Time-and-temperature-dependent conduit wall porosity: A key control on degassing and explosivity at Tarawera volcano, New Zealand


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A B S T R A C T

The permeability of volcanic conduit walls and overlying plug can govern the degassing and explosivity of eruptions. At volcanoes characterized by a protracted history of episodic volcanism, conduit walls are commonly constructed of quenched magma. During each successive eruptive phase, reheating by ascending magma can modify the porosity, permeability and H₂O content of the conduit wall rocks and overlying plug. We investigate whether the unusual explosivity of the 1886 basaltic eruption at Tarawera volcano is related to the heating and degassing of the AD1314 Kaharoa rhodolite rocks, through which it erupted. We heat cores of perlitic Tarawera dome rhyolite to 300 °C–1200 °C for 30 min to 3 days at atmospheric pressure. We characterize time (t)- and temperature (T)-dependent variations in porosity, volatile content and texture through SEM image analyses. We also directly measure pre- and post-experimental connected and isolated porosity and water content. We identify four textural/outgassing regimes: Regime 1 (T ≤ 800 °C, t ≤ 2 h), with negligible textural changes and a significant loss of meteoric water (1.4–0.72 wt.% H₂O); Regime 2 (800 ≤ T ≤ 1100 °C, t ≤ 6 h), with cracking and vesicle growth and a 5–10% increase in connected porosity; Regime 3 (800 ≤ T ≤ 1200 °C, t ≥ 30 min), with healed cracks, coalesced and collapsed vesicles, and overall reduced porosity; and Regime 4 (T ≥ 1200 °C, t ≥ 30 min), with a collapse of all connected porosity. These regimes are governed by the temperature of the event (T) relative to the glass transition temperature (Tₐ) and the time scale of the event (t) relative to a critical relaxation time for structural failure of the melt (τᵣ). We identify a quantitative transition from predominantly brittle behavior such as cracking, which enhances connected porosity and permeability, to viscous processes including crack healing and vesicle collapse, which act to reduce connected porosity. Applied to the 1886 basalt eruption at Tarawera, we show that progressive heat transfer ultimately reduced the open porosity and permeability of the conduit walls, thereby partially sealing the conduit and reducing volatile loss. We argue that this mechanism was an underlying reason for the exceptional explosivity of the 1886 eruption. We further suggest that textural changes associated with reheating could explain some of the cyclic deformation and degassing observed at many lava domes preceding explosive eruptions.

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1. Introduction

Individual volcanic eruptions can shift rapidly in style from relatively quiescent lava dome extrusion to explosive eruptions (e.g. Sparks, 2003; Voight et al., 1999). These shifts are generally attributed to variations in physical properties such as gas content, viscosity, vescularity, wall rock permeability, and crystallinity (e.g. Gonnermann and Manga, 2007; Jaupart, 1998; Melnik and Sparks, 2002). To date experimental and theoretical studies have focussed on variation of these properties as magma rises and decompresses (e.g. Baker et al., 2006; Gardner, 2007; Hammer and Rutherford, 2002; Larsen et al., 2004; Proussevitch et al., 1993; Takeuchi et al., 2009; Yoshimura and Nakamura, 2008) or is sheared (Gonnerman and Manga, 2003; Lavallee et al., 2007, 2008; Okamura et al., 2010; Smith et al., 2009; Tuffen et al., 2003, 2008). Natural pumice, dome rocks and experimentally decompressed glasses show huge textural variations in their permeable vesicle and crack networks (Jaukait, 1998; Michaut and Sparks, 2009; Mueller et al., 2008; Rust and Cashman, 2004; Saar and Manga, 1999; Takeuchi et al., 2008; Westrich and Eichelberger, 1994; Wright et al., 2009; Yoshimura and Nakamura, in press). Other experimental studies investigate the effects of temperature on water speciation, solubility, and magma viscosity (Stolper, 1989; Yamashita, 1994; Zhang et al., 2007). Rocks from conduit walls also exhibit variation in porosity (Kennedy et al., 2005; Rust et al., 2004; Stasiuk et
al., 1996). Yet, no studies have addressed the isobaric time-dependent textural changes of conduit walls in response to reheating. Here we address two key questions: 1. How is degassing of the ascending magma affected by changes in permeability of the conduit walls during reheating? 2. To what extent can the release of volatiles from the reheated wall rock contribute towards the eruption?

The vents for many explosive eruptions are often plugged by fractured, vesicular lava domes or partially filled conduits (e.g. Johnson and Lees, 2000; Voight et al., 1999). Surprisingly, almost no attention has been given to the effects of the hot rising magma on the behaviour (i.e. evolution of texture and volatile content) of the lava that plugs the volcanic conduit. This is despite observations that show temperature rises in older lava domes prior to eruption (Wooster and Kaneko, 1997) that correlate with eruption style (Sahtetapy-Engel and Harris, 2009). We propose that reheating can influence the degassing of both the older plug and the rising magma. Magmatic water and resorbed meteoric water dissolved in glass within the old plug may be available for degassing and vesiculation. This vesiculation in turn affects the porosity and the permeability of the plug and the ability of the rising hot magma to degas.

The 1886 Tarawera eruption (Cole, 1970; Nairn, 2002), is one of only a few examples of basaltic plinian eruptions (Houghton et al., 2004). At Tarawera, basalt erupts through a pre-existing dome complex and silicic conduit system (Carey et al., 2007). This may be a common occurrence at bimodal vents, however, descriptions of bimodal vent exposures are absent in volcanological literature. Detailed stratigraphic studies at Tarawera have tracked the shifting eruption centres and fragmentation level and documented interaction with groundwater and the pre-existing hydrothermal system (Carey et al., 2007; Houghton et al., 2004; Sable et al., 2006, 2009). The effect of rhyolitic conduit wall recycling during this eruption has also been discussed (Rosseel et al., 2006).

Conduit wall permeability is an important variable in explosive basaltic eruptions (Houghton and Gonnerman, 2008) but has not been investigated experimentally. We use laboratory experiments on the Tarawera rhyolitic lava to show that during an eruption the wall rock permeability is both time- and temperature-dependent, as is the release of volatiles from the wall rocks into the erupting magma. We argue that a reduction in the permeability of the conduit walls as a result of reheating hindered outgassing and increased the explosivity of the 1886 Tarawera eruption.

2. Methodology

2.1. Sampling

We collected samples that contained rhyolite and basalt from the proximal deposits of 1886 basaltic fissure eruption. The motivation for this sampling was to collect samples that show evidence for heat transfer between basalt and rhyolite. Enclave samples were collected from along the rim of the fissure on the summit of Tarawera lava dome (Fig. 1). We limited our samples to the crystal rich 1314 AD lava dome xenoliths/enclaves and excluded the older crystal-poor xenoliths (Carey et al., 2007). We chose a single large homogeneous glassy and perlitic sample of the lava dome to use for our experimental starting material. From this sample we drilled cylindrical cores 1 cm in diameter by 2 cm in length, which were then left in an oven overnight at 100 °C to remove surface water from the samples. The precise dimensions, density and porosity of samples were measured prior to, and following, heating experiments. The volume of these porous cylindrical cores was calculated from averages of replicate (n = 3) measurements of diameter and length. This volume and the sample mass were used to calculate the bulk density ($\rho_{\text{bulk}}$) of the core. Skeletal (or framework) density ($\rho_{\text{skeletal}}$) is obtained by measuring sample volume via helium pycnometry. Connected porosity ($\phi_{\text{connected}}$) (Table 1) was calculated from skeletal and bulk density from the relationship: $\phi_{\text{connected}} = 1 - (\rho_{\text{bulk}}/\rho_{\text{skeletal}})$. We obtained values of dense rock equivalent (DRE) density for rock by crushing three cores and performing pycnometry on the resulting powders. All experimental cores have the same average DRE density (2.34 g/cm$^3$ ± 0.10 g/cm$^3$). Using this average value for powder density we compute total and isolated porosity (Table 1) as: $\phi_{\text{total}} = 1 - (\rho_{\text{bulk}}/\rho_{\text{powder}})$ and $\phi_{\text{isolation}} = (\rho_{\text{bulk}}/\rho_{\text{skeletal}}) - (\rho_{\text{bulk}}/\rho_{\text{powder}})$ (Michol et al., 2008). Porosity values have an uncertainty of up to 4% associated with the largest porosities measured by pycnometer (Michol et al., 2008).

2.2. Experiments

We first heated the furnace to 300–1200 °C (± 10 °C) at atmospheric pressure. This represents the expected temperature range of conduit rocks in contact with the erupting basalt (Rosseel et al., 2006). Cylindrical samples (described subsequently) in ceramic crucibles were placed in the centre of the furnace for a specified time

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2.3. S.E.M.

Natural samples and samples from each experiment were coated in carbon and examined using the Philips XL30 electron microscope (SEM) (Beam power 15 kV and setting “Spot 6”) at the Earth and Ocean Sciences Department of UBC. We compare the pre- and post-experimental textures of the samples. When imaging both natural and cored samples we avoided surfaces affected by the rock saw and corer.

2.4. Water content

We ground and sieved samples, including crystals, to 100 μm for bulk water analysis from experiments lasting 30 min, 2, 4, 17.5 and 36 h. These samples were analyzed at ALS Actlabs using a Leco induction furnace combined with spectral analysis of the emitted volatiles. The sample (~0.3 g) is thermally decomposed in a resistance furnace (ELTRA CW-800) in a pure nitrogen environment at 1000 °C, causing release of volatiles, including both H2O− and H2O+.

Some natural and post-experimental samples were thinned and polished on both sides to make wafers suitable for analysis of volatiles with the FTIR. FTIR analysis was undertaken at the University of Oregon following the technique described in Wright et al. (2007). During analysis, we made every attempt to avoid areas containing crystals and vesicles. Total H2O contents and the speciation of H2O in the glass were determined using the absorbances of combination bands at 5230 cm−1 (molecular H2O) and 4520 cm−1 (OH−). At low total H2O contents (~0.5 wt.%; Stolper, 1982), molecular H2O is not detectable, in which case total water content was measured using the 3570 cm−1 band, representing the fundamental OH− stretching vibration. Absorbances were converted to concentrations using the Beer–Lambert law and absorption coefficients at 5230 cm−1 and 4520 cm−1 from Zhang et al. (1997) and at 3570 cm−1 from Stolper (1982). A glass density of 2.34 g/cm3 was obtained from pyrometry. Sample thickness was determined using a micrometer and varied between 300 and 900 μm. Results from the 3570 cm−1 band and totals from the 5230 cm−1 and 4520 cm−1 bands (Table 1) agree to

Table 1

Results of high temperature experiments performed on natural samples summarized as experimental conditions, and properties of starting materials and run products. The number in brackets in the FTIR column represents the number of measurements the mean value is based upon.

| Sample name | Time (t) | Temp. (T) | Relax. time | Log visc. | Connected porosity | Isolated porosity | Total porosity | Leco induction H2O (%) | FTIR (5230 + 5230 peaks) total H2O (wt.%) | FTIR (3570 peak) total H2O (wt.%) | FTIR (4520 peak) OH− (wt.%) | FTIR (5230 peak) mol. H2O (wt.%) | FTIR image mol. H2O (wt.%) |
|-------------|----------|-----------|-------------|----------|-------------------|-------------------|---------------|------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|---------------------------------|
| Original h  | 20       | 100       |             | 0.28     | 0.00              | 0.28              | 1.62          | 0.81 (4)               | 0.78 (2)                        | 0.09 (4)                        | 0.73 (4)                        | 0.16–1.30                      |
| Original j  | 20       | 100       |             | 0.21     | 0.08              | 0.29              | 1.63          | 0.81 (4)               | 0.78 (2)                        | 0.09 (4)                        | 0.73 (4)                        | 0.16–1.30                      |
| BKM06T800  | 2        | 800       | 30          | 0.28     | 0.04              | 0.29              | 1.64          | 0.82 (4)               | 0.78 (2)                        | 0.09 (4)                        | 0.73 (4)                        | 0.16–1.30                      |
| BKM06T900  | 0.5      | 900       | 1           | 0.31     | 0.03              | 0.35              | 1.66          | 0.83 (4)               | 0.78 (2)                        | 0.09 (4)                        | 0.73 (4)                        | 0.16–1.30                      |
| BKM06T1000 | 0.5      | 1000      | 1           | 0.33     | 0.04              | 0.35              | 1.67          | 0.84 (4)               | 0.78 (2)                        | 0.09 (4)                        | 0.73 (4)                        | 0.16–1.30                      |
| BKM06T1100 | 0.5      | 1100      | 1           | 0.36     | 0.04              | 0.39              | 1.67          | 0.84 (4)               | 0.78 (2)                        | 0.09 (4)                        | 0.73 (4)                        | 0.16–1.30                      |
| BKM06T1200 | 0.5      | 1200      | 1           | 0.40     | 0.04              | 0.40              | 1.67          | 0.84 (4)               | 0.78 (2)                        | 0.09 (4)                        | 0.73 (4)                        | 0.16–1.30                      |

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within 0.004 wt.%. Multiple spots were analyzed where possible and the number of spots per measurement is shown in Table 1 using number in brackets after the mean measurement.

Additionally, we conducted FTIR spectroscopic imaging of a fragment of the original dome material to investigate molecular H₂O distribution. This fragment broke naturally along perlitic cracks during further thinning of the samples, which was carried out to reduce the number of crystals and vesicles in the wafers. Images (each 350 × 350 μm) of the fragment were collected using a Varian Inc. Lancer Focal Plane Array (FPA) camera attached to a Varian PFS Stingray 700 Micro Image Analyser spectrometer and UMA 600 microscope at the Institute for Research on Earth Evolution (IFREE), Japan Agency for Marine Earth Science and Technology (JAMSTEC). For more detailed discussion of FTIR spectroscopic imaging see Wysoczanski and Tani (2006). The thickness of the fragment was determined using reflective light spectra and the wavelength of the resulting interference fringes following the method of Wysoczanski and Tani (2006) and Nichols and Wysoczanski (2007). A refractive index of 1.50 was used for rhyolite (Liu et al., 2005; Long and Friedman, 1968). From 89 reflection spectra across the image, the average thickness was 62 μm (1σ = 1 μm), and this value was used for all calculations. The glass density used was the same as for the spot analyses. Variations in glass density as a result of variations in H₂O concentration result in a maximum error of ±0.01 wt.% on the molecular H₂O contents. Owing to the thinness of the sample, the combination bands at 5230 and 4520 cm⁻¹ used to measure molecular H₂O and OH⁻ were below detection (see Supplementary Figure A1). As a result, the spectroscopic image for molecular H₂O represents the absorbance of the fundamental bending of molecular H₂O at ~1630 cm⁻¹. See Supplementary Figure A1 for additional discussion of the use of this band. Total H₂O contents were obtained from the peak at 3570 cm⁻¹ as described earlier.

3. Description of natural samples

Our 30 natural samples vary in relative proportions of rhyolite and basalt (Fig. 2). In contrast to Rosseel et al., 2006, we have chosen to classify samples on the basis of vesicularity rather than bomb type. The samples described subsequently are all from the AD1314 lavas or dykes; wall rocks from older Tarawera lavas (Carey et al., 2007) were not sampled. These young lithics can be split into four vesicularity types. Type 1 are glassy samples of the Tarawera lava dome that we use in the experiments (Fig. 2a). Types 2–4 are samples from the 1886 plinian basaltic fall deposit and contain both rhyolite from the AD 1314 Tarawera lava dome and basalt from the 1886 eruption. The samples vary from rhyolite “lithics” coated in basaltic scoria to completely remelted rhyolite enclaves entirely contained within basaltic spatter (Rosseel et al., 2006) (Fig. 2b–d).

Type 1 samples are representative of the Tarawera lava dome; these samples were collected from outcrops within the 1886 crater within the Tarawera lava dome (Fig. 1). Samples are variably devitrified and texturally perlitic and show a range of vesicularities. Vesicles show a range of vesicle sizes and shapes (Fig. 2a). We chose a typical large glassy block without obvious devitrification but with perlitic cracks (Fig. 2a) for our experiments. All the drilled cores are from the one sample and show a total porosity variation of 28–30 vol.%. The dome is 15–35% crystalline (Cole, 1970), and contains phenocrysts of plagioclase, quartz, biotite and minor amphibole, orthopyroxene and Fe–Ti oxides. SEM images show curvilinear perlitic microcracks at 50 μm intervals in the glass (Fig. 2a(ii)) and unknown secondary minerals on glass surfaces. Water content of the starting material measured by FTIR is 0.8 wt.% and by Leco induction is 1.6 wt.% (see Fig. 5b).

Type 2 samples were collected from the basaltic fall deposits draping the Tarawera lava dome, exposed on the margin of the crater. Samples are 1–10 cm chunks of vesicular rhyolite (Fig. 2c), partially coated in (~10% by volume) basaltic scoria (~1 cm clast diameter).

The basaltic scoria is weakly attached to the surface of the rhyolite. Where basaltic scoria is absent, cracks are visible on the surfaces of the rhyolite. The rhyolite has total porosity estimated to range between 40 and 60% (using proportion comparison charts), only one sample was measured at 53% (Table 1). SEM images show almost no microcracks and complex vesicle shapes with evidence of vesicle wall retraction and vesicle collapse (Fig. 2c (ii)).

Type 3 samples were collected from the same basaltic fall deposits as Type 2 and the talus slopes below this fall deposit. Pyroclasts contain 10–90 vol.% basalt. Samples are up to 30 cm in diameter, and visibly very vesicular (>60%), one sample was measured at 73% total porosity with individual vesicles up to 5 mm. The surfaces of these samples are covered in basaltic scoria and spatter which is strongly attached to the surface of the rhyolite (Fig. 2c). Adjacent to the basaltic spatter, the outer few mm of the rhyolite is vesicle poor and appears locally as black obsidian (Fig. 2c). SEM images show no microcracks and in contrast to the Type 1 dome rock, the vesicles are spherical (Fig. 2c (ii)).

Type 4 samples were collected from spatter-rich areas in the basaltic fall. These pyroclasts are >90% basalt (Fig. 2d). These samples contain small (<5 cm) dense enclaves of rhyolite within basaltic spatter. The rhyolite blebs are generally too small to measure porosities, however, we estimate the total porosity is consistently <25%. SEM images show a range of textures, with prominent small spherical and irregular vesicles (Fig. 2c (ii)).

4. Experimental results

4.1. Qualitative results: SEM

We present the results of the experimental heating of 32 Type 1 lava dome cores with the aim of understanding the potential consequences of reheating conduit filling material. SEM images showing time- and temperature-dependent changes in the microstructure of the samples are shown in Figure 3. Generally, the textures progress from cracked vesicular glassy original (Type 1) textures, to an increase in vesicles with cracked walls, to a connected network of un-cracked vesicles, and finally to less vesicular isolated vesicles.

Experiments lasting 2 h at 300–700 °C show no noticeable difference in texture from the original samples. However, the time series from experiments at 800–1200 °C show remarkable differences in the geometry and size of cracks and vesicles relative to the initial state.

At 800 °C, samples heated for ~6 h are characterized by open cracks. In addition, the widest cracks are spatially associated with dome structures (related to vesicle growth) (Mungall et al., 1996) on the surface of the glass which are 100 μm in diameter (Fig. 3a). In samples heated for 24 h cracks are still apparent and some cracks appear more curved and crack edges appear slightly rounded (Fig. 3b).

At 900 °C, samples heated for ~6 h (Fig. 3c) are texturally similar to the 800 °C runs (Fig. 3a). However, at 24 h, cracks are commonly curved. The dome structures appear to have surfaces made of a thin film of glass and relict cracks and rafts of original surfaces exist on some domal structures (vesicles) (Fig. 3d), suggesting that films of melt have allowed vesicles to grow.

In experiments conducted at 1000 °C, the edges of cracks are distinctly rounded indicating (re-)melting (Fig. 3e). In the sample heated for 24 h at 1000 °C most of the glass is smooth with angular cracks only occurring in crystals. In addition, completely smooth domes occur with only a vague hint of relict cracks (Fig. 3f). The domes appear as connected vesicles, however these vesicles are not bulbous, and some appear deflated (Fig. 3f).

At 1100 °C, a sample heated for 40 min shows similar textures to the sample heated for 24 h at 1000 °C (Fig. 3f and g). The surface of the sample is smooth with bulbous vesicle-like domes, and no relict cracks (Fig. 3g). The surface of the sample heated for 6 h at 1100 °C is completely smooth and cracks are no longer visible in the glass. Most
of the glass is bulbous and appears to be connected vesicles (Fig. 3g). The glass in the sample heated to 1100 °C for 24 h also is smooth and has no cracks. Vesicles are not as apparent on the surface of the sample (Fig. 3h).

Finally, the sample heated for 30 min for at 1200 °C (Fig. 3i) shows similar textures to the sample heated for 6 h at 1100 °C (Fig. 3g). Glass is smooth and unfractured and contains connected vesicles (Fig. 3i). The sample heated at 1200 °C for 24 h has smooth surfaces similar to that heated to 1100 °C (Fig. 3h), however fresh surfaces show some isolated vesicles exist beneath the smooth surface (Fig. 3j). All original textures were obliterated.

In summary, experiments at lower temperatures and/or for short times generally produced samples with open cracks. Experiments at higher temperatures and/or for longer times show evidence of melting, crack annealing, the growth of films of melt, and the inflation and deflation of vesicles.

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Fig. 3. SEM images of experimental run products produced by heating of natural samples over fixed periods of time. a) Experimental run product for 800 °C over 6 h is largely texturally unchanged from the original. However, cracks have begun to open especially in areas around domed glass associated with subsurface vesicles. b) Run product for 800 °C over 24 h shows similar cracking associated with sub-surface vesicles; also shows convex areas and some cracks appear to be rounded. c) Run product for 900 °C over 6 h shows open cracks associated with sub-surface vesicles. d) Heating at 900 °C for 24 h causes formation of new vesicles that feature films of melt that heal old fractures and create rafts of the old vesicle wall on the surface of the new vesicle. e) Sample heated at 1000 °C for 6 h shows angular fracture edges rounding and annealing. f) Sample heated at 1000 °C for 24 h shows almost no vestiges of the earlier cracked surfaces, however, existing dome-like structures appear deflated. g) Samples heated at 1100 °C for 6 h feature connected inflated vesicles within the original vesicle wall septa, and no evidence of the original cracked glass. h) Heating at 1100 °C for 24 h produces a sample on which the surface shows no evidence of the original bubbly and cracked surface. i) Run product for 1200 °C over 30 min is similar to g), vesicle walls contain inflated connected vesicles. j) Heating at 1200 °C for 24 h causes the sample to collapse and flow; material had to be chipped out of the crucible and internally it shows spherical isolated vesicles.
4.2. Porosity

Time series of fractional porosity evolution during the experiments are given in Table 1 and illustrated in Figure 4. Both total and connected porosity initially increase with time at all experimental temperatures. The time taken to reach a maximum in total porosity decreases as the experimental temperature increases. This increase in porosity is then followed by a reduction in porosity, where the specific total vs. connected porosity path is dependent upon temperature. Isolated porosity is illustrated in Figure 4c by the vertical distance between any point and the dashed, fully connected porosity line. At ~700 °C, overall porosity and volume were indistinguishable within error from the unheated samples (Tables 1 and Supplementary Table A2).

At 800 °C the connected porosity reaches a maximum of 0.39 in the experiment conducted for 6 h (Fig. 4a). Experimental times greater than 6 h correspond with lower connected and total porosities. Isolated porosity remains less than 0.05 in all experiments at 800 °C.

The time series at 900 °C also shows an initial increase in connected and overall porosity, connected porosity then fluctuates and reaches a maximum at 17.5 h of 0.37. Thereafter the connected porosity drops monotonically to 0.33 at 36 h (Fig. 4b). Again, the isolated porosity remains similar during this time series (Table 1).

The experiments at 1000 °C have a maximum total and connected porosity of 0.46 at 2 h. Longer experiments correspond with smaller porosities (Fig. 4c). This time series shows higher isolated porosities (up to 0.11) compared to 800 and 900 °C, particularly the experiments for 2, 6 and 24 h.

At 1100 °C a maximum porosity of 0.43 occurs at 30 min. Thereafter, the connected porosity drops sharply to 0.23 by 24 h. Despite the consistent decrease in connected and overall porosity, isolated porosity varies considerably and shows a substantial peak (~0.1) at 6 h (Fig. 4d).

Samples heated at 1200 °C collapsed and lost their cylindrical form (textural collapse), the sample heated for 30 min remained generally cylindrical whilst the sample heated for 24 h collapsed completely.

In summary, at all experimental temperatures an initial increase in total porosity is followed by a decrease.

4.3. Water content

The water content of the glass in the original sample (starting material) varies between 1.6 and 0.8 wt.% as measured by Leco induction furnace and by FTIR spot analyses (N = 4), respectively (Fig. 5b, Table 1). These differences in measured water contents between Leco Induction furnace and FTIR are absent at higher temperatures and lower water contents. The FTIR spectroscopic image of the absorbance for the molecular H2O band (Fig. 5a) supports this, with absorbance increasing, by up to a factor of 7 compared to the interior of the fragment, at the perlitic margins. Notwithstanding the qualifying statements of Zhang et al. (1997) (described in the figure caption of the Supplementary Fig. A1) this represents an increase in molecular H2O content from approximately 0.16 to 1.30 wt.%. All time series show that >75% of bulk water measured by Leco induction was lost within 0.5 h. Over 2 h at 500 °C the bulk water content shows a systematic decrease as temperature increases from 1.6 wt.% to around 0.22 wt.%, which is close to the detection limit of this instrument (Fig. 5b). Within the uncertainty of the measurements, repeated experiments at temperatures above 500 °C showed no change in bulk water content. FTIR spot measurements using the 3550 cm−1 band support the same general pattern of rapid water loss and show a mean initial water content of 0.78 wt.% was reduced to 0.06% after 6 h at 800 °C (Fig. 5b). Additionally, these FTIR spot analysis support continued degassing to 0.01 wt.% after 6 h at 1100 °C.

4.4. Results summary

Our observations and measurements of the rhyolite lava prior to experimentation illustrate that the lava has an intricate network of perlitic cracks (Fig. 2a) and connected vesicles (connected porosity 0.26 and isolated porosity 0.04). In addition, a significant portion of the water contents of the original dome glass are not accounted for by the water species measured by either spot or FTIR mapping. Water contents that are effectively mapped by FTIR are shown to be strongly enriched along the margins of the perlitic cracks (Fig. 5a). Our experimental results show that our samples, where heated above 500 °C lose most of their water within 2 h. However, changes in total porosity do not occur until 800 °C. All time series above 800 °C showed a small but continued degassing and an initial increase in connected porosity associated with cracking and vesicle growth, followed by a decrease in porosity associated with vesicle collapse. Generally, the hotter the experiment the earlier in the time series the porosity decrease occurred. Textural changes are summarized in Fig. 6. experiments are subdivided into samples with (1) no textural changes, (2) cracking/ inflation, (3) vesicle collapse/ deflation, and (4) textural obliteration. This subdivision is used to develop degassing and deformation regimes which are discussed below.

5. Discussion

5.1. Comparison between natural and experimental samples

By comparing the textures of the naturally heated samples and the experimentally heated samples we constrain the porosity and, by inference, the permeability history of the dome and conduit wall rocks in response to reheating. In general, there is a striking similarity between the textures of type 2–4 samples and the higher temperature and long timescale experiments we performed (Figs. 2 and 3). The textural similarities support our estimates of the timescales and temperatures of reheating that were derived from the eruption timescale (Keam, 1988). However, we do not imply that the thermally driven textural changes in the natural samples occurred in-situ at the conduit margin; most of these erupted as bombs and probably followed the thermal history described in Rosseel et al. (2006). The open cracks and cracked vesicle walls of lower temperature and shorter timescale experiments were not seen in our suite of naturally heated samples, although this may be due to sampling bias.

The type 2 samples (Fig. 2b) have textures and porosity similar to experiments for 24 h at 1000 °C (Fig. 3f). In both these rocks, glass surfaces are smooth, cracks have healed but vesicles remain small and only partially inflated.

The type 3 samples (Fig. 2c), appear similar to experiments above 1100 °C (Fig. 3g and h) and show no evidence of the original textures of the type 1 rock (Fig. 2a). However, the experiments do not exhibit the large vesicle sizes and high porosities of the natural type 3 samples (Fig. 2d). In our experiments above 1100 °C, at timescales longer than 40 min, vesicles coalesced, connected with the ambient pressure and collapsed. The large size of the natural samples allowed more vesicle growth and coalescence, and the developments of a larger isolated porosity before depressurizing to ambient pressure. Such an internal pressure could prevent collapse of the sample and loss of porosity (Yoshimura and Nakamura, 2008). Alternatively, vesiculation of these samples could have been aided by decompression as they erupted.

The type 4 samples are similar to experiments at 1200 °C for timescales greater than 30 min, no original vesicle textures resembling the type 1 sample remain. The amoeboid outer shapes and dense textures of the type 4 samples imply that they fully melted. Vesicles are generally small, implying that larger vesicles coalesced, collapsed or escaped by migration through the melt.
The similarity of textures shared by the experimental and natural rocks implies that reheating drove vesiculation and vesicle collapse in the natural samples. Natural type 2 samples found in the basaltic scoria reached temperatures of 1000 °C for at least 24 h, whereas type 3 and 4 samples found in spatter imply higher temperatures and/or longer hot residence times (Carey et al., 2008; Rosseel et al., 2006; Sable et al., 2009). In summary, the comparison of Figures 2 and 3 show that the temperatures and timescales of reheating during our experiments were appropriate for the eruption, and that large variations in conduit wall porosity occurred during the eruption. An ongoing study of the scoria and spatter filled basaltic dyke margins exposed in the base of the fissure support this porosity variation. Additionally, this fieldwork constrains the thermal impact of the basalt to be between a few millimetres to tens of centimetres, in some areas we identified a distinct 5 cm thick welded portion; and these detailed field descriptions will be the subject of a future publication.

5.2. Degassing, and deformation regimes

Degassing and deformation proceeded differently in our experiments relative to other experimental studies involving hydration of melts (Baker et al., 2006; Gardner, 2007; Larsen et al., 2004; Takeuchi et al., 2009; Yoshimura and Nakamura, 2008). Our samples were composed predominantly of glass naturally hydrated by water along perlitic cracks at temperatures well below $T_g$ (Denton et al., 2009). As a result this water can be outgassed at temperatures well below $T_g$ (Tuffen et al., 2010). Isotopic studies indicate that this water is likely to be meteoric in origin (DeGroat-Nelson et al., 2001; Friedman and Smith, 1958; Friedman et al., 1966; Shane and Ingram, 2002). An additional contrast to previous experiments is that our experimental sample already had a high proportion of connected porosity through cracks and vesicles. This distinction is important because the relationship between permeability and porosity is different during vesicle growth and vesicle collapse (Michaut and Sparks, 2009; Rust and Cashman, 2004). A sample that has undergone vesicle collapse will have a higher ratio of permeability to porosity than a sample of similar porosity that has not undergone vesicle collapse (Michaut and Sparks, 2009; Mueller et al., 2008). The complexities of this relationship are most apparent when illustrated by up to 6-fold increase in permeability over the porosity range of 30–40% (Wright et al., 2009). The permeability/porosity relationship is further complicated by the presence of cracks with the ability to open and to heal (Yoshimura and Nakamura, in press). However, all our experiments had the same starting texture and we are confident that incremental changes in open porosity correlate with changes in permeability.

Fig. 4. Experimental results summarized as total porosity vs. connected porosity and labelled by experimental times. Open headed arrows follow a temporal evolution of increasing experimental time. A dotted 1:1 line indicating equality between connected and total porosity is shown on all plots; the vertical distance between each data point and this line measures the isolated porosity within each sample. a) Time series at 800 °C; inset of solid headed arrows shows how porosity is affected by various relevant processes. b) Time series at 900 °C. c) Time series at 1000 °C. d) Time series at 1100 °C.
On reheating, our experiments imply that the porosity and permeability structure of lava domes or plugs will pass through a series of regimes, the nature of which depend on 1) the temperature of the event; 2) the time spent at that temperature (Fig. 6a); and 3) the volatile content and original vesicularity of the rock being heated. Physically, the character of the change in porosity and, by inference the permeability, depends on whether dome rocks respond in a viscous or brittle way to heat transfer from the newly erupting magma. Thus, the key parameter is the glass transition, which depends on the temperature relative to a critical relaxation time for structural failure of the melt. Consistent with experimental results, we take this critical time to be proportional to the viscous relaxation time \( \tau_r = \mu/G \) (Webb and Dingwell, 1990) (Table 1). Here, \( G = 10^{10} \) Pa is the elastic modulus, \( \mu \) is the melt viscosity at a given temperature and water content and \( C \) is a constant to be determined from our experiments. We take \( T_g \) to be the temperature corresponding to a viscosity of \( 10^{1.4} \) Pa s which correlates well with the calorimetric \( T_g \) (Giordano et al., 2008). The value of \( T_g \) is calculated to be 740 °C for this rhyolite glass composition (Nairn et al., 2004) at 0.1 wt.% H\(_2\)O using the viscosity model of Giordano et al. (2008) (see Supplementary Table A2). The transition from brittle to viscous behaviour (Regimes 1 to 3 in Fig. 6) depends on the temperature (\( T \)) relative to \( T_g \) and the time scale for the experiment (\( t \)) or eruptive event.

To gain additional insight we replot the data in Figure 6a in terms of \( T \) normalized to \( T_g \) and \( t \) normalized to \( \tau_r \) (Fig. 6b). The data collapse to a power law defining the transition from brittle to viscous behaviour of the form \( T/T_g = C(t/\tau_r)^{0.046} \), where \( C \approx 0.86 \). This result shows that whether or not brittle processes govern the final porosity of the sample depends on the response time of the melt, which is governed by its strongly temperature-dependent viscosity, relative to the time scale of the thermal forcing applied in the experiment.

We use our experimental degassing and deformation data to propose four degassing and deformation regimes (Fig. 6). In Regime 1 (\( T \leq 800, t \leq 2 \) h; \( T/T_g > C(t/\tau_r)^{0.046} \)) degassing occurs without significant deformation (Fig. 6). The rock maintains its vesicular structure connected by microcracks and ruptured vesicle walls (Fig. 2a, b) and consequently maintains a high proportion of connected to total porosity (Fig. 4). No visible changes to this structure could be seen with the SEM. Between 0 and 700 °C we attribute this initial H\(_2\)O degassing to (1) release of water captured in micropores and/or low temperature alteration hydrous minerals (Denton et al., 2009), and (2) diffusion of resorbed meteoric water out of the perlitic margins (Fig. 5a) of the glass (Tuffen et al., 2010). 0.20–0.05 wt.% magmatic water remains dissolved in the glass interior (Fig. 5). This initial period of degassing had no impact on the textures of the rock as the rock was not sufficiently above its calculated glass transition temperature (Giordano et al., 2008) and ductile deformation did not occur.
Regime 2 \((800 \leq T \leq 1100 \, ^\circ C, \tau \leq 6 \, h, T/T_g < 0.1 \, wt.\% H_2O)\) is a transitional regime showing both brittle and viscous deformation (Fig. 6). Degassing of magmatic water \((<0.1 \, wt.\%)\) continues (Fig. 5) and vesicle and crack expansion cause 5–20% increases in overall porosity (Fig. 4). Observations show the porosity changes are due to local viscous vesicle growth and coalescence with concurrent brittle cracking in areas of high strain (Fig. 3a–c). Our experimental samples have initial porosities of 0.29–0.31, and connected porosity of 0.27; the growth of isolated vesicles frequently impinge on other vesicles (Fig. 3a–c), resulting in coalescence and increasing connected porosity with relatively small increases in isolated porosity (Fig. 4a). In addition, the opening of cracks (Fig. 3b, c) connects isolated vesicles, increasing the connected porosity and reducing isolated porosity (Fig. 4a).

In Regime 3 \((800 \leq T \leq 1200 \, ^\circ C, \tau > 30 \, min, T/T_g < 0.46)\) porosity loss occurs due to vesicle deflation (Fig. 6). This regime is characterized by crack sealing and vesicle collapse leading to a transient but generally decreasing overall porosity (Fig. 4). Both processes involve viscous relaxation and appear to occur at similar temperatures and timescales. Vesicle collapse is driven as the internal pressure of the vesicle can no longer balance the surface tension and gas leaks out of the vesicles through the connected porosity. Crack healing occurs as melt connects opposite sides of a crack (Fig. 3e,f). Healing of cracks reduces connected porosity and increases isolated porosity (Fig. 4). However, generally, in our experiments observations of crack healing correlate with decreases in overall porosity which can only be explained by vesicle collapse (Figs. 3 and 4). Occasional small increases in porosity are seen in this regime and we interpret these to be due to continued degassing and vesicle growth, as larger coalesced vesicles collapse.

In Regime 4 \((T \geq 1200 \, ^\circ C, \tau > 30 \, min, T/T_g > 0.46)\) textural collapse occurs due to viscous flow of the melt in response to gravity (Fig. 6). Although we could not measure the residual porosity of these run-products, SEM image analysis shows a significantly reduced porosity comprising isolated pores (Fig. 3k) implying a much reduced permeability.

The temperature/time space of each of these regimes is ultimately controlled by the viscosity of the melt/glass framework in the dome lava which is itself strongly influenced by the dissolved water content (inset Fig. 6a). The calculated dry value for \(T_g\) of this melt (see Supplementary Table A2) is 770 \(^\circ C\). The equilibrium volatile content of the glass will be controlled by pressure and therefore the “openness” and porosity of the volcanic system. An open, degassed system such as the system reproduced by our experiments has a high \(T_g \) of 740 \(^\circ C\) \((-0.1 \, wt.\% H_2O)\) and is relatively difficult to remelt and initiate Regimes 2, 3 and 4. However, a closed system with a higher ambient pressure and equilibrium volatile content will have a depressed \(T_g\). For example, this melt with 1 wt.\% H_2O has a calculated \(T_g \) of 606 \(^\circ C\) (see Supplementary Table A2, Giordano et al., 2008). Therefore, if such a system is reheated it may achieve Regimes 2, 3, and 4 at lower temperatures.

The style of hydration, the original texture, and the amount of time a reheated rock/magma spends in Regime 1 will strongly influence the relationship between water content and its calorimetric \(T_g\), and ability to continue to other regimes. Similarly the amount of time a rock spends in permeable Regimes 1 and 2 will influence the ability of any magma beneath it to degas. For these reasons heating rates and partially closed systems become important considerations and avenues for future research.

6. Conclusions and implications for Tarawera

The frozen magma forming the conduit walls will be reheated during the ascent and eruption of fresh magma. The time and temperature dependent textural changes in the plug or walls have implications for the monitoring of degassing and deformation of active volcanoes, and on the resultant style and magnitude of an eruption. In particular, a key issue is whether reheating leads to the production or destruction of permeability in the conduit walls. For example, enhanced permeability facilitates outgassing, which can reduce the overpressure in the intruding magma, and favour effusive volcanism (e.g., Quane et al., 2009). In contrast, reduced permeability can inhibit outgassing, leading to greater overpressure in the new magma and an increased likelihood for explosive volcanism.

From our experiments, the evolution of wall rock texture during an eruption depends on the temperature and duration of the event (Fig. 6). In Regime 1 \((T/T_g < Ct/\tau_r^{0.046})\), there is degassing of rehydrated meteoric water from the rhyolite. For Tarawera dome rock, up to 1.4 wt.\% water could be released by conduit wall and old dome rocks into the erupting magma during reheating. In Regime 2 \((T/T_g < Ct/\tau_r^{0.46})\) both ductile and brittle deformation cause increases in connected porosity and inflation. This increase in connected porosity should correlate with an increase in permeability and hence aid the degassing of the fresh magma below. Regime 3 \((T/T_g < Ct/\tau_r^{0.46})\) is a result of vesicle collapse and porosity...
reduction. Although a connected pore structure still remains, permeability will be reduced progressively closing the degassing system (Westrich and Eichelberger, 1994; Yoshimura and Nakamura, 2008; Yoshimura and Nakamura, in press). Regime 4 produces a generally isolated pore structure and permeability would be significantly decreased causing pressure build up, that could give rise to explosive eruptions.

The magnitude of thermally-induced permeability change is expected to depend on the length scale \( l = 2/\sqrt{\kappa} \), where the thermal diffusivity \( \kappa \) of potentially foamed rhyolite dome rocks is order 10^-7 m^2 s^-1 (Bagdassarov and Dingwell, 1994). For Tarawera, over the proposed 5–24 hour magma rise and eruption time (Houghton et al., 2004; Keam, 1988) \( l \) is likely to be approximately 8–25 cm (Fig. 6a). Indeed, on-going field studies of the margins of basalt dykes in the lava domes reveal a 5 cm thick remelted layer of rhyolite with reduced porosity. This 5 cm thick remelted layer is remarkably consistent with the eruption timescale and thermal diffusion length scale (Fig. 6a). However, detailed descriptions of these dyke margins are not yet complete and beyond the scope of this paper.

For the 1886 eruption, we propose a reheating trajectory of 5 h (Fig. 6a) which is consistent with (1) the eruption timescale (Houghton et al., 2004; Keam, 1988), (2) the presence of all three naturally heated sample types, and (3) the thicknesses of remelted dyke margins.

The volume of AD 1314 Tarawera rhyolite affected is difficult to estimate due to uncertainties in the subsurface geometry of these and older rhyolite lavas and feeder dykes, and their relationship to the basalt dyke (e.g. Fig. 4 in Carey et al., 2007). We take the basaltic fissure to be 8 km long (Fig. 1) in map view and 0.2–8 km deep (the lower bound assumes that the basalt interacts only with the shallow dome; the upper bound implies interaction with a conduit over the full depth to the 8 km rhyolitic magma source (Shane et al., 2008)). Over the 5 hour eruption duration the intruding magma will heat and degas the surrounding rhyolitic wall rocks to a distance of 8 cm (Fig. 6a). From these dimensions and assuming that the wall rocks contain 1.4 wt.% H2O, reheating would release around 6.4×10^6 kg of H2O. From our experiments, the majority of this volume would be released within the first 2 h of reheating and may significantly contribute to precursory signs of eruption, although it may not be significant in terms of the total volatile budget of the eruption.

The similarity between the textures in the rhyolite erupted from Tarawera volcano and the experiments defining Regime 4 imply that conduit-wall permeability may have been catastrophically reduced during the early stages of the Tarawera eruption. We propose that such a sealing of the system may have produced a closed system and aided the unusually explosive basaltic eruption of Tarawera in 1886. This offers an additional and complimentary explanation to the driving mechanisms suggested by previous authors e.g., phreato-magmatic interactions and bubble/microlite coupling (Carey et al., 2007; Sable et al., 2009).

Finally, our work also has implications for eruptions related to re-intrusion of similar temperature magma. A closed pressurized lava dome and conduit with 1–3 wt.% H2O (e.g. Burgisser et al., 2010) could potentially depress \( T_g \) (Fig. 6a inset) and hence allow melting associated with the re-intrusion of magmas of similar composition (Giordano et al., 2008), i.e. a 700 °C rhyolite could re-intrude and melt a water rich rhyolite with a \( T_g \) depressed to 600 °C. Our work predicts that as an old lava dome or plug heats up it will initially inflate, partially degas, then deflate, and seal up leading to continued pressurization and possibly eruption (Fig. 6). This sequence may be followed by another reheating event and lead to the cyclic deformation observed at many lava domes (Johnson et al., 2008; Matthews et al., 1997; Voight et al., 1999).

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