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Modeling magmatic accumulations in the upper crust: Metamorphic implications for the country rock

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ABSTRACT

Field exposures of magma chambers tend to reveal contact metamorphic aureoles in the surrounding crust, which width varies from few centimeters to kilometers. The igneous accumulation not only increases the temperature around it, but also weakens its surrounding country rock beyond the brittle–ductile transition temperature. The formation of a ductile halo around the magmatic reservoir may significantly impact into the stability and growth of the magma chamber, as well as into potential dyke injections and processes of ground deformation. In this paper, we examine how a magmatic accumulation affects the country rock through the combination of petrologic and thermal perspectives. For this, we numerically modeled (i) the conductive cooling of an instantaneously emplaced magma chamber within compositionally representative pelitic and carbonate upper crusts, and (ii) the corresponding changes in the viscosity of the host rock potentially leading to ductile regimes. We consider basaltic to rhyolitic magma chambers at different depths with oblate, prolate and spherical geometries. The resulting temperature field distribution at different time steps is integrated with crustal metamorphic effects through phase diagram modeling. Our results indicate that the geometry of the magma accumulations plays a dominant role in controlling the local metamorphic and thermal effects on the country rocks. They conclude that (i) the combination of relatively simple geothermal models with petrologic datasets can generate first order predictions for the maximum metamorphic grade and geometry of magma chamber aureoles; (ii) the possible changes in the mechanical properties of the country rock are not necessarily linked to the petrological changes in contact aureoles; and (iii) the present rheologic outcomes may be used in further studies of magma chamber stability and integrity, which may favor the understanding of the melt transfer throughout the crust.

1. Introduction

The past few decades have been marked by an important investigation into quantifying the thermal effects of igneous accumulations on the surrounding host rock (e.g. Lovering, 1935, 1936, 1955; Jaeger, 1959; Carrigan, 1983, 1988; Cui et al., 2001; Wang et al., 2012; Pla and Álvarez-Valero, 2015). This effort is crucial to understanding and evaluating related phenomena, such as contact metamorphism processes (e.g. Cook and Bowman, 1994; Cui et al., 2001; Annen et al., 2006; Bufe et al., 2014) and the maturation of organic matter in the vicinity of magma reservoirs (e.g. Galushkin, 1997; Fjeldskaar et al., 2008). The correlation between the results of mathematical models and geothermometers, such as vitrinite reflectance and fluid inclusions (Galushkin, 1997; Wang, 2012 and references therein), has widely demonstrated the capability of these studies to reconstruct the thermal evolution of host rocks after an igneous accumulation.

Numerous petrological observations and numerical simulations indicate that crustal aureoles due to contact metamorphism encompass a variety of widths, ranging from few centimeters to some kilometers (e.g. Bowers et al., 1990; Johnson et al., 2011; Peacock and Spear, 2013; Bufe et al., 2014). The extent and degree of the generated aureole depend on many parameters, including: temperature, composition and thermal properties (i.e. thermal conductivity, specific heat capacity) of the magma and country rock; the degree of crystallization of the magma; the size and geometry of the reservoir; the latent heat of fusion for the crystallizing magma; and the amount of heat energy consumed by endothermic metamorphic reactions taking place in the aureole (Johnson et al., 2011; Peacock and Spear, 2013). Additionally, at shallow crustal levels, hydrothermal processes may play an important role in incorporating cations and H_2O into the system, as well as redistributing the heat released from the cooling magma reservoir that can affect contact metamorphism equilibria (e.g. Peacock and Spear, 2013).

In parallel to the work focused on the mineralogical and petrological consequences of igneous accumulations, several studies have attempted to elucidate and quantify their effects on the country rock. For instance,

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the mathematical models by Dragoni and Magnanensi (1989); Jellinek and De Paolo (2003); Del Negro et al. (2009) and Gregg et al. (2013) revealed changes in the host rock's rheology due to the thermal effects of the cooling magma reservoir. Other authors indicated that these changes in the country rock rheology may influence magma chamber growth, rupture (i.e. dike intrusion), or even caldera collapse events (Jellinek and De Paolo, 2003; Gregg et al., 2012, 2013; Gottsmann and Odbert, 2014). Indeed, recent works by Gregg et al. (2012, 2013) define the thermal alteration of the host rock to be a key parameter controlling the dynamics and genesis of caldera-forming eruptions of giant magmatic reservoirs. The temperature increase of the country rock translates into a considerable decrease of its viscosity and Young's modulus, causing host rock rheology to switch from brittle to ductile (Gregg et al., 2012). Depending on the size and shape of the magma reservoir, as well as on new magma influx, the "ductile aureole" formed around the magma chamber then inhibits potential dike injections into the surrounding rock, creating positive feedback in the magma accumulation process by increasing the chamber's stability and growth rate (Jellinek and De Paolo, 2003; Gregg et al., 2012, 2013). Hence, if no dykes are injected, when more magma arrives into the reservoir, its volume increases, along with corresponding thermal effects on the country rock. This process favors the survival and growth of the thermal aureole.

Traditionally, the thermal alteration of both the mechanical properties and petrological features of the host rock due to magmatic accumulations has been treated separately (e.g. Dragoni and Magnanensi, 1989; Cui et al., 2001; Gottsmann and Odbert, 2014). However, assumptions of the geometries, temperature distributions, and rheology in the plumbing system of the magma chamber require input from petrological studies. Likewise, the complete interpretation of any petrological dataset requires key numerical parameters constrained by magma chamber rheology (e.g. Álvarez-Valero et al., 2015).

The final objective of the present research is to understand the metamorphic signatures of different magmatic reservoirs in the host rock and to examine whether or not these are related to particular changes in its mechanical properties. We aim to link the petrological changes observable in the field—as mineral equilibria of contact metamorphism—to possible variations in the country rock viscosity, and vice versa. Does the country rock always undergo a potential transition in its rheology (from brittle to ductile)? If so, is this systematically associated with certain petrological changes?

For this, we carry out a series of heat transfer models of the conductive cooling of different instantaneously emplaced magma chambers in three representative pelitic and carbonate upper crusts. For the cooling models, we solve the conductive heat transfer equation using the Finite Element Method (FEM) in an axisymmetric medium. We examine magma chambers of rhyolitic, andesitic, and basaltic composition in oblate, prolate, and spherical reservoirs. Using the modeled temperature field distribution in the host rock through time, we estimate its change in viscosity due to the temperature increase. Finally, we predict the equilibrium mineral assemblage by overlaying the corresponding phases from pressure–temperature (P–T) pseudosections of the representative crustal bulk compositions.

2. Methodology

2.1. Heat transfer models

The temperature distribution within the magma chamber and the host rock is calculated using FEM, by solving the pure conductive heat transfer equation, and assuming the effects of viscous heating and pressure-volume work are negligible (see Rodríguez et al., 2015 for further details):

$$\rho Cp\frac{\partial T}{\partial t} + \nabla(-k\nabla T) = Q \tag{1}$$

where the equation parameters refer to density (ρ), specific heat capacity at constant pressure (Cp), temperature (T), time (t), thermal conductivity (k) and Q denotes heat sources other than viscous heating (see Table 1 for details of the thermal and physical parameters).

We consider the modeled magma chamber to be emplaced immediately before it begins cooling. The completely liquefied magma reservoir (Gelman et al., 2013) is assumed to intrude in a single episode without further replenishment (Gutiérrez and Parada, 2010). The geometric modeling, mesh discretization and numerical computations are carried out with COMSOL Multiphysics v4.4 software package (http://www.comsol.com). To simulate the solidifying magma, we use the *heat transfer with phase change* module, which allows us to solve Eq. (1) after setting the properties of a phase-change material (from liquid to solid) according to the Apparent Heat Capacity formulation (AHC) (see Rodríguez et al., 2015 and references therein). The AHC formulation is considered to be the best representation of a naturally occurring wide phase-change temperature interval, as occurs during magma cooling (Rodríguez et al., 2015). The latent heat of crystallization is accounted for by increasing the heat capacity of the material within the phase change temperature range. For further details on the methodology, the reader is referred to the work by Rodríguez et al. (2015). A comprehensive report detailing one of the models, which includes all settings within COMSOL (e.g. model properties; physics settings; geometry; mesh), as well as an example of one of the models, are available from the corresponding author upon request (COMSOL Multiphysics commercial software V4.4 or greater and *heat transfer with phase change* module are required).

The performed FE models are axisymmetric and constructed over a cylindrical coordinate system with positive z values related to altitudes above sea level (Fig. 1a). The magma chamber geometry is oblate, spherical or prolate in shape with height h and width w (Fig. 1a). The computational domain corresponds to a section of the crust with a 40 km radius stretching to a depth of 40 km below sea level (Fig. 1a). The FE mesh consists of between 200,000 and 400,000 linear triangular elements up to 250 m in size farther from the magma reservoir, and 2 m in size near the edge of the magma chamber (Fig. 1b). Time steps taken to perform the calculations range from less than a year at the beginning of the cooling processes, up to 300 years to speed calculations once reached the metamorphic peak. We have tested that the chosen time steps do not affect the results obtained.

The selected starting magmatic compositions for the numerical simulations correspond to three representative types that range from alkaline basalt to rhyolite (Table 2). The former, typical of intraplate magmatism, is taken from Polat (2009). The intermediate magma is of andesitic composition and from a subduction zone taken from Yogodzinski et al. (1994). The rhyolitic composition is from an archetypal case of pelitic crustal melting taken from Sensarma et al. (2004). Depending on the model, pressure at the magma chamber center ranges from 1 to 3 kbar (Supplementary material, Table S1).

The melt (θ) and solid (ϕ) fractions, as well as the thermal properties of the crystallizing magmas, are determined using the Rhyolite-MELTS code (Gualda et al., 2012). Results are reported in the Supplementary material (Table S2–S7) and are used as input data for the models. We simulate isobaric cooling at 1, 2 and 3 kbar pressure, with redox conditions one log unit above the quartz–fayalite–magnetite (QFM) oxygen buffer.

Using the thermal properties provided by MELTS, we explicitly account for the temperature dependence of thermal diffusivity (κ) and heat capacity (Cp). Incorporating a temperature-dependent diffusivity is critical due to its strong influence on the temperature-dependence of thermal conductivity ($k = \rho \cdot Cp \cdot \kappa$) at high temperatures (Nabelek et al., 2012). Average magma density ρ_{magma} values obtained from MELTS are 2344 kg/m^3, 2440 kg/m^3 and 2694 kg/m^3 for the rhyolitic, andesitic and basaltic magmas, respectively. For most models (Supplementary material Table S1), we consider a crustal density of 2700 kg/m^3 (Whittington et al., 2009), a temperature-dependent