



Granitic melt viscosity and dike formation

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Abstract—Melt viscosity is an important variable controlling the rates of granitic melt separation and transport from its source region. Viscosities of granitic melts are significantly affected by changes in temperature and volatile content. Addition of 2 wt% H₂O to anhydrous peraluminous and metaluminous granitic melts reduces viscosity by six orders of magnitude at 800°C which is equivalent to the effect of raising the temperature by 500°C. Addition of 10 wt% more H₂O only decreases melt viscosity by approximately 1.5 additional orders of magnitude. At higher temperatures the effect of H₂O addition on melt viscosity is less significant, at lower temperatures more significant. The effect of pressure on hydrous granitic melt viscosities at crustal conditions is insignificant in comparison. Between 0.5 and 1.5 GPa the viscosity at 1000°C increases by 0.4 log units, or less; this effect can be ignored in most petrological calculations. Using newly acquired viscosity data at low H₂O contents, the Baker model for the calculation of metaluminous and peraluminous granitic melt viscosities has been updated and expanded. The new model can predict hydrous (0.3–12.3 wt% H₂O), granitic (69–77 wt% SiO₂) melt viscosities between 577 and 1200°C at crustal pressures. The new granitic melt viscosities have been used to investigate the possibility of granitic melt transport via diking. Diking can be an efficient mechanism of damp granitic melt transport through the crust for granitic melts containing a minimum of 2 wt% H₂O if the lower crust has been sufficiently heated. Modeling the thermal history of basaltic intrusions and surrounding lower crust demonstrates that a sheet-like intrusion 5 km thick, or a suite of intrusions with the same aggregate thickness that were rapidly emplaced, can provide enough heat to allow granitic magmas to migrate through much of the crust in dikes. The generation of granitic dikes associated with thinner intrusions appears unlikely. © 1998 Elsevier Science Ltd. All rights reserved

INTRODUCTION

Silica-rich partial melts form in the deep crust, separate from residual, refractory crystals, and ascend to higher levels where they are emplaced as granites or erupt as rhyolites. This has been the consensus of igneous petrologists based upon decades of field, analytical and experimental research (Carmichael *et al.*, 1974; Clarke, 1992). However, the mechanisms responsible for the separation and transport of granitic composition melts and the rates at which these processes operate remain objects of active debate and investigation. Is gravitational compaction the dominant mechanism of melt extraction (McKenzie, 1984, 1985), or is it controlled by structures and tectonic forces (Sawyer, 1991, 1993)? Once melt is separated from its residuum, is it transported to higher levels by diapirs (Mahon *et al.*, 1988) or by dikes (Clemens and Mawer, 1992; Petford *et al.*, 1993)? Are the rates at which these processes operate geologically slow (millions of years), fast (tens of thousands of years) or instantaneous (less than 1000 years)?

One of the most influential variables in equations describing melt segregation and transport is viscosity. In virtually all equations for the separation and transport of melts viscosity appears as a retarding factor (McKenzie, 1985; Mahon *et al.*, 1988; Petford *et al.*, 1993; Rubin, 1995). Because granitic melt viscosities can vary by many orders of magnitude they play an important role in controlling the rates of separation

and transport. In some cases, viscosity may control the type of mechanism responsible for melt extraction and transport (Rubin, 1995). Knowledge of granitic melt viscosities is essential to our understanding of the processes involved in the extraction and transport of granitic magmas.

The dominant mechanism of granitic melt transport through the crust is now generally accepted to be diking (Clemens and Mawer, 1992; Petford *et al.*, 1993). However, Rubin's (1995) numerical study of the physics of dike initiation and propagation found that long dikes filled with melts of granitic composition were unlikely to form because melts would freeze within a short distance of their source. These calculations are strongly dependent upon melt viscosity. Rubin calculated viscosities using the model of Shaw (1972) which, as discussed below, is inaccurate for granitic melts with low H₂O concentrations. Thus, it was deemed necessary to reinvestigate Rubin's model using new, accurate calculations of granitic melt viscosities.

This contribution begins with a brief review of granitic melt viscosities and the variables which affect them. An improved version of the Baker (1996) model for the calculation of granitic melt viscosities is then presented and used to reinvestigate Rubin's (1995) model for dike propagation. This reinvestigation demonstrates that diking can, at least in some circumstances, provide an effective means of granitic melt transport through the crust.

GRANITIC MELT VISCOSITIES

Numerous studies of melt viscosities have demonstrated the importance of temperature on the viscosity of silicate melts (Richet and Bottinga, 1995). In a similar manner to most substances, the viscosity of silicate melts decreases as temperature increases. Depending upon melt composition, a change in temperature of 100 K can result in more than an order of magnitude change in viscosity (Fig. 1). Over short temperature intervals (a few hundreds of degrees) the viscosity of most silicate melts can be described by an Arrhenius relationship:

$$\ln \eta = \ln \eta_0 + \frac{E_a}{RT} \quad (1)$$

where η is the viscosity, η_0 a constant, E_a the activation energy, R the gas constant and T the temperature in K. Over larger temperature intervals viscosities of melts and glasses do not follow the linear Arrhenius relationship, but instead form curves which can be described by (Richet and Bottinga, 1995):

$$\ln \eta = \ln A_e + \frac{B_e}{TS_{\text{conf}}} \quad (2)$$

In this case A_e is also a constant and B_e is an energy term similar to the activation energy in the Arrhenius equation. The inclusion of the configuration entropy, S_{conf} , in the denominator of the second term of equation (2) reflects the entropy of the melt or glass due to different configurational states, more of which become accessible at higher temperatures (Richet and Bottinga, 1995). This entropy term can be determined from viscosity measurements or, in some cases, independently by calorimetric techniques (Richet and

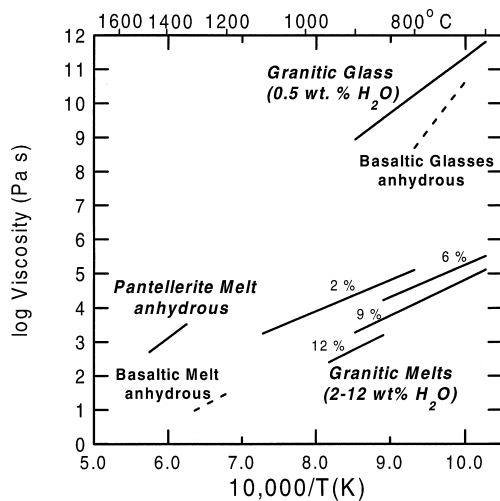


Fig. 1. Viscosities of some basaltic and granitic melts and glasses. Basalt melt data are from Shaw (1969) and basalt glass viscosities are from Taniguchi (1993). Hydrous granitic melt viscosities are from Shaw (1963), 6 wt% H_2O , Burnham (1964), 9 wt% H_2O , and Persikov (1991), 2 and 12 wt% H_2O . Anhydrous pantellerite (peralkaline granite) viscosities are from Scarfe (1977) and granitic glass viscosities are from Murase and McBirney (1973).

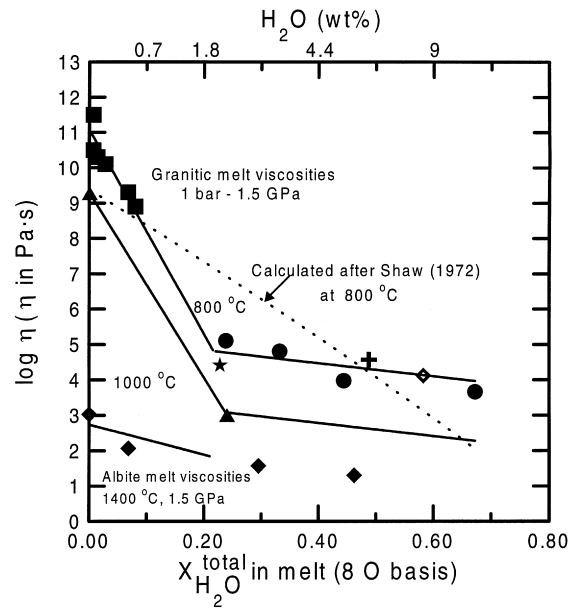


Fig. 2. Viscosities of granitic melts used to calibrate revised version of the Baker (1996) model for hydrous granitic melts. Hydrous granitic melt viscosity data at 800°C (squares, filled circles, crosses, stars and open diamond) and albite melt viscosity data at 1400°C (filled diamonds) are from the compilation of Baker (1996). Granitic melt viscosities at 1000°C (triangles) are from Baker *et al.* (submitted), $X_{\text{H}_2\text{O}} \sim 0.23$, and extrapolated from Hess *et al.* (1995), $X_{\text{H}_2\text{O}} \sim 0.0$. Errors in measured viscosities are typically smaller than the size of the data points.

Bottinga, 1995). Although equation (2) is the preferred expression for silicate melt viscosities, in many cases measured viscosities can be accurately represented by equation (1), or some variation of it, over the relatively small range of magmatic temperatures for granitic melts (Fig. 1).

The composition of silicate melts has a strong effect on viscosity (Shaw, 1963, 1965, 1969, 1972; Richet and Bottinga, 1995). At equivalent temperatures the viscosities of anhydrous basaltic melts or glasses are 1–5 orders of magnitude below the viscosities of anhydrous granitic melts or glasses (Fig. 1). However, the viscosities of peraluminous and metaluminous magmatic silicate melts with SiO_2 concentrations greater than approximately 69 wt% appear to be within 0.5 log units of each other at every temperature studied (see discussions in Baker, 1996; Hess and Dingwell, 1996). On the other hand, the viscosities of peralkaline granitic melts containing both F and H_2O are significantly less than those of peraluminous and metaluminous granitic melts with equivalent concentrations of F and H_2O (Baker and Vaillancourt, 1995).

The effect of H_2O on granitic melt viscosities is clearly displayed in Figs 1 and 2. At 800°C the addition of 2 wt% H_2O , or a mole fraction of ~ 0.25 as calculated following Burnham's model (1979) for a melt with a molecular weight of 260 g, results in a six orders of magnitude decrease in granitic melt viscosities. Such a viscosity reduction is equivalent to increasing the temperature of a granitic melt with 0.5

wt% H₂O by approximately 500°C (cf. Fig. 1). The viscosity reducing effect of H₂O on a silicate melt is attributed to the presence of dissolved OH species which are formed by the interaction of H₂O and bridging oxygens between (Si, Al)O₄ tetrahedra to form two non-bridging oxygens with associated hydrogens (Stolper, 1982) which results in reduced melt viscosity. It must be noted, however, that Kohn *et al.* (1989) suggested addition of H₂O only weakens, not destroys, bridging oxygen bonds. At low H₂O contents the OH molecule is the dominant species of H₂O dissolved in the silicate melt and the addition of small amounts of H₂O result in large viscosity decreases. At total H₂O concentrations greater than ~2 wt%, molecular H₂O becomes the dominant species (Stolper, 1982; Silver and Stolper, 1989; Silver *et al.*, 1990). The interaction between the molecular H₂O species and the silicate melt is minor; because of this, the effects of H₂O addition on melt viscosity at H₂O concentrations above ~2 wt% are insignificant compared to the effects at lower H₂O contents (Fig. 2). Viscosities measured at 800, 1000 and 1400°C (Fig. 2) demonstrate that H₂O is increasingly effective at viscosity reduction as temperature is lowered.

The importance of pressure on melt viscosity is minimal at crustal pressures of 0.1 MPa–1.5 GPa (Baker *et al.*, submitted). Viscosity studies at both high, 6–9 wt% (Burnham, 1964; Schulze *et al.*, 1996), and low, 2–3 wt%, H₂O contents (Dingwell, 1987; Baker *et al.*, submitted) demonstrate that the maximum effect of pressure upon hydrous granitic melt viscosity is approximately 0.4 log units. The magnitude of the pressure effect is sufficiently small that it can be ignored in most calculations and replaced by an average viscosity for all crustal pressures.

Because of the impossibility of measuring the viscosity of all granitic melts at every condition of temperature, pressure and H₂O concentration, a model must be built for their prediction. The first such model was that of Shaw (1972) which has been used for the past 25 years. However, as demonstrated in Fig. 2, the model of Shaw fails to accurately predict the viscosity of many damp, less than 4 wt% H₂O, granitic magmas (Baker and Vaillancourt, 1995). In recent years several research groups have published new models for the calculation of hydrous peraluminous to metaluminous granitic melt viscosities (Persikov, 1991; Baker, 1996; Hess and Dingwell, 1996; Scaillet *et al.*, 1996; Schulze *et al.*, 1996). During the writing of this contribution it was discovered that the viscosity calculations of Baker (1996) were inconsistent with the viscosity measurements of Baker *et al.* (submitted). Because of the inconsistency it became necessary to update the model of Baker (1996) for the calculation of granitic melt viscosities.

Baker (1996) found that when the log of the viscosities were plotted as a function of the mole fraction of H₂O in the melt, the viscosity data formed two linear

arrays intersecting at approximately total $X_{\text{H}_2\text{O}}=0.3$. Based on these results an empirical model was proposed that successfully predicted hydrous, 0.3–12.3 wt% H₂O, granitic melt viscosities at crustal pressures and temperatures between 700 and 900°C (Baker, 1996). This model is applicable to peraluminous and metaluminous granitic melts, but cannot be used to calculate the viscosity of peralkaline granitic melts. New experimental results of Baker *et al.* (submitted) combined with those at 800°C for granitic melts, at 1400°C for albite melts and with the viscosity of an anhydrous haplogranitic melt at 1000°C, 1.87×10^9 Pas, which was extrapolated from measurements of Hess *et al.* (1995) at temperatures between 882 and 939°C, have been used to recalibrate the empirical model of Baker (1996). The calibration for granitic melt viscosities with $X_{\text{H}_2\text{O}} > 0.24$ was also updated by calculating the slope based upon fitting 800°C viscosities of melts with $X_{\text{H}_2\text{O}} \geq 0.21$, instead of $X_{\text{H}_2\text{O}} > 0.3$ as done by Baker (1996). The effect of temperature on the intercept of the line was assumed to be the same as found by Baker (1996). The model is based upon 19 viscosity measurements, 17 of which are from Baker (1996). The resulting new expressions for the calculation of granitic melt viscosities are for $X_{\text{H}_2\text{O}} \leq 0.24$:

$$\log \eta = (5.85458 \times 10^{-5} T^2 - 0.119564 T + 31.926) X_{\text{H}_2\text{O}} - 0.0142778 T + 26.8446 \quad (3)$$

and for $X_{\text{H}_2\text{O}} > 0.24$:

$$\log \eta = -1.86185 X_{\text{H}_2\text{O}} - 0.009511 T + 15.6293, \quad (4)$$

where the mole fraction of H₂O is calculated with a molecular weight of the anhydrous granitic melt equal to 260 g (i.e. an 8 oxygen basis, Burnham, 1979), T is the temperature in K and the viscosity is given in Pa s. The number of digits in these expressions is not reflective of the precision of the calculations but is provided to minimize rounding errors.

An independent test of the viscosity model was performed by comparing calculated viscosities with 71 viscosities of granitic and haplogranitic melts measured at temperatures between 577 and 1200°C, and H₂O contents between 0.02 wt% (0.003 mole fraction) and 12.3 wt% (0.67 mole fraction) selected from Hess and Dingwell (1996). None of these measurements were used for calibration of the model. The model demonstrates excellent agreement with measured viscosities (Fig. 3); the average deviation (i.e. the average of the absolute value of the difference between calculated and measured viscosities) is 0.64 log units. The greatest discrepancies between calculated and measured values are found for H₂O-rich, $X_{\text{H}_2\text{O}} > 0.24$, low-viscosity, high-temperature, 1000°C and greater (i.e. outside of the normal conditions of granite petrogenesis), melts which show a maximum deviation of 2.25 log units (Fig. 3). These discrepancies are most probably due to compositional and pressure effects which are not con-

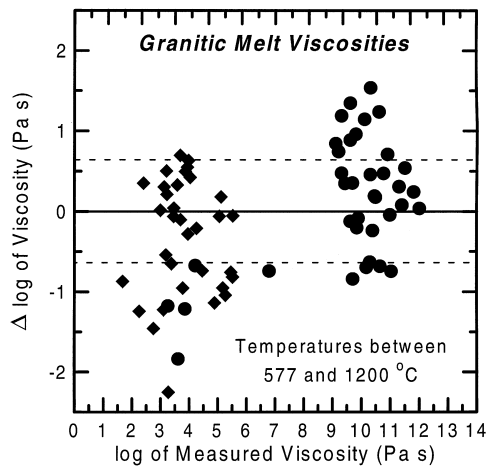


Fig. 3. Comparison of the difference between calculated (using the model in this paper) and measured viscosities. Filled circles are melts with $X_{\text{H}_2\text{O}} \leq 0.24$ and diamonds are melts with $X_{\text{H}_2\text{O}} > 0.24$. Errors are similar to the size of the symbols. The dashed lines represent the average deviation between calculated and measured viscosities.

sidered in this model, or in any of the other recent models for the calculation of granitic melt viscosities (Hess and Dingwell, 1996; Scaillet *et al.*, 1996; Schulze *et al.*, 1996).

The average deviation of this updated model can be compared with the same values for the viscosity model formulated by Hess and Dingwell (1996). Applying the Hess and Dingwell model to the same suite of melts yields an average deviation of 0.34 log units. Although this value is significantly lower than that of this study it is important to note that Hess and Dingwell (1996) formulated their model using all of the melt viscosities available and that their average deviation only reflects the accuracy of their fit to the data. In contrast, the average deviation between the updated model of Baker (1996) and measured viscosities was determined with an independent test using data that were not included in the construction of the model. This test clearly indicates that the updated model of Baker (1996) presented in the paper can be extrapolated to virtually any granitic melt composition between 69 and 77 wt% SiO_2 and is sufficiently accurate to provide viscosity estimates of hydrous metaluminous and peraluminous granitic melts for use in petrogenetic modeling of granitic melt separation from source regions and transport through the crust at temperatures between 577 and 1200°C.

DIKE TRANSPORT OF GRANITIC MELTS OUT OF THEIR SOURCE REGION

Rubin (1995) presented a comprehensive model for the prediction of dike formation and dike length based on the analysis of thermal and mechanical constraints. He identified the following controlling variables for

dike propagation: melt viscosity, η , latent heat of melting, L , thermal diffusivity, κ , heat capacity, c , elastic stiffness, G , excess source pressure, ΔP , the host rock temperature gradient at the contact with the magma source, dT/dx , and the temperature of the magma above its rheological solidus (where the magma no longer behaves as a liquid suspension, but as a solid), $\Delta T = T_1 - T_{\text{RS}}$, where T_1 and T_{RS} are the temperature at the liquidus and the temperature at the rheological solidus. For dikes to propagate away from their source region (i.e. a magma chamber) there must exist a substantial distance, l_0^{min} , of country rocks which are hotter than T_{RS} . Rubin (1995) found that this minimum distance can be calculated:

$$l_0^{\text{min}} = \frac{1}{0.27} \sqrt{\frac{c\Delta T/(L + c\Delta T)}{1 + \sqrt{(L + c\Delta T)/c\Delta T}} \frac{\sqrt{3\eta\kappa}}{(\Delta PG)^2 \sqrt{\Delta P}}}. \quad (5)$$

Only if the zones of country rock with temperatures above T_{RS} are wider than l_0^{min} can dikes propagate significantly beyond the point where the country rocks' temperature falls below T_{RS} . Rubin also found a dimensionless freezing parameter independent of l_0^{min} , β , which relates the rate of magma freezing due to heat loss at the tip of the dike to the rate at which hot magma is supplied from the magma source. Only if β is less than a critical value of ~ 0.15 could transport of magma by diking be possible because hot magma can be supplied to the propagating dike faster than it freezes at the dike tip. The value of β can be calculated exactly for eutectic magmas intruded at their freezing point. For magmas intruded at temperatures above their freezing point, or rheological solidus, such that $\Delta T > 0$, Rubin presented an expression for the modified value of β which must be less than ~ 0.15 for significant dike propagation to occur, $\beta_{\Delta T}$:

$$\beta_{\Delta T} = \frac{2}{\sqrt{\pi}} \frac{c|dT/dx|\sqrt{3\eta\pi}}{L(\Delta PG)^2 \sqrt{\Delta P}} \frac{L}{L + (4/\pi)c\Delta T}. \quad (6)$$

Rubin's (1995) numerical solutions to his equations demonstrated that the dike propagation distance was a function of $\Delta T/|dT/dx|$ and $\beta_{\Delta T}/0.15$. His numerical results relating dike length to $\beta_{\Delta T}/0.15$ are plotted in his Fig. 3 which displays propagation distance divided by twice $\Delta T/|dT/dx|$, not $\Delta T/|dT/dx|$ as in the original publication (Rubin, personal communication 1996). Combining his calculations with the viscosities of Shaw (1972), Rubin found that for granitic melts l_0^{min} was of the order of 10 km and $\beta_{\Delta T}$ far exceeded the critical value of 0.15. The values of each of these parameters demonstrated that transport of granitic melts by diking was not favoured. However, the calculations of Shaw (1972) overestimate the viscosities of damp granitic melts, and a reinvestigation of the propagation of dikes from granitic source regions using new viscosities leads to different conclusions.

Using the 'typical values' from Rubin (1995) for all properties of granitic melts and crustal rocks except for melt viscosities and host rock thermal gradients, the values of l_0^{\min} and estimated dike lengths were calculated for granitic melts containing 2 or 4 wt% H₂O, which span the range of typical H₂O concentrations in damp granitic magmas (Clemens, 1984). Rubin (1995) considered the latent heat to be 300 kJ kg⁻¹, the thermal diffusivity to be 0.6×10^{-6} m² s⁻¹, the heat capacity to be 1.3 kJ kg⁻¹ C⁻¹, the elastic stiffness of the host rocks to be 10¹⁰ Pa and the excess source pressure to be 10^{6.5} Pa. All but the last of these values are based upon measurements or estimations of natural rock properties (Rubin, 1995). The excess source pressure is similar to that created by dehydration melting reactions as proposed by Clemens and Mawer (1992). Calculations were performed for melts with liquidus temperatures of 1000, 900, 800 or 700°C and viscosities calculated from the newly calibrated model. The temperature of the rheological solidus was considered to be 650°C which is ~75°C above the solidus of a granite at 0.8 GPa (Naney, 1983) and is an estimate of where the fraction of crystals may be ~50% by volume. The effect of crystals on melt viscosity was ignored because it was not incorporated into this dike propagation model (Rubin, 1995). The calculations do not provide exact estimates for dike propagation, however they produce at least order-of-magnitude predictions of dike propagation lengths which, as Rubin (1995) repeatedly states, are optimistic.

The value of l_0^{\min} increases rapidly as the temperature of the magma source is decreased or the viscosity of the magma is increased, but appears to become less temperature sensitive near 800°C for melts with 4 wt% H₂O (Fig. 4). At 1000°C the length is approximately

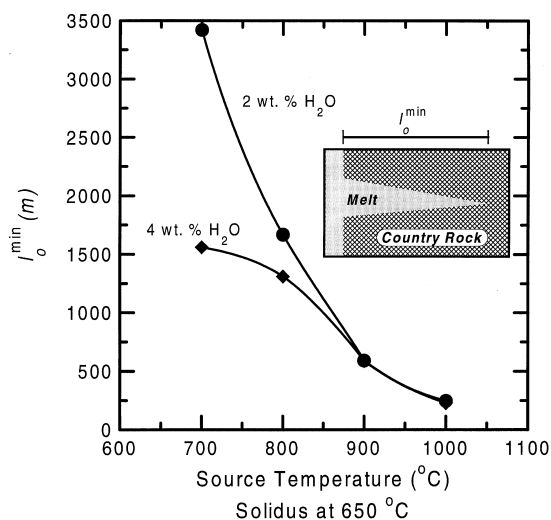


Fig. 4. Calculation of l_0^{\min} vs initial melt temperature at the source of the dike. The l_0^{\min} is the minimum necessary distance between the magma source–country rocks contact and country rocks at the solidus temperature of the intruding magma that will allow dikes to form. The solidus in this and all subsequent diagrams refers to the rheological solidus as defined in the text.

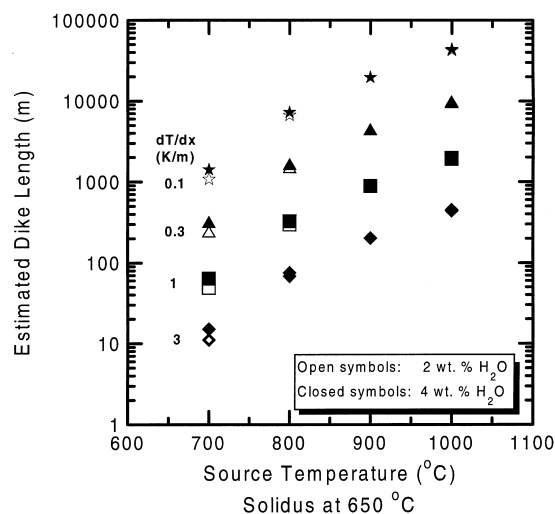


Fig. 5. Calculated dike lengths for granitic melts of various initial temperatures at their source, H₂O contents of 2 and 4 wt%, and thermal gradients of 3, 1, 0.3, and 0.1 K m⁻¹ at the contact between the melt source and the host rocks.

230 m for melts with either 2 or 4 wt% H₂O. With a decrease in source temperature to 900°C the length has almost tripled for melts with 2 or 4 wt% H₂O to nearly 600 m. A further temperature decrease in the source to 800°C approximately triples again l_0^{\min} for both melts, but lower temperatures have little additional effect for melts with 4 wt% H₂O (Fig. 4). However, l_0^{\min} for melts with 2 wt% H₂O at 700°C is twice that of these melts at 800°C.

The calculated dike lengths demonstrate that with the higher temperature, damper melts are expected to propagate further (Fig. 5). However, the effect of H₂O content on dike length is not as significant as in the l_0^{\min} calculations. At a given thermal gradient in the host rocks, the dikes formed from 1000°C source rocks are approximately 1.5 orders of magnitude longer than those formed at 700°C (Fig. 5). Dike length is strongly affected by the host rocks' thermal gradient. For each decrease in the thermal gradient by a factor of approximately one-third the dike length increases by approximately a factor of 10. Dikes formed at 1000°C can only propagate hundreds of meters into rocks with a 3 K m⁻¹ thermal gradient, but are predicted to propagate many tens of kilometers into rocks with a 0.1 K m⁻¹ thermal gradient (Fig. 5). If the thermal gradient is 0.1 K m⁻¹, dikes formed from a source region at 700°C can only propagate approximately 1 km.

The low-temperature results of these calculations are similar to the results of Rubin (1995) because the calculated low-temperature viscosities in both studies are similar, 10⁶ Pa s. In order to extract damp granitic melts from sources at temperatures of 800°C and below, the country rocks must be significantly heated so that they are above the rheological solidus of the magma 1.5–3 km away from the magma source (Fig. 4) and the thermal gradient must be low for dikes to achieve lengths on the order of kilometers (Fig. 5).

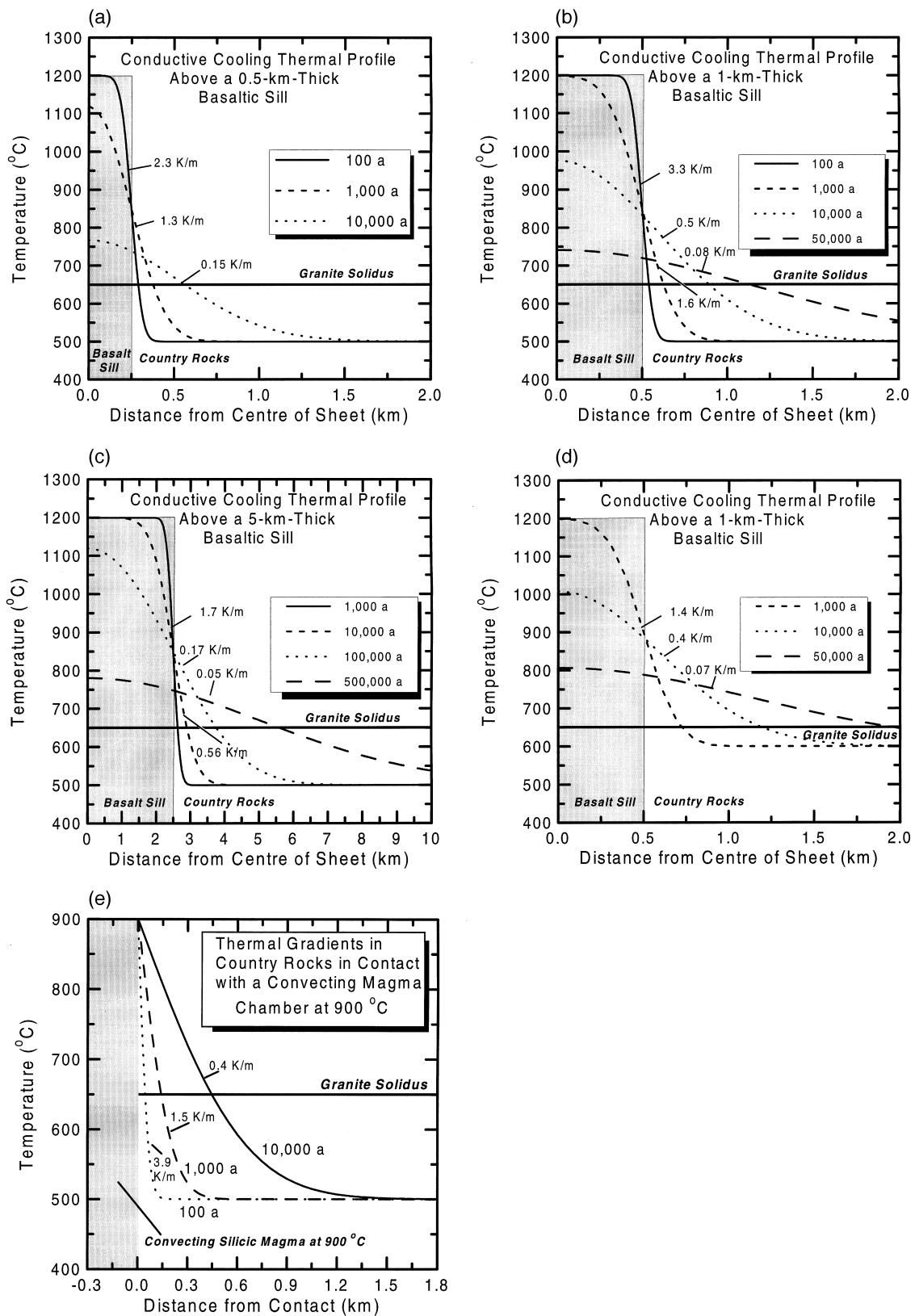


Fig. 6. Modeled thermal evolution near the contacts of igneous intrusions and host rocks. The temperatures and thermal gradients, in $K m^{-1}$, at the contacts are used in conjunction with the results portrayed in Figs 4 and 5 to predict the viability of granitic melt transport by diking. Modeled thermal profiles due to conductive cooling of (a) 0.5-km, (b) 1.0-km and (c) 5.0-km-thick basaltic sills or dikes at various times after instantaneous intrusion into a 500°C crust are presented. (d) displays thermal profiles for a 1-km-thick basaltic dike or sill intruded into a 600°C crust, and (e) gives thermal profiles in country rocks (initially at 500°C) next to a convecting silicic magma chamber 300 m thick.

High-temperature, low-viscosity granitic melts would be more easily extractable from their source regions because of their low thermal entrance lengths (< 1 km, Fig. 4). The calculations suggest that for dikes of damp, granitic melt to propagate kilometers through the crust the magma chamber needs to be at least at $\sim 800^\circ\text{C}$ (Fig. 4) and the thermal gradient at the contact between the magma chamber and country rock needs to be $\sim 0.3 \text{ K m}^{-1}$ or lower (Fig. 5).

The dike propagation calculations have been combined with thermal modeling following Jaeger (1968) and Huppert and Sparks (1988) to investigate the propagation of granitic dikes away from source regions of granitic magmas formed by the intrusion of tabular basaltic sills or dikes in the mid to lower crust because basaltic underplating of the crust has been presumed to be the heat source for granitic magmatism by many researchers (Bergantz and Dawes, 1994). Five illustrative cases have been modeled (Fig. 6). The first three (Fig. 6a–c) involve the instantaneous intrusion of basaltic magma with an initial temperature of 1200°C into a crust at 500°C . The only difference between these three models is the thickness of the intrusion, 500 m, 1 km and 5 km. The fourth model (Fig. 6d) investigates a 1-km-thick basaltic sill or dike intruded into country rocks at 600°C . In these cases the basaltic magma cools by conduction only, and the temperature evolution of the intrusion and surrounding rocks was modeled following Jaeger (1968). The source of granitic melts in all these cases is due to crystallization of the intrusion or partial melting of the wall rocks. The fifth model (Fig. 6e) is based upon the results of Huppert and Sparks (1988) and is an investigation of the conductive thermal gradients in country rocks surrounding a 300-m thick intrusion of convecting granitic magma in country rocks initially at 500°C . This magma is formed by the interaction of a cooling, convecting basalt sill 500 m thick with the lower crust (Huppert and Sparks, 1988). For the purpose of calculating the thermal gradient and temperatures in the host rocks, the time averaged temperature of the convecting silicic magma chamber was estimated from Fig. 5 of Huppert and Sparks (1988) to be 900°C , and the location of the contact between the magma and the country rocks held constant in space at that temperature. In none of these models is the latent heat of crystallization or melting considered. The latent heat liberated by crystallization in the magma chambers would result in slower cooling of the magma chambers and more heating of the host rocks. However, the latent heat of melting in the country rocks surrounding the magma chambers would result in more rapid cooling of the magma chambers and less heating of the host rocks. Although the lack of inclusion of the latent heat of melting and crystallization is obviously a simplification, these illustrative models should provide reasonable estimates of the cooling profiles and ther-

mal gradients in the intrusions and host rocks surrounding them (Jaeger, 1968).

The two first conductive cooling models (Fig. 6a & b) demonstrate that basaltic intrusions of 1 km or less in thickness intruded into country rocks at 500°C are unlikely to result in the formation of significant granitic dikes. The thermal effects of intrusion are limited to within 200 m of the contact with the host rocks, and the thermal gradients near the contacts are between 1 and 3 K m^{-1} at times from 100 to 1000 y. Comparison of these thermal models at 100 and 1000 y with the dike propagation model (Figs 4 & 5) demonstrates that only the high-temperature melts, 900 – 1000°C , formed inside the intrusion have any chance of propagating, but the dike propagation distances are only 10 m to 1 km. With continued cooling of both intrusions, the thermal gradient at the contact decreases which would increase dike propagation distances, but the value of t_0^{min} increases rapidly which decreases the probability of dike generation. After 10,000 y the 500-m thick intrusion appears unable to generate granitic dikes. Comparison of the calculated thermal parameters at 10,000 y for the 1-km-thick intrusion with Figs 4 and 5 indicates that these conditions are favorable for the propagation of km-long dikes with 900°C melts. After 50,000 y the 500-m-thick intrusion has cooled below the granitic solidus, whereas conditions for the 1-km-thick intrusion can allow dikes of granitic melts with 4 wt% H_2O to propagate, but their lengths are likely to be of the order of 1 km. Importantly, all granitic dikes generated from 0.5- and 1-km-thick basaltic intrusions must originate from inside the intrusion and thus contain small-volume melts, on the order of 1% by mass of the initial basaltic magmas, produced by crystallization-fractionation of the basaltic magma.

If the intrusion thickness is increased to 5 km then the formation of granitic dikes from partially melted country rocks is possible (Fig. 6c). In this case, granitic

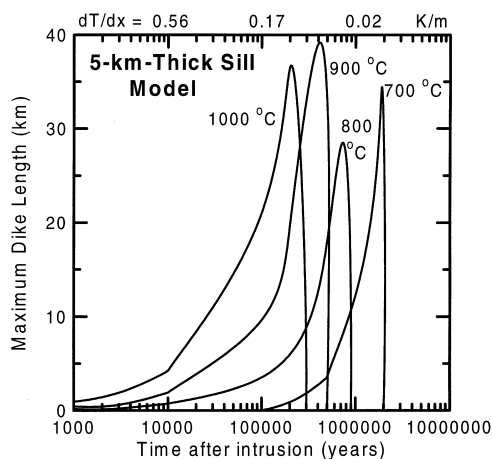


Fig. 7. Calculated granitic dike lengths with time for dikes produced during the intrusion and cooling of a 5-km-thick basaltic intrusion in 500°C crust (see Fig. 6c).

melts produced at all temperatures from 1000 to 700°C can be transported distances between 25 and 40 km, i.e. through a significant portion of the crust (Fig. 7). The volume of granitic melts produced and transported in this case is far greater than in the previous two cases because melts formed in the (presumably) fertile country rocks can be transported by diking, unlike previous cases. The evolution of dike lengths with time (Fig. 7) demonstrates that higher temperature dikes, 1000 and 900°C, are the first to propagate significant distances after 50,000–100,000 y. After a few hundred thousand years the country rocks have been heated sufficiently and the thermal gradient decreased to the point where melts produced at 800 and 700°C can be transported by dike propagation (Fig. 7). These calculations predict that all dikes will have similar maximum lengths.

The importance of country rock temperature on dike propagation is demonstrated in the model of a 1-km thick basaltic intrusion in country rocks with an initial temperature of 600°C (Fig. 6d). For this case melts produced inside the intrusion at 1000°C can escape by diking between approximately 1000 and 10,000 y after intrusion. The lengths of these dikes are predicted to be less than 10 km long. After 10,000 years melts produced at 900°C both inside the intrusion and at its contact with the country rocks may also be transported by dikes which would have lengths reaching a few kilometers. Melts with 4 wt% H₂O produced at 700–800°C can also form dikes if melting occurs near the contact between the basalt and country rocks. Lengths of these dikes appear limited to a maximum of 10 km. Melts with 2 wt% H₂O can only form dikes if the source temperature is 800°C or greater; dike lengths are similar to those of more hydrous melts. Modeling performed with 0.5-km thick intrusions in 600°C country rock demonstrates that no significant diking will occur, whereas 5.0-km thick intrusions in 600°C country rocks are more effective at dike production than when the country rocks are initially 500°C (Fig. 6c).

Calculations of the thermal gradients in the host rocks surrounding a convecting silicic magma chamber at 900°C indicate that dike propagation is not predicted for times of 100 and 1000 y after intrusion (Figs 4, 5 & 6e). Only after 10,000 y does dike propagation seem possible; unfortunately Huppert and Sparks (1988) estimate that the magma chamber will cease convection after 1000 y and cool by conduction. Thus, the propagation of dikes of granitic melt does not seem possible in this example.

Of the five cases investigated, only the 5-km thick intrusion (Fig. 6c) appears to create significant granitic dikes which might be expected to traverse tens of kilometers in the crust and feed growing granitic plutons. In the other two conductive cooling cases involving 500°C crust (Fig. 6a & b) the melts transported by dikes must be formed near the center of the intrusion

in order to achieve lengths greater than the thermal entrance length before the host rock temperature falls below the rheological solidus of the granitic melt. Melts of the country rock formed near the contact are far less likely to escape by the diking mechanism, although these melts could be entrained into dikes originating deep within the intrusion. In the case of the 1-km thick intrusion in 600°C country rocks the formation of granitic dikes is possible, but the maximum calculated dike lengths are 10 km (Fig. 6d). Therefore, this case is not as favorable for granitic dike formation as that of the 5-km thick intrusion (Fig. 6c). Despite the fact that only one of the modeled cases predicts that granitic melts will be transported more than 10 km by the diking mechanism, these simplistic models suggest that in reality dike propagation can play a significant role in the transport of granitic melts. Many other, more complicated, cases can be easily envisioned which would provide further support for dike transport of granitic melts. The models in this paper only calculate the heat effects of individual intrusions injected into the lower crust; based upon the results of the models presented it is easy to extrapolate and envision the complex case of multiple intrusions which would be more effective at heating the lower crust. Multiple intrusions of basaltic magma occurring only once every 10,000 y in the same region of the lower crust could provide sufficient heat to partially melt large volumes of the crust and lower the thermal gradient at the contact between granitic melts and their host rocks, thus favoring dike propagation. More complicated models were not performed in this study, however, because the great number of permutations makes quantitative modeling poorly constrained. Although the results of this study support dike transport of granitic magmas this support should not be misconstrued as evidence against diapiric transport of granitic melts. Recent studies have, however, indicated that diapirs are not efficient transport mechanisms for granitic melts (Clemens *et al.*, 1997)

GEOLOGICAL REALITY: SOME CAVEATS

The model for dike initiation and growth of Rubin (1995) combined with the thermal modeling and new data for granitic melt viscosities discussed above provide important insights into the processes responsible in the formation of granitic dikes in the lower crust. The processes involved in the formation of dikes and melt transport are, however, much more complex than considered in the models, and it is perhaps worthwhile to discuss a few of the geological realities which are not considered by the models. This list is by no means exhaustive; it is meant to highlight some important aspects which must be investigated in future models of magmatic transport.

The first limitation considered is the supply of melt to the propagating dikes. Rubin's (1995) model together with those of Clemens and Mawer (1992) and of Petford *et al.* (1993) require that a reservoir is present to supply melt to the dikes as they propagate. Can melt be extracted from a partially molten region rapidly enough to feed the dikes? Models of the transition between porous flow of melt and diking (Sleep, 1988; Rutter and Neumann, 1995) suggest that melt extraction can be rapid in some cases, but the time necessary to extract 10% volume of melt from a partially melted granite at 900°C by a vein network is slow, $\sim 10^5$ y, if the deviatoric stress, ΔP , is 10^6 Pa (Rutter and Neumann, 1995). Based upon geological observations, the segregation of partial melt from paleosomes to form leucosomes in migmatite terranes appears to occur on the time scale of tens of years (Sawyer, 1991) which suggests that segregation might be rapid enough to feed propagating dikes. Alternatively, we could hypothesize that melts first segregate by some form of compaction mechanism to form magma chambers with dimensions on the orders of meters to tens of meters and that these chambers, which contain a high proportion of melt, form the reservoirs for dikes. These dikes may coalesce at a higher level in the crust to form a larger magma chamber which eventually gives birth to dikes which ascend even higher once enough heat and melt have been delivered to this upper-level magma chamber. Such a process could occur several times between the site of melt generation in the lower crust and the initial emplacement in the middle to upper crust. Many other alternatives must exist; to understand and choose between them demands more complex modeling and further studies of deeply eroded sections of the crust believed to be the source regions for granitic magmas.

Another limitation involves the thermal modeling and transport of heat. The modeling used above is simplistic in that many thermal aspects of magma genesis and transport are ignored: e.g. latent heat of melting and crystallization are not considered, convection is ignored, and the lower crust initially is considered to be at constant temperature over a few kilometers distance. However important these aspects may be, they can be dwarfed by the effects of dike propagation on the lower crust's thermal regime. Any dike propagating from the hot source region into cooler crust will act as a mechanism of heat transport which is potentially the most important one in the system. Although it is possible to consider the heat advected by any single dike (N. Petford, personal communication 1997), unless the number and size of dikes generated through time and the distance they travel is known, their effects on the thermal state of the crust cannot be adequately modeled. As seen in the calculations of Rubin's (1995) model, the thermal state of the crust, in particular the thermal gradient at the contact between the magma

chamber and the country rocks, has a strong influence on dike propagation and must be known.

The dike propagation model discussed for the 5-km thick basaltic intrusion suggests that dikes formed at high temperatures near the contact between the intrusion and the lower-crustal country rocks can propagate significant distances up through the crust. However, the model does not consider the interaction between these propagating dikes and the cooler, but still partially molten, country rocks further away from the contact. Can high-temperature dikes traverse the partially molten region of crust surrounding the intrusion? Or do the propagating dikes entrain lower temperature partial melts from the distal regions of the partial melt zone which mix with the higher temperature melts and are transported into the upper crust? These questions are beyond the scope of the current models of dike propagation, but are clearly important and demand answers.

The effects of cooling, crystallization and the entrainment of restitic material on the viscosity of melts in propagating dikes is another realism which is not considered in many models. Although the effects of temperature (this contribution) and crystals (Bergantz and Dawes, 1994) on melt viscosity can be modeled they have not been incorporated into Rubin's (1995) model. Future models will have to consider these aspects.

The models for dike propagation and transport consider the crust to be a homogeneous body free of any zones of weakness. Such is certainly not the case. The transport of melts away from their source regions may be strongly controlled by the presence of faults and zones of weakness which provide natural pathways for melt migration which will result in substantial differences between the models and nature. Field studies of migmatites (Sawyer, 1991, 1993), and granite transport and emplacement mechanisms (Solar and Brown, 1997), suggest that geological structure and deformation may control the transport of granitic magmas from source regions to their final destinations. Unfortunately, incorporating the realities of crustal structures into mechanisms of melt transport is either impossible in most cases or creates a model so regionally based that it loses its general applicability.

Despite these caveats (and many more which the astute reader can easily imagine) modeling of magma segregation and transport is extremely important and useful. These models, simplistic though they are, provide constraints upon processes which may occur in nature. They enhance our understanding of the mechanisms involved in granitic petrogenesis and point towards new areas of research. The general conclusions of the models presented in this paper are: (1) that only large intrusions (or multiple intrusions) can produce enough heat to allow dike transport of granitic magmas; and (2) that the easiest granitic magmas to transport by dikes are high-temperature ones and that only

small quantities of dissolved H₂O are needed. These conclusions are consistent with field observations and are therefore probably robust. However, models are only models and their results must be compared with detailed studies of field relations and petrochemistry.

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