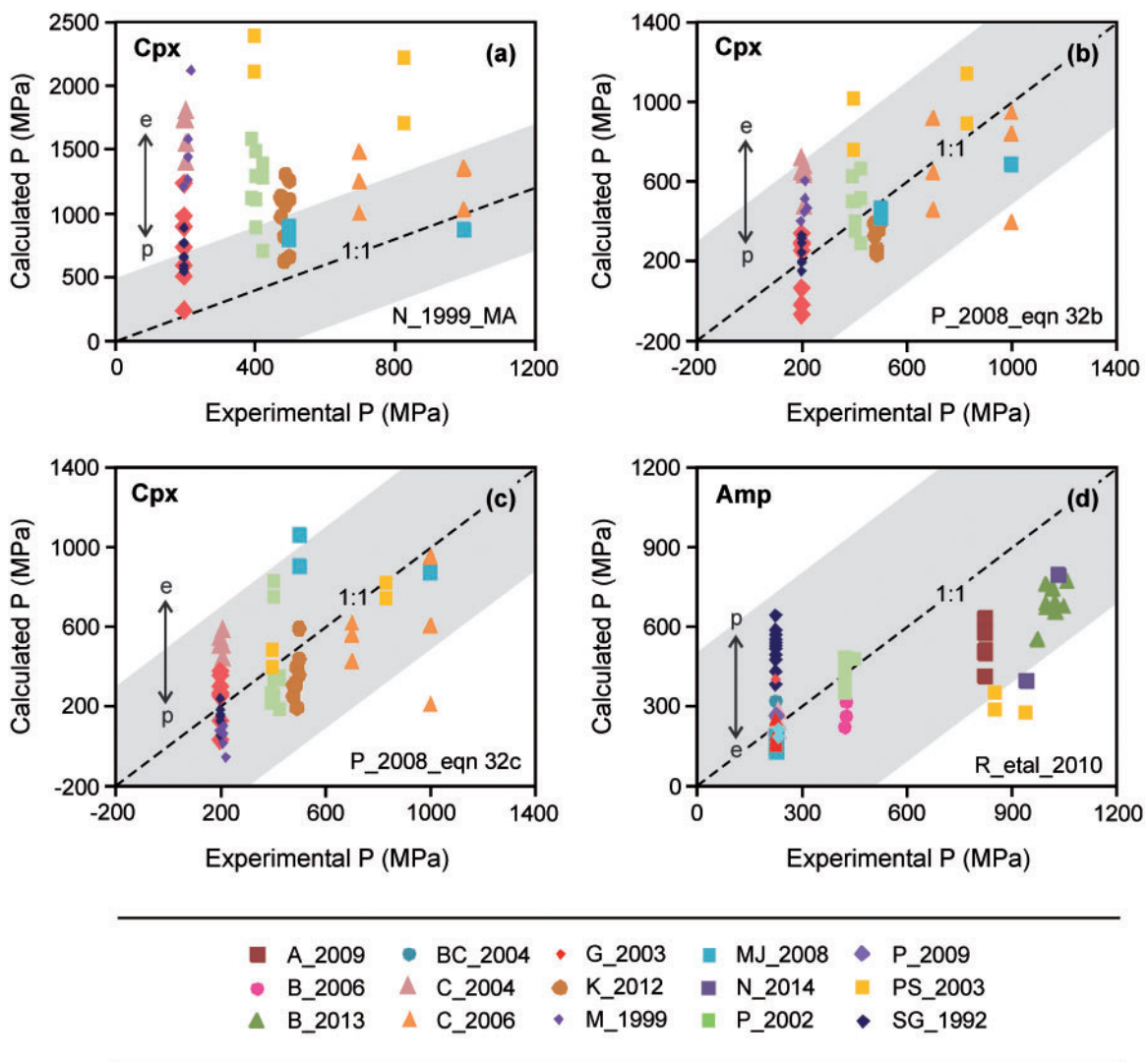
By mimicking the crystallization conditions of natural magma systems, experimental studies can in principle

provide highly accurate estimates for all intensive crystallization parameters. The challenge is, however, to constrain the crystallization conditions for mixed magma systems for which end-member magma compositions are not preserved in the rock record (Pichavant *et al.*, 2007; Blundy & Cashman, 2008). Merapi's crystal-rich, compositionally homogeneous basaltic andesites have been interpreted to represent a thorough mixture of relatively evolved resident and more primitive recharge magmas (Nadeau *et al.*, 2010, 2013; Chadwick *et al.*, 2013; Costa *et al.*, 2013), and thus they provide an excellent case study to address this challenge. To constrain the conditions in Merapi's pre-eruptive reservoir experimentally we needed to identify the natural equilibrium mineral assemblages and the crystallization conditions for the resident and the recharge magmas, respectively. To achieve this objective, we have (1) characterized the mixed mineral assemblages and their compositions in the eruption products of the paroxysmal 2010 eruption of Merapi and (2) performed two series of phase equilibrium experiments to quantify their record of crystallization conditions and processes during both magma storage and recharge.

## SAMPLES, METHODS, AND DATA

The 2010 eruption of Merapi was its largest explosive event since 1872. For this VEI~4 event, eight eruption stages have been distinguished by Komorowski *et al.* (2013): stage 1, precursory unrest; stage 2, vulcanian or phreatomagmatic explosions; stage 3, rapid dome growth and vulcanian explosions; stage 4, paroxysmal, full dome and upper conduit explosion; stage 5, retrogressive old dome collapse; stage 6, convective column fountain collapse; stage 7, rapid dome growth and ash venting; stage 8, continuous ash venting. The stage 7 dome has remained largely intact since its emplacement in 2010, although ash plumes and small volumes of dome clasts were emitted from it in July and November 2013 and in March and April 2014.

We have studied 25 samples of the 2010 eruptive and dome-forming event including the following: (1) black, dense and glassy to scoriaceous dome clasts of stage 4; (2) white pumice of stage 6; (3) light to medium grey, dense to scoriaceous clasts derived from the terminal stage 7 dome. The stage 4 + 6 samples were collected by us in 2010 in Kali Gendol and near the tourist site at Mariam Bunker. The stage 7 samples were collected by S. Byrdina at Pasar Bubar on 19 November 2013, immediately following their eruption in a small explosion that occurred on 18 November 2013 through the dome that was emplaced on 6–8 November 2010 during stage 7. The stage 4 and stage 6 samples were selected because they represent the main erupted magma volume and the most explosive eruption stage, respectively. The stage 7 samples were studied because they have not been previously characterized; they represent the final magma emplaced during the 2010 eruption cycle, and we anticipated that they may yield



**Fig. 1.** Comparison of experimental and calculated crystallization pressures using the composition of (a–c) clinopyroxene and (d) amphibole crystallized from calc-alkaline magmas and the barometers of (a) Nimis (1999), mildly alkaline (MA) calibration; (b, c) Putirka [2008, equations (32b) and (32c)]; and (d) Ridolfi *et al.* (2010). We note that Nimis (1999) explicitly discouraged the use of his barometer for clinopyroxene from calc-alkaline rocks. However, all the calibrations have been employed to estimate crystallization pressure for the Merapi magma plumbing system, even though they have large uncertainties of  $\geq 260$ –500 MPa and systematic errors relating to compositional variations, increasing or decreasing with evolved (e) or primitive (p) magma, melt and mineral compositions as indicated by double-headed arrows. The grey shaded fields highlight calculated pressures within  $\pm 500$  MPa of the known experimental pressures. Detailed evaluations of the thermobarometric methods, their potential and problems have been given by Putirka (2008), Erdmann *et al.* (2014) and Stamper *et al.* (2014). Sources of experimental data: A\_2009, Alonso-Perez *et al.* (2009); B\_2006, Bogaerts *et al.* (2006); B\_2013, Blatter *et al.* (2013); BC\_2004, Barclay & Carmichael (2004); C\_2004, Costa *et al.* (2004); C\_2006, Caricchi *et al.* (2006); G\_2003, Grove *et al.* (2003); K\_2012, Krawczynski *et al.* (2012); M\_1999, Martel *et al.* (1999); MJ\_2008, Mercer & Johnston (2008); N\_2014, Nandedkar *et al.* (2014); P\_2002, Pichavant *et al.* (2002); P\_2009, Pietranik *et al.* (2009); PS\_2003, Prouteau & Scaillet (2003); SG\_1992, Sisson & Grove (1992) (the dataset used is provided in Supplementary Data Electronic Appendix A1).

additional insight into the previously proposed, but cryptic pre-eruptive recharge of Merapi's upper magma system (e.g. Costa *et al.*, 2013). We are confident that the samples classified as stage 7 are derived from Merapi's terminal 2010 dome, because they have abundant calcic ( $\geq \text{An}_{60-70}$ ) plagioclase microlites that are characteristic for the magmas of the 2010 eruptive event, but are unknown from other recent eruption products (Costa *et al.*, 2013; Preece *et al.*, 2013). Magnetite phenocrysts in these samples show trellis-like exsolution, which is consistent with slow cooling

(e.g. Turner *et al.*, 2008) during magma ascent and/or dome residence, but they show no or minor oxidation. In the following analysis we distinguish between stage 4 + 6 and stage 7 samples, based on variable proportions of evolved and inclusion-rich versus primitive and inclusion-poor phenocrysts. The dome clasts and pumice samples of stage 4 + 6, however, have phenocryst populations with comparable textures and compositions, and they are thus described together.

Five selected whole-rock samples (one stage 4 dome clast, four stage 7 dome clasts) were geochemically

**Table 1:** Composition of whole-rock (WR) and glass (Gl) starting material normalized to 100 wt %

St. mat.	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Fe <sub>2</sub> O <sub>3</sub>	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	P <sub>2</sub> O <sub>5</sub>	Total
WR* (100 wt %)	55.11	0.78	19.20	7.69	0.21	2.38	8.39	3.79	2.13	0.32	98.1†
Gl‡ (100 wt %)	57.19	0.69	18.32	6.86	0.12	2.36	8.28	3.81	2.12	0.25	98.9†

St. mat., starting material sample SEM043b.

\*Analyzed by XRF at Saint Mary's University, Canada.

†Pre-normalized totals.

‡Analyzed by electron microprobe at ISTO, France.

The ~11% lower Fe<sub>2</sub>O<sub>3</sub> content determined for the glass as compared with the whole-rock composition may partly reflect Fe loss during glass fusion. Apparent deviations in other element concentrations (e.g. ~4 and 5% for SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub>) between WR and Gl starting materials are considered to reflect analytical and method-related errors.

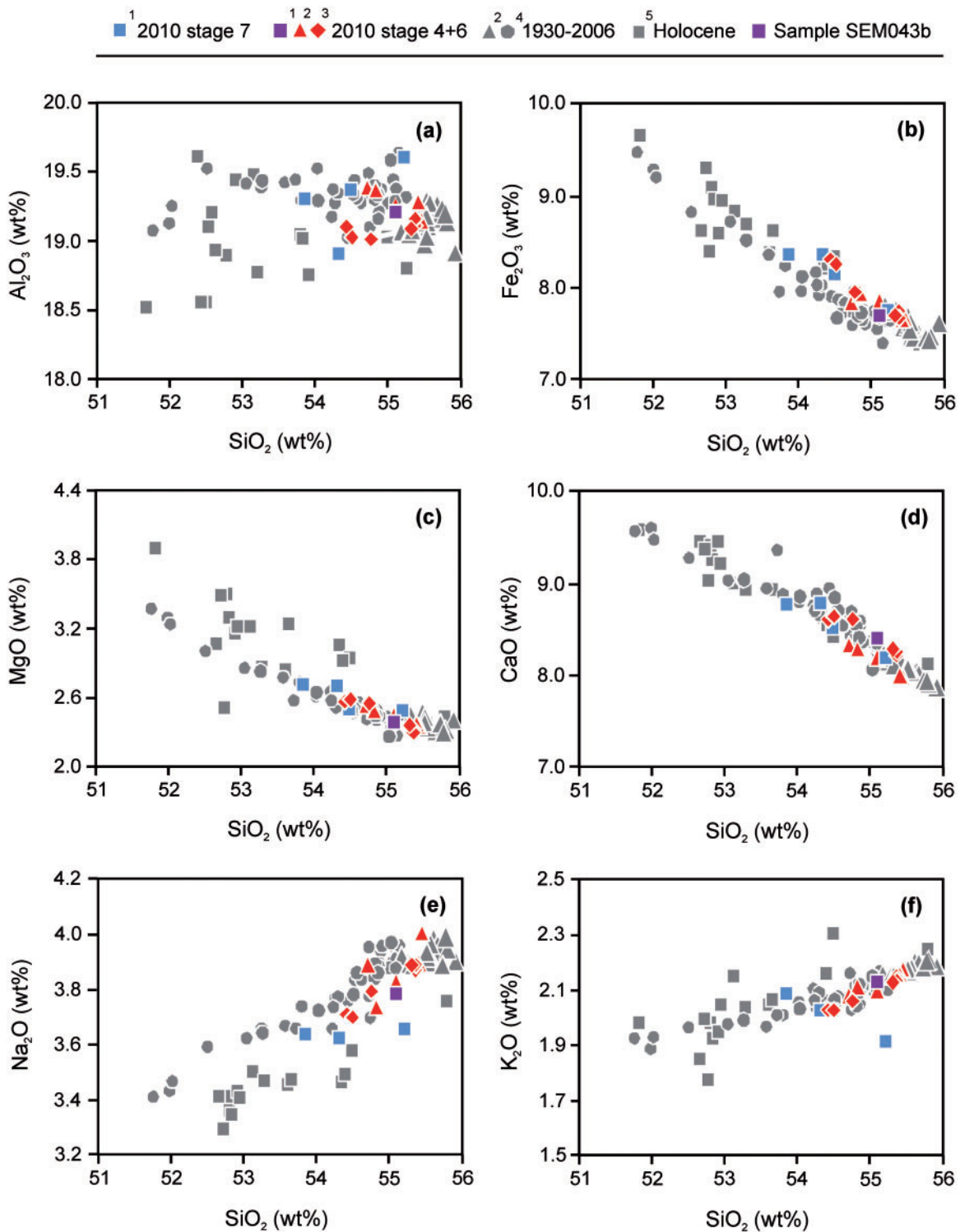
analyzed at the Regional Analytical Centre at Saint Mary's University, Canada, and 25 natural samples were studied in thin section. For our experimental study, we used one of the black, glassy, dense basaltic andesite stage 4 dome clasts (sample SEM043b) as the starting material (Table 1; Fig. 2). We have selected this starting material for the following reasons: (1) the sample represents the main clast type that was discharged in the high-volume pyroclastic flows on 5 November 2010 [i.e. it is a Type 1 clast following the classification of Charbonnier *et al.* (2013), which make up >61 to 84 vol. % and on average 77 vol. % of all stage 4 pyroclasts]; (2) the sample has a whole-rock composition close to the average of all juvenile pyroclasts erupted in 2010 (Fig. 2); (3) the sample has a phenocryst and microphenocryst assemblage of plagioclase–clinopyroxene–orthopyroxene–amphibole–magnetite, which is characteristic of Merapi basaltic andesites (Fig. 3a and b); (4) the sample has mineral rim compositions similar to those of other stage 4 pyroclasts and stage 6 pumice samples investigated (Figs 2 and 4c, g, k; Supplementary Data Electronic Appendix A2; supplementary data are available for downloading at <http://www.petrology.oxfordjournals.org>).

For one series of experiments, referred to as 'melting experiments', we have used a gently crushed powder of the crystal-rich basaltic andesite whole-rock, which was partially melted and partially recrystallized during the experiments. For the second series of experiments, referred to as 'crystallization experiments', we have fused the same rock powder to a homogeneous glass at atmospheric conditions, which was then partially crystallized during the experiments. The melting experiments were performed to characterize the local equilibrium compositions of crystal rims and their host melts for Merapi's crystal-rich basaltic andesites (i.e. accounting for the fact that crystal cores of Merapi's mixed mineral assemblage are not in equilibrium) (Table 2). The crystallization experiments were performed to provide phase compositions that form at equilibrium for the basaltic andesite composition (Table 2), which guided our interpretation of the mineral rim compositions detected in the melting experiments. For both experimental series, Au capsules were first loaded with variable amounts of ultrafiltered, deionized H<sub>2</sub>O and silver oxalate (Ag<sub>2</sub>C<sub>2</sub>O<sub>4</sub> → CO<sub>2</sub>) to yield equilibrium with a

vapour phase with an XH<sub>2</sub>O [H<sub>2</sub>O/(H<sub>2</sub>O + CO<sub>2</sub>)] between ~0.4 and 1 (Table 3), and then filled with the starting rock or starting glass powders at a fluid:solid ratio of ~1:10 to impose volatile-saturated conditions as in Merapi's mid- to upper-crustal magma system (see Nadeau *et al.*, 2010, 2013).

The two series of experiments were run simultaneously in internally heated pressure vessels at the Institut des Sciences de la Terre d'Orléans (ISTO), France. The oxygen fugacity (*f*O<sub>2</sub>) was controlled by loading ~1 bar of H<sub>2</sub> into the vessels, except for one experiment that was performed at the intrinsic *f*O<sub>2</sub> of the vessel at >NNO + 2 (where NNO is the nickel–nickel oxide buffer) (CE-1, Table 2). The *f*O<sub>2</sub> was monitored by a pair of NiO + Ni–Pd solid sensors (see Pownceby & O'Neill, 1994), which indicate that the *f*O<sub>2</sub> of the experiments with H<sub>2</sub> added reached ~NNO + 1.5 to ~NNO + 1.3 in charges with an XH<sub>2</sub>O<sub>V</sub> = 1, and the *f*O<sub>2</sub> was ~NNO + 0.9 to ~NNO + 0.7 in charges with an XH<sub>2</sub>O<sub>V</sub> of ~0.6–0.5 (Table 3). The experimental pressure and temperature were then induced by loading Ar into the vessel, followed by heating at a rate of 10–15°C min<sup>-1</sup>. At the beginning of each experiment, we overstepped the targeted experimental temperature by ~40°C for 2–3 h to promote partial melting and subsequent crystallization of mineral rims that were wide enough for analysis. All experiments were run for ~3 days at 925–1000°C and ~100–400 MPa (Table 3), and ended by drop quench. Temperature and total pressure were continuously monitored; their uncertainties are ± 5°C and ± 2 MPa.

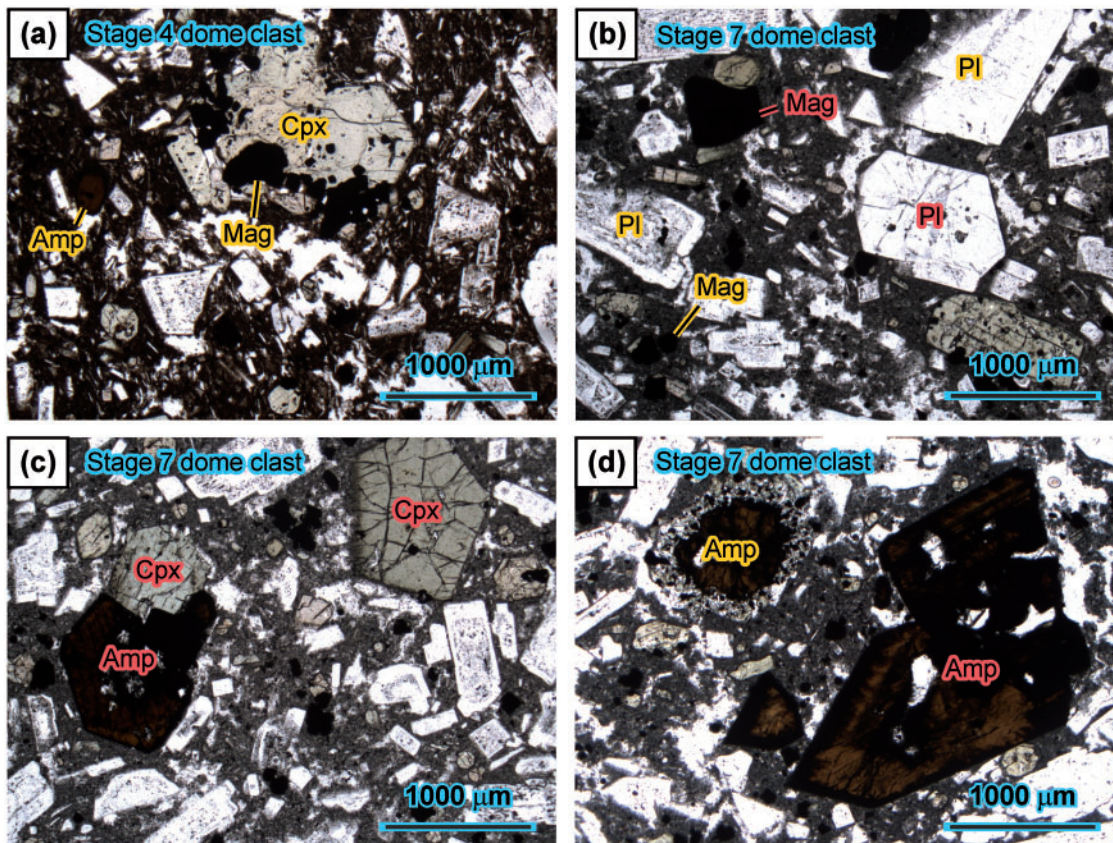
For the natural samples, we distinguish between phenocrysts (>500 μm), microphenocrysts (~100–500 μm) and microlites (>500 μm). For the melting experiments, we distinguish between unreacted crystal cores and newly crystallized inner and outer mineral rims. Mineral proportions in the experimental and natural samples were determined by point counting on backscattered electron images and thin sections, respectively. Mineral, matrix and glass compositions were determined by electron microprobe at ISTO. Average and representative mineral and glass compositional data for the natural samples and the experimental charges are provided in Supplementary Data Electronic Appendices A2 and A3. For the natural samples we determined the phenocryst core and rim compositions for 10 samples and the matrix



**Fig. 2.** Major-element compositional variations of Merapi whole-rocks normalized to 100 wt % anhydrous. The stage 4 + 6 samples (erupted on 5 November) and the stage 7 samples (of the terminal 2010 dome, which were erupted on 18 November 2013) have basaltic andesite compositions, notably lacking a juvenile basaltic component. Data sources: (1) this study; (2) [Preece et al. \(2013\)](#); (3) [Costa et al. \(2013\)](#); (4) [Abdurachman \(1998\)](#); (5) [Gertisser & Keller \(2003\)](#). Sample SEM043b is the sample that was used as the starting material for our experiments.

and glass inclusion compositions for six samples. For the crystallization experiments, we determined mineral and glass compositions for key charges. For the melting experiments, we determined mineral rim and glass

compositions whenever they were large enough for analysis. The electron microprobe operating conditions were set at 15 keV and 10 nA with peak and background counting times of 10–20 s. All minerals were analyzed with a

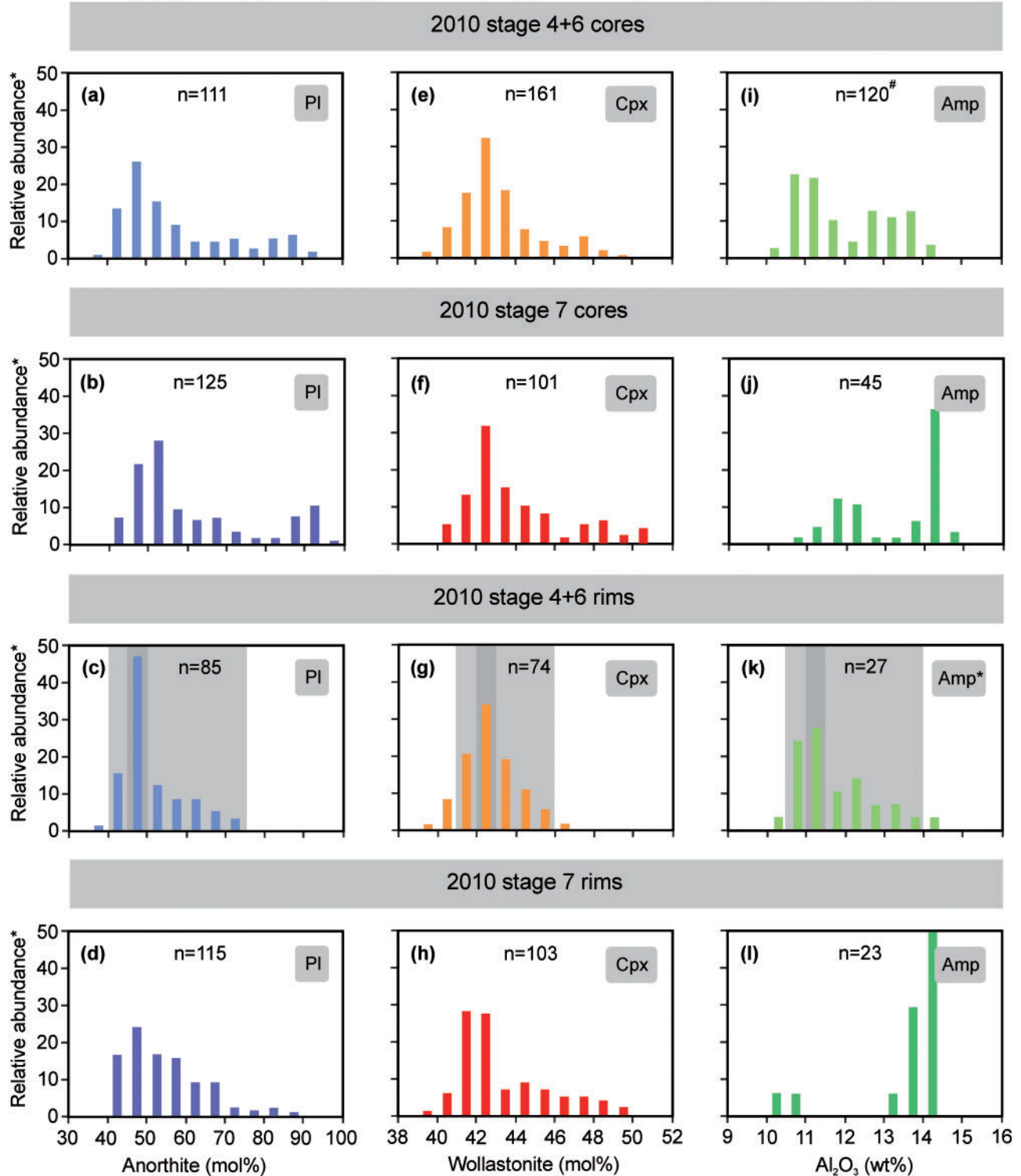


**Fig. 3.** Photomicrographs of typical (a) stage 4 and (b–d) stage 7 samples from the Merapi 2010 eruptive event. All are basaltic andesitic, crystal-rich, and dense dome clasts. Crystals with inclusion-rich zones, resorption features, complex oscillatory zoning and/or evolved compositions form the main assemblages in all samples. Typical examples are highlighted by yellow labels. Crystals with few inclusions, normal zoning and primitive core compositions are more abundant in the stage 7 than in the stage 4 + 6 samples. Typical examples are highlighted by red labels.

focused beam; the experimental and the natural glasses and matrix compositions were analyzed using defocused 20  $\mu\text{m}$  and 50  $\mu\text{m}$  beam diameters, respectively. We analyzed both glass inclusions and the matrix because we anticipated that the matrix compositions would provide a more complete record of pre-eruptive melt compositions and compositional variation. The matrix analyses intentionally included glass domains and microlites, because they were performed to characterize the melt present in the pre-eruptive reservoir prior to microlite crystallization. Alkali migration in glasses under the electron beam was corrected for relative to the hydrous glass standards of Scaillet & Evans (1999). Glass volatile contents, which primarily represent glass  $\text{H}_2\text{O}$  contents, were estimated relative to the hydrous glass standards and electron microprobe totals with an uncertainty of  $c. \pm 1 \text{ wt } \%$ . Glass  $\text{H}_2\text{O}$  and  $\text{CO}_2$  contents were also calculated using VolatileCalc (Newman & Lowenstern, 2002) for an average of predicted volatile solubilities in rhyolitic and basaltic melts (to account for the dacitic melt compositions in Merapi's magma reservoir). The uncertainties on the calculated solubilities are typically better than  $\pm 1 \text{ wt } \%$  for  $\text{H}_2\text{O}$ , but up to  $\pm 1000 \text{ ppm}$  for  $\text{CO}_2$ , whereas relative uncertainties on  $\text{CO}_2$  solubilities at  $< 500 \text{ MPa}$  and in a vapour with an  $\text{XH}_2\text{O}$  of  $\leq 0.5$  are  $< 200 \text{ ppm}$  (see Baker &

Alletti, 2012). Because the calculated glass volatile and glass  $\text{H}_2\text{O}$  contents closely compare (Table 3), and glass volatile contents could not be determined for all experimental charges, we predominantly refer to the calculated melt  $\text{H}_2\text{O}$  contents in subsequent text and figures. The vapour  $\text{XH}_2\text{O}$  compositions ( $\text{XH}_2\text{O}_\text{V}$ ) reported throughout the text and figures are those that prevailed once the experimental melts had reached volatile saturation and the charges had reached equilibrium ( $\text{XH}_2\text{O}_\text{V} = \text{XH}_2\text{O}_\text{eq}$ ; Table 3).

Reported mineral end-member compositions are percentage anorthite  $\{\text{An} = [\text{Ca}/(\text{Ca} + \text{Na} + \text{K})] \times 100\}$  and orthoclase  $\{\text{Or} = [\text{K}/(\text{K} + \text{Ca} + \text{Na})] \times 100\}$  for plagioclase, enstatite  $\{\text{En} = [\text{Mg}/(\text{Mg} + \text{Fe}_{\text{tot}} + \text{Ca})] \times 100\}$ , wollastonite  $\{\text{Wo} = [\text{Ca}/(\text{Ca} + \text{Mg} + \text{Fe}_{\text{tot}})] \times 100\}$  and ferrosilite  $\{\text{Fs} = [\text{Fe}_{\text{tot}}/(\text{Ca} + \text{Mg} + \text{Fe}_{\text{tot}})] \times 100\}$  for clinopyroxene and orthopyroxene, and forsterite  $\{\text{Fo} = [\text{Mg}/(\text{Mg} + \text{Fe}_{\text{tot}})] \times 100\}$  for olivine. Normal and reverse zoning refer to rimward decreasing or increasing Ca, Al, or Mg contents, respectively. Mineral–melt partition coefficients ( $K_D$ ) are  $(\text{Ca}/\text{Na})_{\text{Pl}}/(\text{Ca}/\text{Na})_{\text{Gl}}$  for plagioclase and  $(\text{Fe}/\text{Mg})_{\text{Pl}}/(\text{Fe}/\text{Mg})_{\text{Gl}}$  for clinopyroxene, which are reported as average values for single charges. The mineral abbreviations used in text and figures follow those of Whitney & Evans (2010).



**Fig. 4.** Summary of the compositional variation for (a–d) plagioclase, (e–h) clinopyroxene, and (i–l) amphibole core and rim compositions in stage 4 + 6 and stage 7 samples. The core compositions are bimodal; the rim compositions show a wide range of compositions in which primitive compositions are slightly more common in the stage 7 than in the stage 4 + 6 samples. \*, abundance normalized to 100; *n*, number of original analyses; #, data from Erdmann *et al.* (2014). Dark and light grey shaded fields in (c), (g), and (k) show the mode and range of outer mineral rim compositions for our starting material sample SEM043b.

**Table 2:** Possible experimental and natural equilibrium states

Total equilibrium = equilibrium for all phases and chemical components of the system
Local equilibrium = equilibrium for all crystal rims and host melt, but not for crystal cores and host melt
Partial equilibrium = equilibrium for some, but not all crystal rims and host melt



**Table 3:** Conditions and run products of crystallization experiments (CE) and melting experiments (ME)

Exp.	Charge (vol. %)	XH <sub>2</sub> O <sub>in</sub> * (wt %)	XH <sub>2</sub> O <sub>eq</sub> † (wt %)	ΔNNO‡ (ppm)	Phase assemblage	φ	Melt V§	Melt H <sub>2</sub> O¶ (wt %)	Melt CO <sub>2</sub> ¶ (wt %)
<i>Run 10: 99 MPa, 950° C, fH<sub>2</sub> = 1 bar, t = 85 h</i>									
CE-10	10-36c	1.00	1.00	1.5	Gl, Pl, Cpx, Mag	36	—	3.3	0
CE-10	10-38c	0.80	0.76	1.3	Gl, Pl, Cpx, Mag	50	—	2.8	130
CE-10	10-37c	0.73	0.68	1.2	Gl, Pl, Cpx, Mag, Opx	60	—	2.6	180
CE-10	10-39c	0.60	0.55	1.0	Gl, Pl, Cpx, Mag, Opx	75	—	2.3	240
CE-10	10-41c	0.56	0.51	0.9	Gl, Pl, Cpx, Mag, Opx	87	—	2.2	270
ME-10	10-20m	1.00	1.00	1.5	Gl, Pl, Cpx, Mag	40	3.8	3.3	0
ME-10	10-21m	0.80	0.77	1.3	Gl, Pl, Cpx, Mag, Opx	52	3.5	2.8	120
ME-10	10-22m	0.63	0.56	1.0	Gl, Pl, Cpx, Mag, Opx	58	2.9	2.4	240
ME-10	10-23m	0.53	0.48	0.9	Gl, Pl, Cpx, Mag, Opx	65	2.5	2.1	290
<i>Run 9: 202 MPa, 925° C, fH<sub>2</sub> = 1 bar, t = 80 h</i>									
CE-9	9-31c	1.00	1.00	1.5	Gl, Pl, Cpx, Mag	40	—	4.9	0
CE-9	9-32c	0.81	0.75	1.3	Gl, Pl, Cpx, Mag	52	—	4.0	280
CE-9	9-34c	0.62	0.58	1.0	Gl, Pl, Cpx, Mag	80	—	3.4	480
CE-9	9-35c	0.56	0.54	0.9	Gl, Pl, Cpx, Mag	90	—	3.3	520
ME-9	9-16m	1.00	1.00	1.5	Gl, Pl, Cpx, Mag, Amp	43	5.3	4.9	0
ME-9	9-17m	0.81	0.75	1.3	Gl, Pl, Cpx, Mag, Amp	49	4.8	4.0	290
ME-9	9-18m	0.62	0.54	1.0	Gl, Pl, Cpx, Mag	62	4.0	3.3	520
ME-9	9-19m	0.56	0.49	0.9	Gl, Pl, Cpx, Mag	70	3.8	3.0	590
<i>Run 7: 204 MPa, 950° C, fH<sub>2</sub> = 1 bar, t = 70 h</i>									
CE-7	7-26c	1.00	1.00	1.3	Gl, Pl, Cpx, Mag	25	5.5	4.8	0
CE-7	7-27c	0.90	0.86	1.2	Gl, Pl, Cpx, Mag	42	5.2	4.3	140
CE-7	7-29c	0.72	0.66	0.9	Gl, Pl, Cpx, Mag	66	3.8	3.6	360
CE-7	7-30c	0.60	0.56	0.8	Gl, Pl, Cpx, Mag	80	—	3.2	460
ME-7	7-13m	1.00	1.00	1.3	Gl, Pl, Cpx, Mag	35	5.3	4.8	0
ME-7	7-15m	0.64	0.55	0.8	Gl, Pl, Cpx, Mag	55	3.1	3.2	470
<i>Run 6: 195 MPa, 1000° C, fH<sub>2</sub> = 1 bar, t = 67 h</i>									
CE-6	6-11c	1.00	1.00	1.3	Gl, Pl, Cpx	7	—	4.8	0
CE-6	6-12c	0.90	0.85	1.2	Gl, Pl, Cpx, Mag	30	—	4.2	150
CE-6	6-13c	0.81	0.74	1.0	Gl, Pl, Cpx, Mag	48	—	3.9	260
CE-6	6-14c	0.72	0.65	0.9	Gl, Pl, Cpx, Mag	60	—	3.7	340
CE-6	6-15c	0.63	0.57	0.8	Gl, Pl, Cpx, Mag	72	—	3.3	430
ME-6	6-7m	1.00	1.00	1.3	Gl	0	—	4.8	0
ME-6	6-8m	0.81	0.73	1.0	Gl, Pl, Cpx, Mag	35	4.4	3.9	270
ME-6	6-9m	0.62	0.51	0.7	Gl, Pl, Cpx, Mag	50	3.4	3.1	480
<i>Run 5: 295 MPa, 975° C, fH<sub>2</sub> = 1 bar, t = 67 h</i>									
CE-5	5-16c	1.00	1.00	1.5	Gl	0	—	6.4	0
CE-5	5-17c	0.90	0.83	1.3	Gl, Pl, Cpx, Mag	36	—	5.6	280
CE-5	5-18c	0.82	0.72	1.2	Gl, Pl, Cpx, Mag	43	—	5.0	480
CE-5	5-19c	0.71	0.61	1.1	Gl, Pl, Cpx, Mag	58	—	4.5	640
CE-5	54-20c	0.63	0.54	1.0	Gl, Pl, Cpx, Mag	70	—	4.1	750
ME-5	5-4m	1.00	1.00	1.5	Gl	0	6.4	6.4	0
ME-5	5-5m	0.81	0.71	1.2	Gl, Pl, Cpx, Mag	42	5.5	5.0	480
ME-5	5-6m	0.63	0.47	0.9	Gl, Pl, Cpx, Mag	49	4.4	3.8	860
<i>Run 8: 390 MPa, 1000° C, fH<sub>2</sub> = 1 bar, t = 72 h</i>									
CE-8	8-21c	1.00	1.00	1.4	Gl	0	8.7	7.9	0
CE-8	8-23c	0.80	0.61	1.0	Gl, Pl, Cpx, Mag	27	5.5	5.5	900
CE-8	8-24c	0.71	0.49	0.8	Gl, Pl, Cpx, Mag	38	4.5	4.7	1140
CE-8	8-25c	0.62	0.46	0.7	Gl, Pl, Cpx, Mag	54	—	4.3	1220
ME-8	8-11m	0.81	0.61	1.0	Gl, Pl, Cpx	27	6.0	5.4	890
ME-8	8-12m	0.62	0.42	0.7	Gl, Pl, Cpx, Mag	44	4.7	4.1	1320
<i>Run 1: 200 MPa, 975° C, fH<sub>2</sub> = 0 bar, t = 79 h</i>									
CE-1	1-6c	1.00	1.00	>2-3**	Gl, Pl, Cpx, Mag	16	—	4.9	0
CE-1	1-8c	0.81	0.75	>2-3**	Gl, Pl, Cpx, Mag	49	—	4.0	270
CE-1	1-9c	0.72	0.66	>2-3**	Gl, Pl, Cpx, Mag	63	—	3.7	360
CE-1	1-10c	0.63	0.59	>2-3**	Gl, Pl, Cpx, Mag	76	—	3.4	450

\*XH<sub>2</sub>O<sub>in</sub> = H<sub>2</sub>O/(H<sub>2</sub>O + CO<sub>2</sub>) (mol %) added to the charges.

†XH<sub>2</sub>O<sub>eq</sub> = final XH<sub>2</sub>O of the experiment, calculated from XH<sub>2</sub>O<sub>in</sub>, melt H<sub>2</sub>O solubilities, and experimental glass proportions; XH<sub>2</sub>O<sub>v</sub> reported in text corresponds to XH<sub>2</sub>O<sub>eq</sub>.

‡log fO<sub>2</sub> (experiment) - log fO<sub>2</sub> (NNO), where fO<sub>2</sub> in experiments was calculated from solid NiO + Ni-Pd sensors for variable XH<sub>2</sub>O of the charges.

§Melt V is total melt volatile content, estimated by difference from electron microprobe totals and corrected relative to hydrous glass standards.

¶Calculated using VolatileCalc of Newman & Lowenstern (2002), where summarized H<sub>2</sub>O and CO<sub>2</sub> contents are rounded values; estimated uncertainties are high; that is, up to ~1000 ppm (see Baker & Alletti, 2012).

\*\*Intrinsic fO<sub>2</sub> of the pressure vessel (see Scaillet & Evans, 1999).

Exp., experiment; CE, crystallization experiment; ME, melting experiment; φ, crystallinity; Gl, glass; Pl, plagioclase; Cpx, clinopyroxene; Opx, orthopyroxene; Mag, magnetite; Amp, amphibole.

## RESULTS

### The natural samples

#### Whole-rock composition and petrology

Magmas that have erupted at Merapi in prehistoric to historical times are predominantly basaltic andesite in composition, but occasionally include basalts (Abdurachman, 1998; Gertisser & Keller, 2003; Chadwick *et al.*, 2013; Costa *et al.*, 2013; Preece *et al.*, 2013; Fig. 2). Their SiO<sub>2</sub> contents correlate poorly with Al<sub>2</sub>O<sub>3</sub> (Fig. 2a), negatively correlate with MgO, Fe<sub>2</sub>O<sub>3</sub>, CaO and TiO<sub>2</sub> (Fig. 2b–d), and positively correlate with Na<sub>2</sub>O and K<sub>2</sub>O (Fig. 2e and f). The stage 4 + 6 samples erupted in 2010 and the stage 7 samples of the final 2010 dome have moderately evolved basaltic andesite compositions with ~53.9–55.4 wt % SiO<sub>2</sub> and ~18.9–19.6 wt % Al<sub>2</sub>O<sub>3</sub>, comparable with those of the majority of magmas discharged at Merapi from 1930 to 2006 (Abdurachman, 1998; Gertisser & Keller, 2003; Costa *et al.*, 2013; Preece *et al.*, 2013; Fig. 2). Basaltic magmas or enclaves, however, did not erupt in 2010 (stage 4 + 6) or in 2013 from the terminal 2010 dome (stage 7). Calc-silicate xenoliths are scarce (<<1 vol. %) and none of the samples is enriched in CaO. The compositions of the stage 4 + 6 and 7 samples are broadly comparable, but the stage 7 samples extend on average to slightly more primitive compositions with ~54.7 wt % SiO<sub>2</sub> relative to the stage 4 + 6 samples with ~55 wt % SiO<sub>2</sub> (Fig. 2a).

The investigated rocks, including dense and scoriaceous dome clasts and pumice samples, are crystal-rich with ~40–60 vol. % phenocrysts and microphenocrysts (Fig. 3). Plagioclase (~25–40 vol. %) and clinopyroxene (~5–15 vol. %) form the following: (1) common phenocrysts with complex zoning patterns, evidence of resorption, and inclusion-rich zones (Fig. 3a–c); (2) microphenocrysts that are largely inclusion-free (Fig. 3a–d); (3) rare phenocrysts that are inclusion-poor and largely unresorbed with core–rim step zoning (Fig. 3b–d). Magnetite (≤5 vol. %), amphibole (≤5 vol. %), and orthopyroxene (≤2 vol. %) are minor constituents that form subhedral to euhedral crystals (Fig. 3a–d). Olivine rarely occurs as inclusions <20 μm in size, most commonly in amphibole phenocrysts. The stage 4 + 6 samples characteristically have few inclusion-poor plagioclase and clinopyroxene phenocrysts with simple core–rim zoning (typically <5 vol. %); they contain abundant subhedral, but few euhedral magnetite crystals (Fig. 3a). Largely unresorbed amphibole microphenocrysts are common, whereas unresorbed amphibole phenocrysts are absent (Fig. 3a). Most of the stage 7 samples, in contrast, have a larger proportion of inclusion-poor plagioclase and clinopyroxene crystals with simple core–rim zoning (typically ~5–15 vol. %), and euhedral magnetite crystals (Fig. 3b). They also have largely unresorbed amphibole phenocrysts in addition to microphenocrysts (Fig. 3d).

#### Mineral composition and zoning

Plagioclase, clinopyroxene, and amphibole in the basaltic andesites show a large range in composition, in

which primitive core and rim compositions slightly increase in abundance from stage 4 + 6 to stage 7 samples (Fig. 4; Supplementary Data Electronic Appendix A2). Plagioclase has ~An40–90 core compositions with compositional maxima at ~An50 and ~An90, where the primitive >An80 compositions are characteristic of inclusion-poor crystal cores (Fig. 4a and b). The crystal rims have predominantly ~An50 to ~An70 composition, but range from ~An40 to ~An90 (Fig. 4c and d). Clinopyroxene is augite and less commonly diopside in composition, where cores (Fig. 4e and f) and rims (Fig. 4g and h) have ~Wo40–50 content with compositional maxima around ~Wo42–43. The core–rim zoning patterns of plagioclase and clinopyroxene include normal and reverse zoning, but zoning is most commonly irregularly oscillatory (Fig. 5). Broad plateau zones with evolved composition (i.e. typically with ~An40–50 and ~Wo42–43) and ~10–100 μm wide zones with intermediate composition (i.e. with ≤An70 and ≤Wo50) alternate. The outer ~100 μm of the crystals most commonly show normal zoning, but reverse zoning is

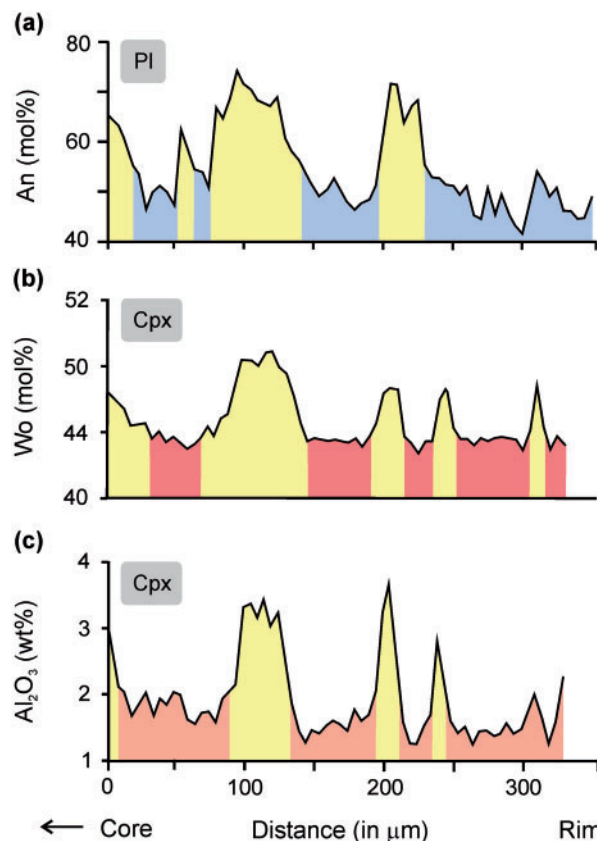


Fig. 5. Typical zoning profiles for (a) anorthite (An) in plagioclase, (b) wollastonite (Wo) and (c) Al<sub>2</sub>O<sub>3</sub> in clinopyroxene from a stage 4 dome clast [same profile in (b) and (c)]. Characteristic features are (1) the evolved plateau zones with ~An50 plagioclase and ~Wo42–43 and ~2 wt % Al<sub>2</sub>O<sub>3</sub> in clinopyroxene, and (2) the irregular and relatively thin, intermediate zones with <An70 in plagioclase and ≤Wo50 and <6 wt % Al<sub>2</sub>O<sub>3</sub> in clinopyroxene.

also developed, particularly for plagioclase and clinopyroxene of the stage 7 samples.

Amphibole microphenocrysts in stage 4 + 6 samples have core compositions with  $\sim 11$  and  $\sim 13$  wt %  $\text{Al}_2\text{O}_3$  (Fig. 4i). The crystals are normally or reversely zoned with rim compositions of  $\sim 10$ – $14$  wt %  $\text{Al}_2\text{O}_3$  (Fig. 4k). Amphibole phenocrysts that are characteristic for the stage 7 samples are Al-rich compared with the microphenocrysts of the stage 4 + 6 samples, with compositional maxima at  $\sim 12$  and  $14$  wt %  $\text{Al}_2\text{O}_3$  (Fig. 4j). The cores with  $\sim 12$  wt %  $\text{Al}_2\text{O}_3$  often have patchy zoning and small olivine inclusions. The cores of the volumetrically predominant phenocryst type with  $\sim 14$  wt %  $\text{Al}_2\text{O}_3$  are largely unzoned except for low-amplitude oscillatory zoning and  $< 200 \mu\text{m}$  wide normally zoned rims. The rims of the stage 7 phenocrysts and core or rim compositions of the stage 7 microphenocrysts were not analyzed, because they are pervasively oxidized. The near-rim compositions of the phenocrysts are, however, predominantly Al-rich with  $\sim 14$  wt %  $\text{Al}_2\text{O}_3$  (Fig. 4l), overlapping with the core and rim compositions of the most primitive stage 4 + 6 microphenocrysts (Fig. 4i and k).

Magnetite with subhedral shape, which is predominant in the stage 4 + 6 and the stage 7 samples, has core compositions with  $\sim 77$  wt % FeO,  $\sim 2.7$  wt %  $\text{Al}_2\text{O}_3$  and  $\sim 8.8$  wt %  $\text{TiO}_2$ . The crystals in the stage 4 + 6 samples are reversely zoned with rim compositions of  $\sim 76$  wt % FeO,  $\sim 2.6$ – $4.0$  wt %  $\text{Al}_2\text{O}_3$ , and  $\sim 7.0$ – $9.2$  wt %  $\text{TiO}_2$ , whereas those in the stage 7 samples show trellis-like exsolution that has overprinted any original zoning. The rare euhedral magnetite crystals, which are characteristic for the stage 7 samples, have  $\sim 74$  wt % FeO,  $\sim 6.0$  wt %  $\text{Al}_2\text{O}_3$  and  $\sim 5.6$  wt %  $\text{TiO}_2$  core composition, but their original zoning is also not preserved, owing to exsolution. Orthopyroxene, which predominantly occurs as microphenocrysts, has  $\sim \text{En}67$  core composition and is largely unzoned. Olivine, which occurs only in the form of inclusions, has an evolved  $\sim \text{Fo}60$ – $70$ , low NiO

( $< 0.25$  wt %), and high MnO ( $\sim 1.0$ – $1.4$  wt %) and CaO ( $\sim 0.1$ – $0.2$  wt %) composition.

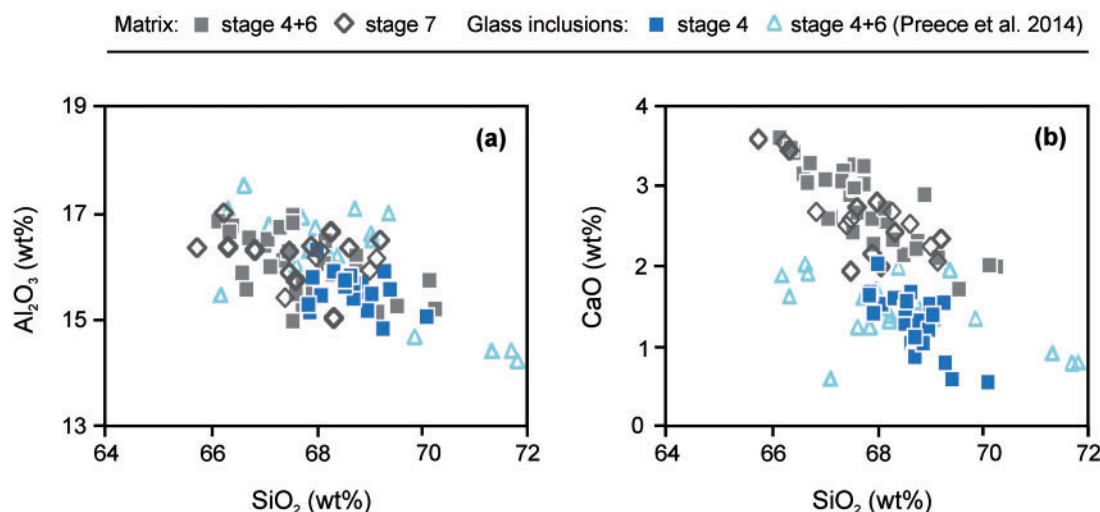
### Matrix and glass inclusion compositions

The matrix (glass plus microlites, analyzed with a  $50 \mu\text{m}$  beam diameter) of the stage 4 + 6 and the stage 7 dome clasts is dacitic with  $\sim 66$ – $70$  wt %  $\text{SiO}_2$ ,  $\sim 15$ – $17$  wt %  $\text{Al}_2\text{O}_3$ , and  $\sim 2$ – $4$  wt % CaO (Fig. 6). Glass inclusions in clinopyroxene phenocrysts have the same range of  $\text{SiO}_2$  and  $\text{Al}_2\text{O}_3$  contents as the matrix (Fig. 6a), but they have lower CaO contents of  $< 2$  wt %, indicating that they were modified by post-entrapment crystallization (Fig. 6b). Their volatile contents calculated by difference from electron microprobe totals are  $\sim 0.5$ – $2.5$  wt % for dome clasts and  $\sim 3$ – $4$  wt % for pumice samples.

## The experimental study

### Run products of the crystallization experiments

The seven crystallization experiments were carried out at  $925$ – $1000^\circ\text{C}$  and  $\sim 100$ – $400$  MPa (Table 3). Six experiments were performed at moderately oxidizing conditions of  $\sim \text{NNO} + 1.5$  to  $\sim \text{NNO} + 0.7$ ; one experiment was performed at  $> \text{NNO} + 2$  (CE-1) (Table 3). The crystallization of charges with an  $X_{\text{H}_2\text{O}_V}$  of 1 to  $\sim 0.5$  produced up to  $\sim 90$  vol. % of euhedral to subhedral crystals. Crystallinities of  $\leq 40$ – $60$  vol. %, which are characteristic for the natural samples, formed in charges crystallized at  $\leq 950^\circ\text{C}$  with an  $X_{\text{H}_2\text{O}_V}$  of  $\geq 0.7$  and in charges crystallized at  $\geq 975^\circ\text{C}$  with an  $X_{\text{H}_2\text{O}_V}$  of  $\leq 0.8$  (Table 3). Plagioclase (7–66 vol. %), clinopyroxene (0–18 vol. %), and magnetite (0–5 vol. %) occur in all run products except for three that crystallized at  $\geq 1000^\circ\text{C}$  and  $\geq 300$  MPa with an  $X_{\text{H}_2\text{O}_V}$  of 1. Orthopyroxene formed in charges crystallized at  $950^\circ\text{C}$  and  $100$  MPa with an  $X_{\text{H}_2\text{O}_V}$  of  $\leq 0.8$ . Amphibole and olivine, however, were not detected in any of the crystallization run products.



**Fig. 6.** Composition of matrix (glass plus microlites, analyzed with a  $50 \mu\text{m}$  beam) and glass inclusions in clinopyroxene phenocrysts of stage 4 + 6 and stage 7 dome clasts. Data shown are from this study and from Preece et al. (2014).

The experimental glasses have a basaltic andesitic to dacitic composition (Fig. 7a). Those that have formed in charges crystallized at 950°C and ~200 MPa with an  $X_{H_2O_V}$  of 0.7 and a melt  $H_2O$  content of ~3.6 wt % approach some of the least evolved matrix compositions of the natural samples with ~66 wt %  $SiO_2$  and ~16.5 wt %  $Al_2O_3$  (Fig. 7a). The experimental plagioclase compositions range from ~An50 to ~An90, compared with the overall more evolved rim compositions of the natural crystals with ~An40 to ~An90 (Fig. 7b). The plagioclase  $K_D$  (Ca–Na) values for crystal rims formed at 950°C and ~200 MPa (CE-7) increase from 3.6 to 5.9 with increasing melt  $H_2O$  and  $X_{H_2O_V}$  of the charges (Supplementary Data Electronic Appendix A3). Clinopyroxene Wo and En contents of the experimental and natural crystals largely match (Fig. 7c), but the  $Al_2O_3$  contents of the experimental crystals are high (>3–8 wt %) relative to the natural crystals (~1.2–7 wt %) (Fig. 7d and e). The clinopyroxene  $K_D$  (Fe–Mg) values for crystal rims formed at 950°C and ~200 MPa (CE-7) vary between 0.19 and 0.30, increasing with melt  $H_2O$  and  $X_{H_2O_V}$  (Supplementary Data Electronic Appendix A3). The experimental and the natural magnetite crystals are comparable in their FeO and  $TiO_2$  contents, whereas the  $Al_2O_3$  contents of the experimental crystals are relatively high (Fig. 7f). We note that crystallization at variable oxygen fugacities (i.e. at strongly oxidizing, >NNO + 2 conditions in CE-1 compared with all other experiments run at ~NNO + 1.5 to NNO + 0.7) affects the  $TiO_2$ , FeO, and MgO contents of clinopyroxene, magnetite, and glass, but that it does not account for their high-Al contents (Fig. 7).

### Run products of the melting experiments

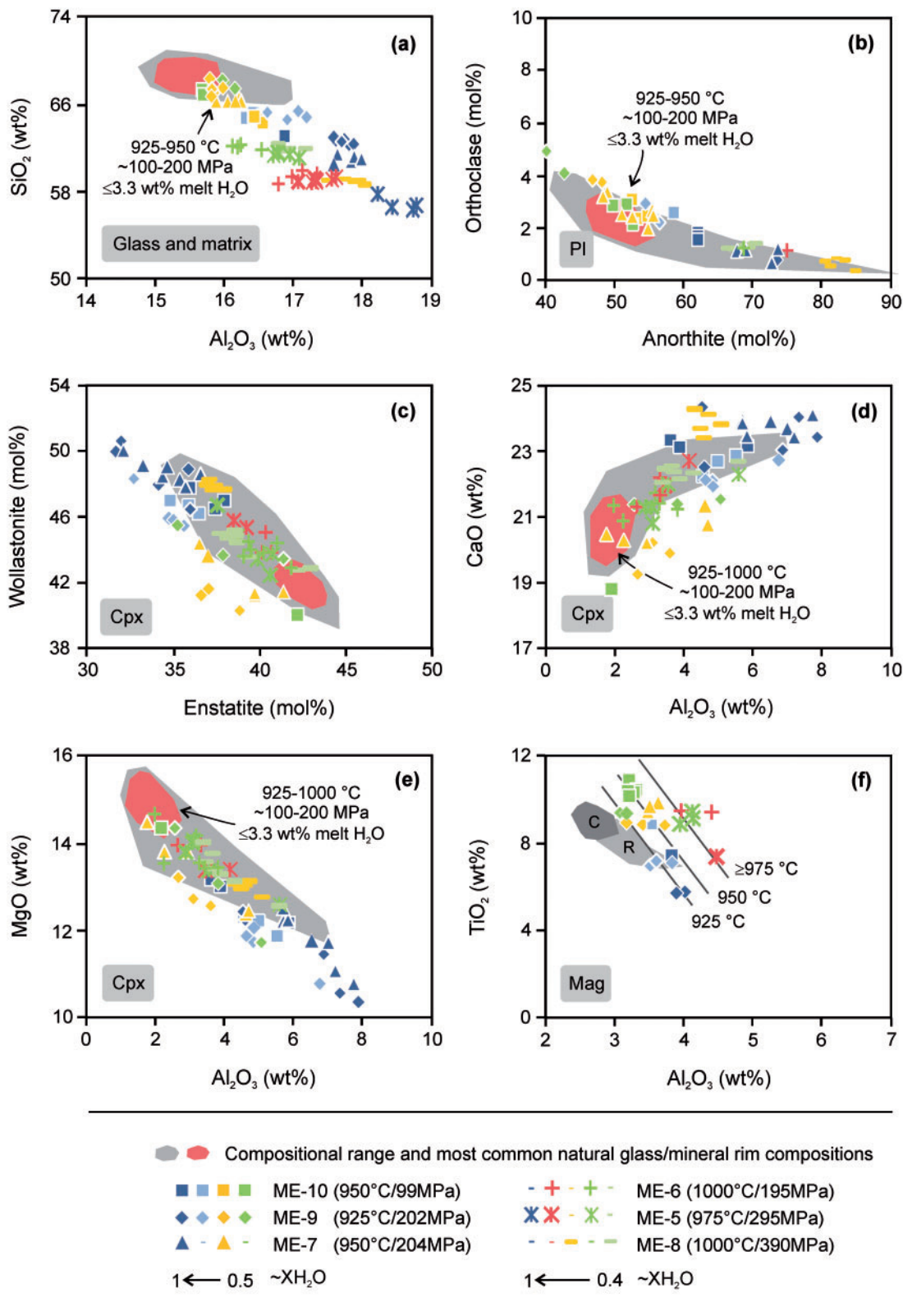
The six melting experiments were performed in conjunction with the crystallization experiments at 925–1000°C, ~100–400 MPa, and at ~NNO + 1.5 to ~NNO + 0.7 (Table 3). Partial melting and partial crystallization of the basaltic andesite rock powders in charges with an  $X_{H_2O_V}$  of ~0.4–1 resulted in run products with 0 to ~70 vol. % crystals (Table 3). Crystallinities of ~40–60 vol. %, which are characteristic for the natural samples, were obtained in charges recrystallized at ≤950°C with an  $X_{H_2O_V}$  of ≥0.6 and in charges run at ≥975°C with an  $X_{H_2O_V}$  of ≤0.7 (Table 3). Plagioclase (~7–66 vol. %), clinopyroxene (~0–18 vol. %), and magnetite (~0–5 vol. %) are present in all experimental products except for those of three charges run at 1000°C or ≥300 MPa and with an  $X_{H_2O_V}$  = 1. Amphibole occurs in charges with an  $X_{H_2O_V}$  of 1 and ~0.8 that were run at 925°C and ~200 MPa, forming clusters of randomly oriented crystals. Orthopyroxene is present in charges recrystallized at 950°C and ~100 MPa with an  $X_{H_2O_V}$  ≤0.8, forming coarse dendritic crystals. The run products of the ≤950°C, ≤200 MPa melting experiments thus reproduce the mineral assemblage of natural samples

from Merapi. Only olivine, which forms rare inclusions in natural amphibole phenocrysts, was not detected.

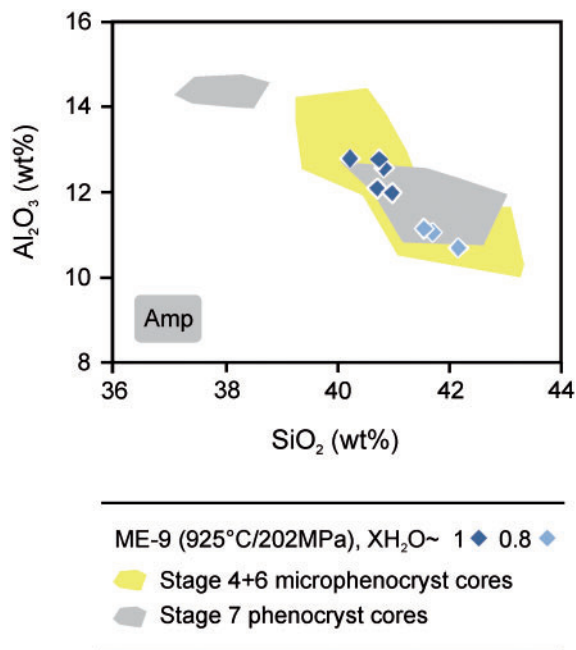
The glasses of the melting experiments are basaltic andesitic to dacitic; those that formed in charges run at ≤950°C and ≤200 MPa with an  $X_{H_2O_V}$  of ~0.5–0.6 match the moderately evolved natural matrix and glass inclusion compositions with ~68 and 16 wt %  $SiO_2$  and  $Al_2O_3$  (Fig. 8a). They have calculated  $H_2O$  and  $CO_2$  contents of ≤3.3 wt % and ≥470 ppm, respectively (Table 3). Glass produced in charges of our ≥975°C and ≥300 MPa experiments, however, deviate from the natural matrix and glass compositions (Fig. 8a). Their  $SiO_2$  and  $Al_2O_3$  contents are <62 and >16.5 wt %, and their calculated melt  $H_2O$  contents are ≥3.8 wt %. Plagioclase crystal rims produced in the experiments have ~An40–85 composition, and thus compare with the range of rim compositions detected for the natural samples (Fig. 8b). Their average  $K_D$  (Ca–Na) values vary between 3.0 and 6.7 (Supplementary Data Electronic Appendix A3). The ~An50 composition, which is most common in the natural crystal rims, is reproduced in charges run at ≤200 MPa with an  $X_{H_2O_V}$  of ~0.5–0.6 and a melt  $H_2O$  content of ≤3.3 wt % (Fig. 8b). Crystals with >An50 to ~An70 rims formed in charges with melt volatile and  $H_2O$  contents of ~3–5 wt % that were run at ≤200 MPa with an  $X_{H_2O_V}$  of ≥0.8 or at ~400 MPa with an  $X_{H_2O_V}$  of ~0.4. However, few cores and inner rims (≤5 vol. %) of newly crystallized plagioclase (<5–10 vol. %) in the low-pressure, ≤200 MPa experiments with a low  $X_{H_2O_V}$  of ≤0.7 also have >An50 to ~An80 compositions, whereas their  $K_D$  (Ca–Na) values are significantly (>2σ) higher than those of the newly crystallized outer rims in the same charges.

Clinopyroxene rim compositions have ~En32–42, ~Wo40–50, ~1.8–8 wt %  $Al_2O_3$ , ~19–23 wt % CaO, and ~10–15 wt % MgO. Their  $K_D$  (Fe–Mg) values vary between 0.16 and 0.38, decreasing with increasing experimental temperature and increasing with experimental melt  $H_2O$  and  $X_{H_2O_V}$ . Clinopyroxene with ~Wo42–43, ~En43, and ≤2 wt %  $Al_2O_3$  that matches the most common natural rim compositions was approached in charges with an  $X_{H_2O_V}$  of ≤0.6 and a melt  $H_2O$  content of ≤3.3 wt % at ≤200 MPa (Fig. 8c–e). Primitive natural clinopyroxene rim compositions, and in particular those with ≥46 Wo, ≥3 wt %  $Al_2O_3$  and ≥22 wt % CaO, are produced at ≤200 MPa in charges with an  $X_{H_2O_V}$  of ≥0.8 and at ≥300 MPa in charges with an  $X_{H_2O_V}$  of ≤0.7 (Fig. 8b–d). Some cores and inner rims of newly crystallized clinopyroxene (<5–10 vol. %) in our ≤950°C,  $X_{H_2O_V}$  ≤0.6 experiments, moreover, have high- $Al_2O_3$  (>2 to 6 wt %), low-CaO (<20 wt %) compositions that are unknown from the natural samples, but their modal proportion is low (<5–10 vol. %). Amphibole, which is a minor but characteristic constituent of Merapi basaltic andesites, formed in our melting experiments at 925°C and ~200 MPa, but was not detected in any of the higher temperature runs. Crystals from charges with an  $X_{H_2O_V}$  of ~0.8 and a melt  $H_2O$  content of ~4.0 wt %





**Fig. 8.** Composition of the melting experiment (ME) glasses and newly crystallized outer mineral rims compared with the natural phase assemblage: (a) glass compositions; (b-e) plagioclase and clinopyroxene rim compositions; (f) evolved magnetite core (C) and rim (R) compositions. The grey fields mark the range of the natural phase compositions; the red fields show the most common compositions. (a) Glass and matrix, (b) plagioclase, (c-e) clinopyroxene, and (f) magnetite compositions. All data are reported in [Supplementary Data Electronic Appendices A2 and A3](#). All experiments were performed at ~NNO + 1.5 to ~NNO + 0.7.



**Fig. 9.** Experimental and natural amphibole  $\text{SiO}_2$  and  $\text{Al}_2\text{O}_3$  core compositions. The experimentally produced amphibole compositions are closely comparable with those of the natural microphenocrysts, but they are significantly more evolved than the main amphibole phenocryst population with  $\sim 14$  wt %  $\text{Al}_2\text{O}_3$ .

have compositions comparable with those of the Al-poor ( $\sim 11$  wt %  $\text{Al}_2\text{O}_3$ ) natural microphenocryst cores (Fig. 9). Amphibole from charges with an  $\text{XH}_2\text{O}_V = 1$  and a melt  $\text{H}_2\text{O}$  content of  $\sim 4.9$  wt %, in contrast, approach the composition of the natural Al-rich ( $\sim 13$  wt %  $\text{Al}_2\text{O}_3$ ) microphenocryst cores (Fig. 9). Magnetite compositions that approach those of the evolved natural crystals form at  $925\text{--}950^\circ\text{C}$  (Fig. 8f). The most evolved compositions, which are preserved in the natural crystal cores, were approached in charges with an  $\text{XH}_2\text{O}_V$  of  $\leq 0.6$ , whereas the natural magnetite rim compositions are approached in charges with an  $\text{XH}_2\text{O}_V$  of  $\geq 0.7$  at  $925$  and  $950^\circ\text{C}$  (Fig. 8f). The natural  $\sim\text{En}67$  orthopyroxene compositions are reproduced at  $950^\circ\text{C}$  and  $\sim 100$  MPa in charges with an  $\text{XH}_2\text{O}_V$  of  $\leq 0.8$ .

## DISCUSSION

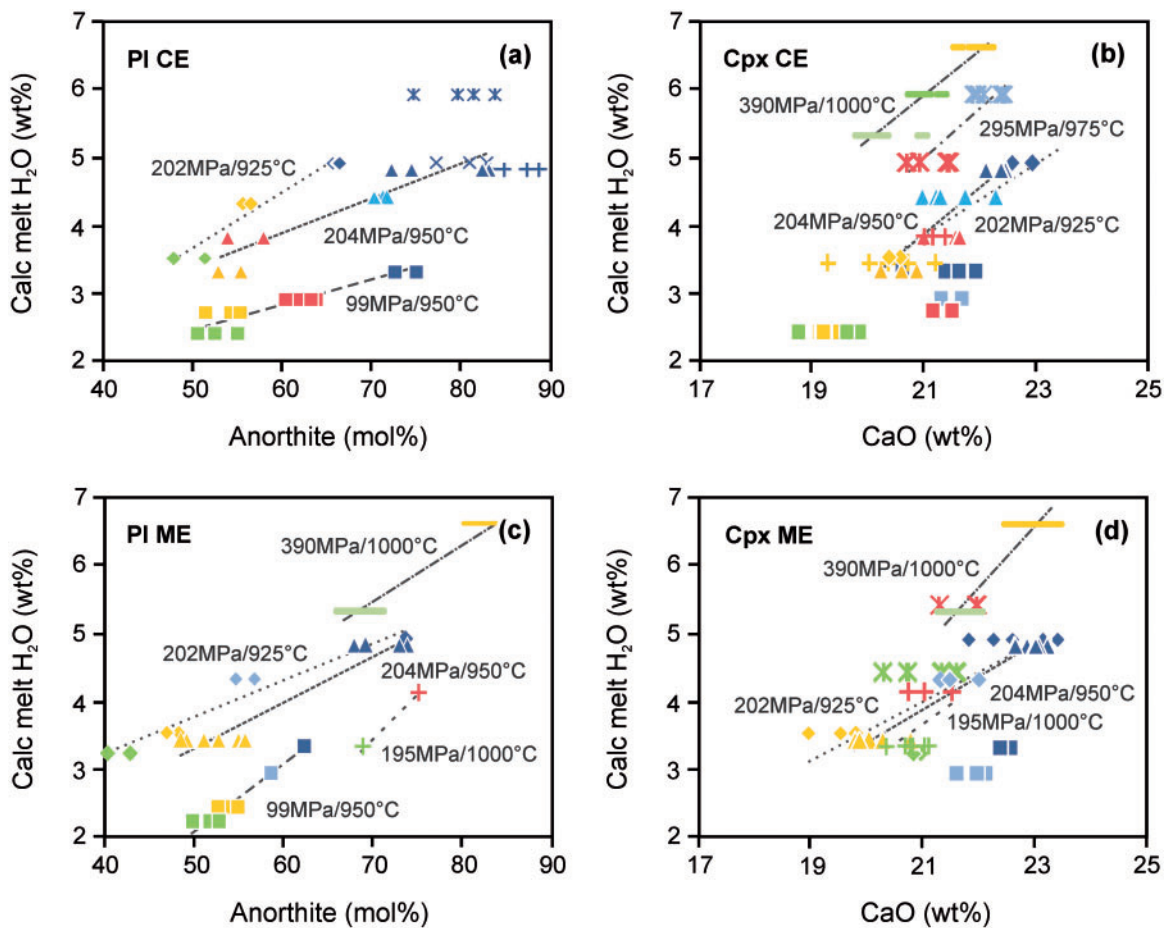
We first evaluate the conditions of equilibrium in our crystallization and melting experiments and their potential for constraining the crystallization conditions in Merapi's pre-eruptive magma reservoir. We then compare the natural and the experimentally produced mineral assemblages and compositions (1) to constrain the crystallization conditions of the resident magma, (2) to characterize the components and conditions of the pre-eruptive magma recharge, and (3) to evaluate the processes that control the variation in crystallization conditions. We focus our discussion on interpreting the evolution of magmas involved in Merapi's 2010 eruption, but note that whole-rock compositions, mineral

assemblages, and mineral and glass compositions are largely invariant in pyroclasts erupted between at least 1930 and 2010 (Berthommier, 1990; Abdurachman, 1998; Chadwick *et al.*, 2013; Costa *et al.*, 2013; Preece *et al.*, 2013). We therefore concur with previous studies that Merapi's shallow magma plumbing system has undergone no radical modifications in the last  $\sim 100$  years, whereas variations in extensive parameters (e.g. volumes of resident and recharge magmas or the amounts of carbonate assimilation) may have been important (e.g. Borisova *et al.*, 2013; Costa *et al.*, 2013; Troll *et al.*, 2013).

## Experimental constraints

### Results from the crystallization experiments

The run products of our crystallization experiments show systematic modal and compositional variation as a function of experimental conditions (Fig. 10a and b; Table 3; Supplementary Data Electronic Appendix A3). Plagioclase anorthite content increases with increasing melt  $\text{H}_2\text{O}$  content and with experimental temperature (Fig. 10a). Average plagioclase  $K_D$  (Ca–Na) values also systematically increase with increasing melt  $\text{H}_2\text{O}$  content from 3.7 to 6.4 for melt  $\text{H}_2\text{O}$  contents of  $\sim 3.6\text{--}5.9$  wt % (CE-7; Supplementary Data Electronic Appendix A3), and they are comparable with values determined by previous studies for compositionally comparable systems (e.g. Sisson & Grove, 1993; Pichavant *et al.*, 2002; Parman *et al.*, 2011). Clinopyroxene CaO content also increases with melt  $\text{H}_2\text{O}$  content, but decreases with increasing pressure, whereas it shows little variation as a function of experimental temperature (Fig. 10b). Average clinopyroxene  $K_D$  (Fe–Mg) values of between 0.20 and 0.30 decrease with temperature and increase with increasing melt  $\text{H}_2\text{O}$  content (Supplementary Data Electronic Appendix A3), and they also are comparable with values determined in other studies (e.g. Sisson & Grove, 1993; Pichavant *et al.*, 2002; Parman *et al.*, 2011). We suggest that both the systematic compositional variation of the crystallization experiment run products and partition coefficients that are comparable with literature values indicate a close approach to equilibrium in our experiments as intended (see Table 2). That the run products failed to reproduce the natural mineral and glass compositions could indicate that the temperature, pressure,  $\text{XH}_2\text{O}$ , or  $f\text{O}_2$  conditions did not match those of the natural system or that our starting material represents a non-liquid composition. We rule out the interpretation that the explored temperature, pressure, and  $\text{XH}_2\text{O}$  conditions are inadequate, because the crystallinities characteristic for the natural samples and the melt  $\text{H}_2\text{O}$  contents were well reproduced. We also rule out the interpretation that the moderately oxidizing crystallization conditions (at an  $f\text{O}_2$  of  $\sim\text{NNO} + 1.5$  to  $\sim\text{NNO} + 0.7$ ) are the cause of the compositional mismatch between the crystallization experiment run products and the natural samples, because the Fe–Mg–Ti compositions of the experimental and the natural clinopyroxene, magnetite, and glasses are well matched. The unmatched high Al



**Fig. 10.** Systematic compositional variation of plagioclase and clinopyroxene as a function of experimental conditions in the crystallization experiments (a, b; mineral core compositions) and melting experiments (c, d; outer mineral rim compositions). Symbols in (a) and (b) correspond to those of Fig. 7, and those in (c) and (d) correspond to those in Fig. 8. Compositional trends are indicated for charges with variable  $X_{H_2O}$  and melt  $H_2O$  contents crystallized at the indicated pressure and temperature conditions. That the run product compositions of the melting experiments are more weakly correlated than those of the crystallization experiments reflects our use of rock powder as a starting material, which introduced some compositional variation between and within the charges. Calc melt  $H_2O$  is melt  $H_2O$  content calculated using VolatileCalc of Newman & Lowenstern (2002).

contents of minerals and glasses, in contrast, are not affected by variations in  $fO_2$  (Fig. 7). We therefore suggest that our starting material represents a non-liquid composition; that is, we infer that the sample from which we fused our starting glass comprises accumulated cognate, antecrystic and xenocrystic components and mixed liquids, which is consistent with Merapi's inferred open-system evolution (e.g. Chadwick *et al.*, 2007, 2013; Nadeau *et al.*, 2011, 2013; Borisova *et al.*, 2013; Costa *et al.*, 2013; Van der Zwan *et al.*, 2013; Erdmann *et al.*, 2014). As a result, we cannot use the phase proportions or compositions from our crystallization experiments to constrain the conditions in Merapi's pre-eruptive magma reservoir; instead, we use their compositions below to interpret the mineral rim compositional variation detected in our melting experiments.

#### Results from the melting experiments

Compared with the crystallization experiments, melting experiments of the type performed have the advantage

that they partially preserve the original crystal assemblage that was present in the magmas stored within the pre-eruptive reservoir, including xenocrysts, antecrysts, and accumulated or mixed phenocryst populations. The experimental conditions closely match the crystallization conditions in the pre-eruptive reservoir if (1) all microlites that crystallized during magma ascent have melted and (2) all phenocrysts in the charges and their rim compositions match those of the natural samples. Melting experiments are thus suitable for characterizing the late-stage crystallization conditions in Merapi's crystal-rich pre-eruptive reservoir; however, it has to be kept in mind (and is considered further below) that the matrix of the starting rock powders may nevertheless contain components of the resident and the recharge melts. As in the crystallization experiments, the outer mineral rim and glass compositions produced in our melting experiments show systematic modal and compositional variations as a function of experimental conditions and partition coefficients comparable with those determined by previous studies (e.g. Sisson & Grove,



1993; Pichavant *et al.*, 2002; Parman *et al.*, 2011). Plagioclase anorthite content (Fig. 10c) and average  $K_D$  (Ca–Na) values (3.0–6.7) increase with increasing melt  $H_2O$  content,  $X_{H_2O_V}$ , and with experimental temperature (Supplementary Data Electronic Appendix A3). The detected compositional variation of the outer plagioclase rims in single charges is nearly always within  $\pm 2$ –3% (rarely within  $\pm 4$ %) of the average  $SiO_2$  and  $Al_2O_3$  contents, indicating that local equilibrium was approached throughout the charges. Clinopyroxene CaO contents increase with melt  $H_2O$  content and  $X_{H_2O_V}$  (Fig. 10d), whereas average  $K_D$  (Fe–Mg) values (0.15–0.38) decrease with increasing experimental temperature and increase with melt  $H_2O$  content and  $X_{H_2O_V}$  (Supplementary Data Electronic Appendix A3). The detected compositional variation of the outer clinopyroxene rims in single charges varies within up to  $\pm 5$ % of the average  $SiO_2$  and CaO contents (whereas minor components such as  $Al_2O_3$  have a variance of up to 50%). The major element compositional variation of clinopyroxene outer rims within single charges is larger than that of the outer plagioclase rims, but nevertheless is relatively homogeneous within charges. However, some of the cores and inner rims of newly crystallized plagioclase and clinopyroxene in low-pressure ( $\leq 200$  MPa), low- $X_{H_2O_V}$  ( $\leq 0.7$ ) experiments clearly record partial disequilibrium. The  $>An_{50}$  to  $\sim An_{80}$  compositions and disequilibrium  $K_D$  values of inner zones of new plagioclase, and equilibrium  $K_D$  (Fe–Mg) values, but high- $Al_2O_3$  ( $>2$  to 6 wt %), low-CaO ( $<20$  wt %) compositions of inner zones of new clinopyroxene, differ by  $>2\sigma$  from the compositions of outer crystal rims (which is unknown from the natural samples). They are interpreted to record early stages of experimental crystallization probably during heating and at high fluid/melt ratios, and they are therefore not considered in the subsequent discussion. Their effect on the composition of the reactive system is nevertheless considered to be limited, given that they have a low proportion of  $<5$ –10 vol. % of all newly crystallized plagioclase and clinopyroxene. That the mineral rim and glass compositions of the melting experiments nevertheless follow compositional trends that are comparable with those of the crystallization experiments (Fig. 10c and d versus Fig. 10a and b), that plagioclase–melt and clinopyroxene–melt partition coefficients are comparable with equilibrium values, and that outer rim compositions within single charges have relatively homogeneous compositions (Supplementary Data Electronic Appendix A3) is taken as evidence that the experiments approached local equilibrium conditions as intended.

In the following sections, we therefore use the mineral rim and glass compositions produced in the melting experiments to infer the crystallization conditions of the resident magma in Merapi's pre-eruptive reservoir, and to approximate the conditions of the pre-eruptive recharge. We also compare the crystallinities between the natural and the experimental samples, but not their mineral proportions, because we cannot accurately

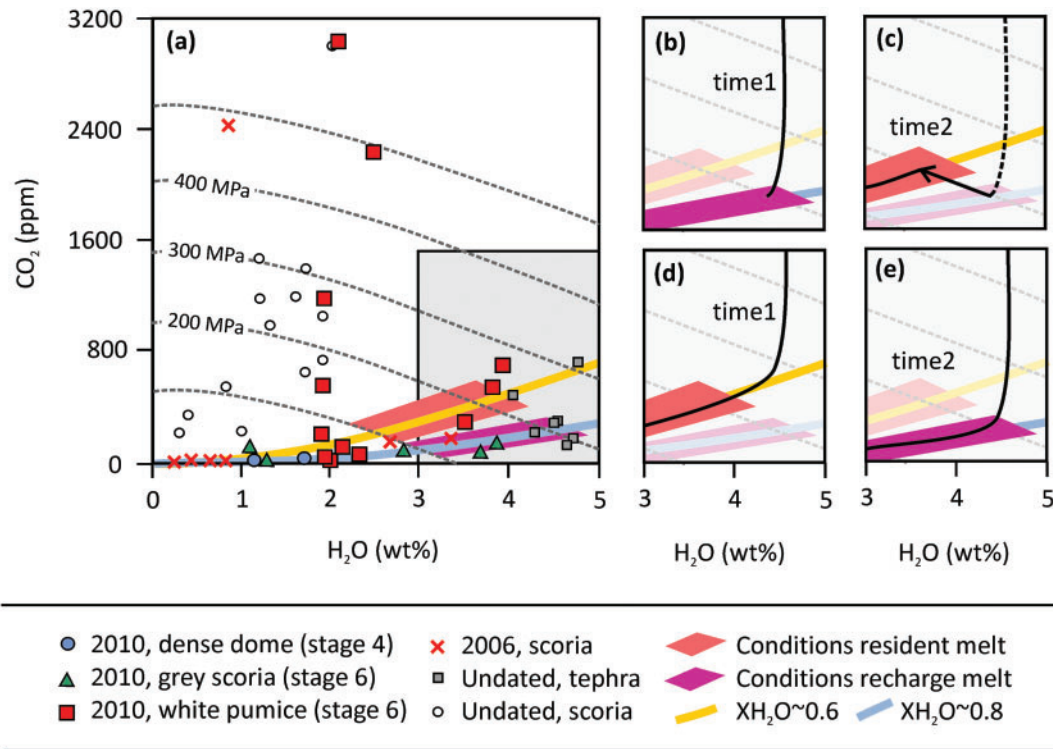
determine the volume of, for example, co-crystallizing plagioclase and clinopyroxene in the natural basaltic andesites (this would require extensive compositional mapping). We suggest that the conditions of the resident magma can be closely constrained by this approach, because the compositionally primitive mineral zones (and microlites) that crystallized from the recharge magma remain unmelted. We further suggest that the compositional variation detected in our melting experiments can be used to closely approximate the recharge crystallization conditions in Merapi's pre-eruptive reservoir, because the compositional difference between the resident melt (inferred to correspond to the evolved matrix and glass inclusion compositions with  $\sim 68$  wt %  $SiO_2$  and  $\sim 15.5$  wt %  $Al_2O_3$ ; Fig. 6a) and the recharge melt (inferred to approach the primitive matrix and glass inclusion compositions with  $\leq 66$  wt %  $SiO_2$  and  $\geq 17$  wt %  $Al_2O_3$ ; Fig. 6a) is limited and the effect on our estimate is thus minor. Previous crystallization experiments with basaltic and basaltic andesitic starting materials with a compositional difference of  $\leq 3$  wt %  $SiO_2$  and  $\leq 1$  wt %  $Al_2O_3$  but otherwise identical crystallization conditions, for example, yielded equivalent plagioclase and clinopyroxene compositions in charges in which both minerals co-crystallized [e.g. 79-35g # 12 and 82-66 #5 of Sisson & Grove (1992) or SAR3 and SAR16 of Parman *et al.* (2011)]. We therefore infer that the results of our melting experiments can be used for estimating both the intensive parameters of the resident magma and those of the recharge magma. We acknowledge that estimating the crystallization conditions on the basis of our melting experiments is more uncertain than for estimates based on crystallization experiments (owing to our use of a starting rock powder that introduced some compositional variation between charges), but that uncertainties are lower than the increments at which we have varied the intensive parameters (i.e. up to  $\pm 25^\circ C$  and  $\pm 75$  MPa). Uncertainties in the calculated melt  $H_2O$  contents (Table 3) are  $\pm 1$  wt %, whereas those of the calculated melt  $CO_2$  contents (Table 3) are large, up to  $\pm 1000$  ppm (see Baker & Alletti, 2012). Fluctuations of  $H_2O$  and  $CO_2$  in Merapi's pre-eruptive reservoir, however, are inferred on the basis of mineral compositional variations, and not on the basis of the calculated  $CO_2$  values.

### Shallow, $H_2O$ -poor pre-eruptive magma storage

Consistent with previous studies, we suggest that the pre-eruptive magma storage conditions in Merapi are recorded by the compositionally evolved mineral zones and by the evolved matrix (glass plus microlites) and glass inclusion compositions in the erupted, crystal-rich basaltic andesites (Table 4; see also Costa *et al.*, 2013; Nadeau *et al.*, 2013; Erdmann *et al.*, 2014). Crystallinities of  $>40$  to  $\sim 60$  vol. % and the predominant,  $\sim An_{50}$  natural plagioclase and Al-poor clinopyroxene compositions are simultaneously reproduced in our melting experiments performed at  $925$ – $950^\circ C$ ,  $\sim 100$ – $200$  MPa,

**Table 4:** Components and inferred crystallization conditions for Merapi's magma plumbing system

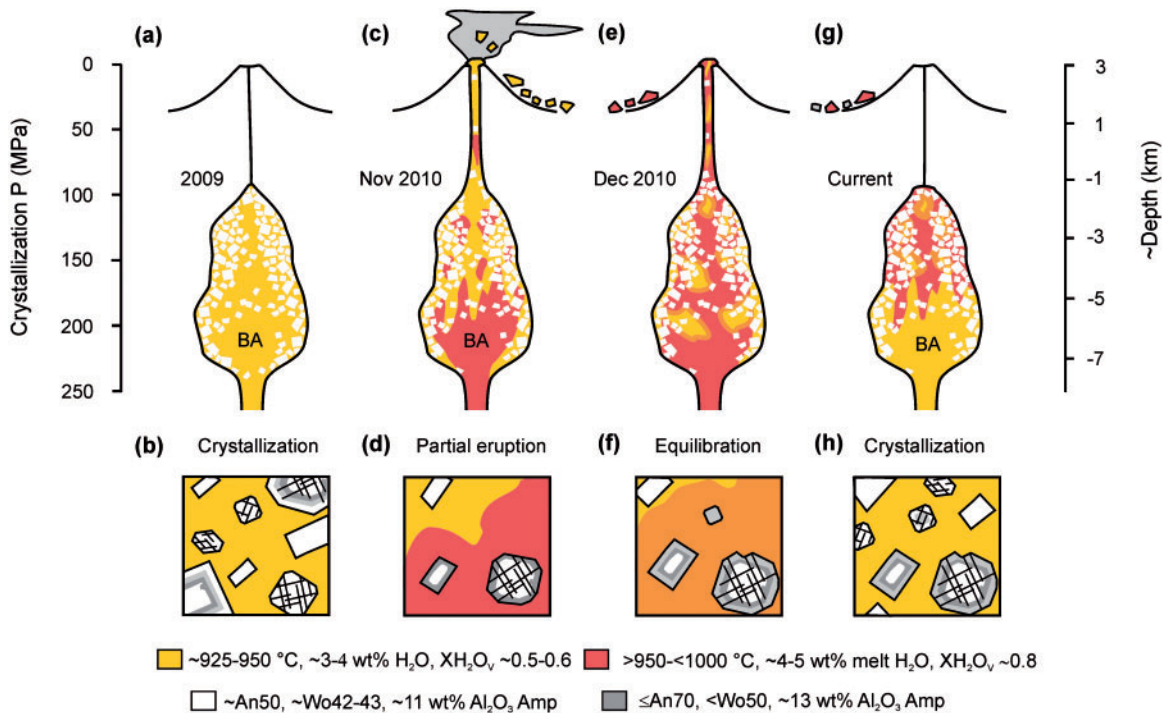
Reservoir: Magma:	Pre-eruptive (main) Resident	Pre-eruptive (shallow) Resident	Pre-eruptive Recharge	Deep, unconstrained Recharge
<i>Components</i>				
Pl	~An50	~An40	≤An70	~An90
Cpx	~Wo42–43	≤Wo42	≤Wo50	~Wo48
Amp	~11 wt % Al <sub>2</sub> O <sub>3</sub>	—	~13 wt % Al <sub>2</sub> O <sub>3</sub>	~14 wt % Al <sub>2</sub> O <sub>3</sub>
Mag	~3 wt % Al <sub>2</sub> O <sub>3</sub>	~3 wt % Al <sub>2</sub> O <sub>3</sub>	≥3.5 wt % Al <sub>2</sub> O <sub>3</sub>	~6 wt % Al <sub>2</sub> O <sub>3</sub>
Opx	—	~En67	—	—
Ol	—	—	—	>Fo60–70
Melt	~68–69 wt % SiO <sub>2</sub>	~69–70 wt % SiO <sub>2</sub>	~66 wt % SiO <sub>2</sub>	?
<i>Conditions</i>				
Melt H <sub>2</sub> O content	~3–4 wt %	~3 wt %	~4–5 wt %	?
XH <sub>2</sub> O <sub>v</sub>	~0.5–0.6 ± 0.1	~0.5–0.6 ± 0.1	~0.8	?
Temperature	~925 to 950 ± 25°C	~925 to 950 ± 25°C	>950 to <1000°C	?
Pressure	~200 ± 75 MPa	≥100 MPa	~200 ± 75 MPa	?



**Fig. 11.** Glass inclusion compositions, estimated resident and recharge melt H<sub>2</sub>O–CO<sub>2</sub> contents, and inferred volatile evolution. (a) Glass inclusion compositions from Nadeau *et al.* (2013) and Preece *et al.* (2014); redrawn from Preece *et al.* (2014) with our estimates for resident and recharge melt compositions and vapour isopleths at an XH<sub>2</sub>O<sub>v</sub> of ~0.6 and ~0.8. (b–d) Insets shown in (a). (b, c) Scenario 1: Merapi's pre-eruptive reservoir is filled by magmas with an XH<sub>2</sub>O<sub>v</sub> of ~0.8 (b, time 1) that become relatively H<sub>2</sub>O-poor as a result of CO<sub>2</sub>-rich volatile fluxing during magma storage (c, time 2). (d, e) Scenario 2: the reservoir is filled by magmas with an XH<sub>2</sub>O<sub>v</sub> of ~0.6 during inter-eruptive periods (c, time 1) and by magmas with an XH<sub>2</sub>O<sub>v</sub> of ~0.8 during pre-eruptive periods (d, time 2). We favour the interpretation of Scenario 2, proposing that magmas ascending during inter-eruptive periods (at time 1) approach a vapour-buffered degassing path, whereas magmas ascending during pre-eruptive periods (at time 2) approach a closed-system degassing path.

with an XH<sub>2</sub>O<sub>v</sub> of ~0.5–0.6, and at a melt H<sub>2</sub>O content of <3.3 wt % (Table 3; Fig. 8b, d and e). The evolved matrix and glass compositions of the natural samples with ~68–69 wt % SiO<sub>2</sub> and ~15.5 wt % Al<sub>2</sub>O<sub>3</sub> are also closely approached at these experimental conditions (Fig. 8a). Their calculated melt H<sub>2</sub>O and CO<sub>2</sub> contents of ≤3.3 wt % and ~240–590 ppm, moreover, match the volatile contents of some of the natural glass inclusions analyzed by Preece *et al.* (2014) (Fig. 11). Amphibole with compositions approaching those of the evolved

microphenocrysts of our stage 4 + 6 samples formed in charges run at 925°C with an XH<sub>2</sub>O<sub>v</sub> of ~0.8 and an estimated melt H<sub>2</sub>O content of ~4.0 wt %, but not at 950°C. Magnetite comparable with the evolved core composition of the natural samples also formed in charges run at 925°C (Fig. 8f). In combination, the mineral and glass compositions and crystallinities indicate that Merapi's pre-eruptive reservoir crystallizes at predominantly ≤200 MPa, at a temperature of ~950°C and ≥925°C, and with a melt H<sub>2</sub>O content of ~3–4 wt % (Table 4; Fig. 12a



**Fig. 12.** Sketch summarizing the estimated crystallization conditions for Merapi's pre-eruptive magma storage reservoir, its location and the main processes proposed for its evolution prior to recent historical eruptions, showing a timeline leading up to the 2010 eruption. (a, b) During inter-eruptive storage, Merapi's basaltic andesitic magmas crystallize an evolved mineral assemblage. (c, d) During pre-eruptive magma recharge, the reservoir is replenished by magma with an H<sub>2</sub>O-rich melt and vapour phase, crystallizing mineral zones of intermediate composition. (e, f) The intensive parameters of magma parcels of the original resident and the recharge magma equilibrate. (g, h) Inter-eruptive recharge by magma with an H<sub>2</sub>O-poor melt and vapour phase begins. The inferred time scales of this evolution need to be assessed by detailed mineral composition diffusion modelling. We further note that the geometry of the magma storage zone is unconstrained by our data, except that we infer that the erupted magma and its phenocrysts must have collected over a pressure range of ~100 MPa, equivalent to a depth range of ~4-5 km. BA = basaltic andesite.

and b). Experimental orthopyroxene of composition ~En68, which closely corresponds to the natural ~En67 crystal compositions, crystallized at 950°C and ~100 MPa in equilibrium with an XH<sub>2</sub>O<sub>v</sub> of ~0.6 and a melt H<sub>2</sub>O content of ~2.4 wt %, further indicating that Merapi's pre-eruptive reservoir extends to levels shallower than a depth equivalent to 200 MPa (Tables 3 and 4; Fig. 12). Consistent with the experimental assemblage, we further propose that the natural plagioclase crystals with compositions of ~An40-50 and clinopyroxene crystals with ≤Wo42 and ≤19.5 wt % CaO have formed in this shallow reservoir region, in agreement with the interpretations of Costa *et al.* (2013). The melt H<sub>2</sub>O and CO<sub>2</sub> contents of ~2.4 wt % and ~240 ppm calculated for these conditions also match those detected in several natural glass inclusions (see Preece *et al.*, 2014). That the Fe, Mg, and Ti contents of glass, clinopyroxene, orthopyroxene, and magnetite produced in charges run at an *f*O<sub>2</sub> of ~NNO + 1.0 and an XH<sub>2</sub>O<sub>v</sub> of ≤0.6 replicate the natural mineral and glass compositions indicates moderately oxidizing crystallization conditions (Table 3).

We emphasize that the evolved mineral and glass inclusion compositions characteristic of Merapi basaltic andesites were in equilibrium and that they all record

crystallization in the pre-eruptive reservoir. Previous studies, in contrast, have suggested that crystals and glass inclusions record magma evolution at deep and shallow levels, respectively, and thus that their record of crystallization conditions is largely decoupled (Nadeau *et al.*, 2013; Preece *et al.*, 2014). Our estimated magma storage temperature of ≥925-950 ± 25°C is consistent with the crystallization temperatures calculated using evolved amphibole compositions and the Ridolfi *et al.* (2010) or the Ridolfi & Renzulli (2012) calibrations (i.e. ~920-970 ± 50°C; Costa *et al.*, 2013; Nadeau *et al.*, 2013; Erdmann *et al.*, 2014). Calculated clinopyroxene crystallization temperatures of ~1025 to 1060 ± 59°C (Preece *et al.*, 2014) using the thermometer of Putirka *et al.* (2003) are comparatively high, but the lower estimated crystallization temperatures are within the error of our estimate. The inferred shallow pre-eruptive storage at ≥100-200 ± 75 MPa indicates that Merapi's pre-eruptive reservoir is located at a depth of ≥4.5 to ~9 ± 3 km, and thus within the carbonate-dominated upper crust (see Smyth *et al.*, 2005), if we assume an average overlying crustal density of 2242 kg m<sup>-3</sup> (see Tiede *et al.*, 2005) and lithostatic pressure conditions. If the reservoir was under significant overpressure (see Costa *et al.*, 2013; Genareau *et al.*, 2014), then we infer

that it is located at even shallower levels (e.g. at  $\geq 2.5$  to  $7 \pm 3$  km for an overpressure of  $\leq 50$  MPa). A shallow, ephemeral magma storage zone at  $\sim 1.5$ – $2.5$  km below Merapi's summit, which has been proposed (Ratdomopurbo & Poupinet, 1995, 2000; Saepuloh *et al.*, 2010; Nadeau *et al.*, 2013) but also contested on the basis of geophysical data and petrological data (Beauducel & Cornet, 1999; Müller & Haak, 2004; Costa *et al.*, 2013), is not evident from our results. Our pressure estimates, however, are consistent with seismic and ground deformation data, which suggest magma storage at depths of  $\geq 5.5$  to  $\sim 8.5$  km below Merapi's summit, from at least 1992 to 2010 (Ratdomopurbo & Poupinet, 1995, 2000; Beauducel & Cornet, 1999; Budi-Santoso *et al.*, 2013). Our results are also consistent with glass inclusion barometry, which has been used to infer re-equilibration of magmas at  $\leq 265$  MPa and depths of  $\leq 10$  km (Preece *et al.*, 2014), but they differ significantly from previous estimates based on amphibole and clinopyroxene barometry, which have suggested a main zone of pre-eruptive magma storage at  $> 100$ – $400$  or at  $\sim 400$ – $500$  MPa (Gertisser, 2001; Chadwick *et al.*, 2013; Costa *et al.*, 2013; Preece *et al.*, 2014). This result is not surprising, considering the uncertainties of the barometric methods employed (e.g. Fig. 1; Nimis, 1999; Putirka, 2008; Erdmann *et al.*, 2014). Crystallization pressure calculated for clinopyroxene using the Nimis (1999) barometer, which is not recommended for calc-alkaline compositions, yields moderate to high calculated pressures of  $\sim 400$ – $500$  MPa for Merapi's main clinopyroxene crystal population if the crystallization temperature is moderately overestimated [i.e. at  $1000^\circ\text{C}$  in estimates of Chadwick *et al.* (2013)], and lower calculated pressures of  $\sim 200$ – $400$  MPa if the crystallization temperature is largely overestimated [i.e. at  $1050^\circ\text{C}$  in estimates of Preece *et al.* (2014)]. For clinopyroxene that crystallized from more evolved magmas than the Merapi basaltic andesites, even higher pressure overestimates would result (Fig. 1a). That a large range in crystallization pressure is calculated with the Nimis (1999) calibration (e.g. Chadwick *et al.*, 2013) is inevitable if the natural clinopyroxene crystals formed over a range of temperatures, but average clinopyroxene crystallization temperatures are assumed (e.g. over- and underestimates of crystallization temperatures by  $20^\circ\text{C}$  propagate into a 100 MPa decrease or increase in the calculated pressure; Nimis, 1999). The clinopyroxene crystallization pressure calculated for Merapi's magma system using the geobarometer equations (32b) and (32c) of Putirka (2008) yields pressure estimates of predominantly  $\sim 200$ – $300$  and  $\sim 100$ – $200$  MPa (Preece *et al.*, 2014). These agree well with our estimate, reflecting the fact that the barometers yield values that approach known crystallization pressures for clinopyroxene crystallized from mafic-intermediate magmas at moderately low temperatures (Fig. 1b and c). We point out, however, that the values calculated for Merapi are (1) based on largely overestimated

clinopyroxene crystallization temperatures of  $1025$ – $1058^\circ\text{C}$ , which increases the calculated pressure values, and (2) based on estimated primitive melt composition that corresponds to Merapi basaltic andesite whole-rock compositions (instead of dacitic compositions as recorded by glass inclusions), which decreases the calculated pressure values. The pressure estimates for the Merapi system based on the amphibole barometers of Ridolfi *et al.* (2010) and Ridolfi & Renzulli (2012) yield minimum values of  $\sim 250$ – $400$  MPa, reflecting that the method calculates intermediate pressures for amphibole crystallized in equilibrium with dacitic melts (Fig. 1d; Erdmann *et al.*, 2014).

We emphasize that using single mineral thermometers or barometers to estimate the crystallization conditions of mixed magma systems has no advantage over experimental constraints, because any compositional variations in the magma and melt that complicate the experimental estimates equally affect the thermobarometric estimates. It is, moreover, important to keep in mind that errors on the calculated conditions may significantly exceed the standard errors of the methods, if the employed mineral thermometers or barometers were calibrated based on compositions or intensive parameters (e.g.  $P$ ,  $T$ ,  $f\text{O}_2$ , or  $X\text{H}_2\text{O}$ ) unlike those of the investigated system.

The geometry of Merapi's shallow magma system cannot be constrained on the basis of our results or previous barometric estimates, but we emphasize that the erupted magma and its phenocryst cargo was assembled over a significant pressure and depth range of  $\sim 100$  MPa and  $\sim 4$ – $5$  km. The pre-eruptive magma reservoir outlined in Fig. 12 highlights this depth range, but we cannot resolve if the magma storage zone is a single reservoir or a number of shallow sub-reservoirs connected by dykes as proposed by Chadwick *et al.* (2013) and Nadeau *et al.* (2013). An elliptical, subhorizontal magma source at a depth of  $\sim 8$ – $5$  km was predicted for the 1996–1997 eruption on the basis of global positioning system (GPS) deformation and tilt data (Beauducel & Cornet, 1999); this is more consistent with a plumbing system consisting of several sub-reservoirs, but remains to be confirmed. In the future, currently installed continuous GPS and tiltmeters (CVGHM/BPPTK and DOMERAPI projects) will provide tighter constraints.

### Pre-eruptive, $\text{H}_2\text{O}$ -rich magma recharge

The recharge of Merapi's magma storage reservoir by a larger than average volume of volatile-rich mafic magma has been suggested in several previous studies to have played a key role in triggering and driving its paroxysmal 2010 eruption (e.g. Budi-Santoso *et al.*, 2013; Costa *et al.*, 2013; Jousset *et al.*, 2013; Preece *et al.*, 2013, 2014). Petrological studies have considered phenocrysts with primitive compositions and An-rich microlites as evidence for this recharge, but have

pointed out that the juvenile materials erupted in 2010 all have a homogeneous basaltic andesite composition (Costa *et al.*, 2013; Preece *et al.*, 2014; Fig. 2). More important is that the stage 7 samples, which have a higher proportion of pristine phenocrysts with primitive core and rim compositions that record the pre-eruptive recharge (Figs 3 and 4), also are closely comparable in composition with the stage 4 + 6 samples. We therefore suggest that the recharge magma of the 2010 event was basaltic andesitic and close in composition to the resident magma, at least in terms of major elements. This is important, because it indicates that magmas crystallize to a crystal-rich mush and mix in Merapi's shallow pre-eruptive reservoir, but that they do not significantly fractionate at this level. If any significant differentiation takes place, it must be at deeper levels of the magma plumbing system.

The most direct evidence for magma recharge and mixing within Merapi's pre-eruptive reservoir is the assemblage of normally zoned plagioclase, clinopyroxene, amphibole and magnetite phenocrysts with primitive core compositions [i.e.  $\sim$ An90,  $\sim$ Wo48,  $\sim$ 14 wt %  $\text{Al}_2\text{O}_3$  amphibole, and  $\sim$ 6 wt %  $\text{Al}_2\text{O}_3$  magnetite; Table 4; see also Costa *et al.* (2013) and Nadeau *et al.* (2013)]. The evolved  $\sim$ Fo60–70 composition, low NiO (<0.25 wt %) and high CaO (>0.1 wt %) and MnO (>1 wt %) contents of rare olivine inclusions are characteristic for crystallization from basaltic andesitic magmas at <1000°C (e.g. Sisson & Grove, 1992; Moore & Carmichael, 1998), and we thus interpret them as recorders of magma evolution in Merapi's magma storage system and not as mantle-derived xenocrysts. We further consider the Al-poor ( $\sim$ 12 wt %  $\text{Al}_2\text{O}_3$ ) amphibole phenocrysts of the stage 7 samples as part of the primitive recharge assemblage, suggesting that they crystallized by replacing early stage olivine phenocrysts; this explains their common olivine inclusions, patchy zoning, and their high  $\text{SiO}_2/\text{Al}_2\text{O}_3$  contents as compared with the predominant population of Al-rich amphibole phenocrysts. Similar textures have been reported, for example, from andesites formed by mixing of evolved and primitive magmas (Foley *et al.*, 2013; Kiss *et al.*, 2014). The characteristic  $\sim$ An90 plagioclase and  $\sim$ Wo48 clinopyroxene core compositions of Merapi's phenocrysts are reproduced in our  $\text{H}_2\text{O}$ -rich charges crystallized at 400 MPa and 1000°C, whereas amphibole, magnetite, or olivine that are part of the recharge assemblage did not form in the experiments. Whether this indicates that they crystallized from a magma that differed in composition from our starting material (e.g. from a basaltic magma or from one that was uncontaminated by calc-silicate components), or whether they crystallized at pressure and temperature conditions that were not explored by us (e.g. at >400 MPa or at  $\leq$ 400 MPa and <1000°C) needs to be established in future studies. The low proportion ( $\leq$ 15 % of all phenocrysts) of the primitive, normally zoned recharge phenocrysts in Merapi's basaltic andesites is notable, but is comparable with the proportion of

recharge phenocrysts observed in other compositionally homogeneous, crystal-rich magmas such as those that have erupted in recent decades at Mount St Helens (Berlo *et al.*, 2007; Streck *et al.*, 2008; Cashman & Blundy, 2013).

We consider it possible that some of the An-rich plagioclase and Ca-rich clinopyroxene crystals in the Merapi magmas may be derived from the disaggregation or the assimilation of calc-silicate xenoliths, as has been inferred on the basis of plagioclase  $^{87}\text{Sr}/^{86}\text{Sr}$  compositions (Chadwick *et al.*, 2007), plagioclase and pyroxene  $\delta^{18}\text{O}$  compositions (Troll *et al.*, 2013), calcic glass inclusions in some phenocrysts (Borisova *et al.*, 2013; see also Deegan *et al.*, 2010), and whole-rock trace element,  $^{87}\text{Sr}/^{86}\text{Sr}$ , and  $\delta^{18}\text{O}$  compositions (Chadwick *et al.*, 2007; Borisova *et al.*, 2013; Troll *et al.*, 2013). We note, however, that calc-silicate xenoliths are rare in the Merapi basaltic andesites erupted in 2010 ( $\ll$ 1 vol. %) and that clinopyroxene approaching wollastonite in composition has not been detected in any significant proportion (Borisova *et al.*, 2013; Costa *et al.*, 2013; Preece *et al.*, 2014; our data), whereas it forms one of the main constituents of locally occurring calc-silicate xenoliths (e.g. Chadwick *et al.*, 2007; Troll *et al.*, 2013; our unpublished data). More important is that Merapi's basaltic andesites have 'typical' CaO contents (e.g. <10 wt % at  $\sim$ 52 wt %  $\text{SiO}_2$ ; Fig. 2d) compared with those of other calc-alkaline basaltic andesites (e.g. Ambrym, Picard *et al.*, 1994; Soufrière Hills, Murphy *et al.*, 2000; Tatara San Pedro volcanic complex, Dungan *et al.*, 2001; Mont Pelée, Pichavant *et al.*, 2002). Small amounts (<2 wt %) of carbonate assimilation, as inferred by Chadwick *et al.* (2007), are consistent with the CaO contents and other compositional and mineralogical features of the Merapi magmas. Large amounts of inferred carbonate assimilation of up to  $\sim$ 20 or 15–40 wt % (see Borisova *et al.*, 2013; Troll *et al.*, 2013) should, however, clearly result in strongly CaO-enriched whole-rock compositions, as Costa *et al.* (2013) have previously shown.

Volumetrically more abundant than the primitive phenocrysts are intermediate zones and rims on phenocrysts of the resident magma and rims on the recharge magma phenocrysts with  $>$ An50 to  $\sim$ An70 and  $>$ Wo43 to  $\sim$ Wo50 composition, and amphibole microphenocrysts with  $\sim$ 13 wt %  $\text{Al}_2\text{O}_3$  cores (Fig. 6). We infer that they crystallized in equilibrium with a melt that was close in composition to the least evolved matrix and glass inclusions in Merapi basaltic andesites with <67 and probably  $\geq$ 65 wt %  $\text{SiO}_2$  (Fig. 7a). We rebut the interpretation that the intermediate plagioclase and clinopyroxene compositions and the Al-rich amphibole microphenocrysts represent crystallization at elevated pressures, involving large-scale convection and down-cycling, because this process is unlikely to produce the following features: (1) the commonly observed saw-tooth-like zoning patterns in plagioclase and the characteristic compositional plateau zones of plagioclase and clinopyroxene; (2) the texturally and compositionally

bimodal amphibole population; (3) the mineral rims with intermediate compositions (they should have developed evolved overgrowths during their return flow to shallower levels of the reservoir). The crystallization of the  $\leq$ An70 plagioclase,  $\leq$ Wo50 clinopyroxene, and amphibole microphenocrysts with  $\sim$ 13 wt %  $\text{Al}_2\text{O}_3$  may thus record crystallization in the pre-eruptive reservoir following magma recharge at an elevated temperature and/or at an elevated melt  $\text{H}_2\text{O}$  content.

Plagioclase with a characteristic  $\sim$ An70 composition, which matches the compositions common in intermediate zones and rims of the natural crystals, is produced in our melting experiments at  $1000^\circ\text{C}$ , and at the same 200 MPa pressure,  $\text{XH}_2\text{O}_V$  of  $\sim$ 0.5–0.6, and a melt  $\text{H}_2\text{O}$  content of  $\sim$ 3–4 wt % that we infer for the pre-eruptive crystallization conditions of the resident magma (Table 3; Fig. 8b). Magnetite with  $\geq$ 3.5 wt %  $\text{Al}_2\text{O}_3$ , which matches the rim composition of the natural crystals, also forms at these conditions (Fig. 8f). However, the Wo and  $\text{Al}_2\text{O}_3$ -rich compositions of clinopyroxene or amphibole that are characteristic of the recharge magma do not crystallize at these conditions (Fig. 8c and d). That the  $\sim$ An70 plagioclase and  $\leq$ Wo50 clinopyroxene, Al-rich amphibole microphenocrysts, and Al-rich magnetite rims record magma recharge with an increase in temperature alone therefore seems unlikely. In contrast, the assemblage with  $\sim$ An70 plagioclase, clinopyroxene with  $\leq$ Wo50 and  $\geq$ 4 wt %  $\text{Al}_2\text{O}_3$ , amphibole with  $\sim$ 13 wt %  $\text{Al}_2\text{O}_3$ , and magnetite with  $\geq$ 3.5 wt %  $\text{Al}_2\text{O}_3$  is produced in our melting experiments at  $925\text{--}950^\circ\text{C}$ ,  $\sim$ 200 MPa, an  $\text{XH}_2\text{O}_V$  of  $\geq$ 0.8, and with a melt  $\text{H}_2\text{O}$  content of  $\sim$ 4–5 wt % (Fig. 8b–f). The most primitive natural matrix and glass inclusion compositions with  $\sim$ 66 wt %  $\text{SiO}_2$  and  $\sim$ 17 wt %  $\text{Al}_2\text{O}_3$  are also closely approached at these conditions (Figs 6 and 8a). In combination, we therefore take the intermediate plagioclase, clinopyroxene, amphibole, and magnetite compositions to estimate that the  $\text{XH}_2\text{O}_V$  and melt  $\text{H}_2\text{O}$  content of the recharge magma were  $\geq$ 0.8 and  $\sim$ 4–5 wt %, and we infer that its temperature was  $>950$  and  $<1000^\circ\text{C}$  (Fig. 12c and d).

It is notable in this regard that  $\text{H}_2\text{O}$ -rich glass inclusions in natural samples from Merapi, which are inferred to represent, or closely approach, entrapment compositions, have up to  $\sim$ 5 wt %  $\text{H}_2\text{O}$  and  $\leq$ 800 ppm  $\text{CO}_2$  (Preece *et al.*, 2014). These inclusions with  $\geq$ 2.5 wt %  $\text{H}_2\text{O}$  have compositions indicating equilibration with a vapour phase characterized by an  $\text{XH}_2\text{O}_V$  of between  $\sim$ 0.6 and  $\sim$ 0.8 (Fig. 11a; data from Preece *et al.*, 2014). This variation in glass  $\text{H}_2\text{O}$  and  $\text{CO}_2$  contents and  $\text{XH}_2\text{O}_V$  could reflect the following: (1) analytical uncertainties; (2) analysis of compositionally heterogeneous glass domains and/or micron or submicron volatile inclusions [as proposed by Preece *et al.* (2014) for their  $\text{CO}_2$ -rich, but not for their  $\text{H}_2\text{O}$ -rich glass inclusion data]; (3) melt inclusions (glass) that variably re-equilibrated (leaked) during magma ascent. However, the compositional variation may also record the primary variations in melt composition and  $\text{XH}_2\text{O}_V$  that we infer on the

basis of plagioclase and clinopyroxene compositions; that is, melt entrapment during inter-eruptive magma residence in equilibrium with an  $\text{XH}_2\text{O}_V$  of  $\sim$ 0.6 and melt entrapment during pre-eruptive magma recharge in equilibrium with an  $\text{XH}_2\text{O}_V$  of  $\sim$ 0.8 (Fig. 12).

### Origin of the inferred volatile variation in Merapi's pre-eruptive magma reservoir

Our experimental study indicates that the Merapi pre-eruptive magma reservoir is contained within carbonate-dominated strata; previous experimental studies have characterized the release of  $\text{CO}_2$  by carbonate assimilation in Merapi magmas (Deegan *et al.*, 2010; Blythe *et al.*, 2015). High  $\delta^{13}\text{C}$  ( $\text{CO}_2$ ) gas emissions from summit fumaroles during Merapi's eruption in 2006 have been interpreted as evidence for carbonate assimilation (Troll *et al.*, 2012), which at that time may have contributed significant amounts of  $\text{CO}_2$  to Merapi's volatile budget. High  $\text{CO}_2$  emissions prior to the 2010 eruption may also reflect  $\text{CO}_2$  liberated by carbonate assimilation (Borisova *et al.*, 2013; Blythe *et al.*, 2015), but they may equally reflect  $\text{CO}_2$  released from the recharge magmas (Suroño *et al.*, 2012; Costa *et al.*, 2013). Phenocryst glass inclusions with high  $\text{CO}_2$  contents may furthermore record  $\text{CO}_2$  fluxing related to carbonate assimilation (Deegan *et al.*, 2010; Nadeau *et al.*, 2013; Blythe *et al.*, 2015), but the locally high  $\text{CO}_2$  contents of glass inclusions trapped in Merapi's phenocrysts are more likely to represent heterogeneous post-entrapment  $\text{CO}_2$  distribution (Preece *et al.*, 2014). In agreement with others, we therefore propose that Merapi's volatile phase comprises contributions from the carbonate-dominated upper crust, but that its largely uniform composition predominantly derives from a steady-state magma source comprising mantle and subducted sediment components (Allard, 1983, 2013; Allard *et al.*, 2011; Troll *et al.*, 2012).

That gas fluxing or degassing within Merapi's magma plumbing system accounts for most of the variations in  $\text{XH}_2\text{O}_V$  in the pre-eruptive reservoir is hence the most likely scenario. One possibility is that magmas reach Merapi's pre-eruptive reservoir with an  $\text{XH}_2\text{O}_V$  of  $\geq$ 0.8 (Fig. 12). During storage they are then fluxed by  $\text{CO}_2$ -rich volatiles from deeper levels of the magma system, decreasing the reservoir's  $\text{XH}_2\text{O}_V$  to  $\sim$ 0.5–0.6 (Fig. 12c). Such  $\text{CO}_2$ -rich volatile fluxing of shallow reservoirs has been inferred for a number of arc volcanoes, including Mount St Helens and Soufrière Hills (Blundy *et al.*, 2010; Cashman & Blundy, 2013; Cassidy *et al.*, 2015; Riker *et al.*, 2015). For the Merapi system this process would require that  $\text{CO}_2$  fluxing of the pre-eruptive reservoir is spatially fairly homogeneous, given that the largest phenocryst and microphenocryst volume crystallizes at these conditions. The characteristic plateau zones of clinopyroxene and plagioclase phenocrysts would further require that the reservoir's  $\text{XH}_2\text{O}_V$  reaches a steady state at  $\sim$ 0.5–0.6, which could indicate that the inflowing vapour is close to this composition or

that the influx of CO<sub>2</sub>-rich vapour diminishes or ceases once the volatile phase of the pre-eruptive reservoir has reached this composition. In this scenario, Merapi's deeper magma plumbing system behaves as an open system, releasing volatiles, whereas the pre-eruptive reservoir behaves as a closed or semi-closed system during inter-eruptive periods, trapping some (or the majority) of the volatiles derived from deeper levels (Fig. 12c).

Another possible scenario is that magmas ascending through Merapi's magma supply system follow different degassing paths—open, vapour-buffered and closed-system degassing paths—during inter-eruptive and pre-eruptive recharge of the reservoir. During inter-eruptive periods, magmas reach the pre-eruptive reservoir with an XH<sub>2</sub>O<sub>V</sub> of ~0.5–0.6 (Fig. 12), following or approaching a vapour-buffered degassing path, along which both H<sub>2</sub>O and CO<sub>2</sub> partially exsolve. During pre-eruptive recharge, in contrast, magmas ascend into Merapi's pre-eruptive reservoir following or approaching an open-system degassing path, releasing a comparatively large amount of CO<sub>2</sub>, and thus reaching the reservoir at a higher XH<sub>2</sub>O<sub>V</sub> of ~0.8 (Fig. 12d). A comparable, although longer-term switch from closed-system to vapour-buffered degassing paths has been inferred for magmas that sourced the 1980 Mount St Helens' plinian eruption and its subsequent effusive 1980–1986 eruptions on the basis of both mineral assemblages and glass inclusion CO<sub>2</sub> and H<sub>2</sub>O compositions (Blundy *et al.*, 2010; Riker *et al.*, 2015). That the supply system of Merapi's pre-eruptive reservoir has switched between degassing trends in such a fashion prior to 2010 and other recent eruptions (as recorded by equivalent mineral compositions and zoning patterns) could relate to changes in the magma ascent and supply rate, which ultimately trigger Merapi's small-volume eruptions. We consider this scenario plausible, because it is consistent with all the petrological and compositional data available, and because it can also account for the strong degassing of CO<sub>2</sub> observed prior to the onset of the 2010 eruption (e.g. Surono *et al.*, 2012). However, additional glass and fluid inclusion data and analyses of inter-eruptive and pre-eruptive gas emission data are urgently needed to further evaluate the inferred variation of volatile compositions and the proposed scenarios for volatile evolution.

## CONCLUSIONS

The key findings of our study on pyroclasts erupted at Merapi, Indonesia, in 2010, and our experiments on one of its crystal-rich basaltic andesites are as follows.

1. Experiments that partially melt and partially recrystallize coarse-grained powders can constrain the crystallization conditions for both resident and recharge magmas in mixed systems by reconstructing local equilibrium conditions between mineral rim and host melt composition, if both magmas

and their melt fractions are closely comparable in composition. The performance of standard crystallization experiments that reconstruct mineral and melt compositions at total equilibrium for the same starting material, and thus for a system with a similar composition, can be used to ascertain that the phase compositions produced in the melting experiments indeed represent local equilibrium.

2. Mount Merapi's current pre-eruptive reservoir hosts crystal-rich magma that is repeatedly recharged by basaltic andesite magma with a closely corresponding composition. The magmas partially crystallize at  $\geq 100\text{--}200 \pm 75$  MPa and thus at a depth between *c.* 4.5 and 9 km. They are stored at  $\sim 925\text{--}950 \pm 25^\circ\text{C}$  with melt H<sub>2</sub>O contents of ~3–4 wt %, and an XH<sub>2</sub>O<sub>V</sub> of ~0.5–0.6. Recharge magmas that mix with the resident magma of the reservoir have higher melt H<sub>2</sub>O contents of ~4–5 wt %, a higher XH<sub>2</sub>O<sub>V</sub> of ~0.8, and a higher temperature of >950 to <1000°C. These variations in melt H<sub>2</sub>O and XH<sub>2</sub>O<sub>V</sub> of the resident and recharge magmas may reflect variable degrees of open-system degassing of the magmas en route to the pre-eruptive magma storage reservoir, which may in turn record variations in magma ascent and supply rates from the deep plumbing system.

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## SUPPLEMENTARY DATA

Supplementary data for this paper are available at *Journal of Petrology* online.

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