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Crystallization and saturation front propagation in silicic magma chambers

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ABSTRACT

The cooling and crystallization style of silicic magma bodies in the upper crust falls on a continuum between whole-chamber processes of convection, crystal settling, and cumulate formation and interface-driven processes of conduction and crystallization front migration. In the end-member case of vigorous convection and crystal settling, volatile saturation advances downward from the roof and upward from the floor throughout the chamber. In the end-member case of stagnant magma bodies, volatile saturation occurs along an inward propagating front from all sides of the chamber. Ambient thermal gradient primarily controls the propagation rate; warm ($\geqslant$40 °C/km) geothermal gradients lead to thick (1200+ m) crystal mush zones and slow crystallization front propagation. Cold (<40 °C/km) geothermal gradients lead to rapid crystallization front propagation and thin (<1000 m) mush zones. Magma chamber geometry also exerts a first-order control on propagation rates; bodies with high surface to magma volume ratio and large Earth-surface-parallel faces exhibit more rapid propagation and thinner mush zones. Crystallization front propagation occurs at speeds of greater than 10 cm/yr (rhyolitic magma; 1 km thick sill geometry in a 20 °C/km geotherm), far faster than diffusion of volatiles in magma and faster than bubbles can nucleate, grow, and ascend through the chamber. Numerical simulations indicate saturation front propagation is determined primarily by pressure and magma crystallization rate; above certain initial water contents (4.4 wt.% in a dacite) the mobile magma is volatile-rich enough above 10 km depth to always contains a saturation front. Saturation fronts propagate down from the magma chamber roof at lower water contents (3.3 wt.% in a dacite at 5 km depth), creating an upper saturated interface for most common (4–6 wt.%) magma water contents. This upper interface promotes the production of a fluid pocket underneath the apex of the magma chamber. If the fluid pocket grew faster than rates of escape into the wall rock, fluid accumulation and hydro-fracturing could possibly trigger an eruption.

1. Introduction

The dynamics of upper crustal (<10 km deep) magma chambers exert first-order controls on the textural evolution of shallow intrusions and volcanic eruptions. Magma chamber processes were initially viewed as the emplacement of a batch of magma that underwent fractional crystallization (e.g., Bowen, 1928). Over time, models evolved to consider magma chamber processes to be open-system, with recharge events (O'Hara, 1977), dynamic changes in magma properties (Sparks and Huppert, 1984), wall-rock assimilation, and fractional crystallization (DePaolo, 1981). Spera and Bohrson (2004) conceptualized magmatic processes as complex systems of generation, pooling, eruption, and recharge, with varied assimilation. Magma chambers may periodically undergo convection with crystals forming plumes and/or separating from the silicate liquid to form adcumulates and orthocumulates (Shaw, 1965; Bartlett, 1969; Sparks et al., 1984; Worster et al., 1990). Magma bodies may also stagnate, with crystals captured by inward-growing solidification fronts (Marsh, 1989a, 1996). The relative roles of convection, inward solidification, and "mushification" have been widely debated and must vary in time and from system to system (see Brandeis and Marsh, 1989; Marsh, 1989b; Huppert and Turner, 1991; Huber et al., 2009).

Volatile content exerts first-order controls on magma chamber dynamics. The volatile budget of most arc and continental magmas consists dominantly of water with lesser concentrations of CO_2/CH_4, SO_2/H_2S, F_2, and Cl_2. Water content is of particular importance because water depresses solidi and liquidi by as much as 200–300 °C and can increase isothermal partial melt volumes by an order of magnitude (Clemens and Vielzeuf, 1987; Hirschmann et al., 1999). Water-rich rhyolitic melts have up to 10 orders of magnitude lower viscosity than equivalent dry melts (Shaw, 1965).

Water contents of volcanic deposits provide an estimate of H_2O concentration at the time of eruption (and in explosive eruptions at volatile saturation). These vary widely with magma composi-

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tion: primary Hawaiian (ocean island basalts – OIB) and mid-ocean ridge basalts (MORB) typically erupt effusively and contain less than 1 wt.% H_2O, whereas back-arc and subduction related basalts are wetter, containing 1–3 wt.% (based on glass inclusions; Sun et al., 1979; Johnson et al., 1994). Water contents of evolved volcanic arc andesites and dacites from explosive eruptions range from ~3–6.5 wt.% H_2O: Mt. St. Helens Holocene lavas contained 3.5–6.5 wt.% H_2O (based on experimental petrology; Gardner et al., 1995a, 1995b), the 1996 Soufriere Hills eruption contained 4.25 wt.% water (based on experimental petrology; Barclay et al., 1998), and the 1991 Pinatubo magma contained 6–6.5 wt.% water (based on experimental petrology; Scaillet and Evans, 1999). The 5000+ km^3 Fish Canyon Tuff (FCT) dacite also contained ~5 wt.% water at eruption (based on experimental petrology; Johnson and Rutherford, 1989). High-silica rhyolites associated with explosive, caldera-forming eruptions, such as the Bishop Tuff, typically have the highest water contents. The Bishop Tuff has water contents that range from 2–6 wt.% (based on glass inclusions in quartz; Anderson et al., 1989; Wallace et al., 1995). Some rhyolites can reach as high as 8 wt.% (e.g., rhyolites associated with the Pine Grove porphyry; Johnson et al., 1994). These volatile contents are average contents; crystallization can drive volatile contents locally higher and fluid phases can completely segregate from the magma to form fluid pockets. The generation of separate fluid phases in magma chambers and conduits is now recognized as an important, potentially eruption triggering process (Shinohara, 2008).

This paper considers the effects of inward solidification from magma chamber walls (after Marsh, 1989a, 1996) on residual melt water contents and saturation. This paper uses a numerical simulation to model the growth of saturation fronts along magma chamber sidewalls and discusses the implications these fronts can have for magmatic convection, metal concentration, and volcanic eruption.

The goal is to investigate the evolution of homogenous, silicic systems that can produce caldera-forming "super-eruptions" and giant porphyry metal deposits. Debate is ongoing as to whether magma rapidly intrudes the crust (Glazner et al., 2004; Cloos and Housh, 2008; Annen, 2009) to form ephemeral chambers or consolidates over timescales >100 kyr (e.g. Bachmann et al., 2007; Gelman et al., 2013). The batholithic magma chambers of interest form in crust that has thermal gradients of $\geqslant 30\,^{\circ}C/km$, though the effects of colder gradients are also considered. The model considers the fate of these chambers as magma flux wanes to zero (Suppl. Fig. 1). This analysis is not intended to encompass steady-state, long-lived volcanic-arc magmas with regular chamber recharge.

2. The model

This model simulates the cooling, crystallization, and water saturation of a simple mid-crustal intrusion. The numerical model is a two-dimensional (pseudo-3D) finite-difference mesh in cylindrical coordinates extending from the surface to 12 km depth, divided into equally spaced 20 m cells. Each cell retains information about pressure, temperature, magma state, crystallinity and water content. The magma body is initially intruded at its liquidus into wall rock at a constant geothermal gradient. Liquidus temperatures range from 840–865 °C for rhyolites and 1050–1060 °C for dacites (Table 1); simulation initial temperatures are 865 and 1050 °C respectively. The magma loses heat conductively according to Eq. (1) (adapted for cylindrical coordinates with rotational symmetry from Eq. (1) in Annen and Sparks, 2002):

$$\rho Cp\frac{\partial T}{\partial t}+\frac{\partial X}{\partial t}\rho L=\frac{\partial}{\partial r}\left(k\times r\frac{\partial T}{\partial r}\right)+\frac{\partial}{\partial z}\left(k\frac{\partial T}{\partial z}\right) \quad (1)$$

where T is temperature, t is time, r is the radial (horizontal) and z is the vertical dimension, ρ is density, Cp is heat capacity, X is melt fraction, L is latent heat of fusion, and k is thermal conductivity. Thermal properties have been chosen conservatively to maximize magma longevity. A 25 yr time-step is used – stability is guaranteed by use of alternating-direction Crank–Nicolson approximations (Crank and Nicolson, 1947). The finite-difference approximation and code are adapted from those developed by Ketcham and Carlson (2012; Appendix 1) for chemical diffusion. Thermal gradient, initial water content, chamber geometry, and magma composition are all variables. The top surface of the mesh is held constant at 0 °C. The bottom surface of the mesh is held constant according the initial temperature at 12 km. The right-hand boundary is no-flux; the model is rotationally symmetric around the left-hand boundary (see Fig. 1 for a representation of initial conditions).

Table 1
Model thermal and compositional parameters.

Variable	Value	Units
Magma heat capacity (Cp)	1600[a]	J/kg K
Magma density (ρ)	2250[a]	kg/m^3
Magma thermal conductivity (k)	1[a]	W/m K
Dacite latent heat of fusion (L)	270,000[b]	J/kg
Rhyolite latent heat of fusion (L)	35,000	J/kg
Crustal heat capacity (Cp)	950[c]	J/kg K
Crustal density (ρ)	2700[d]	kg/m^3
Crustal thermal conductivity (k)	2.75[d]	W/m K
Fit parameter a	2565	dimn.
Fit parameter b_{Al}	−1.997	dimn.
Fit parameter b_{Fe}	−0.9275	dimn.
Fit parameter b_{Na}	2.763	dimn.
Fit parameter c	1.171	dimn.
Fit parameter d	−14.21	dimn.

Oxide (wt.%)	Fish canyon tuff dacite[e]	Bishop tuff rhyolite[f]
SiO_2	67.91	75.5
TiO_2	0.46	0.21
Al_2O_3	15.44	13
FeOt	3.45	1.1
MgO	1.08	0.25
CaO	3.015	0.02
MnO	0.04	0.95
K_2O	4.02	5.55
Na_2O	3.785	3.35
P_2O_5	0.195	0.06
Liquidus (°C)	1050–1060	840–865
Solidus (°C)	685–740	715–790

Fit parameters from Moore et al. (1998). Liquidus and solidus from MELTS. Other sources are as follows:

[a] Spera (2000); [b] Annen and Sparks (2002); Gualda et al. (2012b); [c] Vosteen and Schellschmidt (2003); [d] Turcotte and Schubert (2002); [e] Bachmann et al. (2002); [f] Hildreth (1979).

Magma crystallinity is an evolving function of pressure, temperature, chemical composition, oxygen fugacity, and water content. The model uses the MELTS (Ghiorso and Sack, 1995; Asimow and Ghiorso, 1998) and Rhyolite-MELTS (Gualda et al., 2012a) thermodynamic databases to calculate crystallinity for a dacitic and rhyolitic composition between 0 and 3 kbar at 500 bar intervals, with intermediate values linearly interpolated (Suppl. Fig. 2a, b). The Fish Canyon dacite was modeled with MELTS Ubuntu 5.0 32-bit using the chemical composition listed in Table 1, 2 wt.% initial H_2O, and an oxygen buffer fixed at QFM + 3. Initial oxygen fugacity of the Fish Canyon magma is unknown, however Johnson and Rutherford (1989) report Fe–Ti oxide pairs yield log $fO_2 \approx -12.4$ at the eruption temperature of ~760 °C, hence the choice of the highly oxygenated buffer. The Bishop rhyolite was modeled with Rhyolite-MELTS Ubuntu 1.0.1 32-bit using the chemical composition listed in Table 1, 3 wt.% initial H_2O, and an oxygen buffer fixed at NNO. Fe–Ti oxide pairs indicate that the Bishop magma