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Thermo-mechanical reactivation of locked crystal mushes: Melting-induced internal fracturing and assimilation processes in magmas

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ABSTRACT

Thermal reactivation of locked crystal mushes in the upper crust is a fundamental step towards volcanic eruptions of crystal-rich magmas. Models of such reactivation events indicate that partial melting of the crystalline framework is energetically costly and lead to average crystallinities that are lower than those observed in many erupted crystal mushes. Here, we show that internal overpressurization of the mush induced by small amounts of melting (10–20%) breaks the crystalline framework by microfracturing and allows for efficient unlocking of the mush. Hence, this melting-induced overpressurization, enhanced by addition of gas in wet magmatic systems, plays an important role in generating volcanic deposits with crystal contents close to the rheological lock-up (~50 vol% crystals) by accelerating the incorporation of highly crystalline parts of the magma chamber (self-assimilation). It can also participate in disintegrating pieces of country rock that are commonly scavenged in magmas, leading to bulk assimilation of crustal lithologies in shallow reservoirs.

1. Introduction

Crystal-rich volcanic deposits with obvious textural and geochemical evidence for pre-eruptive reheating imply that thermal reactivation is an important process in triggering the eruptions of shallow crustal magma reservoirs. The observed resorption textures and/or significant chemical zoning displayed by multiple mineral phases in a number of deposits are commonly interpreted as the result of a reheating event and unlocking of a crystal mush associated with the underplating intrusions of more mafic magma (Bachmann et al., 2002; Bachmann et al., 2007; Couch et al., 2001; Koyaguchi and Kaneko, 2001; Snyder, 2000). Some of the largest silicic eruptions preserved in the rock record, the Monotonous Intermediates (large dacitic ignimbrites with homogeneous whole-rock compositions); (Hildreth, 1981), provide formidable examples of such reactivated crystal mushes.

The main hypothesis for the reactivation of crystal-rich reservoirs (i.e. their transition from rheologically locked-up to sluggishly convecting magma body), is partial melting of the mineral assemblage associated with the emplacement of new magma intrusion at or near the base of the mush (e.g. (Bachmann et al., 2002; Bachmann et al., 2007; Couch et al., 2001; Murphy et al., 2000; Nakamura, 1995; Pallister et al., 1992)). Substantial amounts of enthalpy are required for the reactivation (Huber et al., 2010a; Huber et al., 2010b), and, for large systems, a single recharge event is generally not sufficient (the mafic intrusion has to be comparable in size with the mush). Moreover, the inefficient transfer of enthalpy between the intrusion and the mush fraction subjected to partial melting makes this process energetically unfavorable (Dufek and Bergantz, 2005; Huber et al., 2010b). (Huber et al., 2010b) also showed that the typical average crystallinities of erupted mushes are higher than expected from simple thermal models (i.e. melting models tend to overestimate the amount of melting required to reactivate a crystal mush), as conduction and convection models predict a large crystallinity reduction in the lowermost part of the mush when the melting/reactivation front reaches the top of the magma body. Therefore, we hypothesize that energy provided by the intrusion is not uniquely thermal. Exsolved volatiles injected by the intrusion and the volume change associated with the solid-melt phase transition in the mush can both provide some potential mechanical energy (overpressure).

In this study, we use a 1D multiphase thermal model coupled to a mechanical model to test the effect of the in-situ overpressurization associated with melting and the injection of volatiles on the reactivation of crystal mushes. As the reactivated part of the mush reaches a critical melting-induced overpressure (between 10^6 and 10^7 Pa; set here as 10^7 Pa; (Rubin, 1998)), the overlying mush is subjected to internal fracturing, the extent of which depends on the efficiency of the fracturing event to relieve the excess pressure. The style of fracturing is expected to

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be a large number of microfractures rather than dikes, as (1) the stress state at the scale sampled by large dikes is overall compressive (due mostly to melting) and (2) the presence of a low viscosity compressible phase (exsolved volatiles) decreases the efficiency of the stress transmitted to the fracture tip. These fracturing episodes are expected to loosen up a small volume fraction of the overlying locked-up mush and, consequently, increase the volume of magma opened up to wholesale convection. The main advantages of these fracturing episodes are (1) it increases the energy efficiency required to reactivate crystal mushes, as melting provides most of the overpressure, and (2) it increases the average crystallinity of the these systems once reactivated. Both are required to better fit natural observations (Huber et al., 2010b).

This mechanical process of fracturing associated with melting applies to mush reactivation as a self-assimilation process ("defrosting" the overlying locked mush, observed both in volcanic; e.g., (Bachmann et al., 2002; Couch et al., 2001; Mahood, 1990) and plutonic sequences; e.g., (Paterson and Janousek, 2008; Robinson and Miller, 1999; Wiebe et al., 2007)), but can also play a role in assimilating country rocks in the case of magma evolution by AFC (Assimilation Fractional Crystallization; numerous papers, but see e.g., (Bowen, 1928; Daly, 1933; DePaolo, 1981; Taylor, 1980)). As blocks of crystalline wall rocks are incorporated into hot magmas, the mechanism of melting-driven fracturing will lead to enhanced disaggregation, allowing for rapid dissemination of small xenoliths and xenocrysts in the magma.

In the next section, we present the physical model used in this paper, describing shortly the thermal model (based on (Huber et al., 2010a)) and introducing the mechanical model (based and modified from (Huppert and Woods, 2002)). We also discuss the expected fracturing mechanism of the mush (large number of microcracks rather than one or a few large dikes) once a critical overpressure is reached. Finally, we present results from our numerical calculations and discuss the implications in the scope of the reactivation of large crystal-rich magma bodies.

2. Physical model

This section describes the model used to solve for the thermal and mechanical evolution of a silicic crystalline mush underplated by an intrusion of more mafic magma. We start by discussing the multiphase heat transfer between the intrusion and the mush and proceed next to a simple mechanical model based on mass conservation and linear elasticity to describe the evolution of the pressure during the reactivation of the mush.

2.1. Thermal model

The thermal model is similar to (Huber et al., 2010a) (illustrated schematically in Fig. 1). For simplicity, the intrusion is emplaced at once and the volume ratio between the mush and the intrusion is a free parameter in our calculations. The composition of the intrusion is set to andesitic and the mush to dacitic. The mush is assumed to be initially saturated with volatiles and we neglect volatile dissolution or exsolution associated with reheating and partial melting of the mush. We also assume that, by the time the intrusion reaches its emplacement depth (2×10^8 Pa), its volatile fraction mostly consists of water as a large fraction of the CO_2 already degassed and escaped towards the surface. The rate of cooling, crystallization and volatile exsolution in the intrusion is controlled by the heat transfer in the mush.

2.1.1. Underplating magma

We use a simplified relationship to describe the heat balance for the underplating intrusion

$$\frac{dT}{dt} = -\underbrace{\frac{q_{out}}{c_i\rho_i H_i}}_{\text{sensible heat}} + \underbrace{\frac{L_i}{c_i}\frac{\partial\chi_i}{\partial t}}_{\text{latent heat}}, \quad (1)$$

where the subscript i refers to the intrusion, H_i, L_i, c_i, ρ_i and χ_i are respectively the thickness, the latent heat of crystallization, the specific heat, the density and the crystallinity of the underplating magma body. q_{out} is the heat transfer from the intrusion to the mush and therefore couples the magma body in terms of heat transfer. This heat balance equation assumes a well-mixed intruding magma with negligible thermal gradients which is a relatively good approximation when a fluid body with strong temperature dependent viscosity convects. We use stagnant-lid convection scalings to calculate the convective heat flux out of the intrusion using a temperature and crystallinity-dependent viscosity (Huber et al., 2010a). The crystallinity–temperature relationship for an andesitic intrusion (likely composition for recharge in continental arcs) is calculated with MELTS (Ghiorso and Sack, 1995) using the major element composition listed in (Parat et al., 2008) for the Huerto andesite (San Juan Volcanic Field, Colorado). We assume an initial water content of 6 wt.% H_2O. We also use MELTS to parameterize the exsolution of volatiles (water only in this case, see section below) as function of temperature (see (Huber et al., 2010a)).

We assume that the transport of exsolved volatiles from the intrusion to the mush can be described by a multiphase Darcy equation

$$q_{H_2O} = \frac{kk_r\Delta\rho g}{\mu_g}, \quad (2)$$

where k and k_r are the permeability of the magma and the relative permeability for the volatile phase, respectively, $\Delta\rho$ is the density contrast between the melt and the volatile phase and g is the acceleration due to gravity. The permeability is related to the crystallinity by the Carman–Kozeny relation (Bear, 1988)

$$k = A\frac{(1-\chi_i)^3}{(1-(1-\chi_i))^2}, \quad (3)$$

where the constant A is set to 2×10^{-12} m^2, leading to a permeability $k=10^{-12}$ m^2 at $\chi_i=0.5$. Many empirical and theoretical expressions have been derived for the dependence of the relative permeability on the volatile volume fraction, we use the following expression

$$k_r\left(S_g\right) = S_g^4, \quad (4)$$

where S_g is the volume fraction of exsolved volatiles. For a more detailed description of the volatile transport model, the reader is referred to (Huber et al., 2010a).

2.1.2. Crystal mush

The mush above the intrusion has initially a crystallinity ranging from 51 to 67%. We solve for the mass conservation of volatiles (initially exsolved from the intrusion)

$$\frac{\partial\left(\rho_g(1-\chi_{\text{mush}})S_g\right)}{\partial t} = -\frac{\partial\left(\rho_g u_{d,z}\right)}{\partial z}, \quad (5)$$

where $\rho_g=\rho_{H_2O}$, χ_{mush} is the crystallinity of the mush, S_g is the pore volume fraction occupied by the volatiles and $u_{d,z}$ is the z-component of the Darcy velocity for the volatile phase and z is parallel to gravity (increases upwards). The boundary condition at the interface between the mush and the intrusion requires matching the volatile flux (see Eq. (2)). We assume here that the melt is volatile-saturated and that no further degassing or dissolution of volatiles occurs within the mush. The flux of volatiles is obtained from a multiphase Darcy equation

$$u_{d,z} = -\frac{k_{\text{mush}}k_{\text{rmush}}}{\mu_{H_2O}}\frac{\partial}{\partial z}\left[\left(\rho_{H_2O}(z)-\rho_{\text{melt}}\right)gz\right] \quad (6)$$