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Notes

Melt fracturing and healing: A mechanism for degassing and origin of silicic obsidian

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ABSTRACT

We present water content transects across a healed fault in pyroclastic obsidian from Lami pumice cone, Lipari, Italy, using synchrotron Fourier transform infrared spectroscopy. Results indicate that rhyolite melt degassed through the fault surface. Transects define a trough of low water content coincident with the fault trace, surrounded on either side by high-water-content plateaus. Plateaus indicate that obsidian on either side of the fault equilibrated at different pressure-temperature (P - T) conditions before being juxtaposed. The curves into the troughs indicate disequilibrium and water loss through diffusion. If we assume constant T , melt equilibrated at pressures differing by 0.74 MPa before juxtaposition, and the fault acted as a low- P permeable path for H_2O that diffused from the glass within time scales of 10 and 30 min. Assuming constant P instead, melt on either side could have equilibrated at temperatures differing by as much as 100 °C, before being brought together. Water content on the fault trace is particularly sensitive to post-healing diffusion. Its preserved value indicates either higher temperature or lower pressure than the surroundings, indicative of shear heating and dynamic decompression. Our results reveal that water contents of obsidian on either side of the faults equilibrated under different P - T conditions and were out of equilibrium with each other when they were juxtaposed due to faulting immediately before the system was quenched. Degassing due to faulting could be linked to cyclical seismic activity and general degassing during silicic volcanic activity, and could be an efficient mechanism of producing low-water-content obsidian.

INTRODUCTION

Obsidian is typically a vesicle-poor and crystal-poor glassy silicic volcanic rock (e.g., Eichelberger and Westrich, 1981). Pyroclastic obsidian generally has relatively high water contents (~0.5 and ~1.5 wt%; e.g., Castro and Dingwell, 2009; Eichelberger and Westrich, 1981; Rust and Cashman, 2007) when compared to similar lavas (~0.3 and 0.1 wt%; e.g., Eichelberger et al., 1986; Friedman, 1989). Generation of relatively degassed but poorly vesicular obsidian is poorly understood and is ascribed to gas loss either through vesicles (e.g., Eichelberger et al., 1986; Okumura et al., 2009) or along fractures in magma (Gonnermann and Manga, 2003, 2005, 2007; Rust and Cashman, 2007; Rust et al., 2004; Stasiuk et al., 1996). Details of this mechanism remain poorly understood, but volatile concentration measurements support fracture degassing (Rust and Cashman, 2007; Tuffen et al., 2010).

Small-scale faults and welded breccias have recently been described in obsidian conduits, lavas, and pyroclastic blocks (Gonnermann and Manga, 2003, 2005; Goto, 1999; Rust and Cashman, 2007; Rust et al., 2004; Tuffen et al., 2003; Tuffen and Dingwell, 2005). Evidence suggests a process of repeated fracturing and healing of

melt (Tuffen et al., 2003; Tuffen and Dingwell, 2005) that may be characteristic of rhyolitic eruptions, as supported by numerical models (Gonnermann and Manga, 2003).

Here we investigate rhyolite melt degassing through fractures. We carried out synchrotron Fourier transform infrared spectroscopy (SFTIR) transects at a high angle to a rehealed fault in a juvenile obsidian pyroclastic bomb from the Lami pumice cone on Lipari, Italy. Water content profiles suggest degassing through fractures with significant implications for obsidian formation and conduit degassing.

SAMPLE DESCRIPTION AND METHODS

We report results from a healed fault in the glassy rind of one bread-crust pyroclastic obsidian sample (LO1) from the Lami cone (Fig. 1), on the flank of the Monte Pilato pumice cone. These are part of the rhyolitic, explosive, and effusive Monte Pilato–Rocche Rosse sequence (e.g., Cortese et al., 1986; Dellino and Volpe, 1995). The wall of the Lami cone reveals ~1-m-thick obsidian-rich layers interlayered with white tube pumice-rich layers (Fig. 1C) interpreted as fallout coeval with the eruption of the Monte Pilato pumice cone (Cortese et al., 1986). Clasts in obsidian-rich layers have chemistries similar to those of pumice counterparts and later obsidian lava (Table DR1 in the GSA Data Repository¹). Obsidian-rich layers

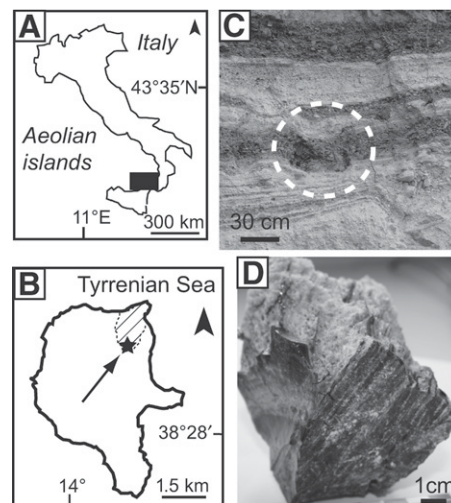


Figure 1. A: Location of Aeolian Islands and Lipari, southern Italy. B: Sketch of Lipari Island highlighting location of Monte Pilato–Rocche Rosse sequence (line pattern) and Lami cone (star) in northeast; black arrow points to Lami cone. C: Overview of exposure of Lami deposits showing dark layers of obsidian interlayered with white pumice layers. Circle highlights sag structure and location of sample LO1. D: Photograph of fragment of sample LO1 with black, outer bread-crust obsidian rind at base and white pumiceous interior at top.

are exclusively formed by bread-crust clasts, lapilli to bomb size, and lapilli-size obsidian clasts similar to Pele's tears. Bread-crust obsidian bombs have >1-cm-thick black glassy rinds and foamed interiors that indicate post-fragmentation expansion, and form sag structures in underlying layers, suggesting ballistic emplacement (Figs. 1C and 1D).

The glassy rind of sample LO1 preserves a healed fault plane truncating folded microlite-rich bands and pristine glassy bands (Fig. 2); fault here is used to describe a fracture plane that accommodated movement. Microlites are thin, >20- μ m-long needles composing at most ~1 vol%. Some fold hinges have relatively undeformed vesicles. The healed fault is >3 cm long in a block ~10 cm across (Fig. 1D); the block lacks evidence of brecciation. SFTIR transects

¹GSA Data Repository item 2011037, Table DR1 (chemical composition for glass analyzed by electron microprobe), is available online at www.geosociety.org/pubs/ft2011.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

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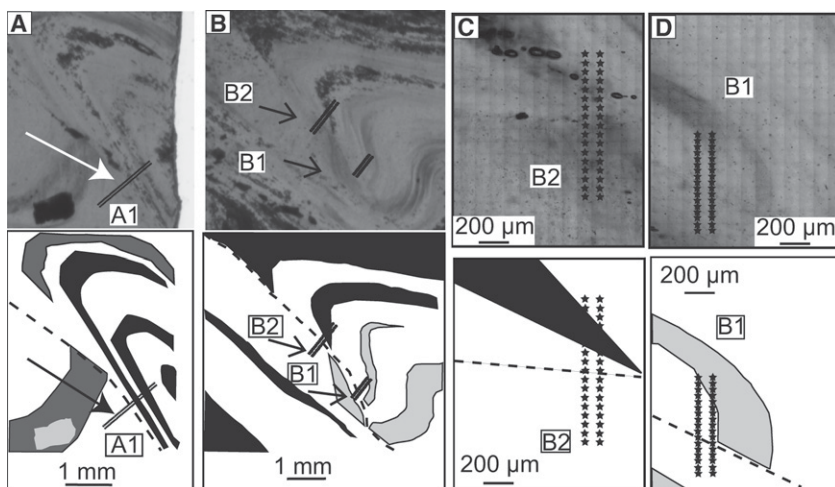


Figure 2. A: Photomicrograph of wafer A. **B:** Photomicrograph of wafer B. In A and B, double lines (arrow) mark positions of three synchrotron Fourier transform infrared spectroscopy transects (A1, B2, and B1) across fault truncating flow banding. Line drawings illustrate flow folding patterns; dashed lines mark fault traces. **C, D:** Enlargements and line drawings of B showing transects B2 and B1. Cloudy bands on glass are microlite rich of similar mineralogy everywhere (>20- μ m-long needles of pyroxene); dark ellipses at fold hinges are vesicles.

were carried out in two doubly polished wafers nearly perpendicular to the fault trace (Figs. 2A and 2B), which is at a right angle to the wafer surface (except for transect B2, where the fault plane dips $\sim 83^\circ$ into the wafer, measured using an optical microscope). Wafer thickness varied between 900 and 1000 μ m, measured with a micrometer with precision of ± 5 μ m. We report water concentrations along three transects (Figs. 2 and 3A) using a 10- μ m-diameter infrared beam at intervals of 40 μ m (for transects A1 and B1) and 60 μ m (for transect B2). For each transect, points were analyzed along two parallel lines 80–100 μ m apart. Water concentrations are reported as a single traverse, using the

average of the two adjacent analyses (Fig. 3A). These were used to compute uncertainty (gray bands in profiles in Fig. 3A), which is <3% of measured values. Errors associated with thickness measurements or microlite influence are smaller than variations from repeated measurements along the two parallel transects. We used 128 scans for each spectrum and analyzed the background after every 10 measurements. Total H_2O concentrations were determined from the intensity of the broad 4500 cm^{-1} and 5200 cm^{-1} absorption bands, and total water concentration (H_2O_{total}) was calculated as in Zhang et al. (1997). Peak heights were approximately three times greater for the OH⁻ peak (4500 cm^{-1}) than

for the H_2O peak (5200 cm^{-1}). Carbon dioxide concentration was below detection.

RESULTS

SFTIR Water Profiles

SFTIR transects define two plateaus of nearly constant water content separated by a V-shaped trough coincident with the fault trace (Fig. 3A). Water content for transect A1 decreases rapidly from the plateaus toward the trough over a region ~ 480 μ m wide. The difference in water content between the plateau on the left (0.66 wt% H_2O) and the right (0.73 wt% H_2O) is ~ 0.07 wt%, and as much as 0.12 wt% higher than the trough (0.61 wt% H_2O). The slope from plateau to trough is steeper and sharper on the right than on the left. Similarly, the V-shaped trough in B2 is ~ 480 μ m wide, but the plateau on the left (0.76 wt%) and right (0.78 wt%) sides have closer values, up to ~ 0.10 wt% higher than the trough (0.68 wt%). In contrast, water content along the third transect, B1, does not define a plateau, and water decreases continuously over ~ 600 μ m.

DISCUSSION

The volatile content of a given melt is controlled by temperature (T) and pressure (P). Previous studies of degassing have focused on decompression (Gonnermann and Manga, 2007); here we also consider temperature variation.

Troughs and Water Diffusion Profiles

Water content curves are fitted by diffusion profiles (Fig. 3B). Here we quantify diffusion into the fault (Fig. 2) following Castro et al. (2005, 2008). We consider that water diffusion began upon fracturing and ceased upon healing. This assumes negligible time for post-healing diffusion, implying rapid quenching (Fig. 3B), as supported by preservation of delicate water diffusion curves. Water diffusivity, D , in silicate melt has Arrhenian dependence on T , negligible dependence on P , and is a function of H_2O content (C) and distance from the fault trace, x (Ni and Zhang, 2008). We solved numerically the one-dimensional (1-D) diffusion equation:

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial x} \left[D(C, T, P) \frac{\partial C}{\partial x} \right], \quad (1)$$

where $D(C, T, P)$ is the diffusivity of H_2O in rhyolite melt (Ni and Zhang, 2008). We assumed either: (1) constant T everywhere and determined the best diffusion time (Fig. 3B), or (2) similar time for diffusion on both sides of the fault, and calculated T (Fig. 4).

For constant T models, calculations were carried out at 50 $^\circ C$ intervals between 750 and 950 $^\circ C$. The initial water content for each diffusion profile is set to the average of each plateau

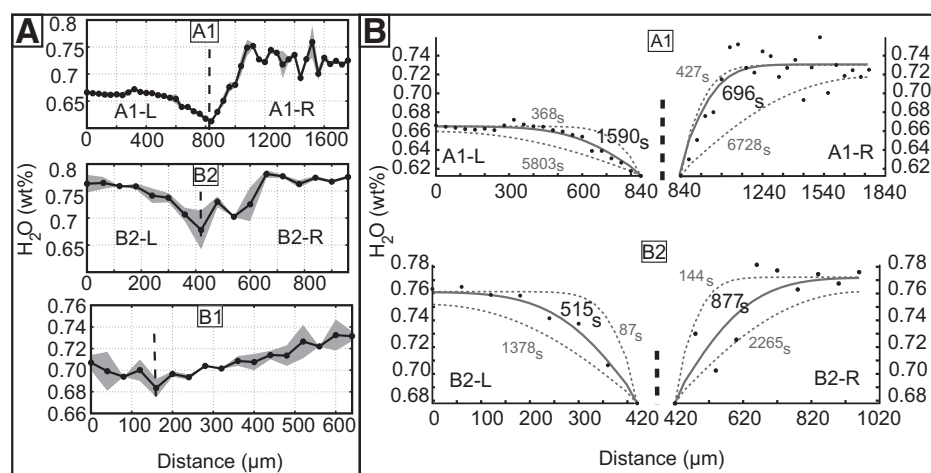


Figure 3. A: Synchrotron Fourier transform infrared spectroscopy water content profiles for transects A1, B2, and B1. Gray bands are average lateral variations in measurements and represent uncertainties. **B:** Measurements and water diffusion profiles for different diffusion times. A1-L and A1-R are left and right side synchrotron profiles in A1; B2-L and B2-R are left and right side profiles in B2. Dashed lines represent position of fault trace.

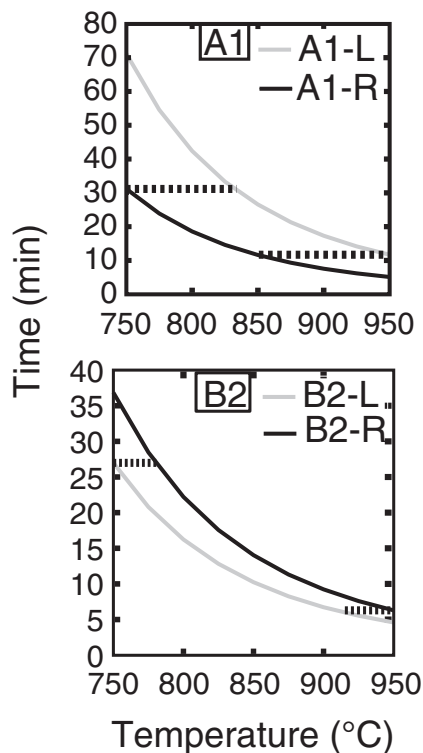


Figure 4. Diffusion times for temperature between 750 °C and 950 °C. Best-fit diffusion times for left and right sides of sample A1 (A1-L and A1-R) and B2 (B2-L and B2-R). Dashed lines mark different temperature for each profile for similar diffusion time. Time scales longer than ~30 min and shorter than ~10 min require temperatures beyond those modeled.

(Fig. 3B). The fault trace has the lowest water content in the profile and water flux in the plateau boundary of the model is defined as $\partial C/\partial x = 0$ (e.g., Castro et al., 2005). The best-fit model is the one that minimizes mean absolute error. For variations in measured water contents, uncertainties for modeled diffusion times vary between 20% and 35% of reported values. Multiple fractures (Fig. 2B) could explain the lack of plateaus in transect B1 (Fig. 3A), for which diffusion times or equilibrium conditions cannot be determined and so we do not report on it any further.

For constant T , calculated diffusion times vary from 73.7 min at 750 °C to 3.9 min at 950 °C (Fig. 4). For an eruptive temperature of 850 °C (Davì et al., 2009), diffusion lasted between 10 and 30 min for the different curves. Longer diffusion times can be reached for unreasonably low temperatures (Fig. 4). There is an issue, however, with results assuming constant T : they yield different best-fit diffusion times for each side of the fault (Fig. 3B). This requires melt on each side to be exposed to diffusion into the fault for different time spans, which we deem unlikely.

Our second approach was to equate diffusion time on either side of the fault and assume that

T , and therefore D , varied. Figure 4 and Table 1 show that to attain equal diffusion times, temperature differences across the fault would have been 80–100 °C for A1, and 30–40 °C for B2. Furthermore, Figure 4 indicates that melts at 750 and 830 °C on either side of the fault in A1 would produce the diffusion curves over a time span of 30 min. These temperatures are close to those estimated independently for equilibration of the plateau water contents (see following).

The higher value of thermal diffusivity compared to water diffusivity implies that thermal equilibrium would be reached before significant water diffusion. Thus, if water diffusion took place at different temperatures, it would have taken place before the two melt portions were juxtaposed by the fault.

Calculations of the evolution of water content on a fault plane with initially 0% water, surrounded by two high-water-content plateaus on either side, reveal that post-healing diffusion raises water content on the fault to the measured value of 0.6% in <60 s. This indicates the sensitivity of the trough, and that post-healing diffusion longer than 60 s would have given rise to a shallower, U-shaped trough. Thus, post-healing diffusion implies that water content on the fault would have been lower at healing, and that P - T estimates for the fault are conservative: lower P and higher T could have prevailed before healing.

Plateaus Indicative of Early Equilibrium

Plateau pairs in A1 and B2 indicate that water content in the small volumes of melt on either side of the fault had at some point reached equilibrium with surrounding P - T conditions. However, in A1 (Fig. 3) the plateau on the left (0.66 wt% H₂O) contains less water than the one on the right (0.73 wt% H₂O). Because the two sides are in contact, they could not both have been in equilibrium with surrounding P - T simultaneously. We therefore infer that fault displacement brought them into contact after they equilibrated in regions with different P - T conditions. Thus, water contents in melts across the fault were both out of equilibrium with conditions imposed by the fault and out of equilibrium with the surroundings after faulting.

Assuming that the small portion of melts were at identical and constant T before faulting, we estimate the difference in equilibration pressure (ΔP) by translating the average water content of each plateau to pressures, using established relationships (Newman and Lowenstern, 2002). Considering magmatic pressure (i.e., no overpressure), ΔP can be further translated to depth difference (ΔZ). For a rhyolite density of 2323 kg/m³ (calculated for the Lami composition with ~0.7 wt% H₂O at ~3.8 MPa and T of 750–950 °C; Spera, 2000) and $T = 850$ °C, the water difference between the plateaus in A1 of

TABLE 1. ΔT FOR FIXED TIMES

A1 time (min)	ΔT (°C)	B2 time (min)	ΔT (°C)
11	101	6	41
15	96	10	37.5
20	90	15	34.5
25	86	20	32
30	83	25	31

Note: Temperature variations (ΔT) on each side of the fault in A1 and B2 assuming diffusion curve pairs had equal times to develop.

~0.07 wt% corresponds to a ΔP of 0.74 MPa, for CO₂ = 0 ppm (Newman and Lowenstern, 2002) and a ΔZ of ~32 m. These are maximum values since melts could have been overpressurized or at different temperatures. Using estimated ΔP , ΔZ , and diffusion times, we estimate relative melt decompression rate and relative vertical velocity across the fault (“relative” here refers to one side in relation to the other, rather than absolute values). Typical times between 10 and 30 min yield a relative decompression rate between 1.5 and 4.5 MPa h⁻¹ and a relative vertical velocity between 1.8 and 4.8 cm s⁻¹.

Alternatively, assuming constant P , we investigate whether T variations explain water content variations. For $P = 3.8$ MPa (Newman and Lowenstern, 2002), the plateau pair in A1 at 0.66 wt% H₂O and 0.73 wt% H₂O would have been in equilibrium if T was ~850 °C and 750 °C, respectively. For these values, diffusion curves on either side of the fault would both be resolved for an identical diffusion time of 30 min (Table 1). This is unlike the case where we varied P , for which each side had a different diffusion time. Furthermore, for $P = 3.8$ MPa also within the fault plane, the 0.61 wt% H₂O in the trough would have been in equilibrium for $T = 950$ °C (Newman and Lowenstern, 2002). This suggests that shear heating on the fault (Hess et al., 2008) could explain the origin of the trough. As for pseudotachylites, the total amount of energy due to shearing is sufficient only to raise temperatures within a narrow band.

Our simple approach assumes fixed water content in the deep part of the trough, and that water in the surroundings diffuses toward this value. Our two end-member cases indicate that the trough in A1 could represent either the minimum P site (in equilibrium at 3.2 MPa for $T = 850$ °C), or the maximum T site (in equilibrium at 950 °C for $P = 3.8$ MPa). In summary, SFTIR results indicate water disequilibrium at two levels: (1) parcels of magmas had reached equilibrium, as indicated by plateaus, but faulting displaced them to new conditions. Melt on either side of the fault either equilibrated ~30 m apart vertically (for constant T), or had ~100 °C temperature difference (for constant P). They were then brought together by faulting and were no longer in equilibrium. (2) The fault plane

imposed new, dynamic P - T conditions and a permeable path for volatiles, giving rise to diffusion profiles that indicate active faulting and healing for 10–30 min. Solutions of diffusion curve pairs for similar times require temperature differences of ~ 100 °C, similar to that estimated for pre-faulting plateaus equilibration at constant P . The loss of volatiles from the melt through diffusion inhibits vesiculation. Temperature increase due to shear heating eases vesiculation, but also favors volatile loss into faults due to increased diffusivity.

Autobrecciation and Melt Degassing: Obsidian Formation

Nonexplosive fragmentation of ascending magmas has been inferred based on clastic textures (Tuffen et al., 2003), supported by earthquakes during dome-forming silicic eruptions (Goto, 1999), and modeled for magma ascent in conduits (Cordonnier et al., 2009; Gonnermann and Manga, 2003). We have shown that degassing occurs through fracturing in rhyolitic melts, supporting suggestions that repeated fracturing and healing assists degassing, suppressing vesiculation and favoring obsidian formation (Gonnermann and Manga, 2005; Rust and Cashman, 2007; Rust et al., 2004; Tuffen et al., 2010).

If this is so, fracturing and degassing should proceed at rates compatible with magma movement in the conduit. We note that faulting and healing durations, here constrained to 10^2 – 10^3 s, are less than but broadly comparable with estimates of the time scale for repeated fracturing and healing from a simple fracture and welding model (Tuffen et al. 2003), and compatible with interevent time for hybrid and long-period earthquake events during dome eruptions ($10^{2.5}$ – $10^{4.5}$ s; Voight et al., 1999), possibly linked to repeated fracturing and healing (Tuffen et al., 2003). We postulate that degassing due to repeated fracturing and healing events in conduits could explain the temporal shift from explosive to effusive eruption common in silicic systems (Castro and Dingwell, 2009; Eichelberger et al., 1986; Fink, 1983), including those in Lipari.

CONCLUSIONS

Results show that water diffused from the melt into a low- P or high- T fault plane. Water diffusion into the fault lasted between ~ 10 and 30 min before healing and quenching. Different water plateaus determined for melts on either side of the fault indicate that each side had achieved equilibrium in terms of water content before being brought together by faulting into new P - T conditions. The duration of the process matches roughly that of the repeat time of earthquakes during silicic lava eruptions (Tuffen et al., 2003). Results indicate that pervasive fracturing is a potential mechanism for degassing of rhyolite magma, explaining the origin of obsid-

ian and the commonly observed temporal shift from explosive to effusive eruptions.

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