

# Evidence for the development of permeability anisotropy in lava domes and volcanic conduits



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## ABSTRACT

The ease at which exsolving volatiles can migrate through magma and outgas influences the explosivity of a volcanic eruption. Volcanic rocks often contain discrete discontinuities, providing snapshots of strain localisation processes that occur during magma ascent and extrusion. Whether these features comprise pathways for or barriers to fluid flow is thus of relevance for volcanic eruption and gas emission modelling. We report here on nine discontinuity-bearing andesite blocks collected from Volcán de Colima, Mexico. We present a systematic porosity and permeability study of fifty cores obtained from the blocks collected, and interpret the genetic processes of the discontinuities through detailed microstructural examination. Bands in pumiceous blocks were inferred to be relicts of inhomogeneous bubble expansion which, despite significantly increasing porosity, do not markedly affect permeability. Other discontinuities in our blocks are interpreted to be shear strain-induced flow banding, cavitation porosity, and/or variably healed fractures. In each of these cases, an increase in permeability (up to around three orders of magnitude) was measured relative to the host material. A final sample contained a band of lower porosity than the host rock, characterised by variably infilled pores. In this case, the band was an order of magnitude less permeable than the host rock, highlighting the complex interplay between dilatant and densifying processes in magma. We therefore present evidence for significant permeability anisotropy within the conduit and/or dome of a volcanic system. We suggest that the abundance and distribution of strain localisation features will influence the escape or entrapment of volatiles and therefore the evolution of pore pressure within active volcanic systems. Using a simple upscaling model, we illustrate the relative importance of permeable structures over different lengthscales. Strain localisation processes resulting in permeability anisotropy are likely to play an important role in the style, magnitude, and recurrence interval of volcanic eruptions.

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## 1. Introduction

Terrestrial volcanism is the surface expression of buoyancy-driven ascent of magma through the Earth's crust. Throughout its journey from depth to the surface, processes such as decompression (e.g. Proussevitch and Sahagian, 1998) and thermal vesiculation (Lavallée et al., 2015) force the exsolution (degassing) of magmatic volatiles such as H<sub>2</sub>O and CO<sub>2</sub> (e.g. Liu et al., 2005). Whether through an interconnected bubble network in the magma (e.g. Eichelberger et al., 1986; Okumura et al., 2009), laterally into the edifice (e.g. Jaupart and Allègre, 1991; Jaupart, 1998; Collombet, 2009), or through fracture networks in magma, the edifice rock, and lava domes (e.g. Stasiuk et al.,

1996; Gonnermann and Manga, 2003; Rust et al., 2004; Edmonds and Herd, 2007; Castro et al., 2012; Cabrera et al., 2011; Lavallée et al., 2013; Pallister et al., 2013; Gaunt et al., 2014), the capacity for gases to migrate in a volcanic system is thought to influence its explosive potential. Effective outgassing is predicted to lessen the explosive potential of a volcano by inhibiting the build-up of pore overpressures within magma, whereas the ascent of poorly outgassed magmas is generally assumed to culminate in catastrophic explosive eruptions (as discussed by many authors, e.g. Eichelberger et al., 1986; Woods and Koyaguchi, 1994; Rust et al., 2004; Edmonds and Herd, 2007; Mueller et al., 2008, 2011; Nguyen et al., 2014; Castro et al., 2014; Okumura and Sasaki, 2014; Gaunt et al., 2014). The genesis and longevity of outgassing pathways are thus expected to constitute a critical parameter dictating eruptive behaviour.

Magma is inherently heterogeneous. The physico-chemical properties of magma evolve in time and space due to a host of interrelated

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processes, such as crystallisation (e.g. Caricchi et al., 2007; Vona et al., 2011; Chevrel et al., 2013, 2015), vesiculation (e.g. Bagdassarov and Dingwell, 1992; Burgisser et al., 2010; Okumura et al., 2013), and chemical differentiation (e.g. Giordano et al., 2008). In turn these processes are affected by variations in shear stress and strain rates across the conduit (Papale, 1999; Gonnermann and Manga, 2003; Caricchi et al., 2007) or during dome extrusion (e.g. Smith et al., 2001; Buisson and Merle, 2002; Cashman et al., 2008). Evidence for this variability is abundant in volcanic environments, where extruded material can show wide ranges in crystal size and assemblage (Cashman, 1992; Blundy et al., 2006) and porosity and pore diameter (Kueppers et al., 2005; Mueller et al., 2005, 2008; Shea et al., 2010; Lavallée et al., 2012; Farquharson et al., 2015; Burgisser et al., 2011), in turn influencing permeability and strength (e.g. Zhu et al., 2011; Kolzenburg et al., 2012; Kendrick et al., 2013; Farquharson et al., 2015; Heap et al., 2015a; Schaefer et al., 2015; Lavallée et al., in press). Magmas can also preserve discrete heterogeneities, evidence for strain localisation, in erupted material (e.g. Tuffen and Dingwell, 2005). Strain localisation processes in magma include extensional fracturing (e.g. Heiken et al., 1988; Stasiuk et al., 1996; Tuffen et al., 2003; Tuffen and Dingwell, 2005; Castro et al., 2014), frictional melting (e.g. Kendrick et al., 2014a, 2014b; Plail et al., 2014), brecciation and gouge formation during shear fracturing (e.g. Pallister et al., 2013; Cashman and Cashman, 2006; Kennedy and Russell, 2012), cavitation (e.g. Smith et al., 2001), and flow banding (e.g. Gonnermann and Manga, 2003, 2005).

Structural or textural discontinuities in porous media—especially where these heterogeneities are manifest as discrete tabular or (sub) planar features, such as stylolites and compaction bands—have often been observed to comprise conduits for or barriers to fluid flow, or more complex combined architectures (Caine et al., 1996; Baud et al., 2012; Lavallée et al., 2013; Heap et al., 2014c; Gaunt et al., 2014; Deng et al., 2015). The nature of these heterogeneities—their genesis, abundance, size (thickness and lateral extent), and orientation—can have a significant influence on the physical properties of the containing medium. For example, the scale and ratio of strain distribution and strain localisation in crustal fault zones have a marked effect on their permeability, resulting in four end-member regimes: localised conduits, localised barriers, distributed conduits, and combined conduit-barrier systems (e.g. Chester and Logan, 1986; Bruhn et al., 1990; Caine et al., 1996; Schultz and Fossen, 2008; Bense and Person, 2006). Due to the pre-established importance of fluid transport in volcanic systems, as well as the prevalence of preserved heterogeneities in erupted materials, it is of significance whether discontinuities in magma can in fact act as barriers to or conduits for fluid flow. To date there are few studies that explore the influence of magmatic heterogeneities on permeability, and even fewer that offer laboratory data. For example, Bouvet de Maisonneuve et al. (2009) present measurements on banded and non-banded pumice, showing that the former tends to be markedly less permeable in their dataset (by over an order of magnitude, depending on the orientation). Cabrera et al. (2015), Castro et al. (2012), and Berlo et al. (2013) discuss the permeability of rhyolite containing tuffsite (defined as fractures in magma or rock infilled with transported juvenile clasts and lithic fragments, e.g. Saubin et al., in press). Kolzenburg et al. (2012); Lavallée et al. (in press), and Kendrick et al. (2016) provide laboratory and field permeability measurements of andesite containing tuffsites and fractures. These studies conclude that the permeability of these features is—at least transiently—higher than that of the host rock mass, implying that they may serve as preferential routes for the outgassing of magmatic volatiles. Similarly, Gaunt et al. (2014) report a strong permeability anisotropy between the central and peripheral conduit of Mount St Helens volcano (USA) due to the juxtaposition of discrete lithofacies: a result of inhomogeneous ascent-driven strain localisation in the magma. Kendrick et al. (2014a) show that pseudotachylites (frictional melts) from Soufrière Hills volcano (Montserrat) are notably less porous and permeable relative to the

host rock and therefore may act as barriers to fluid flow in volcanic systems.

This study comprises a systematic examination of the porosity and permeability of discontinuity-bearing andesites to determine whether they comprise effective pathways for—or indeed barriers to—fluid flow in a volcanic system. We first show and describe nine discontinuity-bearing andesite blocks collected during a recent field campaign (May–June 2014) at Volcán de Colima (Mexico). We display the array of observed microstructures corresponding to these features, before outlining some potential mechanisms for their genesis. The influence of these discontinuities on the physical rock properties is then shown, and the attendant implications for volcanic activity, modelling, and field interpretations are outlined. The dataset of this study complements the large pool of data previously obtained on rocks from Volcán de Colima without discontinuities (e.g. Mueller, 2006; Kolzenburg et al., 2012; Lavallée et al., 2012; Kendrick et al., 2013; Richard et al., 2013; Heap et al., 2014b; Farquharson et al., 2015; Lavallée et al., in press).

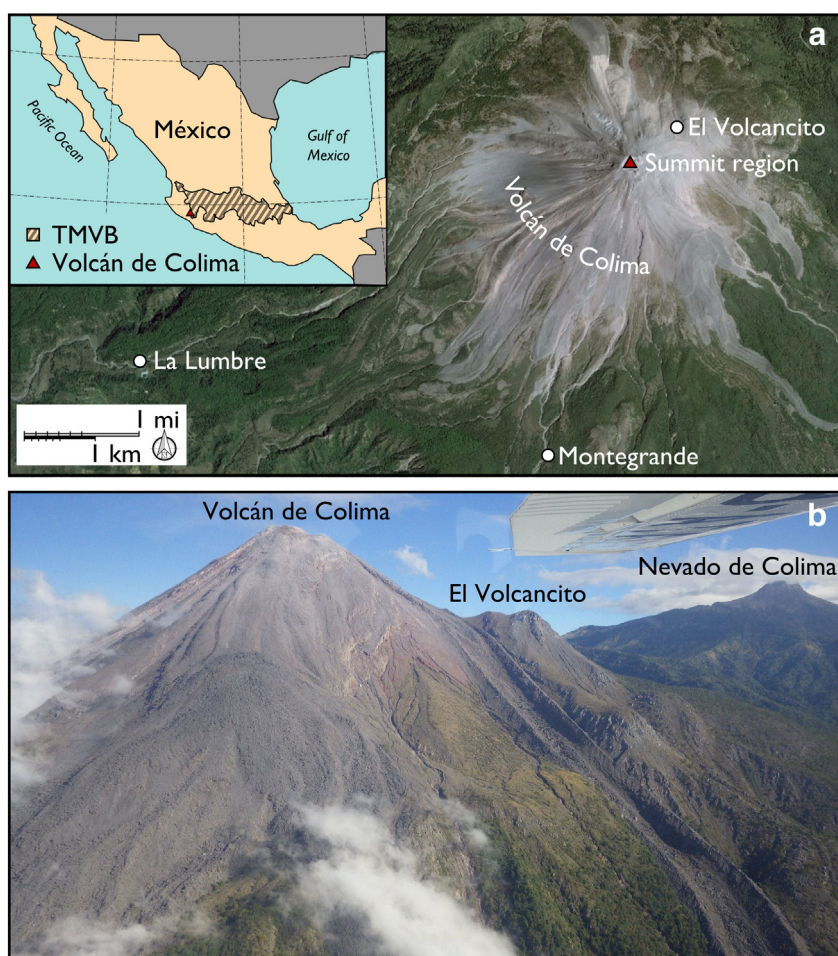
## 2. Case study: Volcán de Colima

Volcán de Colima, located at the western margin of the Trans-Mexican Volcanic Belt (Fig. 1a), is an active andesitic stratovolcano displaying a range of eruptive styles. Activity at the volcano underwent a significant increase from 1998, with two eruptive periods being defined: 1998–2011 and 2013–current. During these eruptions, periods of lava flow emission and dome extrusion have been typically interspersed with cycles of explosive activity (Varley et al., 2010a, 2010b; James and Varley, 2012; Lavallée et al., 2012). A longer term cycle has been identified culminating in larger-scale sub-Plinian or Plinian events approximately once every century (Luhr, 2002). At the time of writing, Volcán de Colima had recently produced its largest pyroclastic density current (PDC) since the last catastrophic event in 1913 (Capra et al., 2015). During 10–11 July 2015, a significant increase in the effusion rate promoted a partial collapse of the growing dome, along with a portion of the crater rim on two occasions. A PDC, resulting from this activity, travelled down the Montregrande ravine and reached a distance of 10.7 km from the volcano summit. Widespread ashfall and the possibility of further pyroclastic density currents and lahars prompted the evacuation of a number of communities in the surrounding area, as well as a pre-emptive cessation of air traffic from Colima airport (La Coordinación Nacional de Protección Civil, 2015a, 2015b). In February 2016, an overflight of the volcano confirmed that a new lava dome is growing within the summit crater, signalling renewed effusive activity.

The rocks that erupted since 1998 have been remarkably homogeneous in terms of their chemical composition, being andesitic with a SiO<sub>2</sub> content typically between 58 and 61 wt.% (e.g. Reubi and Blundy, 2008; Savov et al., 2008). Despite its petrographic constancy, the recent eruptive diversity of Volcán de Colima has given rise to a broad array of erupted material: field studies (Mueller, 2006; Farquharson et al., 2015; Lavallée et al., in press) have found porosities of edifice-forming material to range between 0.02 and 0.73 (i.e. 2 and 73%). Similarly, the permeability of this material has been observed to differ by up to four orders of magnitude for any given porosity, and almost seven orders of magnitude over the whole dataset (Mueller, 2006; Kolzenburg et al., 2012; Kendrick et al., 2013; Richard et al., 2013; Heap et al., 2014b; Farquharson et al., 2015). During a recent field campaign (May–June 2014), discontinuity-bearing rocks were collected from three sample sites (shown in Fig. 1a): two debris-flow tracks on the southwest and south flanks of the volcano (La Lumbre and Montregrande respectively), and the flanks of the parasitic vent Volcancito (Fig. 1b).

## 3. Materials and methods

Nine blocks were sampled from the sites shown in Fig. 1a. These blocks were collected so as to represent a variety of observable planar



**Fig. 1.** Location of Volcán de Colima in Mexico. (a) Google Earth™ satellite image of Volcán de Colima, showing each of the sample sites Volcancito, Montegrande, and La Lumbre (approximately 0.8, 3.5, and 6.7 km from the active vent, respectively). Inset shows the geographic location of Volcán de Colima within the Trans-Mexican Volcanic Belt (TMVB). (b) An aerial view of the main summit of Volcán de Colima, also showing the parasitic vent Volcancito, and the older, extinct member of the volcanic complex, Nevado de Colima.

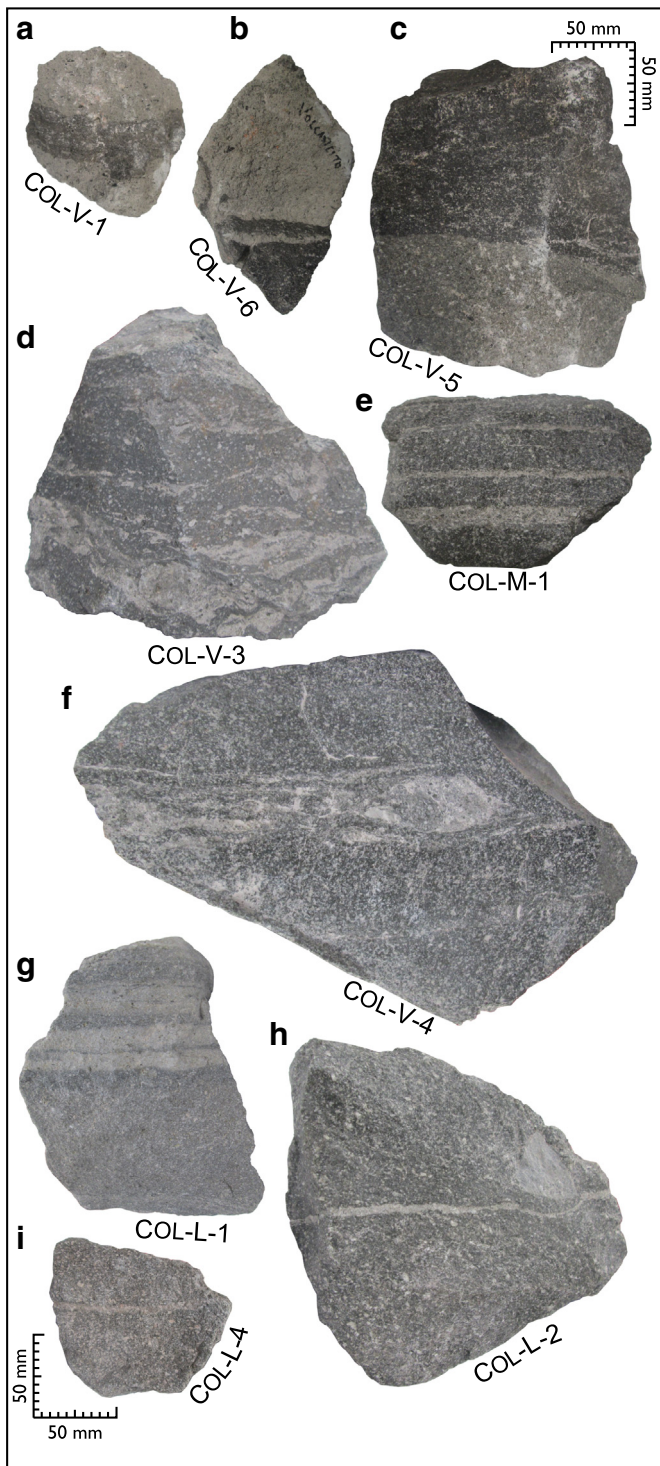
and subplanar heterogeneities (although this range is not necessarily exhaustive with respect to the features discernible in the field). Whilst we cannot offer a quantitative assessment of the abundance of these various features, we highlight that they were commonly observable at each of the sampling sites and elsewhere on the volcano. We also draw attention to the fact that the sites marked in Fig. 1a correspond to the final location of each block after emplacement and potential remobilisation. As such, we cannot draw specific inferences as to their origin based on the sampling location. Furthermore, we acknowledge that this sampling method means that neither the *in situ* orientation of the blocks nor their emplacement conditions—i.e. extrusion rate, style of activity etc.—can be assessed. The blocks collected for the purposes of this study were generally smaller than 300 mm in length, but we highlight that similar discontinuities can be seen on a metre scale. The nine blocks are shown and described in Fig. 2 and Table 1, respectively.

Cylindrical core samples with a diameter of 20 mm were prepared from each block and precision-ground so that their end faces are flat and parallel. The length of each core (from 25 to 40 mm) was dependent on the initial block size. The length to width ratio of our samples was therefore greater than one in each case (length to width ratios lower than one are not recommended for laboratory permeability measurements). Where possible, cores were extracted such that the feature of interest was either parallel or perpendicular to the core axis (i.e. in the *y* and *z* directions, respectively: Fig. 3a). Again where possible, additional cores of the host rock were obtained so as not to include the feature of

interest. Altogether, 50 samples were prepared from the nine initial blocks (shown in Appendix A). All subsequent measurements were performed after the cores had been oven-dried at 40 °C under a vacuum for a minimum of 48 h.

Connected gas porosity ( $\phi$ ) was determined using helium pycnometry (AccuPyc II 1340 from Micromeritics). Gas permeability  $k$  was measured under 1 MPa of confining pressure using a steady-state benchtop permeameter, with nitrogen gas as the permeant. Whilst we appreciate that measurement under a low confining pressure may not necessarily represent the *in situ* conditions, one must note that imposing higher pressure requires presupposing a depth of origin for each sample. Similarly, whilst room-temperature experiments may not capture additional complexities of fluid migration in high-temperature conduit magma, attempting to account for this in the experimental design means presuming a single elevated temperature at which to measure permeability. Importantly, the imposed experimental conditions (fixed pressure and temperature) allow us to compare the permeability of the different samples measured in this study. A schematic of the permeameter is shown in Fig. 3b and a close-up of the coreholder setup in Fig. 3c. The volumetric flow rate was measured at a range of pressure differentials across the sample ( $\nabla P$ : between 0.001 and 0.2 MPa), to assess the need for correction due to inertial effects (i.e. non-Darcian flow). Where necessary, Forchheimer (1901) or Klinkenberg (1941) corrections were applied to the apparent (measured) permeability to derive the permeability of the samples. Thin sections of all but one of the samples were also prepared in order to explore





**Fig. 2.** Collected discontinuity-bearing andesite blocks are shown in panels (a) to (i). Scale is the same for all blocks, and images are aligned so that the discontinuity is oriented more-or-less horizontally in each case. V, M, and L in each sample name refer to the sampling sites of Volcancito, Montegrando, and La Lumbre, respectively. We refer the reader to Table 1 for a detailed description of each block.

and analyse the microstructure using a Scanning Electron Microscope (SEM); due to the similarity of the two pumiceous samples, COL-V-1 and COL-V-6 (Table 1; Fig. 2a, b), the microstructure was only examined for the latter (both blocks were collected from the same fall deposit; as such we deem it probable that the macroscopically similar banding in these samples is characteristic of the same genetic process, rather than

two separate mechanisms being coincident in the same deposit). Digital image analysis of photomicrographs was performed in order to estimate 2D porosities both inside and outside of the features using the open-source image-processing program ImageJ (see Schneider et al., 2012). For select samples, X-ray Fluorescence (XRF) was used to assess compositional differences between the band and the corresponding host rock.

#### 4. Microstructural observations

This section outlines microtextural and microstructural observations of each sample. We also provide image-derived 2D porosities (see Table 2). Inferred genetic processes—based on our observations—are discussed in a later section.

Block COL-V-6 is a pumiceous block made up of distinct dark and light grey portions (Figs. 2a; 4a–c). The 2D porosity is around 0.23 in the dark grey portion, and approximately 0.50 in the lighter coloured regions (Table 2). In the less porous dark grey portions, pores are often equant and typically around 50  $\mu\text{m}$  in diameter, but can be 50–100  $\mu\text{m}$  (Fig. 4a, b). In the lighter coloured, (i.e. more porous) regions, pore diameters are typically larger—generally around 100  $\mu\text{m}$ , and occasionally up to 500  $\mu\text{m}$ —and are often non-equant (Fig. 4c). The long axis of pores in the lighter coloured material is preferentially oriented perpendicular to the interface between the layers, with these pores (the 100  $\mu\text{m}$  range) having an aspect ratio of approximately 2:1. This is in contrast to the pores in the darker coloured portion, which show a preferable orientation parallel to this interface (although this tendency is much less pronounced). The boundary between the light grey and dark grey regions is sharp. We also observe occasional jigsaw-fit broken crystals (Fig. 4b).

Block COL-V-5 (Figs. 2c; 4d–f) is made up of distinct light grey and dark grey portions. Both regions contain microcracks, and porosity is equivalent in both the lighter and darker coloured parts of the block ( $\phi = 0.09$  in each case: Table 2). In the light grey region, pores are generally from 10 to 100  $\mu\text{m}$  in diameter and are usually rounded (Fig. 4e). In contrast, pores in the dark grey region tend to be smaller in size (between 10 and 50  $\mu\text{m}$ ) and amoeboid—rounded but highly irregular—in shape (Fig. 4f). We thus infer that the difference in colour is due to the contrasting pore size and pore size distribution, rather than a significant difference in porosity. The boundary between these two portions is difficult to distinguish on the microscale (as evidenced by the low-magnification SEM image shown in Fig. 4d). Microstructural examination of the interface indicates that the transition between regions dominated by different pore sizes (and shapes) occurs over  $\sim 2000 \mu\text{m}$ . We also note the presence of cm-size xenoliths within the sample.

Block COL-V-3 contains tapered bands or lenses ( $\phi = 0.29$ ) up to approximately 500  $\mu\text{m}$  in width within a dense host rock ( $\phi = 0.01$ ) (Fig. 4g and h). The dense host rock contains pores, generally less than 10  $\mu\text{m}$  in diameter, and microcracks (Fig. 4h). The relatively higher porosity in the bands is responsible for their paler colour relative to the host material. The transition between the host rock and the more porous bands occurs over about 100  $\mu\text{m}$  in each case (Fig. 4h). The bands consist of variably sized (from 50 to 100  $\mu\text{m}$  down to less than 10  $\mu\text{m}$ ) angular particles with varying sphericity (for example, we note the presence of abundant platy or needle-shaped particles). Consequently, the inter-granular porosity is irregularly shaped (Fig. 4i). Throughout the host material, we observe micro-scale crystallites within the glassy groundmass; this texture—the so-called “feathery” texture of Horwell et al. (2012, 2013)—is shown in more detail and discussed in Section 5.4.

The host rock of block COL-V-4 is glassy and low-porosity ( $\phi \approx 0.06$ ). As well as microcracks, the porosity of the host rock is made up of subequant pores that tend to be in the region of 10–50  $\mu\text{m}$  in diameter (Fig. 4j), although some are as large as 300  $\mu\text{m}$ . The host rock contains a number of slightly anastomosing porous bands, from 1 mm up to 10 s of mm in thickness, containing a porosity of around 0.32 (Table 2,



**Table 1**

Block (hand-sample) descriptions, including reference to the corresponding photographs in Fig. 2.

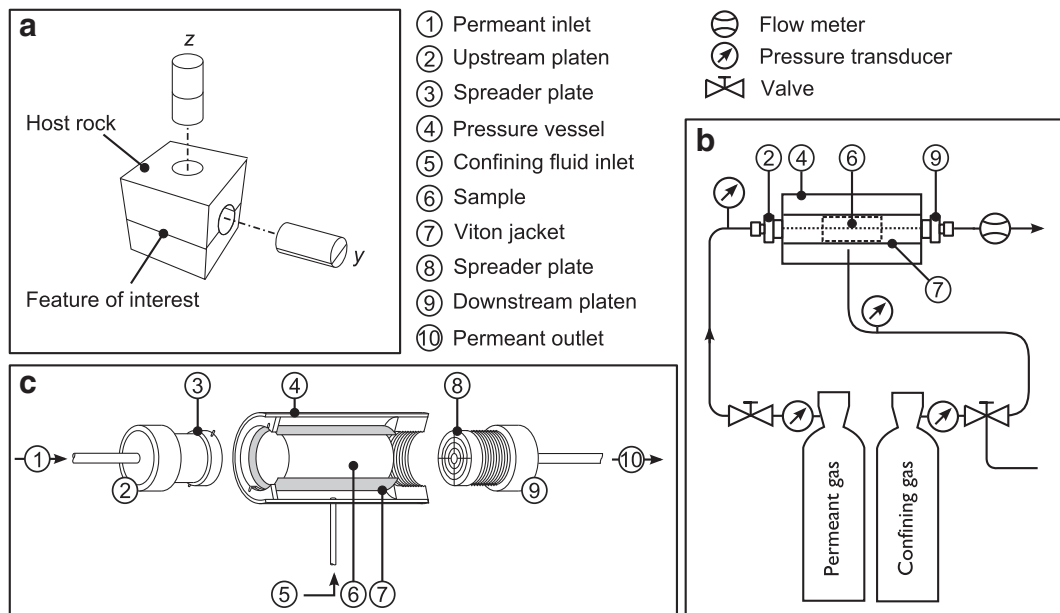
Block name	Photograph	Collection location	Block description
COL-V-1	Fig. 2a	Volcancito	Pumiceous clast containing denser, darker grey portions, approximately 40 mm thick.
COL-V-6	Fig. 2b	Volcancito	Pumiceous clast containing recurring denser, darker grey regions, from 10 to 40 mm thick.
COL-V-5	Fig. 2c	Volcancito	Dense rock displaying two competent portions defined by their difference in colour (a light grey side and a dark grey side). Several small (cm-scale) xenoliths can be observed in the sample.
COL-V-3	Fig. 2d	Volcancito	Dense, dark grey rock containing variously anastomosed lenses of lighter coloured material, ranging from around 5 to 40 mm in thickness.
COL-M-1	Fig. 2e	Montegrande	Dark grey clast containing three light grey tabular bands ranging from 5 to 15 mm in thickness.
COL-V-4	Fig. 2f	Volcancito	Large dense and glassy block, containing numerous anastomosing 1–10 mm thick bands of light grey, variably friable material. Some evidence of oxidation can be observed within the bands.
COL-L-1	Fig. 2g	La Lumbre	Dense rock containing three tabular bands which are light grey in colour. Bands are approximately 5 mm in thickness.
COL-L-2	Fig. 2h	La Lumbre	Dense glassy block containing a single through-cutting band, approximately 5–10 mm thick. The band appears to curve around a xenolith.
COL-L-4	Fig. 2i	La Lumbre	Dark grey clast containing a tabular band 5–10 mm thick. The band is typically light grey, with brick-red oxidation visible.

Fig. 4j–l). Thus the relatively higher porosity within the band makes it appear lighter in colour relative to the dense surrounding rock. There is a sharp contact between the host rock and the band, occurring over a few tens of microns. The interior structure of the band ranges from zones of subangular, granular material at the margins (characterised by a particle size rather than a pore size: particles are generally around 20  $\mu\text{m}$  diameter), to more coherent material towards the centre of the band containing rounded, amoeboid pores, which are generally on the order of 10 to 50  $\mu\text{m}$  in diameter (Fig. 4k, l). There are also large *en échelon* tensional fractures ( $\sim 50 \mu\text{m}$  wide and  $\sim 400 \mu\text{m}$  in length) which cut through the band, oriented approximately  $45^\circ$  to the band (Fig. 4k), and can be seen to skirt around crystal boundaries. Within the granular margins of the band, there are also patches of coherent material, typically adjacent to large phenocrysts (Fig. 4k).

Three bands can be discerned in block COL-M-1, each approximately 5 mm in width (Fig. 2f), two of which are shown in Fig. 4m. The heavily

microcracked host rock contains a porosity of 0.01, in contrast to the bands which have a porosity of around 0.20. Accordingly, the bands appear paler in colour than the host rock. We also observe feathery groundmass textures (see Horwell et al., 2012, 2013) in the host rock as well as crystalline silica, darker in colour than the groundmass (discussed further in Section 5.4). Porosity inside the bands tends to be highly irregular in shape (Fig. 4n). One population of pores is evident in the range of 10–50  $\mu\text{m}$  in diameter, with other, larger pores (100 s of  $\mu\text{m}$  in diameter) occurring adjacent to large rigid crystals (Fig. 4n, o). Despite the bands being very tabular, the transition between each band and the host rock tends to be relatively diffuse, occurring over a distance of approximately 500  $\mu\text{m}$  (Fig. 4n, o).

COL-L-1 is a dense glassy block (with a host rock porosity of approximately 0.03) containing tabular bands with relatively higher porosity (around 0.10), manifest in the difference in colour observable in the hand sample. The porosity of the host rock typically comprises small



**Fig. 3.** Anisotropic permeability measurements on banded rocks. (a) Orthogonal cores are obtained from each block with respect to the plane of the feature of interest. (b) Schematic of the benchtop steady-state permeameter used in this study. Both the permeant (pore fluid) and confining pressure gases are nitrogen. The sample is inserted into the annular viton jacket, which is secured within the pressure vessel shown in (c). The up- and downstream platens are screwed into position until both spreader plates are in firm contact with the sample. A confining pressure can then be applied to the sample by filling the annulus (the void space between the vessel and jacket) with gas. With this setup, the pressure differential  $\nabla P$  between the upstream pressure transducer and atmospheric pressure downstream of the sample can be monitored at different volumetric flow rates, measured using the downstream flowmeter, during which time the system vents to the air. The confining pressure imposed on the sample and jacket (in these tests, 1 MPa) ensures that the through-flowing pore fluid does not leak around the sides of the sample. To ensure microstructural equilibration, the sample was left at this confining pressure of for a minimum of 1 h prior to each test. Similarly, each measurement was only made once the gas flow through the sample had achieved steady-state equilibrium over time. Schematics not to scale.

**Table 2**SEM-derived band and host-rock porosities, including the difference between the two ( $\Delta\phi$ ).

Sample	Band porosity	Host porosity	$\Delta\phi$
Col-V-6	0.23	0.50	−0.27
Col-V-5	0.09	0.09	0.00
Col-V-3	0.29	0.01	0.28
Col-V-4	0.32	0.06	0.26
Col-M-1	0.20	0.01	0.19
Col-L-1	0.10	0.03	0.07
Col-L-2	0.18	0.01	0.17
Col-L-4	0.03	0.13	−0.10

pores (<20  $\mu\text{m}$  in diameter) and microcracks (Fig. 4p–q). The material inside the bands is made up of subangular, prolate fragments with their long axes generally <50  $\mu\text{m}$  in length, interspersed with relatively large phenocrysts up to 500  $\mu\text{m}$  in diameter (Fig. 4q–r). We also note dark patches of cracked crystalline silica. The transition between host and band occurs over a width of up to 1 mm (Fig. 4q).

Col-L-2 is another dense, glassy block (with a host rock porosity of approximately 0.01) containing a 3–5 mm-thick tabular band containing a porosity around 0.18 (Fig. 4s) (the high-porosity band is correspondingly lighter in colour than the rest of the sample). The shape of the intergranular porosity within the band is amoeboid, typically 100  $\mu\text{m}$  or less in diameter, but occasionally up to around 200  $\mu\text{m}$  (Fig. 4t–u), whereas the porosity in the host rock consists of equant pores no greater than 100  $\mu\text{m}$  in diameter, and abundant microcracks. Notably, the band appears to deflect around a xenolith in the host material (Fig. 4s, inset). There is a discrete interface between the host rock and the porous band (Fig. 4t–u). Throughout the groundmass, we observe acicular lath-like crystals, darker than the rest of the groundmass constituents (discussed further in Section 5.4).

Finally, Col-L-4 is a block with a host rock porosity around 0.13; the porosity of the host rock is made up of irregularly-shaped pores—from around 100 to 1000  $\mu\text{m}$  in diameter—and abundant tortuous microcracks. The band visible in the block (Figs. 2i, 4v inset) is characterised by a region of relatively lower porosity ( $\phi \approx 0.03$ ). Much of the band appears reddish in colour at hand-sample scale, as well as in the prepared thin section. The inner walls of the pores at the margins of the low-porosity band are defined by numerous high aspect ratio (platy or needle-like) crystals. Towards the interior of the band, the porosity is characterised by the interparticulate space between this platy or needle-like material (Fig. 4w, x), or spherical pores containing variably granular, angular to rounded particles

from 10 to 50  $\mu\text{m}$  in diameter (Fig. 4w). In the centre of the bands, porosity exists as ~200  $\mu\text{m}$ -diameter patches of subrounded pores that are generally <20  $\mu\text{m}$  in diameter. The boundary between the host rock and the relatively lower porosity band is very diffuse (the band is difficult to discern on the microscale, Fig. 4v), occurring over around 2000  $\mu\text{m}$  in some areas.

## 5. Microstructural interpretations

The various features described above provide snapshots of a range of interrelated conduit or dome processes. However, the microstructures that develop due to any given process may well overprint earlier textures. To add further complication, the extent of expression of any specific texture is likely to depend heavily on the time allowed for the responsible mechanism to operate. As such, we do not ascribe any one definitive genesis to the features discussed in this study but refer to the microstructure to bolster the interpretations we regard as most likely in each instance.

### 5.1. Banding in pumiceous samples

XRF analysis of a banded pumiceous block (Col-V-1: Fig. 2a) indicates that there is a negligible difference in bulk composition (of the major and trace elements analysed) inside and outside of the bands, suggesting a predominantly physical process in their genesis. This is in contrast to a number of other studies which discuss banding in pumice—for example Venezky and Rutherford (1997); Hall et al. (1999), and Bouvet de Maisonneuve et al. (2009)—which attribute observed heterogeneities to the intermingling of magma batches of distinct composition or variations in magma differentiation. Burgisser et al. (2010) show macroscopically similar banded pumice collected at Soufrière Hills Volcano, Montserrat, ascribing the heterogeneity of porosity within individual samples to corresponding variations in water content. Based on our porosity data and microstructural observations (notably, band-perpendicular bubble elongation: Fig. 4b), we surmise that the constituent material underwent a certain degree of extension or decompression-driven bubble expansion assisted by viscous deformation (*i.e.* above its glass transition  $T_g$ ), but not to a degree sufficient to wholly fragment the sample. This process has been observed in shock-tube experiments (*e.g.* Martel et al., 2000; Cagnoli et al., 2002; Namiki and Manga, 2005), which demonstrate that decompression can cause volatile exsolution from a supersaturated liquid, followed by bubble growth and coalescence.

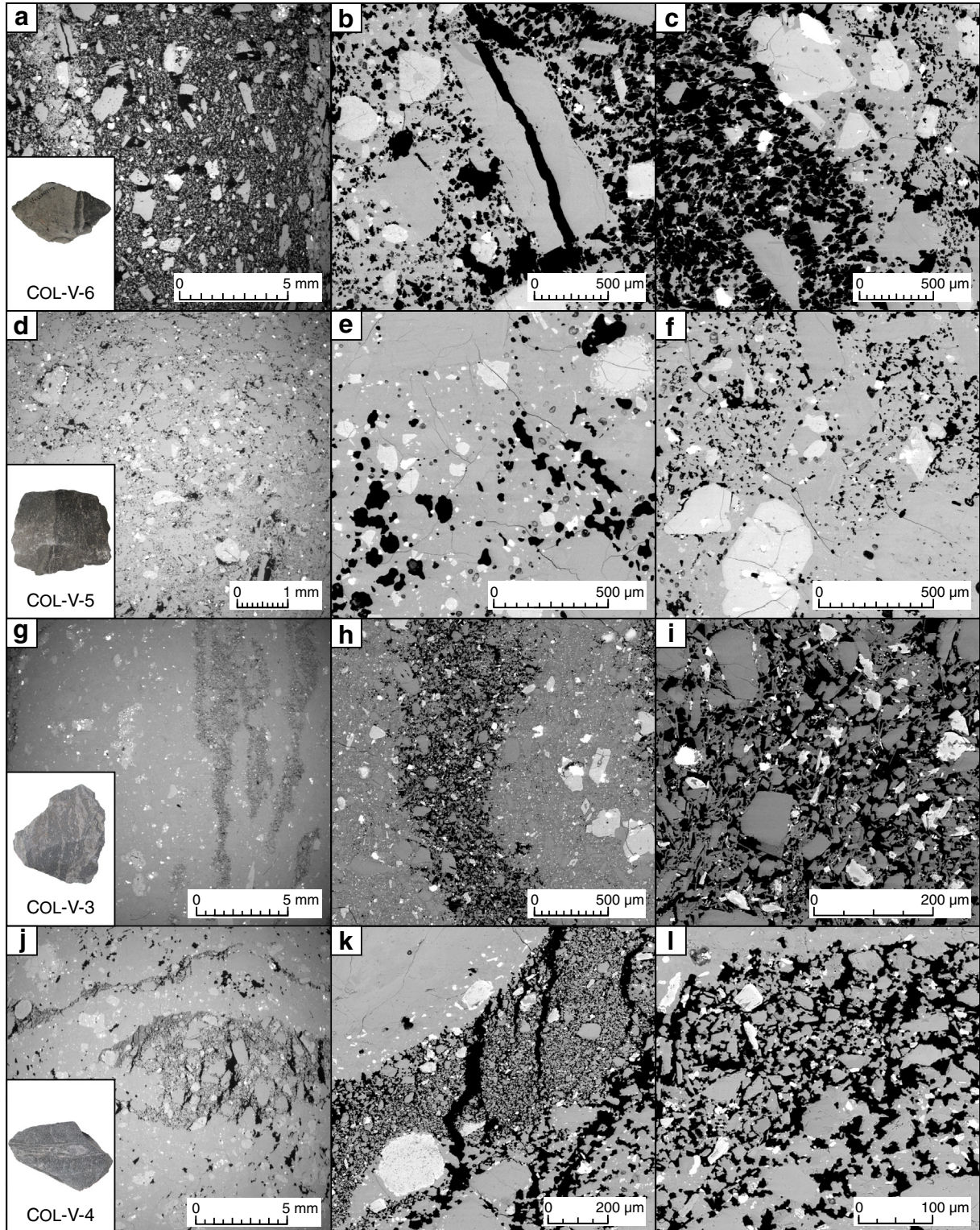
Band-parallel elongation of pores in the denser material (Fig. 4b) may be indicative of coincident compression associated with the propagation of a rarefaction wave (*i.e.* pure-shear flattening of neighbouring bubbles during bubble expansion: Wright and Weinberg, 2009);

**Fig. 4.** Examples of microstructure for eight samples. Each sample is shown inset, with three panels showing microstructure at varying scales. Porosity is shown as black in these and all subsequent SEM images. Note that the left-most panel of each row is an image of the whole thin section, cropped in order to avoid the visual distortion (*i.e.* the “fisheye” effect) resulting from the low magnification. (a) Sample Col-V-6, showing recurring bands of less-porous material within a highly porous sample. (b) Large phenocryst in Col-V-6 which has been extensionally fractured and pulled in two by approximately 100  $\mu\text{m}$ . (c) The transition from high porosity (left-hand-side) to low porosity (right-hand-side) regions of Col-V-6. Much of the porosity in the left-hand-side appears elongated perpendicular to the band (*i.e.* elongated horizontally in the image), whereas evidence of band-parallel directionality (vertical elongation in the image) can be seen in the lower-porosity region (right-hand-side). Refer to text for more discussion. (d) SEM image of sample Col-V-5. Note that the transition between pale and dense regions appears discrete at hand-sample scale (inset), but is less apparent in the microstructure. (e) Typical porosity within the paler part of the block (Col-V-5). It is notably more spherical or amoeboid, as well as typically larger, than the porosity shown in (f), which comprises the darker region of the same block. (g) The variably anastomosed lenses of higher porosity of Col-V-3. A closer image (h) shows that the transition from dense to porous material occurs over a distance on the order of 100  $\mu\text{m}$ . Inside the bands of Col-V-3 (i) we observe incipient (“point-to-point”) welding, and possible evidence of needle-like vapour-phase crystallisation textures. (j) Sample Col-V-4, wherein discrete bands of increased porosity can be easily discerned. Closer images show extensional *en échelon* porosity (k) and variably-sized granular material (l). From the edge of the band towards its centre, an evolution from nonwelded fragments towards a partially welded medium can be seen (k, l). (m) Sample Col-M-1, wherein two of the three porous bands can be seen (oriented vertically). A closer image of the discontinuity is shown in (n), where it can be seen that phenocrysts intrude into the band. In (o), “feathery” groundmass textures can be seen, indicative of devitrification within Col-M-1. Panel (p) similarly shows the alternating bands of dense and relatively porous material in Col-L-1. The transition between these regions is diffuse, occurring over approximately 500  $\mu\text{m}$ , shown in (q). As with sample Col-M-1, panel (r) shows that the largest pores in Col-L-1 are associated with phenocrysts. The band in sample Col-L-2 is shown in (s). The boundary between the host rock and the fracture is shown in greater detail in (t) and (u). Notably, the groundmass is entirely interconnected in (u), indicating that sintering of transiently granular material is well advanced. In the final sample, Col-L-4, the feature is manifest in a low-porosity band aligned vertically in (v) and the hand-sample image, inset. The band is notably diffuse, encompassing the relatively dense patch running more or less vertically through the centre of the image. In (w), the progression from relatively spherical or amoeboid pores to almost entirely infilled porosity can be discerned in more detail as one moves from the bottom to the top of the image. Within many of the pores of Col-L-4, variably-sintered material and vapour-phase crystallisation textures can be observed, as shown in (x).



however we can infer that the process was rapid and predominantly extensional as we observe evidence of crystal parting at the band–host rock interface (Fig. 4c). Comparable breakage of crystals was reported by Kennedy et al. (2005), who attribute similar bands in pumice to lateral conduit implosion following the propagation of a fragmentation front through the magmatic column. Such textures could conceivably also arise during magma ascent in the conduit, or during phases of

lava dome-extrusion: structures that are inherently prone to collapse and depressurisation (e.g. Fink and Kieffer, 1993; Navon et al., 1998; Voight et al., 2006). Indeed, variably-vesiculated layers of pumice and glass observed at Inyo Dome and Big Glass Mountain, both in the USA, illustrate the complex textural heterogeneities that can develop due to dome depressurisation (Fink and Manley, 1987; Castro et al., 2005).





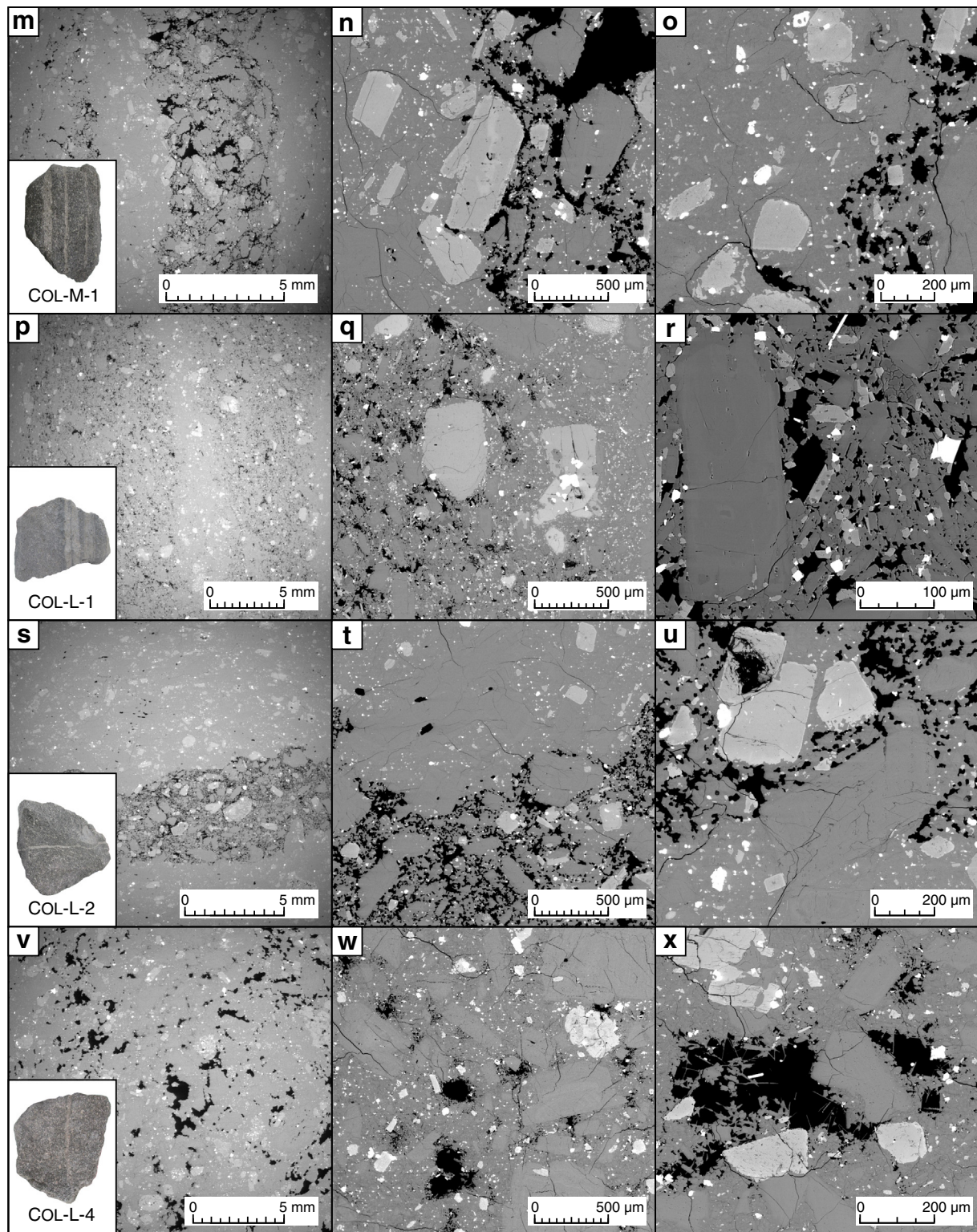


Fig. 4 (continued)

### 5.2. Shear strain-induced flow banding

The development of laminar flow due to vertical and horizontal gradients in temperature, pressure, and shearing mode or internal friction (e.g. Denlinger and Hoblitt, 1999; Mastin, 2002; Rust and Manga, 2002; Rust et al., 2003; Gonnermann and Manga, 2005, 2007) can ultimately partition magma properties such as crystal content, water content, or porosity, resulting in heterogeneous deformation of the

magma even over microscopic scales (e.g. Wright and Weinberg, 2009; Laumonier et al., 2011; Lavallée et al., 2013). Plug or piston flow within a magma conduit (e.g. Gonnermann and Manga, 2003; Hale and Mülhaus, 2007) dictates that shear strain in magma should be higher at the periphery of the conduit than in the centre (the effect of a variable stress profile across a conduit on a non-Newtonian magma rheology containing crystals and bubbles favours strain localisation, and in particular simple shear, in regions of highest strain rate near



the conduit margin; Lavallée et al., 2007; Dingwell et al., 2015). Similarly, large strains—and large gradients thereof—have been estimated from actively extruding domes (e.g. Cashman et al., 2008), and are expected to be prevalent in dome- and spine-building systems (e.g. Griffiths and Fink, 1993; Buisson and Merle, 2004; Hornby et al., 2015). Moreover, additional stresses are likely to be pervasive within an active volcanic system, for example due to cumulative edifice construction and loading of the basement (e.g. De Vries and Borgia, 1996; De Vries and Merle, 1998; Gerst and Savage, 2004; Roman et al., 2004; Odbert et al., 2015). Deformation induced solely by simple shear (which is to say an isochoric or isovolumetric process) would not result in the creation or destruction of porosity (i.e. dilation or densification of the magma), but in inhomogeneous deformation of the porous melt and modification of the permeability (Wright et al., 2006). This is in line with our observations of sample Col-V-5 (Figs. 2c, 4d–f): whilst the pore size distribution is notably different between the competent light and dark grey regions, the connected porosity is equivalent in both (~0.09; see Table 2). Further, the inclusion of xenoliths in the block may be indicative that energetic shearing has occurred (Rust et al., 2004). This interpretation is given weight by the fact that XRF analysis of this sample yields comparable bulk composition in both the light grey and dark grey portions, suggesting a predominantly physical process rather than a compositional one. Nevertheless, alternative physical mechanisms cannot be discounted. For example, finite element modelling of shallow-vent exogenous dome growth predicts the concurrence of magma batches with contrasting degassing histories, manifest in juxtaposed regions of differing porosity (Massol and Jaupart, 2009).

Purely laminar flow in a conduit is, however, not necessarily the norm (Costa and Macedonio, 2003). Rather, some degree of volumetric change may be associated with shear strain, resulting either in transtension (shear and extension) or transpression (shear and compression). Viscous-to-brittle transtension of fluid—whereby the vapour phase of a liquid is formed due to a local pressure or stress differential—is also termed cavitation (e.g. Gruzdkov and Petrov, 2008), and potential evidence for this process exists in the microstructure of, for example, Col-V-4, Col-M-1, and Col-L-1. Below, we describe some of these characteristic features—shown in more detail in Fig. 5—analogue to those often observed in material science (such as ceramic studies: Chokshi, 1997), and texturally similar to those described in dacitic dome rocks from Unzen, Yakedake, and Daisen volcanoes (all in Japan) by Smith et al. (2001). For example, *en échelon* extensional porosity can be seen within some of the porous bands, as shown in Fig. 5a (Col-V-4; see also Fig. 4j, k). These high aspect ratio pores skirt crystal boundaries, comparable to the “crack-like” porosity of Smith et al. (2001). We further note that these features are comparable in form and formation to experimentally-created “tension gashes” in pure melts (A. Kushnir, pers. comm.) and extensional fracturing in crystalline magma under uniaxial compression (Lavallée et al., 2013). Moreover, Fig. 5b (Col-M-1) and Fig. 5c (Col-V-4) show where porosity has preferentially developed along the borders of large phenocrysts, which can be interpreted as further evidence of cavitation porosity (this can also be observed in sample Col-L-1; see Fig. 4q, r). Notably, Fig. 5c shows a tail of low-porosity adjacent to a phenocryst. Although the *in situ* stress field cannot be known, we interpret this feature as an indicator of localised differences in the stress field—i.e. a pressure/stress shadow—which is a prerequisite condition for cavitation.

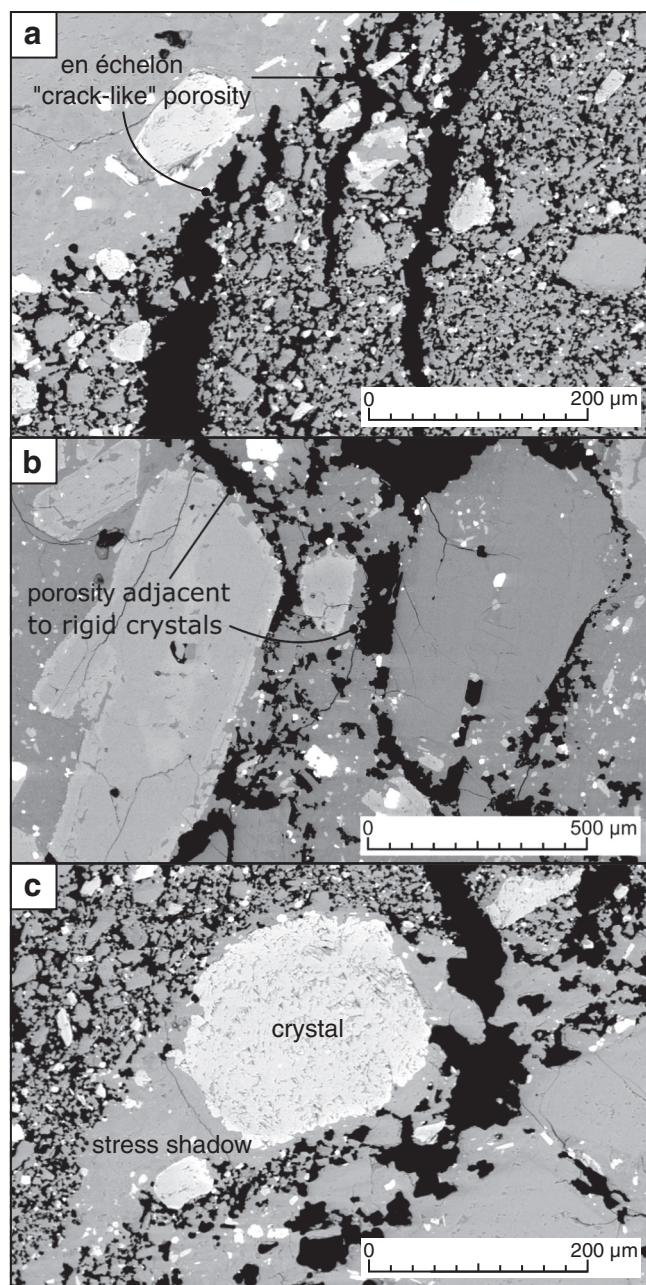
### 5.3. Magma fracture and sintering

Brittle behaviour in magma is typically ascribed one of two mechanisms: magma fragmentation resulting from a pore pressure that exceeds the tensile strength of the magma (i.e. a fragmentation threshold: e.g. Alidibirov and Dingwell, 1996; Sparks, 1997; Zhang, 1999; Spieler et al., 2004; Koyaguchi et al., 2008), or magma failure

due to rapid shearing surpassing a critical strain rate in areas of strain localisation (e.g. Woods and Koyaguchi, 1994; Marti et al., 1999; Papale, 1999; Gonnermann and Manga, 2003; Melnik et al., 2005; Lavallée et al., 2008, 2013). In both cases, failure occurs at the glass transition because the strain rate cannot be relaxed viscously and exceeds the elastic limit of the material (Dingwell, 1996). The former process is associated with relatively rapid magma decompression (as outlined in Section 5.1) or gas waves or slugs fluxing through the magmatic column (e.g. Michaut et al., 2013). As for the latter, a critical strain rate criterion may be met at any depth in the conduit or within extruding lava domes, resulting in magma fracture (as discussed by, e.g. Dingwell, 1996; Papale, 1999; Webb and Dingwell, 1990; Tuffen et al., 2003; Edmonds and Herd, 2007; Lavallée et al., 2008; Tuffen et al., 2008; Cordonnier et al., 2012), if the melt viscosity prevents sufficient stress relaxation during rapid magma ascent (Goto, 1999; Papale, 1999; Gonnermann and Manga, 2003).

The sharp transition we observe between the host material and the bands present in many of our samples (Col-V-3, Col-L-2, Col-V-4), hints at brittle failure of the magma. Examples of the well-defined contact are shown in Fig. 6a and f. In each of these cases, these fractures are also filled with variably granular material, as we may expect if ash-sized material is generated on the fault plane or transported through the fracture. Once a fracture is generated in rock or magma, the conditions may exist to allow subsequent viscous sintering or hot pressing of fragmental material within the fracture (e.g. Stasiuk et al., 1996; Tuffen et al., 2003; Kolzenburg et al., 2012; Vasseur et al., 2013; Wadsworth et al., 2014; Heap et al., 2014a, 2015c), either *in situ* or after some degree of transportation (Tuffen and Dingwell, 2005). As well as having widespread industrial applications (e.g. Sousa et al., 2015), viscous sintering—the agglutination of melt particles above their glass transition temperature—is common in a range of volcanic environments, examples of which include rhyolitic dykes and conduits (e.g. Tuffen et al., 2003; Tuffen and Dingwell, 2005; Okumura and Sasaki, 2014), welded block-and-ash flow deposits (e.g. Michol et al., 2008; Andrews et al., 2014; Heap et al., 2014a), lava spatter and spatter ramparts (e.g. Mellors and Sparks, 1991; Sánchez et al., 2012), localised agglutination of scoria (Wadsworth et al., 2015), conduit-filling pyroclastic material (e.g. Kolzenburg and Russell, 2014), rheomorphic and lava-like ignimbrites (Branney et al., 1992; Streck and Gruner, 1995; Lavallée et al., 2015), and even as a consequence of the deposition of volcanic ash in jet engines (e.g. Song et al., 2014). More specifically, the sintering of “transiently granular” material (Wadsworth et al., 2014) has been described in terms of fracture healing in magma (e.g. Tuffen et al., 2003; Heap et al., 2015c). Viscous sintering of volcanic material requires the temperature and the timescale of the process to be such that the material remains in a liquid regime (above the glass transition  $T_g$  of its melt phase), whereupon the melt structure can relax and heal particulate material via diffusion (e.g. Vasseur et al., 2013). The process encompasses three separate stages (e.g. Vasseur et al., 2013; Wadsworth et al., 2014): (1) point-to-point connection of melt “necks” between granular materials (incipient sintering or welding), (2) a variably porous but melt-dominated medium (partial sintering or welding) and, (3) a nonporous melt and crystal phase (i.e. fully sintered or welded).

Microstructural evidence of the first two of these stages can be observed in a number of our samples, examples of which are shown in Fig. 6. Discrete angular granular material is typically observed at the inner margins of fractures (Fig. 6c, h). Advancing towards the centre of each sintered band (i.e. from Fig. 6c → e and h → j), we observe that fragments are progressively more rounded, and often connected by necks of glassy groundmass, indicating melt relaxation associated with incipient to partial sintering. This process is reflected in increasingly connected groundmass and correspondingly isolated porosity, as the material transitions from fragments within void space to a porous melt wherein the original angular nature of the fractures cannot be discerned. Pores also appear increasingly spherical or amoeboid within the centre of either band (Fig. 6e, j), illustrative of the relaxation of the



**Fig. 5.** Evidence of shear-induced transtension (cavitation) in banded samples. (a) High aspect ratio *en échelon* porosity, indicative of extensional shearing in Col-V-4. In (b) and (c), porosity can be seen preferentially formed around the edge of large phenocrysts (in samples Col-M-1 and Col-V-4, respectively). (c) Local differences in stress are evidence by a tail of non-fragmental material, reflecting a stress shadow adjacent to a phenocryst.

internal surface area of the porosity with ongoing sintering (e.g. Mackenzie and Shuttleworth, 1949; Wadsworth et al., 2014). Comparison of samples Col-V-4 and Col-L-2 (Fig. 6a–e; f–j) illustrates that these

sintering textures are generally more developed in the latter sample, suggestive of higher temperatures or greater depths (e.g. Wadsworth et al., 2014; Heap et al., 2015c), or longer timescales (Vasseur et al., 2013). Thus we can observe differing degrees of sintering both within a single fracture and between different samples.

#### 5.4. Evidence for gas transport

The low-porosity band within the sample Col-L-4 (Figs. 2i and 4v–x) shows pores partially filled with variably-sintered fragmental material (Fig. 7a; Table 2). Based on our observations we speculate that the progressive pore infilling is the result of ash deposition from circulating hot ash-laden fluids. The hot and “sticky” ash particles first adhere to and coat the pore walls (as in Fig. 7b), subsequently reducing pore throat apertures, which may well encourage further agglutination and filling of the pores (Fig. 7c). The originally angular and fragmental ash particles (Fig. 7a) gradually lose their shape as surface tension reduces the curvature of the melt/fluid interface, promoting the rounding of particles (Fig. 7b, c; see Vasseur et al., 2013). Eventually the pore becomes completely filled and the particles continue to densify through sintering. At this stage, the characteristics of the original granular material and the original pore are mostly eradicated; however, relics of the originally granular material are defined by remnant microporous patches (Fig. 7d, and inset).

Other than in sample Col-L-4, we do not observe definitive evidence for the transport of fragmented magma or fault gouge—whether juvenile or from previous explosive activity—in our samples. Nevertheless, recent studies (Kolzenburg et al., 2012; Kendrick et al., 2016; Lavallée et al., in press) suggest that this process may be common at Volcán de Colima in materials derived from either the dome or conduit. However, we do observe variable oxidation of the bands in some hand samples (Col-V-4, Col-L-4; see Table 1), indicative of through-flow of hot magmatic fluids. This is in agreement with the findings of Plail et al. (2014), who interpreted changes in metal compositions in banded volcanic rock as a result of magmatic gas flow.

Further, textural evidence of silica polymorphs, both within bands and in the host rock groundmass, may point to transport of magmatic volatiles through porous discontinuities. Fig. 8a shows sample Col-M-1, highlighting the darker-coloured crystalline silica. The diagnostic “fish-scale” microcracking texture is indicative of cristobalite having undergone the  $\beta$ - to  $\alpha$ -cristobalite transition (e.g. Damby et al., 2014), often observed in volcanic dome rocks from, for example, Inyo Domes, USA (e.g. Swanson et al., 1989), Mount St Helens, USA (e.g. Blundy and Cashman, 2001), and Gunung Merapi, Indonesia (e.g. Kushnir et al., 2016). In sample Col-L-2 (Fig. 8b) we observe lath-like crystals we interpret to be tridymite (see Blundy and Cashman, 2001). Fig. 8c shows an example of a sample exhibiting a feathery groundmass texture (sample Col-V-3), characterised by microscale crystallites occurring within the groundmass glass. The growth of this texture is thought to represent variable devitrification of the initially glassy groundmass (Horwell et al., 2012, 2013). We also observe needle-like or platy textures within pores and bands of some samples, an example of which is shown in Fig. 8d (sample Col-L-4). These textures are consistent with prismatic cristobalite growth observed in lavas from the dome of Soufrière Hills volcano (Montserrat) by Horwell et al. (2013), both in terms of form

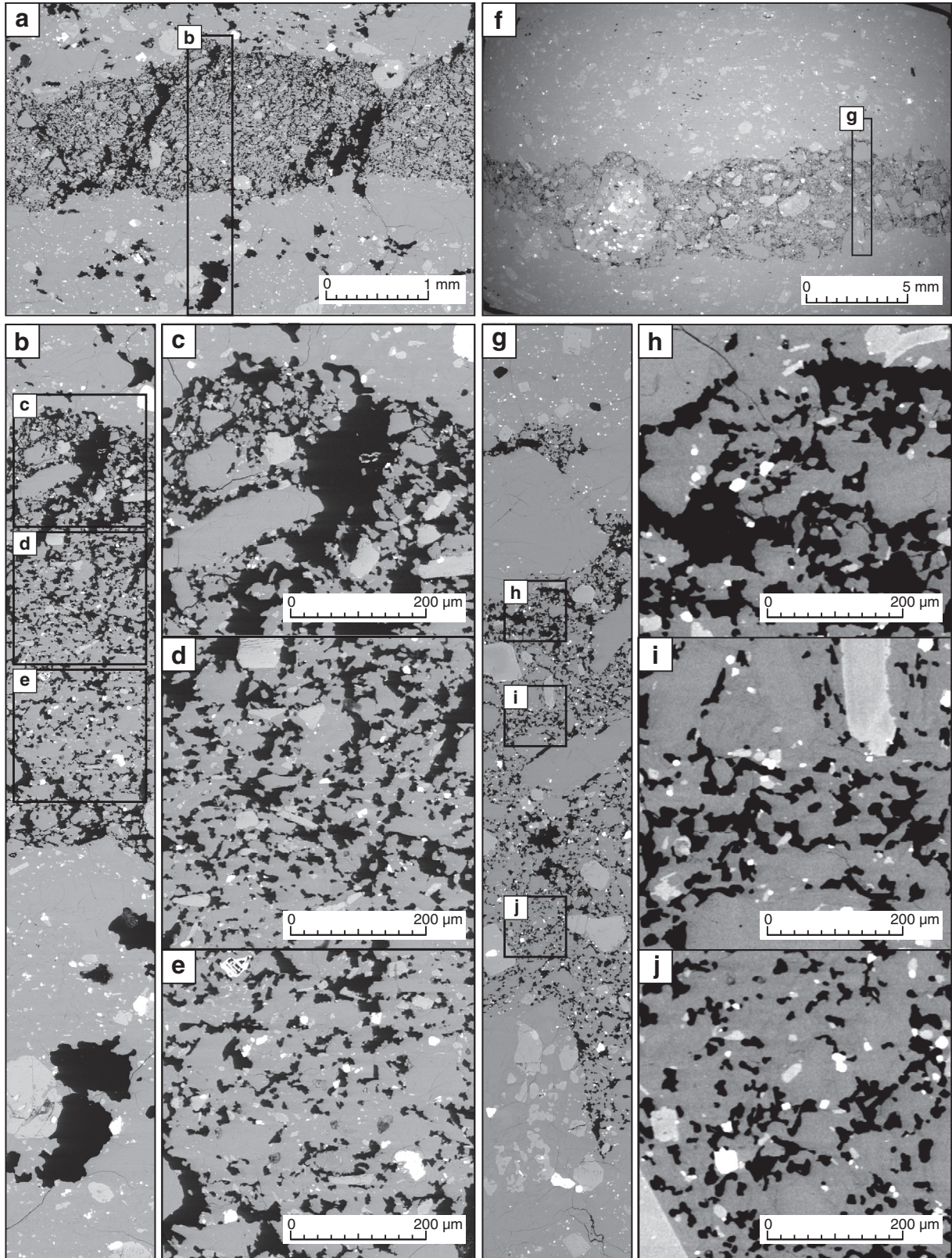
**Fig. 6.** The evolution of sintering of granular fracture material. (a) Planar fracture in sample Col-V-4. (b) Detailed SEM images (c–e) are acquired from the edge and near the centre of the band. (c) At the edge of the fracture, discrete ash-sized fragments can be discerned, most of which retain their angular form. For example, a cluster of discrete grains can be seen in the top-left of the panel. (d) Some rounding of fragments can be seen, and there is an increased connectivity of the groundmass. Both are indicative of incipient sintering. Characteristic “necks” can be observed forming between some of the formerly angular fragments; i.e. where there is a concave curvature of connecting groundmass between particle bodies, in contrast to the convex curvature of the particle surfaces. (e) Closer to the centre of the band, the groundmass is better connected, and porosity is increasingly more isolated relative to (c) and (d). This is evident throughout the panel, where numerous pores <20 µm in diameter can be discerned with little or no connectivity with their neighbouring pores. (f) Planar fracture in sample Col-L-2. (g) The location of detailed images (h–j) is shown, the first at the edge of the band, the latter two towards the centre. (h) As in (c), some discrete granular material can be observed within the void space, particularly in the bottom-right of the image. However, rounding of the fragments is more pronounced, suggesting more advanced sintering than their counterparts shown in (c). (i) Towards the middle of the fracture, the microstructure has transitioned from variably granular material within void space to a competent groundmass with variably connected porosity. Note that the pore shape is generally amoeboid to spherical. Necking textures are pronounced in the centre-right and bottom-left of the image. (j) Similar textures to (i) are evident, but the image indicates notably poorer connectivity of the pores, suggestive of more advanced sintering. Note that the characteristic shapes of the original granular material cannot be discerned.



and tendency to occur within pores and fractures in the rock. These authors interpret the latter fact as evidence for vapour-phase mineralisation.

Vapour-phase crystallisation of silica polymorphs, namely tridymite and cristobalite, provides strong evidence that transiently

connected porosity (e.g. cavities or discrete fractures in magma) does indeed serve as a transport network for element-rich magmatic gases. Tridymite and cristobalite are metastable at high temperatures and low pressures (e.g. [Deer et al., 1996](#); [Damby et al., 2014](#)), indicating that the processes involved in forming or infilling these



fractures and pores must be occurring in relatively shallow regions of the dome or conduit. This is in agreement with the findings of Ball et al. (2015), who suggest that hydrothermal alteration and precipitation will be preponderant in talus-mantled lava domes, where these processes are enhanced due to favourable heat and permeability conditions.

Similarly, Cashman et al. (2008) show evidence of tridymite in dacite from Mount St Helens (USA), and estimate that its presence necessitates relatively low pressures (between around 5 and 25 MPa). The observation that porosity is created and destroyed by physical and chemical processes in the upper conduit and surficial dome emphasises the importance of understanding localised spatio-temporal variations in permeability.

## 6. Permeability anisotropy

Our measurements, and the attendant discussion in the previous section, highlight that various conduit and dome processes manifest themselves differently: they may increase or decrease porosity, or porosity can remain effectively constant. The following section discusses the consequent influence on the permeability of these discontinuity-bearing samples, which can be distinguished into four main categories: (1) an increase in porosity may exert only a negligible influence on permeability, (2) an increase in permeability may arise even when porosity is essentially unchanged, (3) an increase in porosity can be accompanied by an increase in permeability, or (4) a decrease in porosity can yield a decrease in permeability.

In the majority of the collected blocks, porosity within the band or fracture is higher than or equivalent to that of the host rock mass (Table 2). The exceptions are the banded pumiceous samples (COL-V-6 and COL-V-1: Figs. 2a, b, 4a–c) and sample COL-L-4 (Figs. 2i, 4v–x). The 2D porosity data derived from SEM photomicrographs (Table 2) agree with the connected gas porosity measurements performed on cylindrical cores obtained from each block (Table 3). Gas permeability of all core samples is also given in Table 3, where we also indicate whether the heterogeneity—if present—was parallel or perpendicular to the core axis (hence the direction of fluid flow). Permeability as a function of porosity is shown in Fig. 9. In general, high porosities are associated with high permeabilities, and low porosities with low permeabilities, although the precise relation differs between the high (>0.30) and low (<0.20) porosities (as similarly commented on by Heap et al., 2014b; Farquharson et al., 2015; Kushnir et al., 2016).

The permeability of a heterogeneous porous medium is termed here the “equivalent” permeability  $k_e$  (see Renard and De Marsily, 1997 for a review). For sample sets where we have at least one core containing a band parallel to the core axis, and at least one band-free core, we are able to decouple the equivalent permeability in order to estimate the permeability of the band or fracture itself via a simple two-dimensional parallel-plane model. We assume that the (“intact”) host permeability  $k_i$  is that of the feature-free core; thus the permeability of the band or fracture  $k_f$  is calculated as a function of the sample area  $A_s$ , the band area  $A_f$ , and the host and equivalent permeabilities, such that

$$k_f = \frac{k_e A_s - k_i (A_s - A_f)}{A_f}. \quad (1)$$

This method has previously been used to calculate the permeability of tensile fractures in variably-porous andesites from Ruapehu (New Zealand) (Heap and Kennedy, in review). Note that this model assumes that  $A_f$  is small compared to  $A_s$ . Table 4 gives the parameters used, as well as the mean band or fracture permeabilities, for each of the rocks where this calculation was possible. Further, we show the “relative” permeability  $k_r$  between the host rock and the discontinuity, given

here by the fractional change in permeability with respect to the host value

$$k_r = \left( \frac{k_f - k_i}{k_i} \right), \quad (2)$$

where a positive value means that a feature is more permeable than its corresponding host rock, whilst a negative value indicates the reverse. The data of Tables 3 and 4 show that the presence and orientation of a discontinuity have a variable influence on the permeability of our samples: permeability can increase, decrease, or remain roughly constant depending on the properties of the host rock and the feature. These scenarios are discussed below.

We will first examine the banded pumice. Data presented in Fig. 9b and Table 3 show that permeability of pumiceous material (COL-V-1 and COL-V-6) varies by less than an order of magnitude: despite ranging between ~0.35 and ~0.65 porosity, permeability ranges from  $5.44 \times 10^{-14}$  to  $3.42 \times 10^{-13} \text{ m}^2$  in these samples. As observed in previous studies (e.g. Farquharson et al., 2015; Kennedy et al., 2015; Kushnir et al., 2016), the permeability–porosity relationship of porous volcanic rocks can be described by a relatively low power law exponent above a certain threshold in porosity. This is thought to be due to the fact that above this threshold effective fluid pathways (a so-called “backbone” of permeability) must—necessarily—already exist. Thus, a large increase in porosity by rapid bubble expansion, as inferred in Section 5.1, is likely to be associated with only a marginal increase in permeability. Similarly, we may assume that pure-shear flattening of these bubbles will not significantly decrease permeability until a threshold level of compaction.

Sample COL-V-5 has been interpreted as preserving a simple shearing process. We note that the permeability of the darker coloured side ( $1.37 \times 10^{-14} \text{ m}^2$ ) is measurably higher than that of the lighter coloured portion ( $2.90\text{--}6.24 \times 10^{-14} \text{ m}^2$ ; Table 3; Fig. 9a). Thus, in samples where the core axis is parallel to the discontinuity, the permeability is controlled by that of the most permeable region (on the order of  $10^{-14} \text{ m}^2$ ). Conversely, when the core axis is perpendicular to the discontinuity, the permeability is comparable to the lower value, on the order of  $10^{-16} \text{ m}^2$  (Fig. 9a). As the fraction of connected porosity shows negligible variation, we hypothesise that the differences in permeability must be driven in this case by some other corresponding parameter. As noted previously, the pore size and pore size distribution differs between the relatively high- and low-permeability regions. In turn this presumably influences the mean aperture of fluid pathways, as well as other factors such as the relative tortuosity of each pathway. Local differences in the stressing history of the magma may therefore provide a partial explanation for the wide variability in volcanic rock permeability observed in nature (e.g. Mueller et al., 2005; Mueller 2006; Farquharson et al., 2015).

Many of the discontinuity-bearing samples discussed herein are considered to represent variably-sintered fractures. For one of these—sample COL-V-4 (Fig. 4j–l)—we obtain fracture permeabilities on the order of  $1.0 \times 10^{-13} \text{ m}^2$ , approximately three orders of magnitude greater than the mean permeability of non-fractured cores of this sample ( $1.6 \times 10^{-16} \text{ m}^2$ ). It is thus evident that these fractures comprise a highly efficient conduit for fluid flow, despite being partially healed (experiments suggest that the unhealed fracture permeability of a 20 mm-diameter sample containing one through-going fracture is in the vicinity of  $10^{-11} \text{ m}^2$ ; Heap and Kennedy, in review). This supports the inference of previous authors (e.g. Edmonds and Herd, 2007; Castro et al., 2014), that magma fracture is an effective means of outgassing volatiles from conduit-dwelling magma. However, fractures observed in some of our other samples show evidence of more mature sintering (summarised in Fig. 6f–j), and are inferred to originate from greater depths or suggest longer healing timescales. It has been previously suggested through modelling that sintering timescales decrease



with depth due to the correspondingly high magmatic (and lithostatic) pressure and low effective viscosity (Russell and Quane, 2005; Quane et al., 2009; Heap et al., 2015c). However, it has been

noted that volatile resorption at low pressures may actually serve to accelerate viscous sintering (Sparks et al., 1999) by locally reducing the melt viscosity (Hess and Dingwell, 1996). Viscous sintering results in densification and a reduction of the connectivity of the porous network (e.g. Vasseur et al., 2013), and is commonly recorded in volcanic environments (e.g. Grunder and Russell, 2005; Quane et al., 2009; Kolzenburg and Russell, 2014). This process correspondingly reduces permeability (Heap et al., 2015c; Wadsworth et al., 2016), ultimately tantamount to the destruction of fluid pathways in magma.

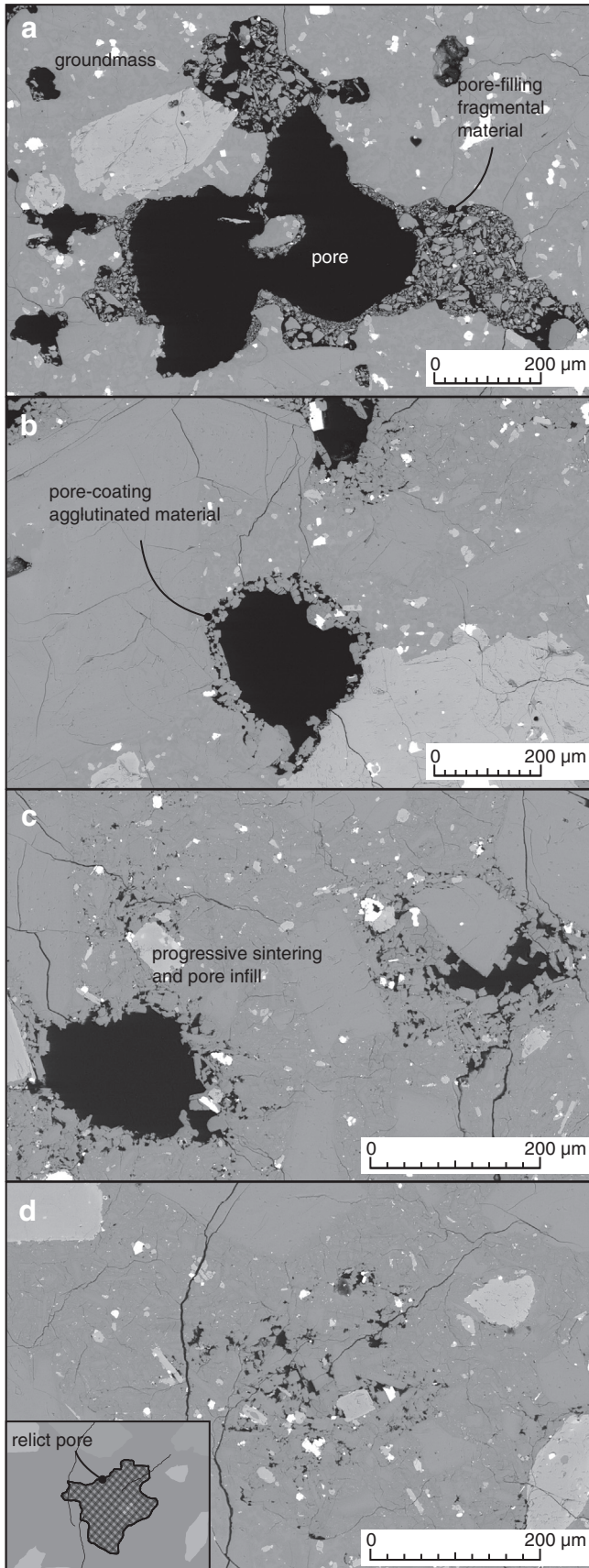
The infilled and relatively low-porosity band of sample Col-L-4 (Figs. 4v–x; 7), yields a fracture permeability of  $6.7 \times 10^{-14} \text{ m}^2$ , approximately an order of magnitude lower than that of the surrounding rock mass ( $2.4 \times 10^{-13} \text{ m}^2$ ); correspondingly the relative permeability  $k_r$  is negative (Table 4). To preferentially permit the passage of hot ash-laden fluids (as discussed in Section 5.4), this now low-porosity band used to be, in all likelihood, a band of higher porosity and permeability than the surrounding host rock. Although the preserved features studied herein only provide snapshots of dynamic processes, this indicates a strong time-dependence of permeability in magma: once a fracture is generated, fault gouge and preferentially transported fragmental material will quickly sinter due to their small size (e.g. Wadsworth et al., 2015). In turn, permeable fractures, cavities, or bubble networks, may progressively decrease in permeability until they are of comparable or, in some instances, lower permeability than the original magma.

In summary, we typically observe that if a feature acts as a conduit for fluid flow, then the orientation of the feature parallel to fluid flow has a markedly greater influence on permeability than when the feature is normal to fluid flow. On the other hand, a feature acting as a barrier for fluid flow will exert its maximum influence when oriented perpendicular to fluid flow. This is due to the difference between fluid flow in parallel and serial systems. Simply put, when offered the choice of parallel layers, fluid will preferentially flow through the layer that is most permeable, and the overall permeability is thus governed by the most permeable element. Contrastingly, in a serial system, fluids must flow through each and every layer, and permeability is thus controlled by the layer with the lowest permeability.

## 7. Implications

### 7.1. Outgassing and volcanic activity

Fig. 10 summarises the influence of processes discussed in the previous two sections. Magma fracture causes relatively large changes in magma permeability, over relatively small increments in porosity (Fig. 10a). Indeed, our data suggest that permeability may increase by three orders of magnitude or more for a change in porosity of only around 0.03 or 0.04 (Table 3). Similarly, shear-induced transtension or cavitation of magma is predicted to yield a local increase in porosity and permeability (Fig. 10a); our data for sample COL-M-1, for example, indicates that for comparable porosities (between 0.05 and 0.07), permeability of the magma can be increased by around two orders of magnitude if these porous discontinuities are preferentially oriented (Table 3; Fig. 9a). By contrast, densifying processes—e.g. viscous sintering or transpression—result in decreases in both porosity and



**Fig. 7.** Progressive pore infill in sample COL-L-4, from near the margin (a) towards the centre (d) of the band. (a) A pore, partially infilled with fragmental material. (b) Agglutinated material adhering to the pore wall. Textures shown in (c) indicate that the infilling material has progressively clogged up pores in this sample. Finally, in (d) the pore is almost entirely filled in, and the infilling material is well sintered. The previous extent of the pore is preserved as a patch of microporosity (sketched in inset).



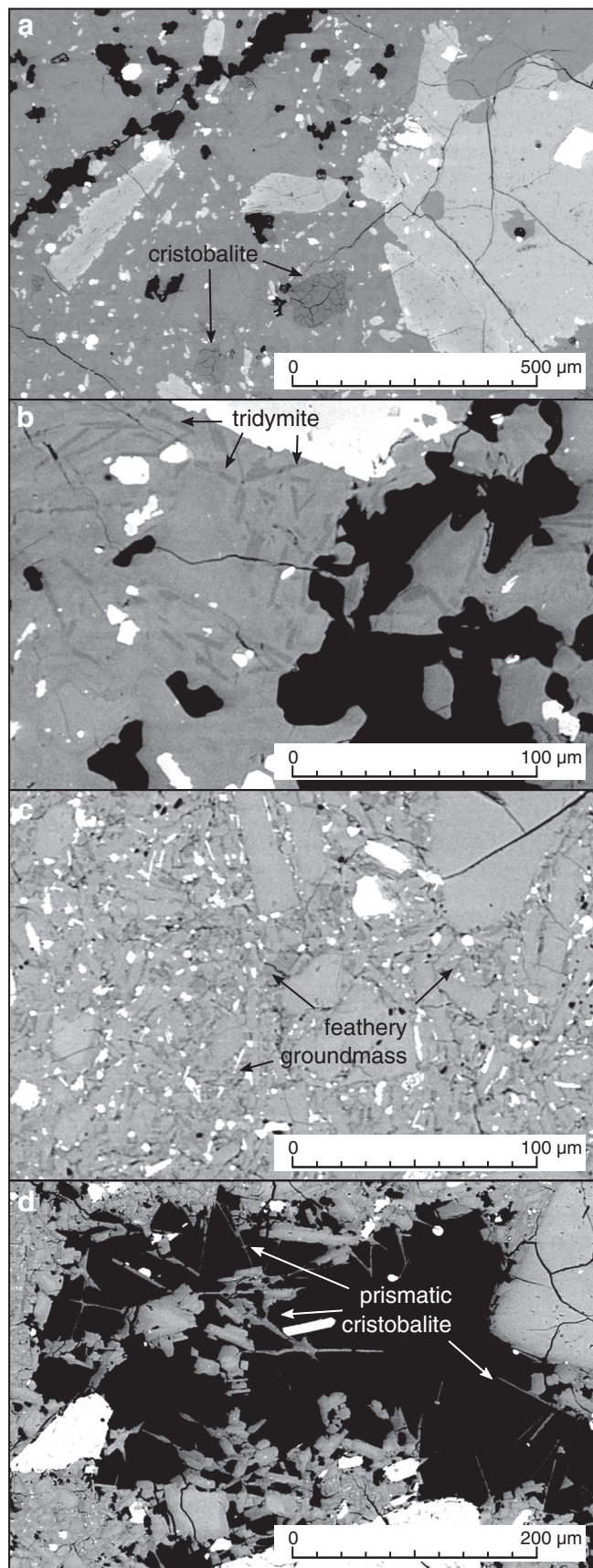
permeability (Fig. 10b). Fig. 10a and b further illustrate the contrasting influences of bubble expansion and the collapse of foaming magma. Because the magma is already highly permeable—due to its high initial

porosity—increases or decreases in porosity of tens of percentage points (e.g. from  $\phi \approx 0.30$  to  $\phi \approx 0.60$ ) are not expected to have a marked influence on the permeability of the magma (less than one order of magnitude difference).

It is important to note that—irrespective of the processes acting on the physical properties of a specific clast or unit of magma (Table 3; Fig. 9)—we observe that the permeability–porosity relationship of heterogeneity-bearing rocks (and indeed, the permeability–porosity relationship of the features themselves) appears always to follow a trend remarkably similar to that of andesitic rock free of macroscopic heterogeneities (shown in Fig. 10c). It follows that at low magma porosities (i.e. lower than around 0.15 or 0.20: see Fig. 9a), permeability evolution is dominated by the generation and healing of fractures, just as in low-porosity volcanic rocks the permeability evolution is strongly governed by the generation or closure of microcracks (Vinciguerra et al., 2005; Nara et al., 2011; Heap et al., 2014b, 2015a; Farquharson et al., 2015). At higher porosities, the influence of bubble growth, bubble coalescence, and expansion-driven fragmentation becomes dominant (Fig. 9b). Thus, dilatant or densifying processes acting on the magma will have a greater marginal influence at low initial porosities than at high initial porosities. In short, our results indicate that an increase or decrease in porosity or permeability will move the magma more-or-less along the paths indicated in panels a and b of Fig. 10.

Our results show that permeability can be increased locally by as much as three orders of magnitude due to dilatant processes (see Fig. 9, Table 3), even if partial sintering has occurred. Such efficient fluid pathways may play an extremely important role in the outgassing of magmatic volatiles, in turn decreasing the propensity for fluid pressurisation and violent explosive behaviour in active volcanic systems. Fracture geometry (i.e. length and aperture) will influence the initial gas volume that can enter and escape through the fracture. If a fracture constitutes a pathway for magmatic fluids to outgas from the system (e.g. into the edifice rock: Heiken et al., 1988; Kolzenburg et al., 2012) then the continuous through-flow of gas has been posited to sustain an open fracture, as long as there exists a persistent supply of gas from beneath which is decoupled from the melt (plus crystal) phase of the magma (e.g. Rust et al., 2004; Plail et al., 2014). Indeed, increased pore fluid pressures are thought to occur, albeit transiently, in particle-filled fractures within magma (e.g. Castro et al., 2014). We anticipate that relatively large volumes of gas can be decoupled from the magma and thus flow through these fractures: a significantly more efficient means of transporting volatiles than bubble migration through a viscous silicate melt, or indeed diffusion-scale processes. We highlight that fractures in magma can allow outgassing even at very low porosities.

As magma fracturing will preferentially occur where strain rates are thought to be highest (e.g. Papale, 1999; Gonnermann and Manga, 2003; Edmonds and Herd, 2007), permeable fractures are likely to contribute to a “damage halo” zone surrounding the central conduit (e.g. Lavallée et al., 2013; Gaunt et al., 2014; Young and Gottsmann, 2015). Microstructural evidence presented here hints that these fractures occur both in shallow regions (textural evidence of silica polymorphs in fractures point to high-temperature, low-pressure conditions) and deeper in the system (advanced sintering textures observed in some samples indicate high-temperature, high-pressure environments), suggesting that this halo may serve to bleed off volatiles from the magma and promote outgassing into the edifice rock or through near-conduit



**Fig. 8.** Textural evidence for silica polymorphs. (a) Dark grey patches are cristobalite crystals, showing characteristic “fish-scale” cracking in sample Col-M-1. (b) Dark grey, high aspect ratio crystals are acicular tridymite (sample Col-L-2). (c) The groundmass of sample Col-V-3 has a “feathery” texture, indicative of devitrification of the groundmass (see Horwell et al., 2012, 2013). (d) High aspect ratio crystals inferred to be prismatic cristobalite are often seen growing into pore spaces (sample Col-L-4).

**Table 3**

Connected gas porosity and gas permeability data for core samples. The table also indicates whether the core was obtained parallel or perpendicular to the feature in the sample, and whether or not the feature was present in the core.

Sample	Gas porosity $\phi$	Gas permeability $k$ ( $\text{m}^2$ )	Core orientation with respect to flow direction	Feature
COL-V-1-Y	0.35	$7.98 \times 10^{-14}$	Parallel	Present
COL-V-1-Z	0.43	$5.44 \times 10^{-14}$	Perpendicular	Present
COL-V-6-Z <sub>1</sub>	0.43	$3.42 \times 10^{-13}$	Perpendicular	Present
COL-V-6-Z <sub>2</sub>	0.44	$3.40 \times 10^{-13}$	Perpendicular	Present
COL-V-6-Z <sub>3</sub>	0.62	$1.18 \times 10^{-13}$	Perpendicular	Absent
COL-M-1-Y <sub>1</sub>	0.07	$5.16 \times 10^{-14}$	Parallel	Present
COL-M-1-Y <sub>2</sub>	0.06	$4.18 \times 10^{-14}$	Parallel	Present
COL-M-1-Z <sub>1</sub>	0.06	$1.46 \times 10^{-16}$	Perpendicular	Present
COL-M-1-Z <sub>2</sub>	0.05	$2.40 \times 10^{-16}$	Perpendicular	Present
COL-L-1-Y <sub>1</sub>	0.08	$2.84 \times 10^{-16}$	Parallel	Present
COL-L-1-Y <sub>2</sub>	0.03	$1.36 \times 10^{-17}$	Parallel	Absent
COL-L-1-Z <sub>1</sub>	0.03	$1.29 \times 10^{-18}$	Perpendicular	Absent
COL-L-1-Z <sub>2</sub>	0.08	$4.24 \times 10^{-16}$	Perpendicular	Present
COL-V-5-Y <sub>BLACK</sub>	0.13	$1.05 \times 10^{-14}$	Parallel	Present
COL-V-5-Y <sub>BOTH</sub>	0.11	$3.03 \times 10^{-15}$	Parallel	Present
COL-V-5-Y <sub>GREY</sub>	0.15	$6.24 \times 10^{-16}$	Parallel	Absent
COL-V-5-Z <sub>BLACK</sub>	0.15	$1.37 \times 10^{-14}$	Perpendicular	Present
COL-V-5-Z <sub>GREY</sub>	0.13	$2.90 \times 10^{-16}$	Perpendicular	Absent
COL-V-5-Z <sub>BOTH</sub>	0.12	$2.48 \times 10^{-16}$	Perpendicular	Present
COL-L-4-Y <sub>1</sub>	0.13	$2.40 \times 10^{-13}$	Parallel	Absent
COL-L-4-Y <sub>2</sub>	0.09	$3.18 \times 10^{-14}$	Parallel	Present
COL-L-4-Z <sub>1</sub>	0.13	$1.11 \times 10^{-14}$	Perpendicular	Present
COL-L-4-Z <sub>2</sub>	0.15	$3.30 \times 10^{-13}$	Perpendicular	Absent
COL-L-4-Z <sub>3</sub>	0.14	$6.21 \times 10^{-14}$	Perpendicular	Present
COL-L-4-Z <sub>4</sub>	0.14	$1.76 \times 10^{-13}$	Perpendicular	Present
COL-V-3-Y <sub>1</sub>	0.07	$3.45 \times 10^{-16}$	Parallel	Present
COL-V-3-Y <sub>2</sub>	0.07	$8.03 \times 10^{-16}$	Parallel	Present
COL-V-3-Y <sub>3</sub>	0.06	$1.56 \times 10^{-16}$	Parallel	Present
COL-V-3-Y <sub>4</sub>	0.09	$1.22 \times 10^{-15}$	Parallel	Present
COL-V-3-Z <sub>1</sub>	0.10	$1.41 \times 10^{-16}$	Perpendicular	Present
COL-V-3-Z <sub>2</sub>	0.08	$4.10 \times 10^{-16}$	Perpendicular	Present
COL-V-3-Z <sub>3</sub>	0.10	$8.98 \times 10^{-16}$	Perpendicular	Present
COL-V-4-Y <sub>1</sub>	0.09	$4.23 \times 10^{-14}$	Parallel	Present
COL-V-4-Y <sub>2</sub>	0.09	$1.56 \times 10^{-14}$	Parallel	Present
COL-V-4-Y <sub>4</sub>	0.08	$1.78 \times 10^{-14}$	Parallel	Present
COL-V-4-Y <sub>5</sub>	0.13	$4.70 \times 10^{-14}$	Parallel	Present
COL-V-4-Y <sub>6</sub>	0.08	$2.41 \times 10^{-14}$	Parallel	Present
COL-V-4-Z <sub>1</sub>	0.06	$3.00 \times 10^{-16}$	Perpendicular	Absent
COL-V-4-Z <sub>2</sub>	0.08	$2.99 \times 10^{-16}$	Perpendicular	Present
COL-V-4-Z <sub>3</sub>	0.04	$1.29 \times 10^{-16}$	Perpendicular	Absent
COL-V-4-Z <sub>4</sub>	0.08	$2.50 \times 10^{-16}$	Perpendicular	Present
COL-V-4-Z <sub>5</sub>	0.09	$1.70 \times 10^{-16}$	Perpendicular	Present
COL-V-4-Z <sub>6</sub>	0.04	$1.32 \times 10^{-16}$	Perpendicular	Present
COL-V-4-Z <sub>7</sub>	0.07	$2.53 \times 10^{-16}$	Perpendicular	Present
COL-V-4-Z <sub>8</sub>	0.06	$2.83 \times 10^{-16}$	Perpendicular	Present
COL-V-4-Z <sub>9</sub>	0.03	$9.77 \times 10^{-17}$	Perpendicular	Absent
COL-V-4-Z <sub>10</sub>	0.03	$1.05 \times 10^{-16}$	Perpendicular	Absent
COL-L-2-Y <sub>1</sub>	0.01	$1.77 \times 10^{-16}$	Parallel	Absent
COL-L-2-Y <sub>2</sub>	0.05	$3.09 \times 10^{-14}$	Parallel	Present
COL-L-2-Y <sub>3</sub>	0.05	$3.30 \times 10^{-14}$	Parallel	Present

fracture networks along the length of the conduit. Fault gouge and ash entrained in escaping fluids may accelerate sealing or closure of pathways (e.g. Tuffen et al., 2003; Saubin et al., in press; Kendrick et al., 2016). However, it has been shown numerically and experimentally that the relatively slow post-fracture recovery of magma strength may actually promote recrudescence fracture events (Heap et al., 2015b, 2015c), which in turn may serve to regulate lava dome eruptions (Heap et al., 2015c; Kendrick et al., 2016).

The extraction of magmatic volatiles due to shear strain-induced cavitation—as we observe evidence for herein—has similarly been posited to result in premature embrittlement of magma (Smith et al., 2001), assisting the recurrence of fracture events. Repeat faulting events in magma have been proposed as a mechanism for earthquake generation during magma ascent (e.g. Tuffen and

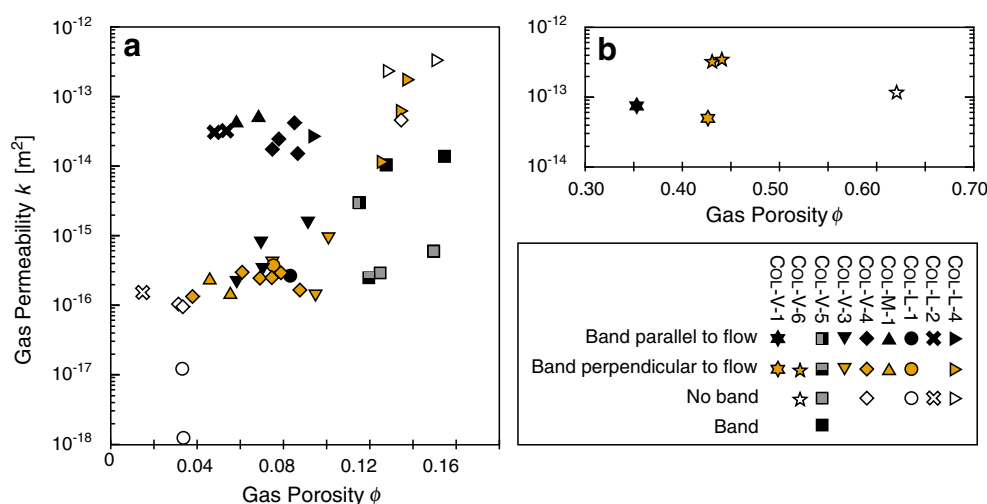
Dingwell, 2005; Neuberg et al., 2006; Lavallée et al., 2008; Tuffen et al., 2008; Varley et al., 2010a, 2010b; Kendrick et al., 2014a; Heap et al., 2015c).

We observe a permeability reduction of an order of magnitude relative to the host material in the sample containing a band of lower porosity characterised by pores that have been variably infilled with ash-sized particles that have sintered to the pore walls. One inference that may be drawn from this is that barriers to flow (viscous densification, sintering of fractures, and infill of pores with granular material or vapour-phase crystallisation) may trap exsolved volatiles and allow the build-up of pore pressure. Iterative densification events may compartmentalise parcels of magma, create pore overpressures, and result in complex permeability anisotropy within the conduit or the dome. We anticipate that the preferential flow of ash-laden fluid along porous networks—that is, interconnected bubble chains or fractures—will cause a time-dependent permeability anisotropy at the periphery of the conduit. The permeability reduction associated with this process (Figs. 7, 9a) means that a pathway that preferentially transmits fluids can consequently evolve into a barrier: the relative permeability  $k_r$  (Eq. (2)) will transition from a positive to a negative value. Consequently, fluids will be forced to find another route through the magma.

As previously described, the strain profile across a volcanic conduit or dyke is typically assumed to vary, with strains and strain rates being highest at the margins (e.g. Gonnermann and Manga, 2003). Consequently, we may expect fractures in magma to be oriented subparallel to the direction of magma flow (thus in a traditionally-envisaged volcanic conduit, magma fractures will be predominantly oriented subvertically). As such, permeability-increasing fractures will be preferentially oriented in the conduit to facilitate annular or halo outgassing. However, effective sintering of such fractures could promote a build-up of overpressure, by limiting lateral migration of volatiles from the conduit into the edifice or an intermediate fractured zone. There also exists field evidence for the generation of tensile fractures that propagate laterally into the edifice (see Heiken et al., 1988; Goto et al., 2008). In this case, features acting as pathways for flow will likely improve outgassing into the country rock, whereas barrier-forming features will not majorly influence diffuse edifice outgassing. Similarly, large-scale fractures throughout a lava dome are often oriented vertically to subvertically (e.g. Walter et al., 2015) and will likely serve to effectively bleed off volatiles. Due to the inherently fractured nature of many lava domes, we anticipate that low-permeability features resulting from local hydrothermal alteration, mineral precipitation, or ash deposition (e.g. Edmonds et al., 2003; Ball et al., 2015) may not significantly reduce their overall outgassing potential. However, such sealing of permeable pathways may be of great importance in the shallow conduit, where restricting the migration of an exsolved gas phase could rapidly elevate pore pressure in the system, reducing structural integrity and increasing the likelihood of explosive behaviour.

Furthermore, the length of time that features with high relative permeability can act as effective outgassing networks relies on the interplay between the incumbent stress field, the through-flow of gas (e.g. Rust et al., 2004) and effective viscosity- and pressure-dependent densification via compaction or sintering (e.g. Quane et al., 2009; Kolzenburg and Russell, 2014; Vasseur et al., 2013; Heap et al., 2015c) as well as ancillary processes such as the transport of ash and fault gouge (e.g. Tuffen et al., 2003; Kendrick et al., 2016), and hydrothermal mineral precipitation (e.g. Edmonds et al., 2003). We emphasise that providing accurate timescales for volcanic processes—even those that are relatively well understood—is inherently problematic due to the lack of knowledge regarding the formation conditions (e.g. depth, temperature, local strain rates) or the state of the initial material (porosity, viscosity, grain size distribution etc.). In reality, all of the processes discussed herein are likely to be





**Fig. 9.** Connected gas porosity ( $\phi$ ) and gas permeability ( $k$ ) data for all fifty core samples. (a) Low-porosity samples ( $\phi < 0.20$ ). (b) High-porosity samples ( $\phi > 0.30$ ). Note the difference in scale of the x-axis between either panel. Different blocks are indicated by different symbols, whilst the presence or orientation of a feature is given by the symbol colour. Measurement error is less than the symbol size in each case. Refer to text for discussion.

concurrent within comparable subaerial volcanic systems. The relative proportion of dilatant vs. densifying processes may heavily influence transitions between eruptive regimes and govern, at least partially, repose periods of Vulcanian explosive activity (Quane et al., 2009; Collinson and Neuberg, 2012): dilatational processes should bridle eruptive behaviour and lengthen repose times, whilst sintering and densification should augment the increase of pore pressure, and decrease repose times. Modelling recurrence intervals during periods of intermittent explosive activity is often fraught with complexity (e.g. Varley et al., 2006), much of which may be attributed to the contrasting influences of dilatant and densifying processes throughout volcanic systems. However, this interplay remains poorly understood, and we encourage its description through systematic sets of controlled experiments.

## 7.2. Considerations for scaling and modelling

The blocks collected for this study were of hand sample size. However, even cursory field observations indicate that variably tabular discontinuities in volcanic material can be seen on much greater lengthscales. Moreover, in the initial stages of their genesis, these features may be orders of magnitude smaller than those discussed herein. Such intra-clast heterogeneity may well be responsible for the large variation in permeabilities measured in volcanic rocks of similar porosities (e.g. Mueller et al., 2005, 2008; Farquharson et al., 2015). As the permeability of fractured volcanic rocks is greatly

scale-dependent (Heap and Kennedy, in review), it is of importance that the permeable structures discussed in this contribution are discussed in this context.

Although accurate constraints of subsurface volcanic architecture are rare, results from drilling projects suggest that surficial vent and crater systems on the order of 10s or 100s of m in diameter can belie conduits of relatively narrow width (Mastin and Pollard, 1988). For example, Noguchi et al. (2008) describe the dacitic conduit of Mount Unzen (Japan) as ranging from 4 to 40 m in diameter to a depth of 2 km. This is in agreement with studies of exposed feeder dykes (e.g. Keating et al., 2008; Galindo and Gudmundsson, 2012), which show that, deeper than around 50 m, solidified dykes tend to have diameters of only a few m. As a consequence, even relatively small discontinuities—such as those described herein—may constitute a non-negligible portion of the conduit width. Whilst the thickness of such features as a proportion of the conduit width will certainly decrease as the conduit widens, we anticipate that in many cases their abundance and longevity will be augmented in shallower systems.

In our study, the equivalent permeability of each block is heavily dependent on its initial size and the area of the discontinuity and the ratios of their permeabilities, just as Eq. (1) is strongly influenced by the ratios of  $A_s$  to  $A_f$  and  $k_i$  to  $k_f$ . The partially sintered fracture of COL-V-4, for example, was found to have a permeability of  $1.04 \times 10^{-13} \text{ m}^2$ , yielding an equivalent permeability of  $2.94 \times 10^{-14} \text{ m}^2$  in a core sample of approximately 20 mm

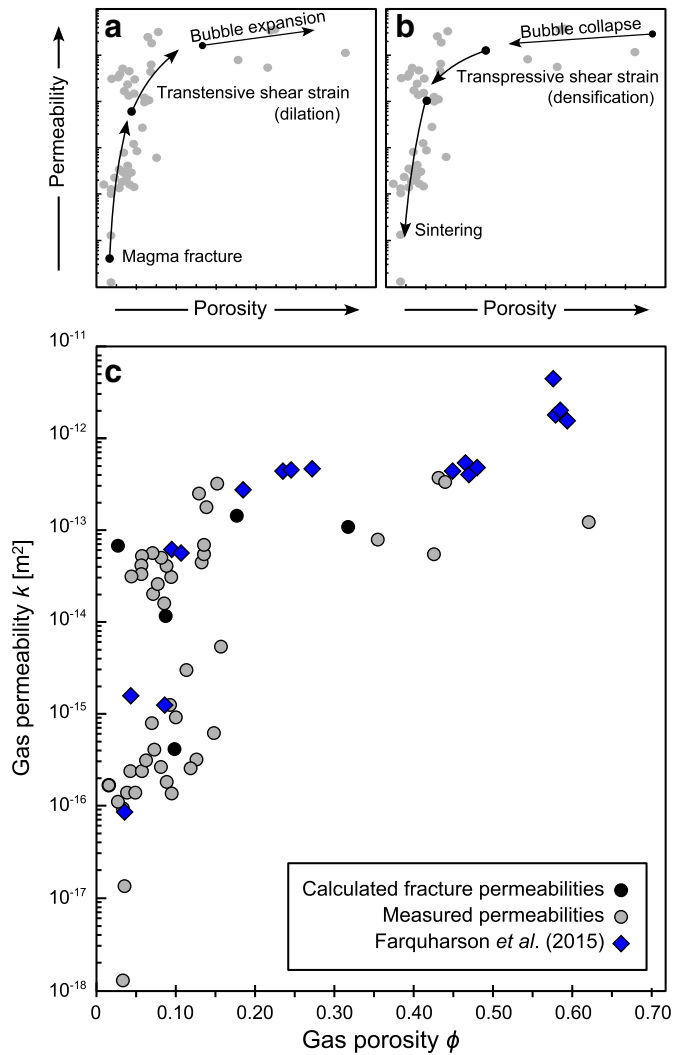
**Table 4**  
Mean permeabilities and dimensions used to calculate the permeability of the feature for five samples. Note that we use “fracture” permeability to refer to the permeability of any of the discussed features, regardless of genesis.

Sample	Intact permeability $k_i$ ( $\text{m}^2$ )	Equivalent permeability $k_e$ ( $\text{m}^2$ )	Sample area ( $\text{mm}^2$ )	Fracture area ( $\text{mm}^2$ )	Fracture permeability ( $\text{m}^2$ )	Relative permeability $k_r$
COL-V-5	$1.37 \times 10^{-16}$	N/A <sup>a</sup>	N/A <sup>a</sup>	N/A <sup>a</sup>	$1.21 \times 10^{-14}$	87.32
COL-V-4	$1.58 \times 10^{-16}$	$2.94 \times 10^{-14}$	312.84	74.1581	$1.04 \times 10^{-13b}$	657.23
COL-L-1	$1.36 \times 10^{-17}$	$2.84 \times 10^{-16}$	312.87	128.23	$3.90 \times 10^{-16}$	27.68
COL-L-2	$1.77 \times 10^{-16}$	$3.19 \times 10^{-14}$	310.92	58.52	$1.37 \times 10^{-13c}$	773.01
COL-L-4	$2.40 \times 10^{-13}$	$3.18 \times 10^{-14}$	308.19	44.80	$6.67 \times 10^{-14}$	−0.72

<sup>a</sup> For COL-V-5 (Fig. 2c) it was possible to obtain cores containing only the host material or only the band (see Table 3). Thus we have a discrete measure of the band permeability relative to the initial material, and no calculation is necessary.

<sup>b</sup> Value shown is mean value determined from fracture permeabilities of two samples COL-V-4-Y<sub>1</sub> and -Y<sub>2</sub>. Values range from  $1.67 \times 10^{-13}$ – $4.12 \times 10^{-14}$ .

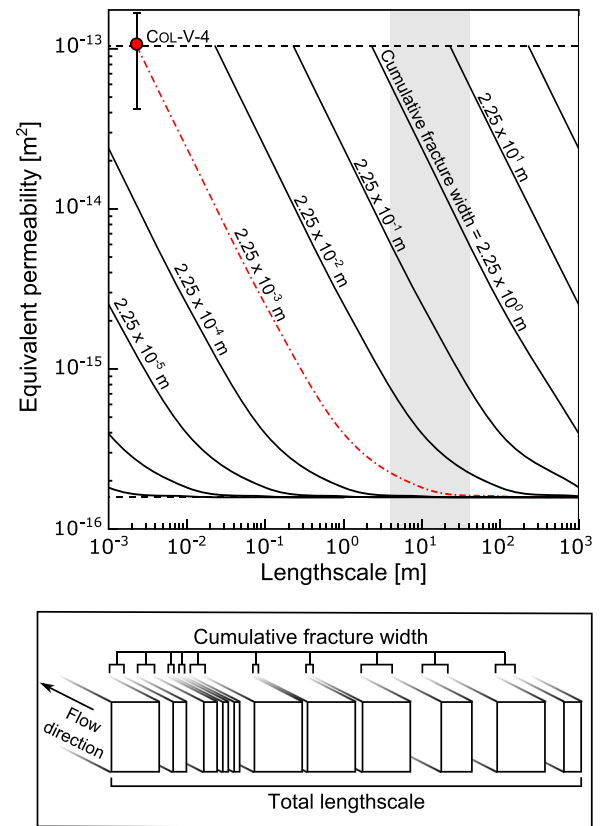
<sup>c</sup> Value shown is mean value determined from fracture permeabilities of two samples COL-L-2-Y<sub>1</sub>, and -Y<sub>2</sub>. Values range from  $1.32 \times 10^{-13}$ – $1.42 \times 10^{-13}$ .



**Fig. 10.** The influence of magmatic processes on the development of permeability, as predicted from our microstructural interpretations. (a) Dilatant (porosity-increasing) processes, including magma fracture, shear-induced extension, cavitation, and bubble expansion, exhibit a non-linear effect on permeability: at low porosities, large increases in permeability may be observed and vice versa. (b) Compactant or densifying processes—such as sintering, pore infill, and bubble collapse—similarly exert a large influence when the initial porosity is low, and only a slight influence when the initial porosity and permeability are high. (c) Collated permeability and porosity data. Despite the inclusion or orientation of the heterogeneities discussed throughout this study, the permeability–porosity data follow essentially the same trend as that of collected band-free samples (data taken from Farquharson et al., 2015). Further, the data for the bands themselves (calculated permeability from Eq. (1) in concert with 2D SEM-derived porosities: from Tables 2, 4) also follow this trend. Measurement error is less than the symbol size in each case.

diameter (Table 4). To upscale these data, we employ a 1D version of Eq. (1), wherein  $A_s$  and  $A_f$  are substituted for  $w_s$  and  $w_f$ : the width of the sample and fracture, respectively:  $k_f = k_e w_s - k_i(w_s - w_f)/w_f$  (as in Heap and Kennedy, in review). Hence, a fracture of 2.25 mm width in a sample 10 m wide (with the same host permeability:  $1.58 \times 10^{-16}$  m<sup>2</sup>) would result in an equivalent permeability of  $1.81 \times 10^{-16}$  m<sup>2</sup>, not significantly higher than that of the host material. In a sample 1000 m in width, the overall influence of the same fracture would be effectively zero ( $k_e = 1.58 \times 10^{-16}$  m<sup>2</sup>).

In contrast, if the fracture width approaches that of the considered lengthscale, then its influence on equivalent permeability will be correspondingly greater. Similarly, an increase in the



**Fig. 11.** The scale-dependence of permeability anisotropy. Equivalent permeability (Eq. (1)) is shown for a fracture or set of fractures within porous magma. As the permeabilities of the fracture and the host material remain constant (defined by the upper and lower dashed lines respectively), equivalent permeability relies on both the considered lengthscale and the cumulative fracture width (i.e. number and width of fractures). Measured value of sample Col-V-4 is shown for reference. Note that  $k_f \geq k_e \geq k_i$ . The shaded area corresponds to the width of a volcanic conduit (see Noguchi et al., 2008); red dashed line describes the influence of a 2.25 mm fracture with the fluid transport properties of that in sample Col-L-4. 1D upscaling model defines flow in parallel cracks, such that the total lengthscale is divided into parallel blocks or plates, shown inset. Refer to text for more discussion. (For interpretation of the references to color in this figure legend, the reader is referred to the online version of this article.)

number of fractures within a given volume of rock or magma will increase the equivalent permeability to a greater or lesser extent, depending on their geometries and fluid transport properties. These concepts are illustrated in Fig. 11, using the example of Col-V-4 (equally, data from other features—Table 4—could be analysed using this method). The lengthscale refers to the total width of material considered, for example the width of a volcanic conduit at a given depth. For reference, the range of widths measured by Noguchi et al. (2008) is shaded grey in Fig. 11. We highlight that even a single fracture with the thickness and fluid transport properties of that in Col-L-4 can increase the equivalent permeability of a conduit a few m in width (i.e. the intersection of the red line with the shaded region). However, the same fracture in a conduit with a diameter of 40 m has a negligible influence. In this simplified context, adding new fractures or widening an existing fracture has the same result, namely increasing the “cumulative fracture width”. We highlight that in reality, increasing or decreasing the width of a fracture will introduce complexities associated with non-laminar flow (e.g. Heap and Kennedy, in review), which are not accounted for in Fig. 11. Nevertheless, Fig. 11 demonstrates the significance of considering upscaling from laboratory samples to outcrop-scale and larger. Accurately incorporating scale



effects remains an outstanding challenge in volcanology, especially in transferring experimental data to eruption models and hazard mitigation policies.

To advance our understanding of volcanic processes and better predict eruptive scenarios requires the development of models which effectively reproduce eruption characteristics observed in nature (e.g. Melnik, 2000; Burgisser and Gardner, 2004; Collombet, 2009), especially the transitions and timescales between explosive and effusive periods (e.g. Jaupart, 1996, 1998; Collinson and Neuberg, 2012). Based on our study, we suggest that future models assessing conduit dynamics and permeability evolution consider the propensity for permeability anisotropy within the magma column and lava dome on multiple lengthscales. Given that such heterogeneities are posited to occur primarily at the conduit margins or in high-strain regions of extruding domes, this could be modelled by stochastic inclusion of transient variations in magma permeability, by a coupling of permeability and shear strain across the width of the conduit, or by the calculation of the (time-dependent) equivalent permeability of fracturing conduit margins (as touched on in Fig. 11 and the attendant discussion).

### 7.3. Field interpretations

We note that in the absence of microstructural examination, interpretation of some textural or structural heterogeneities encountered in volcanic materials can be nontrivial. Despite being visually similar at a hand sample scale, our study has shown that these features are not equally efficient at transmitting fluids. Thus we highly recommend that any interpretation of heterogeneity-bearing volcanic rocks—especially with respect to their influence on permeability and the attendant implications for outgassing—should be buttressed by microstructural description and/or measurements of permeability or porosity.

As a final comment: due to the accretionary growth of stratovolcanoes (e.g. Odbert et al., 2015), explosively or effusively erupted lava eventually comprises some portion of the edifice. Thus, if fluid flow pathways or barriers are erupted and preserved—as was the case with our sample set—we may also expect them to increase permeability anisotropy and influence fluid partitioning within the edifice. We anticipate their presence throughout the upper edifice at Volcán de Colima and indeed at other comparable volcanic systems (such discontinuity-bearing rocks can be seen at throughout the Taupō Volcanic Zone in New Zealand for example). However, a quantitative analysis of their appearance at the surface could be somewhat misleading. For example, it has been noted that features that provide effective outgassing pathways are, by definition, less likely to be erupted out of the volcano than ineffective (or nonexistent) pathways (Plail et al., 2014). Similarly, Castro et al. (2012) intimates that the pyroclastic origin of tuffisite-bearing bombs indicates that systems containing highly permeable discontinuities can still be subject to significant overpressure development. Further, the preservation of clasts containing fractures may be poor, especially if the fracture is poorly-sintered. Nevertheless, we stress that discontinuity-bearing volcanic materials can provide incredibly informative snapshots of otherwise unobservable magmatic processes, making their continued study a valuable pursuit in volcanology.

## 8. Concluding remarks

Tabular heterogeneities in andesitic rocks—of both explosive and effusive origin—can be found in abundance at Volcán de Colima (Mexico), providing frozen-in snapshots of magmatic processes. Alongside an examination of their microstructure, a systematic laboratory study of the physical properties of a selection of these

heterogeneity-bearing blocks allows us to glean important information regarding the likely geneses of these features within a volcanic system, as well as their influence on the physical properties of the magma (at least upon eruption). Bands were inferred to form in high-porosity pumice blocks ( $\phi$  around 0.30 and higher) due to inhomogeneous bubble expansion in magma. Despite a significant influence on the porosity of these samples, permeabilities of banded and non-banded pumice were found to be comparable (of the order  $10^{-13} \text{ m}^2$ ). This is thought to be due to the pre-established pathways of effectively connected porosity. In lower porosity blocks ( $\phi < 0.20$ ), features are preserved that increase porosity and permeability relative to the surrounding material. We suggest that these discontinuities may be the frozen relicts of dilatant processes such as viscous cavitation or magma fracturing. We find that these fractures can increase in permeability by about three orders of magnitude relative to the host rock, when they are oriented parallel to fluid flow. Evidence of subsequent fracture healing can also be observed in our microstructure, and our measurements indicate that fracture permeability decreases with more advanced sintering. Differences in the extent of sintering, and the presence of silica polymorphs in some samples, suggest that these features were likely formed at different depths in the conduit. The evidence therefore suggests that fractures exist—although in some cases, only temporarily—along an extensive portion of the conduit, supporting the concept of a permeable outgassing halo surrounding the conduit. Notably, textures interpreted herein as the infill of pores with transiently granular material (such as gas-transported volcanic ash) indicate that porous networks may subsequently become barriers to fluid flow over time, especially when oriented normal to fluid flow. This highlights a complex interplay between dilatant and densifying processes in magma, as well as the time-dependent evolution of its physical properties. We have presented evidence for significant permeability variation within conduit magma and dome at Volcán de Colima, and we anticipate that the features described here will also exist at similar andesitic stratovolcanoes worldwide, at a range of lengthscales. We contest that localised permeability heterogeneities must be a critical parameter influencing the evolution of pressures within active volcanic systems, and use a simple upscaling model to illustrate the evolving significance of permeability heterogeneities over different lengthscales. It is likely that, in turn, these features will likely influence the eruptive regime and recurrence intervals of explosions or degassing events at active volcanoes.

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## Appendix A



**Fig. A1.** Photographs of the core samples measured in this study. Scale shown in the top left is the same for each panel. Sample names indicate sampling location (V, M, and L refer to Volcancito, Montegrande, La Lumbre, respectively: Fig. 1a), and coring direction (Y and Z correspond to samples cored parallel and perpendicular to the feature of interest: Fig. 3a).



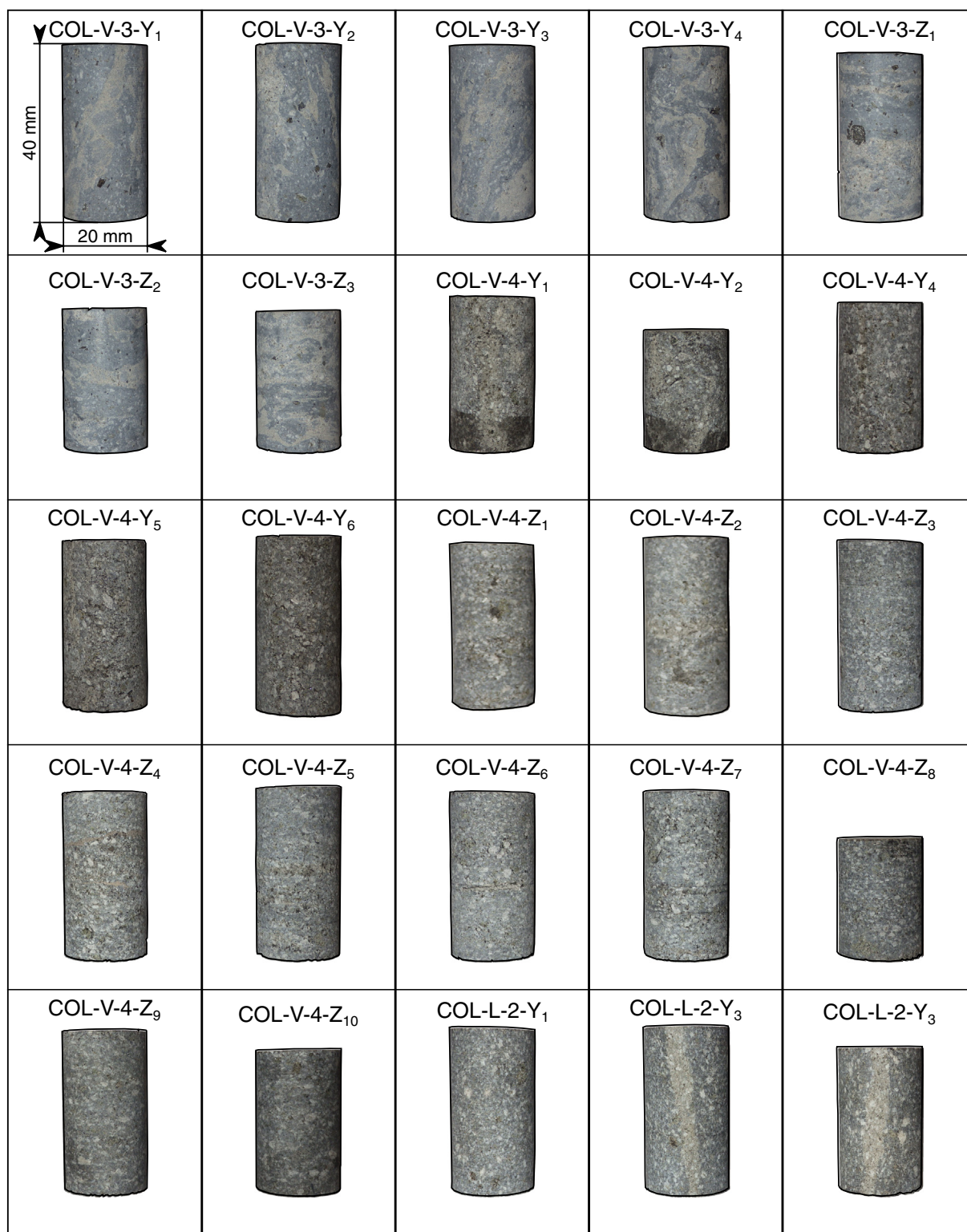


Fig. A1 (continued)

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