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A volcanic degassing event at the explosive-effusive transition

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[1] A sequence of five Vulcanian explosions followed a lava dome collapse in July 2003 at Soufrière Hills Volcano. Each explosion occurred at $\sim t = 190\ n^{4.3}$ where $n = 1-5$ and t is the time (s) since the decompression rate peak during the collapse. Instead of a sixth explosion at the predicted time, a rapid emission of 97×10^3 kg SO_2 was observed by a spectrometer network. This event represents the transition from explosive to effusive activity. After the last explosion, high magma ascent rates were maintained, but the critical overpressure explosion criterion was not reached. Instead, degassing and crystallisation in the upper conduit caused horizontal gradients in viscosity and flow rate, and brittle failure at the walls when the rate of shear strain exceeded a critical value. Development of a permeable shear zone allowed gas release, relief of overpressure and a return to effusive lava-dome building.
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1. Introduction

[2] Transitions between lava dome growth and explosive activity occur at silicic volcanoes on short timescales and with little warning [e.g., *Stix et al.*, 1993]. Our understanding of such events has been achieved through observation of deformation [*Voight et al.*, 1999; *Widiwijayanti et al.*, 2005], magma rheology [*Sparks et al.*, 2000] and petrological studies [*Couch et al.*, 2003; *Clarke et al.*, 2007] and the results compared to models of conduit magma flow [*Melnik and Sparks*, 1999, 2002a; *Clarke et al.*, 2002]. Vulcanian explosions occur when the conduit is suddenly decompressed, or when the gas overpressure in rising magma reaches a threshold value. At Soufrière Hills Volcano, explosive activity has typically taken place after a major dome collapse, when the conduit has been decompressed. Repetitive explosions are triggered when a critical gas overpressure (typically a few MPa at a few hundred m depth) is exceeded in growing bubbles in the rapidly rising magma [*Melnik and Sparks*, 2002a]. This increase in overpressure causes inflation of the ground surface between explosions [*Voight et al.*, 1999; *Widiwijayanti et al.*, 2005]. After an explosion, the system attempts to reach a steady state by magma ascent and vesiculation, but will be interrupted when the overpressure once again exceeds the tensile strength of the magma; the system thus fails repeatedly. Vulcanian explosions that occur in response to decompression during a lava dome collapse are associated with the eruption of dense clasts (>2000 kgm^{-3}), interpreted as a degassed plug [*Voight et al.*, 1999; *Diller et al.*, 2006; *Clarke et al.*, 2007]; subsequent explosions in a sequence are associated with pumiceous ejecta [*Druitt et al.*, 2002].

[3] Melt containing dissolved volatiles at a high pressure is exposed to atmospheric pressure [*Alidibirov*, 1994] and a fragmentation wave moves down the conduit, quenching the magma and ejecting a mixture of pyroclastic material and gas. The explosion ceases when the fragmentation wave reaches magma that does not satisfy the conditions necessary for fragmentation [*Zhang*, 1999]. Vulcanian explosions are short-lived owing to the high viscosity of the melt, which retards bubble growth and prevents sustained fragmentation [*Melnik and Sparks*, 2002b]. The transition from explosive to effusive activity occurs when the criteria for explosions to occur are no longer met. This may be due to stagnation of magma flow up the conduit [*Woods and Koyaguchi*, 1994] or gas loss through the conduit walls [*Jaupart and Allègre*, 1991]. Measurements of volcanic gases have thus far failed to illustrate such processes directly. Gas emitted during Vulcanian explosions is difficult to quantify with ground-based measurements owing to the large amounts of ash emitted during the event, which typically precludes spectroscopic measurements due to scattering. The principle volatile species, H_2O, is present in large and variable amounts in the background air, making it a challenging target for spectroscopic measurements. An opportunity presented itself, however, during a sequence of Vulcanian explosions following a large lava dome collapse in July 2003 [*Herd et al.*, 2005], to measure the amount of SO_2 emitted during a degassing event that followed the fifth explosion of the sequence, using the Montserrat Volcano Observatory network of UV scanning spectrometers [*Edmonds et al.*, 2003]. These measurements illustrate some aspects of the transition from explosive to effusive eruptive behaviour at Soufrière Hills Volcano.

2. SO_2 Flux Data Acquisition and Processing

[4] The UV scanning spectrometer network at Soufrière Hills Volcano has been described in detail elsewhere [*Edmonds et al.*, 2003]. Three scanning spectrometers are positioned 3–5 km downwind of the volcano (Figure 1). At each site, a UV spectrometer (Ocean Optics S2000) is connected to a scanning assembly, driven by a stepper motor. The optical part of the assembly comprises a quartz window to receive the radiation and a right-angle prism to reflect it into a 10 cm-long telescope, which focuses it into a fiber optic cable and hence into the spectrometer. The assembly scans vertically through the volcanic plume at a rate of 0.5–3°/s, from 08:00 to 16:00 each day, acquiring UV spectra every 0.5–3 s. The spectra record UV intensity in the window 303–315 nm, a range that encompasses a

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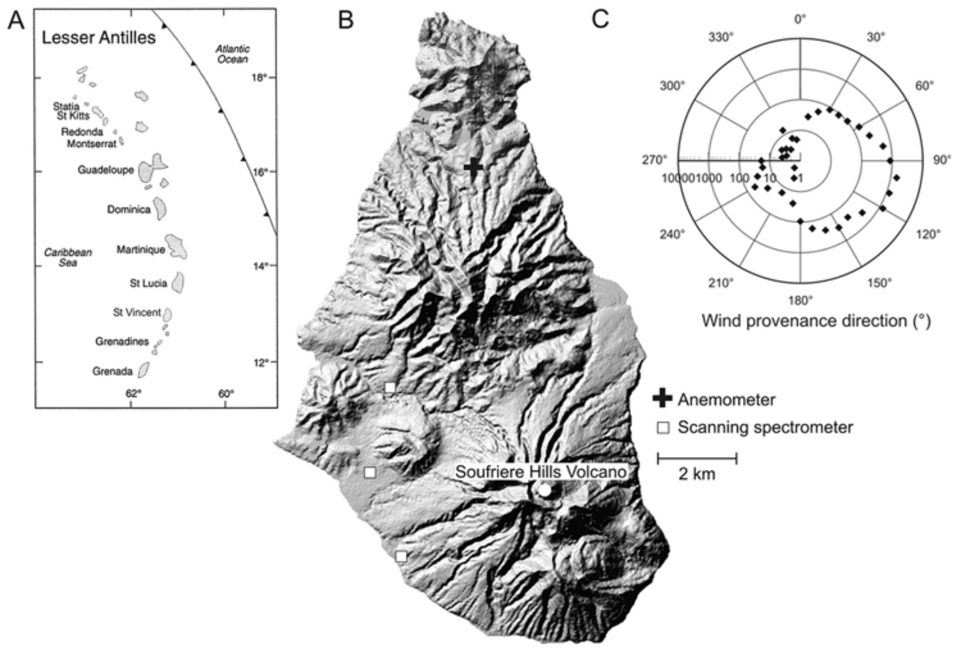

Figure 1. (a) Map of the Lesser Antilles Arc. (b) Map of Montserrat to show the Soufrière Hills Volcano and the scanning spectrometers. (c) Hourly-averaged wind direction measured by an anemometer in northern Montserrat (location marked on Figure 1a) between 1 October 1996 and 1 October 1997.

characteristic absorption feature for SO_2, with a resolution of 0.07 nm. The spectra are processed using Differential Optical Absorption Spectroscopy (DOAS) [*Platt*, 1994], whereby SO_2 absorbs radiation at a wavelength j and:

$$I_j = I_{j,0} \exp(lc\varepsilon) \tag{1}$$

where I_j is the intensity at a wavelength j, $I_{j,0}$ is the intensity of the background spectrum at wavelength j, l is the path length, c is the gas concentration and ε is the absorptivity. "Dark" spectra are subtracted from the measured spectra to eliminate electrical noise. The spectra are normalised by a background spectrum (one that contains no absorption due to SO_2), which eliminates the atmospheric "signature" from the spectra and isolates only the volcanic component, of which SO_2 is dominant. A high-pass filter is applied to the spectra and the result subtracted, leaving only the high frequency component of the absorption feature. The absorbance is calculated from:

$$A = \log\left(I_j \middle/ I_{j,0} \right) = lc\varepsilon \tag{2}$$

The absorbance is compared to a published cross-section [*Vandaele et al.*, 1994] and the fit used to calculate the concentration-path length of SO_2 (in units of ppm.m). The set of spectra that make up one "scan" through the plume are then used to calculate a flux of SO_2. The concentration-pathlengths are integrated across the plume by multiplying their horizontal components by the length of the horizontal segment of plume at each step. The total SO_2 amount in the 2-D plume section is then multiplied by plume speed (estimated using the wind speed at plume height) to yield a flux. The errors on the measurements are estimated as +30 and −15%. The scanning spectrometers were buried in ash during 13–15 July and therefore were not in operation during the Vulcanian explosions.

3. Results and Discussion

[5] The five Vulcanian explosions during 13–15 July 2003 occurred at progressively longer intervals after the peak in the lava dome collapse [*Herd et al.*, 2005]. They were followed by a strong SO_2 degassing event, recorded on 16 July 2003 (Figures 2a and 2b). The degassing event was not associated with an explosion, but with an increase in gas plume vigour and ash venting, although not so much as to impede spectroscopic measurements. The degassing event followed the power trend defined by the previous 5 explosions and occurred at a time, t (after the initial conduit decompression, in s) approximately equal to $t = 190\ n^{4.3}$ where n is equal to 6 (Figure 2c). The degassing event lasted for around 53 minutes, during which time 97×10^3 kg SO_2 were emitted (an average flux of 30 kg/s), with a peak measured flux of 69 kg/s at 13:24 UT 16 July. This is small compared to SO_2 emissions during large lava dome collapses, which have been associated with releases of up to 200×10^9 kg SO_2 over 1–2 hours [*Prata et al.*, 2007] but an order of magnitude larger than both the background rate of degassing on 16 July 2003 (2–4 kg/s) and the average SO_2 emission rate for 2003 (6.9 kg/s) (Figures 2a and 2b). The emitted gas is likely to have been made up of other volatile species aside from SO_2. Gases emitted from the lava

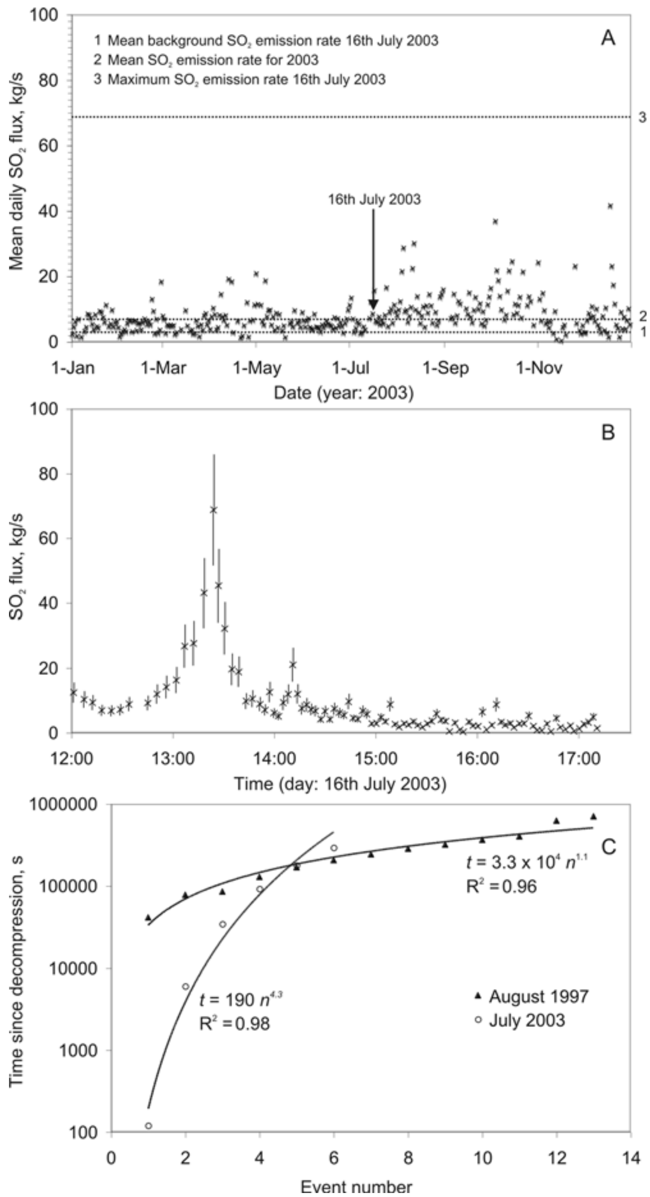

Figure 2. (a) The average daily SO_2 emission rate from Soufrière Hills Volcano, Montserrat, from 1 January to 31 December 2003; also marked are the mean SO_2 emission rate for 2003 and the background and maximum SO_2 emission rates on 16 July 2003. (b) The SO_2 degassing event on 16 July 2003, in kg/s against time. (c) Plot to show the intervals between the initial decompression (at the peak of the lava dome collapse) and each explosion/event against event number, *n*, where values 1–5 are Vulcanian explosions and 6 is the degassing event on 16 July. Also shown is the sequence of Vulcanian explosions that occurred in August 1997.

dome in 1996 contained H_2O and SO_2 in the mass ratio 6.7 [*Hammouya et al.*, 1998]; we therefore estimate that $\sim 650 \times 10^3$ kg H_2O vapour was also emitted during the event.

[6] Vulcanian explosions in August 1997 occurred at approximately regular intervals after the 3 August 1997 lava dome collapse, in contrast to the explosions in July 2003, which occurred at progressively longer intervals (Figure 2b). It has been proposed that the 1997 explosions involved repeated magma ascent and shallow stagnation, followed by the evacuation of typically approximately 3×10^5 m^3 DRE material from the conduit [*Druitt et al.*, 2002]. Magma injection into the base of the conduit and a constant or increasing magma chamber overpressure is clearly required to sustain this rate of eruption. The July 2003 explosions were, in contrast, a consequence of little or no magma injection in the conduit from the magma chamber and a declining magma chamber pressure. A 2-year pause in eruption followed the cessation of explosive activity.

[7] In order to assess whether degassing of magma in the upper conduit caused the degassing event, we calculate the mass of gas that would be produced by a magma cylinder 400–900 m high and 20–30 m in diameter (the dimensions of a "degassed cap" [*Voight et al.*, 1999; *Widiwijayanti et al.*, 2005]). The volume of melt ($\sim$5–15 vol% of the andesite at shallow depths in the conduit) is $0.5–2.5 \times 10^6$ m^3 and the difference in S concentration between melt inclusions (which represent pre-eruptive melt) and matrix glasses (post-eruptive melt) is at most 10–20 ppm. In fact, evidence suggests that S does not degas from the andesite syn-eruption at all [*Edmonds et al.*, 2001]. Complete degassing of the upper part of the conduit would only therefore yield a maximum of $2–9 \times 10^3$ Kg SO_2, an order of magnitude less than observed. This suggests that either a) a much greater volume of andesite contributed to the degassing event (more than is contained within the conduit) or b) the S existed in a free vapour phase beneath the degassed cap, where magma vesicularity is expected to be a maximum. The latter is most feasible; resident magmas in arc chambers are usually saturated with a fluid phase in equilibrium with the resident magma owing to the high initial volatile content of arc magmas (derived from the subducting slab) and the high crystallinity and the shallow depth of the chamber, which results in the exsolution of all but the most soluble species. Most of the S will therefore exist in the vapour phase, rather than in the melt. Such a pre-eruptive fluid phase has been proposed as a source for the high emissions of SO_2 during large silicic eruptions such as Pinatubo in 1991 [*Wallace and Gerlach*, 1994]. Additional evidence for a free vapour phase in the magma chamber comes from borehole dilatometers, which recorded an expansion of the magma chamber caused by the growth of existing vapour bubbles as a consequence of the decrease in lithostatic load [*Voight et al.*, 2006].

[8] Given the existence of a volumetrically-significant pre-existing vapour phase in the conduit, we next assess how it escaped to the atmosphere to cause the gas emission event. We consider two possibilities: non-explosive partial fragmentation; and gas loss at wall shear zones (models A and B; Figure 3). The degassing event may represent a "failed" explosion (Model A). The overpressure criterion for the onset of a Vulcanian explosion [*Melnik and Sparks*, 2002a] may have been barely met. Instead of explosive fragmentation, localized failure of bubble walls and the development of permeability in the form of interconnected chains of bubbles may have developed, allowing a degassing event to take place that relieved the overpressure and prevented explosive activity. Measurements on SHV products illustrate that permeability can change by orders of magnitude for very small changes in porosity: permeabil-

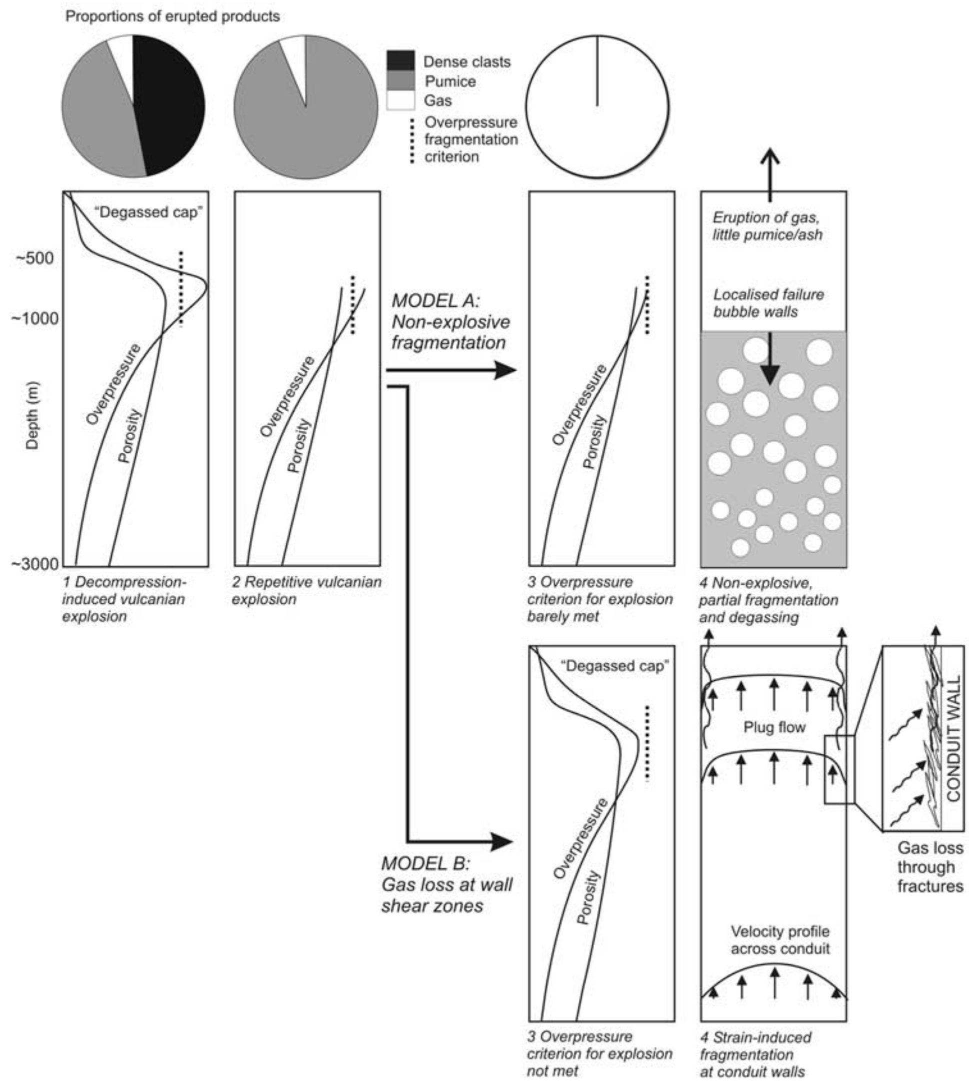

Figure 3. Schematic diagrams to illustrate the degassing event mechanism at Soufrière Hills Volcano. A decompression-induced Vulcanian explosion occurs when overpressure at the base of a degassed plug exceeds the tensile strength of the plug. A large proportion of dense clasts are erupted. Subsequent repetitive Vulcanian explosions occur when a critical overpressure is exceeded inside growing bubbles in rapidly ascending magma. Typically, these explosions are associated with the eruption of a high proportion of pumiceous material. The degassing event (characterised by the emissions of mostly gas, with a small amount of ash) may be explained by either non-explosive fragmentation (model A) or gas loss at conduit wall shear zones (model B). The models are discussed in the text.

ities range from 28×10^{-15} m^2 (at 21.5% porosity) for a lava block to 5092.9×10^{-15} m^2 (at 33% porosity) for vesicular lava [*Melnik and Sparks*, 2002b]. It is likely then that only very small changes in the porosity structure would be required in order to allow gas escape. The degassing event ended when either the localised, non-explosive fragmentation ceased; or when the amount of gas available to degas in the bubble network was depleted; or both.

[9] Our preferred model involves gas loss at the conduit wall zones (model B; Figure 3). As well as a vertical viscosity gradient in the conduit, horizontal gradients are also predicted to exist; cooling, degassing and crystallinity are expected to be highest at the conduit walls [*Sparks et al.*, 1999; *Tuffen et al.*, 2003]. The conduit wall zones are therefore susceptible to brittle failure at a relatively low rate of shear strain. Rapid development of a permeable shear zone might therefore be a viable mechanism to release large volumes of SO_2-rich gases and hence prevent explosive activity (Figure 3). Shear-induced "fragmentation" at conduit walls depends on melt viscosity and the strain rate of the magma [*Papale*, 1999; *Gonnerman and Manga*, 2003]. As melt viscosity increases owing to degassing and crystallisation, strain rate increases at the conduit walls. The threshold for strain-induced fragmentation scales with the critical conduit shear-strain rate ($\dot{\gamma}$), which is proportional to magma flux Q and inversely proportional to conduit radius R:

$$\dot{\gamma} = Q \Big/ \pi R^3 \tag{3}$$

and relates to the relaxed melt viscosity μ whereby

$$\dot{\gamma} = CG_{\infty}\mu^{-0.9} \qquad (4)$$

where C is a fitting parameter equal to $0.01^{-0.1}$ and G_{∞} is the elastic modulus at infinite frequency (equal to 10 GPa) [*Gonnerman and Manga*, 2003]. At a given Q and R, shear-induced fragmentation is predicted to occur once μ exceeds the threshold value given by equation above. For high-viscosity magma ($\mu = 10^{12}-10^{14}$ Pas; [*Sparks et al.*, 2000]) shear-strain rates of $10^{-5}-10^{-3}$ s^{-1} are required, which can be achieved at magma ascent rates of 0.01–0.5 m/s through a conduit of radius 15 m (equivalent to magma fluxes of 7–350 m^3/s). Effusion rates during lava-dome building are typically within the range 0–10 m^3/s [*Sparks and Young*, 2002]. Shear-induced conduit-wall-zone fragmentation thresholds are therefore readily achieved (particularly if the conduit radius is <15 m) and it is likely that this mechanism not only accounts for the observations presented here, but represents a common mode of gas loss both at SHV and other similar volcanoes. This mechanism is supported by observations of brittle failure and annealing cycles in rhyolitic magma in a frozen vent in Iceland [*Tuffen et al.*, 2003]; the proposal that tilt cycles during periods of high magma extrusion rate in 1997 were caused by shear stresses at the conduit walls [*Green et al.*, 2006]; and that hybrid earthquakes may be triggered by brittle failure of magma at the conduit walls at Soufrière Hills Volcano [*Tuffen et al.*, 2003; *Neuberg et al.*, 2006]. Observations of gas emitted from the base of spines and shear lobes have been commonplace at Soufrière Hills [e.g., *Sparks et al.*, 2000]; gas and ash emission from ring-shaped fractures at the vent-edges of Santiaguito volcano, Guatamala, suggest magma plug-flow and gas loss along conduit wall shear zones [*Bluth and Rose*, 2004]; and fault gouge at the margins of dacite spines at Mount St. Helens indicate that brittle failure of magma takes place [*Cashman et al.*, 2007].

[10] The abrupt cessation of the SO_2 release may represent either the depletion of the vapour source; or an annealing event, whereby the fracture network seals itself in the manner described by *Tuffen et al.* [2003] for brittle-ductile transitions in rhyolite. Melt crystallinity is typically >95% after microlite crystallisation, which makes annealing unlikely, although through-flow of magmatic volatiles (mostly H_2O vapour) through the wall shear zones [*Rust et al.*, 2004] may elevate the volatile concentration of the melt and hence allow a return to visco-elastic deformation and welding, in the manner described by *Sparks et al.* [1999].

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