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Deciphering Magma Chamber Dynamics from Styles of Compositional Zoning in Large Silicic Ash Flow Sheets

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INTRODUCTION

The understanding of the dynamic processes in magmatic systems has grown and changed markedly in the last decade. Old models for magmatic systems as vats of near-liquidus material have been revised by observations from seismology (Sinton and Detrick 1992), crystal chemistry and zoning (Davidson et al. 2007) and the geochronology and geochemistry of both plutonic and volcanic systems (Hildreth 2004; Charlier et al. 2007; Miller et al. 2007; Peressini et al. 2007; Walker et al. 2007). New views emphasise magmatic systems as temporally dominated by crystal mushes (magma bodies with a high fraction of solid particles; see definition in Miller and Wark 2008), that wax and wane in temperature and crystallinity, and are subject to significant open system processes (Charlier et al. 2007; Hildreth and Wilson 2007; Walker et al. 2007; Bachmann and Bergantz 2008). These processes, such as magma reintrusion, mixing, gas sparging, and subsequent thermal rejuvenation, may be significantly more important in producing the characteristics of a magmatic system than the previous closed-system, near-liquidus behaviour would predict. There are a number of recent reviews that summarize these observations (for example see Eichelberger et al. 2006; Bachmann et al. 2007b; Lipman 2007). Our aim here is to illustrate how this new perspective to magma dynamics is motivated by observations of heterogeneities (or lack thereof) in erupted rocks (and to a lesser amount in plutons).

Owing to rapid withdrawal and quenching of magma during explosive volcanic eruptions (hours to a few days), large-volume (>1 km^3) pyroclastic deposits (also referred to as *ignimbrites* or *ash-flow tuffs*) provide an instant image of the state of the magma chamber evacuated during eruption. A first-order observation that characterizes these pyroclastic deposits of intermediate to silicic composition is that many do not tap into chemically homogeneous reservoirs but show compositional and thermal zoning, with early-erupted material differing significantly from late-erupted material (e.g., Lipman et al. 1966; Lipman 1967; Smith 1979; Hildreth 1981). In detail, nearly every eruptive product is different (see Hildreth 1981 for a comprehensive classification), but three general groups can be defined as follows (Table 1 and Fig. 1): 1) deposits showing quasi monotonic (or linear) gradients in composition and/or temperature, 2) deposits showing abrupt gradients in composition, with gaps that can reach several wt% in major elements over a very narrow stratigraphic interval, and 3) deposits showing no significant compositional or thermal gradients. Our objective is to provide a synopsis of zoning patterns preserved in ignimbrites as a means for understanding magma dynamics in silicic systems. We will argue that retaining and/or producing homogeneity by convective stirring is actually more challenging than preserving gradients, particularly in viscous, *crystal-rich*, silicic magma chambers, and

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Table 1. Characteristics of some well-studied ignimbrites.

	Abruptly zoned units	Linearly zoned units	Homogeneous units
Examples	*Normally zoned:* Purico ignimbrite, Crater Lake ignimbrite, Katmai 1912, Los Humeros, Timber Mountain/Oasis Valley caldera complex *Reversely zoned:* Whakamaru ignimbrite, Bonanza Tuff	Bishop Tuff, Bandelier Tuff, Huckleberry Ridge, Tuff La Primavera Tuff, Taylor Creek Rhyolite, Loma Seca Tuff, Laacher See ignimbrite, Campanian Ignimbrite, Aso Ignimbrite	Fish Canyon Tuff, Lund Tuff, Cerro Galan ignimbrite, Atana ignimbrite, Oranui ignimbrite
Gradients in:			
Major element	Abrupt gaps (up to 20 wt% SiO_2)	Generally limited (few wt% at most)	Not measurable
Trace element variations	Significant	Significant	Not measurable
Isotopic ratios	Limited or absent in some cases, and important in others	Significant, and often correlated with Differentiation Index	Some heterogeneities at the whole-rock scale
Crystallinity	Abrupt change (few percent to >30 %). More mafic endmember generally more crystal-rich	Significant increase from early-erupted to late-erupted (~1 to >20 vol%)	Not measurable
Temperature	Up to 100 °C in cases but absent in others	Reported (up to 100 °C)	Not measurable
Gas content	Top richer in volatiles	Top richer in volatiles	Not measurable

Linearly zoned units: Hildreth 1981; Christiansen 1984; Grunder and Mahood 1988; Mahood and Halliday 1988; Johnson 1989; Duffield et al. 1995; Christiensen and Halliday 1996; Davies and Halliday 1998; Knesel et al. 1999, Lipman 1967; Wolff and Storey 1984; Worner and Schmincke 1984b; Worner and Schmincke 1984a; Wolff 1985; Wolff et al. 1990; Civetta et al. 1997

Abruptly zoned units: Whitney et al. 1988; Chesner 1998; Lindsay et al. 2001; Schmitt et al. 2001 Bacon 1983; Bacon and Druitt 1988; Hildreth and Fierstein 2000; Lipman et al. 1966; Varga and Smith 1984; Brown et al. 1998; Bindeman and Valley 2003

Homogeneous units: Francis et al. 1989; Lindsay et al. 2001; Bachmann et al. 2002; Maughan et al. 2002; Dunbar et al. 1989; Wilson et al. 2006.

require two conditions: (1) density filtering of material allowed to enter the growing silicic reservoir, and (2) thermal buffering close to the haplogranite eutectic.

Types of gradients in ignimbrites

Precisely reconstructing the zonation in a magma chamber from erupted deposits is difficult. Low sampling density, complexities in the depositional patterns induced by shifting eruptive vent sites and fluctuations in eruption rates (e.g., Wilson and Hildreth 1997), partial mixing during magma ascent in conduits (Blake and Ivey 1986a,b; Trial and Spera 1992) and the time-dependence of the dispersal of fragmented materials (e.g., Neri et al. 2003) will scramble many of the finer details of any spatial patterns of composition within the chamber prior to eruption. Hence only first-order observations, such as how early-erupted material differs from late-erupted material, can be considered as robust expressions of pre-eruptive

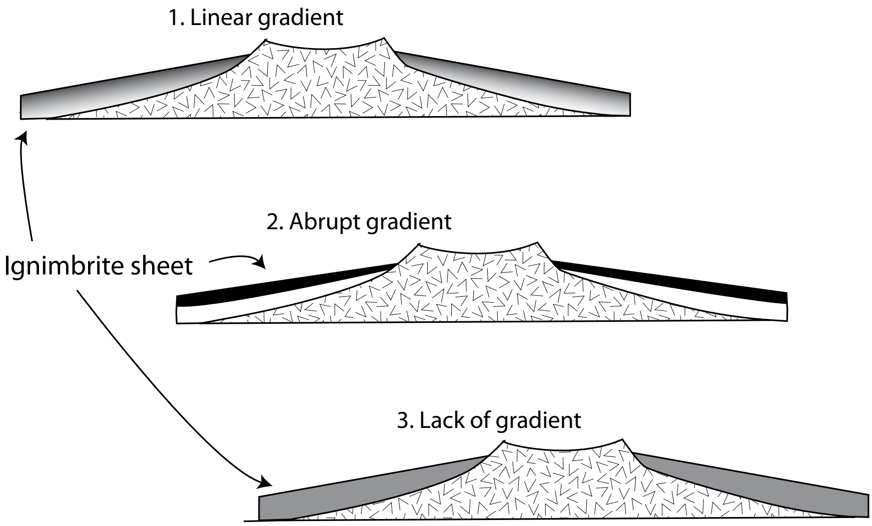


Figure 1. Schematic illustration of the three types of zoning patterns that commonly occur in ignimbrite (or ash-flow sheets) (modified from Bachmann and Bergantz 2008).

conditions for inter-deposit comparison. However, by sampling *unfragmented material* (pumice or lava blocks) from multiple localities in a given ignimbrite, an approximation of the pattern of gradients that existed in the magma chamber (whether abrupt, nearly continuous or absent, Fig. 1) can be obtained.

It is important to appreciate that by using the term "gradient," we are referring to a *horizontally-averaged quantity*. In detail the variations in composition, temperature, and crystallinity associated with convective stirring can locally vary significantly over short distances (envisage mixing two different flavors of honey with different colors), but these intensive variables will on-average produce a vertical gradient in a gravity field if the different phases have variable densities. Below we provide a short synopsis of the different types of gradients, then assess the role of convection in magma chambers and combine the two to explain how distinct types of zoning, or the lack thereof, emerge.

Abrupt gradients

Step-like changes in the whole-rock chemical composition of pumices up to several wt% for certain major elements has been documented in many deposits of explosive volcanic eruptions (e.g., Brophy 1991, Fig. 2). Striking color changes make these abrupt variations visibly manifest (see Bacon and Druitt 1988; Druitt and Bacon 1989; Eichelberger et al. 2000 and Fig. 2 for pictures), and they are commonly associated with large changes in crystallinity (the more mafic deposits being more crystal-rich, Bacon and Druitt 1988; Hildreth and Fierstein 2000).

Linear gradients

When plotted on geochemical variation diagrams, many silicic ignimbrites present a nearly continuous array in pumice composition (e.g., Smith and Bailey 1966; Hildreth 1979; Halliday et al. 1984; Grunder and Mahood 1988; Streck and Grunder 1997; Wolff et al. 1999; Milner et al. 2003, Fig. 3) from most evolved, early-erupted to least-evolved late-erupted deposits, although, in some volcanic units, small compositional gaps have been reported (Fridrich and Mahood 1987; Streck and Grunder 1997). The observed chemical variations are particularly striking for trace elements, but also involve major elements, isotopic rations, crystallinity,

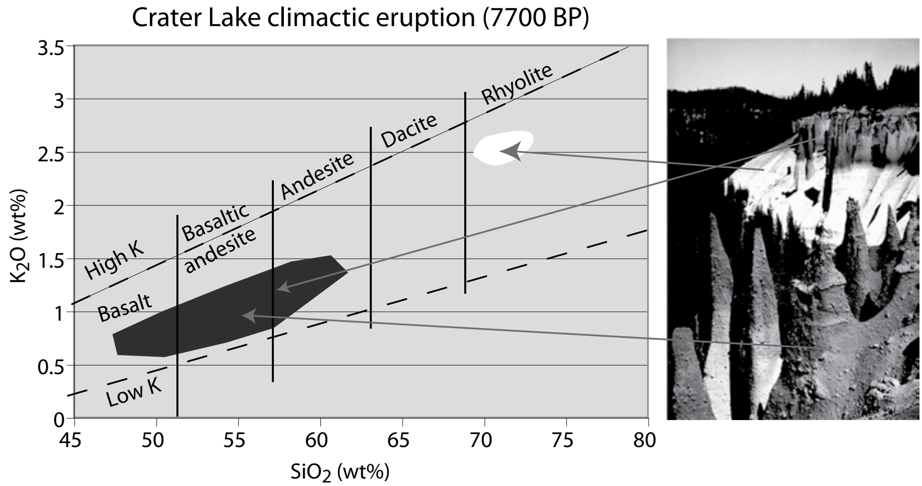


Figure 2. Compositional gap in the Crater Lake ignimbrite, obvious both on chemical plots and in the field. (modified from Bacon and Lanphere 2006).

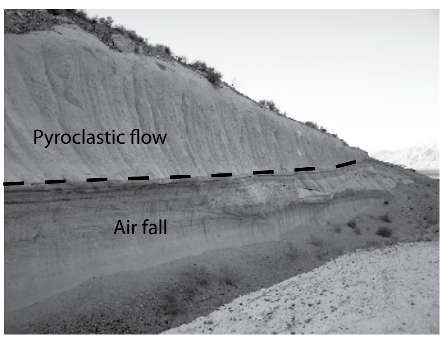
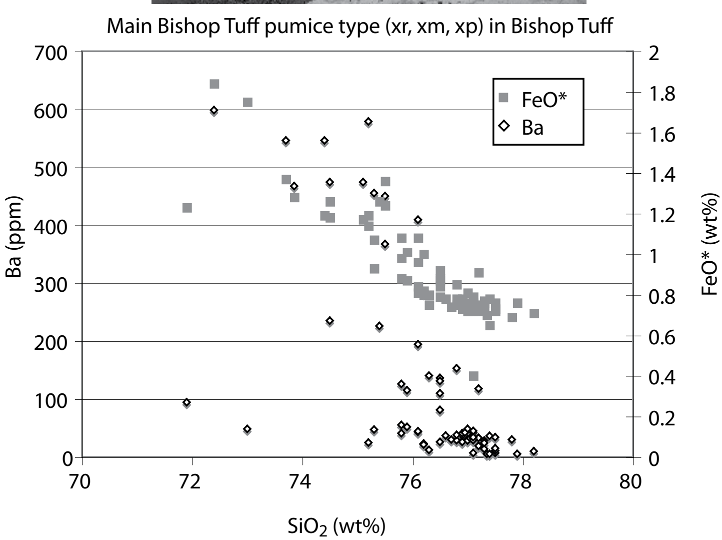


Figure 3. Compositional spread in pumices of the Bishop Tuff, showing progressive depletion in Ba and FeO* as a function of SiO_2 (modified from Hildreth and Wilson 2007).



temperature, and gas content (Table 1). These variations in most magmatic tracers imply that fairly continuous compositional and thermal gradients are present in the magma reservoir tapped by the eruption (no information on their spatial distribution though).

As an example of a typical linearly-zoned system, the Bishop Tuff shows: (1) an increase in crystal content and decrease in volatile content from early- to late-erupted material (Hildreth 1979; Wallace et al. 1999), (2) lack of significant major element zoning, but more than twofold variation in incompatible trace elements and extreme depletion in compatible trace element (e.g., Sr, Ba) in early-erupted material (Hildreth 1979; Michael 1983; Hervig and Dunbar 1992; Knesel and Davidson 1997; Anderson et al. 2000), (3) variations in ε_{Nd}, $^{206}Pb/^{204}Pb_i$, and $^{87}Sr/^{86}Sr_i$, but homogeneity in oxygen isotopes (Bindeman and Valley 2002), and (4) a ΔT of ~100 °C from early to late-erupted material, documented with Fe-Ti oxides (Hildreth and Wilson 2007), oxygen isotope thermometry (Bindeman and Valley 2002), and Ti-in-Quartz thermometry (Wark et al. 2007).

Lack of gradients

Among a number of well-described zoned ignimbrites, several examples stand out as having very homogeneous whole-rock characteristics (including major and traces elements, isotopic ratios, crystallinity, and temperature; Fig. 4). Most of these homogeneous units belong to the *Monotonous Intermediates*, a group of large (>1,000 km^3), crystal-rich (up to 45 vol% crystals), dacitic units erupted in mature continental arcs (Hildreth 1981, Francis et al. 1989; de Silva 1991; Chesner 1998; Lindsay et al. 2001; Bachmann et al. 2002; Maughan et al. 2002). However, some homogeneous examples also occur in rhyolitic, crystal-poor units (Dunbar et al. 1989; Wilson et al. 2006).

Figure 4. A section of intracaldera Fish Canyon Tuff (in Canyon Diablo, CO), showing a lack of any kind of gradient from top to bottom. Up to 1 km thick section is exposed on the northern flank of the resurgent dome (La Garita mountains), with no base and top exposed, implying a thickness significantly greater than 1 km (estimated at an average of 2 km for the entire 80×30 km La Garita caldera collapse area; Lipman 2000).

MECHANISMS FOR GENERATING GRADIENTS IN SILICIC MAGMA CHAMBERS

Following the recognition that many large-volume ignimbrites originate in compositionally complex magma reservoirs (Lipman et al. 1966; Lipman 1967), numerous models have been proposed to explain the origin of these gradients. Below we summarize some key elements of these models, and discuss them in light of recent advances in our understanding of the dynamics of viscous, multi-phase mixtures (Bergantz and Ni 1999; Jellinek and Kerr 1999; Jellinek et al. 1999; Bergantz 2000; Bergantz and Breidenthal 2001; Phillips and Woods 2002; Burgisser et al. 2005; Dufek and Bergantz 2005). We will first summarize some aspects of the kinematics of convection in magmatic systems.

The role of convection

As noted by Grout almost a century ago (Grout 1918) and many workers since (Morse 1986; Marsh 1988; Rudman 1992; Simakin et al. 1997; Simakin and Botcharnikov 2001), convection in magma reservoirs is unavoidable as long as a dominantly-liquid (*but not necessarily crystal-poor*) batch of magma is present. The most significant source of density changes that provides the potential energy for convection are associated with phase change; examples include the formation of crystals and/or bubbles. Magmatic systems are also open to new injections (with both similar and different chemical compositions) and can assimilate their wall-rocks (DePaolo 1981; Davidson and Tepley 1997; Bohrson and Spera 2001; Thompson et al. 2002; Beard et al. 2005), leading to additional temperature and crystallinity (and hence buoyancy) fluctuations within the reservoir (Keller 1969; Bergantz 2000; Murphy et al. 2000; Bachmann et al. 2002). Thus, convective dynamics within magma reservoir are the result of both (1) *in situ* density changes arising from cooling and/or volatile exsolution and (2) magma reintrusion and/or volatile flux.

Convective stirring in magmatic systems is likely to be highly intermittent and arise from many or all of the above mentioned buoyancy sources during the lifetime of a reservoir (which can last several hundreds of thousands of years; e.g., Reid et al. 1997; Brown and Fletcher 1999; Vazquez and Reid 2004; Bacon and Lowenstern 2005; Simon and Reid 2005; Bachmann et al. 2007a). Geological data require that any particular model must accommodate the following observations: a) magma bodies cool on conductive timescales and rarely assimilate large amounts of their margins by rapid heat transfer (e.g., Carrigan 1988; Barboza and Bergantz 2000; Dufek and Bergantz 2005; Petcovic and Dufek 2005), b) heterogeneous crystal populations (particularly in age population of zircons) at the margins and core of plutonic suites generally preclude simple, monotonic sidewall and roof crystallization (Eichelberger et al. 2006a; Miller et al. 2007; Walker et al. 2007), c) open system processes and secular cooling produces intermittent convection with Rayleigh numbers high enough ($>10^5$) where conditions for chaotic convection are (at least temporarily) obtained.

In these complex, open-system magma reservoirs, convection has a dual nature: it reduces the scales and intensity of heterogeneities (e.g., Oldenburg et al. 1989; Coltice and Schmalzl 2006), but can *also* produce gradients in intensive variables and composition if the source of buoyancy in itself generates heterogeneities (Bergantz and Ni 1999; Jellinek and Kerr 1999; Jellinek et al. 1999, Figs. 5 to 8). The following paragraphs attempt to clarify this dual nature.

It is widely understood that mixing requires advection of material, which acts as a source of localized high strain, reducing length scales of heterogeneities and allowing a significant volume of material to be serviced by the mixing process. Subsequently, diffusion can operate effectively once the average distance between heterogeneities have been reduced below the diffusive length scale (e.g., Coltice and Schmalzl 2006). The most efficient mixing systems will be those that can produce the greatest amount of strain at the widest range of scales, typically by the rotational motion of eddy-like structures that both entrain fluid and wind them.

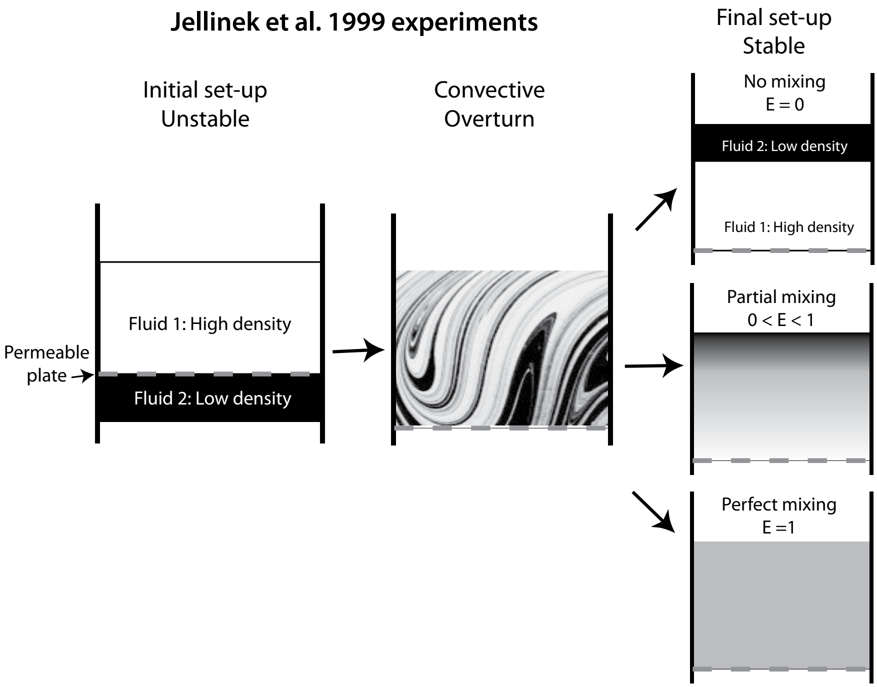


Figure 5. Schematic diagram of mixing experiments described by Jellinek et al. (1999), starting from an initial, gravitationally unstable set-up, inducing an overturn, and leading to a final mixture, whose mixing efficiency will depend on the density and viscosity ratios of the two fluids.

The Reynolds and Rayleigh numbers are the most important diagnostics of the dynamic and kinematic conditions within in a magma body. A very general form of the Rayleigh number is (Jellinek et al. 1999):

$$\mathrm{Ra} = \frac{BH^4}{\nu\kappa^2} \tag{1}$$

where B is the buoyancy flux, H the height of the system, ν the kinematic viscosity of the ambient magma and κ the diffusivity. The Reynolds number is traditionally defined as:

$$\mathrm{Re} = \frac{uL}{\nu} \tag{2}$$

where u is a scaled velocity, L the matching length scale and ν the kinematic viscosity. It can also be written as (Jellinek et al. 1999):

$$\mathrm{Re} = \frac{B^{1/3}H^{4/3}}{\nu} \tag{3}$$

(assuming that the velocity is scaled on a balance of inertia and buoyancy, a problematic assumption for small Reynolds numbers but that does not impact our overall conclusions). The buoyancy flux appearing in both the Rayleigh and Reynolds numbers will be a complex function of the conduction-controlled rate of cooling, the geometry (as it impacts cooling rate), the resulting density changes associated with phase change, and thermal buffering with latent heat (see, for example, Jellinek and Kerr 2001).

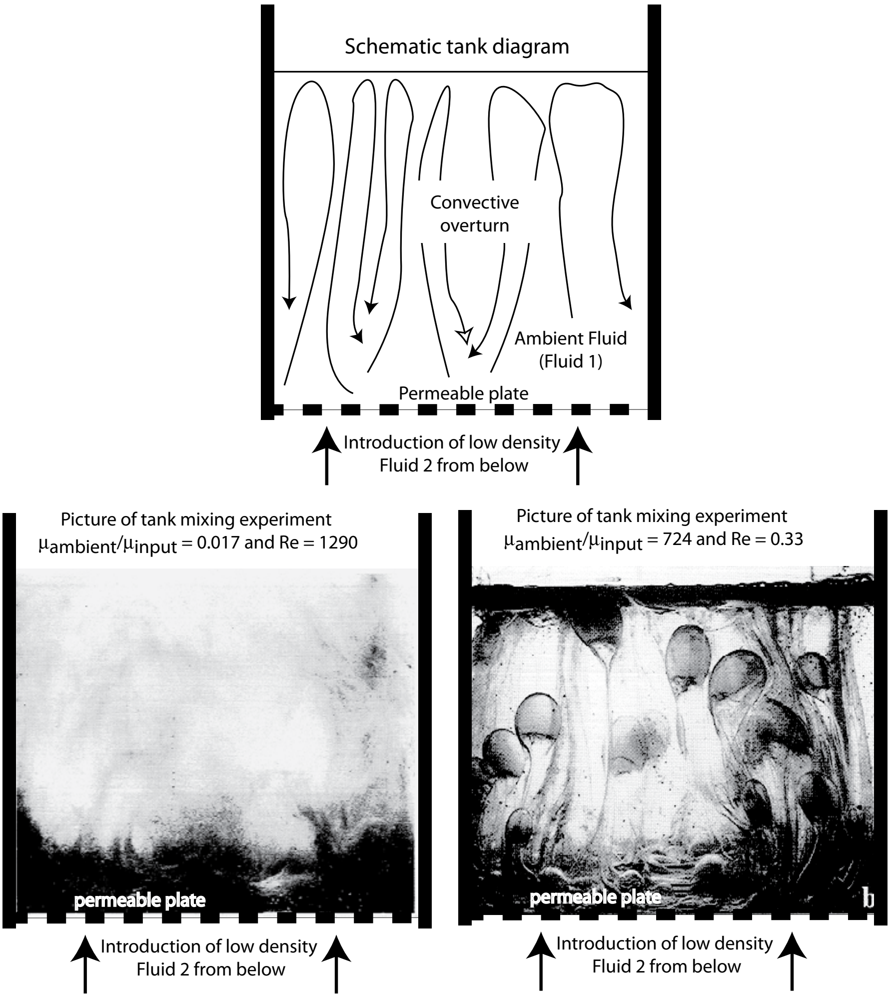


Figure 6. Illustration of the overturn process in the mixing experiments of Jellinek et al. (1999).

There are generally three regimes where the Reynolds number produces a distinct flow template: (1) laminar, (2) transitional and (3) fully turbulent. In the laminar regime (Re <1), buoyancy forces are balanced by viscous forces (viscosity controls the dynamics of the flow), while in the turbulent regime (Re ~>1000), the buoyancy force is balanced by inertial forces (onset of true turbulence). Intermediate values between these two (1 < Re < 1000) indicate that both viscosity and inertia play a role, and this is the most difficult range to generalize because the flows are not self-similar. This means that the structure of the flow is not ordered at the largest scales: a small change in Reynolds number can produce an intermittent flow regime that is not similar to the one that preceded it.

At high values of the Reynolds number (~1000), the mixing transition is reached, and there is a full spectrum of eddy sizes from largest scales set in part by the geometry of the system (initial and boundary conditions), to the smallest scales of eddies set by the viscosity. For systems with Reynolds numbers at or above the mixing transition, mixing is fast and

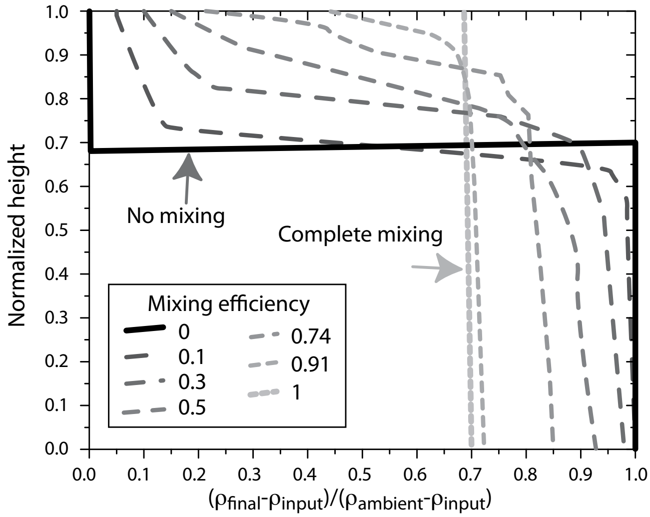


Figure 7. Mixing efficiency for fluids at different densities and viscosities as defined by experiments of Jellinek et al. (1999), illustrated in a 2D cross section of a tank.

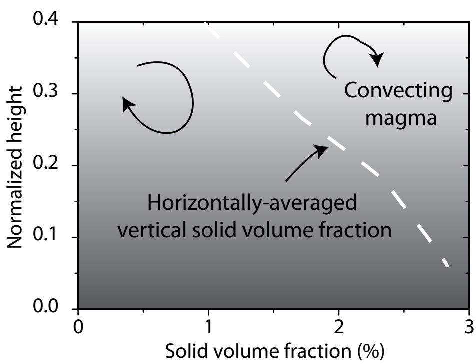


Figure 8. Horizontally-averaged vertical gradient in crystallinity within a reservoir (modified from Bergantz and Ni 1999). These results were obtained by numerically simulating cooling of an originally solid-free liquid, triggering precipitation of dense, solid phases at the roof. Once the dense roof layer reaches a critical crystallinity, solid-rich plumes collapse, leading to the growth of a stable gradient in solid fraction within the convecting, liquid body. See Bergantz and Ni (1999) for details.

thorough, as mixed material is quickly passed among eddies of many sizes (Broadwell and Mungal 1991; Dimotakis 2005), thus efficiently breaking-down existing gradients. Jets and plumes produced in explosive eruptions reach this Reynolds number.

For Re <1000 (low to transitional Reynolds numbers), there is not a statistically complete set of eddy sizes in the flow. Only a few eddy sizes dominate and the high-strain stirring is of limited efficiency. If the Rayleigh number is high enough, such situations are prone to "chaotic" convection (e.g., Ottino 1989; Ferrachat and Ricard 1998; Petrelli et al. 2006). The onset of chaotic conditions can be determined by the appropriate Rayleigh numbers, although the values will depend on the specific system under consideration. Generally, however, as long as the Rayleigh number is above about 10^5, chaotic conditions can be obtained, even with a low Reynolds number (Coltice 2005).

Chaotic mixing leads, on the short term, to islands of well-stirred material in a domain of otherwise poorly stirred material (see overturn illustration in Fig. 5). If kept active long enough, this kind of flow can lead to efficient mixing (unless there are strong contrasts in viscosity and density; Ottino 1989). However, if the source of buoyancy and convective stirring itself induces heterogeneities, then complete homogeneity cannot be reached.

Consider *discrete, mechanically passive*, chemical heterogeneities such as a local patch within the magma system that has a distinct character, such as a small difference in crystallinity, temperature or composition (or all three). If there is chamber-wide overturn driven by cooling, heating or reintrusion from the boundaries, this patch will be homogenized with the rest of the magma reservoir in a few overturns in a chaotically convecting system (e.g., Coltice 2005) even at low Reynolds number; it does not require sustained vigorous fluid circulation. If the viscosity contrast between the patch and the magma reservoir is large, the homogenization may take longer but even crystal multiphase aggregates can be eroded and dissipated by shear (Powell and Mason 1982; Petrelli et al. 2006).

On the other hand, if new heterogeneities are the sources of the convective process itself (such as dense crystal plumes originating from a cold ceiling, or reintrusion of low density magma from below), the system cannot reach perfect homogeneity (except at extremely high mixing efficiencies in high Re number situations). As heat exchange with the surroundings never completely stops, new heterogeneities in temperature, inducing variations in crystal content are constantly created, impeding complete homogeneity in crystallinity (and therefore composition); new gradients are re-established as old ones are consumed. More generally, any process taking place at the chamber boundaries that leads to variations in trace elements, crystal content, or temperature can export that variation into the magma body by convection (e.g., Marsh 1989) and produce a gradient in these quantities, while also homogenizing pre-existing gradients. Therefore, even in a chemically closed (but cooling) system, perfect homogeneity cannot be achieved for waning systems.

As illustrated in numerical and laboratory experiments (Figs. 5 to 8), convective processes can readily generate gradients as they transport material and heat across a magma system. Any density instability that originates at a boundary (e.g., crystal-rich plume falling from the top, a gas-rich region that rises from the bottom, or a reintrusion of less dense material from below) will all produce a gradient as it passes through the system. As material at the boundary rises or falls, this process will entrain surrounding material (see Bergantz and Breidenthal 2001), become diluted, and deliver this now mixed material to the other boundary; it also leaves a "wake" of material that is a mix of the initial falling or rising plume and the surrounding magma. If the process is repeated, it will produce and sustain a (horizontally averaged) gradient, as it has been exemplified for a number of applications to magmas (Bergantz and Ni 1999; Jellinek and Kerr 1999; Jellinek et al. 1999; Ruprecht et al. 2008).

The ability of rising or falling plumes to create gradients in the surrounding magma has been called the "mixing efficiency" (Linden and Redondo 1991; Jellinek and Kerr 1999; Jellinek et al. 1999). It is defined on the basis of an initially unstable stratification, such as a less dense layer entering the bottom, allowed to mix with the over-lying fluid as it penetrates upward. One endmember is that the plumes rise or fall through the ambient magma with no mixing, simply changing position, and the least amount of initial gravitational potential energy is retained in the final configuration. The second endmember is perfect mixing, leaving no gradient but a uniform mixture over the entire vertical span of the system. Between these two extremes, the overturn process produces a stably stratified system and the vertical variation in the density, and hence concentration, from the mixing of the penetrating and ambient fluids produces a gradient (Fig. 7). To quantify this, the mixing efficiency, E, is defined as:

$$E = \frac{P_f - P_{\min}}{P_{\max} - P_{\min}} \tag{4}$$

where $P_{\max}$ is the *final*, maximum value of the potential energy associated with perfect mixing (which yields no gradient), $P_{\min}$ is the *final* value associated with no mixing as the two layers simply exchange places and P_f is the actual final potential energy of the system following overturn:

$$P_f = \int_0^H g\left(\rho_a - \rho(z)\right) z dz \tag{5}$$

g the scalar acceleration of gravity, ρ_a the density of the ambient, or resident fluid and $\rho(z)$ the horizontally averaged vertical density profile after overturn. It is important to note that this definition of the mixing efficiency is a measure of the (horizontally averaged) vertical gradient. The mixing efficiency has a value for zero if the two fluids slide past each other with no mixing, and unity if the two fluids mix perfectly creating a system with no horizontally-averaged vertical gradient. So a value of the mixing efficiency of say 0.5, does not imply that 50% of fluid A and 50% of fluid B are everywhere well mixed. It means that there is a gradient in density (and hence concentration) from top to bottom with a slope of 0.5 and with equal amounts of the resident and intruding material at only one vertical position.

For silicic magmas, there will be a wide range of buoyancy flux B depending on the source of buoyancy (thermal, compositional, phase change), but reasonable values will vary between 10^{-7} and 10^{-11}. Thus, for a given height, kinematic viscosity and diffusivity, one can estimate the approximate Rayleigh and Reynolds numbers across a range of compositions (using Eqns. 1 and 2). As shown in Figure 9**,** the Reynolds number generally remains low (<10) even for the largest possible examples (up to 5 km thick) and most dynamic scenarios. Therefore, the mixing efficiency will always be low in magmatic systems, and for heterogeneities related to magma cooling, perfect homogeneity cannot be reached.

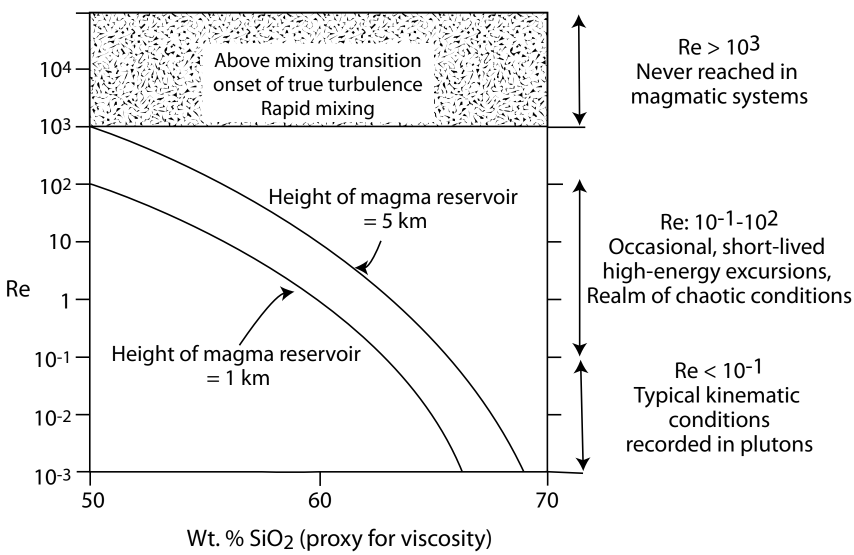


Figure 9. Reynolds number as a function of composition for different magmas reservoirs.

Origin of heterogeneity in silicic magmas: convective stirring vs. phase separation

On the basis of the arguments expressed above, the presence of zoning or heterogeneities in a volcanic unit can represent two endmembers: (1) incomplete, or partial convective mixing of two magma batches or (2) chemical differentiation or distillation by crystal-liquid separation of an initially more homogeneous system. These two scenarios are similar to the concepts of "arrested homogenization" and "progressive unmixing" of Eichelberger et al. (2000), although we stress that convective stirring of different magma batches does not necessarily lead to homogenization (and therefore does not need to be "*arrested*").

Convective interactions of compositionally variable magma batches present in the same reservoir are clearly common in the geological record. Magmatic plumbing systems favour it, and a number of ignimbrites record the mechanical encounters between magmas of from different sources (Smith 1979; Hervig and Dunbar 1992; Mills et al. 1997; Bindeman and Valley 2003; Eichelberger et al. 2006b; Davidson et al. 2007; Knesel and Duffield 2007). However, many volcanic deposits display chemical heterogeneities that have been interpreted as being *dominantly* induced by crystal fractionation within a given reservoir (*in situ* differentiation, e.g., Thompson 1972; Hildreth 1981; Michael 1983; Brophy 1991; Francalanci et al. 1995; Thompson et al. 2001; Hildreth 2004). A magma residing in a cooler reservoir will undergo crystallization and volatile exsolution. As the new phases produced have different densities, the system will have the tendency to differentiate, or unmix, into two or more batches with distinct characteristics. Therefore, zoning patterns present in deposits with an *in situ* differentiation signature represent *progressive unmixing* of an initially more homogeneous magma body (e.g., McBirney 1980; Eichelberger et al. 2000).

Both scenarios of heterogeneization can occur at different time scales depending on the physico-chemical conditions in the magma reservoirs. They are explored below (when applicable) in the development of the different types of gradients in magma chambers.

Mechanisms to generate abrupt gradients

An abrupt juxtaposition of distinct magma compositions (and/or crystal content) can easily be explained by the interaction of two distinct magma batches with different characteristics (e.g., Eichelberger et al. 2006b). However, abrupt compositional gaps are also found in units showing geochemical evidence for *in situ* differentiation (Bacon and Druitt 1988; Druitt and Bacon 1989; Brophy 1991; Hildreth and Fierstein 2000). Due to the small size of crystals in magmatic systems (typically around 0.1 to 5 mm) and the high viscosity of SiO_2-rich melts (10^4 – 10^6 Pa·s), separating crystals from its melt is a slow process (Sparks et al. 1984; Reid et al. 1997; Anderson et al. 2000; Eichelberger et al. 2006a), particularly if the magma is undergoing some convective stirring (Martin and Nokes 1989; Burgisser et al. 2005). Therefore, crystal-melt separation is enhanced when convection stops, which usually occurs when a crystal-bearing magma transforms into a locked crystalline mush (at ~50 vol% crystals for low strain rates; Vigneresse et al. 1996; Petford 2003; Rosenberg and Handy 2005; Caricchi et al. 2007; Champallier et al. 2008). Once the crystallinity is high enough (> ~70%), permeability becomes so low that crystal-melt separation by compaction cannot occur on geologically reasonable timescales (McKenzie 1985; Bachmann and Bergantz 2004). Hence, the most favourable crystallinity window appears to be around the rheological threshold (~50-60 vol% crystals), when convection has stopped but the system remains permeable enough for the interstitial melt to be expelled from the compacting crystalline framework (Thompson 1972; Brophy 1991; Thompson et al. 2001). The low-density interstitial melt can then accumulate above the mush, and generate a nearly crystal-free cap (Bacon and Druitt 1988; Bachmann and Bergantz 2004; Hildreth and Wilson 2007; Walker et al. 2007, Fig. 10). In the event of the formation of a crystal-poor cap by interstitial liquid extraction from a crystalline mush, eruption of both the crystal-poor cap and its underlying mush during the same eruptive episode will lead to the observed abrupt gradient present in many systems (Fig. 11).

THE CASE OF THE ABRUPTLY ZONED CRATER LAKE ERUPTION

Despite the wide compositional gap between the nearly-aphyric Crater Lake rhyodacite and the co-erupted crystal-rich andesite and clear evidence for a chemically open-system (e.g.,

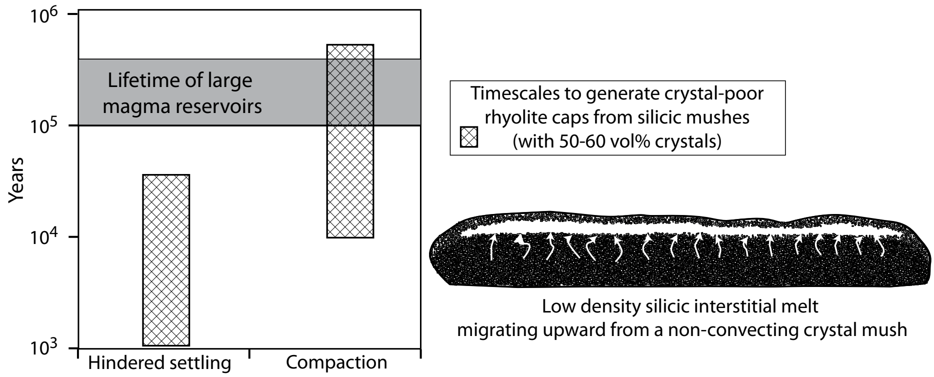


Figure 10. Range of timescales of interstitial liquid extraction (by hindered settling and compaction) for typical silicic mushes with crystallinities of 50-60 vol% (for more details, see Bachmann and Bergantz 2004).

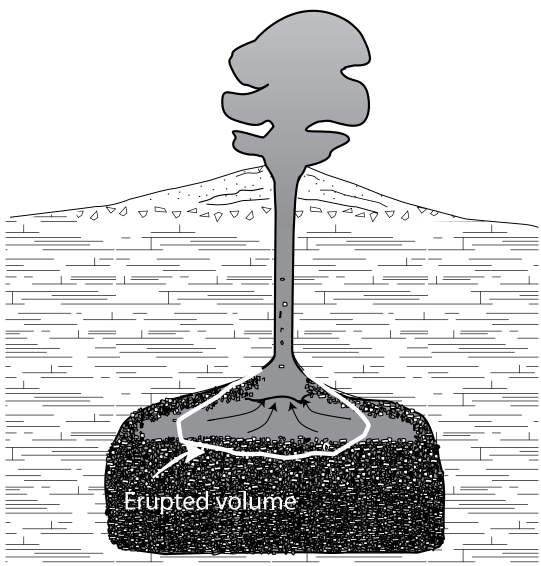


Figure 11. Development of a crystal-poor horizon above a crystalline mush (and below a crystallizing roof layer). When such a systems in tapped by eruption, the least-viscous, crystal-poor material is expected to erupt first. When exhausted, the viscous-crystal-rich portion can be drawn-out, rapidly leading to chocking of the eruption. In natural deposits, such as the 7700 BP Crater Lake ignimbrite, one expects the commonly observed sequence of a crystal-poor base covered by a relatively thin, crystal-rich, co-magmatic layer.

low and high Sr magma types), the two compositional layers have strong affinities, which suggest a genetic link by crystal fractionation (Bacon and Druitt 1988; Druitt and Bacon 1989): (1) the rhyodacite has trace element contents that require extensive fractional crystallization. (2) The interstitial glass trapped within the andesitic layer is very similar in composition to the rhyodacite. (3) Temperature and oxygen fugacity are nearly identical in both compositional layers. These chemical and thermal affinities, in addition to the physical proximity, indicate that the eruption tapped a magma reservoir similar to the setting illustrated in Figure 11. The zoning pattern was induced by crystal-melt separation.

Mechanism to generate linear gradients

As for the abrupt gradient case, the two models for generating linear gradients in magma chambers are: *in situ* differentiation processes (Hildreth 1981; McBirney et al. 1985; Trial and

Spera 1990; Marsh 2002) and magma mixing by new addition from below (Smith 1979; Hervig and Dunbar 1992; Eichelberger et al. 2000; Knesel and Duffield 2007). The dynamic template for the latter hypothesis has been investigated by a number of authors (Sparks and Marshall 1986; Frost and Mahood 1987; Oldenburg et al. 1989); incomplete mixing leads to imperfect chemical blending (often referred to as mingling), and stratification in the chamber (Jellinek et al. 1999). However, the mechanisms leading to stratification by the *in situ* evolution process is more controversial. The most commonly accepted mechanism is the "sidewall crystallization" hypothesis, involving the extraction of interstitial liquid from a crystallizing boundary layer at the cold margin of a magma chamber (Chen and Turner 1980; McBirney 1980; Rice 1981; Huppert and Sparks 1984; Spera et al. 1984; McBirney et al. 1985; Wolff et al. 1990; de Silva and Wolff 1995; Spera et al. 1995). Although this mechanism may play a role in the generation of chemical and physical heterogeneities in some cases, several considerations preclude sidewall crystallization being the dominant differentiation template in crustal magma chambers.

a. Bodies of eruptible magmas in large silicic chambers are sill-like, having a low wall to roof (or floor) ratio, rendering sidewall crystallization inefficient (de Silva and Wolff 1995)

b. Little evidence for sidewall crystallization is preserved in extensively mapped and studied plutonic bodies (e.g., McNulty et al. 2000; Barnes et al. 2001; Zak and Paterson 2005; Eichelberger et al. 2006a; Miller et al. 2007; Walker et al. 2007).

c. The fact that most crystals found in silicic magmas (except for some rhyolites and aplites) are complexly zoned requires that crystals undergo complex transport paths and circulation. This is inconsistent with the monotonic sidewall crystallization hypothesis, where crystals would not be available to circulate and respond to changes in the magmatic environment.

d. The vertical stacking of several convecting magma "layers" is inherently unstable as drag and entrainment should occur at the interfaces between the different magma batches, leading to re-blending at a rate similar to estimated differentiation rate (see Davaille 1999; Gonnermann et al. 2002; Bachmann and Bergantz 2004 for more details in this process).

Thus, the generation of a linear gradient by a "box filling" mechanism of double-diffusive convection in a largely fluid reservoir appears unlikely in magmatic situations. We argue that mixing of heterogeneities by sluggish convection is the most common process of generating linear gradients in magma chambers. Linear gradients will certainly develop when two different magmas interact, but such gradients can also develop *in situ* as crystallization and differentiation continue even during periods of closed-system evolution, generating heterogeneities in crystallinity and composition (e.g., Bergantz and Ni 1999; Bergantz 2000; Couch et al. 2001) that will be progressively smeared over the entire chamber.

THE CASE OF THE LINEARLY ZONED BISHOP TUFF

Numerous models have been proposed to explain the distinctive characteristics of the Bishop Tuff. The most recent and comprehensive treatment of this unit is Hildreth and Wilson (2007). They propose a model for the zonation of the Bishop Tuff in which different pockets of rhyolitic melt are segregated from a large, subjacent crystal mush. As the conditions within the mush vary over time, extracting interstitial liquid at different stages would have lead to rhyolitic pockets with slightly different chemical characteristics. This model appears plausible, but does not take into account the effect of convection, which must have occurred in a system showing a thermal gradient of ~100 °C (Hildreth 1979; Hildreth and Wilson 2007; Ghiorso and Evans 2008).

On the basis of the observations mentioned above and new constraints provided by chemical and thermal heterogeneities in quartz crystals, which require timescales of less than 10^2-10^4 years (Bindeman and Valley 2002; Wark et al. 2007), the characteristics of the Bishop Tuff seem best explained by a model very similar to the one proposed by Hildreth and Wilson (2007), but with the requirement of some late convection stirring in the rhyolite cap induced by a hotter reintrusion (leading to the development of a thermal gradient and the growth of bright rims on quartz from the late-erupted material; Wark et al. 2007). So both convective stirring just prior to eruption, and progressive unmixing to produce the rhyolite, seems to have played a role in the evolution of the Bishop Tuff magma (see Fig. 12).

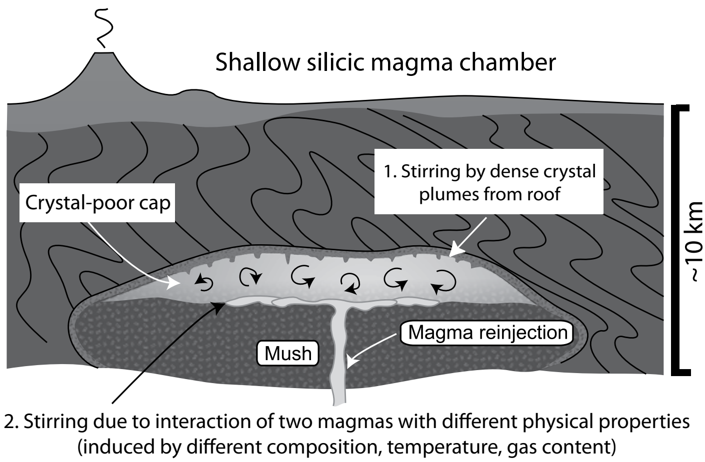


Figure 12. Schematic illustration of the two main possible sources of heterogeneities in shallow, crystal-poor, silicic magma chambers, such as the Bishop magma reservoir.

The generation of homogeneity

In light of the discussion above, homogenizing at the whole-rock scale a crystal-rich silicic magma, the most viscous silicate liquid on Earth (Scaillet et al. 1998), appears unlikely. We consider below possible mechanisms to reach homogeneity in viscous silicic magmas. These ideas are largely based on studies of homogeneous volcanic systems, such as the Fish Canyon magma body (see following section), but stress that more work is needed to better constrain some of these issues.

The largest silicic magmas bodies (10,000+ km^3) are certainly constructed incrementally (e.g., Deniel et al. 1987; Petford et al. 2000; Lipman 2007). To obtain these large homogeneous masses of magma in such a clearly open-system situation, new magmatic additions need to be either (1) similar in composition to the growing reservoir, or (2) efficiently blended by mechanical stirring. Furthermore, crystal-melt separation (which will recreate heterogeneities) should be hampered (Fig. 13).

All three requirements can be met if the following conditions are obtained. As a new magma batch intrudes in the upper crust, it will quickly start crystallizing in the cold, low-pressure environment. Once crystallinity increases, further changes in temperature will be slowed by the latent heat release by crystallization, and increasingly ineffective conductive cooling (Koyaguchi and Kaneko 1999). Therefore, magma batches are expected to remain as crystal mushes (with crystallinities > 50 vol%) for most of their time above the solidus.

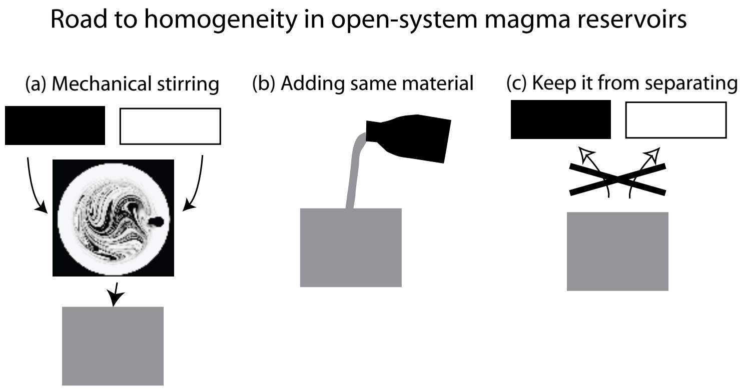


Figure 13. Illustrations of possible ways to render and/or keep open-system magma bodies homogeneous.

Magmatic additions to this growing *silicic* magma reservoir are restricted to either (1) similar composition (Hervig and Dunbar 1992; Eichelberger et al. 2000), released by a filtering lower crustal MASH zone (Hildreth and Moorbath 1988) or (2) more mafic composition (hotter, less viscous, and generally denser), which will mostly pond beneath the low-density, silicic mush (Miller and Miller 2002; Wiebe et al. 2002; Harper et al. 2004). This latter situation is prone to rejuvenation and self-mixing as the mafic magma acts as a hot plate (Couch et al. 2001; Bachmann and Bergantz 2006). As the magma body never remains for long in a crystal-poor situation, crystal-melt separation can only occur by interstitial melt extraction.

THE CASE OF THE FISH CANYON MAGMA BODY

The Fish Canyon magma erupted during the largest known silicic volcanic eruption, the crystal-rich Fish Canyon Tuff (45 vol% crystals) and its satellite units erupted from the same magma chamber (the Pagosa Peak Dacite and Nutras Creek Dacite, Lipman et al. 1997). This system is well known for its whole-rock homogeneity (at the scale of the hand sample and larger), in major and trace element composition, mineralogy, modal abundance, isotopic ratios, temperature, and water content (Whitney and Stormer 1985; Johnson and Rutherford 1989; Bachmann et al. 2002; Charlier et al. 2007). Similarly homogeneous units have been described in detail in the Andes (Francis et al. 1989; Lindsay et al. 2001), and in the Great Basin, USA (Maughan et al. 2002). However, at the mm scale, The Fish Canyon magma shows an enormous range in composition (both major, trace and isotopic), implying complex open system behaviour and excursions in intensive variables (Bachmann and Dungan 2002; Bachmann et al. 2005; Charlier et al. 2007). Of particular note: (1) different patches of glass analyzed by microdrilling with a mm spacing are heterogeneous in $^{87}Sr/^{86}Sr_i$; also, (2) biotite crystals have $^{87}Sr/^{86}Sr_i$ higher than any other component in the magma (including the glass around them) and thus indicate a provenance from the Precambrian wall rocks surrounding the magma chamber (Charlier et al. 2007). Therefore, as these biotites were derived from the disaggregation of blocks of material from the walls and are now found in the Fish Canyon magma, *some* crystal dispersal by convection is required. On the basis of Sr diffusive reequilibration of these Precambrian biotites, this dispersal had to occur in less than 10,000 years prior to eruption.

One possibility to dissipate gradients in crystal-rich situations, as anticipated by the "defrosting" hypothesis (Mahood 1990), is reheating of locked silicic mushes by interaction with hot, intrusive magmas. As it has been described in multiple natural systems (Murphy et al.

2000; Couch et al. 2001; Bachmann et al. 2002; Bachmann and Bergantz 2006; Hildreth and Wilson 2007), cooling silicic magmas reach a crystal mush state (crystal content >50-60%) that becomes rigid at low strain rates (convecting currents stop). If sufficient heat is added by a more mafic re-intrusion *from below* (a late reheating event of about 50 °C is recorded by the mineral phases in the Fish Canyon magma (Bachmann and Dungan 2002; Bachmann et al. 2002; Bachmann et al. 2005), the crystalline framework will progressively re-melt from the bottom until becoming liquid again (able to sustain convection). The stirring agent in such case is a buoyancy force generated not only by heating, but more importantly by the melting of the high crystallinity barrier at the bottom of the chamber (thermal expansion accounts for only 1-10% of density variation due to crystallinity changes) AND by the injection of volatiles from the commonly gas-rich new magma (Bachmann and Bergantz 2003, 2006). This re-heating can create an inverted density stratification that could evolve to full-scale self-mixing if the reheating event can supply enough energy.

An important observation made in rejuvenated systems is that they appear not to have significantly mixed with the more mafic magma that acted as a heat source; convective mixing is limited to the reactivated mush (self-mixing of Couch et al. 2001). In the Fish Canyon system, only a few mafic enclaves have been preserved in the late erupted intracaldera facies of the Fish Canyon Tuff (Bachmann et al. 2002) although a large basaltic andesite unit (the Huerto Andesite) erupted immediately after the Fish Canyon Tuff eruption (Parat et al. 2005, 2008). This absence of chemical mixing can be understood if buoyancy terms are compared. As density variations due to compositional changes are of the order of a few percent (~10 times larger than buoyancy forces driven by thermal expansion during a reheating of around 100 °C), reheating alone will not be sufficient to invert the compositional density gradients and enable large scale convection involving both mafic and silicic magmas.

THE CASE OF THE HOMOGENEOUS GRANITOIDS

Most large silicic plutons (such as those found in the Sierra Nevada Batholith) are very similar to the Monotonous Intermediates. We consider the largest of them (e.g., the Half Dome Granodiorite in the Tuolumne Intrusive Suite; Bateman and Chappell 1979) as unerupted equivalents of units such as the Fish Canyon magma. We argue that they follow the same incremental growth scenario, undergoing periodic intrusions of (a) evolved magma of similar composition blending in the growing reservoirs and (b) more mafic intrusions ponding at the base of the mush (Wiebe and Collins 1998; Robinson and Miller 1999; Waight et al. 2001; Miller and Miller 2002; Wiebe et al. 2002), which induce convective stirring due to addition of heat and gas from below (Wiebe et al. 2007). The only difference between plutonic and volcanic rocks is that the former cooled slowly to full solidification, allowing time to undergo some local crystal-liquid separation. Such late crystal-melt separation, leading to evolved granitic cupolas on top of large granodioritic bodies, is observed in several cases of well-exposed plutonic sections (Johnson et al. 1990; Barnes et al. 2001; Walker et al. 2007; Wiebe et al. 2007).

CONCLUSIONS

Many caldera-forming (large-volume) ignimbrite sheets display chemical and thermal heterogeneities, reflecting evacuation from a compositionally zoned shallow reservoir. These chemically and thermally complex reservoirs are an expected consequence of open-system processes that are common in magmatic systems. Mixing of magma batches with different physical properties (density, viscosity), assimilation of wall rocks, and internal phase changes (crystallization, gas exsolution) related to cooling and decompression all lead to chemical and thermal gradients within the reservoirs.

In contrast to common belief, convection is not an efficient homogenizing agent at all scales in viscous magmatic systems and can produce long-lasting gradients. Recent experiments and numerical studies have shown that heterogeneities (in composition, crystal content, temperature) will arise (or be preserved) as sluggish convection stirs the system. If left active long enough in a chaotic mode, convection can produce homogeneity, except in the case of thermally-induced heterogeneities, which are continuously re-established in cooling magma bodies. Therefore, in most magmatic situations, gradients are inescapable. This inference is on par with volcanic deposits from explosive eruptions (ignimbrites), which, in many cases, show heterogeneities in their whole-rock composition, crystallinity, temperature and volatile contents. Looking at the ignimbrite record, we conclude that most crystal-poor magma chambers will preserve heterogeneities as the system undergoes sluggish convection.

Paradoxically, the most viscous of these magmas (the crystal-rich Monotonous Intermediates) and many of the exposed fossil magma chambers (upper crustal silicic plutons) are remarkably homogeneous at the hand sample scale. If complete homogenization is the result of stirring by low Re convection, one would expect the most viscous magmas to be the least homogeneous. This apparent contradiction can be resolved by recognizing that homogeneous units are evolved, crystal-rich mush zones. Many silicic magmas persist as near-solidus crystal mushes, due to a combination of slow conductive cooling and latent heat buffering temperature close to the eutectic. We argue that large silicic crystal mushes mostly grow by addition of compositionally similar magmas, released by lower to mid-crustal MASH zones. As a result, the base-line homogeneity is dictated by processes occurring in source regions, and is modified by open-system events in the upper crust. These mushes block denser mafic reintrusions at their bases, and keep them from thorough mixing. This hot underplating triggers periodic stirring through large-scale overturn of the reservoirs by self-mixing and gas sparging. Such periodic chamber-wide stirring is well illustrated in erupted crystal-rich units (the Monotonous Intermediates) that show thorough convective stirring in the last 1,000-10,000 years prior to eruption.

Plutons cannot record a high-energy state, as they must cool slowly to their solidii. They, nonetheless, commonly appear fairly homogeneous at the hand sample scale, and we propose that (1) they are also stirred periodically by reintrusion events, and (2) are kept fairly homogeneous by the sluggishness of crystal-liquid separation in silicic systems. However, plutons do partially unmix by crystal-liquid separation late in their histories, as evidenced by highly evolved granitic caps on top of several large plutonic bodies (Johnson et al. 1990; Barnes et al. 2001; Miller and Miller 2002; Walker et al. 2007).

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