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RHEOLOGY OF GRANITIC MAGMAS DURING ASCENT AND EMPLACEMENT

Nick Petford
School of Earth Sciences and Geography, Kingston University, Surrey KT1 2EE, United Kingdom; email: n.petford@kingston.ac.uk

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■ **Abstract** Considerable progress has been made over the past decade in understanding the static rheological properties of granitic magmas in the continental crust. Changes in H_2O content, CO_2 content, and oxidation state of the interstitial melt phase have been identified as important compositional factors governing the rheodynamic behavior of the solid/fluid mixture. Although the strengths of granitic magmas over the crystallization interval are still poorly constrained, theoretical investigations suggest that during magma ascent, yield strengths of the order of 9 *kPa* are required to completely retard the upward flow in meter-wide conduits. In low Bagnold number magma suspensions with moderate crystal contents (solidosities $0.1 \leq \phi \leq 0.3$), viscous fluctuations may lead to flow differentiation by shear-enhanced diffusion. AMS and microstructural studies support the idea that granite plutons are intruded as crystal-poor liquids ($\phi \leq 50\%$), with fabric and foliation development restricted to the final stages of emplacement. If so, then these fabrics contain no information on the ascent (vertical transport) history of the magma. Deformation of a magmatic mush during pluton emplacement can enhance significantly the pressure gradient in the melt, resulting in a range of local macroscopic flow structures, including layering, crystal alignment, and other mechanical instabilities such as shear zones. As the suspension viscosity varies with stress rate, it is not clear how the timing of proposed rheological transitions formulated from simple equations for static magma suspensions applies to mixtures undergoing shear. New theories of magmas as multiphase flows are required if the full complexity of granitic magma rheology is to be resolved.

INTRODUCTION

The emplacement of granitic magmas in Earth's crust marks the end of a coupled thermomechanical process involving partial melting, segregation, and ascent of silicate melts and suspensions. In all cases, the mechanical behavior of partially molten rock plays a key role (Petford et al. 2000). Insight into the processes that take place is somewhat difficult to generalize, as many mechanisms occur simultaneously and interdependently, with a tendency toward more brittle behavior with time. Aspects that have a bearing on the overall behavior and evolution of molten rock include seepage flow, heat flow, phase transitions, deformation mechanisms of both fluid and solid phases, the physics and rheology of multiphase suspensions, and chemical processes. Progress in understanding the interplay in this multiphysics problem is being made all the time—examples include analogue deformation experiments (e.g., Rosenberg & Handy 2001), semi-quantitative assessments of magma rheology (Vigneresse et al. 1996, Barboza & Bergantz 1998, Vigneresse & Tikoff 1999), strength of partially molten aggregates (Rutter & Numann 1995, Renner et al. 2000, Rosenberg 2001), and fully analytical treatments (Koenders & Petford 2000). Field and experimental studies have shown that deformation can enhance significantly the rate of melt segregation and transport in partially molten rock (Brown & Rushmer 1997), both during initial partial melting (Sawyer 1994) and during magma emplacement, and that granitic magma accumulates preferentially in structurally controlled sites across a range of tectonic settings (e.g., Handy et al. 2001).

This review is split broadly into two sections. The first is concerned mainly with recent work on the rheology of magmatic suspensions as applied to the crystallization of granitic magmas. The main results of these studies are reviewed, showing how new approaches based on developments in the physics of multiphase flow have been applied, with some success, to modeling the mechanical behavior of magmas during ascent and emplacement. An important distinction can be made between both regimes in that for rapid ascent, where flow is continuous, an approximately adiabatic heat balance ensues, whereas the emplacement stage is marked by monotonic cooling. It is under the latter conditions that rheological changes in the magma will be most extreme. A discussion on emplacement then follows, highlighting the role of fabric development in models for magma rheology. The viscosity-reducing effects of volatile phases and magmatic enclaves as kinematic markers of bulk magma rheology are discussed, and a mechanical comparison is made between silicic rocks of similar composition and physical properties emplaced at the surface at high strain-rates. Only the rheological behavior inside the magma body is considered here; readers interested in the rheological changes in the surrounding country rock during emplacement are referred to other sources (e.g., Kerrick 1991, Handy et al. 2001). Finally, this review deals mostly with the rheology of granitic magmas undergoing freezing. This is because (*a*) an informed discussion of the rheology of anatectic rocks would require a separate review and (*b*) as shown by Vigneresse et al. (1996), partial melting and crystallization, although complementary, should not be regarded as symmetrical processes (e.g., Wickham 1987), as proposed rheological transitions and thesholds in static systems do not map directly from one porosity interval to the other (Figure 1).

Multiphase Flows

Magmas are prime examples of flows that involve the transport of solids and gases (and other fluids) by a separate carrier phase. Such flows are called multiphase, and have attracted much recent attention due to their important range of engineering

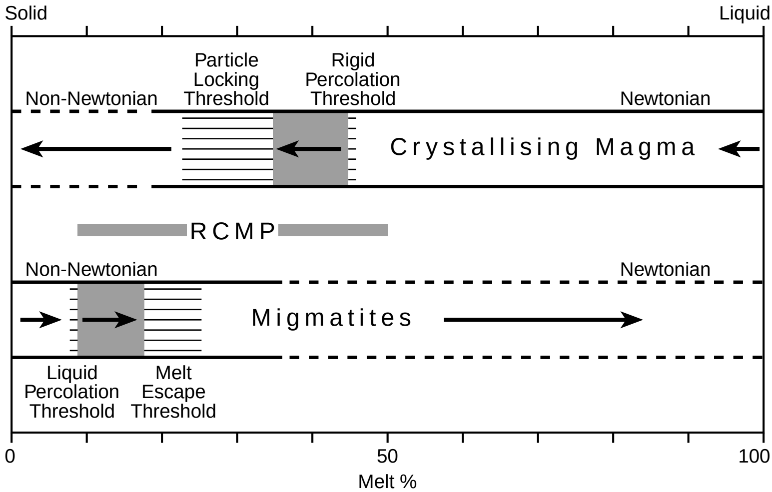

Figure 1 Relationship between magma crystallization and partial melting based on the rheology of granitic magma suspensions proposed by Vigneresse et al. (1996). The rheological critical melt percentage (RCMP) relates strictly to partial melting (migmatites) only. Estimates of the position of the RCMP as the percent volume of melt increases varies accordingly: 20 $\pm$ 10% (Arzi 1978), 30%–35% (van der Molen & Paterson 1979), 35%–50% (Wickham 1987), and $\geq$50% (Miller et al. 1988).

applications. The defining characteristic of multiphase flows are their complex behavior, for which a large technical literature now exists (Gidaspow 1994). In modeling and simulation, the continuous phase, and for low mass loading the dispersive phase, are treated as Eulerian (Ferziger & Perić 1999). Where the number density of the dispersed phase is large, the influence of particles on the fluid motion becomes significant and must be taken into account in any explanation of the bulk behavior of the mixture.

Numerical techniques for simulating multiphase flow require that computation of particle and fluid trajectories are done simultaneously and iteratively and are still in a relatively unadvanced state, although progress is being made all the time (Crowe et al. 1998). Further complications arise due to the interactions between particles and walls and when phase changes take place in the carrier fluid. Although the application of multiphase flow theory to magma suspensions is still in its infancy, it has the potential to bring new insight into complex geological flow behaviors (Bergantz & Barboza 2003). Some examples of multiphase flows employing the concept of "granularity" are set out in later sections.

Magma Emplacement Mechanisms

The problem of developing a generic rheological model for the emplacement of granitic magmas is complicated in part by competing ideas on how the magmas ascend through the lithosphere. The emplacement of granitic magma in Earth's crust has a long history of controversy (see Pitcher 1993), with debate currently polarized between those who favor a traditional diapiric ascent mode and the so-called dykists, who argue for ascent in narrow conduits. In the context of this article, several important and practical differences between the two mechanisms can be recognized. First, in dyke ascent a clear distinction is made between vertical flow during ascent and switching to predominantly horizontal flow during emplacement (e.g., Clemens et al. 1997), whereas for diapiric rise, this distinction is blurred because ascent and emplacement are regarded as essentially the same (e.g., Miller & Paterson 1999). Second, the rates of magma ascent in dykes are up to 10^9 greater than diapiric velocities, leading to the possibility that some dyke magmas become superheated during rise and will resorb solids. The rheological effects of such behavior are discussed later.

Another important difference between the two ascent mechanisms is the geometry of the intrusion at emplacement. Dyke-fed intrusions are expected to be typically sheet-like, with a low aspect ratio (e.g., McCaffrey & Petford, 1997, Cruden & McCaffrey, 2001), whereas diapirs, with long vertically extending tails, tend towards high aspect ratios (Weinberg & Podladchikov 1994). Clearly, the longer a pluton of given volume and emplacement depth can stay above its solidus, the more chance it has of interacting with the regional stress field during emplacement. The significant point here is that tabular plutons will cool faster than diapiric ones of similar mass due to their larger (approximately 30%) surface area to volume ratio. In the following sections, the rheology of granitic magmas during ascent and emplacement are treated in turn. Where possible, I consider effects in a parcel of magma close to a vertical and horizontal surface such that the scale of the problem is largely independent of macroscopic geometry of the boundary (straight or curved). Unfortunately, it is hard to assess the rheological effects of increasing crystal load on diapiric motion using a Stokseian formation, as changes in the viscosity of the magma have a negligible effect (see Batchelor 1970, Petford et al. 1994). Thus, changes in granitic magma rheology during the ascent phase are concerned mostly with focused flow in narrow conduits or along crustal discontinuities.

RHEOLOGICAL THRESHOLDS: REAL OR IMAGINARY?

Magma viscosity can increase by a factor of 10^{13} over a temperature interval of about 200°C, with most of the increase due to the effects of mass loading of the melt phase by crystals (Pinkerton & Stevenson 1992). There are a number of experimentally derived equations that make predictions about the behavior of rigid, generally spherical particles suspended in a viscous fluid, and the results of a survey by Jinescu (1974) are summarized by Vigneresse & Tikoff (1999). Unfortunately, the phenomological nature of the results do not allow one to arrive at a fundamental explanation of the rheological behavior of a suspension. Indeed, it is well recognized that establishing the mechanical properties of suspensions, both static and in a state of shear, is nontrivial, and no simple expression can be used to describe their rheology with constantly varying concentrations (e.g., Jeffrey & Acrivos 1976). However, an impression of the changing rheology of suspensions as the solid phase increases can be obtained from the well-known equations put forward by Einstein (1906) and Roscoe (1952) and are shown in Figure 2. In its original formulation, the Einstein equations were valid for small numbers of particles (Bird et al. 1960), although extrapolation to higher mass loading appears valid in some cases (Shaw 1965, Jeffrey & Acrivos 1976). Lejeune & Richet (1995) suggest the relationship can be extended to crystal contents up to 40%, and experimental work by Scaillet et al. (1997) shows that for some hydrous magmas, this may hold for more than 90% of the crystallization interval. However, problems still exist with extrapolation to concentrations ≥50%, along with other complicating factors, including particle shape, size, and mean size distribution. This notwithstanding, curves showing a sudden and large increase in relative viscosity over a small porosity interval helped inspire the work of Arzi (1978), who introduced the concept of a rheological critical melt percentage (RCMP) at about 25% to 30% melt as a fundamental property of partially molten materials. The experimental work of van der Molen & Paterson (1979) appeared to support evidence for a RCMP, and the idea was put forward for a critical melt fraction (CMF) at crystal contents (solidosity) of $0.3 \geq \phi \geq 0.5$, thought to represent a fundamental rheological barrier to melt extraction (see the review by Wickham 1987). Similar curves appear to explain some more recently derived experimental data on magma rheology (e.g., Lejeune & Richet 1995). However, the experiments of Rutter & Neumann (1995) on the Westerly granite (3 kbar, strain rates of $10^{-7}s^{-1}$) have cast some doubt on earlier work in support of a CMF by showing that the strength of the material decreased gradually with increasing melt fraction up to 50% volume. Based on a compilation of six partial melting experiments on granite systems reviewed by Bagdassarov & Dorfman (1998), the relative (or effective) viscosity of the mixture can be approximated from $\log \eta_{rel} = 8 - 5.9n$ for low to intermediate melt fractions (porosities, n) of 0 to 0.5. This curve is plotted in Figure 2 for comparison with the Einstein-Roscoe & Spera (2000) curves for high (1–0.7) melt fractions and various packing densities. Although the extrapoloated curve from the partial melting experiments overestimates the relative viscosity at high melt fractions, there appears to be some degree of convergence at solidosities around 30% for both the simple cubic ($\phi_0 = 0.52$) and body centered ($\phi_0 = 60$) cubic packing conditions, and at higher mass loading (approximately 40%), the modified Spera curve.

Brown & Rushmer (1997), who reviewed the mechanical behavior of migmatites during deformation, cautioned against the presence of a CMF on the basis of field studies, whereas Wolf & Wyllie (1991) have shown that melt connectivity in partially molten amphibolite can be achieved after just 2% melting. Most recently, Rosenberg (2001) has reviewed experimental and natural case studies of the deformation of partially molten granite, pointing out that the sigmoidal shapes of

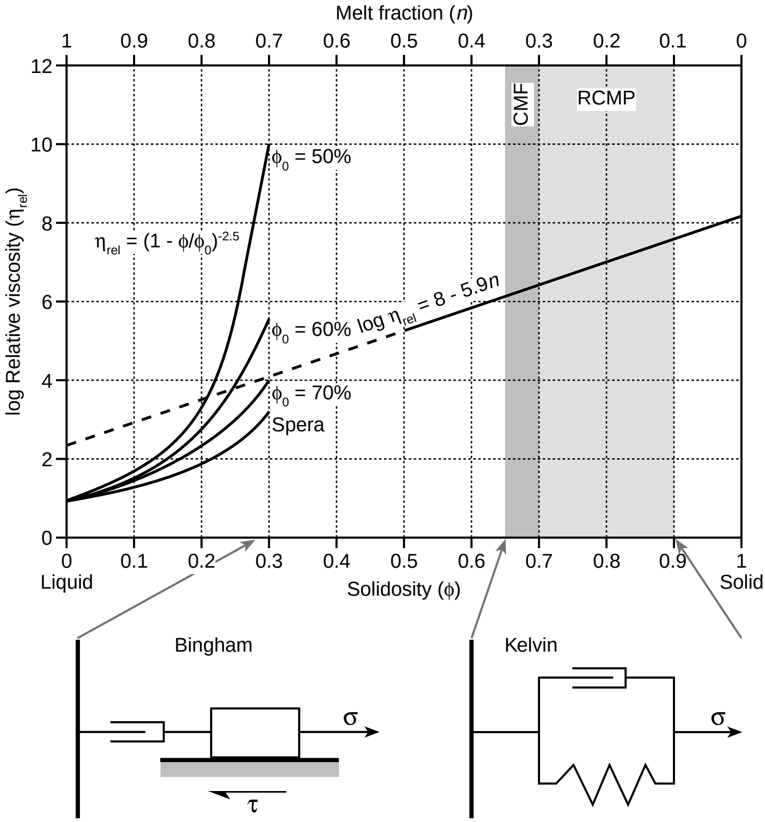

Figure 2 Composite plot showing various experimental and empirical curves used to constrain the change in relative (or effective) viscosity η_{rel} of magma suspensions with increasing crystal content (solidosity). The family of curves at high melt fractions ($\phi = 0$–0.3) are calculated using the suspension viscosity equation for three maximum packing values (ϕ_0) of 70%, 60%, and 50% (see Pinkerton & Stevenson 1992 and Spera 2000 for a review of the coefficients used and their justification). Also shown (Spera 2000) is a modified suspension viscosity curve that takes into account melt trapped between particles $\Psi_M = 1 - \phi[1 - \phi_0]/\phi_0$, where $\eta_{rel} = \eta_{magma}/\eta_{melt} = (1 - \phi/\Psi_M)^{-5/2}$. The solid line is a compilation by Bagdassarov & Dorfman (1998) of six partial melting experiments, valid for $0.5 < \phi < 1$, extrapolated to $\phi = 0$ (*dashed*). The positions of the RCPM and CMF are from Arzi (1978) and van der Molen & Paterson (1979). Inserts show diagramatically some of the possible rheological behaviors of the mixture across the crystallization interval (see text for further discussion).

the Arzi-type RCMP curves at low melt fractions depend to a large degree on the experimental conditions, in particular, the melt viscosity. In summary, it is probably fair to say that for static systems, the data are sufficiently ambiguous that the presence of a critical melt fraction at some fixed threshold cannot be ruled out entirely. Recent treatments of suspensions (e.g., Cates et al. 1998) that exhibit nonequilibrium transitional behavior from solid to fluid-like states due to processes such as particle jamming and associated refluidization (fragile media) may also be applicable to magmas and should be explored accordingly.

ONSET OF NON-NEWTONIAN BEHAVIOR A further assumption made in the extrapolation of experimental models to real magmas is that the viscosity of the melt phase is Newtonian. Clemens & Petford (1999) have summarized the range in viscosity of granitic melts for water contents typical of leucogranitic and tonalitic systems. The results are summarized in Figure 3, where crystal-free granitic melt viscosities are of the order 10^4 Pa s. A similar result was obtained by Scaillet et al. (1998). However, as pointed out by Spera et al. (1992), if the melt phase is non-Newtonian, or if the crystals are not elastic (e.g., Bagdassarov & Dorfman 1998), the strength of the mixture will be strain-rate dependent. Unfortunately, the influence of strain rate on melts and magma rheology is still poorly understood, although experimental work on partially molten mantle rocks suggests that even at small amounts of melt ($\lesssim$5%), a combination of melt-enhanced grain boundary diffusion and sliding in both the diffusion and dislocation creep regimes can cause significant weakening of the mixture (Kohlstedt & Zimmerman 1996).

It has been proposed that granitic magma suspensions will behave as Bingham materials at some critical solid content (generally around 30%–40%), at which the static mixture assumes a yield strength that must be overcome for flow to proceed (e.g., Dingwell et al. 1993, Fernandez & Gasguet 1994, Paterson et al. 1998). Numerous studies have attempted to identify the exact point of transition from Newtonian to non-Newtonian behavior as a function of crystal content and yield strength. Estimates vary somewhat, but regardless of magma composition, this transition is in the region 30%–50% (maximum) crystals. Examples from the literature, mostly from work on basaltic systems, include Ryerson et al. (1998) ($\phi >$ 25%), Shaw (1965), Pinkerton & Stevenson (1992) ($\phi >$ 30%), Lejeune & Richet (1995) ($\phi >$ 40%), Hoover et al. (2001) ($\phi >$ 25%), and Kerr & Lister (1991) ($\phi >$ 50%). The last two studies also highlight the important role of particle shape in determining the onset of yield strength. It should be noted that there are a number of possible types of non-Newtonian flow behavior exhibited by unidirectional shear flows, and that non-Newtonian does not necessarily imply an increase in effective viscosity. For example, shear thinning and dilatant flow are examples where the viscosity decreases with increasing strain rate (Bird et al. 1960). Indeed, highly cited Bingham flow is just one example of a range of possible mechanical responses to increasing mass loading of the mixture. Other types of possible rheological behavior include Maxwell (elastic response followed by viscous flow) and Kelvin bodies, where both elastic deformation and viscous flow of the mixture occur in

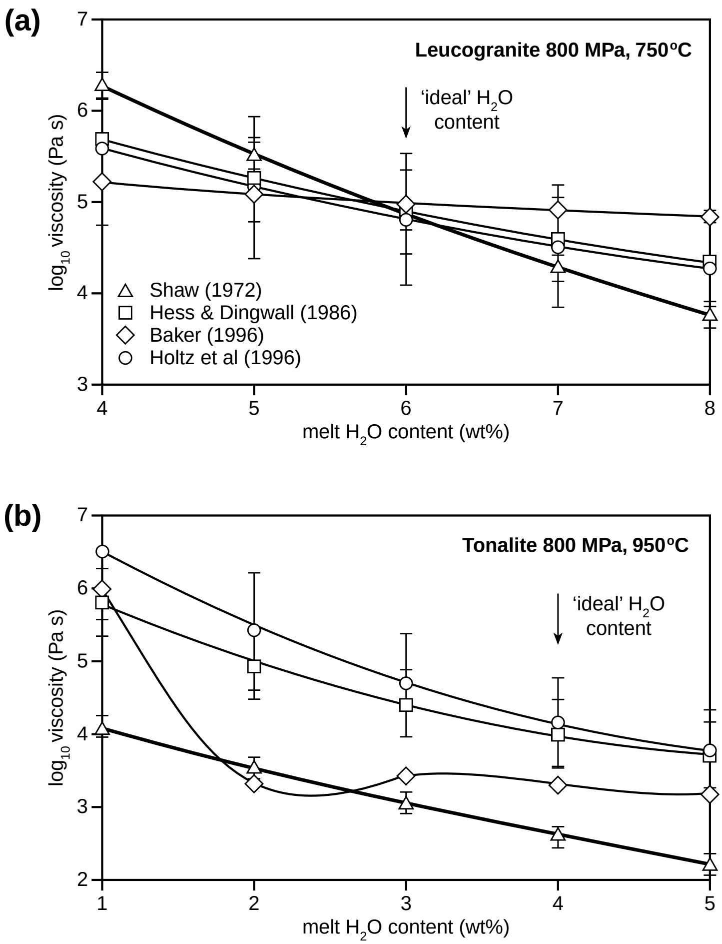

Figure 3 Compilation of predicted melt viscosity curves for a tonalite ($SiO\%_2$ = 64.2 wt%) and leucogranodiorite (SiO_2 = 75.3 wt%) derived from various experiments as a function of increasing water content in the melt, along with the Shaw (1972) curve for comparison (after Clemens & Petford 1999).

parallel. The latter has been shown to be important in the mechanical behavior of water-saturated clays over short time periods at moderate porosities, where strain is taken up initially in the viscous fluid (Middleton & Wilcock 1994). Both may be applicable to the rheology of magmatic suspensions at some time during their deformation history (Figure 2).

BRITTLE DEFORMATION A consequence of increasing the solid fraction in the melt is that once a continuous framework is formed, the suspension is capable of transmitting deviatoric stresses. The precise timing of this is again a matter for debate, but depending on crystal shape, is likely to be at an excess of 50% solids (Kerr & Lister 1991). This value corresponds generally to the point of "lock-up" of the system (e.g., Marsh 1981). It has been suggested that at high strain rates, such a mixture is capable of deforming as a brittle (elastic) solid, despite the presence of up to 50% melt (Hallot et al. 1996). The critical strain rate corresponding to brittle behavior for crystal-free glasses and highly crystallized magmas is $> 10^{-5}\mathrm{s}^{-1}$ (Webb & Dingwell 1990). However, as pointed out by Miller et al. (1988), at high differential stresses, the yield strength of the skeleton framework may be exceeded, resulting in a sudden loss of contiguity and corresponding reduction in strength. Fernandez & Castro (1999a) have presented evidence for simultaneous brittle and viscous behavior in granite plutons from Iberia, but again (as with static models for a RCMP), rigorous constitutive models needed to describe the mechanical response of the magma during shear are lacking. Brittle deformation is commonly observed in microgranular enclaves (see below). Fracturing of enclaves is most likely where the flow rate exceeds several m/year (Williams & Tobisch 1994).

To summarize, models based on empirical studies, as attractive as they are, cannot capture the full range of likely mechanical behavior of crystal mushes nor the time-dependent nature of the mixture as it deforms. What is required is a new approach, either experimental or theoretical, that can inject the required subtlety into the problem. One rapidly advancing field of science that may provide new theoretical insight is work currently done in the field of multiphase flow. Some possible strategies for dealing with applications to magma flow are set out in the following sections.

MAGMA (DYKE) ASCENT

FLOW OF BINGHAM MAGMA Newtonian fluids moving with constant viscosity (no-slip boundary condition) at low Reynolds number result in the familiar parabolic profile characteristic of viscous channel flow (e.g., Bird et al. 1960, p. 45). This is the situation described in simple models of dyke ascent of basaltic and granitic melts, and although robust under these simplifying assumptions, the presence of crystals above a critical concentration is likely to effect the flow properties of the ascending magmas through the introduction of a yield strength (Shaw 1965). The effects of yield strength on flow rates of granitic magmas are now examined. In their review on the rheology of magmas at sub-liquidus temperatures, Pinkerton & Stevenson (1992) presented equations that were used to modify the experimental measurements of Murase et al. (1985) on the Mt. St. Helens dacite, and obtained a yield strength of 370 Pa. Using this value as an example, the effects of yield strength (0−500 Pa) on the flow rates of granitic magma are shown in Figure 4 for two cases of varying magma viscosity (Figure 4*a*) and density contrast (Figure 4*b*). It is apparent that the effect of yield strength on the transport velocity is negligible at these values in comparison with the large effect of melt viscosity. For isoviscous flow, the effect on mean flow rate is slightly more pronounced, with velocity decreasing with increasing yield strength. However, this effect is more than offset by increasing the density contrast driving the flow. For granitic dykes of similar widths and melt viscosities, yield strengths >9 kPa, 20 times greater than the value reported for the Mt. St. Helens dacite are required to retard flow completely (Petford & Koenders 1998). Note that in the above example, the yield strength is supplied simply as a coefficient for illustrative purposes and does not imply a specific crystal content for the flow.

GRANULAR TEMPERATURE MODEL In simple pipe flow, the shear field varies across the radial dimension, from a maximum at the wall to zero at the center. Scaled experiments designed to reproduce the flow of magma suspensions during ascent and emplacement have been made by Bhattacharji (1966). The experiments (an example is reproduced in Figure 5) show that flow is accompanied by axial migration of solids towards the center of the flow (the Bagnold effect), even for moderately low particle contents of approximately 15% by volume. Flow segregation during magma ascent and emplacement can be important in producing chemical differentiation in magmas (Philpotts 1990) and highlights the importance of understanding the mechanical effects arising from flow of magma. Despite being a purely phenomenological description, the Bagnold effect is cited as the dominant mechanism for flow segregation of crystals or other suspended material commonly observed in solidified dykes and sills (e.g., Komar 1972, Philpotts 1990). Given (as shown already) that deviations from simple Newtonian behavior are expected as the fraction of solids in the flow increases, it is important to be able to quantify and predict the evolution of magmatic flows under increased mass loading. In the past couple of years, papers have appeared in the fluid mechanics and micromechanics literature that help explain the mechanisms occurring in sheared suspensions. In this new work, a measure of the fluctuations of the particle velocity is introduced that is regarded as a temperature field. The temperature is defined as the kinetic energy of the particles associated with the mean quadratic deviation from the average velocity $\langle \mathbf{V} \rangle$; that is, if the fluctuation velocity of particle with number (j) is

$$\tilde{\mathbf{V}}^{(j)} = \mathbf{V}^{(j)} - \langle \mathbf{V} \rangle, \tag{1}$$

then the granular temperature of a material point with volume V is

$$T = \frac{1}{V} \sum \frac{1}{2} m^{(j)} (\tilde{\mathbf{V}}^{(j)})^2 \tag{2}$$

(Koenders et al. 1996). Particle fluctuations arise because in shearing a suspension, particles are forced to avoid one another and to roll and translate in order to satisfy the mean imposed motion. The energy associated with these irregular movements is much greater than that to be expected on purely Brownian grounds. Further insight in the order of magnitude of these irregularities has been much enhanced by the development of numerical simulations of spherical particles in Newtonian fluids (Leighton & Acrivos 1987, Koh et al. 1994, Nott & Brady 1994), and prolate spheroids (Claeys & Brady 1993).

Although the theory is well known for dry suspensions and water-saturated flows, it is less developed for low Bagnold number viscous flows in pipes with

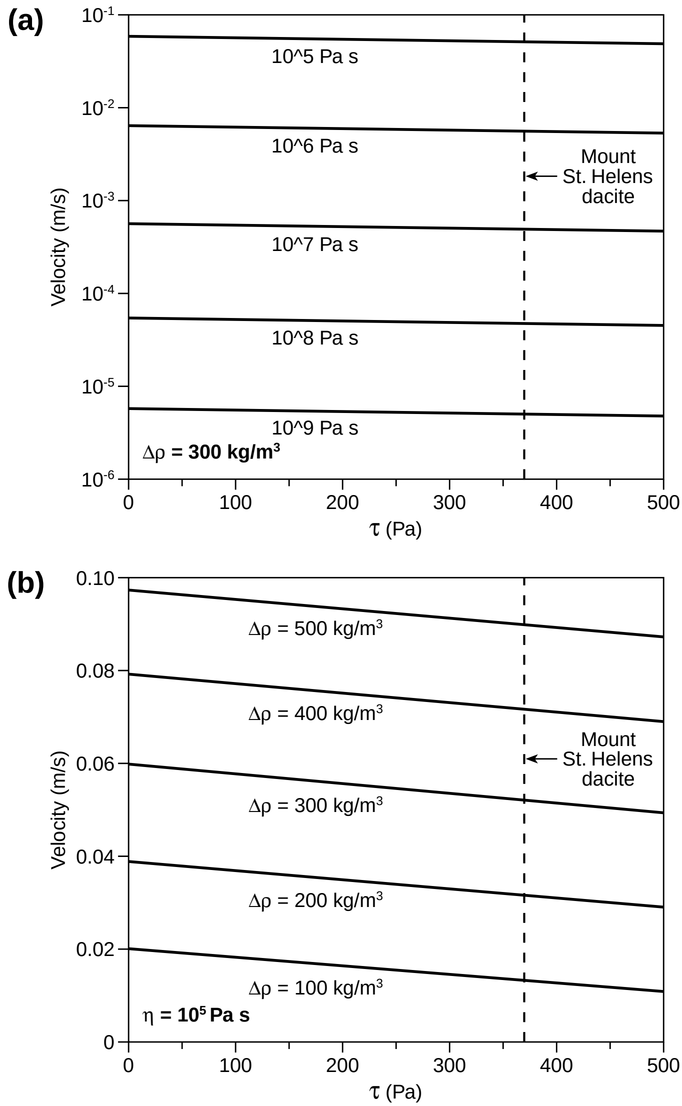

Figure 4 Plots showing the effect of yield strength (τ) on the mean flow velocity of a granitic magma in a dyke of width 6 m, for varying melt viscosity (*a*) and density contrast (*b*). The estimated yield strength of the Mt. St. Helens dacite ($\tau = 370$ Pa) is shown for reference. Note the small effect of yield strength on the average magma ascent velocity.

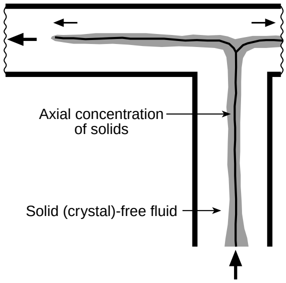

Figure 5 Summary of model experiments by Bhattacharji (1966) to investigate flow differentiation in (mafic) magmas during ascent and emplacement for an initial solids content of 15% by volume. The axial concentration of solid particles (crystals) mimics clearly field observations on porphyritic dykes where highest concentrations of phenocrysts (highest effective viscosities) occur towards the centers of the flow.

rough walls (e.g., dykes). To this end, following an approach first proposed by McTigue & Jenkins (1992), Petford & Koenders (1998) made these multiphase flow equations applicable to flow of viscous magmas. To do this, several important modifications to the boundary conditions, namely the introduction of an isotropic fluid constitutive law much like those of a dense gas, were introduced. Thus, their approach in modeling magma as a granular material involved a balance equation for the fluctuational energy, equivalent to the heat equation for a gas. In their model, which unlike that of McTigue & Jenkins (1992) had rough boundaries, the particles in the suspension were considered a dense gas with mechanical and thermal properties (see Chapman & Cowling 1970). The energy fluctuations of the particles due to shearing are captured in the granular temperature. Most importantly, particle diffusion takes place when a temperature gradient is present.

Results summarizing the flow simulations pertinent to viscous multiphase flow in granitic dykes are shown in Figure 6*a–c* for three initial solidosities (ϕ) of 10%, 20%, and 25%. The accompanying solidosity variations, velocity, and particle (crystal) fluctuation velocity intensity are plotted as a function of the nondimensional cross streamwise parameter (note that a continuum solution is unavailable outside the mean free path boundary). The results show the granular velocity fluctuation intensity [$\hat{W}(\hat{y})$] is highest closest to the dyke walls, as expected from theory (Figure 6*a*). In response, high solidosities (crystal loads) develop at the centre of the dyke (Figure 6*b*). Intriguingly, a plateau develops in the center of the dyke, much like a plug in Bingham flow (Figure 6*c*), but note that the mechanism here is entirely diffusional as the magma has no inherent strength (Petford & Koenders 1998). The outcome of the calculation is that high granular "temperatures" are obtained near the dyke walls, whereas low values appear at the center, thus allowing diffusion of suspended crystals towards the axis of the conduit.

Although the results were promising, the formulation has limitations. For example, at higher crystal loads ($>$about 0.40), the theory is deficient as it holds only

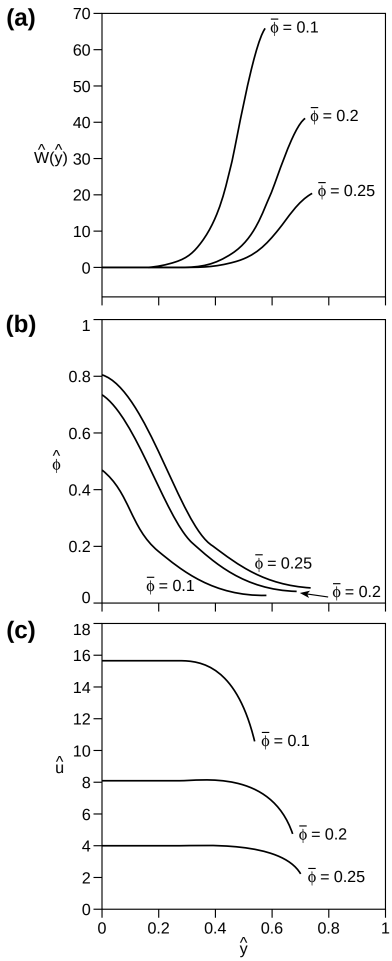

Figure 6 (*a*) Profile showing the calculated fluctuation velocity for three average solidosities ($\bar{\phi}$) of 0.1, 0.2, and 0.25 across a dyke half width with margin at $y = 1.0$. Calculated particle velocity fluctuations are highest close to the dyke margin, implying high granular temperatures in this region. (*b*) Variation in solidosity (crystal content) for the three average solidosities. The predicted increase in particles toward the centre of the flow imply a diffusion effect similar to that seen in flow differentiated dykes (Bagnold effect). (*c*). Normalized particle velocity profile showing the development of an unexpected plug flow (unexpected in the sense that no yield strength was assumed, and the plug-like flow profiles arose spontaneously as a consequence only of particle diffusion).

for small (<0.30) solidosities where the lubrication approximation is invalid. The measure of the energy loss (e) during an encounter between two grains is also an unknown parameter in magmas. The model assumes the grain surfaces are very smooth, but this may not be entirely relevant to crystals in magmatic flows if a distinct angularity is present. Highly anisotropic velocity fluctuations may also pose mathematical problems due to the elliptical nature of the equations. However, although the theory at present contains a number of coefficients as well as the underlying assumptions of dense gas theory, the key physical mechanisms that take place in magma flow are represented, allowing an impression of the relevant phenomena to be assessed. The variation of temperature, velocity, and solidosity fields are all as expected when the average solids density is varied. Although more work is required to obtain the exact constitutive values of the interactive parameters, viscous fluctuation theory has clear application in other important geological processes (e.g., debris flows and pyroclastic flows) where grain interactions play an important role in governing the rheology of the mixture (Figure 7).

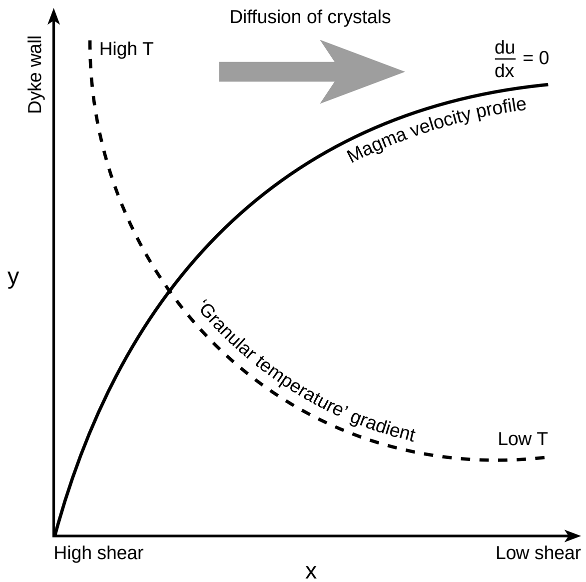

Figure 7 Summary, based on results shown in Figure 6, showing the gradients in velocity (du/dx) and granular "temperature" acting on a magma flowing vertically in a dyke. During shear, crystals diffuse away from the dyke margins, where particle-particle collisions (granular temperatures) are high, toward the center of the dyke (low shear), where collisions are minimal (low granular temperature). The resultant granular temperature gradient allows diffusion of particles from the margin to center of the flow (flow differentiation).

EMPLACEMENT

Dyke ascent of granitic magmas is likely to be "clean and fast" with relatively little change in rheology over the transport distance (Clemens et al. 1997). In contrast, the emplacement process is likely to be rheologically "messy," with large changes in mechanical properties occurring over a relatively small temperature interval (Hutton 1988). Different parts of the crystallizing system may well possess different rheological properties simultaneously. For example, granitic magmas may be emplaced at temperatures in excess of 900°C (Clemens & Petford 1999), and in the presence of a porosity (intergranular melt), thermally activated mechanisms such as dislocation motion and twinning will govern the brittle fracture strength of the mixture (e.g., Kohlstedt & Zimmerman 1996, Rosenberg 2001). Laboratory mechanics studies on the Westerly granite show that above approximately 300°C, and strain rates $>10^{-8}s^{-1}$, rock strength is reduced significantly by high pore fluid pressures. For crystallizing granitic magmas at high but declining melt fractions (approximately 50%), it is assumed that the melt phase in the mush is continuous and that grains are everywhere wetted. However, as crystallization proceeds and temperature falls, this situation is likely to change. We are thus faced with the problem of how to proceed with a rigorous investigation into the effects of competing factors, including strain rate, changing melt composition, and melt fraction on the rheological behavior of the mixture. Solution of this problem requires a combination of field, experimental, and theoretical investigations. Some of the strategies that have been employed are set out below.

Magmatic Fabrics

Changes in magma rheology during crystallization will have a profound effect on the development of magmatic fabrics (foliations and lineations) in plutonic rocks, and a wide range of crystal fabrics have been described [for comprehensive reviews, see Paterson et al. (1998) and Vernon (2000)]. However, it is not clear how best to invert the mechanical information preserved in magmatic fabrics to give clues about magma rheology during emplacement. This is due largely to the fact that fabrics record kinematic histories, whereas the rheology of a material changes in accord with the laws of dynamics (that is, the direction and magnitude of forces acting on it). For example, with reference to Figure 2, does the alignment or tiling of crystals reflect Newtonian, Bingham, or Kelvin (or Maxwell) rheologies? Despite their origin, there is a general consensus, based on evidence from field, microstructural, and AMS (anisotropy of magnetic susceptibility) studies (see Bouchez et al. 1997), that (*a*) magmatic fabrics form late in the crystallization history of plutonic rocks, (*b*) they record generally the last increment of finite strain during magma emplacement, and (*c*) preserve little or no memory of the ascent stage. One mechanism commonly invoked as a means of aligning crystals into a magmatic fabric is convection. Convective instabilities in magmas can arise in principle by a number of mechanisms, including simple heat transfer, buoyancy effects due to changes in composition, and forced convection due to mechanical disturbance arising from intrusion of new magma batches or dykes into a crystallizing chamber. Although it is likely that thermal (free) convection will occur in magmas that are superheated, there is some doubt as to whether large-scale, chamber-wide convective motions can arise in solidifying fluids (Marsh 1996; also Bergantz 1991), despite supercritical Rayleigh numbers (see also Tritton 1988, p. 176, on the irrelevance of the Ra number in natural systems). However, there is good field evidence to suggest that local (e.g., meter-scale) instabilities due to convective motions are important in fabric and structural development in some plutons (e.g., Barrière, 1981, Weinberg et al. 2001).

The physical conditions prevailing during crystallization also have bearing on the timing of magma fabric development during emplacement. Eutectic magmas will develop most of their textures very late in the crystallization interval (say in the last 10%), whereas H_2O-CO_2-bearing magmas may have more linear crystallization trends, with larger amounts of crystals forming over a wider temperature interval. In these instances, the mineral fabrics will have longer to record the total strain (see Scaillet et al. 1997, who argue that estimates of strain rates based on the interpretation of shape preferred fabrics in granitic plutons should be treated with caution).

The effects of volatiles on magma rheology is discussed in a later section.

Magma Mingling

The emplacement of granitic magmas is often accompanied by coeval mafic to intermediate magmas that form in intimate association with the host rock throughout its crystallization interval (e.g., Didier & Barbarin 1991). Although the timing and relationship between (*a*) magmatic fabric development and magma rheology and (*b*) the amount of strain experienced by the magma can be problematic using conglomerations of crystals only, magmatic enclaves offer a means of recovering some (albeit limited) information about viscosity contrasts and the strain history of the pluton while still partially molten. It is a recurrent observation that magma mingling results in enclaves with lobate margins, and that the final shape of the enclave contains kinematic and physical information, most importantly the viscosity and density contrasts with their host, at the time of formation (Fernandez & Gasquet 1994, Fernandez et al. 1997, Scaillet et al. 2000). Once account has been taken of the temperature effects of magma mingling, namely the heat capacities of the magmas and latent heat of crystallization (Sparks & Marshall 1986), many techniques used in the analysis of rocks deformed entirely in the solid state can be applied, with some caution, to make some statements about the rheology of both materials (e.g., Smith 2000). Where pure melts of different compositions begin to interact, the viscosity of the lower temperature melt is higher than the hotter one. This would be the case where basalt at its liquidus temperature is intruded into granitic melt. Note, however, that this is an end-member case and that again the role of crystals is paramount. Hallot et al. (1996) investigated the rheologies of partially crystallized magmas of different compositions, and the geometrical effects that can arise, including (Newtonian) Saffman-Taylor fingering and associated non-Newtonian instabilities such as dendrites. Wavelength analysis using a method first proposed by Ramberg (1981) of structures formed due to gravitational instabilities along the margins of mafic enclaves suggest that at the time of intrusion, the granitic host was commonly less viscous by an order of magnitude than the intruding microgranular material (Bremond d'Ars & Davey 1991). Such viscosity inversions may be commonplace. Recent numerical work (Bergantz 2000) shows that on the larger scale, intrusive contacts between separate magma bodies are liable to collapse by internal wave breaking on a timescale $t \approx L^2/\kappa$, where L is a length scale and κ is the thermal diffusivity, and unlikely to survive intact unless there is a high absolute viscosity in the host magma.

Deformation of a Magma Mush

A current theme in this review has been to seek an answer to the question, "How does a crystallizing magma mush respond to deformation?"

In an attempt to answer this, Koenders & Petford (2000) proposed that densely packed magma suspensions, where crystal contents exceed 50%, are good examples of heterogeneous granular materials with a stiffness that evolves according to a deviatoric evolution rule [note that this usage of granular flow differs from that of Paterson (1995), who used the term to mean the movement of solid particles in a fluid assisted by intergranular diffusion at low strain rates]. The relevant geological situation is shown in Figure 8 for a crystallizing magma near the floor of a chamber. The rheological divisions and geometry of the resulting solidification front are from Marsh (1996). The main deformation mechanism in such materials where interaction between particles is elasto-frictional is one of sliding along conjugate planes (Koenders 1997), the result of which is shear-induced dilatancy in the porous mush. On this assumption, a mathematical model based on Biot's (1941) equations of linear poroelasticity, modified to take into account the effects of dilatancy (see also Koenders et al. 2001), was developed for a layer of deforming magma at high crystal contents. Recast for the consolidation of a compressible fluid moving in a porous material with a position-dependent permeability $k(\mathbf{x})$, porosity (melt fraction) ϕ, and melt compressibility β (Koenders & Petford 2000), the governing equation describing a porous layer (magma mush) undergoing shear is

$$\frac{\partial}{\partial y}\left(k(y)\frac{\partial p}{\partial y}\right) = n\beta'\frac{\partial p}{\partial t} + \frac{\partial \dot{v}}{\partial y}. \qquad (3)$$

The solution to Equation 3 depends largely on the rheological assumptions put forward for the stiffness (dilatant modulus) tensor and the permeability, and the boundary conditions. An analytical solution for the simple case where both these variables are constant is given in Koenders & Petford (2000) for a porous layer with crystal contents ranging from 50% to 70%, with an impermeable top and a weak base (excess pore fluid pressure = 0). They provided a natural scaling (in length and time) for the problem and showed that the excess pore pressure in the

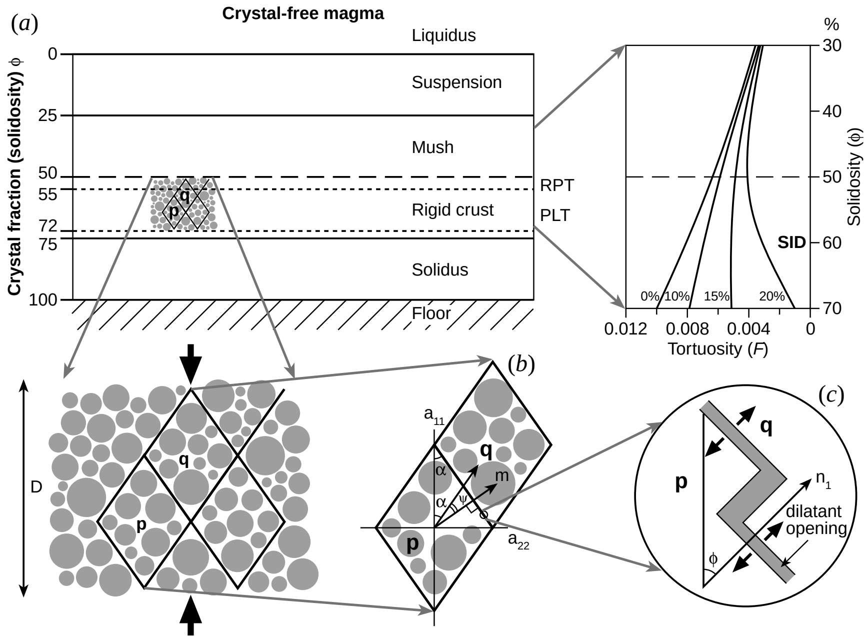

Figure 8 Summary of the multiphase flow processes that may occur in a crystallizing mush at the base of a crystallizing granitic pluton during pure shear. The rheological divisions between the solidus and liquidus are from Marsh (1996). Shown for reference are the rigid percolation (RPT) and particle locking thresholds (PLT) proposed by Vigneresse et al. (1996) for crystallizing magmas. The sequence *a–c* shows in progressively more detail the conceptualized geometry and microstructural effects of shear-induced dilatancy between two coupled but rigid domains (*p* and *q*) in a vertical region of mush ($0.3 < n < 0.5$). If the top of the region is virtually impermeable [pressure gradient $\partial p/\partial y(0) = 0$] and the bottom is regarded as entirely fluid (infinite permeability), then all components of the incremental stiffness tensor relating the solid framework *a* vanish. Assuming no excess pore pressure or skeletal stress in the vicinity of $y = \mathrm{D}$, then $d(t) = 0$ with $p(D) = 0$ and the shear stress rate is defined by $\dot{c}(t)$, and a solution to Equation 3 can be found. The insert shows the relationship between tortuosity and porosity for various percentages of fluctuations in packing density of grains in the mush. Although the range of packing density fluctuation in nature is not known, the effect is likely to be important and strongest at crystal contents ≥50% (i.e., in the region of mush most sensitive to the effects of shear-induced dilation).

deforming layer as a function of stress rate conforms to

$$p = \frac{p^* R \dot{c}_0 D^2}{(k_0 a_1)}. \tag{4}$$

Although the geometry investigated was limited to one dimension, the essence of a number of geophysically relevant situations were captured in this way. For example, the maximum value of pressure at the top of the region is at $t_0 = 1.6D^2(1 + a_1 n\beta)/(k_0 a_1)$. Permeability is known to vary with the grain size (d) and melt viscosity η through $k = Fd^2n^3/\eta$, where F is the tortuosity of the porous matrix and n the porosity (e.g., Dullien 1992). It has also been identified as an important parameter in rheological models of a critical melt fraction (Renner et al. 2000). Taking the stiffness (dilatant modulus) for a densely packed granular material in the range 10^7–10^9 Pa, and a melt viscosity of approximately 10^5 Pa *s* (Clemens & Petford 1999), the time constant is of the order of $t_0 \approx 0.01D^2/d^2$. For a layer thickness (D) of several meters this is approximately 40 days. Note that the time constant is independent of the shear stress rate ($\dot{c}$). Finally, the long-term flow rate ($k_0\partial p/\partial y$) in the middle of the layer is

$$t \to \infty\colon k_0\frac{\partial p}{\partial y} = \frac{-DR\dot{c}_0}{2a_1}, \tag{5}$$

and the change in excess pore fluid pressure as a function of shear stress rate at long times is given by

$$p(y,\infty) = R\dot{c}_0\frac{D^2 - y^2}{2k\theta}. \tag{6}$$

This set of equations marks an important improvement on previous semiquantitative models of granitic melt flow in porous rock (e.g., Vigneresse et al. 1996) in that they provide a link between strain rate and coupled pressure changes in both the solid and melt phase. Magma compressibility (due to the presence of gas) is also taken in to account. In particular, Equation 6 allows a sensitivity analysis of the effect of the transport properties of the mush (expressed through the permeability) to be assessed as a function of strain rate. Results showing the distribution of the excess melt fluid pressure as a function of strain rate are shown in Figure 9 for various positions (y) through a 10-m-thick layer. Figure 9*a* shows two curves corresponding to different strain rates for a fixed permeability of $k = 10^{-15}$ m^2. The maximum excess pore fluid pressures occur at position $y = 0$ and increase with increasing strain rate from an average of approximately 10^2 *Pa* at $\dot{e} = 10^{-14}$ s^{-1} (plate tectonic rates) to >1 MPa at strain rates of 10^{-10} s^{-1}, typical of emplacement loading conditions. Note that increasing the grain size from 1 mm to 5 mm results in a drop of $O(10^4)$ in excess pore fluid pressure. Figure 9*b* shows the effect of varying solids concentration on the mechanical response of the mixture for a fixed strain rate of 10^{-10} s^{-1} and grain size of 5 mm. The effect of increasing permeability of the mush on the development of excess pore fluid pressure is strong, with lowest excess pressures at highest porosities (50%). With increasing solids (crystal) content, the excess pore pressure also increases. In physical terms, this behavior reflects volume changes caused by shear-induced dilatancy in the granular framework, which is more easily achieved at a given shearing stress at lower porosities. At strain rates typical of pluton emplacement (10^{-10} s^{-1}), this becomes a significant effect, creating zones of transient low pressure that act to draw the melt into the mush from above or below. Local flow in the mush produced in this way may

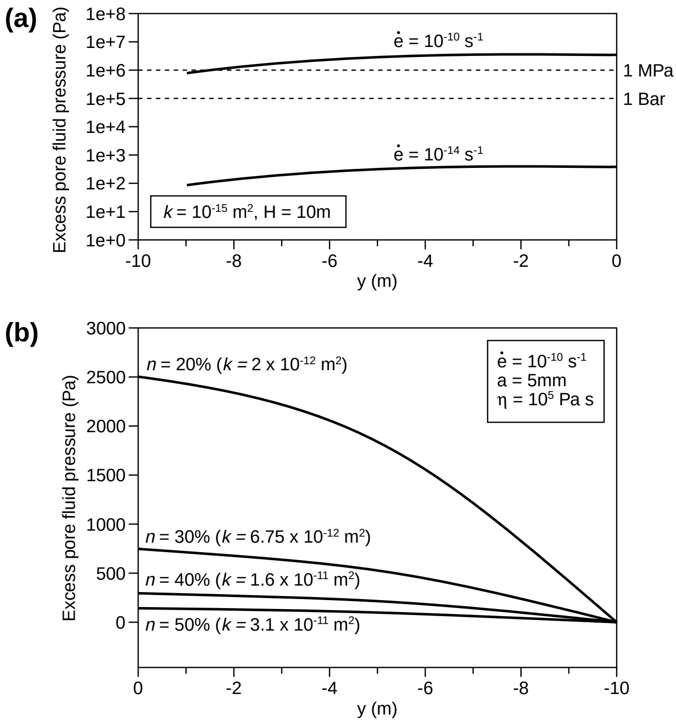

Figure 9 Results based on Equation 6 showing the change in excess pore fluid (melt) pressure in the deforming porous layer depicted schematically in Figure 8, as a function of vertical position (*y*). (*a*) Highest excess pore fluid pressures scale with shear stress rate and can theoretically exceed 1 MPa. (*b*) The effect of increasing crystal loads (decreasing permeability) in the mush results in excess pore fluid pressure at a given rate of strain.

lead to the rotation and alignment of suspended particles, and local blocking and chocking of the mushy zone analogous to filtration effects (cakes) formed during some industrial processes involving two-phase flow (Petford & Koenders 2003).

Three-Phase Mixtures

Volatiles (mostly H_2O) are an important component of many subvolcanic magmatic systems and in deeper-level water-rich plutons. Their effect can be compositional or mechanical. In a series of experiments to determine the rheology of cooling granitic magmas in the presence of a discrete volatile phase, Scaillet et al. (1997) showed the crystallization paths over a range of temperatures (920°C–750°C) and initial H_2O contents (7-4.5 wt% H_2O) were eutectic, with only 20–30 wt% crystals after 90% cooling. Although these H_2O contents are higher than the "ideal" values of Clemens & Petford (1999) for tonalite and leucorganodiorites, the effect was instructive, with the increase in melt viscosity less than twofold, whereas the magma viscosity remained within $O(10^1)$ of the initial value throughout most of the crystallization interval. This is shown graphically in Figure 10, where the "dry" viscosity, calculated using suspension theory, and the magma viscosity based on the curves shown in Figure 3 as a function of increasing water content, are compared for two granitic compositions. The results show the strong modifying effect of water on the bulk viscosity, which during crystallization counters the effects of falling temperature and increasing mass loading that both tend to increase the bulk viscosity. Such trends show that simple models of granite magma rheology based on Einstein-Roscoe theory should be treated with caution. Not all volatiles reduce magma viscosity. In contrast, Scaillet et al. (1997) found that both CO_2 and fO_2 can act to increase the viscosity of the melt phase during crystallization, and they speculated that during slow (diapiric) magma ascent, the level of emplacement may be sensitive to the oxidation state of the magma.

At low pressures, volatiles will be exolved from the melt, forming bubbles, that produce a mechanical effect (e.g., Bagdassarov & Dingwell 1992). In a three-phase mixture of bubbles and solids in a melt carrier phase, the effective viscosity can increase or decrease depending upon the stress rate (Spera 2000). The Capillary number (Ca), the ratio of viscous forces to interfacial surface tension (γ), is defined as

$$Ca = \frac{\eta \dot{e} r}{\gamma}, \tag{7}$$

where $\dot{e}$ is the shear rate and r is the bubble radius. Natural systems define a range of Ca numbers from $1 \leq Ca$ to $\gg$1, and an increase in bubble content from 0% to 50% can result in a factor of 10 decrease in relative magma viscosity. Again, as with static models for two-phase suspensions, strain rate is crucial in determining the rheomorphic behavior of the system. Thus, at low rates of shear, bubbles act as nondeformable inclusions (analogous to solids), and the melt viscosity increases with increasing solids content. At high shearing rates, the bubbles deform, resulting in a decrease in the viscosity of the solid-fluid mixture with increasing bubble content (Spera 2000). At Ca numbers close to unity, viscous dissipation in and around bubbles becomes important as they are deformed by shearing (Rust & Manga 2002). The decrease in solubility of H_2O and CO_2 with decreasing pressure can lead to the limited development of silicate foams (melt-vapor emulsions) characterized by bubble contents in excess of 60%. Although not fully relevant to plutonic systems, such foams may develop during very late stage fractionation of high-level plutons and in some pegmatite dykes.

Emplacement and Rheology of Silicic Volcanic Rocks

Useful rheological comparisons can be made with volcanic rhyolitic and dacite domes and lava flows and intrusive magmas, which are especially relevant at the crossover between hypabassal (subvolcanic) dykes and sills. As with deeper-seated

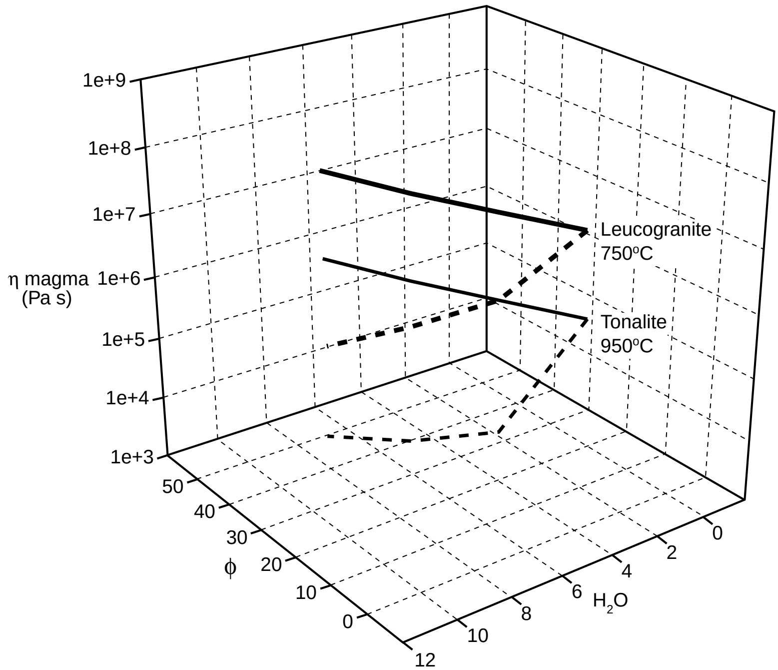

Figure 10 Modification of Figure 2 showing the calculated change in magma viscosity with increasing solids content for the two compositions (tonalite and leucogranite, 800 *MPa*) shown in Figure 3. The two curves for both compositions show the increase in viscosity due to increasing solids where the melt viscosity is fixed (*solid line*) and the effect of H_2O on the viscosity of the melt phase (*dashed line*), resulting in an overall decrease in magma viscosity with increasing mass loading.

plutonic rocks, studies show that the relative viscosities of crystallizing lavas vary by several orders of magnitude due to differences in crystal concentrations and shear stress rate. A major difference between plutonic and extrusive magmas is the strain rate during flow. In general, emplacement of silicic materials at the surface of the earth is characterized by much higher strain rates (10^{-2} to 10^{-6} s^{-1}; Chadwick et al. 1988), than for granitic plutons (approximately 10^{-10} s^{-1}; Fernandez & Castro 1999b). The effects of volatiles are also likely to be far more relevant at lower pressures (see previous section). Taking again the example of the Mt. St. Helens dacite, Pinkerton & Stevenson (1992), using a reinterpretation of measurements by Muarse et al. (1985), show that for solidosities of 0.693, the apparent viscosity of 3×10^{12} Pa s at a strain rate of 10^{-10} s^{-1} drops by three orders of magnitude to 3.5×10^9 at a high strain rate of 1 s^{-1}, with a change in apparent viscosity from 47,730 to 45. Similar calculations at lower solids contents ($\phi = 0.39$) and varying water contents show a factor of three variation in apparent viscosity with strain rate. These changes are significant for plutonic systems, where the strain rates accompanying magma emplacement, especially at high transport rates, require them to be well in excess of plate tectonic values, or those based on time-averaged fault slippage (Petford et al. 2000). Given the strong control of shear stress on the rheology of suspensions, it is of importance to constrain the peak rates of deformation during the emplacement of granitic magmas in the continental crust. It may be significant that many of the deformational and textural features seen in crystal-rich rhyolite lavas (for example the Inyo chain, California), with calculated relative viscosities of between approximately 7–15 Pa s and estimated yield strengths of just 100–200 Pa, record strain rates of approximately 10^{-7} s^{-1}. Given the similarity between these structures and those in plutonic environments where magmas have comingled, it is tempting to speculate that averaged rates may be punctuated by periods of high shear. Such events would be accompanied by rapid dilatancy and local flow phenomena predicted from the models of Koenders & Petford (2000).

SUMMARY

The rheological changes that take place in a granitic magma during ascent and emplacement are summarized in Figure 11. Three ascent modes are implied: 1. rapid dyke ascent, 2. ascent at constant melt fraction, and 3. ascent accompanied by crystallization. Calculated flow rates in dykes suggest that the ascent stage (1) is likely to be adiabatic (Clemens et al. 1997). If so, and the superheated melts are able to resorb entrained material, resulting in a reduction in mass loading as ascent proceeds, then a number of interesting effects may follow. One is a drop in relative viscosity, resulting in faster magma ascent rates. Such behavior would constitute a negative feedback. However, as pointed out by Bagdassarov & Dorfman (1998), if the resorbed solid phase is quartz or a feldspathic mineral, the increase in SiO_2 content in the fluid carrier phase may lead to an increase in melt viscosity that may balance, or exceed, the mechanical effects of reduction in particle number density. Alternatively, resorbtion of ferromagnesian minerals would increase the structural parameter NBO/T, the ratio of nonbridging oxygen to the total oxygen content, leading to a decrease in the viscosity of the melt phase. If these minerals are H_2O-bearing, further reductions may follow. Curve 2 shows a hypothetical ascent path where the crystal load is constant. Flow in a dyke would result in the flow differentiation effects described earlier, with migration of particles towards the core of the flow. This is the picture modeled by Bhattacharji (1966) (Figure 5). Magma emplaced carrying a suspended load may be followed by a period of crystal sedimentation and development of layering as the flow rate drops prior to full crystallization. Curve 3 (possible diapiric ascent) is marked by continuous crystallization. Only in this case will fabric orientations provide information on the ascent process.

A consensus is emerging that most granitic magmas will be emplaced either crystal free or with ≤ 0.50 solids. The emplacement stage is marked by large changes in magma rheology and strength as crystallization proceeds, and is

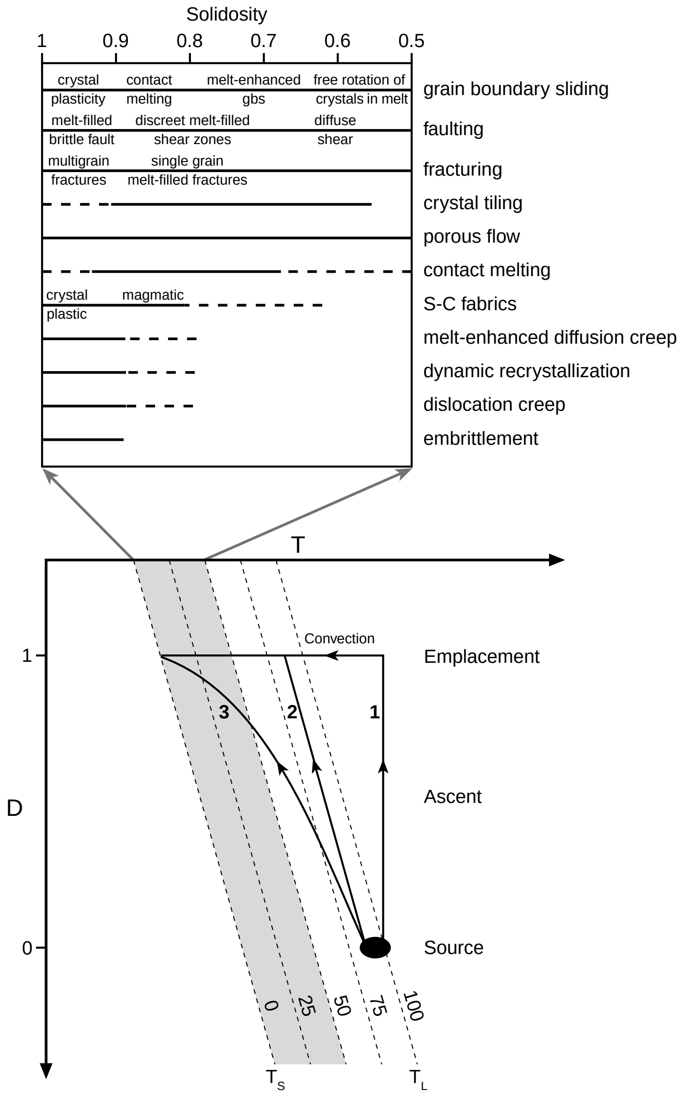

Figure 11 Diagram summarizing the different processes likely to affect the rheological behavior of granitic magma during channelized flow (ascent) and emplacement, based on some of the ideas reviewed in the text. Adiabatic ascent (1) results in superheating of the melt. Shear-enhanced diffusion governs the mechanical component of the upwards flow (2) with moderate crystal loads (up to 25%). Crystallization during ascent (diapiric rise?) is depicted by line 3. Cooling, solidification, and increasing relative viscosity will dominate the emplacement phase and govern the resultant fabrics and structures (summarized after Paterson et al. 1998).

reflected in the fabrics that develop (Figure 11; summarized from Paterson et al. 1998). Undoubtedly, various rheological thresholds will be crossed, although it is still not clear how these might be affected by the combined effects of volatile content and deformation accompanying the emplacement process. Shear-induced pressure changes in the crystallizing mush can now be calculated for porosities in the range 0.5–0.3, along with important microstructural parameters governing flow in the deforming mush, such as the tortuosity, fluctuations in packing density, and grain size. A move away from models of magma rheology and emplacement constrained by averaged rates of fault motion is recommended.

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