The Magma Reservoirs That Feed Supereruptions

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The vigor and size of volcanic eruptions depend on what happens in magma reservoirs in the Earth's crust. When magmatic activity occurs within continental areas, large reservoirs of viscous, gas-rich magma can be generated and cataclysmically discharged into the atmosphere during explosive supereruptions. As currently understood, large pools of explosive magma are produced by extracting interstitial liquid from long-lived "crystal mushes" (magmatic sponges containing >50 vol% of crystals) and collecting it in unstable liquid-dominated lenses.

KEYWORDS: magma chamber, supereruption, rhyolite, evolution of Earth's crust, ignimbrite

INTRODUCTION

The rocky innards of the Earth locally melt to produce magma, which sometimes erupts as cataclysmic explosions. The energy released by these eruptions is matched only by that of large meteorite impacts. Whether large pools of magma generate huge volcanic eruptions, whose volume and duration can have global consequences for humanity, or solidify slowly within the Earth's crust to form plutons (bodies of magmatic rock that crystallize at depth beneath volcanic areas) depends primarily on what happens in magma reservoirs in the upper 10–15 km of our planet. These reservoirs are built over time spans of 10^5–10^6 years and cannot be observed directly. What goes on inside them, and what dictates their fate? We are only beginning to unravel their inner workings.

MAGMA RESERVOIRS, CHAMBERS, AND MUSHES

How Do We Know They Exist and What Do They Look Like?

The way magma behaves in reservoirs is strongly affected by the percentage of solid particles (crystals) present. Depending on the pressure, temperature, and chemical composition, the amount of crystals within partially molten regions of the Earth can range from 0 to 100%. When the crystal fraction in a magma (silicate liquid + crystal mixture) is between 0 and ~50%, the magma can flow. At a greater crystal fraction (resulting from cooling, for example), a critical mechanical threshold—thought to occur at around 50–60 vol%—is reached. At this stage, the crystals start touching each other, eventually forming a rigid skeleton, and the molten silicate–crystal mixture cannot flow (or erupt) anymore (Marsh 1981). Therefore, we define a **magma chamber** as a continuous region in which *eruptible* magma (crystal fraction <50%) is stored. We refer to a rigid or semi-rigid magma composed of a framework of touching crystals and interstitial liquid as **crystal mush** (mush can be somewhat mobile when close to its mechanical lock-up point, but it acts mostly as a "rigid sponge"—cf. Hildreth 2004). Magma chambers and mushes together form **magma reservoirs**.

Magma chambers and reservoirs are not directly observable because they form in the Earth's crust (or upper mantle). However, their existence is established by at least three lines of evidence:

1. Indirect observations using geophysical methods, foremost of which is seismic tomography (Miller and Smith 1999 and FIG. 1)
2. Exposed plutonic bodies (particularly those that crystallized in the upper ~5–15 km of the Earth's crust), which are believed to represent exhumed ("fossil") magma chambers (e.g. Miller and Miller 2002 and FIG. 2).
3. Large-volume volcanic eruptions (supereruptions), which imply that integrated pools of *eruptible* (dominantly liquid) magma are periodically present in the crust (FIG. 3).

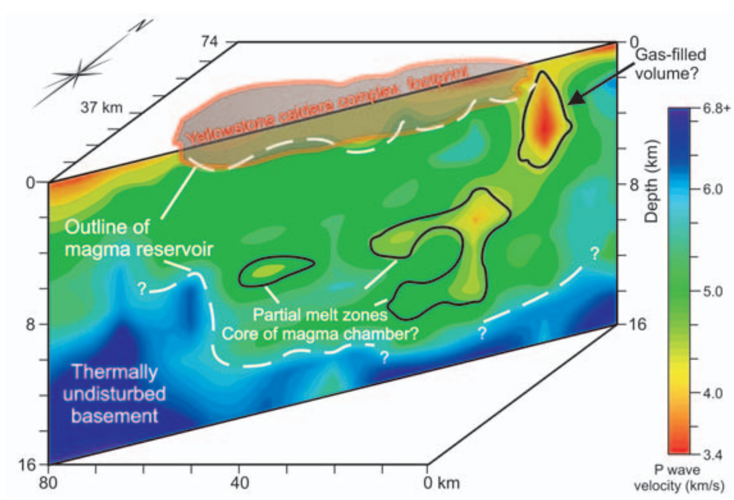

FIGURE 1 Tomographic image (slice picture obtained by analyzing the velocities of P-waves, Vp, generated by earthquakes) of a magma reservoir in the Yellowstone area (modified from Miller and Smith 1999). The warm colors indicate hotter (lower Vp velocities), partially molten regions (the potential magma reservoir) in the shallow crust beneath the caldera.

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On the basis of the observations provided by the different views of magma reservoirs listed above, we can make the following deductions:

- When conditions allow it, magma reservoirs can become very large (at least 5000 km^3, which corresponds with the largest-known erupted magma volume for a single eruptive episode), most likely by the addition of magmatic increments of various sizes and compositions over extended periods of time (e.g. Lipman 2007).
- The largest chambers appear to be elongated lenses with aspect ratios (thickness/length) of 1/5 to 1/10 (Fig. 4). This geometrical information can be inferred not only from geophysical images of active magmatic provinces (Fig. 1) and well-exposed plutons (Fig. 2), but also from the horizontal dimensions of calderas (i.e. depressions generated by the catastrophic emptying of underlying magma chambers during supereruptions; Fig. 3). These volcanic collapse structures provide an approximation of the horizontal extent of magma chambers and can be used to estimate the thickness of the erupted part of the magma chamber if the volume of erupted products is determined. The largest-known calderas have surface areas of 1000–3000 km^2 and volumes of 1000–5000 km^3; therefore, erupted magma lenses are on the order of a few kilometers (generally 1–2 km) thick.
- Magma reservoirs can form at different depths in the crust, preferentially at major transitions in rock density or strength (e.g. at the base of the crust and lithological breaks). Interestingly, most eruptions—both super- and normal-sized—appear to be fed from magma chambers at pressures of 1000 to 3000 bars (4–10 km depth; e.g. Chesner 1998; Lindsay et al. 2001; Bachmann et al. 2002). The rest of this paper will focus on what happens in these reservoirs.

WHAT PROCESSES OCCUR IN MAGMA RESERVOIRS?

Magmatic Differentiation

Most supereruptions involve rhyolitic magma rich in silica (>72 wt% SiO_2) and in volatile constituents (particularly H_2O). Both components contribute to the magma's explosive character: during decompression, H_2O produces low-density bubbles that can expand catastrophically, and the high silica content renders the liquid very viscous, trapping bubbles and making the bubbly mixture extremely buoyant. Two major processes contribute to the high SiO_2 and H_2O contents of these rhyolites, which are among the most chemically differentiated (differentiation refers to processes that generate different magma compositions from a single parent magma) of all magma types. The first is the incorporation of SiO_2- and H_2O-rich liquids from partially melted crustal rocks into hotter, SiO_2-poor (less-evolved) magmas as they ascend. The second, which probably plays an even bigger role in shallow, large reservoirs of evolved magma, is referred to as crystal–liquid fractionation.

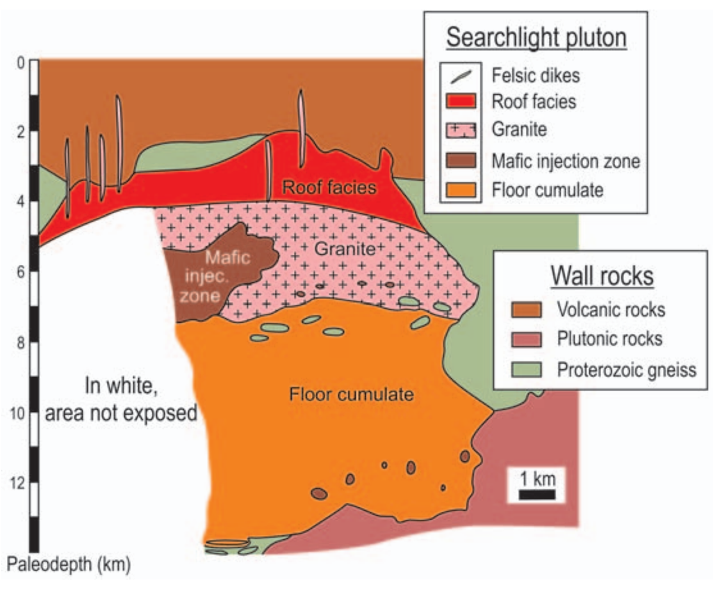

Figure 2 Cross section of a fossil magma chamber: the Searchlight pluton, Colorado River Extensional Corridor, Nevada (modified from Miller and Miller 2002). Tilting of large fault blocks during rapid extension in the area in the Miocene exposed a roof-to-floor cross section of this pluton in the southern Nevada desert.

How does crystal–liquid fractionation work, and how does it produce rhyolitic magma? It is in essence a distillation process. Magmas are complex, multicomponent, multiphase mixtures. When thermodynamic conditions (e.g. temperature or pressure) change, the different chemical components are redistributed into new phases (crystals, liquids, or gases). Silicon and water tend to be more concentrated in the liquid portion of magma. Due to the lower density of the liquid portion, it may be separated gravitationally from denser and less-Si-rich crystals, leading to progressive enrichment of Si and some other elements (including H_2O) in magmas as they ascend and construct the upper part of the Earth.

Two principal mechanisms (not mutually exclusive) cause the gravitational separation of crystals from liquid: (1) in solid-dominated systems (>50% crystals), interstitial liquid is extracted from crystal-rich residue by compaction (i.e. solid-state deformation of the crystal framework; McKenzie 1985), and (2) in liquid-dominated systems (<50% crystals), dense solid particles settle to the floor of the chamber. Both mechanisms are extremely slow in silica-rich magmas because density contrasts between liquids and crystals are generally small (a few tens of percent at most), crystals are small (rarely larger than 5 mm), and the silicate liquid is very viscous (up to a billion times the viscosity of room-temperature H_2O!). For example, using Stoke's Law, a 1 mm^3 (~box shaped) crystal would take ~10 years to sink 1 meter in a rhyolitic liquid with a viscosity of 10^5 Pa s. Moreover, there is an additional complexity in liquid-dominated systems: slow convection currents driven by the density differences associated with the presence of crystals and with recharging—the arrival of new, hotter, and less-dense magma—commonly occur in the chamber (e.g. Grout 1918). These currents tend to stir the magma, keeping crystals in suspension.

How do we study the magmatic processes that occur in these chambers? Due to the geological instantaneity of supereruptions (hours to perhaps years; Wilson 2008 this issue), their products, especially giant ignimbrites (or ash-flow tuffs—deposits emplaced by hot avalanches generated during explosive eruptions), provide an invaluable snapshot of the state of the chamber at a given time. By assembling observations on a number of volcanic systems that evacuated their chambers at different times during their evolution, a time-integrated reconstruction of the magmatic puzzle can be made.

Chemical Patterns in Ignimbrites as Clues to Magma Chamber Dynamics

Since their recognition in the late 1950s and early 1960s (e.g. Smith 1960), a number of giant ignimbrites have been carefully studied. Despite local variations, they fall roughly into three categories on the basis of their whole-rock chemical composition (Fig. 5; see Hildreth 1981 for more details):

FIGURE 3 Aerial picture of the 26.9-million-year-old Creede caldera in the San Juan volcanic field, Colorado (e.g. Lipman 2007), taken from a small airplane. Snowshoe Mountain, a post-caldera resurgent dome, is ~15 km in diameter. Red dashed line represents approximate caldera rim.

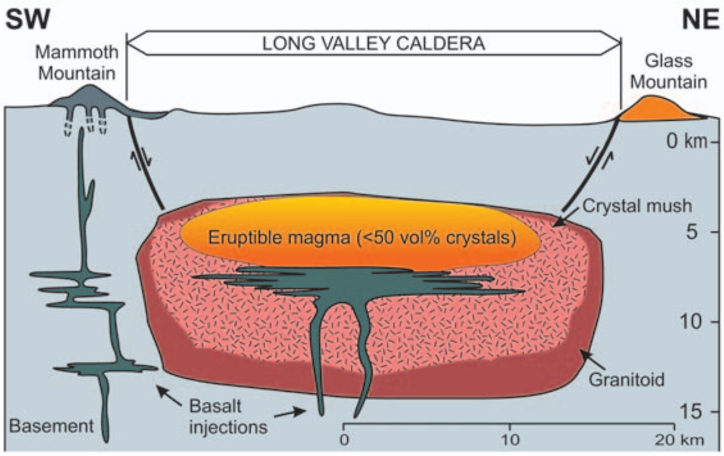

FIGURE 4 Simplified cross section of one of the most studied caldera systems in the world: the Long Valley caldera, California (modified from Hildreth 2004; Hildreth and Wilson 2007). Vertical scale and relative volumes are approximate.

Group 1 – These sheets of explosively erupted magma show gradational compositional zoning from early- to late-erupted deposits. Eruptions generally start by tapping crystal-poor rhyolitic melts (which form the base of the deposits) and end with more crystal-rich, less-differentiated magma types. Group 1 ignimbrites are the most common products of supereruptions. Well-studied examples include the Bishop Tuff, California (Hildreth and Wilson 2007); the Bandelier Tuff, New Mexico (Wolff and Ramos 2003); and the Huckleberry Ridge Tuff, Wyoming (Wilson 2008).

Group 2 – Unlike the first group, these volcanic deposits have almost no compositional gradients. They are further subdivided into two subcategories: (a) crystal-poor rhyolites, such as units of the Taupo Volcanic Zone (Wilson 2008) and (b) crystal-rich dacites (crystal fractions up to 45 vol% and SiO_2 contents on the order of 65–70 wt%), such as the Fish Canyon Tuff (Bachmann et al. 2002) and the Atana ignimbrite (Lindsay et al. 2001). Individual erupted volumes of these crystal-rich, homogeneous dacites are up to five to ten times larger than erupted volumes of Group 1 rhyolites. Based on their high crystal content, we consider these to be erupted mush.

Group 3 – This group of ignimbrites, observed in slightly smaller eruptions (erupted volumes on the order of 10s to 100s of km^3), shows abrupt gaps in composition between early- and late-erupted deposits (e.g. Crater Lake; Bacon and Druitt 1988). As with Group 1 ignimbrites, eruptions start with crystal-poor rhyolites, but shift abruptly to crystal-rich, less-differentated compositions. Although Group 3 ignimbrites have not (yet?) been documented in supereruption deposits, they provide valuable, complementary information about how magmas are stored in reservoirs.

Large Mushes as Rhyolite Nurseries

Although they erupt infrequently, Group 2 dacites are a common magma type in the evolution of the upper continental crust (Lipman 2007). They resemble in every respect (compositionally, mineralogically, and texturally) the most abundant constituent of the upper crust, granodioritic plutons. Units such as the Fish Canyon Tuff (up to >5000 km^3) are believed to represent the erupted equivalents of these plutons. This connection makes these Group 2 ignimbrites a particularly informative window into the later, most mature stages of magmatic activity at the largest scales.

The erupted mushes of Group 2 reflect the presence of large, crystal-rich storage zones of rhyolitic liquid in the Earth's crust. These storage zones can be considered as rhyolite nurseries: not only are these gigantic, mush-like magma bodies common, but, while they are hot and active, up to 50% of their volume is interstitial rhyolitic liquid that can be retained for periods of up to several hundred thousand years (see Reid 2008 this issue). As these crystal-rich mushes are much more voluminous than crystal-poor rhyolites, they can generate the largest examples of Group 1 ignimbrites; only ~10–20% of the stored interstitial rhyolitic liquid is required to be extracted from the crystal framework to form such ignimbrites. However, as mentioned previously, crystal–liquid separation is extremely slow in these magmatic systems. How then, can interstitial liquid in these mushes be extracted fast enough to form large volumes of *highly eruptible,* crystal-poor rhyolitic liquid on geologically reasonable timescales?

The Mush Model

Several mechanisms to separate liquid from crystal residues have been proposed during the last century. They include simple settling of individual crystals, downward flow of crystal-laden magma in plumes, formation and ascent of pockets of lower-density liquids along the crystallizing sidewall of the chambers, and interstitial liquid extraction by compaction. Most of these mechanisms have been discussed since the earliest studies of magmatic systems (Grout 1918; Bowen 1928) and have been reconsidered in more detail by many authors since (see, for example, Chen and Turner 1980; McBirney 1980; McKenzie 1985; Marsh 2002). They all probably play a role in contributing to the final character of mushes and liquids extracted from them, but they generally are too inefficient to generate the volumes of crystal-poor rhyolitic liquid required to produce Group 1 ignimbrites (see Bachmann and Bergantz 2004 for an in-depth account of these hypotheses).

Recently, a model incorporating several of these dynamic processes and using the mechanism of expulsion of interstitial liquid from crystal frameworks (Brophy 1991) has been applied to these silicic mushes (FIG. 6; Bachmann and Bergantz 2004; Hildreth 2004; Hildreth and Wilson 2007). The model is based on the assumption that there is a crystal fraction window (between ~45 and ~65 vol% crystals) in which the separation of crystals and interstitial liquid is particularly efficient. When stored magmas contain less

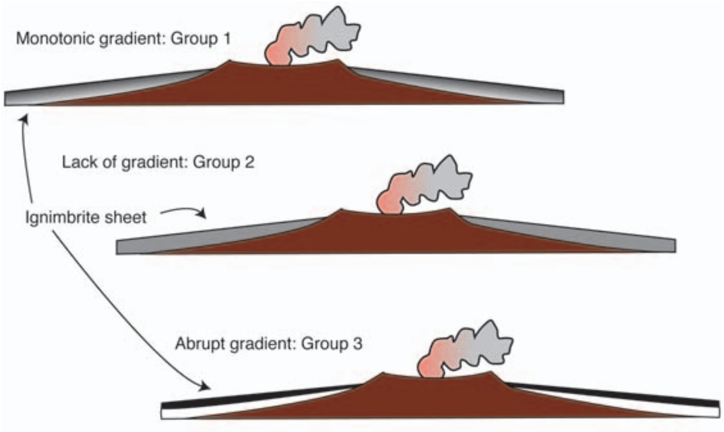

FIGURE 5 Schematic illustration of the three most common types of compositional patterns in pyroclastic volcanic deposits (ignimbrites): (**Group 1**) gradients are monotonic, with deposits showing a nearly continuous range in composition; (**Group 2**) some deposits, in particular the largest ignimbrites, display a conspicuous lack of chemical gradients at the hand sample scale; (**Group 3**) abrupt changes in composition are observed in certain ignimbrites.

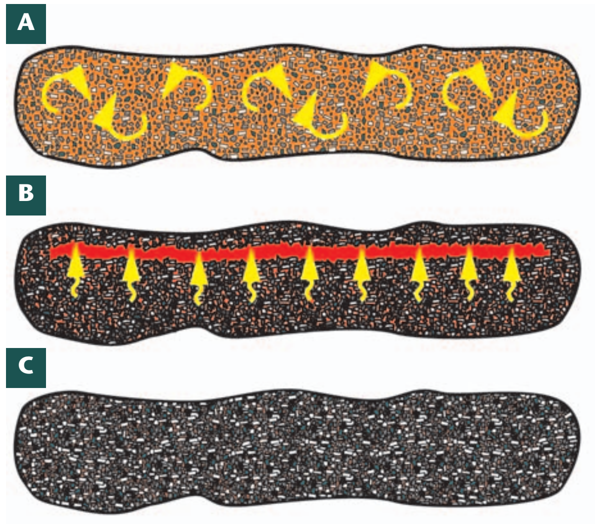

FIGURE 6 Schematic illustration of the evolution of a mushy magma reservoir (Bachmann and Bergantz 2004; Hildreth 2004). (**A**) Low-crystallinity stage (<45 vol% crystals): most crystals are kept in suspension by convection currents. (**B**) Medium-crystallinity stage (~45–60 vol% crystals): the absence of convection and the high permeability provide a favorable window for crystal–melt separation. (**C**) High-crystallinity stage (>60 vol% crystals): the permeability is too low for high-viscosity melt to be extracted efficiently by compaction.

than ~45 vol% crystals, convection currents efficiently stir the magma chamber (i.e. the crystals are homogeneously distributed). However, when the crystal fraction approaches ~65 vol%, separation of crystals from liquid occurs only by compaction; separation is extremely slow, even at geological timescales, in such viscous, low-permeability systems (McKenzie 1985). Therefore, physical separation of crystals and liquid should be most efficient when convection just comes to a halt, stopping the disruptive stirring effect; at that point, the liquid fraction in the mush is high enough for a combination of mechanisms, such as crystal settling, microrearrangement of the crystal framework (microsettling), and high-permeability compaction (see Bachmann and Bergantz 2004 for more discussion on these processes) to bring about the efficient separation of crystals from the high-viscosity liquid. This model also takes advantage of the fact that dacitic mush bodies (a) have low aspect ratios (low height/length), limiting the distance over which the viscous liquid has to travel upward, and (b) survive in the upper crust for >100,000 years (Reid 2008) providing enough time for extraction to occur. Calculations of liquid extraction rates and timescales for rhyolitic cap formation above large mush bodies (Bachmann and Bergantz 2004) are in broad agreement with the longevities of mushes.

The mush model (FIG. 6) provides a framework to explain some of the characteristics of each group of ignimbrites described above. Past explanations for the origin of the chemical heterogeneities in Groups 1 and 3 (zoned ignimbrites), for example, have invoked interaction between recharging and resident magma batches within the magma chamber (Smith 1979; Hervig and Dunbar 1992; Eichelberger et al. 2000); indeed, different degrees of mixing between compositionally diverse batches of magma can lead to the observed complexities in chemical zoning of erupted products (arrested homogenization of Eichelberger et al. 2000). However, several chemical indicators reveal that compositionally distinct early-erupted and late-erupted parts of some deposits at least partly relate to each other by crystal–liquid fractionation within the magma reservoir (in situ differentiation; Hildreth 1981, 2004). The mush model pertains to this latter scenario of in situ differentiation within the reservoir.

Both Group 1 and Group 3 represent the mature stage of the mush model. In the largest systems, a liquid-rich rhyolitic cap forms above a mush, but the process can take time (up to several tens of thousands of years). During this protracted extraction period, fluctuations in temperature and crystal fraction are expected to occur within the mush, generating some heterogeneities. Therefore, incremental extraction of different batches of interstitial melt from the mush (e.g. Hildreth and Wilson 2007) and/or slow stirring *in the cap* by sluggish convection currents can produce fairly continuous gradients in composition and crystal fraction. Abrupt transitions occur (Group 3) when the top of the crystal-rich mush below the cap is tapped during an eruption, leading to the observed rapid shift from crystal-poor to crystal-rich deposits in the field (e.g. Bacon and Druitt 1988).

In contrast, the absence of zoning in the crystal-rich units of Group 2 can be explained by invoking either an immature or a reactivated ("rejuvenated") stage of the mush model. A large portion of the mush itself is erupted, without the presence of a significant rhyolitic cap. Therefore, the absence of zoning requires a situation in which magmatic stirring has been efficient enough either (1) to keep the magma as a homogeneous suspension (no formation of a cap) or (2) to redigest the rhyolite cap, erasing the memory of the extraction stage. Homogeneous, crystal-poor rhyolites (such as those found in the Taupo Volcanic Zone; Wilson 2008) may also be explained by efficient convective stirring in a situation where a large amount of heat is added to the crust by hot-magma recharge events below the silicic magma chambers (the Taupo Volcanic Zone has the highest heat flux of any volcanic arc), leading to vigorous convection currents.

MUSH REJUVENATION AS AN ERUPTION TRIGGER

The main triggering mechanisms for giant eruptions are obviously important beyond pure academic interest. A necessary condition for the eruption to occur is the pressurization of the chamber above a critical value, allowing dike propagation in wall rocks (e.g. Jellinek and DePaolo 2003). Both internal and external triggers are possible. External triggers refer generally to the collection of tectonic stresses acting on the magma chambers and are likely to play a role in many supereruptions (e.g. Lindsay et al. 2001). However, internal overpressurization mechanisms (saturation of the silicate liquid in gas and formation of bubbles, recharge by new input magma) can be important for smaller eruptions (e.g. Pallister et al. 1992), and there is no reason to believe that they cannot act as triggers in large systems.

A number of volcanic deposits present evidence for reheating and partial remelting of their crystals prior to eruption. Several authors have suggested that recharge events can rejuvenate crystal-rich mushes that are otherwise too viscous to flow or erupt (e.g. Murphy et al. 2000; Bachmann et al. 2002). New magma batches are constantly generated in the mantle or deep crust and during ascent can encounter existing rhyolitic–dacitic magma reservoirs. These recharges can add heat and gases to overlying mush, without mixing thoroughly with it, and can result in some partial remelting of the crystal network (Bachmann and Bergantz 2006). Adding new magma to the chamber, in addition to transforming some crystalline material into liquid, leads to an increase in volume and overpressurization (in gas-saturated systems) at a rate that may overcome the strength of the wall rocks and trigger (or help trigger) an eruption. In addition, a consequence of unlocking the crystal framework could be rapid convective overturn in the magma chamber, creating additional stresses on the surrounding rocky container. If mush rejuvenation is confirmed as an important process in triggering volcanic eruptions, monitoring whether active systems are cooling down or heating up (as is being done at present at Yellowstone; see Lowenstern et al. 2006) could be critical in predicting the eruption of an awakening giant magma chamber.

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