

# Non-linear feedback loops in the rheology of cooling-crystallising felsic magma and heating-melting felsic rock

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## ABSTRACT

At least six major parameters control the rheology of partially molten systems: melt content, rate of melt production, reaction to strain of the solid component, reaction to strain of the molten component, temperature and chemical composition of the source rocks. We examine their interactions to understand the rheology of partially molten rocks and partly crystallised magmas. This paper especially focuses on the rheology in the transitional domains between two pairs of thresholds that bracket a transitional regime between a solid-state and a fluid behaviour during melting and crystallisation, respectively. We review related information and point out non-linear effects that develop during heating of melting rocks and cooling of crystallising magmas. Owing to the non-linear interactions, positive or negative feedback loops accelerate or damp the system. Melt in migmatite experiences shear-softening which, along with strain partitioning, facilitates melt segregation. Conversely, the augmenting number of rigid crystals during cooling increases the suspension viscosity (shear-hardening), which soon inhibits magma movement. Those effects reinforce the asymmetry between solid to melt and melt to solid transitions. They severely contradict the concept of one rheological critical melt percentage valid for both melting and crystallisation transitions. Fabric acquisition competes with nucleation and crystal growth, thus leading to hysteresis of the stress-strain rate curves. Implications to field observations include horizontal magma segregation, magma extraction and successive magma intrusions.

**Keywords:** migmatites, magma crystallisation, partly molten rocks, rheology, stick-slip behaviour, fabrics, hysteresis loops

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## INTRODUCTION

Complex interactions between solid and molten phases along with applied stress and resulting strain in partially molten rocks are challenging students of rheological processes. For example, compaction models of magma extraction (Mc Kenzie, 1984) do not explain discontinuous and irregular melt segregation observed in natural shear zones and more sophisticated mathematical formulations fail to explain how melt is extracted in a volume large enough to feed a granitic pluton (Brown *et al.*, 1995; Petford *et al.*, 1993). Experimental studies of partial melting in rocks have chiefly aimed at estimating melt-producing reactions for a given magma type (e.g. Clemens & Vielzeuf, 1987; Patiño-Douce & Beard, 1995) but the small sample sizes are not conducive to assessing mechanical processes during melting. In contrast, experimental deformation of melting rocks constrains mechanisms that accommodate the bulk sample deformation (Rushmer, 1995; Rutter & Neuman, 1995; Kohlstedt *et al.*, 2000). From all issued studies, mechanisms and the amount of melt segregation appear to be extremely sensitive to boundary conditions.

Magma segregation in the continental crust has been approached through the study of migmatites. These metamorphic rocks were partly molten and the crystallised melt forms leucosomes associated with residual components, the melanosome (Mehnert, 1968; Ashworth, 1985; Brown, 1994). Microstructures are clues to deformation mechanisms active during crystallisation (Vernon, 2000; Rosenberg, 2001), but often disappear with further melting. Another approach to understanding the behaviour of partly molten rocks has involved fabric studies of plutonic rocks (Bouchez, 1997). However, plutonic fabrics mostly record the waning strain history that obliterated earlier strain patterns. Therefore, they do not give clues to the magma physical state during segregation and subsequent ascent from the source.

**Partially Molten crustal Rocks** (migmatites and crystallising felsic magmas, here simply referred as PMR) are two-phase systems relevant to investigations of complex behaviours. One phase is a mobile viscous fluid (migmatitic or residual melt) and the second is the stiff viscoplastic or rigid material (restitic matter or crystals). Taken as solid-liquid systems, attempts to get rheological information from experimental and theoretical studies of concentrated suspensions are frustrating for their application to melting and crystallisation (reviews in: Adler *et al.*, 1990; Liu & Masliyah, 1996). In effect, these studies consider successive static situations, each valid for a given particle concentration. However, the continuously changing fraction of the fluid phase with respect to the solid matrix during melting or crystallisation of rocks alters the transport property of the whole system. Connection of melt pockets during progressive partial melting builds up channels through which magma can flow (Laporte & Watson, 1995). Magma crystallisation, on the other hand, builds a crystalline framework through which applied stresses can be transmitted. Therefore, PMRs are transitional stages that cannot be directly compared to suspensions, mushes or slurries in which the solid phase content has no time variation (Fowler, 1987).

Besides temperature and chemical composition of the source, four major physical effects interact and cause non-linear behaviour of PMR. These effects concern (i) material and stress transfer; (ii) strain partitioning between the solid and fluid phases; (iii) melting and crystallisation rates; and (iv) non-linear interactions between each system component, each also presenting non-linear behaviours. The non-linear interactions explain the inadequacy of classical models for rocks undergoing state transitions (Barboza & Bergantz, 1998; Renner *et al.*, 2000).

This paper first reviews key findings pertinent to the rheology of PMR (see also Miller *et al.*, 1988; Vigneresse *et al.*, 1991; Nicolas, 1992; Renner *et al.*, 2000; Rosenberg, 2001). Aiming to investigate implications of non-linear effects on the bulk behaviour of PMR, we examine the non-linear responses of each component of the melt-solid system to stress. We explore these responses under increasing and decreasing temperatures, thus simulating evolutions of melting rocks and crystallising magmas, respectively. During melting, positive feedback loops internally amplify and enhance horizontal melt segregation. Conversely, melt segregates in crystallising magmas under action of negative (counteractive) loops. The discontinuous and irregular motion of melt is controlled by its rheological contrast with the solid phase and affects the subsequent amount, periodicity and composition of extracted melt. We were unable to derive a unique equation or "model" that would fit all melt-involving states, but we suggest several ways of tracking the behaviour of PMR. The present paper

focuses on felsic rocks because we know many examples of migmatites and plutons. The general concepts should also apply to other rock types, with the restriction that thresholds and viscosity have different values, and stress effects different amplitudes.

## **MATTER AND STRESS TRANSFER THRESHOLDS**

Structural studies of PMR often refer to Arzi's (1978) rheological critical melt percentage (RCMP). The viscosity of a dilute suspension of rigid particles in a viscous fluid increases with the number of particles up to a threshold (the RCMP) beyond which aggregates of solid particles lock the system. This concept has been applied to melting rocks by inferring process symmetry (Van der Molen & Paterson, 1979). However, a single and symmetric RCMP for melting and crystallisation does not match geological observation (Vigneresse *et al.*, 1996). We suggest using different thresholds for migmatite and crystallising magma.

### *Liquid versus Rigid Percolation Thresholds*

Melting and crystallisation are transitional states whose rheology can be approached with the percolation theory (Stauffer, 1985), which addresses the transport properties of a given network. During partial melting, melt connection along grain boundaries corresponds to a bond-percolation. The bond is the linking film or tube permitting melt movement from pocket to pocket. Conversely, grain attachment during crystallisation is a site-percolation, a term that refers to the creation of a continuum of touching particles, each one being located at a site. In bond- and site-percolations, the volume of material ensuring connection differs. The bond-percolation threshold is significantly smaller than the site-percolation threshold because it is easier to build links along continuous grain boundaries than attach particles of finite size (Stauffer, 1985). For melting rocks and crystallising magmas, these situations have been coined the Liquid Percolation Threshold (LPT) and Rigid Percolation Threshold (RPT), respectively (Vigneresse *et al.*, 1996; Fig. 1). In felsic magmas, measured and calculated values of 8 volume % of melt for the LPT and 55 volume % of solid for the RPT vary by a few % according to the actual mineralogy (Vigneresse *et al.*, 1996). Both thresholds should show up in the rheological parameters since the shear modulus becomes null when stresses are transmitted through the melt.

### *Melt Escape and Particle Locking Thresholds*

The cohesive strength of migmatites disappears at about 20-25 volume % melt, after connection of melt pockets. The Melt Escape Threshold (MET) marks the loss of cohesion of the source rock during melting (Fig. 1). In crystallising magmas, close packing of phenocrysts locks the solid framework at much higher crystal proportions (70-75 volume %) than the onset of interactions between solid particles, which results in a loose packing (Vigneresse *et al.*, 1996). The Particle Locking Threshold (PLT) corresponds to total locking. The exact threshold values depend on external factors such as composition and applied stress.

## **NON LINEAR MELTING AND CRYSTALLISATION RATES**

### *Melting in source rocks*

Melt content and production are non-linear in space and time because melting rocks include several minerals that react differently to temperature variations (Fig. 2).

Incipient melting is governed by the free energy of crystals and thus varies from grain to grain (Laporte & Watson, 1995). The lower free energy a mineral pair presents, the easier melting occurs, such as at triple (in two dimensions) or quadruple (in three dimensions) junctions. Minerals and fluids may be randomly spread in rocks but a given assemblage of minerals with similar melting capacities is irregularly distributed. For instance, a rock with 30% quartz in randomly dispersed small grains has not the same melt production, at a given temperature, as a rock of same composition with quartz concentrated in a single domain. Accordingly, melting is heterogeneously distributed.

Experimental melting has shown that production of more melt than the MET only occurs when water-saturated pelitic, biotite-muscovite- or biotite-bearing rocks leave the stability field of biotite, above 820 °C (Vielzeuf & Holloway, 1988; Patiño-Douce & Johnston,

1991; Gardien *et al.*, 1995; Patiño-Douce & Beard, 1995). For a given bulk composition, the increase in melt content is not rate constant with temperature but varies as different mineral-out reactions are overstepped (e.g. Vielzeuf & Holloway, 1988). Around 800°C, heating of about 75°C suffices to metasediments to go from the LPT to the MET, yielding about 25 % melt. Conversely, above 850°C, further heating by about 200 °C is required to increase melt by 25 %. These variations are reflected in melt production curves, which are concave downward for metasediments and concave upward for metaigneous rocks (Fig. 2).

Rocks exhibit, on a small (centimetric) scale, compositional variations to which their bulk response to temperature, including melting, is very sensitive (Fig. 3). Thermodynamic data constrain the melting temperature of any given mineral, but eutectic behaviour buffers melting as soon as this mineral is mixed with other phases (Johannes & Holtz, 1996). In the system quartz-orthoclase-albite-anorthite-H<sub>2</sub>O, cotectic surfaces of each eutectic sub-system plunge toward a minimum (Johannes & Holtz, 1996). The starting composition on the cotectic surfaces and the orientation of the slope on this surface control the chemical evolution of the melt (Fig. 3). Consequently, the amount of melt changes considerably for small variations in temperature and chemical composition of the source rock. Introduction of a fluid further lowers the melting temperature of minerals (Mehnert, 1968; Johannes & Holtz, 1996). Therefore, water generated by dehydration of hydrous minerals is immediately consumed for melting. Since fluids in rocks are heterogeneously distributed, melt production in migmatites is non-linear in space and time, even if a linear temperature distribution is assumed.

#### *Crystallisation of magmas*

The bulk crystallisation rate in magma chambers varies spatially because it depends on heat exchange between magma and host rocks. In particular, the crystallisation front of a pluton advances irregularly towards its core from tortuous margins where a crystal suspension becomes rapidly a mush and ultimately forms a rigid envelope. This envelope insulates the residual magma (Marsh, 1996) resulting into an unequally distributed growth-rate.

Besides, magma crystallisation is essentially non linear with temperature. Changes in slope of melt fraction curves (Fig. 4) mark the appearance or disappearance of mineral phases.

## **STRAIN PARTITIONING**

The contrasting rheology between coexisting phases (new crystals versus residual melt or restitic crystals versus new melt) in PMR induces strain partitioning at the grain-scale (solid grain/adjacent viscous medium) and crustal-scale (non molten country rocks/partially molten body). Partitioning is fundamental because it increases the vorticity, hence the non-coaxial component of deformation within the weak melt (Vigneresse *et al.*, 1996; Vigneresse & Tikoff, 1999). This non-coaxial component addresses two important rheological issues: viscosity and shear softening or hardening (called shear thinning or shear thickening, respectively, in the literature on physics of materials).

#### *Strain partitioning during partial melting*

The melt film between mineral grains triggers strain partitioning as soon as stress is applied because the melt viscosity is several orders of magnitude lower than that of grains. Indeed, a weak phase within a deforming matrix is a major cause of strain localisation (Burg and Wilson, 1987; Olgaard, 1990; Bazant, 1988; Herwegh *et al.*, 1997). Strain rate in the melt is thus faster than in the solid matrix, hence allowing melt redistribution (Dell'Angelo & Tullis, 1988; Bagdassarov *et al.*, 1996). Observations on migmatites suggest that melts, represented by leucosomes, tend to be channelled into shear zones (Brown *et al.*, 1999). Occurrence of shear zones contributes to bulk strain softening of the rock mass; channelling in these deformation sites involves changes in grain size and/or water content that modify the strain-stress curve. An important consequence is reduction of the bulk viscosity during deformation to accommodate fast shear strain localised into the melt. A lower viscosity takes up more strain into the melt whose displacement accelerates, which in turn induces further non-coaxial deformation. Shear softening, is important at strain rates faster than  $10^{-4.5} \text{ s}^{-1}$  for magma compositions that range from nephelinite to rhyolite (Webb & Dingwell, 1990). As a

corollary, syn-deformation melting fundamentally has rheological consequences since it leads to faster extraction of less viscous melt.

#### *Strain partitioning during magma crystallisation*

Strain partitioning mostly concerns the strong crystals during magma crystallisation. Crystal-poor magmas are viscous media in which shear-hardening results from crystals interacting while they rotate (Jeffery, 1922; Barnes, 1989; Jezek *et al.*, 1994; Arbaret *et al.*, 1997; Hoffman, 1998). At higher crystal contents, crystals form clusters that behave as a single particle with a new aspect ratio, and thus different rotation rate (Tikoff & Teyssier, 1994; Joseph *et al.*, 1994). Further crystallisation results in a non-cohesive skeleton of solid particles whose rigidity is sufficient to sustain external constrictive but not extensive stress (Guyon *et al.*, 1990); in addition, the skeleton cannot transmit shear stress because particles are not tied to each other. While crystals aggregate, dilatancy becomes necessary in crystal-rich magmas because crystals must disconnect to override and pass each other before rearranging in denser packing (Barnes, 1989; Bashir & Goddard, 1991). The entire system becomes constrained once all crystals touch each other, which marks closed packing. At higher crystal concentration, increased contacts over-constrain the system and the ability of crystal to move is reduced by several orders of magnitude (Moukarzel, 1999). In summary, particle rotation, strain partitioning and dilatancy impart to non-cohesive aggregates some strength but a permanently evolving rheology.

### **RHEOLOGY OF PMR**

Rheology is expressed either in a stress/strain ( $\sigma/\epsilon$ ) curve whose instantaneous slope is the shear modulus, or in a stress/strain rate  $\sigma/\dot{\epsilon}$  curve whose instantaneous slope is viscosity:

$$\eta = \sigma / \dot{\epsilon} \quad (1)$$

(e.g. Ranalli, 1995). In both cases, the instantaneous slope refers to either the linear slope if material is Newtonian or to the local property if the ( $\sigma/\dot{\epsilon}$ ) curve is non-linear. We first consider solid matrix and melt separately, and then try to understand how they can influence each other's rheology when they coexist in a PMR.

#### *Rheology of solid state rocks in sub-melting conditions*

Partial melting of crustal rocks takes place at 700-1000 °C and 500-900 MPa (Mehnert, 1968; Ashworth, 1985). The solid-state rheology of rock-forming minerals under these pressure-temperature conditions is typically written:

$$\dot{\epsilon} = A \sigma^n \exp(-Q / RT) \quad (2)$$

where A is a grain-size sensitive constant, n the power-law coefficient, Q the activation energy, R the perfect gas constant and T the absolute temperature (Kirby & Kronenberg, 1987; Carter & Tsenn, 1987; Wilks & Carter, 1990).

We selected experimental calibrations on granite, amphibolite and felsic granulite because they probably represent the bulk upper, intermediate and lower crusts, respectively (Table 1). To infer values of the equivalent viscosity, we first computed the differential stress ( $\sigma$ ) with respect to temperature from 700 to 1000 °C and strain rate from  $10^{-16}$  to  $10^{-6}$  s<sup>-1</sup>, which are parameters relevant to natural melting conditions (Fig. 5). In this window, stress at a constant strain rate decreases with temperature for all rock types. It is nearly identical for amphibolite and granulite, and overlaps the high strain rate stress in granite. We will use the amphibolite parameters to model the behaviour of mid-crustal source rocks.

#### *Melt rheology*

Computed viscosity of molten silicate (Bottinga & Weill, 1972; McBirney & Murase, 1984; Baker, 1998; Petford & Clemens, 1999) matches experimental measurements (Ryan & Blevins, 1987; Ryerson *et al.*, 1988; Spera *et al.*, 1988), which themselves fit models for granitic melts (Goto *et al.*, 1997). Viscosity values of felsic melt, supposed to be Newtonian,

range from  $10^6$  to  $10^8$  Pa.s. We considered an Arrhenian variation with temperature following the law:

$$\eta = A_n \exp(-E_0 / RT) \quad (3)$$

in which  $A_n$  is a constant and  $E_0$  the activation energy of the system (Spera *et al.*, 1988). Heating from 800°C to 900°C corresponds to one order of magnitude reduction of the viscosity. This is negligible compared to the range of viscosity variation following changes of stress intensity (Fig. 5). Therefore a Newtonian approximation is valid enough between 700 and 1000°C and at strain rates  $< 10^{-4.5} \text{ s}^{-1}$ . We adopted a stress-independent viscosity ranging from  $10^6$  to  $10^8$  Pa.s. In a linear relationship between strain rate and a stress of about  $10^7$  Pa (10 MPa), these viscosities correspond to strain rates of  $10^{-1}$  and  $10^1 \text{ s}^{-1}$ , respectively.

However, shear softening reduces viscosity by two orders of magnitude at strain rates higher than  $10^{-4.5} \text{ s}^{-1}$  (Webb & Dingwell, 1990). This effect is responsible for the break in the slope of viscosity isovalues (Fig. 6) and departure from the Newtonian behaviour (Dingwell *et al.*, 1996).

#### *Rheology of migmatites during melting*

To express the temperature/strain rate relationship in terms of viscosity, we extract from Figure 5 the stress values obtained between 700 and 900°C, each value corresponding to one strain rate through viscosity (Eq. 1). We obtain a parallelogram that contains every possible stress vs. strain rate point of migmatites in the considered temperature range (Fig. 6). The parallelogram covers the first-order rheology of crustal rocks undergoing melting because it integrates both the rheology of crustal materials (Eq. 2) and the rheology of silicate melts (Webb & Dingwell, 1990). The lines of equivalent Newtonian viscosity are deduced from the stress and strain rate coordinates. Those passing through the parallelogram field range from  $10^{22}$  to  $10^{13}$  Pa.s.

In a second step, we restrict temperatures between 750 and 850°C, the incipient melting temperature of a granitic protolith and the dehydration melting temperature of biotite, respectively (Johannes & Holtz, 1996). We also constrain the differential stress between 5 and 50 MPa, considering that higher figures would produce fracturing in the matrix and lower ones would produce extremely slow strain rates ( $< 10^{-14} \text{ s}^{-1}$ ). The corresponding lines bound a small parallelogram that spans strain rates from  $10^{-14}$  to  $10^{-7} \text{ s}^{-1}$  (Fig. 6). The equivalent viscosity of melt-containing rocks lies now between  $10^{20}$  and  $10^{14}$  Pa.s. By contrast, the viscosity of the small amount of melt staying in the matrix, below the LPT, is  $10^6$  to  $10^8$  Pa.s (Petford & Clemens, 1999).

One can now discuss consequences on the rheology of progressively melting migmatites. As long as the rock contains few melt, crystal grains of the matrix remain attached to each other. Stress applied to one grain is transmitted to its neighbours up to some correlation length that marks the range of crystal interaction. In the melt, interaction and chemical exchange are possible only by diffusion over a shorter correlation length than stress in the matrix. Thus, the bulk rheology of low-melt migmatites is nearly that of the matrix. Migmatite bodies showing similar structural orientations as unmolten country rocks support this proposition (Weber *et al.*, 1985; Barbey *et al.*, 1990; Sawyer, 1991). However, this bulk behaviour does not explain local and independent melt mobility indicated by leucosomes cutting regional structures (e.g. Hopgood, 1999), which we will comment on later.

#### *Viscosity and rheology of crystallising magma*

The growing number of crystals and dilatancy increase the viscosity until particle aggregation inhibits movement in crystallising magma. Previous studies investigating how these and other processes interact and affect viscosity refer to the Einstein (1906) and Roscoe (1952) expression:

$$\eta = \eta_0 (1 - \Phi / \Phi_{\max})^{-n} \quad (4)$$

This power law equation implies that the viscosity ( $\eta$ ) of a suspension depends on the

volume fraction ( $\Phi$ ) compared to the maximum possible concentration ( $\Phi_{\max}$ ) of solid particles in a fluid of viscosity ( $\eta_0$ ). Since ( $\eta$ ) increases with particle concentration, hence with cooling, magma in course of crystallisation should show shear hardening. In fact, the complex evolution of viscosity must be considered under various situations. (i) At low shear rate, the exponent  $n$  is around 2.5, the system behaves as a dispersed suspension and shear-hardening occurs for a given effective volume fraction while particles start clustering (Joseph et al., 1994). (ii) At faster strain rates, a more complex exponent ( $n\Phi_{\max}$ ), equal to about 1.8, replaces  $n$  (Krieger & Dougherty, 1959). The suspension of particle clusters must have a lower viscosity than at low strain rate and undergoes rheofluidization (the viscosity increases less than it should under a constant strain rate, Quemada, 1998). (iii) At a significant number of crystals, packing fixes viscosity at a plateau value and magma behaves as a granular medium with an intrinsic strength.

We first consider the latter case. For a dry or wet granular medium the static friction angle ( $\phi$ ) relates tangential ( $\tau$ ) to normal ( $\sigma$ ) stress:

$$\tau = \sigma \tan \phi \quad (5)$$

where  $\tan \phi$  is a coefficient that depends on friction at grain contacts (Scholz, 1990; Marone, 1998). It commonly ranges between 0.85 and 0.65, resulting in a friction angle of about  $30^\circ$  (Chester, 1995; Blanpied *et al.*, 1995). In dynamic conditions, the friction coefficient varies with the displacement velocity, the slip history and the sample porosity (Marone, 1998). In addition, dilatancy implies that a new angle  $\varphi$  is defined, which sums the particle/particle friction angle and the parameters a particle must overcome to slide freely over adjacent particles before close packing.

If magma becomes too sluggish, two mechanisms compete.

- i) Particle aggregation lowers the melt/crystal surface energy, which is the difference between twice the solid/liquid surface energy and the solid/solid surface energy (Laporte & Watson, 1995). The aggregation kinetic is limited by the collision frequency between the particles, which directly depends on both the particle concentration ( $\Phi$ ) and the shear rate.
- ii) Particle growth tends to minimise the melt/crystal surface energy by narrowing the size distribution of aggregates and smoothing surfaces (Voorhees, 1985). It is diffusion limited with a characteristic length of about the individual particle equivalent radius. The characteristic diffusion time is about 1000 s for about 1 mm big crystals and chemical diffusivity of about  $10^{-9} \text{ cm}^2\text{s}^{-1}$  around  $900^\circ\text{C}$  (Johannes & Holtz, 1996). Consequently, textural coarsening is likely to take place in magma that cannot flow over long distances owing to lack of space or fast cooling.

Experiments with up to 30-40 % crystals (Lejeune & Richet, 1995) are consistent with Eq. 4. They imitate the large-scale behaviour of a crystallising magma whose viscosity increases with decreasing temperature. The viscosity of a leucogranitic magma close to its solidus increases by about one order of magnitude for  $< 50^\circ\text{C}$  cooling. However, this bulk response does not apply to small-scale behaviour and does not take into account dilatancy (Onoda & Liniger, 1990).

Strict application of Eq. 4 states that viscosity increases with the solid concentration up to the maximum packing ( $\Phi_{\max}$ ) (Adler *et al.*, 1990; Liu & Masliyah, 1996). Close to  $\Phi_{\max}$ , particles are in contact. Clusters then deform only above some critical stress level, but are destroyed if the strain rate is fast and the associated stronger flow rearranges particles in closer packing.

Developing the power function of the Einstein-Roscoe equation in series indicates that crystallising magma has a yield stress and should be considered under a constant strain rate description

$$\eta = \eta_0 (1 - \Phi / \Phi_{\max})^{-n} \approx \eta_0 (1 + n \Phi / \Phi_{\max}) \quad (6)$$

Since high strain rate leads to a strong readjustment of the packing (Onoda & Liniger, 1990; Rogers et al., 1994), we assume that  $\Phi_{\max}$  depends on some strain rate  $\dot{\epsilon}_0$ , which writes:

$$\Phi_{\max} = \Phi_0 \dot{\epsilon}_0 \quad (7)$$

leading to

$$\eta = \eta_0 + \eta_0 \Phi / \Phi_0 \dot{\epsilon}_0 \quad (8)$$

Multiplying both sides by strain rate, and putting all second right hand members in a constant  $\alpha(\Phi_0, \dot{\epsilon}_0)$ , yields the equivalent equation:

$$\sigma = \sigma_f + \eta_0 \Phi \alpha(\Phi_0, \dot{\epsilon}_0) \dot{\epsilon} \quad (9)$$

This equation describes a Bingham flow with a yield stress ( $\sigma_f$ ) (Adler *et al.*, 1990). Yield stress is important for low applied stress and hence low strain rate. The few experiments on basaltic or dacitic lavas measured yield strength values of 17 to 78 Pa (Gauthier, 1973; Pinkerton & Norton, 1995; Cashman *et al.*, 1999). The yield strength of granitic magma is not known, but evidence of Bingham behaviour is recognised from phenocryst fabrics varying from the border to the centre of small veins (Shaw, 1965; Komar, 1976). Strain is irregularly distributed in space and time. It depends on the transmission of stress from the ambient stress field, and on friction between particles. The yield strength decreases with crystallisation, thus explaining crystal accumulation at rims of granitic intrusions (Pons *et al.*, 1995; Brun *et al.*, 1990).

## BEHAVIOUR LOOPS

Combined strain partitioning and non-linear rheology produce feedback loops that bolster the non-linearity of PMR. Positive feedback loops amplify processes and are effective during partial melting. Conversely, negative loops over-dampen the rheology of crystallising magma.

### *Positive feedback loops in melting*

Interactions between melt-production rate, strain partitioning and shear softening act as positive feedback loops enhancing melt segregation (Fig. 7). Heating produces more melt and lessens the melt viscosity; a lowered viscosity partitions more strain into the melt, which increases strain rate; shear softening diminishes further the already low melt viscosity. This process combination favours melt extraction.

The bulk viscosity of matrix-melt systems remains controlled by the relaxation of the matrix when melt percolates it. The bulk viscosity should decrease with increasing amount of melt, hence with temperature, because the melt viscosity is lower than that of the matrix. Once there is enough melt to break matrix cohesion, the bulk viscosity becomes nearly the melt viscosity.

To examine the influence of deformation on the feedback loops, we modelled the quantity of melt extracted under coaxial and non-coaxial deformation (Fig. 8). In the absence of tectonic deformation, heat controls melt production and the only applied force is gravity. The system is static and magma extraction stops rapidly after extraction due to coaxial compaction has removed the overabundant magma. Magma is easily extracted when shear deformation is applied, though the extracted volume is erratic in time (Fig. 8), an extraction behaviour that has been observed in analogue models (Barraud *et al.*, 2001).

### *Negative feedback loops in magma crystallisation*

In cooling magmas, interactions between shear hardening and strain partitioning generate negative feedback loops that ultimately choke the system (Fig. 9). With cooling, the growing proportion of crystals adds stiffness to the magma while melt viscosity increases. A robust estimate, using Eq. 3, indicates that viscosity of a leucogranitic magma increases by 2 orders of magnitude (only one with Arrhenian viscosity) when the crystal content increases from 50% to 66 %. This roughly corresponds to only 20°C cooling (Fig. 4). The combination of cooling and increasing number of crystals raises viscosity by 3 orders of magnitude. Under constant applied stress, the strain rate is correspondingly reduced by 3 orders of magnitude.

This feedback loop counteracts strain partitioning (Fig. 9) and thus reduces magma mobility, slowing down the magma flow. With the onset of crystallisation, magma becomes less able to internally reorganise and will cool down close to where it is. Correspondingly, magnetic susceptibility in felsic magmas reflects the last strain increment (Bouchez, 1997). Though transposition to strain is not direct, anisotropy values, commonly lower than 3 %, do

not reflect high strain rates. They rather indicate that magma is transported en masse with little internal strain, after it has upwelled to upper crustal levels.

## DISCUSSION

### *Rheological difference of melting and crystallising PMR*

Formulation of bulk properties of composite systems commonly uses arithmetic or geometric average values of each material property. For instance a system composed of two materials with viscosity  $\eta_1$  and  $\eta_2$  has an arithmetic mean viscosity  $\eta_a$ :

$$\eta_a = \Phi \eta_1 + (1 - \Phi) \eta_2 \quad (10)$$

and a geometric mean viscosity  $\eta_g$ ,

$$1/\eta_g = \Phi / \eta_1 + (1 - \Phi) / \eta_2 \quad (11)$$

where  $\Phi$  is the percentage of the phase with subscript 1. These equations are the Voigt and Reuss approximations, respectively (Guéguen & Palciauskas, 1990). They correspond to a system under constant strain rate (Eq. 10) or under constant stress (Eq. 11). Focusing on the rheological differences of PMR, we can approach state equations for melting and crystallisation transitions with some intrinsic assumptions.

For melting, we start from the matrix rheology and the melt viscosity calculated above. Implicitly, we assume that the two-phase material is treated at constant stress and thus strain partitioning exists between melt and matrix. For crystallisation, we take viscosity from Eq. 4, thus formulating the system under constant strain rate.

We use tonalite and pelite as extreme rock types for which melt content in function of temperature is known (Vielzeuf & Holloway, 1988; Rutter & Wyllie, 1988). Computing an average viscosity for both rock types, we obtain different curves for melting and crystallisation transitions; these curves do not show the superposition anticipated by the RCMP concept (Fig. 10). They trace a cycle suggesting hysteresis, thus a different mode of energy loss during melting and crystallisation transitions (Brokate & Sprekels, 1996). The large viscosity change, around 700°C for pelite and at 950°C for tonalite, reflects the Liquid Percolation Threshold during melting; the viscosity change around 775°C (pelite) and 1025°C (tonalite) during crystallisation reflects the Rigid Percolation Threshold.

### *Continuous versus discontinuous motion of one phase in PMR*

We have mentioned secant leucosomes indicating expulsion of melt from migmatites. Similarly, magmatic fabrics provide evidence for bulk laminar flow at slow strain rates (Bouchez, 1997; Paterson *et al.*, 1998). However, metre-scale disturbances of the larger scale fabrics and heterogeneous enclave patterns (Tobisch *et al.*, 1997) show that the bulk flow is locally perturbed. Depending on the scale, the description of PMR thus relates either to a liquid, or to a viscous solid, or both in any proportion. Besides scale lengths, the time span for melting or crystallisation to take place is very different.

Melt segregation through strain partitioning in migmatites drives the system from a stage of non-connected and heterogeneously distributed melt below the MET to a stage of connected melt films focusing strain. The evolution occurs at slow melting rate and slow strain rate because the system is governed by the matrix rheology. The discontinuous and repetitive melt motion obtained using Lagrangian modelling is similar to stick-slip motion (Fig. 8). It represents an alternating switch from matrix-dominated to melt-dominated rheology. In crystallising magma, melt escape obeys the discontinuous motion of the solid crystalline framework. Regions of smaller friction coefficient concentrate deformation into a smaller volume, inducing a feedback loop that localises strain (Barnes, 1989; Marone, 1998) and local extensional zones act as sinks for the residual melt.

## GEOLOGICAL RELEVANCE

### *Magma extraction*

We argue that magma extraction results from both melting (temperature effect) and rheology (tectonic stress and intrinsic property of the melt).

In crustal rocks melt is essentially generated from melting of muscovite, quartz and plagioclase (Thompson, 1982). Resulting leucosomes lack ferromagnesian minerals (e.g. amphibole, biotite). Therefore, melting quickly followed by melt ascent should not provide magma of granitic composition. En-route assimilation of ferromagnesian minerals from intruded country rocks is unrealistic in terms of volume. Consequently, the migmatite leucosome-melanosome system must undergo further melting that involves ferromagnesian minerals to produce a granitic melt (Sawyer, 1998). For example, breakdown of biotite liberates water and increases the melt volume, which allows for overcoming the MET above which melt may escape (Petford *et al.*, 2000).

Competition between horizontal and vertical segregation has an influence on melt productivity. According to Le Chatelier's principle, fast melt production slows down the melting rate. Conversely, fast melt extraction increases melt productivity. However, if the melt is extracted upwards, it rapidly gets disconnected from its source, which becomes depleted in material adequate for melting and cannot produce more melt. If melt segregates horizontally and remains close to its source, low melting rate may produce a lot of melt with a homogeneous composition. We conclude with Sawyer (1994) that migmatites crystallised with a high melt fraction denote low ambient deformation, unable to quickly extract significant melt volumes. On the contrary, source regions with melt productivity high enough to produce magma and feed granitic plutons should have been submitted to intense deformation. The remaining source region would comprise restitic rocks, devoid of melt (i.e. granulites).

### *Equilibrium versus disequilibrium compositions*

Strain partitioning induced by horizontal tectonic stresses enhances the horizontal separation between melt and residual phases which is further reinforced by the melt flow channelled between unmolten layers (Vigneresse & Burg, 2000). Horizontal segregation has two major consequences. First, melt may chemically re-equilibrate because it remains for some time at constant pressure and temperature conditions while concentrating at some level. Second, upward flow of felsic magma cannot be reduced to a pervasive percolating fluid through a more or less porous crust because melt is collected before extraction in dykes (Petford *et al.*, 2000).

Melt production and segregation influences the equilibrium textures. For instance, if both melting and melt extraction rates are fast, melt can hardly re-equilibrate with its matrix (Sawyer, 1994). A robust estimate of the influence of melt circulation in a migmatite can be obtained using the Péclet number, which defines the ratio of diffusion versus advection effects (Crank, 1975). For this purpose, we consider chemical diffusion of elements that are present within the leucosome.

A migmatitic section of length  $L$  is idealised by a series of leucosome veins separated by a distance  $d$  and in which melt segregates with a velocity  $v$  (Fig. 11). The time ( $t_{adv}$ ) for melt to segregate is given by:

$$t_{adv} = L / v \quad (12)$$

By comparison, a rough estimate of the time ( $t_{diff}$ ) for an element ( $s$ ) to diffuse is given by:

$$t_{diff} = d^2 / D_s \quad (13)$$

where ( $D_s$ ) is the diffusion coefficient of the considered element.

The Péclet number ( $Pe$ ) is the inverted ratio between the two time quantities:

$$Pe = v d^2 / D_s L \quad (14)$$

Disequilibrium occurs when advection (mass transport) is faster than diffusion, or if  $Pe$  is larger than 10 (Ottino, 1989). In migmatites, we adopt a length  $L$  of about 1 m and vein spacing about 0.1 m, which results in a  $d^2/L \approx 0.01$ . In granitic magmas, tracer diffusivities range from  $10^{-16}$  to  $10^{-7}$   $\text{cm}^2\text{s}^{-1}$  (Jambon, 1982). Slow diffusion of Si and Al controls diffusion

of non-alkaline elements that have chemical diffusivity values of about  $10^{-9} \text{ cm}^2\text{s}^{-1}$  around  $900^\circ\text{C}$ , Carmichael *et al.*, 1974). Excepting Li and Na, all other cations have diffusion coefficients smaller than  $10^{-7} \text{ cm}^2\text{s}^{-1}$  at about  $900^\circ\text{C}$  (Johannes & Holtz, 1996). Whereas diffusion time can easily be computed, advection time is not directly measurable and must be estimated. Calculated viscosities yield strain rates faster than  $10^{-4} \text{ s}^{-1}$  in melt whilst strain rates range from  $10^{-14}$  to  $10^{-8} \text{ s}^{-1}$  in the solid matrix (Fig. 6). If one assumes that strain rates for material to move by about 1 m (the section length  $L$ ) convert to velocity, Péclet numbers of ca.  $10^7$  for melt and below  $10^3$  for the matrix are obtained. Those Péclet numbers indicate that advection largely dominates over diffusion. Therefore, melt segregation by melt movement within the matrix leads to melt disequilibrium. It can be argued that segregated melt staying in the matrix can re-equilibrate if the time interval between deformation pulses is long (about 1 Ma) as under a bulk strain rate of  $10^{-10} \text{ s}^{-1}$ . This time interval is very long compared to that between migmatite melting and upwelling (about 1 Ma in total, Vanderhaeghe & Teyssier, 1997). Accordingly, we expect that most migmatite melts show disequilibrium composition (see also Sawyer, 1991).

#### *Fabrics in imbricated magma intrusions*

Interactions between PMR include imbricated magma intrusions from the injection of melt into another melt to the intrusion of a melt into a solid rock (Bédard, 1993; Fernandez & Gasquet, 1994; Hallot *et al.*, 1996; Fernandez *et al.*, 1997). Clearly, different structures should document different crystallisation states between the RPT and the PLT of the older, intruded magma.

When two melt batches interact, mixing leads to chemical homogenisation. New magma dismembers the fabric of the intruded, loosely connected framework of particles; similar new-fabric patterns in both magmas integrate concordant contacts (Fig. 12). The two magmas may partly exchange mobile elements and a profile across a facies (paragenesis) change displays a smooth variation of those elements. A more crystallised, older magma batch begins to sustain stress and resist to new intrusion, maintaining its bulk fabric. However, fabric variations close to contacts reflect mechanical interaction (Vigneresse & Bouchez, 1997). Exchange of incompatible elements is limited, but compaction of the earlier magma expels residual fluids so that near-contact enrichment in Th, Rb, Ba may be expected (Fig. 12). Sharply discordant fabrics indicate that new magma breaks and disrupts a former, consolidated magmatic body. No large-scale chemical exchange should then take place.

Shear zones in a partly crystallised magma compare to proto-faults that disrupt the fabric of a non-consolidated framework of crystals (Brun *et al.*, 1990; Pons *et al.*, 1995; Smith, 1996; John & Stunitz, 1997). They indicate that fracturing is the response of PMR to a new, high strain rate intrusion (Dingwell, 1997).

## **CONCLUSION**

We have examined the relationships that occur in migmatites between melting, melt segregation and deformation in one hand, and in felsic magmas between crystallisation, deformation and crystal interactions on the other hand.

Solid to suspension transitions during melting and crystallisation are not symmetrical. Two thresholds that depend on the respective percentage of melt and solid phase bound three rheological stages. For melting, a transitional rheology takes place between 8 and 20-25 % melt. During crystallisation, the transitional rheology takes place between 50 to 25 % melt. In the transitional domain, the rheology of the two-phase material presents bulk shear softening (reduction of viscosity with shear rate) during melting and shear hardening (increase of viscosity with shear rate) during crystallisation.

Similarities include strain partitioning and non-linearity in time (mostly due to temperature variation) and space (due to variability of chemical composition, including fluids). Non-linearity is responsible for feedback loops. They are positive during melting and negative during crystallisation. In both cases, the loops modify melt extraction. As a consequence, strain-partitioning-driven, horizontal melt segregation due to non-coaxial deformation overcomes buoyancy- and matrix compaction-driven melt segregation. Conversely, during crystallisation, melt segregates only into local dilatant sinks and the negative feedback loops damp large-scale melt segregation.

Stick-slip-like melt motion in PMR results from the coexistence of melt and solid phase components with contrasted rheologies. The motion is discontinuous in space and time, leading to numerous leucosome generations in migmatites. In a crystallising magma, the negative feedback loops counteract this effect and restrict the appearance of proto-faults to the early-crystallised rims of plutons. The cycle of melting/crystallisation shows hysteresis in transport properties, amongst which viscosity is a specific one.

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## FIGURE CAPTIONS

Fig. 1. Thresholds (hatched) and behaviour of PMR during the transition from solid to liquid (bottom) and from liquid to solid (top). The range of RCMP (Arzi, 1978) is given for comparison. We are particularly concerned with the shaded intervals between thresholds.

Fig. 2. Melt fraction as a function of temperature for experimental melting of crustal derived sources. Metapelite from Vielzeuf & Holloway (1988), biotite-muscovite gneiss and biotite-gneiss from Gardien *et al.* (1995), plagioclase-poor metapelite from Patiño-Douce & Johnston (1991) and tonalite from Rutter & Wyllie (1988). LPT and MET thresholds as in Fig. 1. Steps in the non-linear melting curves indicate a new mineral participating in melting.

Fig. 3. Anorthite-Albite-Orthoclase-Quartz (An-Ab-Or-Qz) system with its cotectic surfaces (shaded). Three sources with close compositions (a, b and c) follow different cotectic slopes. As a result, their final compositions (arrows) are different.

Fig. 4. Melt fraction with respect to temperature for experimental crystallisation of different magma compositions. Basalt from Wright & Okamura (1977), leucogranite and wet granodiorite computed from Bouchez *et al.* (1992). RPT and PLT as in figure 2. Breaks in the non-linear crystallisation curves indicate that a new mineral phase enters or leaves the system.

Fig. 5. Differential stress-temperature log diagram computed from the power law rheology for granite ( $\gamma$ ), amphibolite (a) and granulite ( $\gamma_V$ ) (source values in Table 1). For each rock, differential stress is computed for fast ( $10^{-6}$ ), moderate ( $10^{-10}$ ) and slow ( $10^{-16} \text{ s}^{-1}$ ) strain rates. Scale-bar on the right refers to the corresponding curves and respective rock abbreviations. Temperatures between 700 and 1000°C span the range of melting temperatures of crustal rocks.

Fig. 6. Stress-strain rate diagram for PMR. Stress values were extracted from figure 5 for the 700-900°C temperature range. Viscosity is computed in a Newtonian mode as the ratio of stress to strain rate. Restricting temperatures to 750-850°C and differential stress to 5-50 MPa yields the viscosity of the matrix (left hatched box,  $10^{14}$  to  $10^{20}$  Pa.s). Melt viscosity (right shaded box) is between  $10^6$  to  $10^8$  Pa.s, corresponding to strain rate faster than  $10^{-1} \text{ s}^{-1}$  at 5-50 MPa stresses. At strain rate faster than  $10^{-4.5} \text{ s}^{-1}$  (dashed vertical line and hatched domain) shear softening results in a break in the viscosity curves (Webb & Dingwell, 1990). The contrasted viscosity between matrix and melt, and their non-linear difference, rules out mere averaging of matrix and melt viscosities.

Fig. 7. Interacting non-linear processes in migmatite during partial melting. (top left) non linear melting produces more melt with temperature and time during melting; (top right) viscosity decreases with strain rate and shear-softening takes place in the melt; (middle) distribution of molten material with depth is highly anisotropic. All processes combine in positive feedback loops (circled + between two arrows) that accelerate melt segregation. The curves are for a granitic magma composition, as figured in Fig. 2.

Fig. 8. Erratic, stick-slip-like evolution of a numerical model of melt extraction starting with 25 % melt content (Vigneresse & Burg, 2000). ( $\mu$ ) is the ratio of extracted melt percentage versus the initial melt percentage available in the matrix. Various theoretical MET values (in grey) induce transient variations of  $\mu$  followed by a return to a smooth extraction rate.

Fig. 9. Interacting non-linear processes during magma crystallisation. (top left) non-linear crystallisation results in less residual melt fraction with time; (top right) viscosity increases because of shear-hardening due to particle concentration; (middle) melt concentrates towards the core of plutons because of border effects. All processes combine in negative feedback loops (circled - between two arrows) that choke and ultimately stop melt circulation. The curves are for a granitic magma composition, as figured in Fig. 4.

Fig. 10. Viscosity curves computed for a pelite (dashed) and a tonalite (solid line). Melt content computed from figures 2 and 4. Rapid variations of the viscosity reflect the threshold values that depend on the melt fraction. The rheologically significant thresholds and associated drastic changes in viscosity occur at different temperatures for melting and crystallisation. The shift between the melting and crystallisation curves is similar to hysteresis in memory dependent materials.

Fig. 11. Schematic model of PMR of length  $L$  in which melt circulates with a velocity  $v$  along small conduits separated by a distance  $d$ . To be in chemical equilibrium with its matrix, diffusion  $D_s$  of an element  $s$  of the melt must be faster than the time for melt to segregate.

Fig. 12. New intrusion in an already settled magma. a) the new intrusion takes place when the former magma is not yet totally crystallised. Chemical re-homogenisation of the residual liquid occurs. b) the new magma intrudes a yet partly crystallised magma. Flattening occurs at the contacts, segregating incompatible elements from the residual liquid as reported in the granitic massif of Cabeza de Araya (Vigneresse & Bouchez, 1997). c) in case of a fast rate intrusion, brittle fracturing may take place within the non yet crystallised magma. No chemical exchange occurs.

layer	$\log(A)$ $\text{MPa}^{-n}\text{s}^{-1}$	$Q_c$ $\text{kJ mol}^{-1}$	$n$
granodiorite	1.5	212	2.4
amphibolite	-9.11	244	3.7
felsic granulite	-4.82	243	3.1

Table 1. List of the parameters used in model of crustal source material for migmatites (from Kirby & Kronenberg, 1987; Wilks & Carter, 1990) .