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# Phase equilibrium constraints on the viscosity of silicic magmas

## 1. Volcanic-plutonic comparison

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**Abstract.** By using recently determined experimental phase equilibria we show that the viscosity of granitic magmas emplaced at upper crustal levels is approximately constant at  $\sim 10^{4.5}$  Pa s, irrespective of their temperature and level of emplacement. Magmas crystallizing as granitic plutons are not water-poor and thus not more viscous than their extrusive equivalents. Instead, comparison between pre-eruption magma viscosities of extrusive silicic-intermediate and intrusive granitic magmas shows that the former are on average slightly more viscous. Given the typical strain rates in silicic magma chambers, magma rheological behavior is expected to be dominantly Newtonian, bubbles having a minor rheological influence at depth although exceptions can exist. Thus whether a silicic-intermediate magma is erupted or frozen at depth depends primarily on the rheological properties of surrounding terranes or on external tectonic factors, but not on the rheology of the magma itself. However, preeruptive viscosities of extrusive magmas rarely exceed  $10^6$  Pa.s, which suggests that crystal-melt mushes with higher viscosities cannot leave the magma storage regions beneath volcanoes. The narrow range of viscosities displayed by silicic-intermediate magmas results from both the strong control that pressure exerts on volatile solubilities in silicate melts and thermal limitations required to produce acid magmas. Considerations of the relationships between magma crystallinities, bulk  $\text{SiO}_2$ , and preeruptive melt  $\text{H}_2\text{O}$  contents show that the higher the melt  $\text{H}_2\text{O}$  content is the higher the maximum crystallinity that a given magma will be while still being potentially erupted. An empirical correlation is proposed that enables us to estimate preeruptive melt  $\text{H}_2\text{O}$  contents of erupted magmas by knowing their crystallinity and bulk  $\text{SiO}_2$ .

### 1. Introduction

Silicic magmas are often chosen to illustrate the end-member behavior of viscous fluids on Earth, with inferred viscosities that span nearly 10 orders of magnitude ( $10^5$ - $10^{13}$  Pa.s), whereas basalt viscosities cluster tightly around  $10^1$ - $10^2$  Pa.s [e.g., *McBirney and Murase*, 1984]. Such contrasted behavior is still viewed as a major reason for basalts being easily erupted at the Earth's surface while silicic magmas pond at depth. However, recent studies have shown that from a rheological standpoint, granites behave in much the same way as basalts, yet they are colder by 300°-500°C [see *Wall et al.*, 1987; *Clemens and Mawer*, 1992]. They are emplaced at high levels in the crust through fractures, with magma flow rates of  $\sim 0.01$ - $1 \text{ m s}^{-1}$  and with temperatures close to the liquidus when emplaced, filling, in some instances, upper crustal chambers of  $\sim 150 \text{ km}^3$  within a few years [e.g., *Clemens and Mawer*, 1992; *Scaillet et al.*, 1996]. Granitic magmas are nevertheless still widely perceived as failed rhyolites. This view implicitly

assumes that the water contents of the magmatic precursors of plutonic rocks are much lower than that of their volcanic counterparts [e.g., *Johnson et al.*, 1994]. Attempts to determine intensive parameters governing the evolution of plutonic rocks through a classical petrological approach give rise to estimates that are, at best, legacies of near-solidus conditions [e.g., *Whitney*, 1988]. Most of the current viscosity determinations for granitic magmas are thus necessarily fraught with assumptions regarding temperature, melt composition, and crystal content. Early experimental work [*Maaloe and Wyllie* 1975], though pioneering the use of phase equilibrium studies as a working tool for inferring intensive conditions in the evolution of plutonic rocks, has suffered from the lack of analytical methods for measuring the water content of quenched experimental charges and is thus of little use for constraining the rheological properties of granitic liquids and magmas. This is apparent in the compilation of *Clemens* [1984] on the water content of silicic-intermediate magmas, where most phase equilibrium studies carried out on plutonic rocks provide only maximum estimates of melt  $\text{H}_2\text{O}$  contents. More recently, *Whitney* [1988] reviewed experimental evidence for the water content of granitic magmas and concluded that their  $\text{H}_2\text{O}$  contents were more likely in the range 2-4 wt %, although for similar reasons (i.e., lack of analytical tools) the melt  $\text{H}_2\text{O}$  contents (not to be confused with the magma  $\text{H}_2\text{O}$  content) could not be better determined. In contrast, recently

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performed phase equilibrium studies have benefited from the analytical improvements made in the past 10 years in the determination of melt H<sub>2</sub>O contents [e.g., Newman *et al.*, 1986; Delouie *et al.*, 1995; Devine *et al.*, 1995]. This enables a more accurate characterization of the rheological properties of experimental phase assemblages and granitic magmas. In parallel, the new analytical tools for measuring the H<sub>2</sub>O content of quenched glasses, together with well-calibrated thermobarometers for the determination of temperature, pressure, and  $fO_2$ , have also been extensively applied recently to volcanic rocks. The two types of approaches now allow an accurate determination of the major factors controlling magma viscosity (temperature, liquid composition including water content, and crystal content), either, for volcanic systems, in the magma storage zone [e.g., Gardner *et al.*, 1995a] or, for plutonic systems, at the emplacement level of the magmas [e.g., Clemens and Wall, 1981].

In this paper we use recently determined experimental phase equilibria to infer the viscosities of five crustally derived granites during their ascent and emplacement and to compare them to the preeruption melt viscosities of silicic-intermediate extrusive magmas. We show that melt viscosities cluster around  $10^4$ - $10^5$  Pa s irrespective of the either extrusive or intrusive nature, temperature, and H<sub>2</sub>O content of the magmas.

## 2. Method and Rocks Studied

To constrain the viscosity of silicic plutonic rocks, we use the results obtained from the phase relations recently determined on several granites. The granitic plutons are those of Strathbogie, Watergums (Australia), Manaslu, Gangotri (Himalaya) and Jamon (Brazil) [Clemens and Wall, 1981; Clemens *et al.*, 1986; Scaillet *et al.*, 1995; R. Dall'Agnoli *et al.*, Evolution of A-type granite magmas: an experimental study of the lower proterozoic Jamon granite, eastern Amazonian craton, Brazil, submitted to Journal of Petrology, 1998]. Their SiO<sub>2</sub> contents range from 71 to 75 wt % and the rocks are either peraluminous or metaluminous. They are representative of different types of granitic magmatism although all are formed by partial melting of the continental crust. The experimental studies have constrained the  $T$ -H<sub>2</sub>O<sub>in melt</sub>(wt %) conditions and crystal content during the emplacement of the magmas, by comparing the experimental phase equilibria with the natural crystallization sequence as deduced from textural observations. All these studies have invariably confirmed the crystal-poor nature of the magmas during emplacement [see Wall *et al.*, 1987], suggesting that fractionation was limited during ascent and hence that magma composition at its emplacement level (where the magma chamber was built) was close to that produced in the partial melting zone. In other words, these magmas were emplaced in a nearly entirely molten state in upper crustal chambers, where crystallization occurred. The most important aspect of the experimental work is that it has confirmed the water-rich nature (3-7 wt % H<sub>2</sub>O<sub>in melt</sub>) of the magmas parental to these plutonic rocks. These studies thus represent the best opportunity to quantify the viscosity of granites during at least two major steps of their evolution, ascent, emplacement, and crystallization.

For volcanic rocks we restrict our analysis to rocks whose preeruption conditions have been determined either from phase equilibrium experiments (Mount St. Helens dacite, Fish

Canyon Tuff latite, Spor Mountain rhyolite, El Chichón trachyandesite, Pinatubo dacite, and Mt. Pelée andesite) or from the direct measurement of the H<sub>2</sub>O content of melt inclusions, and FeTi oxides for temperature (Bishop Tuff, Cerro Toledo, Lower Bandelier, Taupo Rhyolite, Crater Lake, Pine Grove, Katmai, Fantale, Inyo Dome, and Krakatau). For the Bishop Tuff several recent studies have determined its preeruptive melt H<sub>2</sub>O content and variation during the main eruptive cycles [Anderson *et al.*, 1989; Skirius *et al.*, 1990; Dunbar and Hervig, 1992a; Wallace *et al.*, 1995]. In this study we use melt H<sub>2</sub>O contents given by Anderson *et al.* [1989], Skirius *et al.* [1990] and Wallace *et al.* [1995], yet to avoid redundancy, only those of Skirius *et al.* are reported in table 1 and in figures 1-4. For cases where the determined H<sub>2</sub>O contents of melt inclusions analyzed in a given sample display a dispersion much larger than the analytical uncertainty (for instance, Cerro Toledo) [Stix and Layne, 1996], although the anhydrous major element composition of the glass inclusions remains constant, only the highest values were used for the viscosity calculations. In doing so we assume that H<sub>2</sub>O-poor glass inclusions were affected by syneruptive to posteruptive H<sub>2</sub>O loss, an assumption supported by results obtained from the study of Pinatubo melt inclusions [Westrich and Gerlach, 1992; Rutherford and Devine, 1996; Gerlach *et al.*, 1996; Scaillet and Evans, 1998]. Apart from Macusani, volcanic rocks for which preeruptive H<sub>2</sub>O contents have been inferred only through mineral-fluid or mineral-melt equilibria were not considered here. The reason for this is that such an approach often yields a poorly constrained range of melt H<sub>2</sub>O contents given the uncertainties in experimental calibrations and solution models. These problems, added to the fact that reported temperatures may vary by more than 100°C, make it difficult to choose the appropriate  $T$ -H<sub>2</sub>O<sub>in melt</sub>(wt %) values for calculating melt viscosities. The Macusani volcanics are considered here because pre-eruptive melt H<sub>2</sub>O content and temperature are particularly well constrained [Pichavant *et al.*, 1988a,b] and allow the database to be extended toward temperatures below 650°C (Table 1). The compositional spectrum covered by the volcanic rocks is larger than that for the plutonics, with bulk SiO<sub>2</sub> contents ranging from trachyandesitic to rhyolitic (57-78 wt %). The data set includes peralkaline, metaluminous, and peraluminous compositions, which are associated with extensional (peralkaline) to compressional (peraluminous) regional tectonic stress regimes. Although examples of explosive eruptions largely dominate the data set, effusive volcanic rocks are also represented (Inyo Dome). For the explosive eruptions the samples represent a wide range in ejected masses, with volumes in dense rock equivalent ranging from 0.1 km<sup>3</sup> (Mt. Pelée) up to 3000 km<sup>3</sup> (Fish Canyon Tuff). Taken together, the plutonic and volcanic rocks represent a range of intermediate to silicic magmas involved in a large variety of tectonic settings, which ensures that our conclusions are of general validity.

## 3. Viscosity Calculations

For metaluminous felsic melts that fit the following chemical criteria (wt %), we use the empirical equation of Hess and Dingwell [1996]:  $SiO_2 = 73.2 - 78.6; CaO + MgO + FeO = 0 - 2.34; (2Na + 2K + Mg + Ca + Fe^{2+}) / (2Al + 2Fe^{3+}) = 0.93 - 1.02$ . For

**Table 1.** Calculated Melt and Magma Viscosities for Natural Intermediate to Silicic Plutonic and Volcanic Rocks

Rock	SiO <sub>2</sub> Magma, wt %	SiO <sub>2</sub> Melt, wt %	Crystals, % volume	T, °C	H <sub>2</sub> O Melt, wt %	log $\eta$ Melt, Pa s	log $\eta^a$ Magma, Pa s	Method <sup>b</sup>	References <sup>c</sup>
<i>Plutonics</i>									
Gangotri	72.9	72.9	<10	700	7.0	4.2	4.4	Exp, Sc	1
Manaslu	73.0	73.0	<10	750	5.5	5.0	5.2	Exp, Sc	1
Jamon	71.5	71.5	<10	920	4.5	4.1	4.2	Exp, Sh	2
Stratboghie	70.4	70.4	<10	850	4.5	4.6	4.8	Exp, Sh	3
Watergums	74.5	74.5	<10	900	3.5	4.7	4.9	Exp, Sh	4
<i>Volcanics</i>									
Fantale	71.0	71.0	14	800	4.6	3.9	4.2	IR, Sh	5
Fantale	71.0	71.0	14	800	4.9	3.7	4.0	IR, Sh	5
Pine Grove	75.1	76.4	32	675	7.1	5.3	5.8	IR, HD	6, 7
Pinatubo	64.0	78.3	47	780	6.4	4.5	5.7	Exp, IR, IP, HD	8, 9, 10
El Chichón	57.2	71.4	57	800	5.0	4.5	7.2	Exp, Sh	11
Crater Lake	73.0	73.0	10	885	3.9	4.7	4.9	IR, Sh	12
Crater Lake	73.0	73.0	10	885	4.7	4.2	4.4	IR, Sh	12
St. Helens-1980	62.8	73.5	37	930	4.6	3.7	4.4	Exp, VDB, Sh	13
St. Helens-T	63.2	70.2	31	893	4.6	3.8	4.3	Exp, Sh	14
St. Helens-Wn	67.2	74.8	27	847	4.8	3.8	5.1	Exp, Sh	14
St. Helens-Bi	64.1	72.8	41	913	3.7	4.5	5.3	Exp, Sh	14
St. Helens-Pu	63.3	74.3	46	870	4.3	4.6	5.7	Exp, Sh	14
St. Helens-Ye	65.5	75.6	40	795	5.6	4.5	5.3	Exp, Sh	14
St. Helens-Yn	65.9	75.5	35	768	6.5	4.4	5.0	Exp, Sh	14
St. Helens-Yn	65.8	74.8	32	790	6.3	4.2	4.7	Exp, Sh	14
Fish Canyon	64.4	77.1	40	760	5.0	5.2	6.0	Exp, HD	15
Spor Mountain	72.9	72.9	30	735	6.0	4.8	5.6	Exp, HD	16
Katmai Andesite	58.6	74.3	25	990	1.0	4.1	4.7	IR, IP, Sh	17, 18
Katmai Dacite	64.6	77.4	10	950	2.8	4.5	4.6	IR, IP, HD	19, 18
Katmai Rhyolite	77.4	78.5	1	900	4.0	4.4	4.4	IR, IP, HD	19, 18
Inyo Dome	71.5	71.5	5	897	3.9	4.5	4.5	IP, Sh	20, 18
Inyo Dome	73.7	73.7	5	874	4.3	4.5	4.5	IP, HD	20, 18
Taupo Plinian	74.0	76.0	5	850	4.3	4.9	5.0	IP, Sh	21, 22
Taupo Okaia	74.0	76.0	13	798	5.9	4.4	4.6	IP, Sh	21, 22
Taupo Hapete	74.0	76.0	5	839	4.3	5.0	5.1	IP, Sh	21, 22
Bishop Plinian	77.0	77.0	7	710	6.3	5.2	5.3	IR, HD	23, 18
Bishop Chidago	77.0	77.0	10	720	6.5	5.0	5.2	IR, HD	23, 18
Bishop Mono	77.0	77.0	22	790	4.2	5.2	5.7	IR, HD	23, 18
Cerro Toledo-15-8	72.0	72.0	<5	813	5.3	4.6	4.7	IP, HD	24, 25
Cerro Toledo-15-9	73.4	73.4	<5	813	6.8	4.2	4.3	IP, HD	24, 25
Cerro Toledo-15-11	73.0	73.0	<5	777	6.8	4.4	4.5	IP, HD	24, 25
Cerro Toledo-6-8	73.0	73.0	<5	697	6.7	5.2	5.3	IP, HD	24, 25
Bandelier Upper	73.1	73.1	<5	697	6.2	5.4	5.5	IP, HD	24, 26
Mt. Pelée	61.0	75.0	50	875	5.5	3.9	5.8	Exp, IR, Sh	27
Montserrat	59.0	77.6	40	830	4.7	4.7	5.5	Exp, IR, Sh	28, 29
Krakatau Rhyodac.	69.5	72.0	10	885	4.0	4.4	4.6	VDB, Sh	30
Macusani	73.0	72.0	45	635	7.2	4.9	5.9	Calc, Sc	31

For the Spor Mountain and Macusani rhyolites the H<sub>2</sub>O contents include F, B, and Li

<sup>a</sup>Magma viscosity calculated using the melt viscosity and the Einstein-Roscoe relationship as modified by Marsh [1981] for crystal contents up to 30% volume from equation (2) for higher crystal contents.

<sup>b</sup>Method of H<sub>2</sub>O determination: Exp, experimental phase equilibria; IR, infrared spectroscopy; IP, ion probe, VDB, volatile by difference; Calc., calculated through mineral equilibria. The model used to calculate the melt viscosity is given as follows: Sc, Scaillet *et al.* [1996]; Sh, Shaw [1972]; or HD, Hess and Dingwell [1996].

peraluminous leucogranite ones the equation of Scaillet *et al.* [1996], experimentally calibrated on a natural strongly peraluminous melt, is used. The model of Shaw [1972] was used to calculate silicate melt viscosities of all remaining melts because (1) it is the only one that explicitly takes into account the liquid composition in hydrous systems, and (2) its experimental range of calibration, in terms of melt H<sub>2</sub>O content and temperature, is similar to that relevant for natural granites and for most silicic erupted magmas, as deduced from the studies mentioned in section 2. Although recent work has shown that the Shaw model may either underestimate or overestimate melt viscosities at low water contents [Stein and Spera, 1993; Schulze *et al.*, 1996; Dingwell *et al.*, 1996;

Baker, 1996], good agreement between experimental and calculated melt viscosities is observed in the range of melt H<sub>2</sub>O contents of interest for the present study (mainly, 3-7 wt % H<sub>2</sub>O), differences barely exceeding 1 log unit. Although low melt H<sub>2</sub>O contents do occur in the evolution of magmatic rocks, they are attained mainly in volcanic contexts, during the final stages of degassing. All these processes are not considered in this paper (see, for instance, Dingwell *et al.* [1996]) where only preeruption conditions are taken into account.

It is well established that silicate melt viscosities depend, among other factors, on the strain rate, viscosities at high strain rates exhibiting significant departure from Newtonian behavior [e.g., Webb and Dingwell, 1990; Dingwell and

Webb, 1990; Bottinga, 1994]. The strain rates required to observe non-Newtonian behavior in silicate melts correspond to timescales 3 orders of magnitude lower than the structural relaxation timescale  $\tau$  for the melt [Webb and Dingwell, 1990], the latter parameter being related to the relaxed shear viscosity  $\eta_{\text{melt}}$  by the Maxwell relation:

$$\tau = \eta_{\text{melt}} / G_{\infty} \quad (1)$$

$G_{\infty}$  being the infinite frequency shear modulus of the melt [see Dingwell and Webb, 1990]. Given that  $G_{\infty}$  is constant for most silicate melts ( $\log G_{\infty} = 10 \pm 0.5$  Pa) [Dingwell and Webb, 1990], variations in the relaxation time arise predominantly from melt viscosity. For the range of melt viscosities found in this paper (see below), the structural relaxation time is of the order of  $10^{-5}$ - $10^{-6}$  s. Estimated convective rates in silicic magma chambers are of the order of  $10^{-8}$  s $^{-1}$  [see Spera et al., 1988], which is largely below the threshold for the onset of non-Newtonian viscosity (3 log units below the log of the relaxed strain rate ( $1/\tau = 10^5$  s $^{-1}$ )). In addition, there have been suggestions based both on analogical [e.g., Brandeis and Marsh, 1989] and numerical [e.g., Brandeis and Jaupart, 1986] modeling that convection in silicic magma chambers is weak or nonexistent, convective motions virtually ceasing when crystallization starts [Marsh, 1989]. These facts suggest that the melt phase obeys dominantly a Newtonian behavior under pre-eruption conditions.

For calculating magma viscosities having less than 30% (volume) of crystals we use the Einstein-Roscoe relationship, as modified by Marsh [1981]. Recent studies have shown that this empirical relation holds well at low crystal contents [Pinkerton and Stevenson, 1992; Lejeune and Richet, 1995]. At higher crystal contents, however, this relation is no longer valid [see Pinkerton and Stevenson, 1992], and we use the following equation, as recommended by Dingwell et al. [1993]:

$$\eta_{\text{magma}} = \eta_{\text{melt}} \{1 + 0.75 [(f/f_m) / (1 - f/f_m)]^2\} \quad (2)$$

where  $\eta_{\text{magma}}$  and  $\eta_{\text{melt}}$  are the viscosities of the magma and the melt, respectively,  $f$  is the volume fraction of crystal, and  $f_m$  is the concentration of crystals beyond which the magma has an "infinite" viscosity. Following Marsh [1981] and Pinkerton and Stevenson [1992] we have chosen a value of 0.6 for  $f_m$ . Strictly,  $f_m$  is a function of crystal size distribution, which is unknown for the magmas considered in this report, and thus we have taken the value for a monodisperse suspension. Equation (2) gives relaxed shear viscosities, i.e., the viscosity at low strain rates, in keeping with the inference of low strain rates at work in silicic magma chambers.

In addition to crystals, magmas can contain bubbles. Recent studies of silicic- intermediate arc magmas using either trace element behavior [Wallace et al., 1995] or sulfur yields associated with explosive eruptions [Scaillet et al., 1998] yielded preeruption fluid contents that generally did not exceed 5 wt %, with fluids having H<sub>2</sub>O-rich compositions. For H<sub>2</sub>O densities in the range 800°-900°C and 200-300 MPa, such an amount translates into a fluid volume fraction of ~10%. As with crystals, the effect of bubbles on magma viscosity depends on size distribution and strain rate, among other parameters (i.e., capillary number and viscosity ratio) [Spera et al., 1988; Stein and Spera, 1992; Bagdassarov and Dingwell, 1992, 1993]. The presence of bubbles introduces an additional relaxation time, related to the response of bubbles to

viscous stresses [Bagdassarov and Dingwell, 1992, 1993]. This relaxation time  $\tau_b$  is related to the melt viscosity and bubble radius  $R$ :

$$\tau_b = \eta_{\text{melt}} R / \sigma \quad (3)$$

$\sigma$  being the surface tension. If the strain rate is greater than  $\tau_b^{-1}$ , bubbles behave as rigid particles, like crystals, and the viscosity of the bubble-bearing magma is increased [Stein and Spera, 1992]. Inversely, when the strain rate is slow (i.e., in magma chambers), bubbles will deform viscously and the magma viscosity is decreased [Bagdassarov and Dingwell, 1993]. There are, obviously, no direct measurements of  $R$  at preeruption conditions (excluding fluid inclusions trapped in melt inclusions or crystals). Crystallization experiments of natural magmas saturated with H<sub>2</sub>O-CO<sub>2</sub> ( $\pm$  H<sub>2</sub>S) fluids aimed at defining preeruption conditions generally give bubble radii in the range 10-100  $\mu\text{m}$  [e.g., Scaillet and Evans, 1998]. These are probably minimum values since the timescale of laboratory experiments is likely to be much shorter than the residence time of magmas in the storage region, during which bubbles can coalesce and thus increase in size (but note that large deformations will have the opposite effect; Stein and Spera [1992]). Taking an average melt viscosity of  $10^5$  Pa s, a surface tension of  $0.1 \text{ N m}^{-1}$  [Bagdassarov and Dingwell, 1992], and  $R = 100 \mu\text{m}$  yields  $\tau_b = 100$  s whose inverse is  $10^{-2}$  s $^{-1}$ . This is significantly higher than strain rates in silicic magma chambers, so that bubble-bearing magmas can be predicted to have lower viscosities than bubble-free ones. The experiments performed by Bagdassarov and Dingwell [1992] on a viscous ( $\sim 10^9$  Pa s) rhyolitic melt at low strain rates ( $10^{-5}$ - $10^{-7}$  s $^{-1}$ ) yielded the following relationship:

$$\eta_{\text{magma}} = \eta_{\text{melt}} [1 / (1 + 24.4 f)] \quad (4)$$

where  $f$  is the volume fraction of bubbles. Use of (4), together with the estimate of preeruption fluid contents, results in a maximum decrease of the magma viscosity by a factor of 3. This lowering effect of bubbles upon magma viscosity has been shown to hold also in melt+crystal+bubble suspensions [Bagdassarov et al., 1994], the presence of bubbles counteracting that of solid particles. For instance, a magma having 16% bubbles and 16% sapphire spheres has a shear viscosity almost identical to that of the melt phase alone. Although strain rates in silicic magma chambers are expected to be low, it is instructive to calculate how much magma viscosity will be affected by bubbles under high strain rates. Stein and Spera [1992] have obtained the following empirical equation for calculating the viscosity of magmatic emulsions with a melt viscosity of  $10^5$  Pa s, and low gas content (< 7% vol), at relatively high strain rates ( $0.06$ - $7$  s $^{-1}$ ):

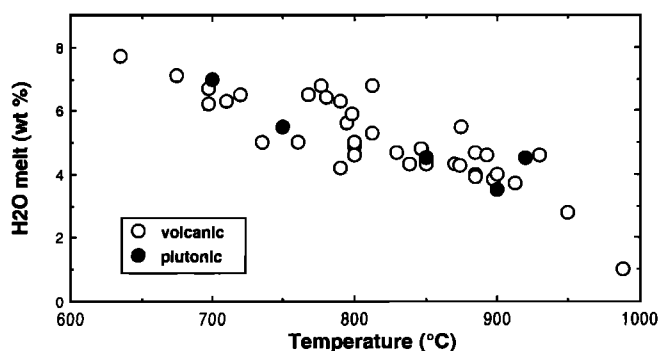
$$\eta_{\text{magma}} = \eta_{\text{melt}} (1 + 13.1 f) \quad (5)$$

where  $\eta_{\text{magma}}$  stands for the viscosity of melt+bubbles and  $f$  is the volume fraction of bubbles. Equation (5) shows that under the storage conditions of silicic magmas, high strain rates will rise by a factor of 2 the magma viscosity. Altogether, these considerations suggest that as far as preeruption conditions are considered, bubbles play a minor role in magma rheology when compared, for example, to that played by variations in crystal contents of erupted magmas (0-60%) or to the viscosity variations arising from changes in melt H<sub>2</sub>O contents. Accordingly, unless otherwise stated, magma viscosities reported here do not take into account the role of bubbles. In

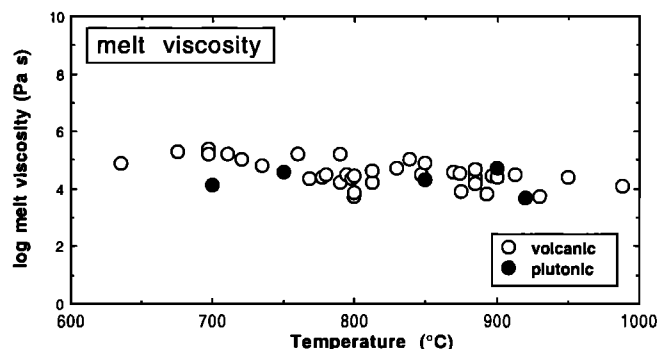
view of the above calculations and considering the possible range of fluid content found in natural magmas (see above), the melt viscosities reported in this work can therefore be considered to be accurate to within a factor of 3.

#### 4. Melt and Magma Viscosities Before Uprise

The calculated melt and magma viscosities of the rocks used in this study, together with the main compositional features necessary for the calculations, are listed in Table 1. In Figure 1 the melt H<sub>2</sub>O contents of granitic magmas are plotted against the temperature, both variables being taken for conditions prevailing during magma emplacement in the upper crust. A similar plot is constructed for the preeruption *T*-H<sub>2</sub>O (wt %) conditions of the volcanic rocks. The overall pattern is a nearly linear inverse relationship between *T* and H<sub>2</sub>O, the drier magmas being the hottest. No difference appears between plutonic and volcanic rocks. In Figure 2 the melt viscosities of granitic magmas are plotted against the emplacement temperature of the magma and compared to the preeruptive melt viscosities of the volcanic rocks. Again, both plutonic and volcanic melt viscosities lie on a single horizontal trend over more than 350°C, virtually spanning the entire temperature range covered by silicic magmas [Ghiorso and Sack, 1991]. It is worth mentioning that the Macusani volcanics, whose estimated preeruptive temperature and melt H<sub>2</sub>O content are 635°C and 5 wt %, respectively [Pichavant *et al.*, 1988a,b], give a liquid viscosity of 7.2 unit log Pa.s; that is, they would plot well above the general trend. However, if the effects on melt viscosity of volatiles other than H<sub>2</sub>O (F, Li, and B) are considered [Dingwell *et al.*, 1992; Baker and Vaillancourt, 1995] and assuming that these effects add up, the preeruptive melt viscosity for the Macusani volcanics is 4.89 log unit Pa s, in good agreement with the calculated viscosities for other less evolved felsic magmas. A similar rationale can be applied to the Spor Mountain rhyolite in which the additive effects of H<sub>2</sub>O and F result in a preeruption viscosity similar to other volcanic rocks as well (Table 1). These examples illustrate the fact that the pattern of Figure 2 is not specific to magmas in which the dominant volatile is H<sub>2</sub>O but also applies to those highly enriched in halogens or other volatile elements. Conversely, it suggests that melts enriched in volatiles other



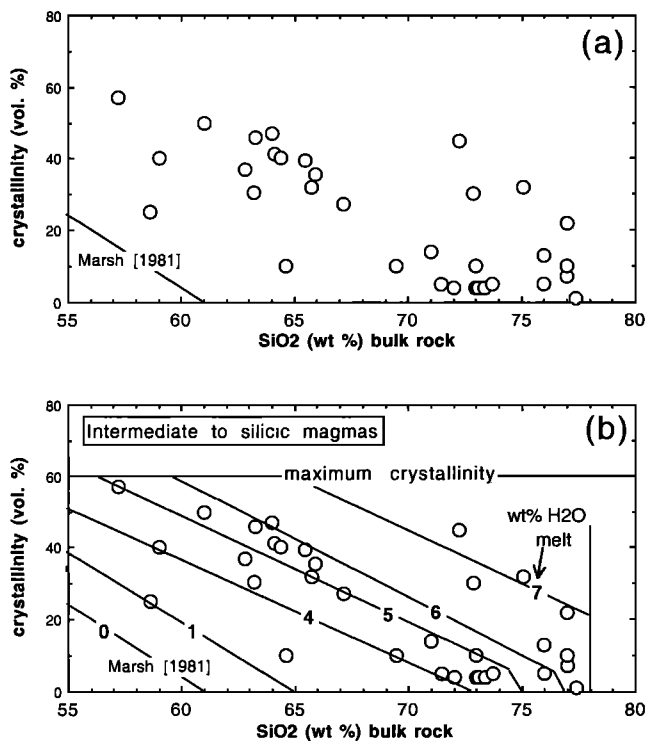
**Figure 1.** Relationship between temperature and melt H<sub>2</sub>O content for both plutonic (solid circles) and volcanic (open circles) rocks. For the plutonic rocks the conditions correspond to magma emplacement, and for volcanic rocks the conditions correspond to preeruption conditions. Source of data is given in Table 1.



**Figure 2.** Melt viscosities for granitic magmas (solid circles), plotted as a function of their emplacement temperature and compared to the preeruption melt viscosities of volcanic rocks (open circles) listed in Table 1. For all granites the melt composition used in the viscosity calculation is that of the bulk composition used in the phase equilibrium studies. See text and Table 1 for details on viscosity calculations.

than water will not segregate from their source earlier (i.e., at lower viscosity) than those containing only dissolved water. The average value of melt viscosity for the whole data set is  $\log \eta_{\text{melt}} = 4.5 \pm 0.5$  (Pa s).

The experimental work indicates that magmas parental to the plutonic rocks were all emplaced at near liquidus conditions, with crystal contents less than 10 wt % and thus that melt viscosities were close to magma viscosities. In contrast, preeruptive crystallinities for volcanic rocks vary from 0 to 60%, displaying a clear inverse relationship with bulk SiO<sub>2</sub> content (Figure 3). This relationship could partly reflect the increase in incubation time for the onset of crystallization in silicate melts as SiO<sub>2</sub> increases. Numerical modeling indicates, however, that crystallization in large magma chambers occurs under quasi-equilibrium conditions with small undercooling, except near the very borders at early stages of the intrusion [Brandeis *et al.*, 1984]. Given that the volcanic rocks used in this study come from magma chambers of moderate to large sizes (1 to several hundred km<sup>3</sup>, the volume of erupted magma representing on average 10-20% of the magma chamber), it is likely that the thermal regime in their storage region was characterized by slow cooling rates and thus that the relationship of Figure 3 is not due to kinetic effects. Therefore, given that both intrusive and extrusive magmas display similar melt viscosities, if the effects of crystals are taken into account, the viscosities of silicic volcanic rocks are on average slightly higher than that of their granitic counterparts (Figure 4). Nevertheless, apart from a single outlier (El Chichón trachyandesite), volcanic rock viscosities plot in a restricted viscosity range, 10<sup>4</sup>-10<sup>6</sup> Pa s. This suggests that magmas more viscous than 10<sup>6</sup> Pa s cannot leave their host chamber but instead freeze at depth. This does not preclude silicic magmas from reaching higher viscosities, however, but highly viscous conditions are likely to be attained only during the last stages of magma ascent and degassing [e.g., Dingwell *et al.*, 1996]. Although magma viscosities are less clustered than melt viscosities, the general pattern with temperature variation is again flat, which indicates that hot magmas are not erupted more easily than cold ones. Our analysis therefore suggests that in many, if not all, instances the preeruption melt and magma viscosities are at, or below, 10<sup>4.5</sup> and 10<sup>6</sup> Pa s, respectively. One apparent exception to this rule is provided



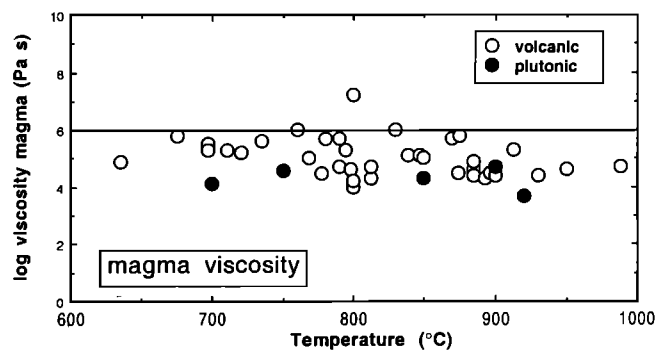
**Figure 3.** Relationship between magma crystallinity and bulk rock  $\text{SiO}_2$  for the volcanic rocks of Table 1, including (a) a plot of the raw data together with the Aleutian lavas trend obtained by *Marsh* [1981] and (b) the same diagram contoured with lines of isomelt  $\text{H}_2\text{O}$  contents. All fits are handmade best estimates.

by the El Chichón magma, whose crystal content is close to that at which the magma reaches the solid behavior field. The calculated magma viscosity is more than 1 order of magnitude higher relative to other magmas and, at such high crystal content, magmas can be anticipated to also have a significant yield strength [Pinkerton and Stevenson, 1992]. Nevertheless, this magma has been erupted, and the eruption was of Plinian type. Interestingly, the restored preeruption fluid content of this magma ranges between 8 and 26 wt %, depending on the  $\text{CO}_2$ - $\text{SO}_2$  ratio used [Scaillet *et al.*, 1998], which implies that the volume fraction of the fluid was ~ 20-50% of the scavenged part of the magma chamber. Considering the role of bubbles on magma rheology as summarized above, we tentatively suggest that the eruption of El Chichón magma was made possible, or at least greatly facilitated, by the large volume of the fluid phase. Specifically, the experimental results of *Bagdassarov et al.* [1994] on the effect of bubbles predict that the viscosity of El Chichón magma is only marginally different from that of the melt phase alone, falling well within the range displayed by the other magmas (Table 1). Finally, though in some instances a relation between eruption intensity and magma viscosity has been clearly established (Mount St. Helens) [Gardner *et al.*, 1995a], no correlation between magma viscosity and eruption volume, or any other parameter, was found.

## 5. Inferring Preeruptive Melt $\text{H}_2\text{O}$ Contents From Crystallinity

The relationship between crystallinity and bulk  $\text{SiO}_2$  (Figure 3) is qualitatively similar to that obtained by *Marsh*

[1981] for a series of basaltic-andesitic lavas erupted from the Aleutians volcanic center of Atka. Both trends are parallel, the one obtained here being displaced toward higher  $\text{SiO}_2$  content (Figure 3). In both cases, volcanic rocks with more than 55-60% crystals are not found, an observation that suggests, as *Marsh* noted, that there is a locking point, characterized by a critical crystallinity, beyond which magma viscosity increases so dramatically that it behaves essentially as a solid, a proposal corroborated by experimental data [e.g., *Pinkerton and Stevenson*, 1992; *Lejeune and Richet*, 1995]. In his analysis of the Aleutians lavas, *Marsh* derived a value of  $10^{4.6}$  Pa s for this viscosity threshold. This value is lower by 0.5 log units than the average magma viscosity obtained in this study for volcanic rocks ( $\log \eta_{\text{magma}} = 5.1 \pm 0.7$  Pa s). However, although the magmatic  $\text{H}_2\text{O}$  contents were not determined, nearly dry conditions (< 1 wt %) were inferred for the Aleutian lavas [Marsh, 1981], as opposed to the volcanic rocks presented here whose average melt  $\text{H}_2\text{O}$  content is ~ 4-5 wt %. The two samples lying in between the Aleutian lavas trend and the one obtained here are the andesite and dacite erupted at Katmai, whose water contents are among the lowest of the data set, 1 and 2.8 wt %, respectively. Conversely, the point plotting well above the general trend corresponds to the volcanic rock richest in water, with 7.1 wt % dissolved  $\text{H}_2\text{O}$  (Pine Grove). Thus we confirm the prediction of *Marsh* that water increases the critical crystallinity at a given  $\text{SiO}_2$  content. In other words, magmas rich in water can be erupted at a higher crystallinity than those poor in  $\text{H}_2\text{O}$ . As proposed by *Marsh*, the relationship between bulk  $\text{SiO}_2$  content and magma crystallinity can be used in a quantitative way to infer preeruptive melt  $\text{H}_2\text{O}$  contents. A generalization of his model has been attempted with our data set and is shown in Figure 3b, where the diagram of Figure 3a is contoured with lines of constant melt  $\text{H}_2\text{O}$  contents. All correlations were assumed to be linear, yet some curvatures probably occur at both ends of the trends. Most of the previous analyses ignore the possible role of bubbles. The overall coherent trend obtained does suggest, however, that in most instances the gas phase abundance is too low to have any appreciable effect on magma rheology under preeruption conditions, as previously argued. Nevertheless, not all the volcanic rocks conform to the lines of isomelt  $\text{H}_2\text{O}$  content, in particular in the high  $\text{SiO}_2$  range. For instance, the Mono unit of the Bishop Tuff is clearly off the general trend (Table 1). Owing to the lack of precise preeruptive temperature constraints, the Lower Bandelier Tuffs



**Figure 4.** Magma viscosities versus temperature for the plutonic (solid circles) and volcanic (open circles) rocks listed in Table 1. Note that plutonic rocks are on average slightly less viscous than volcanic ones.

analyzed by *Dunbar and Hervig* [1992b] could not be used in the present study, but some of the samples would plot in the same area as the Mono lobe. Although, as illustrated by El Chichón, a high amount of fluid could explain the off position of these high-silica rhyolites, the restored fluid content of the Mono unit is not abnormally high (<5 wt %) [*Wallace et al.*, 1995]. This lack of correlation may thus reflect a specific characteristic of the large-magnitude volcanic events that gives rise to ignimbrite deposits, with eruptive dynamics fundamentally different from those of smaller eruptions. It is possible that the locking point rule does not hold for certain eruptions of particularly large magnitude. However, not all ignimbrites plot out off the trends, as exemplified by the Fish Canyon Tuff, one of the largest known ignimbrite deposits that fits well with other eruptive events. Clearly, future work will improve the quality of the fit between crystallinity, bulk SiO<sub>2</sub>, and melt H<sub>2</sub>O content, but we stress that such a diagram (Figure 3) is useful for a first-order estimation of preeruptive melt H<sub>2</sub>O contents of erupted magmas, especially those ejected during eruptions of small to moderate magnitude.

## 6. Dikes, Diapirs, and Emplacement Levels

The rather low viscosities found in this study are consistent with the recent proposal that felsic magma migration occurs through dikes rather than via diapirs [*Clemens and Mawer*, 1992; *Petford et al.*, 1993]. *Emerman and Marret* [1990] have shown that the mode of ascent of magmas (as either a dike or a diapir) can be estimated from the following dimensionless parameter:

$$\lambda = \eta_{\text{magma}} c / d \Delta\sigma \quad (6)$$

where  $c$  is the speed of propagation,  $d$  is the difference between the depth at which deviatoric stress becomes significant and the depth of final emplacement, and  $\Delta\sigma$  is the difference between the least compressive stress (horizontal for dike emplacement) and the other horizontal principal stresses. This relation evaluates the time needed to reorientate a dike relative to its rate of propagation. If  $\lambda$  is much lower than 1, then the magma will propagate upward as a dike without suffering significant reorientation due to the external stress field. Taking  $\Delta\sigma = 10$  MPa,  $d = 10,000$  m,  $c = 1$  m s<sup>-1</sup>, and  $\eta_{\text{magma}} = 10^{4.5}$  Pa s, the average viscosity found in this study for plutonic magmas, gives  $\lambda = 10^{-6.5}$ , which implies that granitic magmas have little chance of ending their crustal travel as diapirs. Note that even if the speed of propagation is decreased by 5 orders of magnitude (i.e.,  $c = 10^{-5}$  m s<sup>-1</sup>), the magma will still ascend as a dike. The parameter  $\lambda$  may approach, however, values close or higher than 1 at very small ascent distances ( $d = 10$ -100 m), and the emplacement level would be close, in fact, to the melt generation level.

The diking mechanism is characterized by rapid flow of the melt, which prevents crystallization and contamination during uprise [*Clemens and Mawer*, 1992], and explains the crystal-poor character of the magmas at the end of their crustal travel. Thus, for magmas that are produced by partial melting of the middle-lower continental crust and since pressure variations of less than 1 GPa have a negligible effect on melt viscosity [*Burnham*, 1964; *Schulze et al.*, 1996; *Scaillet et al.*, 1996], the viscosity-temperature trend of Figure 2 suggests that melt extraction from the partially molten source region also occurs for viscosities around  $10^4$ - $10^5$  Pa s. Viscosities of liquids

produced in fluid-absent melting experiments carried out on crustal protoliths are in the range  $10^3$ - $10^5$  Pa s [*Rushmer*, 1995; *Wolf and Wyllie*, 1995], which supports this conclusion. It should be stressed that the different granites considered in this study, although originating by fluid-absent melting processes, differ in the nature of the main hydrous phase involved in the melting reaction [*Whitney*, 1988]. Yet they all lie on the same viscosity-temperature trend, which indicates that the nature of the melting reaction has little influence on melt segregation and separation from its source region. In the same way, the extent of melting, which may vary depending on the nature and abundance of the hydrous phase breaking down (from 10 to 50%), appears not to be the critical factor triggering upward melt migration. The experimental data on plutonic rocks imply that granitic melts should be nearly perfectly extracted from their source with little entrainment of restites [see *Clemens and Mawer*, 1992]. The fact that there is no crustal, restite-poor, granite magma emplaced with melt viscosities demonstrably higher than  $10^5$  Pa s, suggests that this value also represents a viscosity threshold above which no melt can migrate alone (i.e., without restites), whatever the  $P$ - $T$ -conditions and the state of stress in the source.

The overall viscosity pattern clearly indicates that the volcanic-plutonic dichotomy does not arise from rhyolites being systematically less viscous than granites. It demonstrates also that viscosity cannot control the level of emplacement of silicic magmas. Factors like regional or local tectonic stress fields [e.g., *Scaillet et al.*, 1995] or the mechanical properties of the traversed terranes [e.g., *Hogan and Gilbert*, 1995] are more likely causes for the trapping of silicic magmas in the crust [see *Clemens and Mawer*, 1992]. In contrast to viscosity, density is probably one of the key factors that will control the ultimate level of emplacement of magmas. However, there is no consensus about the actual value of the partial molar volume of water dissolved in silicate melts, as well as about the possible compositional dependency of this parameter [*Lange*, 1994]. In addition, as emphasized in section 3, most silicic-intermediate magmas are saturated in a fluid phase [e.g., *Wallace et al.*, 1995], which seriously complicates density calculations. For instance, assuming for simplicity a crystal+melt density of  $2.4$  g cm<sup>-3</sup> and a fluid (H<sub>2</sub>O) density of  $0.5$  g cm<sup>-3</sup>, the magma density decreases between 2.31 and 2.02 g cm<sup>-3</sup> if its fluid content increases from 1 to 5 wt %, i.e., a density variation similar to that arising from a temperature change exceeding 1000°C [see *Lange and Carmichael*, 1990]. This outlines the dramatic importance that the fluid phase may have on magma density. The large uncertainties that still surround this last factor make the results of density calculations for magmas highly model-dependent, and for these reasons such calculations were not attempted in this study.

## 7. Concluding Statements

One of the striking features of this study is that silicic magmas display not only an upper limit in viscosity but also a lower one. Although the former limit is understandable from a rheological point of view (i.e., the magma becomes too viscous to flow), the reasons for the latter are less obvious. However, silicic magmas with a viscosity lower than  $10^4$  Pa s are apparently not produced. This compilation illustrates that besides their anhydrous composition, two main variables can lower the viscosity of silicic magmas, temperature and volatile contents. However, as discussed below, neither temperature

nor the volatile content can be expected to vary so as to decrease the viscosity of silicic magmas beyond the lower viscosity threshold displayed in Figure 1. In order to decrease significantly the viscosity of silicic magmas, temperatures largely in excess of 1000°C would be required (Figure 2). Nevertheless, at least for modern geological periods (i.e., post-Archaeon), the geothermal regime of the continental crust prevents such high temperatures from being reached, even in thinned continental crust. Ponding of hot mafic magmas at the base of the crust can induce partial melting, but thermal modeling indicates that the temperature of silicic magmas generated in this way hardly exceeds 950°C [Huppert and Sparks, 1988]. In addition, the produced melts, although hot, will also be dry owing to the low H<sub>2</sub>O content of most high-grade metamorphic rocks [e.g., Clemens and Vielzeuf, 1987]. If free H<sub>2</sub>O is present, then melting of crustal lithologies at temperatures around 1000°C will not yield rhyolitic magmas but, instead, andesitic magmas or even more mafic compositions. Protracted fractional crystallization of mafic magmas, on the other hand, can produce silicic derivatives [e.g., Helz, 1987] but only at temperatures below 1000°C, especially under H<sub>2</sub>O-rich conditions [e.g., Sisson and Grove, 1993].

An increase in the water content of the melt to decrease its viscosity is the other possibility to obtain a melt viscosity significantly lower than 10<sup>4</sup> Pa s. However, owing to the range of pressures at which most extrusive magmas are stored (200-300 MPa), the maximum water solubility of a rhyolitic melt will be in the range 6-7 wt %, and the corresponding melt viscosity will be again in the range 10<sup>4</sup>-10<sup>5</sup> Pa s, depending on temperature. Clearly, a silicic melt at 1000°C and with 7 wt % H<sub>2</sub>O will have a very low viscosity but, again, there are no known geological mechanisms able to produce such conditions, in keeping with the strong inverse correlation of Figure 1 that demonstrates that H<sub>2</sub>O and temperature are not independent parameters in magmas. Another parameter to be considered is the presence of volatiles other than H<sub>2</sub>O that have a lowering effect upon viscosity (B or F). However, despite the fact that these elements have an additive effect on viscosity (F and H<sub>2</sub>O) [Baker and Vaillancourt, 1995], the available evidence (Macusani and Spor Mountain) shows that there is an inverse correlation between the abundance of water and that of other fluxing elements, so that the net effect on viscosity is not expected to be large. A possible major exception to this rule is provided by strongly peralkaline rhyolites whose water solubility can be noticeably higher than 6-7 wt % at 200-300 MPa, as shown experimentally [e.g., Linnen et al., 1996; Dingwell et al., 1997], with corresponding low viscosities [Dingwell et al., 1998]. Such magmas may well display viscosities lower than 10<sup>4</sup> Pa s before eruption. Thus, apart maybe from peralkaline magmas, the above lines of evidence suggest that whatever is the mechanism that produces a silicic melt, the magma most inevitably ends in a quite narrow viscosity window, which is defined both by the strong control that pressure exerts on volatile solubilities in silicate melts and by the thermal limitations required to produce acid magmas.

Considering the strain rates in silicic magma chambers, it is expected that the magma rheology falls in the Newtonian domain. In general, the available evidence suggest that the role of bubbles on magma rheology can be neglected, but examples like the El Chichón magma show that the fluid phase may be an important rheological factor. In this respect, we note that

magmas with restored high fluid contents are also those with high crystal contents, which intuitively makes sense, since the amount of fluid must come in part from the crystallization of anhydrous minerals. However, the mechanisms at work in arc magmas are much more complicated than those arising from a simple closed crystallization mechanism. Processes like mixing with mafic magmas can seriously complicate this scenario. This parameter needs to be specifically addressed for each single volcanic event, and we encourage that such type of studies be carried out in the future. Important and immediate applications of such data would be to accurately estimate magma densities. For plutonic rocks there are no, as yet, methods to estimate the potential importance of fluids at the magmatic stage, though some were clearly rich in water at early stages of their evolution [Scaillet et al., 1995]. The diking mode of magma ascent implies high ascent rates and, in turn, high strain rates. Under these conditions, as with eruptive phenomena, bubbles will increase magma viscosity. However, if we take the comparison done in this work at face value, there are no reasons to believe that plutonic magmas display higher fluid contents than their volcanic counterparts. Instead, the current consensus is that crustally derived granitic magmas are generated, and most probably evolved, under fluid-absent conditions [i.e., Clemens and Vielzeuf, 1987].

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