

On the Origin of Crystal-poor Rhyolites: Extracted from Batholithic Crystal Mushes

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The largest accumulations of rhyolitic melt in the upper crust occur in voluminous silicic crystal mushes, which sometimes erupt as unzoned, crystal-rich ignimbrites, but are most frequently preserved as granodioritic batholiths. After approximately 40–50% crystallization, magmas of intermediate composition (andesite–dacite) typically contain high-SiO₂ interstitial melt, similar to crystal-poor rhyolites commonly erupted in mature arc and continental settings. This paper analyzes the feasibility of system-wide extraction of this melt from the mush, a mechanism that can rationalize a number of observations in both the plutonic and volcanic record, such as: (1) abrupt compositional gaps in ignimbrites; (2) the presence of chemically highly evolved bodies at the roof of subvolcanic batholiths; (3) the observed range of ages (up to 200–300 ka) recorded by zircons in silicic magmas; (4) extensive zones of low P-wave velocity in the shallow crust under active silicic calderas. We argue that crystal–melt segregation occurs by a combination of several processes (hindered settling, micro-settling, compaction) once convection is hampered as the rheological locking point of the crystal–melt mixture (≥ 50 vol. % crystals) is attained. We constrain segregation rates by using hindered settling velocities and compaction rates as end-members. Time scales estimated for the formation of >500 km³ of crystal-poor rhyolite range from 10⁴ to 10⁵ years, within the estimated residence times of mushes in the upper crust ($>10^5$ years, largely based on U/Th and U/Pb dating). This model provides an integrated picture of silicic magmatism, linking the evolution of plutonic and volcanic systems until storage in the upper crust, where granitoids become the leftovers from rhyolitic eruptions.

KEY WORDS: *crustal evolution; igneous processes; rhyolite; silicic magmatism; two-phase flow*

INTRODUCTION

Establishing the processes and controls on the generation of magmas with SiO₂ content in excess of 65 wt %

(hereafter referred to as ‘silicic’) remains a major challenge of igneous petrology. In particular, the origin of rhyolites, which are typically the most crystal-poor, despite being the most viscous silicate liquids that the Earth produces, remains elusive. Combinations of two end-member mechanisms are commonly invoked to explain such magmas (Fig. 1a): (1) partial melting of crustal material; (2) fractional crystallization of a more mafic parent. Although systems dominated by crustal melting occur in some settings (e.g. peraluminous volcanics: Clemens & Wall, 1984; Munksgaard, 1984; Pichavant *et al.*, 1988a, 1988b; intra-oceanic rhyolites: Gunnarsson *et al.*, 1998; Smith *et al.*, 2003; igneous provinces related to continental break-up: Green & Fitz-Thomas, 1993; Milner *et al.*, 1995; Riley *et al.*, 2001), many geochemical studies suggest that fractional crystallization plays an important role in the generation of silicic magmas (e.g. Michael, 1983; Bacon & Druitt, 1988; Mahood & Halliday, 1988; Hildreth *et al.*, 1991; DePaolo *et al.*, 1992; Hildreth & Fierstein, 2000; Lindsay *et al.*, 2001; Clemens, 2003). This is particularly true for silicic volcanic rocks, which are known to have a more pronounced crystal fractionation signature than their plutonic equivalents (Halliday *et al.*, 1991; Fig. 1b).

For magmas to evolve by fractional crystallization, differential motion must occur between crystals and melt (two-phase flow, where all crystals are considered to be one effective phase). Despite numerous attempts, no consensus has emerged on the dominant process occurring in the viscous, silicic systems. Three influential hypotheses, all assuming an initially near-liquidus batch of silicate melt, are: (1) crystal settling, commonly advocated since Bowen (1928); (2) convective fractionation in a crystallizing, double-diffusive boundary layer (Chen & Turner, 1980; McBirney, 1980; Rice, 1981; McBirney *et al.*, 1985; Spera *et al.*, 1995); (3) the ‘solidification front

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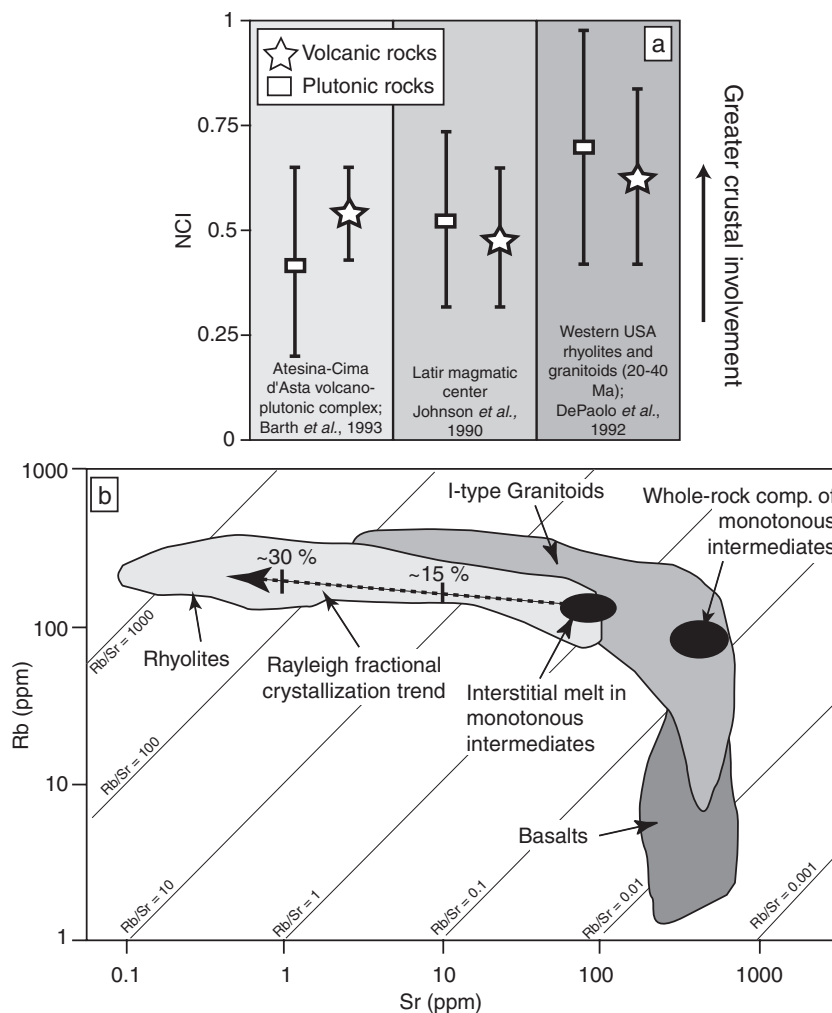


Fig. 1. (a) Variation of the NCI value (Neodymium Crustal Index; DePaolo *et al.*, 1992) for cogenetic volcanic and plutonic rocks from the Atesina–Cima d’Asta volcano-plutonic complex (Barth *et al.*, 1993), Latir magmatic center (Johnson *et al.*, 1990), as well as some 20–40 Ma rhyolites and granitoids from the western USA (DePaolo *et al.*, 1992). (b) Rb vs Sr content for magmatic rocks from several magmatic provinces (western North America, Andes, Great Britain, Australia; modified from Halliday *et al.*, 1991). Dashed line represents a crystal fractionation trend, using the concentration of Rb and Sr in the interstitial melt of the Fish Canyon Tuff (FCT; a typical monotonous intermediate) as initial value and typical bulk partition coefficients for these elements in FCT ($D_{Sr} = 12$, $D_{Rb} = 0.5$; Bachmann, 2001). Tick marks indicate 15 and 30% of Rayleigh fractional crystallization. The more restricted, but overlapping NCI range displayed by the volcanic sequences suggests that the erupted magmas are a subset of a consanguineous petrological suite, with both crystal fractionation and crustal assimilation playing a role in the generation of these magmas. However, the rhyolites record a stronger crystal fractionation signature (higher Rb/Sr ratios).

instabilities’ hypothesis of Marsh and coworkers [see Marsh (2002) for a review]. As we discuss below, none of these processes, taken alone, appears to satisfy the geological observations from voluminous rhyolites.

Another group of two-phase flow mechanisms, which require a higher crystal fraction (≥ 50 vol. % of crystals in the magma), involves the upward percolation of buoyant interstitial melt from a mush, as a result of either the exsolution of a gas phase (gas-driven filter-pressing, Anderson *et al.*, 1984; Sisson & Bacon, 1999), or compaction phenomena (e.g. McKenzie, 1984; Shirley, 1986; Philpotts *et al.*, 1996; Rabinowicz *et al.*, 2001; Jackson

et al., 2003). The genetic link between silicic volcanic units and shallow granitoids (e.g. Lipman, 1984; Halliday *et al.*, 1991; Fig. 1), as well as the number of spatially and temporally associated crystal-rich intermediate magmas and cogenetic rhyolites (e.g. Hildreth & Fierstein, 2000; Lindsay *et al.*, 2001), has led a number of workers to propose the following hypothesis for the generation of crystal-poor rhyolites (e.g. Bacon & Druitt, 1988; Sisson & Bacon, 1999; Hildreth & Fierstein, 2000): could silicic mushes in the upper crust, detected seismically under active calderas and preserved as batholiths or crystal-rich ignimbrites, be the sources for rhyolites? In this

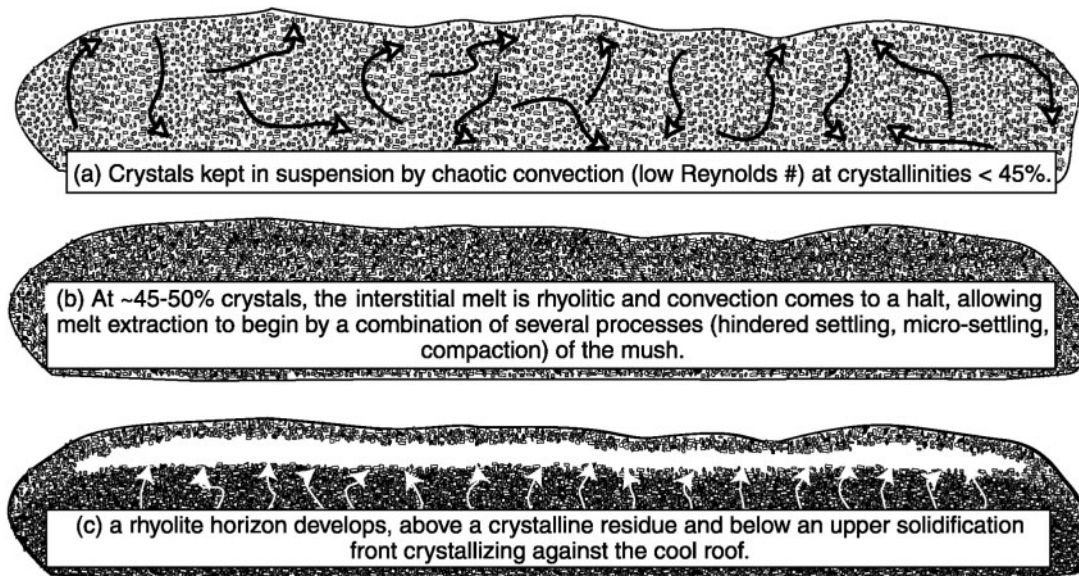


Fig. 2. Schematic illustration of the melt expulsion model for the generation of crystal-poor rhyolites. (a) A large magma body of intermediate composition with a crystal content lower than 45% is likely to retain crystals in suspension as a result of convective stirring. No speculation is made on the initial crystallinity of the mush. (b) At crystallinities of ~45–50%, the interstitial liquid is rhyolitic, and crystals start to form a framework. The associated yield strength impedes convection and allows melt extraction to begin, probably by a combination of several two-phase flow mechanisms (compaction, micro-settling, hindered settling). (c) A rhyolite horizon forms above a crystalline residue, and below a crystallizing roof layer.

study, we employ field, chemical and physical constraints to quantitatively assess whether rhyolites can be produced by this process of melt expulsion from homogeneous silicic mushes once crystals start forming a connected framework (~45–50 vol. % crystal; Fig. 2). Around this rheological transition from liquid to solid, the main mechanism for crystal–melt segregation changes from crystal settling to melt extraction by compaction. Thus, to cover a range of possible time scales, we used hindered settling velocities and compaction rate to show that differential motion between crystals and melt in static mushes close to their rheological transition can occur rapidly enough to form the most voluminous rhyolites (segregation rates of 10^{-1} – 10^{-3} km³/yr). We also suggest that gas-driven filter pressing related to exsolution and/or upward percolation of a low-density phase in these upper-crustal mushes (Sisson & Bacon, 1999; Bachmann & Bergantz, 2003) may enhance crystal–melt segregation but is not the dominant mechanism of melt expulsion from mushes.

THE MAKING OF RHYOLITES

Silicic crystal mushes: rhyolite nurseries

Kilometer-sized granodioritic batholiths are found in the roots of all mature continental arcs. They are so common that the average composition of the upper crust is essentially granodioritic (Taylor & McClelland, 1985). Despite this ubiquity, much controversy still surrounds their

origin (Bergantz & Dawes, 1994; Evans & Hanson, 1997; Martin *et al.*, 1997; Castro *et al.*, 1999; Patiño Douce, 1999; Barboza & Bergantz, 2000; Petford *et al.*, 2000); however, an exhaustive discussion of this conundrum is beyond the scope of this study. We begin with the fact that they exist and focus on the documentation of their magmatic youth provided by the giant, unzoned, crystal-rich, dacitic ignimbrites, referred to as monotonous intermediates (Hildreth, 1981). One of the critical observations made on these ‘erupted batholiths’ and other highly crystalline units of intermediate composition is that at the observed crystallinity (40–45% crystals), their interstitial liquid phase is chemically very similar to crystal-poor silicic magmas (Cashman & Blundy, 2000; Bachmann *et al.*, 2002; Schmitt *et al.*, 2003), making these crystal mushes a natural source and storage site for high-SiO₂ melts in the upper crust. Considering the volume of these units (10^3 – 10^4 km³), extraction of only ~10% of this interstitial melt would be enough to account for the largest crystal-poor rhyolites (10^2 – 10^3 km³). But is there a process capable of separating melt from crystals in these silicic mushes at a fast enough rate to form a voluminous rhyolite cap in geologically reasonable time scales?

Two-phase flow at high crystal fraction

A key assumption of this melt expulsion model is that enough crystallization occurs to produce rhyolitic interstitial melt and that system-wide convection is hampered (Fig. 2b). Importantly, both of these criteria converge at

crystal fractions >0.4 . First, as observed in numerous systems (e.g. Druitt & Bacon, 1989; Bachmann *et al.*, 2002; Schmitt *et al.*, 2003), magmas with intermediate whole-rock composition and >35 – 45 vol. % crystals contain interstitial rhyolitic melt. Second, convection ceases once a rigid crystal framework is attained, which is thought to occur at around 50–55% crystals (Rigid Percolation Threshold, RPT; Vigneresse *et al.*, 1996). The onset of yield strength in crystal–melt suspensions, thereby retarding convection, may even occur at lower crystal fraction (for plagioclase laths, the critical crystallinity ϕ_c is <0.3 ; Philpotts *et al.*, 1998; Saar *et al.*, 2001), even though the homogeneity in whole-rock composition of monotonous intermediates (crystal contents up to 40–45%) suggests that some convective stirring may occur almost up to the RPT.

At crystal fractions around the rheological transition from liquid to solid, the physics of two-phase flow is complex (e.g. Petford, 2003). It is unlikely that the mechanical properties of the crystal–melt mixture are perfectly uniform, leading to locally variable behavior and segregation rates. Moreover, work in the field of liquid-phase sintering (LPS; Niemi & Courtney, 1983) suggests that textural adjustments of individual grains and melt expulsion occur after the formation of a rigid skeleton. This theory has been applied to crystal mushes and referred to as ‘micro-settling’ by Miller *et al.* (1988). To average out local variability, and because segregation rates associated with micro-settling are difficult to estimate, we constrain the relative motion of interstitial melt and crystals using two end-member mechanisms: (1) particle settling in a dense suspension (hindered settling; Davis & Acrivos, 1985); (2) melt expulsion from a compacting porous medium (McKenzie, 1984; Shirley, 1986). Segregation rates for both compaction and hindered settling were calculated using equations (1)–(9) and are reported below, following an assessment of the two most important parameters in controlling the efficiency of segregation, the melt viscosity and the mush permeability [see also Petford (1995) for a discussion of grain size dependence on compaction rate].

Viscosity of high-SiO₂ melts

Since the work of Shaw (1972), water content has been identified as an important compositional variable in controlling magma viscosity. Following the recognition that silicic magmas are water-rich (commonly 6, but up to 8 wt % H₂O; e.g. Sisson & Bacon, 1999), the viscosity of high-SiO₂ rhyolites has been reassessed using recent experimental work (Baker, 1996, 1998; Dingwell *et al.*, 1996; Hess & Dingwell, 1996; Schulze *et al.*, 1996). In particular, the study of Scaillet *et al.* (1998) has shown that, over a wide range of water contents (1–8 wt %) and temperatures (650–1000°C), natural rhyolitic melts

Table 1: Units and variables

Symbol	Description	Value/units
k_ϕ	Permeability	m ²
ϕ	Porosity	0.4–0.5
r	Radius	m
K	Constant in permeability relationship	50–200
μ	Dynamic viscosity of melt	Pa s
ρ_s	Density of solid phase	kg/m ³
ρ_l	Density of liquid phase	kg/m ³
U_{hs}	Hindered settling velocity	m/s
U_{Stokes}	Stokes velocity	m/s
c	Crystallinity	0.5–0.6
δ_c	Compaction length	m
g	Acceleration due to gravity	9.81 m/s ²
ζ	Bulk viscosity of matrix	10 ⁻¹³ –10 ⁻¹⁵ Pa s
η	Shear viscosity of matrix	10 ⁻¹³ –10 ⁻¹⁵ Pa s
w_0	Relative velocity between melt and matrix	m/s
h	Height of mush layer	m
τ_0	Characteristic time for compaction equation	s
t_h	Time to reduce fluid in porous layer by a factor of e	s
h_m	Melt layer extracted in time t_h	m
H	Height of melt column	m
τ_{settle}	Settling time scale	s

(>70 wt % SiO₂) have viscosities clustering around 10^{4.5} Pa s, with a range of 10^{3.7}–10^{5.4} Pa s.

Permeability of silicic mushes

The permeabilities of naturally occurring porous media are difficult to assess. Several empirical equations have been published (e.g. Dullien, 1979), but a commonly used expression is the Blake–Kozeny–Carman equation (McKenzie, 1984; Barboza & Bergantz, 1998; Rabinowicz *et al.*, 2001),

$$k_\phi = \frac{\phi^3 r^2}{K(1 - \phi^2)} \quad (1)$$

where ϕ is the porosity, r is the grain size (radius), and K is a constant (~ 50 – 200 for porosities >0.1 and grain size of ≥ 0.5 mm; Rabinowicz *et al.*, 2001; Jackson *et al.*, 2003). Equation (1) has two degrees of freedom: the porosity and grain size of the system. For the hypothesis tested in this study (two-phase flow in high-porosity crystal mushes), we constrain porosity to a narrow window of ~ 0.4 – 0.5 . However, grain size is likely to vary. It is possible to obtain a minimum value of the

radius using petrographic observations of the dominant phenocryst phases (feldspars, hornblende, biotite, quartz, radius of the order of 1–3 mm) in lava flows and monotonous intermediates (Best *et al.*, 1989; Francis *et al.*, 1989; de Silva *et al.*, 1994; Bachmann *et al.*, 2002; Maughan *et al.*, 2002). However, these lava flows and the pumices found in monotonous intermediates, as well as granodioritic plutons, commonly contain megacrysts and glomerocrysts, up to several centimeters in diameter (Bateman & Chappell, 1979; Lipman *et al.*, 1997; Seaman, 2000; Bachmann *et al.*, 2002), suggesting that the effective grain size of the mushes may be larger (3–5 mm). Using $r = 3$ mm, the permeability is of the order of 10^{-8} m^2 , similar to recent estimates for high-porosity mush in the mantle (Rabinowicz *et al.*, 2001).

Hindered settling

Hindered settling provides a mechanism for clarification of dense suspension. Settling velocities of monodisperse particles in a static, non-dilute suspension (>1 vol. % particles) at low Reynolds number can be predicted by

$$U_{\text{hs}} = U_{\text{Stokes}} \cdot f(c) \quad (2)$$

with

$$f(c) = \frac{(1-c)^2}{(1+c^{1/3})^{5c/3(1-c)}} \quad (3)$$

where U_{hs} , the hindered settling velocity, equals U_{Stokes} , the Stokes settling velocity,

$$U_{\text{Stokes}} = \frac{2r^2g\Delta\rho}{9\mu} \quad (4)$$

where r is radius of the solid particle, g is the acceleration from gravity, $\Delta\rho$ is the density contrast between particle and melt, and μ is the dynamic viscosity of the melt, corrected by the factor $f(c)$, c being the crystal fraction (Barnea & Mizrahi, 1973). As seen in Fig. 3, U_{hs} drops sharply as crystallinity increases, and tends to zero at crystallinities >0.6 . Using settling rates based on monodispersed particles will give an outer bound; polydispersed suspensions are more realistic for magmas, but their settling behavior is poorly known (Davis & Acrivos, 1985).

The hindered settling velocities are calculated assuming constant porosity. However, in silicic mush piles, the porosity will tend to decrease downward as crystals settle. Thus, a question is whether this downward decreasing porosity will greatly reduce settling rates. As a first approximation, one can estimate the maximum vertical distance that crystals must sink from the top of the pile to account for the volumes of the largest erupted rhyolites ($\sim 10^2$ – 10^3 km^3). On the basis of caldera surface, horizontal extensions of large magma chambers are usually

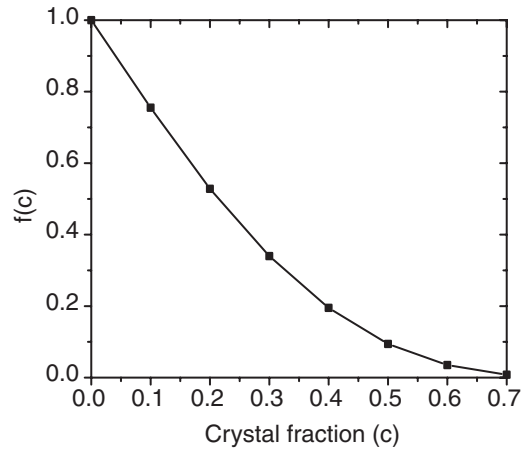


Fig. 3. Evolution of the empirical hindering function $f(c)$ given by Barnea & Mizrahi (1973), showing a sharp decrease of the settling velocity $U_{\text{hs}} = U_{\text{Stokes}}f(c)$ as crystal fraction (c) increases.

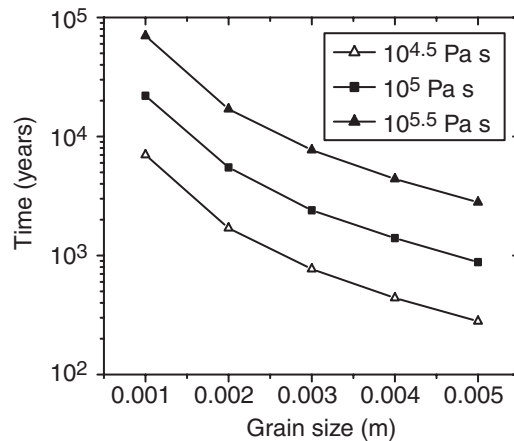


Fig. 4. Time scales required for crystals with a density of 2600 kg/m^3 (feldspar, quartz) and variable radii to sink 500 m in a silicic mush (crystal volume fraction is 0.5) with a melt density of 2300 kg/m^3 , and melt viscosities of $10^{4.5}$, 10^5 , and $10^{5.5} \text{ Pa s}$.

considered to be of the order of 1000 km^2 , and sinking distances of 0.5 km are sufficient to produce 500 km^3 of rhyolite. As mush piles are several kilometers thick, we can assume that the porosity will not change significantly in the upper 0.5 km, and accept the time scale required for a crystal to sink for 0.5 km as a reasonable estimate (Fig. 4).

Grain size and viscosity will be the most important parameters for a given porosity. Taking the upper range of melt viscosities as published by Scaillet *et al.* (1998; $10^{4.5}$ – $10^{5.5} \text{ Pa s}$), Fig. 4 shows that, for crystallinities of 50%, crystals with a density of 2600 kg/m^3 (approximately feldspar, quartz) and radii of 2–3 mm will take 10^3 – 10^4 years to sink 500 m in a rhyolite melt. Increasing the crystallinity to 60% increases the time scales by a factor of 2.7.

Compaction

Assuming that crystals form a deformable, purely viscous network (no account of viscoelastic behavior of the matrix, $De \ll 1$; Vasilyev *et al.*, 1998), time scales necessary for the separation of crystals and melt in compacting magmatic systems can be calculated (McKenzie, 1984, 1985; Shirley, 1986). The compaction length, δ_c , defined as the distance over which compaction occurs above an impermeable boundary, is

$$\delta_c = \left[\frac{\zeta + \frac{4}{3}\eta}{\mu} k_\phi \right]^{1/2} \tag{5}$$

where ζ and η are the effective bulk and shear viscosities of the matrix (both $\sim 10^{-13}$ – 10^{-15} Pa s for mush porosities ≥ 0.3 ; Rabinowicz *et al.*, 2001; Jackson *et al.*, 2003), ϕ is the porosity of the mush, and k_ϕ is the porosity-dependent permeability. w_0 , the relative velocity between melt and matrix (positive upward), is defined as

$$w_0 = k_\phi(1 - \phi)(\rho_s - \rho_l) \frac{g}{\mu\phi} \tag{6}$$

where ρ_s and ρ_l are the densities of the solid and liquid phases, μ is the dynamic viscosity of the melt and g is the acceleration due to gravity. τ_0 , the reference time, t_h , the time taken to reduce the total amount of fluid in the layer by a factor of e , h , the initial thickness of the mush layer, and h_m , the layer of melt extracted in time t_h , are defined as follows:

$$\frac{t_h}{\tau_0} = \frac{h}{\delta_c} - \frac{\delta_c}{h} \tag{7}$$

If $h \gg \delta_c$, then

$$t_h = \frac{\tau_0 h}{\delta_c} = \frac{h}{w_0(1 - \phi)} \tag{8}$$

$$h_m = h\phi(1 - e^{-1}). \tag{9}$$

Using the same range of viscosities as for hindered settling ($10^{4.5}$ – $10^{5.5}$ Pa s) and the same crystal volume fraction (0.5), time scales t_h [equation (8)] required to expel more than 500 km^3 of rhyolite at the top of the system for a grain size of 2–3 mm are of the order of 10^4 – 10^6 years, more than an order of magnitude greater than hindered settling (Fig. 5). At a crystal volume fraction of 0.6, time scales increase by a factor of 1.5.

DISCUSSION

The calculated compaction and hindered settling time scales bracket the segregation rates occurring in silicic

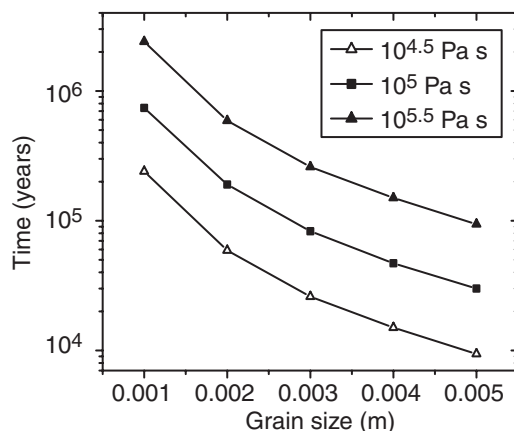


Fig. 5. Time scales required for a 2000 m crystal mush with an average crystal density of 2900 kg/m^3 (average crystal density from the Fish Canyon magma; Whitney & Stormer, 1985), a melt density of 2300 kg/m^3 , and a porosity of 0.5 to produce a $>500 \text{ km}^3$ crystal-poor cap with melt viscosities of $10^{4.5}$, 10^5 , and $10^{5.5}$ Pa s.

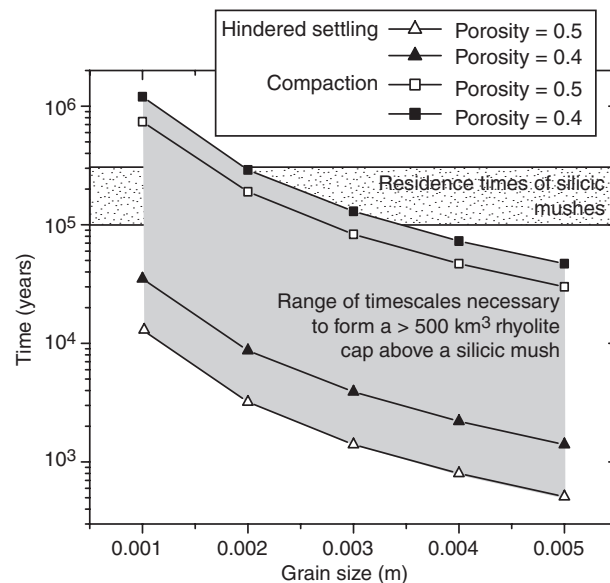


Fig. 6. Time required to segregate $>500 \text{ km}^3$ of crystal-poor rhyolite from a silicic mush by compaction and hindered settling as a function of the grain size of the mush (radius of solid particles) using a melt viscosity of 10^5 Pa s, and porosities of 0.4 and 0.5. Density contrast between melt and crystals is 300 kg/m^3 for hindered settling (using feldspar or quartz densities) and 600 kg/m^3 for compaction (using average crystal density of the Fish Canyon magma; Whitney & Stormer, 1985).

mush with crystal fraction $\geq 0.45\%$. Figure 6 reports data for both processes and suggests a range of possible values. Using 10^5 Pa s as the viscosity of the melt [close to the highest viscosities reported by Scaillet *et al.* (1998)], time scales from 10^4 to 10^5 years are to be expected. These time scales give segregation rates of $\sim 5 \times 10^{-2}$

to $5 \times 10^{-7} \text{ km}^3/\text{yr}$ (10^4 – 10^5 years for rhyolite caps of $\sim 500 \text{ km}^3$); this range agrees with the slightly slower magma output rates in silicic systems estimated by Trial & Spera (1990; 10^{-2} – $10^{-4} \text{ km}^3/\text{yr}$). These magma output estimates of Trial & Spera (1990), calculated using erupted volumes divided by the repose time in many volcanic systems, must include every step in the generation of evolved magmas, and, thus, provide minimum values.

These segregation rates do not take into account the effect of shear-induced expansivity of porous media and the possible melt expulsion (or ‘seepage’) related to it (Petford & Koenders, 2003). We consider that the mush as a whole is immobile with respect to its wall rocks, and that no shear stress other than tectonic loading is acted upon it. Estimated melt flow rates as a result of tectonic loading (10^{-11} – 10^{-12} m/s ; Petford & Koenders, 2003) are similar to or slower than those caused by compaction (10^{-9} – 10^{-10} m/s , this study) and should not influence significantly the results presented here. Had tectonic loading any effect on melt segregation, it would accelerate it, implying that the segregation time scales presented in Fig. 6 are upper bounds.

Longevity of silicic mushes

For crystal–melt segregation to occur, the estimated segregation time scale of 10^4 – 10^5 years has to be shorter than the residence times of high-porosity mushes in the crust. The pioneering work of Halliday *et al.* (1989) on the longevity of silicic magmas suggested residence times up to 700 kyr for Glass Mountain lavas on the basis of Rb/Sr isochrons. These protracted lifetimes, first taken to be residence times of magmas near their liquid, were criticized on a number of grounds, (e.g. extremely delicate heat balance to keep the system liquid for such long periods—Sparks *et al.*, 1990; contamination from wall-rock, giving irrelevant Rb/Sr ages—Knesel *et al.*, 1999). However, starting with Mahood (1990), the idea that magmas could oscillate thermally, allowing retentive crystals to record a long residence time, has been progressively gaining acceptance. Recent advances in *in situ* U/Th and U/Pb dating of zircons have confirmed that minerals in rhyolites can record crystallization ages up to 200–300 kyr prior to eruption (Reid *et al.*, 1997; Brown & Fletcher, 1999; Vazquez & Reid, 2002). This is also supported by the near-equilibrium $^{238}\text{U}/^{230}\text{Th}$ ratios measured in most rhyolites, suggesting differentiation time scales of $>10^5$ years (Hawkesworth *et al.*, 2000; Heumann *et al.*, 2002; Reagan *et al.*, 2003). Independent estimates of magma residence times, such as time scales required to build batholiths with known accumulation rates (Jellinek & DePaolo, 2003) or cool them by conduction (Spera, 1980; Koyaguchi & Kaneko, 1999) also range from 10^5 to 10^6 years.

Such extended residence times would allow enough crystal–melt segregation to form the most voluminous rhyolites, even by slow compaction alone (except for very small grain sizes), if magmas are preferentially stored as crystal mushes. As the model we are proposing requires that high-porosity mushes exist for 10^4 – 10^5 years, the evolution of magma crystallinity with time must be determined. Several arguments, listed below, suggest that crystal mush is the long-lived magma storage state in the crust and provide some justification as to why we did not perform conductive cooling models to assess the importance of cooling on segregation rates.

(1) There is little evidence for the presence of large bodies of crystal-poor intermediate to silicic magmas in the crust. Apart from rhyolites, most intermediate to silicic eruptive products contain >20 vol. % crystals (Marsh, 1981; Ewart, 1982), even though they always tap the least viscous, and, thus, least crystalline magma available in the reservoir. Moreover, high-resolution tomography beneath active calderas (Long Valley—Dawson *et al.*, 1990; Weiland *et al.*, 1995; Valles—Steck *et al.*, 1998; Yellowstone—Miller & Smith, 1999; Central Andes—Schmitz *et al.*, 1999; Zandt *et al.*, 2003; Toba—Masturyono *et al.*, 2001) has not been able to demonstrate the existence of crystal-poor magmas in the present-day crust. Instead, the attenuation of P-wave velocities from 15 to 40% at depths of 5–30 km has been interpreted as a reflection of large, partially molten zones in the middle to upper crust.

(2) Near-liquidus magmas can convect, and therefore cool faster than static mushes, which are constrained by conductive heat loss. However, it must be noted that turbulent convection, and resulting cooling in 10^2 – 10^4 years (Huppert & Sparks, 1988; Koyaguchi & Kaneko, 2000), appears unlikely for cold and viscous rhyolites (for which Reynolds numbers are orders of magnitude below the onset of turbulence).

(3) Lifetimes of mushes can be prolonged by thermal input from below. Recent work on both crystal-rich volcanic rocks and silicic plutons (e.g. Matthews *et al.*, 1999; Robinson & Miller, 1999; Murphy *et al.*, 2000; Bachmann *et al.*, 2002) has shown that ‘rejuvenation’ events (progressive reheating and partial remelting) can occur in mushes following the injection of hotter magma. They are, therefore, expected to survive longer than estimated by monotonic, conductive cooling models, and can remain around the RPT for significant periods of time.

Supporting evidence in the rock record for melt extraction from mushes

Volcanic systems

Spatial and chronological proximity between rhyolites and intermediate crystal-rich magmas in many systems supports the idea of interstitial melt extraction from silicic

mushes. Occurrences of comagmatic crystal-poor and crystal-rich magmas are common in the volcanic record. Young examples in active volcanic arcs include: (1) the Atana–Toconao system in the Altiplano–Puna Ridge Complex, Central Andes, where the rhyolitic Toconao ignimbrite has been interpreted as the evolved cap of the crystal-rich dacitic Atana ignimbrite (Lindsay *et al.*, 2001; Schmitt *et al.*, 2003); (2) the interbedding of crystal-rich dacites and rhyolites in the Toba Tuffs, Sumatra (Chesner, 1998); (3) the Whakamaru group ignimbrites, in the Taupo Volcanic Zone, New Zealand, where a stack of cogenetic, low- and high-SiO₂ rhyolitic ignimbrites show large variations in crystallinity (Brown *et al.*, 1998); (4) two canonical examples of compositional gaps in ignimbrites, the 1912 VTTS eruption, on the Alaska Peninsula, and the Crater Lake eruption, Oregon, in which chemical and isotopic affinities between the early erupted rhyolite and the late-erupted, crystal-rich, more mafic magma suggest expulsion of the rhyolite from the andesite–dacite mush (Bacon & Druitt, 1988; Druitt & Bacon, 1989; Hildreth & Fierstein, 2000). Pyroclastic units characterized by early erupted, crystal-poor rhyolite followed by crystal-rich, more mafic magma may actually be the rule rather than the exception in mature continental arcs: multiple ignimbrites of the well-studied Oligocene sequences in the San Juan volcanic field, Colorado (Steven & Lipman, 1976; Lipman, 2000), the Great Basin ash-flow province (Maughan *et al.*, 2002), and the Southwestern Nevada Volcanic Field (SWNVF; Mills *et al.*, 1997; Bindeman & Valley, 2003) display this pattern. Even peraluminous magmas, which are probably crustal melts, seem to have undergone extensive crystallization and expelled comagmatic, low Sr–Ba–Eu, crystal-poor melt that erupted concurrently with crystal-rich ash-flows (Pichavant *et al.*, 1988*a*, 1988*b*). Lavas flows, such as those erupted at the Medicine Lake center, northern California, can also be zoned from rhyolite to comagmatic, crystal-rich andesite (Brophy *et al.*, 1996; Grove *et al.*, 1997).

Another compelling line of evidence for crystal-poor silicic melts being expelled from a large mushy zone in the upper crust comes from clusters of small-volume rhyolitic lava flows that tapped the same ‘magma chamber’, but were erupted over extended time intervals (~100–500 kyr) and over wide areas (100–1000 km²). Notable examples include the Glass Mountain Rhyolite (Metz & Mahood, 1991), the Coso volcanic field (Bacon *et al.*, 1981; Manley & Bacon, 2000), the Taylor Creek Rhyolite (Duffield & Dalrymple, 1990), the Bearhead Rhyolite (Justet & Spell, 2001), and the Central Plateau Member rhyolites (Hildreth *et al.*, 1991; Vazquez & Reid, 2002). The fact that these cogenetic, rhyolitic lavas cover large areas and erupted over tens or hundreds of thousands of years implies a long-lived, laterally extensive and well-mixed magma reservoir, but the relatively small

volumes of individual units suggest the absence of a large, continuous, crystal-poor body over the entire eruptive period. Rather, these ‘leaks’ tapped localized horizons (or cupolas) or rhyolitic melt that formed by periodic segregation events from an underlying, voluminous and homogeneous mush. In many such cases, however, the crystal-rich domain presumed below rarely or never is tapped. This observation, in concert with the fact that these rhyolites are commonly small and have very high Rb/Sr ratios, may point out areas where the mush has reached a fairly high crystallinity (~70–80 vol. % crystals, on the basis of Rayleigh fractional crystallization with typical partition coefficients for Rb and Sr in these systems; Fig. 1b), reducing the segregation rate and preventing the crystal-rich magma from erupting. Their small volumes may also allow more efficient open-system degassing upon ascent (e.g. Eichelberger *et al.*, 1986; Jaupart, 1998), thus preventing an explosive eruption.

Plutonic systems

Despite the difficulty of identifying specific or progressive magmatic events in large plutonic suites as a result of the time-integrated histories that they record (e.g. multiple intrusive events, post-intrusion deformation, subsolidus processes), plutons also provide support to the hypothesis of melt extraction from crystal mushes. One of the strongest arguments is the preservation of this melt as chemically evolved horizons at the roof of large batholiths (Nozama, 1983; Michael, 1984; Pitcher *et al.*, 1985; Dilles, 1987; Johnson *et al.*, 1989; Barnes *et al.*, 2001; Miller & Miller, 2002). Such horizons (or cupolas) have been interpreted as originating from the underlying, coarser, and more mafic magmas, mainly based on the fact that their composition is consistent with a formation by crystal fractionation from the underlying intermediate magma (Johnson *et al.*, 1989; Barnes *et al.*, 2001; Miller & Miller, 2002). Evidence for melt escape or cumulus textures in silicic plutonic bodies, without the preservation of caps, has also been reported elsewhere (Shearer & Robinson, 1988; Mahood & Cornejo, 1992; Weinberg *et al.*, 2001; Wiebe *et al.*, 2002).

These cumulus textures and melt escape features seem in contradiction with the nearly closed-system evolution and ubiquitous crystal recycling recorded by many silicic plutons [e.g. *in situ* fractional crystallization of Michael (1984)]. Whole-rock compositional homogeneity of the pluton coupled with core-to-rim depletion in Sr, Ba and Ca in plagioclase greater than those that can be attributed to pressure and temperature alone (e.g. Michael, 1984; Blundy & Shimizu, 1991) imply a well-stirred system that records progressive depletion in compatible elements from the interstitial melt as crystallization proceeds. The model of melt expulsion from batholithic crystal mushes predicts that only a small fraction

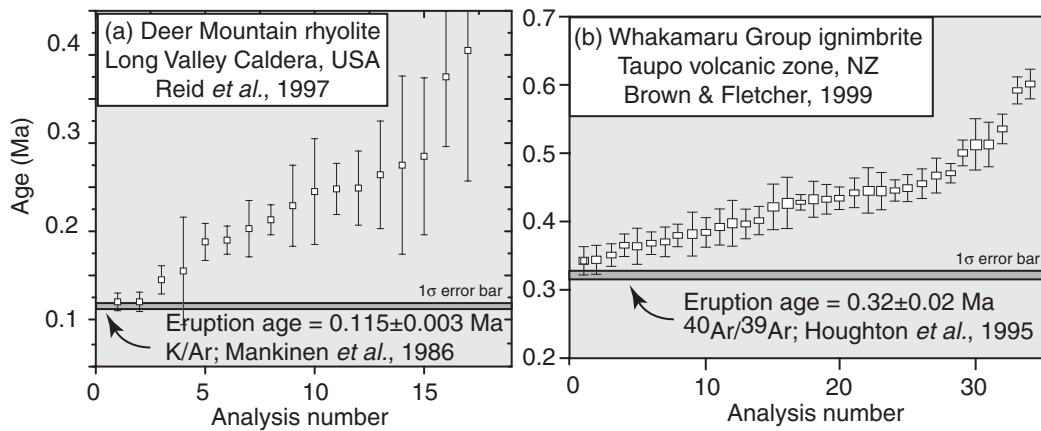


Fig. 7. Ages of single zircons from (a) the Deer Mountain rhyolite (Reid *et al.*, 1997), and (b) the Whakamaru ignimbrite (Brown & Fletcher, 1999), compared with their published K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages (Mankinen *et al.*, 1986; Houghton *et al.*, 1995). The large MSWDs (>8) of both datasets imply a range of zircon ages in excess of analytical precision, suggesting extended periods of zircon crystallization (up to >200 kyr prior to eruption).

(≤10 vol. %) of interstitial liquid is necessary to form the largest erupted rhyolites. Limited melt loss from a mush is capable of reconciling cumulus textures and melt escape features with these *in situ* crystallization trends in plutons (see also Langmuir, 1989, fig. 3).

Extended range of crystal ages

One of the most appealing aspects of melt expulsion from a long-lived mush is the fact that it predicts the extended age range recorded by zircons in silicic magmas. Long crystallization intervals (>200 kyr) are implied by several studies (e.g. Reid *et al.*, 1997; Brown & Fletcher, 1999; Fig. 7), which show large mean standard weighted deviations (MSWD) in the datasets, indicating a scatter in excess of analytical precision. Melt expulsion from the mush is likely to entrain some phenocrysts, in particular small ones such as zircon, as settling velocities of zircon crystals with radii of 10–50 μm in a rhyolitic melt ($\mu = 10^5 \text{ Pa s}$) are of the order of 10^{-10} – 10^{-12} m/s , slightly slower than the melt expulsion velocities in a compacting mush calculated here ($\sim 10^{-9}$ – 10^{-10} m/s). Therefore, the crystal-poor cap inherits crystals that have started growing in the mush, prior to extraction. This model also predicts that smaller crystals should record ages close to the eruption age, whereas the larger ones should have a longer crystallization interval. Such grain size dependent zircon ages were observed by Brown & Fletcher (1999) in the Whakamaru group ignimbrites.

Alternative mechanisms of two-phase flow in magmas

Several models of two-phase flow leading to differentiation of magmas have been proposed. Crystal settling from

near-liquidus magmas has often been invoked, following the influential works of Grout (1918), Bowen (1928) and Wager & Brown (1967). However, the recognition that sizeable volumes of near-liquidus magma stored in a colder environment are destined to convect at low crystal content led to the realization that settling velocities using Stokes' Law [equation (4)] were too fast. Settling velocities in convecting systems were, therefore, assessed. In particular, Martin & Nokes (1988) proposed that a characteristic time scale for crystal settling in convecting magmas is

$$\tau_{\text{settle}} = \frac{9H\mu}{2g\Delta\rho r^2} \quad (10)$$

where τ_{settle} is the settling time scale, H is the chamber height, μ is the magma viscosity, g is the acceleration due to gravity, $\Delta\rho$ is the density contrast between crystals and magma, and r is the radius of crystals. For $\mu = 10^5 \text{ Pa s}$, $\Delta\rho = 500 \text{ kg/m}^3$ and $r = 10^{-3} \text{ m}$, $\tau_{\text{settle}} = 6600 \text{ years}$, indicating that some settling should occur over time scales $>10^4 \text{ years}$. However, the experiments of Martin & Nokes (1988) were performed at particle concentrations of <0.01 wt %, which are unrealistically low for magmas. Koyaguchi *et al.* (1990) have looked at settling in slightly more dense suspensions (0.3 wt % particle; $\sim 0.1 \text{ vol. \%}$), which is still low compared with magmas, as volcanic rocks have 1–50 vol. % crystals, with an average of $\sim 25\%$; Ewart, 1982), but noticed a significant change in the behavior of the system. Although settling removes a fraction of the particles, cycles of overturn occur, remixing most of the layers periodically and increasing clarification time. Settling rates in convecting fluids at geologically realistic crystal fractions (>2–5 vol. % particles) are poorly known, but

it appears likely that equation (10) still provides an overestimation.

Another argument in support of low settling rates in convecting magmas involves the Stokes number of the crystals. The Stokes number (S_T) relates the response time scale of the crystal to the response time scale of the fluid (e.g. Burgisser & Bergantz, 2002) and allows an assessment of the behavior of a particle in a fluid. At $S_T \gg 1$, particle motion is not dictated by the motion of the surrounding liquid, and will follow its own trajectory (settling in this case). In contrast, at $S_T \ll 1$, a particle will closely follow the fluid. As the S_T is vanishingly small for nearly any conditions in silicic magmatic systems, as a result of the small size of the crystals and the high viscosity of the melt ($\sim 10^{-11}$ for 3 mm crystals with convective velocities of 1 m/yr), crystal settling is predicted to be very slow as long as any convective currents subsist.

Observations from large silicic systems agree with the inference that near-liquidus crystal settling is not important in the differentiation of silicic magmas. First, the striking whole-rock homogeneity of the monotonous intermediates (and of the granodioritic batholiths), together with the extremely complex zoning patterns of the crystals they contain, requires magma stirring by convective movements, even at crystal contents up to 45% and melt viscosities of $>10^5$ Pa s (Bachmann & Dungan, 2002; Bachmann *et al.*, 2002; Maughan *et al.*, 2002). Second, evidence for *in situ* crystallization trends in large granitoid bodies (e.g. Michael, 1984) and relict phenocryst cores (particularly obvious in plagioclase) in plutons and volcanic rocks (Blundy & Shimizu, 1991; Robinson & Miller, 1999; Bachmann *et al.*, 2002; Miller & Miller, 2002) requires a well-stirred suspension and recycling of crystals by scouring and re-entrainment of cumulus crystals.

An alternative hypothesis of differential motion between liquid and crystals in near-liquidus magmas is residual melt extraction from a crystallizing, double-diffusive boundary layer (Chen & Turner, 1980; McBirney, 1980; Rice, 1981; Huppert & Sparks, 1984; Spera *et al.*, 1984, 1995; McBirney *et al.*, 1985). The 'sidewall crystallization' hypothesis was particularly influential because it provided an explanation for the zoning pattern observed in silicic ignimbrites. In the case of bimodal magmatic fields (e.g. Iceland), and basaltic sills (e.g. Peneplain Sill, McMurdo Dry Valleys, Antarctica), a process involving instabilities within a solidification front nucleating on the sidewalls and growing inward (e.g. Marsh, 2002), has been proposed. Some melt extraction from a crystallizing sidewall boundary layer may occur in magmatic systems, but we suggest that it is not important in producing the large volumes of rhyolites in continental settings for several reasons, as follows.

(1) Sidewall crystallization requires that crystals are removed from the magma as it evolves. Hence, it is

difficult to explain the presence of old, and complexly zoned phenocrysts in silicic magmas, which requires that early crystallized material does not become isolated in a boundary layer, but remains in the dominant volume or is periodically recycled.

(2) Crystalline boundary layers (or solidification fronts) rarely appear to be preserved in the rock record. Concentric zoning, recognized in a large number of plutons (Bateman & Chappell, 1979; Halliday *et al.*, 1984; Stephens, 1992, 2001), has been used as evidence for sidewall crystallization (e.g. Sawka *et al.*, 1990; Verplanck *et al.*, 1999). However, at least two observations disagree with the interpretation that concentric zoning results from inward crystallization of a large magma body. First, it requires the unlikely assumption that the whole pluton was crystal-poor at some time during its evolution. Although small pockets of 'liquid', with crystal fractions <0.4 certainly existed throughout its hyper-solidus history, the concentric zoning could not represent the crystalline margins of a giant (often >1000 km³) pool of crystal-poor intermediate to silicic magma. Second, the presence of mafic enclaves near the margins of concentrically zoned plutons (e.g. Sawka *et al.*, 1990; Didier & Barbarin, 1991; Molyneux & Hutton, 2000) is not consistent with a mechanically isolated crystallizing boundary layer. Alternatively, concentric zoning could be explained by telescoping of compositionally variable intrusions (John, 1988; Hecht & Vigneresse, 1999; Coulson *et al.*, 2002; Tsuboi & Suzuki, 2003).

(3) Large silicic magmas bodies have sill-like aspect ratios, and producing a voluminous cap of differentiated melt by this mechanism appears difficult, as inclined walls constitute only a small fraction of the chamber surface (e.g. de Silva & Wolff, 1995). Melt escaping from the floor is not a viable alternative, as it would not be able to cross the dominant magma volume without being re-assimilated (Jellinek *et al.*, 1999).

(4) Although the sidewall crystallization model assumes that the presence of a cap over a denser, convecting dominant volume is a stable configuration (Fig. 8a), numerical and analog experiments have suggested otherwise (Oldenburg *et al.*, 1989; Davaille, 1999a, 1999b; Davaille *et al.*, 2002; Gonnermann *et al.*, 2002), as a result of the entrainment and mixing occurring at the interface between the two magmas. Scaling theory (Davaille *et al.*, 2002; Gonnermann *et al.*, 2002) yields a volumetric entrainment rate, which depends primarily on the density contrast (B value; Fig. 8b) and Rayleigh number. Figure 8c and d shows, for a conservative surface of 100 km², that the entrainment rate can be higher than the rate at which evolved magma is generated (e.g. Trial & Spera, 1990), suggesting that the less dense cap can be re-assimilated as fast as it is formed.

The model we are proposing for the generation of continental rhyolites (Fig. 2) is similar to the model

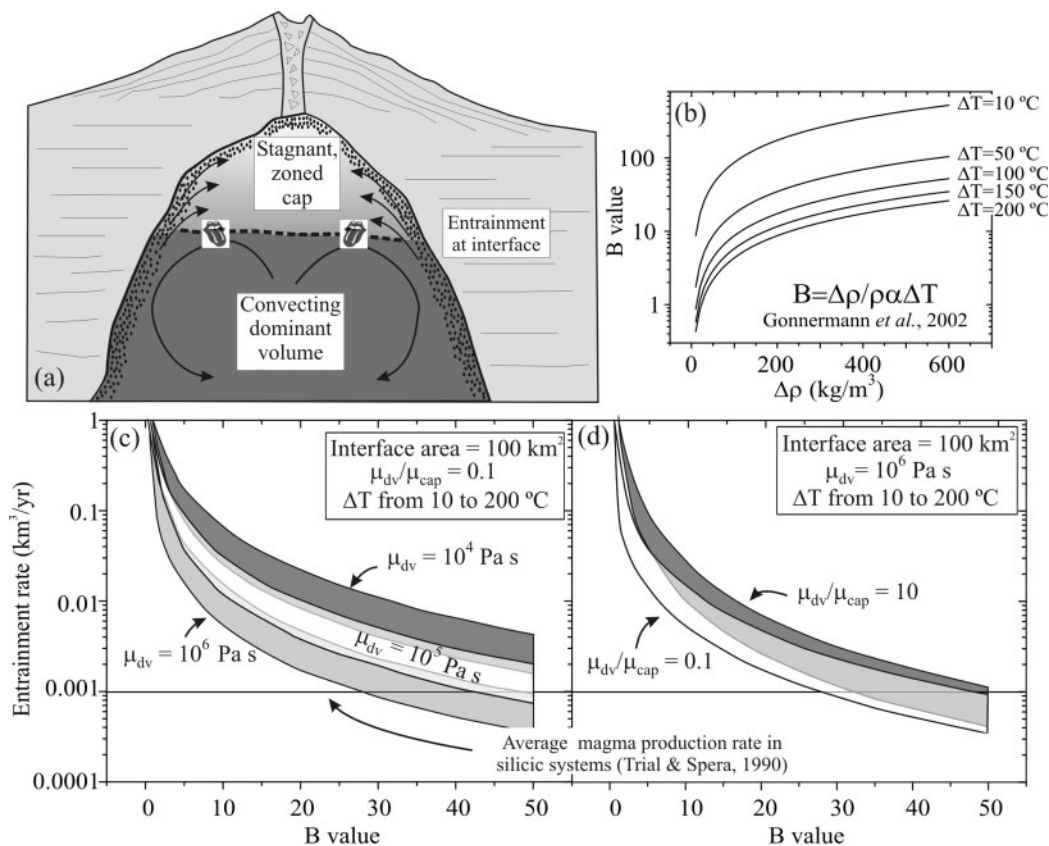


Fig. 8. Upward melt migration of light, residual liquid from a crystallizing, double-diffusive boundary layer is a popular, but problematic, hypothesis for shallow magmatic differentiation [schematic diagram in (a), modified from McBirney (1993)]. This model implies a stably stratified magma chamber, with a stagnant cap overlying the convecting, denser parent magma. In such a system, re-entrainment of the cap by the convecting interior (or dominant volume; dv) is known to occur as a result of viscous coupling at the interface (e.g. Davaille, 1999a; Davaille *et al.*, 2002; Gonnermann *et al.*, 2002). Quantification of the entrainment rate Q for an interface of 100 km² as a function of B is shown for various viscosities of the dominant volume (c), and for different viscosity ratios [<50 , (b)], the entrainment rate is as fast as or faster than an estimate of the rate at which evolved magmas are generated (Trial & Spera, 1990), suggesting that convecting 'stably stratified chambers' are ephemeral.

advocated by Philpotts and co-workers for large mafic bodies (Philpotts *et al.*, 1996; Meurer & Boudreau, 1996; Boudreau & Philpotts, 2002), although we suggest that compaction is not the only two-phase flow mechanism occurring in silicic mushes. As the rheological transition from melt to solid is a complex process, which varies as a function of system properties, strain rate, and shapes of particles (Lejeune & Richet, 1995; Vigneresse *et al.*, 1996; Barboza & Bergantz, 1998; Petford, 2003), it is likely that a large crystal mush encompasses regions slightly above the RPT, and regions just below. Therefore, we propose that segregation occurs as a combination of multiple processes (hindered settling, micro-settling, compaction), once convective currents become negligible. As discussed by Philpotts *et al.* (1996), the crystal-poor layer forming above the mushy region is probably overlain by a crystalline boundary layer at the roof. This hanging crystalline ceiling, denser than the underlying magma, could not

only account for dripping crystal–melt instabilities, producing the trace element and crystallinity gradients recorded in high-SiO₂ rhyolites, but also shed nearly solidified lumps as it becomes locally unstable, providing an explanation to the presence of cogenetic xenoliths in some ash-flow tuffs (e.g. Bacon, 1992; Lowenstern *et al.*, 1997).

Another mechanism potentially able to induce differential motion between crystals and melts in porous media is gas-driven filter pressing, as first proposed by Anderson *et al.* (1984) and expanded by Sisson & Bacon (1999). As many silicic mushes have water-rich interstitial melts (4–6 wt % H₂O on average, but up to 8 wt %), and are commonly stored in the upper crust (5–15 km), a significant fraction of them will reach volatile saturation and exsolve a low-density, low-viscosity fluid phase (e.g. Wallace *et al.*, 1995; Wallace, 2001). Sisson & Bacon (1999) suggested that overpressure following exsolution

of this gas phase could drive some interstitial melt out of the porous matrix. This process might participate in segregating melt from crystalline residue in the water-rich and shallow systems, but some of our recent work (Bachmann & Bergantz, 2003) suggests that it may be rather inefficient in expelling large quantities of rhyolite melt, and it could certainly not explain the generation of rhyolite in relatively dry settings, such as those in the Yellowstone system. We therefore propose that it facilitates segregation in certain situations, but is not the dominant mechanism.

Compositional gradients in magma chambers

Compositional gradients between early erupted and late-erupted material is a ubiquitous observation in ignimbrites. Simplifying the classification of Hildreth (1981) on zoning patterns in pyroclastic deposits, two general classes can be distinguished: (1) monotonic gradient, mainly apparent in trace element concentration (e.g. Bishop Tuff; Hildreth, 1979); (2) abrupt compositional gap, with major element variations up to several weight percent (e.g. VTTS; Hildreth & Fierstein, 2000), often associated with the observed lack of units with intermediate composition in volcanic sequences (the 'Daly Gap'). As this second zoning pattern commonly appears to be related to an abrupt change in crystallinity of the deposits (as noted above), our model of melt expulsion from a crystal mush provides a reasonable explanation for the preservation of such a gap, as already suggested by Thompson (1972) and Brophy *et al.* (1996). However, this model does not provide an explanation for the development of a monotonic gradient.

Although a detailed assessment of the potential mechanisms leading to the development of monotonic chemical gradients in rhyolites is beyond the scope of this study, we believe that our melt expulsion model is not incompatible with the formation of this type of gradient. We propose that they are separate processes, with chemical zoning developing after melt expulsion, and suggest that low Reynolds number convection in the growing cap, coupled with small amounts of wall-rock assimilation (<1 wt %), may provide a plausible model reconciling most characteristics of high-SiO₂ rhyolites, such as the Bishop Tuff (e.g. Hildreth, 1979; Wilson & Hildreth, 1997; Anderson *et al.*, 2000; Bindeman & Valley, 2002; Heumann *et al.*, 2002), and Bandelier Tuff (e.g. Wolff *et al.*, 1999, 2002). As crystal–melt segregation occurs, weak convection in the crystal-poor cap is predicted to begin once thicknesses of a few tens of meters are reached. Unstable thermal boundary layers are estimated to be of the order of a few meters thick, even for very small thermal gradients (Martin *et al.*, 1987; Couch *et al.*, 2001). Convective instabilities can also be created

by dripping of solid–liquid mixtures as crystallization proceeds at the roof (Bergantz & Ni, 1999). Because of the high viscosity of the magma and the conduction-controlled heat loss, convective currents are likely to remain slow, leading to a density stratification, such as roofward decrease in crystals and increase in volatile content (e.g. Jellinek *et al.*, 1999). The trace element gradient recorded in the interstitial melt and melt inclusions (early erupted material has a stronger crystal fractionation signature; Hildreth, 1979; Knesel & Davidson, 1997; Anderson *et al.*, 2000) may be explained by slow 'crystal settling' induced by dripping solid–liquid mixtures from the roof (Bergantz & Ni, 1999), depleting the upper and early extracted rhyolite of its compatible elements.

CONCLUSIONS

At crystallinities $\geq 40\%$, voluminous crystal mushes, preserved as batholiths in all continental arcs, and sometimes erupted as unzoned, crystal-rich ignimbrites, are known to store vast amounts of high-SiO₂ melt, similar in composition to the crystal-poor rhyolites erupted in continental settings. Extraction of this interstitial liquid to form rhyolitic caps has rarely been considered (however, see Johnson *et al.*, 1989; Barnes *et al.*, 2001; Miller & Miller, 2002), because of the common assumption that the high viscosities and low permeabilities of silicic systems would prevent sufficiently high segregation rates to produce the largest rhyolites. Recognizing that silicic bodies can go through a sustained, high-porosity (0.4–0.5), but static stage, has motivated a re-evaluation of efficient crystal–liquid separation in these mushes. Using hindered settling velocities and compaction rates at porosities of 0.4–0.5 to constrain a possible range of segregation rates, the production of the largest known crystal-poor rhyolites can occur in 10^4 – 10^5 years, well within the range of longevities inferred for mushes in the crust. These segregation rates can even be accelerated by gas-driven filter pressing in volatile-rich systems (Sisson & Bacon, 1999; Bachmann & Bergantz, 2003).

This model provides an integrated view of silicic magmatism, in agreement with observations documenting the common magmatic evolution of shallow plutons and volcanic systems (Lipman, 1984; Halliday *et al.*, 1991). Not only are plutons and volcanic rocks spatially and chronologically associated in magmatic centers preserving thick crustal sections (e.g. Latir center in northern New Mexico; Johnson *et al.*, 1990; Atesina–Cima d'Asta complex, Italy; Barth *et al.*, 1993) but petrological indicators, such as the Neodymium Crustal Index (NCI, DePaolo *et al.*, 1992), even suggest that volcanic rocks are a subset of consanguineous magmatic sequences (Fig. 1a), in accord with the low extrusive/intrusive ratios (e.g. Crisp, 1984). However, because of the generally

stronger crystal fractionation signature displayed by silicic volcanic units (Fig. 1b), the models proposed for the generation of granitoids have been inclined towards crustal melting, whereas crystal fractionation was favored for volcanic rocks. Our model of melt expulsion from crystal mushes reconciles these views by coupling their evolution until shallow storage, where melt extraction from crystal mush occurs, imparting a more pronounced crystal fractionation signature in the erupted portions of the systems.

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