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# **Phase equilibrium constraints on the viscosity of silicic magmas with implications for mafic-silicic mixing processes**

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**Abstract.** Isobaric crystallization paths obtained from phase equilibrium experiments show that, whereas in rhyolitic compositions melt fraction trends are distinctly eutectic-like, dacitic and more mafic compositions have their crystallinities linearly correlated with temperature. As a consequence, the viscosities of the latter continuously increase on cooling, whereas for the former they remain constant or even decrease during 80% of the crystallisation interval, which opens new perspectives for the fluid dynamical modeling of felsic magma chambers. Given the typical dike widths observed for basaltic magmas, results of analogue modelling predict that injection of mafic magmas into crystallizing intermediate to silicic plutons under pre-eruption conditions cannot yield homogeneous composition. Homogenization can occur, however, if injection takes place in the early stages of magmatic evolution (i.e. at near liquidus conditions) but only in magmas of dacitic or more mafic composition. More generally, the potential for efficient mixing between silicic and mafic magmas sharing large interfaces at upper crustal levels is greater for dry basalts than for wet ones. At the other extreme, small mafic enclaves found in many granitoids behave essentially as rigid objects during a substantial part of the crystallization interval of the host magmas which implies that finite strain analyses carried out on such markers can give only a minimum estimate of the total amount of strain experienced by the host pluton. Mafic enclaves carried by granitic magmas behave as passive markers only at near solidus conditions, typically when the host granitic magma shows

near-solid behavior. Thus they cannot be used as fossil indicators of direction of magmatic flow.

key words: granite/rhyolite/dacite/andesite/basalts/mixing/viscosity/phase equilibria

Viscosity is one fundamental parameter that controls the physical and chemical behavior of magmatic rocks. Much effort has been invested in the past 30 years to evaluate the magnitude of this parameter in terrestrial magmas as well as its major controlling factors. Since the pioneering study of Bottinga and Weill (1972), soon modified by Shaw (1972) to incorporate the effect of water, there have been several works aimed at constraining the viscosity of various silicate liquids under a wide range of P-T-melt H<sub>2</sub>O content conditions (see Lange, 1994). However, it has been mainly in the last five years that interest in the viscous behavior of magmas, particularly of silicic compositions, has been renewed. This stems partly from the recognised important role that viscosity plays on eruptive phenomena. It is also due to recent technological improvements made both at low and high pressures that have opened new perspectives in the field of laboratory measurement of viscous properties of silicate liquids. In particular, the development of a high performance furnace operated in internally heated vessels (Roux *et al.*, 1996) has allowed viscosity determinations using the falling sphere method with greater accuracy than was possible in previous studies (e.g., Scaillet *et al.*, 1996; Schulze *et al.*, 1996; Holtz *et al.*, 1999). Similarly, the application of the parallel plate viscosimeter tool to hydrous glass cylinders synthesised at high pressures has allowed viscosity determinations in the high viscosity range (e.g., Richet *et al.*, 1996; Dingwell *et al.*, 1996), i.e., in a range not accessible by the more conventional, but time consuming, falling sphere method. Combination of both approaches enables the characterisation of the viscous behavior of supraliquidus liquids over very large temperature ranges,

which in turn allows derivation of thermodynamic properties, at least for some, well behaved, silicate liquids (e.g., diopside; Richet, 1984). These technological improvements have also benefited from recent analytical developments that enable full chemical characterisation of the run products before and after the viscosity determination. In particular, the melt H<sub>2</sub>O content can be now routinely analysed accurately, using for instance the Karl Fischer titration technique, and with a high spatial resolution through micro-infra red spectroscopy analysis (FTIR).

From a more geological perspective, it is well established that the viscosity of magmas depends mainly on temperature, melt water content, melt composition and melt fraction. All of these parameters can vary widely and there is not as yet empirical or thermodynamic model able to predict the controlling factors of the viscosity for all classes of natural magmas. Direct measurement of viscous properties of natural magmas (e.g., Pinkerton & Norton, 1995) are difficult, hazardous and intrinsically restricted to extruded magmas. For plutonic magmas, a laboratory approach is required and we must rely on the knowledge gained from a restricted number of case studies and, provided that these are representative, attempt to find the general rules or behavior of the rheological properties of magmas. As stated above, laboratory investigations have mainly focused on silicate liquids and only a limited number of studies has been devoted to the characterisation of real magmas, i.e. of crystal+liquid+fluid mixtures (e.g., Van der Molen & Paterson, 1979; Arzi, 1978; Rutter & Newman, 1995, Rushmer, 1995). These studies, however, were mainly conducted in the low melt fraction range, with the aim

of understanding the rheological behavior of the partially molten middle to lower crust. Those studies carried out at high melt fractions were aimed at characterizing volcanological, or shallow, regimes (Stevenson *et al.*, 1996). In contrast, the rheological behavior of magmas during crystallisation at depth remains a field largely unexplored. Although the first aim of experimental phase equilibria is not to constrain the rheology of magmas, this approach is one of the most powerful that can be used to evaluate all the critical factors mentioned above. It can be used on both volcanic and plutonic rocks, and for the latter it is the only one that allows the retrieval of magmatic evolution conditions. Its main drawback, however, is that this is a static or near-equilibrium approach by definition, and as such it does not incorporate any potential effect of deformation upon crystallisation. Nevertheless, combining the information gained from phase equilibrium studies on natural rocks together with empirical rheological laws obtained from the study of model or analogue systems is a useful first step for understanding the rheological properties of magmas under relevant P-T-melt H<sub>2</sub>O conditions. In this paper we take this approach, with particular emphasis on the relative rheological behavior of mafic versus silicic magmas. We build upon previous work that has mainly focused on silicic magma rheology, also using a phase equilibrium approach (Scaillet *et al.*, 1997; 1998a). We start this report by first reviewing our current knowledge about the viscosity of hydrous silicate melts as well as the effects of both bubbles and crystals on bulk viscosity. We then use recent phase equilibrium studies carried out on silicic to intermediate magmas (granite/rhyolite to andesite) as proxies for constraining the viscosity

patterns of silicic to intermediate cooling magmas. These patterns are then compared against those of basaltic magma crystallizing either under dry or wet conditions to explore the most favourable conditions leading to thorough mafic-silicic magma mixing. Finally, we discuss the rheological behavior of mafic enclaves found in granitoids in which they are used as strain gauges of the finite strain that affected the host plutonic body.

### **1. Viscosity models and their application to magmas**

From the rheological standpoint any magma can be usefully considered as a three phase system made of a mixture of melt, bubbles of exsolved volatiles and crystals in varying proportions. The rheological behavior of such a polyphase system will result from the interplay of the dynamical response of each of these phases with respect to any applied stress.

Decisive progress in the determination of the viscosity of hydrous silicate melts has been made in recent years and several empirical equations are now available (see Whittington *et al.*, 1999a). These may be valid over a rather large compositional range (e.g., Hess & Dingwell, 1996; Baker, 1998; Holtz *et al.*, 1999) or apply to a more specific (yet widespread) group of granitic magmas (Scaillet *et al.*, 1996 and Whittington *et al.* 1999a, for peraluminous granites). These empirical formulations supersede the classical model of Shaw (1972), especially in the low melt water content range. These new data have shown that hydrous silicic melts are distinctly non-Arrhenian over large temperature ranges (i.e. > 1000°C) although notable exceptions do exist (ie hydrous peraluminous silicic magmas, see Whittington *et al.* 1999a).

However, silicic magmas *sensu lato* display fairly smaller melting/crystallization temperature intervals, typically 100-200°C, over which the temperature dependence of the melt viscosity can be considered as Arrhenian. The melt viscosity is also strongly dependent on the strain rate: at low strain rates, silicic melts display a newtonian behavior, while at high strain rates, non-newtonian behavior is observed, and shear viscosities decrease with increasing strain rates (Webb & Dingwell, 1990). The low strain rates prevailing in plutonic environments imply that the melt in magmas at depths obeys predominantly a Newtonian behavior. However, magmas transported through dykes, be it from the source region or from the plumbing system of volcanoes, may well experience high strain rates at which the onset of non-newtonian behavior (Webb & Dingwell, 1990) is observed. Applications of the above empirical models to such cases will likely give a maximum value for the melt viscosity. The restored fluid content of most volcanic rocks falls in the range 1-5 wt.% (Wallace *et al.*, 1995; Scaillet *et al.*, 1998b). Available experimental data indicate that the role of bubbles in magma rheology is critically dependent on the strain rate (Stein & Spera, 1993; Bagdassarov & Dingwell, 1992). Assuming that volcanic rocks are the extruded equivalents of plutonic magmas, the low fluid content of the former suggests that the rheological role of bubbles in magmas at depth is probably minor (Scaillet *et al.*, 1998a). Only during eruption, or emplacement at very shallow levels, is the bubble fraction of a magma an important rheological factor (see Dingwell *et al.*, 1996).

The influence of crystals on magma viscosity is still a subject of active research. Experiments carried out on model systems have shown that the Eintein-Roscoe equation satisfactorily reproduces the data for crystal contents < 30% in volume (Lejeune & Richet, 1995). Beyond a crystallinity of 30%, magma viscosities are generally observed to increase rapidly up to reaching solid behavior at around 60% crystals. Phase equilibrium studies carried out on silicic magmas have conclusively shown that they arrive in a nearly molten state at their emplacement level, their load of crystals barely exceeding 10% (e.g., Clemens & Wall, 1981; Scaillet *et al.*, 1995; Dall'Agnol *et al.*, 1999; Scaillet & Macdonald, 2000). This indicates that during extraction and ascent, the rheology of silicic magmas is mostly controlled by the melt properties. A recent compilation of silicic magma viscosities has shown that melt viscosities of felsic magmas tightly cluster at  $10^{4.5}$  Pa s, irrespective of the temperature, melt water contents and the plutonic or volcanic nature of the rock (Scaillet *et al.*, 1998a). The main reason for such a constant viscosity is that temperature and melt H<sub>2</sub>O content are interrelated parameters, inversely correlated to each other (Fig. 1). Peralkaline, peraluminous and metaluminous magmas plot on the same general trend which indicates that the eruptability of silicic to intermediate magmas is not compositionally controlled. Similarly, the lack of difference between extrusive and intrusive rocks shows that the fate of silicic magmas, whether volcanic or plutonic, is not due to fundamental differences in viscous properties of the magma. In contrast, host rock rheological properties may exert a critical role on magma stopping upon ascent (e.g, Hogan & Gilbert,

1995). Nevertheless, the fact that there are virtually no volcanic rocks with crystal contents higher than 50% (Marsh, 1981; Scaillet *et al.*, 1998a), corresponding to magma viscosities of  $10^6$  Pa s (Fig. 1), indicates that this is a fundamental rheological barrier, beyond which intermediate to silicic magmas hardly move upward.

In the following, unless otherwise stated, the Shaw (1972) model has been used to calculate melt viscosities since it is still the only one available that can account for melt compositions more mafic than rhyolites (i.e., dacite, andesite and basalt), unlike all recently published models. In addition, despite the fact that several studies have shown that the Shaw (1972) model may yield inaccurate results for some T-H<sub>2</sub>O conditions (e.g., Dingwell *et al.*, 1996), recent experimental data from hydrous acidic, intermediate and basic compositions suggests that this empirical model is quite accurate at high water contents and magmatic temperatures, although it is unreliable at viscosities higher than its intended upper limit of  $10^6$  Pa.s. For example, the viscosity of Crater Lake andesite containing 4 wt.% H<sub>2</sub>O at 1350°C was measured as between 4 and 5 Pa.s by Kushiro (1976), and is reproduced exactly by the Shaw model. Viscosities of synthetic phonolite and trachyte melts containing 5 wt.% H<sub>2</sub>O at 777 and 877 °C, extrapolated from lower temperature measurements (Whittington *et al.*, 1999b) are reproduced to within 0.2 log units by the Shaw model. Similarly, the viscosity measurements performed on a hydrous peraluminous leucogranite melt by Scaillet *et al.* (1996) are reproduced to within 0.2 log units by the Shaw

model, for melt H<sub>2</sub>O contents between 3 and 6.6 wt.% and temperatures between 800 and 1100°C.

For the sake of simplicity, the Einstein-Roscoe equation has been used to calculate magma viscosities over the full crystallisation range, although this equation is strictly valid only at melt fractions in excess of 70 % (e.g. Lejeune & Richet, 1995). However, use of more appropriate equations (see Dingwell *et al.*, 1993) introduces only minor changes in the calculated viscosity, that differ by less than an order of magnitude from those obtained with the Einstein-Roscoe approximation. Given that mafic and felsic magma viscosities may differ by several orders of magnitude, as illustrated below, such relatively minor deviations are unlikely to significantly alter any conclusions derived here.

## **2. Magma viscosity during crystallization**

As phase equilibrium studies have shown, nearly all the crystallization of granitic magmas of crustal origin occurs isobarically, at the level of emplacement (see Wall *et al.*, 1987; Clemens & Mawer, 1992; Scaillet *et al.*, 1998a). Thus, phase equilibrium experiments are particularly well suited for constraining the rheological patterns of cooling silicic magmas.

Crystallization paths can be obtained firstly by recognizing that water has a nearly perfect incompatible behavior and secondly by assuming a closed system (e.g., Scaillet *et al.*, 1997). In all cases studied (see Scaillet *et al.*, 1997), melt fraction trends, illustrated here for two compositions, a peraluminous and a metaluminous one, display a near-eutectic behavior, the

magmas being still 60-80 wt.% liquid at 15 °C above their solidus (Fig. 2, see also Whitney (1988)). The melt and magma viscosity trends corresponding to these melt fraction trends are shown on figure 2. After a short increase, the melt viscosity decreases as crystallization proceeds. In the case of a magma beginning to crystallize with 7 wt.% H<sub>2</sub>O, at 750 °C and 400 MPa (peraluminous rhyolite), the viscosity decreases by nearly 1 order of magnitude during crystallisation (note that the viscosity of this specific case is calculated using the model of Scaillet *et al.* (1996), calibrated on a peraluminous silicic melt, and not with the Shaw model: the decrease observed therefore not due to an inadequacy of the empirical model). In both cases, magma viscosities either remain constant or increase during cooling, but by less than a factor of 10 over more than 80 % of the crystallization interval. The reason for such behavior is simply the lowering effect of water upon viscosity that largely offsets, or at least counteracts, the opposite effect arising from cooling and increasing crystal content during crystallization. This decrease in viscosity can be even more dramatic if, in addition to water, F and B are present in the system. For instance, the addition of 1.5 wt.% F to a peraluminous liquid with 6 wt.% H<sub>2</sub>O decreases the melt viscosity by a factor of 4 (Baker & Vaillancourt, 1995). This rheological pattern is likely to be common to most rhyolites/granites in so far as they have, by definition, a composition close to that of the eutectic and they are invariably water-rich (e.g., Whitney, 1988; Scaillet *et al.*, 1998a). These findings sharply contrast with the widely held belief that viscosity increases during the crystallization of rhyolitic magmas, and raise new questions about the fluid dynamical

properties of felsic magma chambers. Although physical models have not yet considered the possibility of a viscosity decrease during crystallization (e.g., Spera *et al.*, 1982; Brandeis & Jaupart, 1986; Marsh, 1989) the low and constant viscosity of felsic magmas during most of their crystallization history opens the possibility that most of the fractionation mechanisms at work in basaltic magma chambers, such as crystal settling, compaction and thermal or solutal convections, can be effective in felsic magma chambers as well (see Petford, 1993). For instance the compaction length (McKenzie, 1984) of magma with a melt fraction of 40 % and an interstitial melt viscosity of  $10^{3.3}$  Pa.s (corresponding in the case shown above to conditions 10 °C above the solidus for a granitic magma starting crystallizing with 7 wt.% H<sub>2</sub>O; Fig. 2) would exceed several hundred meters whilst the compaction time (McKenzie, 1984), around  $10^3$  years, is well below the lifespan of felsic magma chambers several kilometers across (Spera, 1980). Current physical models that postulate high and increasing viscosities for cooling silicic magmas may not, therefore, faithfully reproduce the fluid dynamics of natural felsic magma chambers.

Similar crystallization paths have been calculated for a dacitic bulk composition (SiO<sub>2</sub> = 65 wt.%), using the experimental results obtained by Scaillet and Evans (1999) on the dacite of Mt Pinatubo, for which variations of melt fraction with T and melt H<sub>2</sub>O contents, as well as melt composition, are all well characterised from phase equilibrium experiments performed at 220 MPa. Two crystallization paths have been calculated, one for starting conditions set at 950°C with 4 wt.% H<sub>2</sub>O in melt, and the second for H<sub>2</sub>O-

saturated conditions, with melt H<sub>2</sub>O contents of 6.9 wt.% at all temperatures (Fig. 3). Although some curvature is apparent, in both cases melt fraction trends display a much more linear relationship with temperature than those obtained with bulk granitic/rhyolitic compositions (Fig. 2). For the water-saturated case, the liquid viscosity rises steadily with lowering temperatures, having increased by nearly 2 orders of magnitude at midway between the liquidus and solidus (Fig. 3). As a consequence, magma viscosity increases throughout crystallisation, by 4 orders of magnitude at 50% of the crystallisation interval. Starting with under-saturated conditions (4 wt.% H<sub>2</sub>O at 950°C) has little effect. The main difference is that, after a short increase, the melt viscosity increases slowly over 80 % of the crystallisation interval, but the magma viscosity rises rapidly, by 5 orders of magnitude at 70 % of cooling (Fig. 3). Crystallisation paths calculated for an andesite composition using the phase equilibrium experiments on the Mt Pelée andesite obtained by Martel *et al.* (1999) are shown in Fig. 4 for two different initial melt H<sub>2</sub>O contents. Although not constrained in the near solidus range, the high temperature part of the crystallisation interval of andesitic magmas shows a linear correlation between melt fraction and temperature. As for the dacite, both the melt and magma viscosities continuously increase during cooling, by 1 and 3 orders of magnitude, respectively, during the first 40 % of the crystallisation interval (Fig. 4).

Thus, at upper crustal levels, the rheological pattern of a dacitic/andesitic magmas does not resemble that of rhyolites, in that their viscosities increase during cooling, irrespective of the initial melt H<sub>2</sub>O

content, although the increase remains moderate during the first half of the crystallization interval. Increasing pressure will increase H<sub>2</sub>O solubility, and it cannot be excluded that, beyond a certain depth, the effect of water will dominate that of an increase in both melt silica content and crystals, as seen for rhyolitic compositions (Scaillet *et al.*, 1997). However, the viscosity reduction with increasing water content is less marked for more basic compositions than for rhyolitic compositions, at least at low temperatures near the glass transition (Whittington *et al.*, 1999b; 1999c). For the purposes of fluid dynamical modelling of shallow dacitic/andesitic magma chambers, the present data show that it is safe to assume that viscosity increases continuously during cooling. The experimental data available for more mafic compositions than silicic andesite does not allow us at present to carry out this type of modelling, but it is clear that, since basaltic magmas obey the same rule (i.e., linear decrease in crystal content with temperature), we can predict that such intermediate compositions will follow a rheological pattern also characterized by a continuous increase in viscosity during crystallization.

### **3. Implications for basalt-rhyolite mixing**

Field evidence clearly shows that basaltic and felsic magmas often come into contact in the crust (e.g., Eichelberger, 1975; Bacon, 1986; Barbarin & Didier, 1992; Wiebe, 1994). The most common situations are due either to basaltic magmas that pond at the base of the crust and subsequently melt lower continental lithologies (e.g., Huppert & Sparks, 1988; Atherton & Petford, 1993) or to the injection of basaltic magmas into upper silicic magma

chambers (e.g., Coulon *et al.*, 1984; Nixon, 1988; Michael, 1991; Pallister *et al.*, 1992; Feeley & Dungan, 1996). In both cases, the extent of mixing between the two end-members will depend, among other factors, on the viscosity contrast (e.g., Sparks & Marshall, 1986; Campbell & Turner, 1986; Frost & Mahood, 1987; Fernandez & Gasquet, 1994; Snyder & Tait, 1995; Jellinek & Kerr, 1999). Intermediate to silicic magmas have a characteristic magma viscosity of  $10^{4.5}$  Pa.s during emplacement or at pre-eruption conditions (Scaillet *et al.*, 1998a). Trends displayed on figure 2 show that this viscosity value is maintained during a substantial part of the magmatic evolution of rhyolitic magmas. We can use this value to evaluate the effects of basalt injection into crystallizing intermediate to silicic magma chambers. Analogue experiments carried out by Campbell and Turner (1986) suggest that the effectiveness of mixing in turbulent fountains at the base of magma chambers can be characterized by the Reynolds number:

$$Re = w d / \eta$$

(2)

where  $w$  is the velocity of the incoming fluid (basalt),  $d$  the width of the input pipe and  $\eta$  the viscosity of the ambient magma (rhyolite). When  $Re$  is  $>70$ , mixing is predicted to be very efficient, whereas for  $Re < 7$  no mixing occurs at all. Adopting the set of values used by Campbell and Turner (1986) for basaltic magma but taking a viscosity of  $10^{4.5}$  Pa.s obtained for silicic rocks (instead of  $10^7$  Pa.s), shows that this mixing criterion is met once the dike width exceeds 10 m ( $Re = 10$ ). This is at the high end of the range measured for basaltic magmas, whose average dike width is close to 1 m (e.g, Wada,

1994). Taking the average magma viscosity obtained for the pre-eruption conditions of volcanic rocks increases the minimum dike width for mixing up to 100 m. High intrusion rates may induce high strain rates at the interface between the two magmas, conditions that may promote the onset of non-Newtonian behavior in silicate melts (Webb & Dingwell, 1990; Dingwell & Webb, 1990) which is characterized by a decrease of melt viscosity as the strain rate increases (Webb & Dingwell, 1990). This decrease may be shared by both magmas, and the experiments do not indicate that acid compositions depart more significantly than mafic ones from non-newtonian behavior in such a way that the viscosity of felsic magma would be decreased more rapidly than the mafic one. Thus, the intrusion of a basalt magma into an upper solidifying silicic magma chamber will hardly result in a compositionally homogeneous intermediate magma. However, injection of basaltic magmas at early stages of the crystallization of a dacitic or andesitic magma may lead to mixing, since the near-liquidus viscosities of the latter are lower by 1-2 log units relative to those at pre-eruptive conditions (Fig. 1), decreasing the critical dike width by a factor of 10 to 100.

The previous analysis considers the case of mixing arising from vigorous injection of basaltic magma into crystallizing silicic ones, where the energy for mixing comes in part from the rate of intrusion of the incoming magma. There are many other instances where basalt and rhyolite are brought together in a more quiescent way, sharing large physical interfaces. Given that heat diffusion is many orders of magnitude faster than chemical diffusion in silicate melts (e.g., Watson, 1981), the first effect of basalt juxtaposition is

to force the silicic magma back up its liquid line of descent, without significant contamination (e.g., Sparks *et al.*, 1977; Sparks & Marshall, 1986). In this case the viscosity contrast across the interface will be one critical factor in determining whether or not wholesale homogenization can occur (Sparks & Marshall, 1986).

Following Sparks and Marshall (1986), Frost and Mahood (1987) extensively modelled such a situation although, owing to the lack of pertinent experimental data, they approximated the crystallinity as an error function of  $T$ . The rheological patterns derived from recent experimental studies and summarized above allow a critical evaluation of such an approximation. Marsh (1981) has shown that it holds well for most mafic compositions though in some cases a linear fit to the data would also give satisfactory results (e.g., the basalt of Makaopuhi, Marsh (1981)). The experimental data show that this approximation is correct for compositions as felsic as dacites, though in detail small differences can be noted (Fig. 3, 4). The modelling of Frost and Mahood (1987) of the interactions between mafic and silicic magmas, the latter having 60 wt.%  $\text{SiO}_2$ , probably closely reproduces the essential physics of the process. For more silicic rocks, however, this error function badly fails in reproducing the overall melt fraction trend (Fig. 2), irrespective of the constant used (i.e. the standard deviation parameter) for modifying the general shape. The error function greatly under-estimates the amount of crystallization that takes place from mid-way in the crystallisation interval down to sub-solidus conditions. For instance, after 80 % of cooling, the error function predicts a melt fraction of 10 % vol, whereas the

experimental melt fractions vary from 60% for a hornblende-biotite rhyolite crystallizing at 300 MPa to nearly 80 % for a peraluminous leucogranite at 400 MPa (Fig. 2). This implies that for cases of injection at subliquidus conditions of the silicic end-member, the amount of mafic magma needed to prevent its being quenched by the silicic one is less than that calculated by Frost and Mahood (1987) for a granitic composition, because there is less heat taken up to melt pre-existing crystals in the silicic magma.

Schematically, the masses obtained by Frost and Mahood (1987) for interacting mafic and granitic magmas should be scaled down by a factor of 5. Although the interactions between both end-members can be modelled, as done by Sparks and Marshall (1986) and Frost and Mahood (1987), there is no particular set of initial conditions (especially the crystallinity of both end members at the time of intrusion, and their relative masses) that is of wide applicability. An evaluation of the potential for mixing between mafic and silicic magmas can, however, be done simply by considering the evolution of the viscosity contrast during crystallization. In doing so we take advantage of the fact that rhyolitic magma viscosities remain constant over a large part of the crystallization interval, which implies that, unless intrusion occurs at near-solidus conditions, variations in the viscosity contrast arise predominantly from the viscosity increase of the mafic magma.

Two limiting cases can be envisaged. One is the mixing of a silicic magma with a dry basaltic one whose rheology during crystallization is well known (e.g. Shaw, 1969). The evolution of magma viscosity ratio during the crystallization of dry basalt and a granite that started to crystallize with 7

wt.% H<sub>2</sub>O shows that, after 40 % of cooling, both magmas display similar viscosities (Fig. 5). Upon further cooling, the basalt viscosity becomes many orders of magnitude higher than that of the rhyolite because the effects of temperature drop and of increase in crystal content cannot be counteracted by that of water. Also, the crystallinity of a basaltic magma increases in a nearly linear fashion as temperature decreases (at least down to ~ 60 vol. % of crystallinity) which implies that the crystal effect on the magma viscosity happens earlier during the cooling of basaltic magmas than of rhyolitic magmas. The net result is that mixing between silicic and dry mafic magmas should be much more effective after substantial crystallization has occurred in the mafic end-member. As shown by previous studies (Frost & Mahood, 1987), the cross-over in viscosity varies with the initial water content of granitic magmas. Decreasing initial H<sub>2</sub>O content down to 4.5 wt.% delays the attainment of the viscous neutral point to 50 % of cooling (Fig. 5). Assuming that the basalt becomes effectively rigid when its crystallinity is 60 % (i.e. convection ceases, see Marsh (1981)), and given that at this point the granitic magma is largely molten (less than 30 % crystals), both magmas will be still able to convect at the viscosity cross-over. This result confirms previous findings obtained from thermomechanical modeling that, from the strict point of view of the viscosity contrast, the intrusion of a dry basalt into a granitic magma may ultimately produce a mixed/mingled rock of intermediate composition (i.e. andesitic or granodioritic), provided that the mass of the intruded basalt prevents its quenching.

The second case consists of mixing a wet basalt with a silicic magma. This case may find natural applications in active subduction zones where a growing body of evidence indicates that some arc basalts may contain a significant amount of dissolved water, up to 8 wt.% (e.g., Sisson & Grove, 1993a,b; Sisson & Layne, 1993; Sobolev & Chaussidon, 1995; Pichavant *et al.*, 2000). Obviously, the presence of water implies that such basalts will undergo a minor magma viscosity increase during cooling relative to their dry counterpart, and reach the point at which basaltic and granitic viscosities become equal later in their crystallization history. To illustrate this point, we use the experimental results that Sisson and Grove (1993b) obtained on their composition 87S35a, assuming 6 wt% H<sub>2</sub>O dissolved in melt at all temperatures, to calculate melt viscosities and use the modal amount of phases present to calculate magma viscosities. In the case of mixing with a normally wet-silicic magma (4.5 wt.% initial H<sub>2</sub>O), the calculation shows that the viscosities of the silicic magmas are always higher than that of the basalt for crystallinities lower than 60%. If the silicic magma contains an initial H<sub>2</sub>O content of 7 wt.%, both magmas will reach similar viscosities, but only very late in their crystallization interval, when the basaltic magma is very close to having a solid-like rheological behavior. The latter example, however, uses the rheological pattern of a granitic magma crystallizing at 400 MPa, reaching melt H<sub>2</sub>O contents in excess of 9 wt.% at saturation (Scaillet *et al.*, 1995), and thus comparatively low magma viscosities. Silicic magma chambers of arc-volcanoes lie most often in the 200-300 MPa pressure range (e.g. Rutherford *et al.*, 1985; Rutherford & Devine, 1996; Luhr, 1990; Martel

*et al.*, 1998; Scaillet & Evans, 1999), in which the water solubility is 7-8 wt.%. In addition, most of the erupted rocks record pre-eruptive melt H<sub>2</sub>O contents in the range of 3-5 wt.% (see Scaillet *et al.*, 1998a). All these facts suggest that the conditions appropriate for thorough mafic-silicic magma mixing in the case of H<sub>2</sub>O-rich basalts are probably never met under shallow crustal levels. The possibility of a causal link between basalt intrusion and explosive eruption of silicic magmas (Sparks *et al.*, 1977) also points to a rather limited time of coexistence between the magmas, on the order of only a few days as petrological evidence suggests (e.g., Nixon, 1988; Pallister *et al.*, 1992). Thus, mixing between wet basalts and rhyolitic magmas under shallow crustal conditions has little chance of producing a homogeneous intermediate magma, as testified by the widespread occurrence of poorly mixed mafic-silicic end-members in arc magmas (e.g., Eichelberger, 1975; Nixon, 1988; Pallister *et al.*, 1992).

#### **4. Implications for the use of mafic enclaves in strain analyses**

The intrusion of small dikes of more mafic composition into a large reservoir of silicic magma may eventually, upon disruption, give rise to isolated mafic enclaves (e.g., Bacon, 1986). Such enclaves are found in many granitoids and are commonly used either for the quantification of finite strain (e.g., John & Blundy, 1993) or for evaluating strain rates (Williams & Tobisch, 1994). Williams and Tobisch (1994) used enclave shapes to calculate magmatic flow velocities, by considering mainly the case where the enclave viscosity is lower than that of the host magma, a situation that implies

overheated silicic magma (temperatures significantly higher than 1000°C). This arises when the mass of injected mafic magma is at least comparable to that of the intruded/underplated one (Sparks & Marshall, 1986; Frost & Mahood, 1987) such as in the case of basalt underplating of the lower continental crust (Huppert & Sparks, 1988). Here we concentrate on the case where the incorporation of small blobs of mafic magmas does not affect the thermal regime of the host magma, extending the analysis done by Williams and Tobisch (1994) to the low temperature range of the crystallisation interval of granitoid magmas. In such a case, beside transient effects related to the local superheating of the silicic magma adjacent to the mafic magma, the equilibration temperature of the resultant mixture will not differ greatly from the pre-intrusion temperature of the resident magma. The viscosity contrasts obtained under such circumstances have been calculated for the intrusion of wet basalts into andesitic, dacitic, and rhyolitic magmas, dacitic into granitic magmas, and andesitic into dacitic magmas, using the experimental results presented above to constrain the rheological patterns of each magma. The case for dry basalt is not shown since its solidus temperature, about 1000°C, is above the liquidus temperatures of most silicic to intermediate magmas which implies that small dikes of dry basalt injected will very rapidly behave as rigid objects with respect to any applied stress. The case of mixing between wet basalt and dacite is shown in figure 6 without normalizing both temperature and viscosity to help understand the basic features common to all cases dealt with here. Two viscosity cross-overs can be observed: the first is attained when the basalt crystallinity reaches 60

%, at around 915°C, which induces a dramatic rise in the enclave viscosity. The second cross-over is reached when the host silicic intermediate is 60 % crystallised, which occurs at about 750°C, at which point the viscosity rapidly rises and becomes equal to that of the basaltic enclave since both magmas basically behave as solids. Obviously, the temperature range between these two viscosity cross-overs is controlled by the liquidus-solidus temperature and the melt fraction-temperature trend specific to each magma, both factors depending on the phase relationships of the magma. The variation of these factors with the chemistry of the mixed magmas is illustrated in the following section.

For basaltic enclaves in either rhyolitic or dacitic magmas (Fig. 7), the rheological patterns are all similar, being characterized by small viscosity contrasts at near-liquidus conditions of the granitoids, followed at intermediate temperatures by a large domain, in some cases more than 50 % of the crystallization interval, during which mafic enclaves are many orders of magnitude more viscous than their surrounding matrix (Fig. 7). As the melt fraction of the host magma decreases, however, the viscosity contrast drops again and becomes nil as soon as the silicic magma crosses the transition from suspension-like to rigid-like behavior, thought to occur at around 30-40 % of melt by volume. In rhyolitic magmas such a transition will occur close to solidus conditions, in keeping with their near-eutectic behavior (around 680°C, Fig. 2), whereas in dacitic bulk compositions, this rheological change will occur at early stages of the crystallization interval (around 750°C at 220 MPa, Fig. 3). These calculations indicate that enclaves of mafic compositions

found in granitoids will have a viscosity several orders of magnitude higher than their host magmas during an extended part of the crystallization interval. For injection of dacitic dikes into crystallizing granitic plutons the same general pattern is observed (Fig. 8), with the viscosity of the two magmas differing by less than 1 order of magnitude during the first 30-40% of the crystallization interval, while the intermediate domain of high viscosity for the mafic enclaves is significantly reduced, as might be expected from the greater similarity in crystallization temperatures and compositions between resident and injected magmas. The case of andesite inclusions in a dacite magma is shown on figure 8, where it is apparent that the rheological behavior of enclave and host is mainly controlled by the H<sub>2</sub>O content of host magma. Andesite enclaves in an H<sub>2</sub>O-saturated dacite will display a large viscosity contrast during a longer part of the crystallisation interval of their host than when included in a H<sub>2</sub>O-undersaturated dacite magma. This is a direct consequence of the lower crystallisation temperature of the H<sub>2</sub>O-saturated dacite relative to the andesite. Although the H<sub>2</sub>O content of the inclusion exerts a minor influence of the viscosity contrast, figure 8 shows that, at near liquidus conditions, the viscosity contrast is smaller when the melt water content of the andesite is the same as in the dacite. Finally, injection of basalt in andesite (Fig. 7), is characterized by a substantial shrinking of the domain with large viscosity contrast. No fundamental difference in the rheological pattern compared to the previous examples is seen when the H<sub>2</sub>O content of the andesite varies, although the peak value in the viscosity contrast is the smallest of all cases considered here. Relative to

the dacite/rhyolite and andesite/dacite mixing cases, however, the intermediate domain with a high viscosity contrast is displaced toward near-liquidus conditions. The various examples displayed in Figs.7-8 invariably show that the potential for mixing between a mafic and a silicic magma is optimum at near liquidus conditions of the intruded silicic end member and lasts for about 50 % of the crystallisation interval in the most favourable case. Repeated injection of small amounts of mafic magma in a near-liquidus felsic magma chamber may thus progressively drive the composition of the ambient magma toward a more mafic bulk composition. In contrast injection of mafic dykes at later stages of the evolution of the felsic magma chamber are less likely to induce severe chemical changes or homogeneous mixtures.

Decreasing the compositional gap between the two end-members produces, on average, a substantial diminution of the domain where the viscosity contrast is very large as seen for instance for the dacite into rhyolite or basalt into andesite cases (Fig. 7, 8). Differences in the rheological contrast are not simply correlated, however, to a specific compositional variable such as the bulk  $\text{SiO}_2$ , as illustrated by the andesite/dacite situation compared to the basalt/andesite one. In the first case, the end member magmas differ only by 5 wt% in  $\text{SiO}_2$  while there is a 10 wt% difference in  $\text{SiO}_2$  between the silicic andesite and basalt. Yet, the viscosity difference during cooling is greater between andesite and dacite than between andesite and basalt (Fig. 7, 8). This difference reflects variable liquidus-solidus temperature ranges as well as variations of crystallinity with temperature of the magmas involved in the process. More specifically, the phase equilibrium data show that, at 200

MPa, whilst the hydrous liquidus of both andesitic and basaltic magmas differ only by 10°C, the difference between liquidus temperatures of dacite and andesite reaches 55°C. For magmas having their crystallinity linearly correlated with temperature (i.e. for any magma more mafic than a dacite), the amount of crystallisation for each decrement of temperature will vary with the crystallisation interval, being larger when the crystallisation interval is smaller. Thus, a dacite with an initial melt H<sub>2</sub>O content of 4 wt% will have a crystallisation interval larger by 120°C than when at H<sub>2</sub>O-saturation. The results show clearly that the melt water content of the host exerts a critical role on the viscosity contrast as illustrated here for the andesite into dacite situation (Fig. 8). Altogether, these examples illustrate that the rheological behavior is hardly predictable from the consideration of the anhydrous composition only. Water appears to be as important as the bulk anhydrous composition of the magmas involved in the mixing/mingling process. This emphasizes the importance of a correct determination of intensive variables for a realistic evaluation of the rheological properties of natural magmas.

Considering now the possibility of using mafic enclaves as strain gauges, the previous analysis shows that mafic enclaves may experience rather complicated magmatic strain histories, reflecting changes from less to more viscous behavior during the crystallization of their host magma. Overall, however, and in particular for rhyolitic magmas hosting enclaves of basaltic composition, enclaves are likely to behave more as rigid objects, relative to their host medium, than as passive markers, especially in the interval of viscous behavior of the granitoid. Experimental evidence suggests

that once the viscosity ratio exceeds 4 (that is 0.6 in log units), enclaves are virtually unaffected by the deformation (see for more details Williams and Tobisch (1994)) and thus cannot record it through changes of shape. This indicates that the use of mafic enclaves as finite strain markers will give only a minimum estimate of the total amount of strain that affected a plutonic intrusion. Williams and Tobisch (1994) concluded that mafic enclaves will mainly record magmatic strain affecting the host magma at near- or super-liquidus conditions, but inspection of figures 7 and 8 shows clearly that they can also record strain during the waning stages of granitoid crystallization. In fact, the state of passive marker for mafic enclaves is attained and maintained for sustained periods only very late in the crystallization history of granitoid magmas, basically when the latter reach solid behavior (Fig. 7, 8). Viscosity cross-overs occurring at early stages are necessarily transient and are unlikely to last for long periods, especially in rhyolitic systems where the latent heat release upon crystallization, and thus the time of cooling, is prominent at near-solidus conditions. Indeed, experimental melt fraction trends obtained in rhyolitic systems (Fig. 2) show that the major part of the crystallization occurs at sub-solidus conditions, which implies that fabrics preserved in granitic plutons have little chance of providing clues about emplacement mechanisms, or about any strain event occurring at early stages (Scaillet *et al.*, 1997). A similar conclusion can be applied to strain trajectories derived from the study of mafic enclaves when they parallel magmatic foliation patterns defined by mineral orientations. They cannot be interpreted solely as remnants of magmatic directions of flow nor in terms of effects related to

emplacement mechanisms, since their shape must reflect to a large extent deformation episodes that took place when their host matrix had a rigid behavior.

## **5. Conclusions**

We have attempted to place some constraints on the rheological properties of natural magmas using recent phase equilibrium data. The strength of the present approach is that full consideration is given to the P-T-melt H<sub>2</sub>O content conditions of evolution of the magma. These constraints are by no means definitive, however, for several reasons. First, the chemistries of magmas have been considered constant during interaction. This is clearly an over-simplification, since numerous field evidence show that chemical interaction takes place between intruded and host magmas (e.g., Debon, 1991; Barbarin & Didier, 1992), decreasing the compositional gap, and thus the rheological contrast, between coeval mafic and felsic magmas. Second, in all the cases handled here, the melt viscosity has been calculated using the Shaw model (Shaw, 1972). As stated in the introduction, this model has been shown to be in error under some conditions for silicic magmas and it cannot be excluded that its application to mafic compositions may also lead to erroneous results. Although considerable progress has been made in the field of viscous behavior of silicic magmas, there have been few recent attempts to investigate the viscosity of hydrous mafic melts, and no direct measurements at magmatic temperatures. We consider that such data are critical for a rigorous evaluation of mixing processes such as those illustrated above and

suggest that future laboratory studies take this need into account. Second, the magma viscosity is here calculated using the melt fraction obtained at equilibrium, yet a deforming magma is by definition out of equilibrium. Available data concerning the effect of deformation upon crystallisation are restricted to dry basaltic composition. These data demonstrate that deformation induces severe changes in the crystallisation sequence of mafic magmas (Kouchi *et al.*, 1986) and it cannot be precluded that the phase relations of felsic magmas are sensitive to deformation as well. The effect of deformation on crystallising silicate liquids is clearly one aspect that should also be rigorously explored in the near future if we wish to correctly characterise the rheological properties of natural magmas seen in a dynamic perspective.

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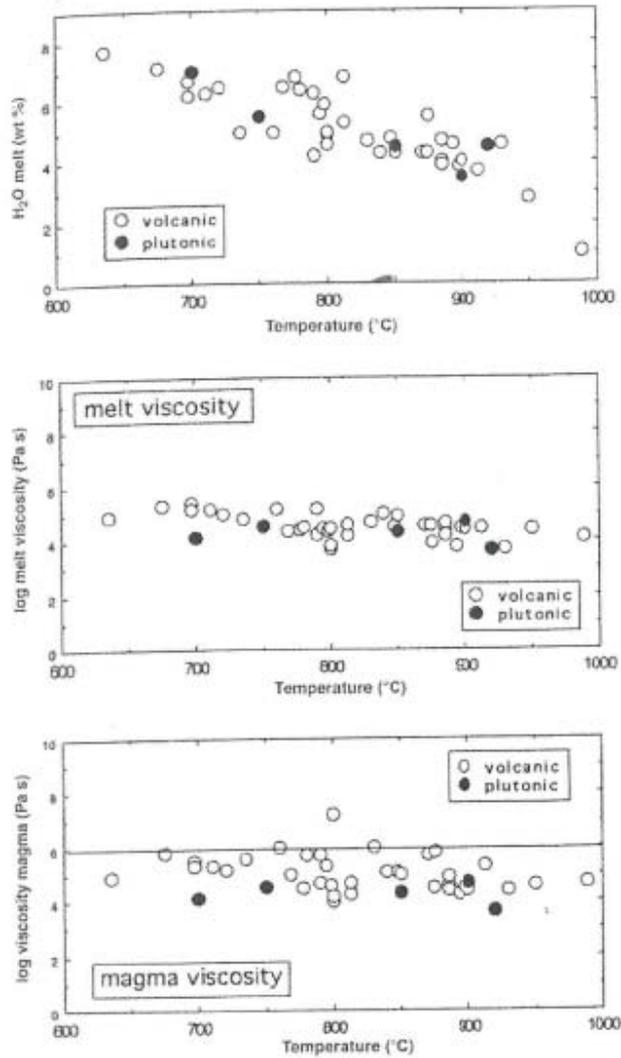
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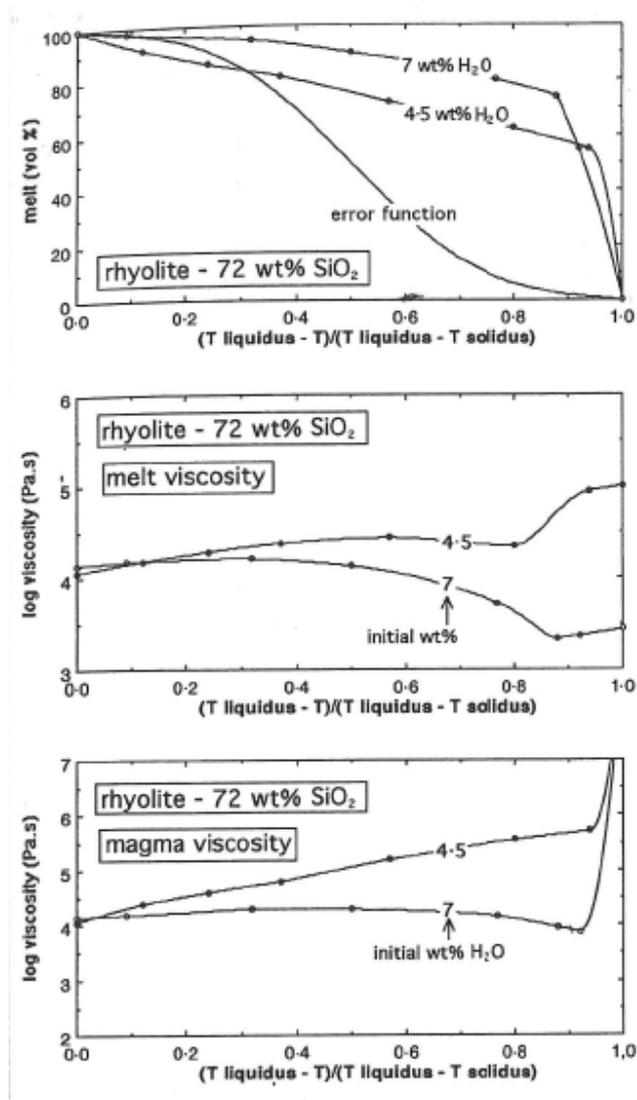
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## Legends of figures

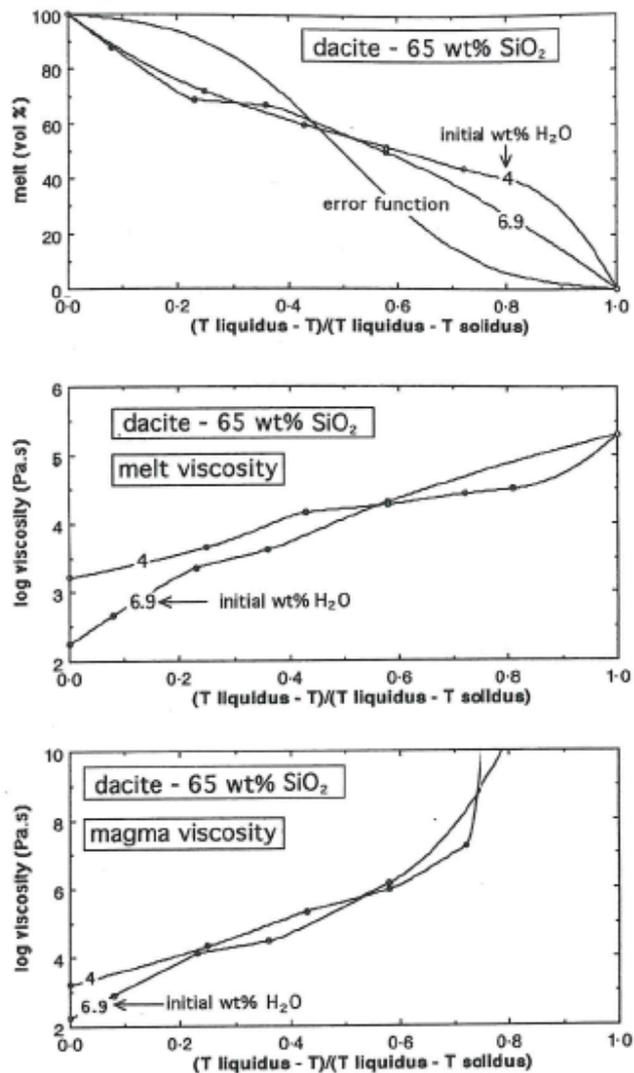


**Figure 1.** Temperature against the melt H<sub>2</sub>O content, melt and magma viscosities of silicic to intermediate magmas at pre-eruption conditions or during emplacement in the upper crust (modified from Scaillet *et al.*, 1998a)



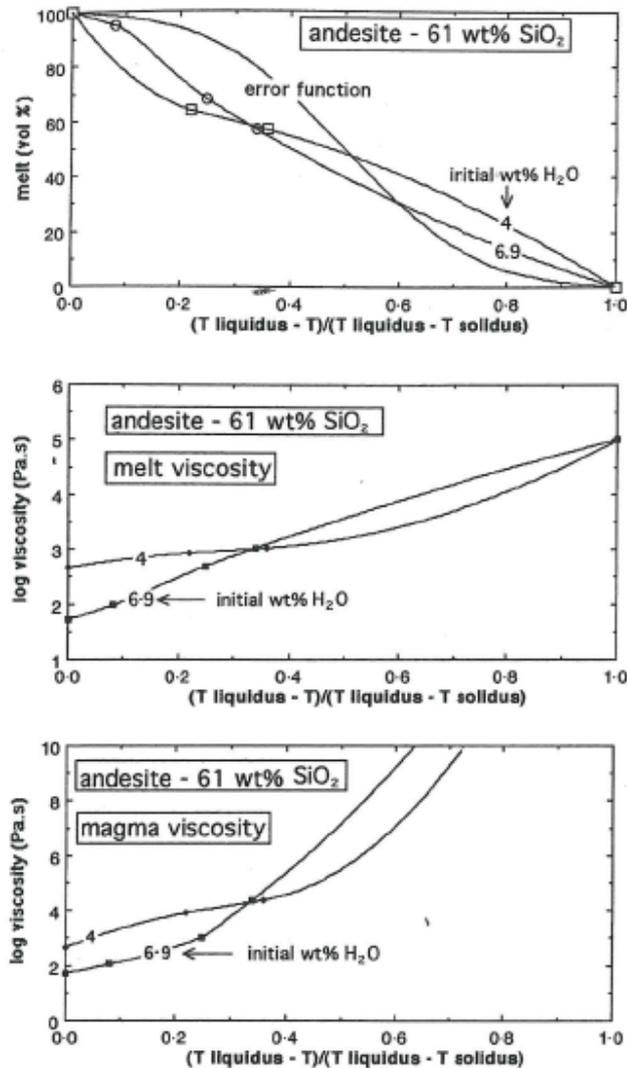
**Figure 2.** Experimental melt fractions, melt and magma viscosity trends for two rhyolitic magmas versus normalized temperature. Numbers on the trends correspond to the initial melt H<sub>2</sub>O content. Also shown is the melt fraction trend obtained with the error function using a standard deviation parameter of 1.5 (Marsh, 1981). Note the constancy or even the decrease of melt viscosity as cooling proceeds. Melt viscosities are calculated using the empirical

equation of Scaillet *et al.* (1996) for the melt with 7 wt% H<sub>2</sub>O and the model of Shaw (1972) for the melt with 4.5 wt% H<sub>2</sub>O. Magma viscosities are calculated using the Einstein-Roscoe equation (e.g., Lejeune & Richet, 1995). Liquidus and solidus temperatures are 920°C-665°C (4.5 wt.% H<sub>2</sub>O initial) and 750°C-630°C (7 wt.% H<sub>2</sub>O initial). See Scaillet *et al.* (1997) for additional details.



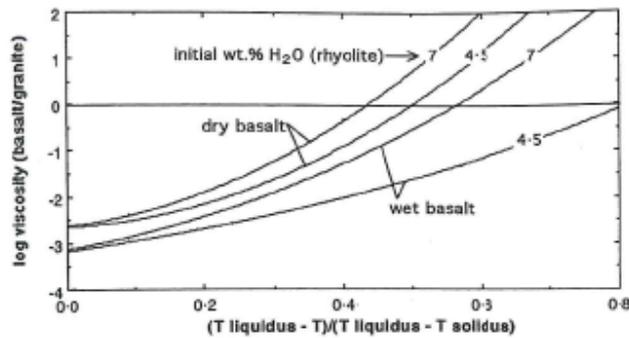
**Figure 3.** Experimental melt fractions, melt and magma viscosity trends for a dacitic magma versus normalized temperature, calculated using the experimental results of Scaillet and Evans (1999). Numbers on the trends correspond to the initial melt H<sub>2</sub>O content. Also shown is the melt fraction trend obtained with the error function using a standard deviation parameter of 1.5 (Marsh, 1981). Note the continuous increase of magma viscosities as

cooling proceeds. Liquidus and solidus temperatures are 1040°C-680°C (4 wt.% H<sub>2</sub>O initial) and 920°C-680°C (6.9 wt.% H<sub>2</sub>O initial). Melt viscosities are calculated using the model of Shaw (1972). Magma viscosities are calculated using the Einstein-Roscoe equation (e.g., Lejeune & Richet, 1995). Liquidus and solidus temperatures are 1040°C-680°C (4 wt.% H<sub>2</sub>O initial) and 920°C-680°C (6.9 wt.% H<sub>2</sub>O initial).

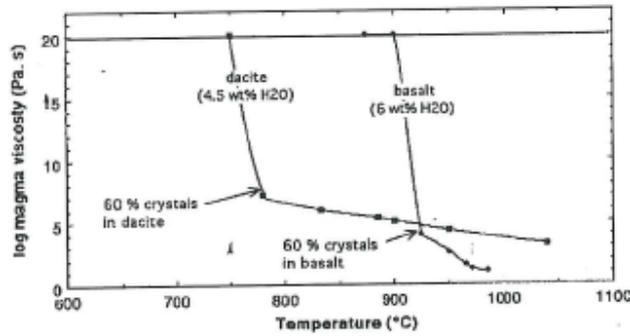


**Figure 4.** Experimental melt fractions, melt and magma viscosity trends for an andesitic magma versus normalized temperature, calculated using the experimental results of Martel *et al.* (1999). Numbers on the trends correspond to the initial melt H<sub>2</sub>O content. Also shown is the melt fraction trend obtained with the error function using a standard deviation parameter of 1.5 (Marsh, 1981). Note the continuous increase of magma viscosities as

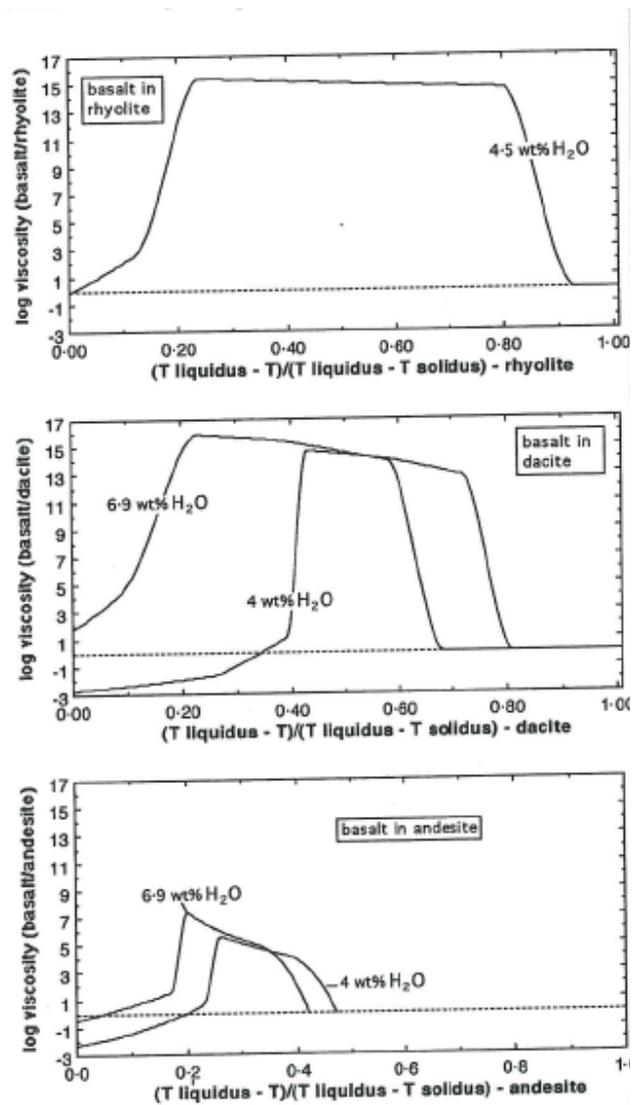
cooling proceeds. Liquidus and solidus temperatures are 1000°C-680°C (4 wt.% H<sub>2</sub>O initial) and 975°C-680°C (6.9 wt.% H<sub>2</sub>O initial). Melt viscosities are calculated using the model of Shaw (1972). Magma viscosities are calculated using the Einstein-Roscoe equation (e.g., Lejeune & Richet, 1995). Liquidus and solidus temperatures are 1000°C-680°C (4 wt.% H<sub>2</sub>O initial) and 975°C-680°C (6.9 wt.% H<sub>2</sub>O initial).



**Figure 5.** Comparison of the evolution of viscosities during the crystallization of basaltic and rhyolitic magmas. The viscosity of the dry basaltic magma is from Shaw (1969), whereas the viscosity of wet basalt has been calculated from the experimental results of Sisson and Grove (1993) using the model of Shaw (1972). Rhyolitic magma viscosities are from figure 2. Numbers reported on the lines correspond to the initial melt H<sub>2</sub>O content of the rhyolitic magma.

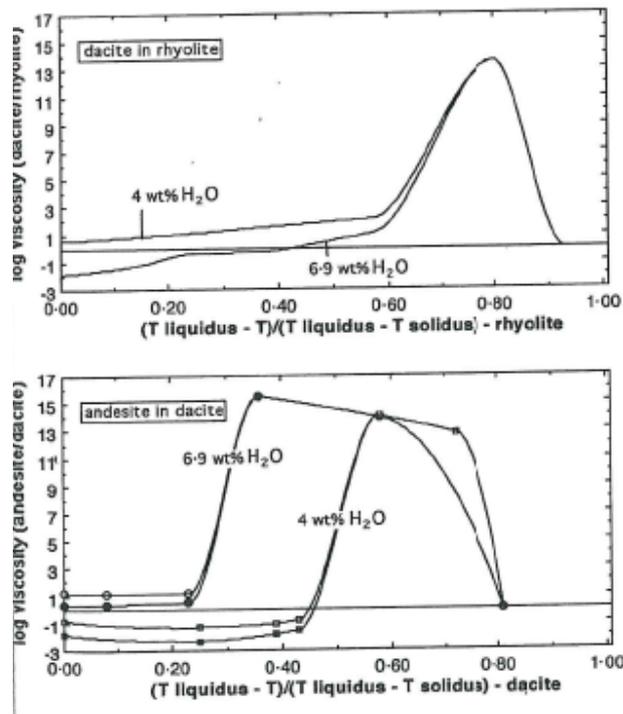


**Figure 6.** Comparison of the viscosity evolution of dacitic and wet basaltic magmas during cooling. Data for the dacite are from Scaillet and Evans (1999) and for the basalt from Sisson and Grove (1993). In both cases the solidus is fixed at 680°C. Melt viscosities are calculated using the model of Shaw (1972). Magma viscosities are calculated using the Einstein-Roscoe equation (e.g., Lejeune & Richet, 1995).



**Figure 7.** Evolution of the viscosity ratio of a thermally equilibrated basaltic enclave and a host rhyolitic, dacitic or andesitic magma. The temperature of the mafic enclave is that of the silicic host. Numbers reported on the lines correspond to the initial melt H<sub>2</sub>O content of the host magma. The viscosity in the solid behavior field (that is for melt fractions lower than 30 vol. %) is taken to 10<sup>20</sup> Pa.s. Liquidus and solidus temperatures of the silicic magmas

are as in figures 2-4.



**Figure 8.** Evolution of the viscosity ratio of thermally equilibrated dacitic enclaves and host rhyolitic magma and of andesitic enclaves and host dacitic magma. The temperature of the mafic enclave is that of the silicic host. Numbers reported on the lines correspond to the initial melt H<sub>2</sub>O content of the host magma. Open symbols are for 4 wt% H<sub>2</sub>O in the andesite enclave, filled ones for 6.9 wt% H<sub>2</sub>O. The viscosity in the solid behavior field (that is for melt fractions lower than 30 vol. %) is taken to 10<sup>20</sup> Pa.s. Liquidus and solidus temperatures of the felsic magmas are as in figures 2-4.