

Crystal Capture and Settling in Sheet-like Silicic Magma Chambers Consisting of Eutectic Melt and Pre-existing Crystals

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Crystal Capture and Settling in Sheet-like Silicic Magma Chambers Consisting of Eutectic Melt and Pre-existing Crystals

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Abstract

In a solidifying magma chamber, settling crystals are captured by an advancing solidification front. The effects of the crystal capture process on the compositional structure of a silicic magma chamber are investigated based on a theoretical model. Previous studies have shown that the capture process causes an S-shaped vertical distribution of captured crystals especially in mafic intrusive bodies. However, the present results show that for a high-silica magma body, consisting of eutectic melt and pre-existing crystals, the vertical concentration of captured crystals changes linearly and a clear compositional gap exists at certain depths.

Keywords: crystal capture, solidification front, magma chamber, high-silica magma

1. Introduction

Crystal settling plays a fundamental role in controlling the compositional evolution of a magma chamber regardless of the presence or absence of convection. Geological, experimental, and theoretical studies have revealed how petrological and geochemical features of intrusive suites are controlled by crystal settling (Gray and Crain, 1969; Fujii, 1974; Marsh, 1988; Martin and Nokes, 1988; Koyaguchi et al., 1990; Mangan and Marsh, 1992; Sparks et al., 1993; Nishimura, 2006, 2009). Convection impedes the settling process because the settling velocity is greatly exceeded by the convective velocity throughout most of the depth of the magma chamber (e.g., Martin and Nokes, 1988; Koyaguchi et al., 1990; Sparks et al., 1993). In general, the thermal Rayleigh number (Ra) for a silicic magma chamber is thought to be extremely large and thermal convection must therefore be highly vigorous (Martin et al., 1987). However, where a melt in a cooling magma chamber has a water-saturated eutectic composition, thermal convection does not occur because there is no temperature gradient within the chamber (i.e., Ra is equal to zero). Moreover, there is no compositional convection

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in the chamber due to the homogeneous composition of the eutectic melt (i.e., the compositional Rayleigh number (R_s) is equal to zero). Hence, crystal settling occurs in a static magma chamber. As the compositions of many silicic magmas are close to the water-saturated eutectic (or minimum) composition (Luth et al., 1964), it is important to consider silicic magma chambers consisting of a eutectic melt and pre-existing crystals, which can be composed of early formed phenocrysts or source rock residues (i.e., restite crystals) (White and Chappell, 1977; Wyborn and Chappell, 1986; Chappell et al., 1987).

Marsh (1988) studied the thermal and compositional evolution of a static magma in a sheet-like chamber, and proposed the following scenarios: (1) a downward-solidifying crust of magma might overtake slowly settling crystals, capture them, and incorporate them into the crust itself when the growth rate of the leading edge of the capturing crust (i.e., the capture front) is higher than the rate of crystal settling; (2) the crust grows systematically slower with time, capturing fewer and fewer crystals; (3) the concentration of preexisting crystals decreases downward and then increases due to basal accumulation. This capturing process has been widely used to explain the modal and chemical variations observed in many sheet-like mafic bodies (Gray and Crain, 1969; Fujii, 1974; Marsh, 1988, 1989, 1996; Mangan and Marsh, 1992).

The ability of crystals to settle through silicic melt is strongly dependent on the melt viscosity. Until the 1990s, silicic melt was commonly thought to have a viscosity close to that of solid rock (e.g., Pitcher, 1993). However, recent experimental studies have established that silicic melts generated over a wide range of crustal pressure–temperature conditions have viscosities in the range of 10^8 – 10^3 Pa s (e.g., Clemens and Petford, 1999; Petford et al., 2000). Therefore, the amount of crystal settling should be estimated for different melt viscosities.

I propose a simple theoretical model that demonstrates how the degree of crystal settling affects temporal and spatial variations in the distribution of preexisting crystals in silicic magma chambers. This study shows that silicic magma chambers form various compositional structures depending on the viscosity of the melt. The implications of this model for compositional structures of granitic suites and rhyolitic magma chambers are discussed.

2. Model

Although multi-component melts generally solidify over a temperature range (typically 100–300°C), melts with a water-saturated eutectic (or minimum) composition solidify at a constant temperature. Emplacement of a silicic magma consisting of a eutectic (or minimum) melt and pre-existing crystals (i.e., restites or early formed

phenocrysts) is often observed within the continental crust (White and Chappell, 1977). Due to a homogeneous liquid density (thermal and compositional) within the magma chamber, the convective motion is suppressed. The compositional evolution of such a solidified silicic magma chamber (i.e., granite) is governed solely by the interaction of the solidifying and migrating crust, and the settling of pre-existing crystals.

Consider a sheet-like magma chamber consisting of a water-saturated eutectic melt and spherical crystals distributed uniformly (volume fraction ϕ_0) at the time of emplacement. The initial crystal fraction ϕ_0 should be consistent with the predictions of recent experimental studies (Clemens and Petford, 1999; Petford et al., 2000), which showed that most granite plutons were initiated as crystal-poor (<30% crystals by volume) liquids rather than largely crystalline magmas. In a eutectic magma chamber, crystallization is confined to the solid–magma boundary analogous to the ice–water transition (Nishimura, 2006). The boundary conditions are $T = T_\infty$ at $z = \infty$ and $T = T_E$ at $z = E(t)$ (Fig. 1); the initial condition is $T = T_\infty$ for $z > 0$ and $E = 0$. At time $t = 0$, the magma begins to solidify from the initial solid–magma boundary ($z = 0$) in the negative z direction. In the surrounding rock or the solidified part of the magma chamber, the heat conduction equation can be expressed as

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \quad (1)$$

where T is temperature, t is time, and κ is thermal diffusivity. Conservation of energy at the solid–magma boundary requires that released crystallization heat is conducted away from the boundary. During the capture process, conservation of energy at the boundary $z = E(t)$ indicates that

$$\rho L(1-\phi_0) \frac{dE}{dt} + \rho L \phi_0 V_S = k \left(\frac{\partial T}{\partial z} \right)_{z=E} \quad (2)$$

where ρ is density, L is the latent heat of solidification, k is thermal conductivity and V_S is the crystal settling velocity, which is negative in the case of downward solidification (Fig. 1 a) and vice versa (Fig. 1b). The second term on the left represents the effect of varying liquid proportions due to crystal settling, which is negligible when the V_S is small. Hence, this term is omitted here for simplification and equation (2) can be rewritten as

$$\rho L(1-\phi_0) \frac{dE}{dt} = k \left(\frac{\partial T}{\partial z} \right)_{z=E} \quad (3)$$

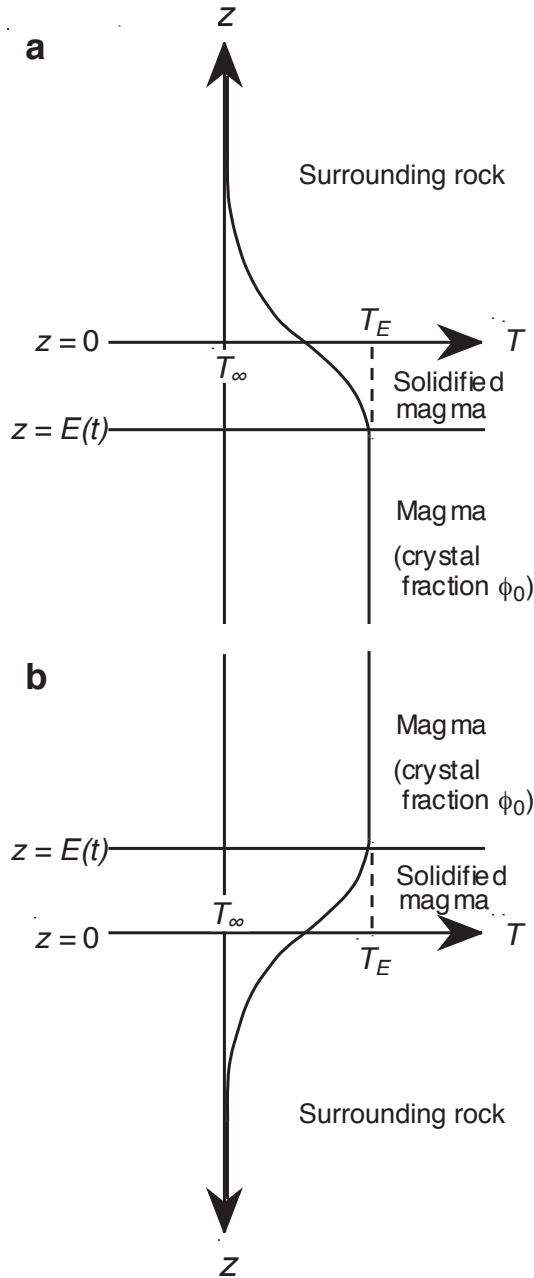


Figure. 1 Boundary conditions of cooling of a sheet-like magma chamber consisting of a eutectic melt and pre-existing crystals (crystal fraction ϕ_0). **a:** downward solidification. **b:** upward solidification. The magma solidifies at $z = E(t)$ and at eutectic temperature T_E . The surrounding rock near the magma is heated above T_∞ by the release of latent crystallization heat, but $T \rightarrow T_\infty$ at greater distances from the magma $z \rightarrow \infty$.

The position of the boundary between solid and magma, $E(t)$, is negative (Fig. 1) and can be expressed using the thermal diffusion length $(\kappa t)^{1/2}$ as

$$E = -2 \lambda \sqrt{\kappa t} \quad (4)$$

where λ is a constant. Combining equation (3), (4), and the solution of equation (1) gives a transcendental equation (Turcotte and Schubert, 1982) for determining λ :

$$\frac{L(1-\phi_0)\sqrt{\pi}}{c(T_e-T_\infty)} = \frac{e^{-\lambda^2}}{\lambda(1+\text{erf } \lambda)} \quad (5)$$

where c is the specific heat.

The settling velocity of a spherical crystal in a concentrated suspension is generally less than the Stokes' velocity mainly because of the equal volumetric flow rate of the displaced fluid relative to which the crystals must move. For uniform crystals, Richardson and Zaki (1979) established that they settle as follows:

$$|V_S| = \frac{a^2 g \Delta \rho}{18 \eta} (1 - \phi_0)^N \quad (6)$$

where a is crystal diameter, g is the acceleration due to gravity, $\Delta \rho$ is the density contrast between the crystal and the liquid, η is the shear viscosity of the liquid, ϕ_0 is the modal fraction of the crystals in suspension, and the constant (N) is empirically determined and varies from 4.6 (Mirza and Richardson, 1979) to 6.55 (Al-Naafa and Selim, 1992).

The compositions of pre-existing crystals and eutectic melt correspond to the mafic end-member and the silicic end-member, respectively. Therefore, the distribution of pre-existing crystals can be regarded as a compositional structure. When the growth rate of a downward-solidifying crust (solid roof) is higher than the rate of crystal settling, the crust might incorporate the settling crystals (Fig. 2a). If the settling rate becomes higher than the migration rate of the solid roof, a clear liquid phase develops under the roof of the chamber and consequently, the fraction of captured crystals, ϕ_E , at the solid–magma boundary becomes zero (Fig. 2b). On the other hand, an upward-solidifying crust (solid floor) can also capture the settling crystals. Unless ϕ_E exceeds a critical fraction, ϕ_{crit} , which is needed for packing, crystal settling would not form a settling layer (i.e., cumulus crystals + intercumulus liquid) on the solid floor (Fig. 2c). During the capture processes (Fig. 2a, c), ϕ_E can be calculated by mass balance at the solid–magma boundary as

$$\phi_E = \frac{(dE/dt - V_S)}{dE/dt} \phi_0 \quad (7)$$

where V_S becomes negative in the case of downward solidification (Fig. 2a) and vice versa (Fig. 2c). The dE/dt is obtained by differentiating equation (4) as follows:

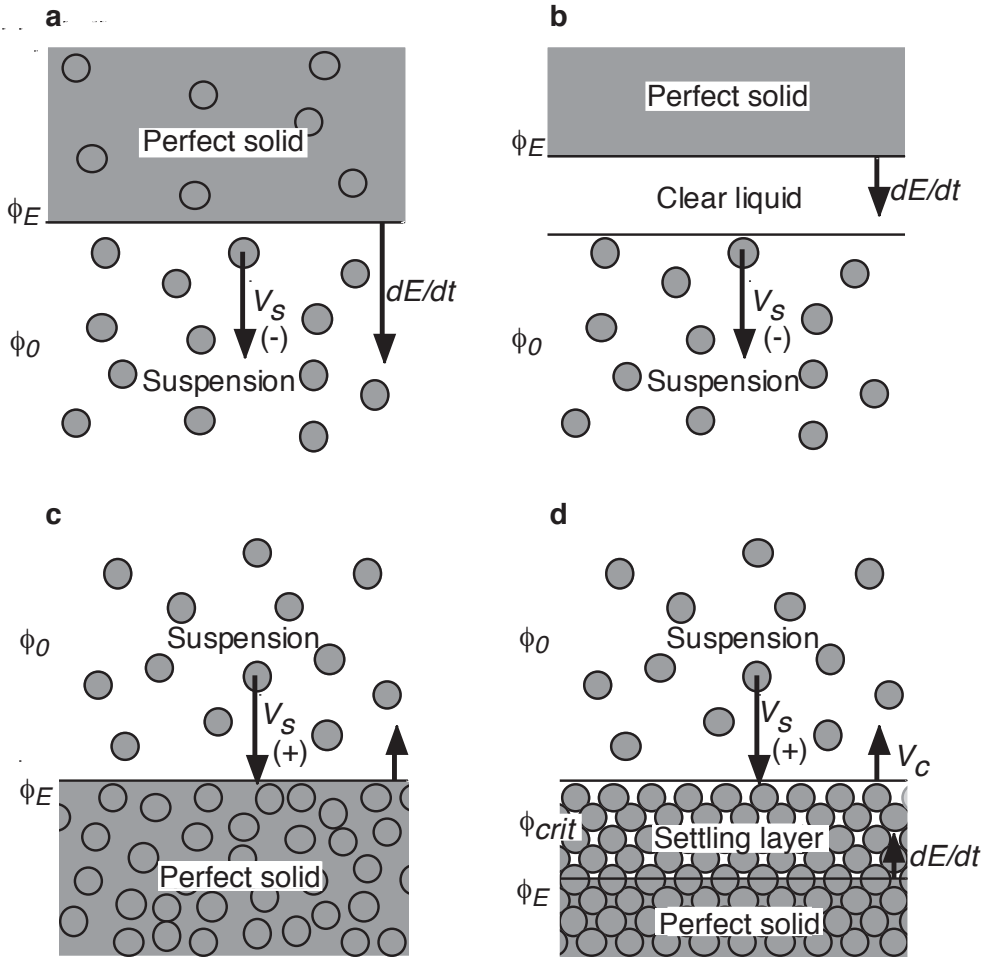


Figure. 2 Schematic illustration of crystal capture processes. **a:** A downward-solidifying crust (solid roof) captures settling crystals as long as the growth rate of the crust exceeds the crystal settling rate. **b:** A clear liquid phase develops under the roof when the crystal settling rate is higher than the growth rate of the crust. **c:** An upward-solidifying crust (solid floor) captures settling crystals unless the concentration of captured crystals, ϕ_E , reaches a critical fraction, ϕ_{crit} , which is needed for packing. **d:** After ϕ_E reaches ϕ_{crit} , a settling layer (cumulus crystals + intercumulus liquid) develops on the solid floor.

$$\frac{dE}{dt} = -\lambda \left(\frac{\kappa}{t} \right)^{1/2} \quad (8)$$

The time t required to move the solid–magma boundary from $z = 0$ to E can be obtained directly from equation (4) as

$$t = \frac{E^2}{4\lambda^2\kappa} \quad (9)$$

Incorporating equations (8) and (9) into equation (7), ϕ_E can be rewritten as

$$\phi_E = \phi_0 - \frac{V_S\phi_0}{2\lambda^2\kappa} E \quad (10)$$

This equation shows that the fraction of captured crystals, ϕ_E , decreases linearly with increasing depth at the downward-moving boundary (V_s is negative) and increases linearly with increasing height at the upward-moving boundary (V_s is positive) (cf., Nishimura, 2006). After ϕ_E reaches ϕ_{crit} at the upward-moving boundary, a settling layer develops on the solid floor of the chamber (Fig. 2d). The growth rate of the settling layer V_C (negative) is represented by a mass balance at the top of the layer (Koyaguchi and Kaneko, 2000) as

$$V_C = \frac{V_S\phi_0}{\phi_0 - \phi_{crit}} \quad (11)$$

It should be noted that the rate V_C has a constant value. The values for the relevant physical parameters are: $k = 7.7 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$, $\rho = 2.4 \times 10^3 \text{ kg m}^{-3}$, $L = 2.5 \times 10^5 \text{ J kg}^{-1}$, $k = 2.4 \text{ W m}^{-1} \text{ K}^{-1}$, $c = 1.3 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$, $T_E = 700^\circ\text{C}$, $T_\infty = 150^\circ\text{C}$, $\phi_0 = 0.2$, $\phi_{crit} = 0.6$, $\lambda = 0.715$ (calculated from equation (5)), $a = 5 \times 10^{-3} \text{ m}$, $\Delta\rho = 500 \text{ kg m}^{-3}$, $g = 9.8 \text{ m s}^{-2}$, and $N = 5$.

The settling of crystals in a silicic melt is strongly dependent on the physical properties of the melt, among which viscosity is the most important (see equation (3)). Experimental studies have shown that the viscosity of natural silicic melts generated at a given crustal pressure is largely independent of temperature (Scaillet et al., 1998). In particular, the eutectic (or minimum) melt on the water-saturated solidus at a given pressure must have a constant viscosity because major oxide contents (including H_2O) and temperature are constant throughout the solidification history. The viscosities of the water-saturated eutectic (or minimum) melts for haplogranite are shown in Fig. 3. The

H₂O content of a eutectic (or minimum) melt decreases and the melt viscosity increases with decreasing pressure.

3. Results and discussion

The distribution of pre-existing crystals is calculated for constant melt viscosities in the range 10^4 – 10^8 Pa s. Figure 4 shows the temporal evolution of the crystal distribution within the initially 1000-m-high magma chamber (left panels) and the resulting compositional structure within the solidified granite (right panels). In the magma chamber with a melt viscosity greater than 10^7 Pa s, a homogeneous distribution of pre-existing crystals is maintained throughout the cooling history (Fig. 4a,c). However, the concentration of pre-existing crystals in the solidified granite decreases downward and then suddenly increases near the center of the pluton (Fig. 4b,d). The gradient in crystal concentration increases with decreasing melt viscosity. Previous studies of the capture processes have shown that in many cases the pre-existing crystal concentration gradually changes along an S-shaped curve (Gray and Crain, 1969; Fujii, 1974; Marsh,

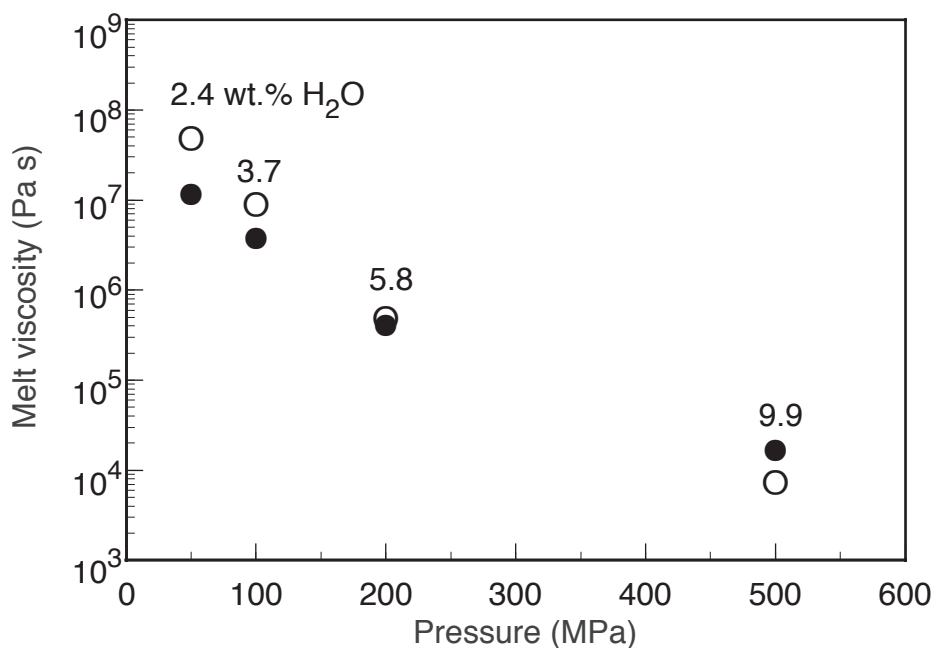


Figure 3 Pressure dependence of the viscosity of water-saturated eutectic (or minimum) melt in the haplogranite system Q–Ab–Or–H₂O. The melt viscosities are calculated from temperature and compositional data by Tuttle and Bowen (1958) and Holtz et al. (1992) using models by Shaw (1972) and Giordano et al. (2008) (open and solid circles, respectively).

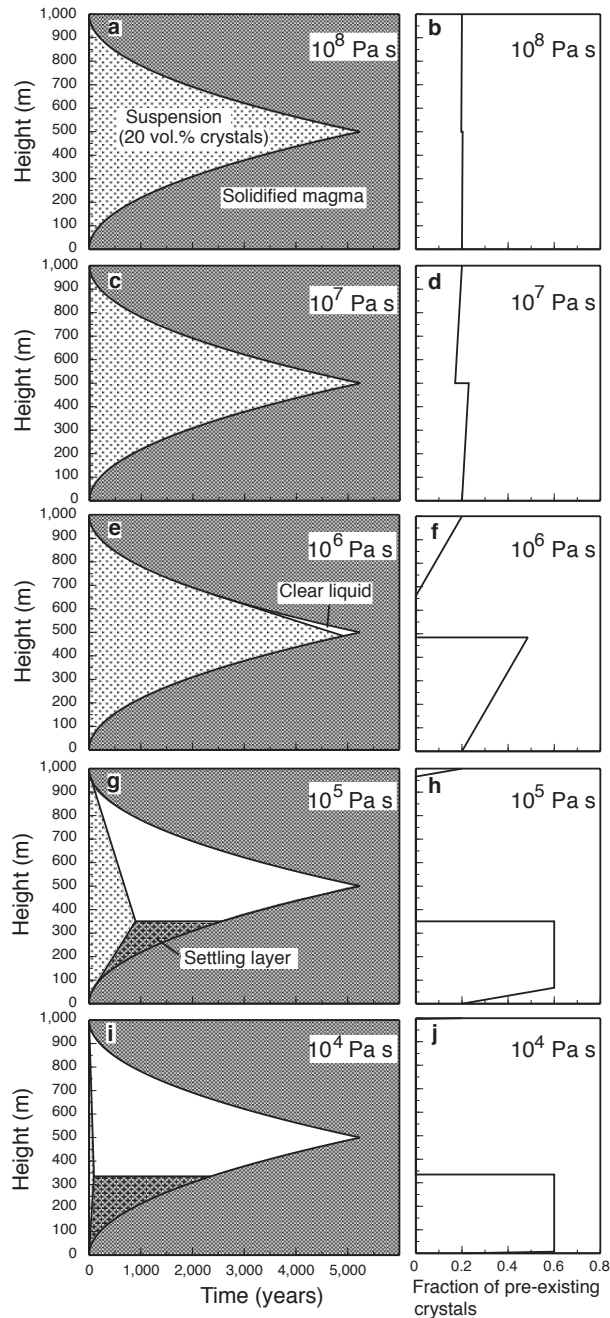


Figure. 4 Temporal evolution of the crystal distribution within magma chambers with different melt viscosities (**a, c, e, g, i**) and resulting compositional structures (distribution of pre-existing crystals) of the resulting granite plutons (**b, d, f, h, j**). A compositional structure is formed within a granite pluton even if a uniform crystal distribution is maintained in the magma throughout its cooling history (**a–d**). A settling layer develops on the floor of the chamber if the melt viscosity is less than 10^5 Pa s (**g, i**).

1988; Mangan and Marsh, 1992). However, the present results show that for high-silica magma bodies, the captured crystal concentration changes linearly, and a clear compositional gap exists at certain depths.

In the magma chamber with a melt viscosity of 10^6 Pa s, a uniform distribution of pre-existing crystals is maintained for a long time ($\sim 3,000$ years) and a clear liquid phase is formed at the last stage of crystallization (Fig. 4e). In this case, the compositional gap near the center of the pluton reaches ~ 50 vol.% (Fig. 4f). In the magma chamber with a melt viscosity lower than 10^5 Pa s (Fig. 4g, i), a clear liquid phase and a settling layer (cumulus crystal + inter-cumulus liquid) develop in a short time span (10 – 10^2 years), forming a vertically zoned magma chamber. The concentration of pre-existing crystals in the granite pluton becomes zero in the upper part of the pluton and a settling layer ($\phi_{crit} = 0.6$) develops on the pluton floor (Fig. 4h, j). Even if the eutectic (or minimum) melt is water-undersaturated, its melt viscosity is almost equal to that of a water saturated eutectic (or minimum) melt at a given pressure (Hess and Dingwell, 1996; Scaillet et al., 1996; Holtz et al., 2001; Nishimura, 2006). Therefore, the calculated results (Fig. 4) represent a good approximation for water-undersaturated high-silica magmas. Petrological studies suggest that some rhyolitic volcanic rocks are erupted from the upper part of a magma chamber that successively crystallizes to form a granitic pluton (e.g., Bachmann et al., 2007). For a 1000-m-thick sheet-like magma chamber with a melt viscosity greater than 10^7 Pa s (shallower level), the rhyolite (area labeled ‘20 vol. % crystal’ in Fig. 4a, c) and the granite (Fig. 4b, d) have almost the same composition. Conversely, crystal-poor rhyolite and crystal-rich cumulative granite are formed at melt viscosities lower than 10^5 Pa s (deeper level).

The calculated time period required for settling layer formation (Fig. 4g, i) is not significantly different from that estimated for a vigorously convecting magma chamber (Martin and Nokes, 1988). Although there are many factors that affect the evolution of magma chambers (e.g., magma recharge, wall-rock assimilation, magma mixing), crystal capture would inevitably occur and an idealized model of this process may help to provide new explanations for the compositional evolution of magmas.

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要 旨

珪長質マグマ溜まりの固化前線における沈降結晶の捕獲

西 村 光 史

共融点メルトと浮遊結晶からなる珪長質マグマ溜まりを仮定し、固化前線による沈降結晶の捕獲が組成の空間構造にもたらす影響について計算を行った。これまでの研究では主に苦鉄質貫入岩体中に見られる結晶捕獲プロセスが調べられ、沈降結晶がS字型のモード分布を形成することが示されてきた。しかし今回の計算の結果、共融点メルトと浮遊結晶からなる珪長質マグマ溜まりでは沈降結晶のモードは直線的に変化し、ある深さにおいて急激な組成ギャップを形成することが明らかとなった。