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## Viscosity of Himalayan leucogranites: Implications for mechanisms of granitic magma ascent

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**Abstract.** The viscosity of the Gangotri Himalayan leucogranite has been experimentally determined between 800° and 1100°C, 300 and 800 MPa, for meltwater contents of 3.98 and 6.66 wt %. The melt viscosity is independent of pressure and shows an Arrhenian behavior relative to temperature within the range of conditions investigated. We present an empirical relation that can be used to determine leucogranite magma viscosities knowing their meltwater content and temperature. This relation together with phase equilibria experiments constrain the viscosity of the Himalayan leucogranites to be around  $10^{4.5}$  Pa s during their emplacement. These viscosities and the widths of dikes belonging to the feeder system are consistent with the theoretical relationship relating these two parameters and show that the precursor magma of the leucogranite was at near liquidus conditions when emplaced within host rocks with preintrusion temperatures around 350°C. Calculated terminal ascent rates for the magma in the dikes are around 1 m/s. Magma chamber assembly time is, on this basis, estimated to be less than 100 years (for a volume of 150 km<sup>3</sup>). In addition, the dynamical regime of the magma flow in the dikes was essentially laminar, thus allowing preservation of any chemical heterogeneity acquired in the source. These results constrain the viscosity of melts formed during the first steps of crustal anatexis, those involving muscovite breakdown, to be also around  $10^{4.5}$  Pa s. Thus compaction may not be the only mechanism of melt segregation in partially melted crustal rocks in view of the very short timescale inferred for magma ascent and emplacement.

### Introduction

The diking mechanism is the widely accepted mode of magma transport for basaltic magmas, as testified by abundant field evidence of basaltic dikes [e.g., Wada, 1994]. In contrast, the common circular plan view of many granitic intrusions is perhaps the argument that has convinced many authors that diapiric uprise is a viable mechanism of transfer for granitic magmas [e.g., Pons *et al.*, 1995] (see also reviews by Paterson and Fowler [1993], Paterson *et al.* [1991], Brown *et al.* [1995]). This mode of ascent went virtually uncontested for decades until Clemens and Mawer [1992] proposed, on thermal and mechanical grounds, that diking was a more viable alternative than diapirism for felsic melt migration [see also Pefferd *et al.*, 1993]. A major criticism of the diapir model is that rates of magma ascent, as obtained from early thermal and mechanical modeling [e.g., Mahon *et al.*, 1988], are extremely low, which should allow for extensive heat loss toward country rocks and thus crystallization at depth, while a number of geologic arguments show that granite intrusions are clearly disconnected from their source zone [Miller *et al.*, 1988]. However, Weinberg and Podlachikov [1994] have shown that rates of diapiric rise can significantly increase (e.g., 100 m/yr) if the crust behaves in a non-Newtonian fashion. Similarly,

the recent mathematical analysis of Rubin [1995] suggests that the thermal viability of granitic dikes should prevent them from propagating far from the source region, thus showing that the diking model has its shortcoming too. Clearly, the question of whether granite ascent through the crust occurs only via dikes or only via diapiric rise is not likely to be resolved through theoretical analysis alone, inasmuch as both the diapir and dike mechanisms need not to be exclusive to each other [Rubin, 1993].

Although the case for granitic dikes is clear by the analyses of Clemens and Mawer [1992] and Pefferd *et al.* [1993], assessment of this mechanism of ascent for natural felsic magmas has been, as yet, seriously hampered by two factors. The first is that in marked contrast to basaltic dikes, well-identified examples of feeder dikes of plutonic systems are exceedingly rare [e.g., Le Fort, 1981; John, 1988; Scaillet *et al.*, 1995a], a direct consequence of the fact that the bottom contact of most plutonic intrusions is not visible. The second is that the determination of the viscosities of magma frozen in dikes is, for obvious reasons, not straightforward. Current attempts [e.g., Pefferd *et al.*, 1993; Wada, 1994] rely on the empirical model of Shaw [1972] which requires knowledge of the melt composition, including its water content, melt temperature, and amount of crystals present at the time of dike intrusion. These parameters (in particular, the water content which most of the time is fixed arbitrarily) are poorly known for most plutonic rocks. The lack of precise determination of magma viscosity, a master variable that may influence whether granitic magmas will ascend as a diapir or as a dike [e.g., Emerman and Marret, 1990; Rubin, 1993], is a weakness in the present status of the diking model that has obscured the

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relationship between dike width and viscosity, particularly for felsic magmas [Petford *et al.*, 1993].

In this study, we use an exceptional geological area where melt migration undoubtedly occurred via diking, as seen in the field, and ultimately filled batholithic sized bodies, to examine theoretical models for magma ascent via diking as applied to granitic magmas. The region of interest is the High Himalaya range where highly dissected relief provides natural cross sections of great vertical extent enabling accurate three dimensional representation of the plutonic bodies [e.g., Scaillet *et al.*, 1995a]. These are the High Himalayan Leucogranites (HHL) which have been extensively studied in the recent years and whose petrogenesis is well known [e.g., Le Fort *et al.*, 1987; Castelli and Lombardo, 1988; Inger and Harris, 1993]. In particular, important parameters that control magma rheology such as temperature, weight percent H<sub>2</sub>O in melt, crystallinity, and melt composition of the magma at the time of its emplacement in the higher levels of the crust are now well constrained by experimental data [Scaillet *et al.*, 1995b]. Rather than using the empirical model of Shaw [1972] we have experimentally determined the melt viscosity of a High Himalayan leucogranite within P, T, weight percent H<sub>2</sub>O conditions relevant to its entire petrogenetic history (as inferred from phase equilibrium and petrographic studies). These viscosity measurements combined with the field data provide the first comparison of the theoretical relationship relating dike width to the magma viscosity [Petford *et al.*, 1993]. We show below that such a relation is entirely consistent with the emplacement of the HHL as near liquidus magmas and that the emplacement time required for complete assembly of the laccolith was only a few years in upper crust whose temperature was around 350°C.

### The Himalayan Leucogranite Laccoliths

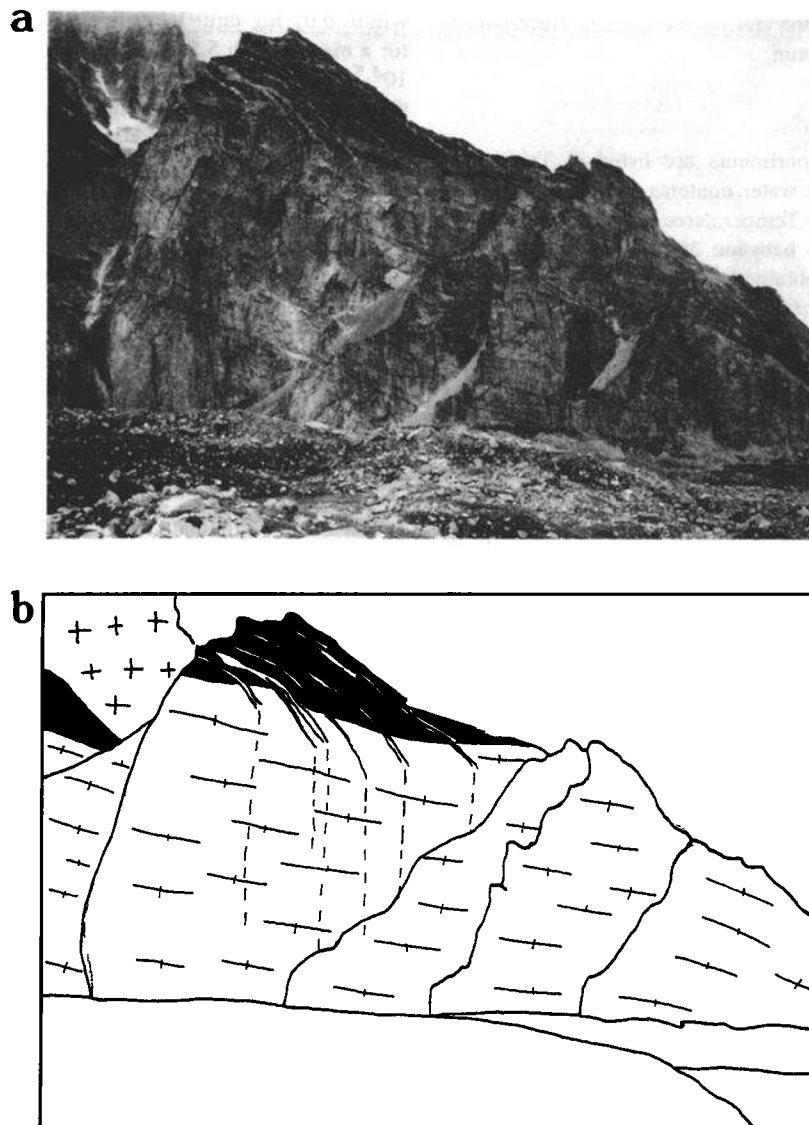
The HHL were produced during the collision between the Indian subcontinent and Eurasia which started circa 65 Ma [Beck *et al.*, 1995]. Geological mapping carried out over the past 30 years has shown that these leucogranites form discrete bodies cropping out regularly all along the 2000 km of extent of the Himalayan range. Each body has invariably a lens shape, being emplaced most of the time within metasedimentary rocks with clear intrusive contacts [e.g., Castelli and Lombardo, 1988; France-Lanord and Le Fort, 1988; Searle *et al.*, 1993; Inger and Harris, 1993; Scaillet *et al.*, 1995a]. In this report we will focus on the Garhwal Himalaya, where both the granite and its feeder system are exposed in the Gangotri region along vertical cliffs of more than 2500 m of relief. Detailed structural studies of this area have shown that magma emplacement took place during extensional tectonism in the Himalayan orogen contemporaneously with, or subsequent to, the crustal thickening process [Searle *et al.*, 1993; Scaillet *et al.*, 1995a]. The feeder system is made up of hundreds of dikes whose vertical extension is at least 1000 m, with thicknesses varying between 10 and 50 m, most of them intersecting the bottom contact of the laccolith (Figure 1). This general disposition led Scaillet *et al.* [1995a] to propose that the intrusive bodies represent genuine laccoliths, having grown by lifting up the overlying metasediment. The heterogeneity in Rb/Sr isotopes suggests that the whole body corresponds to the assembly of different magmas batches which did not undergo any subsequent vigorous mixing process [e.g., Deniel

*et al.*, 1987]. Owing to the general lack of intrusive contacts within the laccolith (i.e., between the different magma batches), the growing time of each individual laccolith must have been extremely short in order to prevent any significant cooling of each individual unit during the growth of the laccolith. Preliminary numerical thermal modeling shows that a 2-km-thick laccolith should cool below its solidus within 30,000 to 100,000 years, depending on the thermal regime of the crust and on the thermal diffusivity value chosen for the magma (B. Scaillet, manuscript in preparation, 1996). These calculations provide an upper bound for the assembly time of the laccoliths at the level of emplacement. Petrographic and experimental studies [Scaillet *et al.*, 1995b] have shown that these magmas were emplaced as crystal-poor melts (< 5 vol %), at temperatures ranging from 800° to 750°C (biotite-muscovite facies) to 750° to 700°C (tourmaline-muscovite facies), with melt water contents varying between 5.5-7 (Bt-Ms) and >7 (Tur-Ms) wt % H<sub>2</sub>O. The pressure of emplacement is constrained to have been between 300 and 400 MPa, on the basis of thermobarometric results obtained on associated thermal aureoles [Guillot *et al.*, 1993], while the pressure of magma generation was between 700 and 1000 MPa [e.g., Pêcher, 1989]. Thus the magma underwent a pressure drop of around 300-400 MPa during emplacement.

### Experimental Determination of Leucogranite Viscosities

#### Experimental Method

The experimental method used to determine the viscosity of hydrous melts is the falling sphere technique. In this study the same procedure as that used by Schulze *et al.* [1996] was followed, and the reader is referred to that work for full technical details. Only the salient features will be repeated here. A hydrous glass cylinder was prepared by melting the rock powder (natural tourmaline-muscovite bearing leucogranite GB4 of Scaillet *et al.* [1995b]) loaded within a Pt capsule together with an appropriate amount of distilled and deionized water. The melting was done at 300-400 MPa, at 1100°-1200°C for 3-4 days to ensure a homogeneous distribution of water. A glass cylinder free of bubbles approximately 4 cm long and 0.5 cm in diameter was obtained. The cylinder was sawed across so as to obtain a long (2-3 cm) and a short (0.5 cm) cylinder. The two cylinders were subsequently stuck together within a new Pt capsule (diameter 0.4 cm), with Pt powder in between. This Pt surface (thickness 1-2 µm) was used as a reference level to measure the position of the Pt spheres before and after the experiment. On the other end of the cylinder, a small layer of glass powder, obtained from crushed fragments of the same glass cylinder, was added. Two Pt spheres were placed within the powdered glass. The Pt capsule was finally welded shut and brought to the desired P-T conditions. After an experiment under a given set of P-T conditions, the next one was performed by using the same glass cylinder simply turned upside down to allow the spheres to settle back in the opposite direction. This procedure allowed us to use the same glass cylinder (with the same water content) for various P and T conditions (up to 10 experiments were made with the same cylinder). The water content of the glass cylinder was measured by Karl Fischer titration (KFT) before and/or after (on both ends of the glass cylinder) a given set of



**Figure 1.** (a) Photograph of the south face of the Shivling peak showing the base of a leucogranite laccolith with the array of feeder dikes that still connect to the bottom of the intrusion. The height of the cliff is about 1300 m. The leucogranite is intruded into black schist that overlies a thick layer of orthogneisses. In the latter, the dikes are nearly vertical, while they are tilted in the schist as a consequence of the late orogenic extensional collapse [see Scaillet *et al.*, 1995a]. (b) Schematic line drawing of the Figure 1a enhancing the geological contours of the main units. Only the upper reaches of the thicker dikes are drawn, when they cross the black schist level: crosses, leucogranite; horizontal crosses, orthogneisses. The intermediate level of schist is black.

experiments and was determined to be constant, within the analytical uncertainty of the KFT technique (0.3 wt % H<sub>2</sub>O). The falling distances of the Pt spheres were measured with an optical microscope equipped with an automated X-Y-stage enabling measurements to be performed with a precision of 0.001 cm. Given the fact that two Pt spheres were used, each experiment gave two independent measurements of the viscosity. The viscosity was calculated from Stokes' law, corrected for border effects (Faxen correction), using the run duration, the radius of the Pt sphere (determined to  $\pm 5 \mu\text{m}$  by optical and weighting methods) and the densities of the Pt sphere (21,450 kg/m<sup>3</sup>) and melts. In our experiments, calculated melt densities vary between 2290 kg/m<sup>3</sup> at 800°C and 6.66 wt % H<sub>2</sub>O and 2310 kg/m<sup>3</sup> at 1100°C and 3.98 wt %

H<sub>2</sub>O. For a fixed meltwater content and temperature this difference in melt density produces a variation of  $2.5 \times 10^{-4}$  log unit on the calculated viscosity. Therefore for convenience, a constant melt density of 2300 kg/m<sup>3</sup> was used in all viscosity determinations. All experiments were performed within an internally heated pressure vessel, pressurized with argon and working vertically. Total pressure is known within 50 bars. The temperature was read by at least three PtRh30 thermocouples and is known to within  $\pm 10^\circ\text{C}$  (including the gradient across the capsule [see Roux *et al.*, 1994]). The length of the glass cylinders allowed us to perform runs of long duration, between about 3 and 78 hours, with settling distances ranging from 0.2 to 1 cm, which virtually eliminates any error arising from the settling of Pt spheres prior to attainment of

experimental conditions and during the quench. Uncertainties for settling times are  $\pm 5$  min.

## Results

The results of 11 experiments are listed in Table 1 in chronological order. Melt water contents of 6.66 and 3.98 wt % have been investigated. Temperatures ranged from 800°C up to 1100°C and pressures between 300 and 800 MPa. In all experiments, viscosities obtained from the two Pt spheres are within 0.02 log unit of each other. Replicate experiments at around 860°C give viscosities that agree within 0.05 log units. Overall, the uncertainty in viscosity measurements is estimated to be better than 0.05 log units. Experiments performed at 860°C and 500 and 800 MPa with a meltwater content of 6.66 wt % H<sub>2</sub>O yield viscosities that are identical within error (Table 1), indicating that pressure has no detectable effect on hydrous melt viscosities within this range of pressure variation, as shown also by *Burnham* [1964] and *Schulze et al.* [1996]. The two sets of experiments display linear behavior in an Arrhenius plot (Figure 2), with correlation coefficients close to 1 (0.999 for 3.98 wt % H<sub>2</sub>O and 0.993 for 6.66 wt % H<sub>2</sub>O). The activation energy of viscous flow decreases slightly with increasing water content, as observed in the haplogranite system [*Schulze et al.*, 1996]. All the experimental data have been fit with the following equation:

$$\log \eta = 16280/T - 7.5461 + [0.59784 - 1235.4/T] \text{ wt \% H}_2\text{O} \quad (1)$$

where  $\eta$  is the viscosity in pascal seconds,  $T$  is the temperature in kelvins and weight percent H<sub>2</sub>O is the water content of the melt. This equation reproduces the experimental viscosities

**Table 1. Experimental Viscosities of Leucogranite Melts**

Run	P, MPa	T, °C	duration, s	log $\eta_1$ , Pa s	log $\eta_2$ , Pa s
<i>6.66 wt % H<sub>2</sub>O in Melt*</i>					
4	500	860	63420	3.58	3.57
5	500	910	18660	3.25	3.22
6	500	855	46860	3.55	3.53
7	500	800	72000	3.94	3.93
8	500	865	28680	3.54	3.54
9	800	860	45900	3.52	3.51
10	500	955	9000	3.00	2.99
11	650	813	75300	3.86	3.86
<i>3.98 wt % H<sub>2</sub>O in Melt<sup>§</sup></i>					
12	300	1100	14880	3.09	3.10
13	300	1001	35770	3.79	3.77
14	300	907	280300	4.45	4.45

P, pressure; T, temperature.

\*Pt spheres of radius 0.0110 cm and 0.0127 cm.

§Pt spheres of radius 0.0132 cm and 0.0136 cm.

within 0.01 log unit. It predicts a viscosity of  $10^{4.6-10^5}$  Pa s for a magma with 5.5 wt % H<sub>2</sub>O at 800°-750°C and of  $10^{4.1-10^{4.5}}$  Pa s for a magma with 7 wt % H<sub>2</sub>O at 750°-700°C. (Note that the Arrhenian approximation is strictly valid only for small temperature ranges. Use of equation (1) well outside the experimental range of calibration may give erroneous results.) These conditions correspond to those during the emplacement of biotite-muscovite and tourmaline-muscovite bearing leucogranites, respectively, as inferred from phase equilibrium experiments [*Scaillet et al.*, 1995b]. However, given the narrow range of these viscosity determinations and for the sake of simplicity, in the following section we will consider average values of 750°C for the temperature of emplacement and  $10^{4.5}$  Pa s for the magma viscosity.

## Implications for the Diking Mechanism

The results obtained in this study provide direct constraints on the relationships existing between dike width and magma viscosity. Dike width is an easily measured parameter in the field [e.g. *Wada*, 1994], and it has long been recognized, at least qualitatively, that the less viscous the magma the narrower the dike. *Petford et al.* [1993, 1994] used the mathematical analysis of *Bruce and Huppert* [1989, 1990] to derive a relation between dike width and magma viscosity:

$$w_c = 1.5[c(T_{sol}-T_w)^2/L(T_{liq}-T_{sol})]^{3/4}(\eta k H/g \Delta\rho)^{1/4} \quad (2)$$

where  $w_c$  is the critical dike width necessary to prevent its thermal lock-up during melt flow,  $T_{liq}$  is the magma initial temperature (750°C), and  $T_{sol}$  is the temperature at which the magma near the dike wall becomes immobile, here taken to be that of the solidus (645°C). Strictly, the temperature at which magma stop flowing differs from that of the solidus but leucogranite magmas have a strong eutectic-like behavior [*Scaillet et al.*, 1996], being 80% liquid 15°C above the solidus, and this difference should not exceed 10°C. The specific heat,  $c$  (1600 J/kg °C), was calculated using the bulk composition and the model of *Lange and Navrotsky* [1992] for all oxides except for water whose partial molar specific heat was taken from *Clemens and Navrotsky* [1986]. The density difference between the magma and the host rock,  $\Delta\rho$ , was calculated with an averaged weighted mean density of 2700 kg/m<sup>3</sup> for the country rocks [e.g., *Corry*, 1988] and a melt density calculated using the bulk composition and partial molar volumes of *Knoche et al.* [1995] for all oxide components except for FeO and H<sub>2</sub>O whose partial molar volumes were taken from *Lange and Carmichael* [1990] and *Holtz et al.* [1995], respectively. The thermal diffusivity  $k$  was set to  $8 \times 10^{-7}$  m<sup>2</sup>/s, and  $g$ , the gravitational constant, was set to 10 m/s<sup>2</sup>. All the previous input parameters of (2) can be considered as well constrained. In contrast,  $T_w$ , the far-field wall rock temperature,  $L$ , the latent heat of crystallization, and  $H$ , the dike length, are more difficult to assess. A minimum value for  $H$  is the present vertical extent of dikes (1000 m), while a maximum value is given by the difference between the depth of magma generation, which is estimated to be between 7 and 10 kbar, and its level of emplacement. Varying  $H$  between 1000 and 20,000 m, keeping all other values constant (with  $L = 3 \times 10^5$  J/kg and  $T_w = 300^\circ\text{C}$ , see next section), increases the critical dike width from about 1 to 2 m, which shows that this parameter has little influence on the estimate of  $w_c$ . Although reasonable, these thicknesses are ~ 1 order of

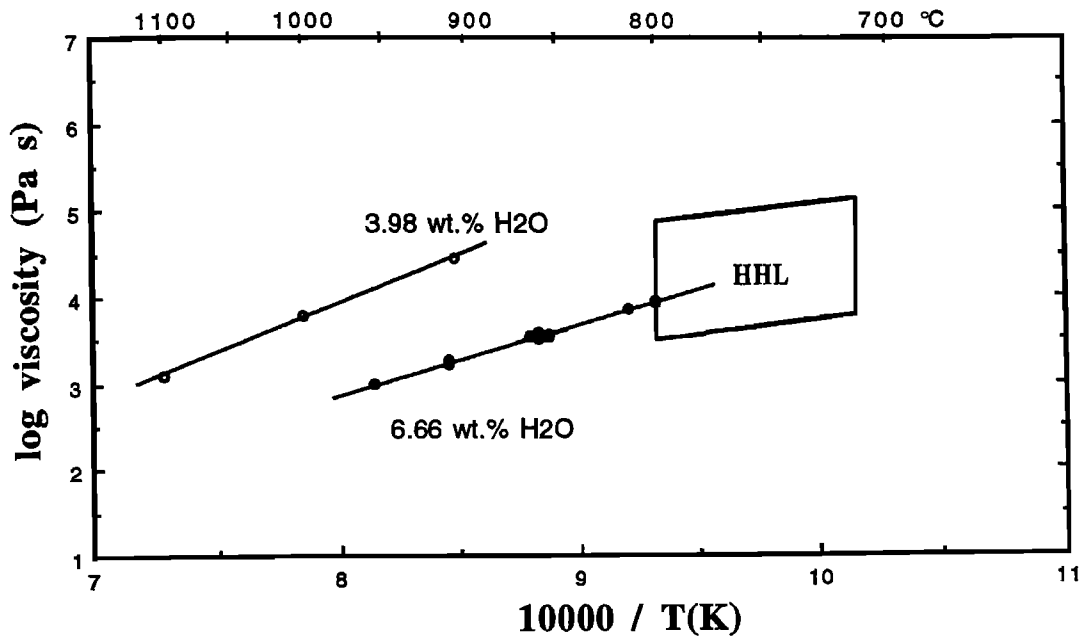


Figure 2. Arrhenius plot of experimental viscosities determined for a High Himalayan leucogranite at two meltwater contents. The box indicates the range of magma viscosities during the emplacement of Himalayan leucogranites, as inferred from phase equilibria and petrographical studies [Scaillet *et al.*, 1990, 1995b] and from the present work.

magnitude lower than those observed (10-20 m, or even more). However, this first calculation is based on the premise that the latent heat is being released uniformly across the crystallization interval. In fact, crystallization paths obtained from phase equilibrium diagrams show that such leucogranites are still 70-80 wt % liquid 15°C above their solidus [Scaillet *et al.*, 1996]. In other words, the contribution of latent heat to the heat budget of the magma during its crystallization is

highly non linear, being minimal at near-liquidus conditions, and a value of  $3 \times 10^5$  J/kg greatly overestimates the amount of heat released by crystallization during dike intrusion. The effect of latent heat on the critical dike width is illustrated on Figure 3 for a magma having a viscosity of  $10^{4.5}$  Pa s at 750°C emplaced in a host rock at 300°C. Also shown is the approximate amount of accumulated heat released by the crystallization of a leucogranite magma between 750° and

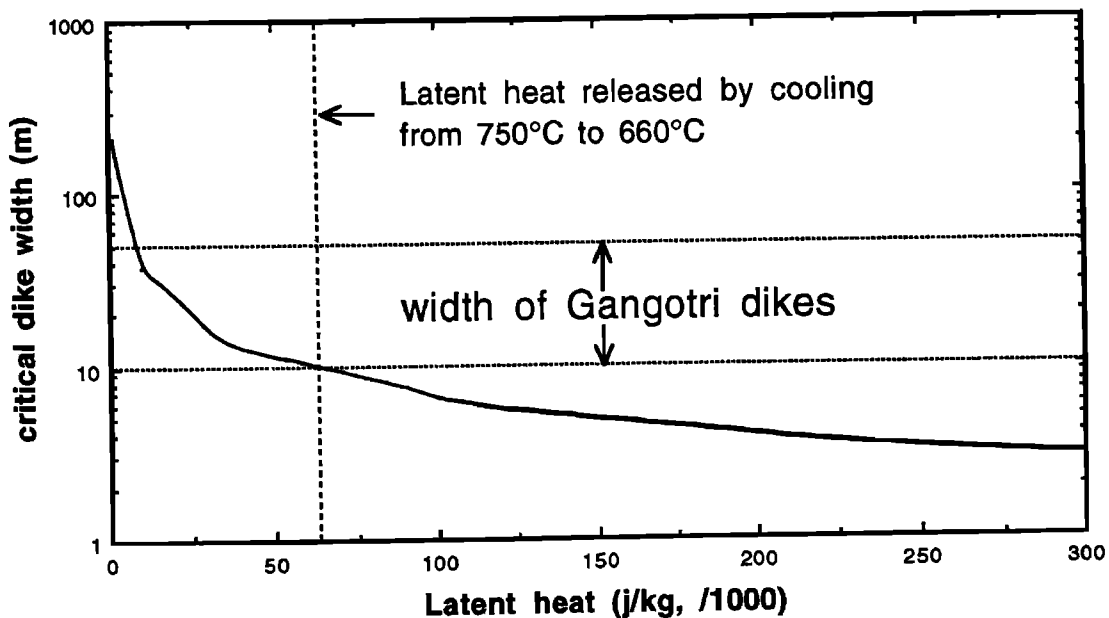


Figure 3. Effect of latent heat of crystallization on the critical dike width for a leucogranite magma at 750°C with a viscosity of  $10^{4.5}$  Pa s. Also shown is the range of widths for the Gangotri dikes that are clearly connected to the base of the laccolith (Figure 1). The approximate amount of cumulated latent heat released during the crystallization of a leucogranite magma between 750° and 660°C is reported.

660°C, as estimated from phase equilibrium experiments [Scaillet *et al.*, 1995b]. Clearly, the preserved dike thicknesses indicate that the amount of latent heat released during dike intrusion was small, probably below  $1 \times 10^5$  J/kg. The dike thicknesses are thus consistent with a magma emplacement at near-liquidus conditions, as shown by petrographic and experimental studies [Scaillet *et al.*, 1990, 1995b; Inger and Harris, 1993]. They also fit with the overall trend established by Wada [1994] between dike thickness and magma viscosity, falling midway (10–50 m,  $10^4$ – $10^5$  Pa s) between those reported for mafic (1 m,  $10^1$ – $10^2$  Pa s) and viscous felsic (100 m,  $10^6$ – $10^7$  Pa s) magmas. The combined use of experimental and field data supports the theoretical approach followed by Petford *et al.* [1993, 1994] and shows that, in this particular case, dike width is a precise indicator of magma viscosity, provided that other controlling factors such as the latent heat budget during crystallization or the host rock temperature (see below) are also well constrained.

### Implications for Magma Emplacement Time

The minimum time required to assemble the magma chamber can be obtained by using the equation of fluid flow in a tabular conduit that approximates the dike geometry. The general form of this equation is [e.g., Petford *et al.*, 1994]

$$V_{ave} = g\Delta\rho w^2 / 12\eta \quad (3)$$

where  $V_{ave}$  is the average velocity of magma flow and  $w$  is the dike width. Taking a minimum value of 10 m for the dike width, the flow velocity varies between 2.6 m/s and 0.3 m/s for a magma with a viscosity  $\eta$  of  $10^{4.1}$  and  $10^{5.01}$  Pa s, respectively. The horizontal length of the dikes is not precisely known but should be no more than a few hundred meters. A single dike with 300 m of horizontal extension could therefore fill a  $150 \text{ km}^3$  magma chamber (the maximum volume of a laccolith in the Gangotri area) in less than 2 years.

As there are probably more than a hundred dikes, the time of chamber building can be considered as instantaneous if all dikes were active together. Thus the 2-km laccolith assembly time should lie between 1 and 100,000 years, the maximum cooling time to attain solidus conditions as obtained from numerical simulations (B. Scaillet, manuscript in preparation, 1996). However, a time at the lower end of this range is presently favored because of the near constancy of dike widths, as preserved now in the field. In fact, equation (2) predicts that the critical dike width will decrease with increasing host rock temperature. This is what is shown on Figure 4, where the relation between critical dike width and host rock temperature is shown for a magma at 750°C and for different values of latent heat. The widths of the Gangotri dikes are reproduced at host rock temperatures below 400°C, with the most reasonable preintrusion temperatures being located at around 350°C. Had the process of magma incoming been protracted (e.g., hundreds of thousands of years), then the host rocks would have had time to warm up, which should have significantly decreased the width of the latest dikes intruded. For instance, with a value of  $T_w = 600^\circ\text{C}$  the critical dike width drops to around 0.5 m (i.e., almost 2 orders of magnitude below observed thicknesses). Significantly, a temperature around 350°C is that expected at a depth of 15 km in a continental crust having a normal geothermal gradient. If the preintrusion temperature was 500°C at 15 km depth, then temperatures of around 1000°C would prevail at 30 km of burial, conflicting with existing thermal models of thickened continental crust [e.g., England *et al.*, 1992], which show that even when the thermal relaxation process is completed (i.e., ~ 60 Myr after the stacking of the crust), temperatures at such depths (in the hanging wall of the thrust sequence) hardly exceed 700°C. In addition, numerical simulations of magma cooling (B. Scaillet, manuscript in preparation, 1996) reproduce the observed metamorphic peak temperatures of around 550°C in the contact aureole only for preintrusion temperatures below 400°C. Temperatures in excess of this value yield much higher metamorphic grades in the aureole, in some cases reaching

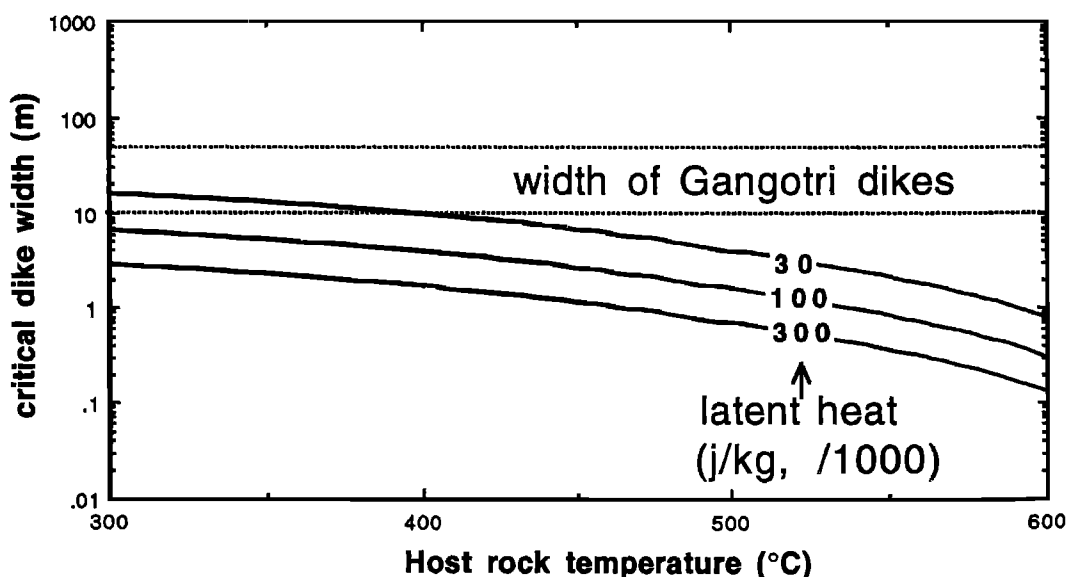


Figure 4. Effect of host rock temperature on the critical dike width for a leucogranite magma at 750°C, with a viscosity of  $10^{4.5}$  Pa s and for three values of latent heat:  $0.3 \times 10^5$ ,  $1 \times 10^5$ , and  $3 \times 10^5$  J/kg. Also shown is the range of widths for the Gangotri dikes that are clearly connected to the base of the laccolith (Figure 1).

conditions of incipient partial melting ( $> 650^{\circ}\text{C}$ ). Such features are lacking in the Gangotri region, and we therefore conclude that magma intrusion in this area took place in an already cold upper crust [see also *Copeland et al.*, 1990], the emplacement process being achieved in an extremely short period, a fact endemic to most laccolithic intrusions [*Corry*, 1988]. This cold environment is in agreement with the hypothesis of *Scaillet et al.* [1995a] of the magma fractures being arrested by collapse folds in the upper crust at a level close to the brittle to ductile transition [see also *Hogan and Gilbert*, 1995].

### Implications of High Rates of Magma Flow

The rates of magma flow suggested by this study are among the highest ever calculated for silicic magmas. These high rates have two major implications. First, such rates of magma flow should minimize, or even impede, any chemical interaction with the country rocks encountered during uprise [*Clemens and Mawer*, 1992]. Thus the magma composition at its arrival level is probably close to that produced by the melting reaction, just before it left the source zone. It follows that these leucogranites can be possibly used as direct probes of the rheological properties of the melt in the melting source, owing to the lack of pressure dependence on melt viscosity. Thus it can be concluded that melt viscosity was around  $10^{4.5}$  Pa s in the source area. The HHL are inferred to have been produced by a melting reaction involving the breakdown of muscovite with melt fractions in the source of about 10-15 % [e.g., *Le Fort et al.*, 1987; *Harris and Inger*, 1992]. The rate of melt segregation of a partially molten rock through a compaction mechanism can be evaluated with the physical model of *McKenzie* [1984]. Calculations done with a melt viscosity of  $10^{4.5}$  Pa s and a porosity of 10% give compaction times of the order of  $10^5$  years [*Wickham*, 1987; *Laporte*, 1994]. Therefore, in view of the extremely short timescales during which magma ascent and emplacement occur, this result indicates that compaction alone is probably not an efficient process for melt segregation in the continental crust [see *Wickham*, 1987].

The second implication concerns the chemical heterogeneity (e.g., Sr isotopes) of leucogranitic magmas. The fluid dynamic regime of a magma flowing in a dike can be assessed through the Reynolds number which is defined as [e.g., *Jaupart and Allègre*, 1991]

$$Re = \rho V_{ave} w / \eta \quad (4)$$

Taking average values of  $10^{4.5}$  Pa s and 20 m for magma viscosity and dike width, respectively, the computed Reynolds number is 5, which is well below the critical value of 2000 beyond which the onset of turbulent flow is predicted. Increasing the dike width up to the maximum observed in the field (50 m) and taking the lowest magma viscosity found in this study ( $10^{4.1}$  Pa s) gives a Reynolds number of 603. Therefore the dynamical regime of the magma flowing through the feeding system of the Gangotri laccolith can be predicted to be dominantly laminar. The lack of turbulence means that the homogenization of an heterogeneous batch of magma during its uprise proceeds mainly through chemical diffusion. The very slow rate of cationic diffusion in silicate melts compared to the rate of magma ascent found in this study

indicates that virtually any chemical heterogeneity inherited from the source will survive the ascent period. Therefore the chemical heterogeneity found at the outcrop scale does not necessarily imply the coexistence of different magma pulses having travelled in separate dikes, but it may also reflect the existence of a former single magma batch that was already heterogeneous in the source region. Besides the implications that such finding may have on the mechanisms and kinetics of partial melting processes occurring in the crust (i.e., equilibrium or disequilibrium melting, rate of melt extraction, see [*Brown et al.*, 1995]) and which are beyond the scope of this paper, a first consequence is that it explains the existence of strong variations in some isotopes (e.g., Sr) while physical boundaries testifying of the coexistence of several magma batches are lacking [*Deniel et al.*, 1987].

### Concluding Remarks

The High Himalayan leucogranites are a most spectacular example of the assembly of large granitic batholiths via diking on very short timescales (i.e., of the order of years). This is not to say that every single granitic pluton or large batholith behaves in the way exemplified above. The HHL are derived from a crustal melting process alone with no mantle input [*Le Fort et al.*, 1987]. In this respect they differ from most of the granites that belong to coastal batholiths such as the Sierra Nevada Batholith in California. In the latter, the mantle component represents a significant fraction, if not a dominant one, of the intrusive rocks. A consequence of this mantle involvement is that Cordilleran granites span a wider compositional range than do the HHL. This compositional diversity implies more complicated petrogenetic stories which in turn open the possibility for having more complex rheological behaviors (i.e., different crystal/melt ratios during extraction, ascent and emplacement). Finally, it needs be stressed that the volume of magmatic material involved in these orogenic belts (from 20% up to 90%, [*Paterson and Fowler*, 1993]) is much more important than that in the Himalayan range, where exposed plutons represent no more than 2% of this intracontinental belt [*Le Fort et al.*, 1987]. A major consequence of this 1 order of magnitude of difference in the volume of intrusive rocks is the heat budget of the orogenic belt. Repeated injections of magmatic bodies over protracted periods associated with large and sustained heat flux at the base of the crust may affect the rheological behavior of the medium through which granitic magmas ascend. In this context, first emplaced magmas may have emplacement mechanism(s) different(s) from latter ones [see *Paterson and Fowler*, 1993]. For instance, it can be envisaged that first injected magmas will traverse a cold crust through dikes, while latter ones will encounter a hotter medium, be it a heated crust or a fully or partially crystallized intrusion. In this case the viscosity contrast between the intruding magma and its hosting rocks may be more favourable (i.e., lower) to a diapiric rise. In comparison, the Gangotri leucogranites represent a short-lived magmatic event with limited thermal effects on the traversed terranes. At a larger scale the spacing between the different major plutons (100-200 km) belonging to HHL belt also precludes thermal interferences between them. Thus, in contrast with what probably occurs in magmatic arcs, the rheological properties of the crust were not significantly affected during the injection of the Gangotri magma (note that



this may not be entirely true for the largest HHL plutons such as the Everest-Makalu or Manaslu granites). In summary, although the Gangotri leucogranites validate diking as a viable mode of ascent for silicic magmas, they do not exclude other mechanisms of granite ascent and emplacement.

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