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Crustal Magmatic Systems from the Perspective of Heat Transfer

Jon Blundy¹ and Catherine Annen

ABSTRACT

Crustal magmatic systems are giant heat engines, fed from below by pulses of hot magma, and depleted by loss of heat to their surroundings via conduction or convection. Heat loss drives crystallization and degassing, which change the physical state of the system from low-viscosity, eruptible melt, to high-viscosity, immobile, partially molten rock. We explore the temporal evolution of incrementally grown, magmatic systems using numerical models of heat transfer. We show that their physical characteristics depend on magma emplacement rates, and that the majority of a magma system's lifetime is spent in a highly crystalline state. We speculate about what we can, and cannot, learn about magmatic systems from their volcanic output.

Keywords: magma; intrusion; granite; eruption; zircon

INTRODUCTION

Heat is the amount of energy that is transferred spontaneously from a hotter body to a colder body. Not surprisingly, in the land of magma, heat is king. A volume of magma contains sensible heat, defined as the heat capacity multiplied by the temperature difference between the magma and its surroundings, and latent heat, which is that released upon crystallization (TABLE 1). Intrusion of magmas into the crust advects heat from depth; this heat is then lost to the crust, and eventually to the Earth's surface, driving crystallization of the magma, which in turn controls its rheology and potential eruptability, and hydrothermal circulation. To understand the thermal evolution of crustal magma bodies is, therefore, central to learning how magmatic systems work. The heat balance is controlled primarily by the rate at which heat is transported (advected) into the crust by buoyant, hot magma generated deeper in the crust or in the mantle. Here we use the terms *magma flux* and *heat flux* for the volumes of magma and the amount of heat, respectively, that enter the system per unit of time. On Earth, advective heat flux due to magma rise is spatially and temporally variable. In this paper, we use simple thermal models to quantify how variations in heat flux control the thermal structure of crustal magmatic systems, and explore the consequences for their chemical and

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physical evolution. Simple models cannot capture all of the complexity present in nature, but, as we show, they afford some surprisingly informative insights.

MAGMA CHAMBERS AND PARTIALLY MOLTEN RESERVOIRS

An enduring concept in igneous geology is the magma chamber, a large shallow-crustal vat of hot, molten rock that can fuel volcanism above, solidify in situ to create a pluton, or both. The magma chamber concept lends itself to the notion, and indeed requires, that large volumes of magma, say several tens of km³, can be emplaced instantaneously into the crust. Instantaneous emplacement on this scale raises a number of mechanical challenges, especially in the shallow crust (Menand 2008). Indeed, there is overwhelming geological and mechanical evidence that magma emplacement into the crust is piecemeal, via a series of dikes or sills emplaced sequentially, rather than in one giant event (de Saint-Blanquat et al. 2011). Evidence includes protracted magmatic histories of plutons, as revealed by zircon geochronology (Coleman et al. 2004; Lipman 2007), internal contacts between successive intrusive pulses in plutons (e.g. John and Blundy 1993; McNulty et al. 1996), and thermal modelling (Glazner et al. 2004; Annen et al. 2008, 2015). Some field examples, such as the Torres del Paine intrusion, Chile (FIG. 1), illustrate beautifully the concept of sequential emplacement, with granitic magma emplaced over 90 ky as a series of thin sheets totaling ~80 km³ (Leuthold et al. 2012).

Seminal, early numerical models of magma bodies (e.g. Bartlett 1969; Spera 1980) focused on the thermal evolution of large molten vats. Recent numerical models of magmatic systems (e.g. Annen et al. 2008, 2015) have embraced instead the piecemeal emplacement concept to build magma bodies with characteristics similar to Torres del Paine. Piecemeal emplacement differs significantly from instantaneous emplacement in terms of transient heat balances and, consequently, the amount of the system that is molten at any one time. In the most extreme case of piecemeal emplacement the mass and heat flux are so low that each pulse cools and solidifies completely before the next one arrives. The total volume of melt at any time is then very small and the system is unable to support large eruptions at all. At low magma fluxes, therefore, the system evolves into a pluton. At higher magmatic fluxes, the proportion of sustained melt is larger and such systems have the potential for eruption (Annen et al. 2015). Establishing the critical values of magmatic fluxes that lead to plutonism from those that lead to volcanism is a central question in igneous geology.

In natural silicate systems the interval between the liquidus (fully molten) and solidus (fully solidified) is rich in complexity. In this melting interval the physical state of the system is related to the phase relations of magma, that is, how the crystal fraction varies with

temperature, or, alternatively, the relationship between heat loss and crystal content. The latter is dictated by the latent heats of crystallization, and the specific heat capacity and thermal conductivity of the magma and its country rocks, through which heat is lost. Definitions of these various parameters, values used for modelling crustal magmatic systems, and their relevant units, are given in TABLE 1.

NUMERICAL MODELLING OF MAGMATIC SYSTEMS

Numerical models of incremental magma emplacement employ a discretized system of cells between which heat is balanced using finite difference methods (Annen et al. 2008). The system can be one-, two- or three-dimensional, with complexity and computational requirements increasing accordingly. The models presented here (FIG. 2A) are 2-D, but have rotational symmetry along a vertical axis, conferring pseudo three-dimensionality. Each magma increment has the form of a cylindrical disc, emplaced beneath all previous emplacements (FIG. 2B). It is equally possible to “over-accrete” new additions or emplace them at random depths, but under-accretion is more typical when the magma is silicic (Leuthold et al. 2012).

Magmatic Parameters

Magma composition determines its phase relations and physical properties. Phase relations can be determined experimentally or calculated using thermodynamic software, such as Rhyolite-MELTS (Gualda et al. 2012). Parameterization of experimental data is preferred to thermodynamic calculations, because of difficulties in correctly calculating the stability and composition of hydrous minerals, such as amphibole or mica. However, thermodynamic models have the advantage that they can be adapted to small changes in, for example, pressure or composition.

Crustal silicic magmas are almost always volatile-bearing, if not volatile-rich. H₂O is the dominant magmatic volatile, and the only one considered here. Dissolved H₂O influences a number of important magmatic properties. It reduces melt density and viscosity; reduces liquidus and solidus temperatures; and exsolves during crystallization. This exsolution process, sometimes referred to as “second boiling”, results from the crystallization of solids containing less volatiles than the parent melt, enriching it further in the volatile constituent. The release of H₂O from solidifying magmatic systems is important in the development of hydrothermal ore deposits and geothermal systems. Crustal silicic magmas typically contain ≥4 wt% H₂O, as inferred from analysis of melt inclusions (Blundy and Cashman 2008). We will use a value of 6 wt% H₂O (TABLE 1) corresponding to saturation at ~200 MPa pressure

(~7 km depth). We track the amount of H₂O released from the system as the magma cools and crystallises.

Our calculations use the parameterized, experimentally-determined phase relations of Caricchi and Blundy (2015) for dacite from Fish Canyon, Colorado (Bachmann et al. 2002), a typical composition for subduction-related silicic magmas. The liquidus and solidus temperatures, and an expression relating crystal fraction to temperature are given in TABLE 1, along with calculated values for melt and solid densities. Magma viscosity increases exponentially with crystal content (Caricchi et al. 2007) and exerts an important control on whether magma is sufficiently fluid (crystal-poor) to erupt. We adopt the simplifying assumption that only magma with $\geq 40\%$ melt by volume (melt fraction, F , ≥ 0.4) is eruptible, consistent with the typical phenocryst content of erupted, crystal-rich dacites (Caricchi and Blundy 2015).

To relate the model results to a measurable geochronological parameter, we calculate the temperature at which zircon becomes saturated in the melt. Zircon is widely used as a geochronometer, due to incorporation of radioactive Th and U into its lattice at the time of crystallization. Intrusive and extrusive silicic magmas reveal spectra of zircon ages that can be used to constrain the longevity of the magmatic systems from which they grew (Charlier et al. 2005; Costa 2008; Caricchi et al. 2014). In our system zircon saturates at $F = 0.45$ at which point the radiometric clock starts. If that parcel of magma becomes heated above $F = 0.45$ zircon may dissolve back into the melt, with an efficiency that depends on kinetics (Harrison and Watson 1983). We consider the two extremes of complete dissolution and no dissolution.

INSIGHTS FROM MODELS

For illustrative purposes our models focus on two systems of different size, and on two end-member scenarios with high and low advective heat fluxes. System radius is either 2.5 km (*small*) or 10 km (*large*). Simulations involve incremental emplacement of 200 m-thick sills at their liquidus temperature and separated by time intervals of 2 ky or 8 ky, corresponding to emplacement rates of 0.1 (*high flux*) and 0.025 m/y (*low flux*) respectively. In nature, sills will be smaller and emplaced more frequently, but numerical models show that the parameter controlling the long-term thermal evolution of the system is the average emplacement rate rather than the exact values of thickness and time interval. A rate of 0.025 m/y is close to the minimum emplacement rate that allows a magma chamber to form (Annen 2011) and corresponds to surface deformation rates measured in some volcanic areas (e.g. Pritchard and Simons 2002). A rate of 0.1 m/y is presented for illustrative

purposes; it is not clear if such a high emplacement rate is realistic. The first sill is intruded at 5 km depth and the system grows downwards until a total thickness of 5 km of magma has been intruded. The system is then left to cool until completely solidified.

FIGURE 2C shows a snapshot of the high-flux, small system at the time emplacement ceases. The snapshot is contoured for temperature and melt fraction, both of which have a dome-like shape due to emplacement by under-accretion with the last, hottest sill at the base of the system. The volume of potentially eruptible magma lies within the core of the dome, bounded by the $F = 0.4$ contour. For different heat fluxes and sizes the temperature structure varies slightly, but the dome-like shape persists. Lower heat fluxes result in a much thinner volume of eruptible magma.

Physical State of the System

In FIGURE 3A–C we show the physical state of the magmatic system for three of the scenarios in terms of: solid rock ($F = 0$); non-eruptible, highly crystalline state ($F < 0.4$); eruptible magma containing zircon crystals ($0.4 \leq F \leq 0.45$); and eruptible, zircon-free magma ($F > 0.45$). All three systems spend most of their lifetime in a predominantly crystalline state. The potentially eruptible volume is almost always less than the overall volume of magma intruded, attaining its maximum value when intrusion ceases. The greatest proportion of eruptible magma (~60%) is achieved for the high-flux, large system; the high-flux small system reaches ~30%; the low-flux, large system just 20%. A fourth scenario of low-flux small systems (not shown) never accumulates eruptible magma except for brief periods immediately following each intrusion. Calculations with a range of system dimensions and crustal depths indicate that a minimum emplacement rate of 0.01 to 0.02 m/y is required to sustain eruptible magma between intrusive pulses (Annen et al. 2015). Below these values all systems are destined to become plutons.

It is heat loss through the roof and walls that limits the development of eruptible magma volumes and ultimately controls the ability of magmatic systems to sustain contemporaneous volcanism. If the system is more deeply emplaced, then heat loss to the surroundings is diminished and larger volumes of eruptible magma accrue. Melting of country rocks is limited to deeper systems and/or unusually fusible country rocks, consistent with the relatively rare observation of partial melting in the aureoles of granitic plutons (e.g. Floess and Baumgartner 2015). The temperature structure in the country rocks, which controls the development of the thermal aureole (e.g. FIG. 2B), depends on emplacement geometry and flux (Annen 2011; Floess and Baumgartner 2015).

Our modelled magma bodies cool exclusively by conduction. If a convective system is established at the roof, for example by heating of groundwater, then heat loss is increased and the volume of eruptible magma at any time is less than for the purely conductive case. Nonetheless, the abiding impression is the difficulty of sustaining large, long-lived volumes of potentially eruptible silicic magma in the shallow crust without high advective heat fluxes. For example, the total volume of eruptible melt in the high-flux, large system (800 km³; FIG. 3B) could support a very large eruption, whereas smaller systems with lower fluxes invariably produce much less eruptible magma. A possible exception is when the evolved interstitial melt becomes efficiently segregated from its crystalline matrix by internal processes such as compaction, crystal-settling and gas-driven filter pressing (Cashman and Giordano 2014), mechanisms could generate crystal-poor rhyolitic eruptions, like those from Long Valley, California (Bachmann and Bergantz 2004). However, large eruptions of crystal-rich dacite require either high heat fluxes or efficient convective stirring of the magma system (Bartlett 1969; Couch et al. 2001), a process not considered in the models presented here.

Efflux of Magmatic Water

The water contents of some magmas can be buffered by the production of hydrous minerals which result from reactions between hydrous melt and anhydrous minerals (Beard et al. 2004). However, magmas with more than ~2 wt% H₂O release water continuously as they solidify. Upon total solidification almost all of the original magmatic water is lost, with only a small amount retained in hydrous minerals, such as amphibole and mica. Fresh, granitic rocks rarely contain more than 0.5 wt% H₂O, despite forming from magmas with as much as ten times this value initially. Getting the water out of solidifying plutons is an interesting fluid mechanical problem, especially as it is hard to find field evidence of fluid-loss pathways through granites, with the notable exception of pegmatite veins and patches. Perhaps such features record the passage of much greater volumes of water than their size would suggest.

FIGURE 3C–E shows how water is released from the three systems as they evolve. The rate of efflux, that is, the slope of the water-lost versus time curves, is roughly constant during the intrusive phase, declining thereafter until the system fully solidifies. However, water continues to be lost from magmatic systems long after they have ceased to grow. This has implications for the development of hydrothermal ore deposits, whose formation is often attributed to volatile loss from granitic intrusions, often during the waning stages of magmatism. Our results are entirely consistent with such an idea.

Plutonism versus Volcanism

A long-standing question is how much magma is erupted from a system versus how much is retained in the form of plutons, i.e. the volcanic to plutonic ratio (VPR). In FIGURE 3G-I we define VPR as the ratio of eruptible ($F > 0.4$) magma to non-eruptible ($F \leq 0.4$) magma and show how it varies over time. Before the system is hot enough for a magma chamber to grow, the VPR oscillates between 0 and a value corresponding to the volume of each pulse relative to the total intruded volume at that stage. After eruptible magma starts to accumulate in the system, VPR evolves with time as a function of the magma flux. In the high-flux, small system $VPR \approx 0.5$ for most of the system's lifetime, dropping exponentially when intrusion ceases. For the high-flux, large system VPR transiently exceeds 1 as eruptible volumes exceed non-eruptible volumes. In the low-flux, large system VPR remains low (typically ≤ 0.2). If all the potentially eruptible magma were to erupt at any given time, our calculated VPR would approximate the extrusive to intrusive ratio of the system at that moment. Published estimates of VPR vary widely (0.03 to 1), with 0.2 a typical value (White et al. 2006), suggesting that low-flux systems are dominant unless, of course, it is much harder to get eruptible melt out of a magmatic system than it is to generate it.

What Can We Learn from Volcanic Zircons?

Advances in our ability to date precisely large populations of zircons from volcanic rocks has heralded a number of attempts to constrain the dynamics of sub-volcanic magma systems in terms of zircon age spectra (Charlier et al. 2005; Costa 2008; Allan et al. 2013). It is tempting to imagine that what comes out, as recorded by zircons, reflects what is present, in terms of magma. We explore this possibility by tracking the volume of melt that is both eruptible and zircon-bearing ($0.4 < F < 0.45$). FIGURE 4 shows the maximum age recorded by an individual erupted zircon crystal as a function of system age. We find a consistent mismatch between the oldest erupted zircon and the longevity of the system that spawned it, regardless of whether zircons redissolve or not. If we consider longevity (L) to be the time during which eruptible melt was present, then within the high-flux small system where $L = 83$ ky, the oldest volcanic zircons are 37 ky for the no-dissolution case and just 9.5 ky for the with-dissolution case. The corresponding values are 65 ky and 34 ky for the high-flux, large system ($L = 172$ ky), and 81 ky and 64 ky for the low-flux, large system ($L = 96$ ky). The greatest zircon ages are attained in magmas erupted after the cessation of growth, but before the oldest batch of eruptible magma solidifies. The age range implied by volcanic zircons always underestimates L , by a factor of 0.1 to 0.8 depending on magma fluxes and zircons' ability to dissolve. This finding may go some way to resolving the conflicting messages of zircons from plutons, wherein the magmatic longevity of a system is faithfully preserved, and from coeval volcanics, which sample only the eruptible portions (Costa 2008). Again there are caveats,

notably if eruptions dredge up partially disaggregated, zircon-bearing plutonic material (Charlier et al. 2005; Costa 2008).

CONCLUSIONS

Field geology, geochronology and heat-transfer considerations suggest it is time to reconsider the physical state of magmatic systems. Incremental growth of crustal magmatic systems confers complexity in terms of eruptible magma volumes, aureole dimensions, hydrothermal mineralization and magmatic fluxes. Simple thermal models afford surprising insights into this complexity. Further developments should include convection within and above the magmatic system, melt-crystal segregation mechanisms, models of ground deformation above magmatic systems, and improved thermodynamic models of magmatic phase relations.

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FIGURE CAPTIONS

FIGURE 1 Torres del Paine Intrusion, Chile. The granite is composed of three units (I-III) each several hundred meters thick separated by sharp contacts. The units themselves are composed of thinner sills with wavy contacts. PHOTOGRAPH AND INTERPRETATION BY JULIEN LEUTHOLD. From Annen et al. (2015).

FIGURE 2 Representative thermal models of magmatic systems. (A) Conceptual model: igneous body grows by addition of disk-shaped sills of fixed radius. Temperatures, melt fractions and zircon ages are computed on a 2-D vertical cross-section through the igneous

body and country rocks. **(B)** Numerical setup: temperatures are calculated by finite difference methods on a numerical grid using cylindrical coordinates. **(C)** Example results: snapshot of the system 192 ky after emplacement of first sill. Colours show temperatures; contours show melt fractions (F). Emplacement of 25 sills, each with thickness 200 m and radius 10 km has been simulated. The time interval between sill emplacement is 2 ky and emplacement rate is 0.1 m/y. The dashed line shows the outline of the entire intrusive body. The volume of eruptible magma is delimited by the $F = 0.4$ contour.

FIGURE 3 Simulated evolution of representative systems over time. The first column is for sills of radius 2.5 km emplaced at 0.1 m/y; second column is for sills of radius 10 km emplaced at 0.1 m/y; third column is for 10 km-radius sills radius emplaced at 0.025 m/y. First row shows the volumes that are: solid ($F = 0$), highly crystalline ($0.4 > F > 0$), mobile and zircon-bearing ($0.45 > F > 0.4$), mobile and zircon-free ($F > 0.45$). Second row shows the mass of H₂O in the system, dissolved in the melt, and the cumulative mass lost as vapour. Third row shows the volcanic to plutonic ratio. The vertical dashed line marks the end of sill emplacement. Simulations run until all melt is exhausted.

FIGURE 4 Simulated zircon ages. Red and green curves show maximum ages of zircon present in eruptible magma ($F > 0.4$); red for the case where zircons dissolve if magma is reheated; green for the case where zircons do not. Blue curve shows maximum age of zircons in the entire igneous system. The abrupt drops in the green curve in **(A)** and **(B)** correspond to solidification at the roof of the system that contained reheated, but undissolved, zircons. After this point, no reheated zircons remain and the green and red curves are coincident. Vertical dashed lines mark the end of magma emplacement, the time when no more eruptible magma ($F > 0.4$) is available, and complete solidification of the system.

TABLE 1 Magmatic parameters used in our models. For the sake of simplicity, we neglect the differences in density and specific heat between solid and liquid, and the temperature dependence of specific and latent heat.	
Specific heat (solid and melt)*	1265 J kg ⁻¹ K ⁻¹
Thermal conductivity	as in Whittington et al. (2009)
Density (solid and melt)	2700 kg m ⁻³
Latent heat*	3.13 × 10 ⁵ J kg ⁻¹
Sill thickness*	200 m

Initial geothermal gradient	25°C km ⁻¹
Emplacement depth	5 to 10 km (under-accretion)
Solidus (T_s)*	670 °C
Liquidus (T_l)*	930 °C
Emplacement temperature	930 °C
H ₂ O solubility (at all T)	6 wt%
Melt fraction*	$1 - (58.21 + 8.908 \sinh(-4.831[(T - T_s)/(T_l - T_s)] + 2.257)) / 100$
*Parameters and equations from Caricchi and Blundy (2015). T is temperature in °C	