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Volcanic eruptions from *ghost* magma chambers

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[1] Recent studies have proposed that magma reservoirs crystallized to a virtually rigid crystal-mush can be partially remelted by diffusion of hot fluids. We show that for a crystal mush with the composition of a K-trachyte from the Campanian Ignimbrite (CI) Eruption, remelting can occur without a significant increase of the magma temperature, but simply by diffusion of H_2O by the magmatic gases feeding the system. The CI origin is not the issue here, but rather the chemical and physical behavior of an almost solidified magma mass left over in a reservoir after a major eruption. To test our hypothesis, we run high pressure/high temperature laboratory experiments to study the kinetics of water diffusion, together with thermodynamics and fluid diffusion modelling. For small diffusivities, or large diffusion time, the remelting mechanism proposed above needs to be replaced by other processes as gas percolation or intrusion of a magmatic mass. **Citation:** Trigila, R., M. Battaglia, G. Sottili, and M. Brilli (2008), Volcanic eruptions from *ghost* magma chambers, *Geophys. Res. Lett.*, *35*, L16304, doi:10.1029/2008GL034579.

1. Introduction

[2] A number of studies have proposed that magma reservoirs crystallized to a virtually rigid crystal-mush can be partially remelted by diffusion of hot fluids, and some volcanic eruptions have been interpreted as induced by this mechanism [e.g., *Couch et al.*, 2001]. In particular, percolation of magmatic gases through the reservoir at near-solidus temperatures has been proposed as the potential heating mechanism [*Bachmann and Bergantz*, 2006]. The magmatic system beneath Campi Flegrei caldera, an active volcanic area near Naples (Italy), is a good analog to study the reactivation of these near-solidus magmatic reservoirs (the so called *ghost* magma chambers), because of its trachytic composition giving to the chemical system a nearly eutectic behaviour [*Fowler et al.*, 2007]. Here, we show that for the studied composition (i.e., a K-trachyte from the Campanian Ignimbrite (CI) Eruption, Campi Flegrei) remelting can occur without a significant increase of the magma temperature, but simply by diffusion of H_2O by the magmatic gases feeding the system. The CI origin is not the issue here, but rather the chemical and physical behavior of an almost solidified magma mass left over in a reservoir after a major eruption. Laboratory experiments and modelling results indicate that the kinetics of this process is (a) compatible with the amount of gases exolving from the magma mass in its way to the surface and (b) fast enough to overcome the solidification of large reservoirs [*Bachmann and Bergantz*, 2003]. These results, suggesting that near-solidus magmatic systems (e.g.; a magma containing at least 85–90 wt% of crystalline phases) can be reactivated without the input of new magma, have implications on the hazard assessment at Campi Flegrei and similar active volcanic regions.

[3] Campi Flegrei caldera formed by collapse of the magma reservoir after two major eruptions: the Campanian Ignimbrite eruption (39 kyr B.P.) and the more recent Neapolitan Yellow Tuff eruption (12 kyr B.P.) [*De Vivo et al.*, 2001]. Geological and petrological evidences indicate that most of the eruptions in the past 12,000 years have been fed by several independent conduits, supplied by the reservoir left behind by the caldera forming eruptions [*D'Antonio et al.*, 1999]. The 1538 AD Monte Nuovo eruption, the most recent unrest episodes as well as the widespread hydrothermal and fumarolic activity testify that the Campi Flegrei magmatic system is still active [*De Natale et al.*, 2006]. Extrapolation of temperature data from geothermal drills indicates a near magmatic environment climbing up to 4 km depth [*Wohletz et al.*, 1999]. Notwithstanding all these evidences, the presence of a molten body beneath the Campi Flegrei caldera is still an open issue [*Vanorio et al.*, 2005].

[4] A magmatic system at near-solidus conditions behaves as a solid body, but if this mass is permeated by the fluids (i.e., H_2O) rising along the feeding dikes, the *liquidus* and *solidus* temperatures of the magmatic system will be controlled by the P, T phase relations for the water saturated system, and may fall as much as 100–150°C depending on the confining pressure. Under these conditions, the magma body will remelt and become eruptible without any increase in the system temperature. To test this hypothesis, a number of laboratory experiments (together with thermodynamics and fluid mechanics modelling) were performed to investigate the behaviour of the magmatic system beneath Campi Flegrei.

2. Laboratory Experiments and Modelling

[5] The obsidian samples used in this study were collected within the Breccia Museo Unit of the Campanian Ignimbrite eruption. The Unit is exposed in Procida Island (near Naples, Italy) and includes fresh, decimetric, obsidian blocks, strictly reproducing the trachytic composition of the predominant pumice deposits of the Campanian Ignimbrite pyroclastic flow. The obsidian blocks show millimetric euhedral sanidine (*San*) phenocrysts (up to 10–15% of the volume) regularly associated with augitic pyroxene (*Cpx*), corroded bytownite (*Pl*), euhedral andesine (*Pl*), ulvo-spinel (*Sp*), biotite (*Bio*) and apatite (*Ap*) (all together 5–8% of the total volume). All these minerals, but the bytownite, crystallized

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in the Campanian Ignimbrite intracrustal reservoir and represent the product of the fractionation of the basaltic trachyandesite melt, taken as indicative of the parent magma source [*Fowler et al.*, 2007]. In Figure 1, we report on the TAS classificative diagram the composition of Campanian Ignimbrite products [*Fowler et al.*, 2007] together with the composition of the studied obsidian block and, for comparison, the composition the scoria glass from the 1538 AD Monte Nuovo eruption.

[6] The experimental and theoretical studies were organized in three steps. In the first step, the phase relations of the magmatic system were modelled, using the MELTS code [*Ghiorso and Sack*, 1995], in the range of pressures (up to 260 MPa) where the Campanian Ignimbrite magma chamber could have possibly been located [*D'Antonio et al.*, 1999; *Wohletz et al.*, 1999; *Fowler et al.*, 2007]. In the modelling, we used the CI obsidian composition (see Figure 1 caption) as input for isobaric MELTS code simulations under equilibrium mode. Particular attention was paid to determine the near-liquidus temperatures of the system under virtually dry ($H_2O^m \approx 0.4$ wt%) or water saturated conditions ($H_2O^m \approx 4.5$ wt% at 100 MPa) and $f0_2$ of the Ni-NiO buffer. These relations were then compared (under the same dry and saturated H_2O^m conditions) with

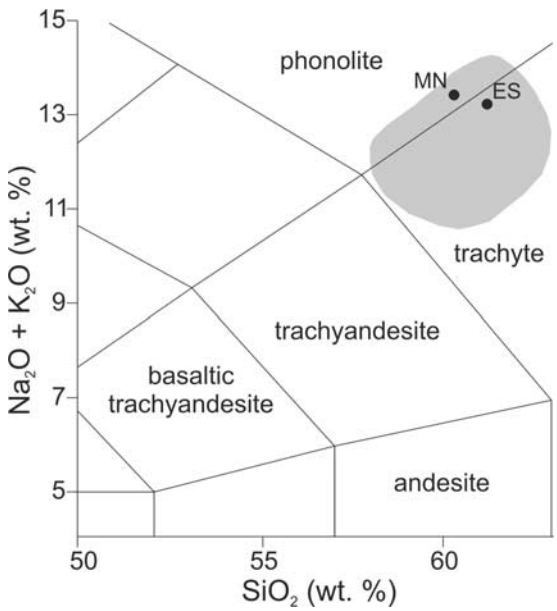

Figure 1. Total alkalies-silica diagram for the erupted composition of the Campanian Ignimbrite (grey area; modified after *Fowler et al.* [2007]). In the diagram we have included, for comparison, the composition of the K-trachytic obsidian block sample (ES) used for the experiments and the composition of the phonotrachytic scoria glass from M. Nuovo 1538 AD eruption (MN). The chemical analyses of both samples (including L.O.I., F and Cl) are the following: (ES) SiO_2 = 61.54, TiO_2 = 0.29, Al_2O_3 = 18.74, FeO_{tot} = 3.30, MnO = 0.18, MgO = 0.24, CaO = 1.79, Na_2O = 6.16, K_2O = 7.09, P_2O_5 = 0.06, F = 0.21, Cl = 0.36, L.O.I. = 0.43; (MN) SiO_2 = 59.70, TiO_2 = 0.42, Al_2O_3 = 19.44, FeO_{tot} = 3.16, MnO = 0.19, MgO = 0.18, CaO = 1.66, Na_2O = 5.84, K_2O = 7.47, P_2O_5 = 0.05, F = 0.39, Cl = 0.62, L.O.I. = 0.48.

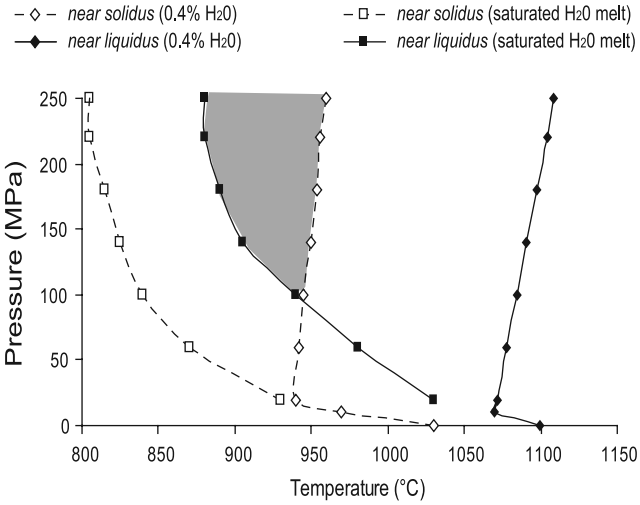

Figure 2. P-T relations of the Campanian Ignimbrite chemical system from isobaric simulations. Near-liquidus (melt > 98%) and crystal-mush (crystalline fraction > 85 wt%) P-T relations of the Campanian Ignimbrite chemical system under almost dry (H_2O^m = 0.4 wt%), and H_2O saturated conditions. It is worth noting that for pressures exceeding 100 MPa, the near-liquidus temperatures of the H_2O saturated system (solid line/solid squares) move below the temperatures of the near-solidus dry system (dashed line/open diamonds). The grey area identifies the P-T range for remelting under water saturated conditions.

those for a severely crystallized system (e.g., crystal fraction > 85 wt%). The results in Figure 2 show a dramatic decrease of the *liquidus* temperatures for the water saturated melt for pressures up to 100 MPa. Above this pressure value, the near-liquidus temperature of the water saturated system moves below the near-solidus temperature of the same system under virtually dry conditions. In Figure 3, the crystallization sequence at constant pressure (100 MPa) is shown for the system under water saturated conditions.

[7] In the second step the sequence of crystallizing phases was cross-checked by 5 equilibrium experiments at $P(H_2O)$ = 100 MPa (see results in Figure 3). The target of these experiments was mainly to estimate the amount of melt in equilibrium with the crystalline phases. Details on the equipment and experimental procedures are given in the section on kinetic experiments. Both the simulated and the experimental assembly given by the sequential entry of *Sp-San-Pl*, followed by rare *Cpx*, *Ap* and *Bio*, duplicates the assembly found for the Campanian Ignimbrite. Our results indicate that more than 60% of the system crystallizes between 890 and 920°C. The physical dominant process is the crystallization of the K-feldspar (*Kfs*). However taking into consideration the amount of liquid from the experiments in the reported range of pressure under equilibrium conditions, we found a general agreement between the experimental and calculated values.

[8] Finally, we estimated the time required for supercritical water to diffuse through the system. Diffusion time scale is a critical parameter for the rejuvenation of the magmatic system. For very small diffusivities, or very large diffusion time, the remelting mechanism proposed above needs to be replaced by other processes as gas percolation

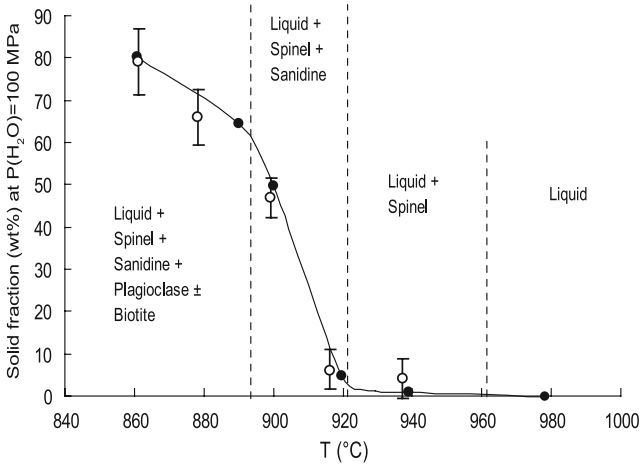

Figure 3. Modelled equilibrium phase relations calculated at $P(H_2O)$ = 100 MPa, using MELTS code [*Ghiorso and Sack*, 1995]. Figure 3 illustrates the rate of crystallization process vs. temperature and the corresponding sequential entry of the essential phases. It is evident the role of K-feldspar (sanidine), whose extensive crystallization between 890 and 920°C drives the system to crystal-mush conditions. Closed circles indicate the temperature values used for the MELTS code simulations. Open circles indicate the experimental temperatures and the respective crystalline assemblages. We measured the glass fraction (vol %) of the experimental samples by modal analysis with a Swift point counter. Error bars are 2 standard deviations.

[*Bachmann and Bergantz*, 2006] or intrusion of a magmatic mass [*Humphreys et al.*, 2006].

[9] The preparation of the starting material was directed to reproduce the possible textures of an almost consolidated magma chamber. For this reason, the finely ground powder of the obsidian sample was melted to simulate the two possible end members of any magmatic system: a glassy sample, representing the quenched liquid matrix, and a crystal rich sample, simulating a rigid crystalline framework. The glassy starting material was prepared within a Mo sheathed alumina crucible by repeated cycles of quick melting (15 min at 1600°C) at atmospheric pressure under controlled Argon atmosphere obtaining a 5 g bead of dry clear glass devoid of bubbles. The crystal rich starting material was set up by keeping the sample within the crucible at super-liquidus conditions (1150°C under controlled Argon atmosphere) for 3 hours, then decreasing the temperature to 950°C in two hours and maintaining the sample at this temperature for 24 hours, before quenching. The bead of crystal rich starting material was finally inspected by optical microscopy in thin section confirming the presence of clear glass and of the assemblage: *Kfs* + *Pl* + *Cpx* + *Bio* + *Sp* + some interstitial glass. Then, samples for the kinetic experiments were produced using a core drill with a 3 mm outer diameter. Cores used in the experiment were 20 mm long and 2.5 mm in diameter.

[10] Experiments were run at constant pressure ($P(H_2O)$ = 100 MPa) and temperature (960°C) for time intervals ranging from 1 to 72 hours (Figure 4). Extra care was used to maintain a controlled P-T environment (pressure uncertainty lower than ±3 MPa and temperature uncertainty lower than ±2°C). The cores were sealed in capsules of gold palladium alloy (Au60/Pd40). A weighted amount of water (in slight excess of the necessary amount to saturate the melt at 100 MPa) was added before charging the capsules. Experimental runs were performed using vertical TZM vessels especially modified for rapid quenching. The sample experimental temperature was measured using an external cromel alumel thermocouple connected to a 0.1% ERO electronic thermo-regulator. Before running the experiments, this thermocouple was calibrated against a thermocouple located in the sample seat. The pressure was measured through a 1% FLUKA analog manometer. The amount of water dissolved in the sample after the experiments was determined by a bulk analysis of the core by a CE CHN-1110 gas chromatographic elemental analyser and checked by ponderal analysis of the leftover excess water. Accuracies in the Elemental Analysis for H_2O gas range from ±10% for samples with 0.1% total H_2O to ±2% for samples with 10% total H_2O content. The precision of the different samples was determined by comparing the measured H_2O peak intensities from three/four repetitions; the Relative Standard Deviations was always less than 3.0%.

[11] A linear diffusion model was fitted to the experimental data to estimate the cores diffusivity D for the solubilised water (Figure 4). The diffusion equation for a problem with radial symmetry (e.g., a cylindrical sample with height h_0 and radius r_0) is

$$\frac{\partial \phi}{\partial t} = D\left[\frac{1}{r}\frac{\partial}{\partial r} r \frac{\partial \phi}{\partial r} + \frac{\partial^2 \phi}{\partial z^2}\right], \quad (1)$$

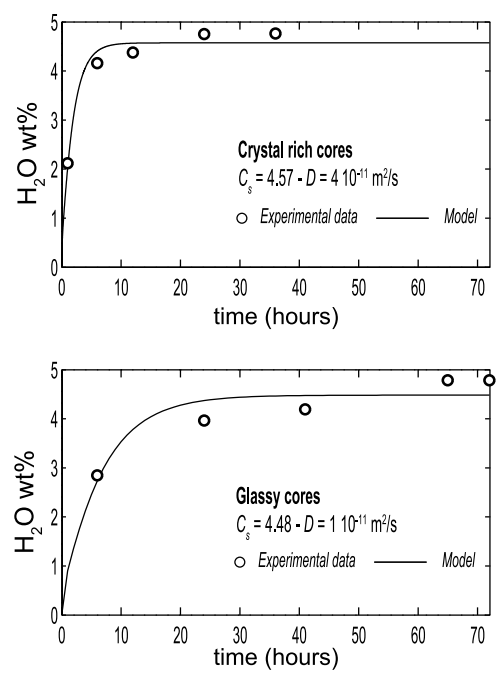

Figure 4. The results of water diffusion experiments performed at the pressure of $P(H_2O)$ = 100 MPa. Note that the hydration times for crystal rich cores are in average one-half of those for the glassy cores at near-solidus temperature. The experimental value of the water saturation content (C_s) is 4.7 ± 0.2 wt%. The linear diffusion model best fit parameters are shown in the legend.

where ϕ is the concentration of water in the sample in dimensions of mass percentage. We can neglect the vertical component in (1) when $r_0/h_0 \ll 1$. For example, if r_0 = 0.00125 m and h_0 = 0.020 m than the vertical component is 0.0625 times the radial component. Keeping this in mind, we can describe the diffusion of water in the cylindrical sample used in the experiments, solving the diffusion equation

$$\frac{\partial \phi}{\partial t} = D\frac{1}{r}\frac{\partial}{\partial r} r \frac{\partial \phi}{\partial r}, \tag{2}$$

subject to the initial condition of dry sample and the boundary condition of no flow at the sample axis of symmetry

$$\phi(r,0) = 0, \qquad \frac{\partial \phi}{\partial r}(0,t) = 0. \tag{3}$$

The solution is

$$\phi = C_s\left\{1 - 2\sum_{n=1}^{\infty}\frac{J_o(\alpha_n r/r_0)}{\alpha_n J_1(\alpha_n)}\exp\left[-\left(\alpha_n^2/r_0^2\right)Dt\right]\right\}, \tag{4}$$

where C_s is the experimental water saturation content (4.7 ± 0.2 wt%), t is the time, r_0 = 1.25 mm is the core radius, α_n is the nth positive zero of the Bessel function J_0 [*Davis and Kirkham*, 1927]. Since we have only measurements of concentration over time, we average (4) over r

$$\phi = C_s\left\{1 - \pi\sum_{n=1}^{\infty}\frac{H_o(\alpha_n)}{\alpha_n}\exp\left[-\left(\alpha_n^2/r_0^2\right)Dt\right]\right\}, \tag{5}$$

where H_0 is the Struve function (see page 496 of *Abramowitz and Stegun* [1972]). The estimated D values of $4 \cdot 10^{-11}$ m^2/s (crystal rich cores) and 10^{-11} m^2/s (glassy cores) are in the range of those reported in the literature for similar igneous compositions [*Brady*, 1995] (Figure 4).

3. Discussion

[12] Our experiments on K-trachyte remelting by water diffusion should be considered as analogue models of magma chamber processes more than scaled models. The transport mechanism is more complex than the diffusive process we are able to reproduce in the laboratory. In a magma chamber, a system of open fractures within the rigid crystalline framework offers the path of least hydraulic resistance, but diffusion from the fractures to the porous matrix would probably constitute an attenuation mechanism that could be highly effective in removing mass from the primary flow channel, thus retarding the advance of the fluids in the system [*Sudicky and Frind*, 1982]. Despite these technical limitations, our experiments still provide valuable insights on the physics of water diffusion in a K-trachyte.

[13] The hydration and remelting for the glassy starting material, documented by a different charge shape, occurs within hours. Water reaches the saturation level, causing complete remelting, within 48 to 60 hours for the glassy cores and in about 24 hours for the crystal rich cores. It is worth noting that the *liquidus* temperature for the anhydrous glassy cores is higher than the experimental temperature being dependent on the amount of water solubilised in the melt at 100 MPa (Figure 4) [*Mysen and Richet*, 2005]. This implies that the glass samples must become hydrated before they can melt. For the crystal rich cores the melting process is associated with the eutectic-like reactions between the diffusing water and the crystalline phases. Fluid diffusion and the cooling of the magmatic system appear to be the competing parameters of the rejuvenation process. For small diffusivities, or large diffusion times, the remelting mechanism proposed has to be replaced by other processes as gas percolation or intrusion of a magmatic mass.

[14] The localization of the gas and/or thermal flux to a limited portion of the near-solidus body could facilitate the intrusion of the gas phase into the roots of the felsic body. This can explain why eruptions of very small size and of suitable composition (e.g.: Monte Nuovo, 1538 AD eruption, Campi Flegrei, Italy) could be originated by the remelting of the crystal mush of a supposedly existing large reservoir.

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