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Ascent mechanism of felsic magmas: news and views

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ABSTRACT: Diapirism has been discredited as a transport mechanism for magmas partly because diapirs seem to be unable to bring magmas to shallow crustal levels (< 10 km) and partly because recent developments in the theory of dyke propagation have shown that sufficiently wide dykes are able to efficiently transport felsic magmas through the crust. However, it is still unclear how felsic dykes grow to widths that allow them to propagate faster than they close by magma freezing. Ultimately, it may be the ability of felsic dykes to grow within the source that controls which mechanism dominates ascent.

The ability of dykes to propagate from the top of rising diapirs depends among other factors on the changing temperature gradient of the wall rocks. The steep gradient around rapidly rising diapirs in the low viscosity lower crust will cause dykes to freeze. As diapirs rise to colder stiffer crust and decelerate, heat diffuses further from the diapir, resulting in shallower temperature gradients that favour dyke propagation. The mechanism may thus swap, during ascent, from diapirism to dyking. Calculations of the thermal evolution of diapirs and their surroundings show that basaltic diapirs may never form because they would be drained by dykes at a very early stage; felsic diapirs may be unable to give rise to successful dykes, whereas diapirs of intermediate magmas may propagate dykes during ascent.

KEY WORDS: diapirs, dykes, magma segregation, magma emplacement, magma transport.

There has been considerable discussion on the mechanisms of magma transport through the continental crust in recent publications (Cruden 1990; Emerman & Marrett 1990; Lister & Kerr 1991; Clemens & Mawer 1992; Cruden & Aaro 1992; Paterson & Fowler 1993, 1994; Weinberg 1994; Weinberg & Podladchikov 1994, 1995; Petford, this volume). Discussion concentrates mainly on two mechanisms: dykes and diapirs. The limiting parameter controlling the velocity of dyke propagation is magma viscosity. For diapirs magma viscosity plays only a minor part in determining the velocity, which is mainly controlled by the much higher viscosity of the diapir's surrounding. The models of Grout (1945) and Ramberg (1967) made diapirs a popular mechanism to explain the ascent of felsic magmas through the crust. This was because diapirs are buoyant and may give rise to voluminous plutons that show geometries similar to those resulting from the models, such as elliptical shape and concentric patterns of foliation (Sylvester 1964; Brun & Pons 1981; Bateman 1985; Courrioux 1987; Ramsay 1989; Cruden & Aaro 1992; Paterson & Fowler 1993). The idea that felsic magmas rise as diapirs was further strengthened by the observation that large granitic dyke swarms are rare compared with basic swarms, suggesting that most felsic magmas would be too viscous to rise rapidly enough to avoid freezing.

The controversy regarding the transport of felsic magmas started in the 1980s when several workers suggested that the crust was too viscous to allow diapirs to rise fast, thus they froze after travelling only a short distance (Marsh 1982; Morris 1982; Ribe 1983; Daly & Raefsky 1985; Mahon *et al.* 1988). Similar results were found when the effects of the thermal softening of wall rocks by hot diapirs (hot Stokes models) were included: the faster ascent due to lower viscosity was counteracted by faster magma cooling. These workers concluded that diapirism of magmas through the crust was a slow process limited by the thermal energy of diapirs, and that to be efficient diapirism requires either an anomalously hot crust or several diapirs heating up a path through the crust. Paterson & Vernon (1995) argued that the latter is just what is observed in nature, with several diapirs following the same path to nest into each other and form the commonly observed pluton zonation. Doubts about diapirs were strengthened by a test carried out by Schwerdtner (1990) in the Archaean crust of Ontario, which showed that diapirism was unlikely to give rise to the structures observed around the studied domes.

The situation at that time was such that it seemed that the viscosity of the crust was too high for diapirs, and that the viscosity of felsic magmas was too high to allow dykes to propagate without first freezing. In the beginning of the 1990s, Lister and Kerr (1991), Clemens and Mawer (1992) and Petford *et al.* (1993) showed that, *given a large enough initial dyke*, felsic magmas can rapidly and episodically rise across the crust through dykes and form large batholiths in the upper crust, despite their high viscosity. The critical width required for felsic dykes to rise without freezing was estimated to be from a few metres to a few tens of metres (Petford *et al.* 1993, 1994; Lister 1995), reasonable values for observed dyke widths (Wada 1994, 1995; Kerr & Lister 1995). The common flow structures observed in and around plutons and batholiths fed by dykes would be caused, not by ascent, but by the emplaced magma warming up the surrounding rocks which then flowed due to the expansion of the magma chamber.

The efficiency of magma transport through dykes, coupled with the common spatial association between plutons and large crustal-scale fault zones, have led several workers to suggest that faults and shear zones are pathways for magmas and are also responsible for creating space for magma emplacement through extension (e.g. Pitcher 1979;

Guineberteau *et al.* 1987; Hutton *et al.* 1990; Hacker *et al.* 1992; Hutton & Reavy 1992; Petford & Atherton 1992; Petford *et al.* 1993; Brown 1994; Ingram & Hutton 1994). Sleep (1975) showed that shallow basaltic magma chambers can be maintained at shallow depths along mid-ocean ridges due to the prevailing extension rates. Only very recently have we learned that crustal extensional rates are able to open space rapidly enough to give rise to a steady-state felsic magma chamber (Hanson & Glazner 1995). A steady-state chamber describes a chamber where magma slowly accumulates and interacts with older (still not totally solidified) magma, in contrast with a collection of small frozen magma batches or dykes that results when extension is too slow. Hanson & Glazner (1995) note, however, that accurate dating is needed to determine if magma chambers grow at appropriate extension rates or faster. The problem of space for the emplacement of large magma volumes is by no means solved (e.g. Paterson & Fowler 1993, 1994; Weinberg 1994; Schwerdtner 1995; Petford, this issue). That magmas find or open space for themselves does not seem to be a problem; the space problem lies in understanding how this is done. Once this is understood, we might be able to fully understand the ascent and emplacement of granites. Diapirs make their own space by imposing flow on the surrounding crust. The width of the strain aureole is controlled by the temperature distribution and the rheology of the wall rocks (Weinberg & Podladchikov 1995). Paterson and Fowler (1993), studying the aureoles of several plutons, suggested that the observed shortening was insufficient to account for the pluton volume and that therefore mechanisms other than the viscous flow of the wall rocks were necessary to create that space. However, Weinberg (1994) and Schwerdtner (1995) pointed out that volume estimations resulting from shortening measured at the margins of essentially two-dimensional plutons cannot be integrated into three dimensions with confidence, and that the method used by Paterson and Fowler (1993) is unable to yield reliable estimates of the volume displaced by diapirs. For plutons fed by dykes it is clear that space for magma emplacement needs to be opened and several mechanisms are available, such as faulting, stoping, doming of the roof, caldera subsidence and viscous flow of the wall rocks. As suggested by Petford *et al.* (1994), emplacement may depend on the temperature and rheology of the crust and the rate of magma flow into the chamber. Buddington (1959) and, more recently, Paterson and Fowler (1993) suggested that more than one mechanism may be required to explain the ascent and emplacement of shallow level batholiths and that different mechanisms may dominate ascent at different depths.

Counterbalancing the recent tendency in publications to favour dykes as the most effective ascent mechanism, Rubin (1993a, b, 1995) pointed out that a fundamental question remains to be answered by the proponents of the dykes-ballooning model: how do felsic dykes survive the early stages of propagation without freezing? As the velocity of dyke propagation increases linearly with dyke length (and decreases linearly with magma viscosity), small dykes starting from the magma source and progressing into rocks at subsolidus temperatures will propagate slowly and magma freezing may clog the dyke and stop propagation. Freezing will halt most rhyolitic dykes soon after intrusion into rocks at subsolidus temperature (Rubin 1995). However, hot rhyolites (100°C above its solidus) may successfully develop dykes given low magma viscosity (10^4 Pa s) or high magma pressure (> 10 MPa), or low temperature gradients in the surrounding rocks (< 5°C/km), or some combination of these (Rubin 1995). These conditions limit felsic dykes to narrow and rather anomalous situations. Regarding diapirism, recent work has shown that diapirs may be more efficient than previously thought (Weinberg & Podladchikov 1994, 1995). These workers showed how the strain rate softening of power law crust, rather than thermal softening (hot Stokes models), may allow diapirs to rise fast enough to reach upper crustal levels before freezing. Strain-rate softening relies on the diapir's velocity to decrease the crustal viscosity rather than the heat content of the rising magma, allowing diapirs to rise faster without considerably increasing their cooling rate (Weinberg & Podladchikov 1994). These workers showed, however, that individual diapirs would require anomalous crustal temperatures or extremely low viscosity upper crust to reach shallow crustal levels (< 10 km).

Although considerable progress has been made in understanding magma transport through the crust, the theories of dyking and diapirism are still unable to answer several fundamental questions. I start this paper by discussing the problem of dyke initiation, suggesting conditions that may favour or inhibit dyking. Results by Rubin (1993a, 1995) are applied to a rising diapir to determine if at any point during rise dykes would successfully leave the diapir and drain it of its magma. The results suggest that a swap in ascent mechanism may occur for magmas of viscosity of intermediate values ($\approx 10^3$–10^6 Pa s). Finally, I discuss a few other questions regarding dykes and diapirs. Firstly, the proximity of plutons and fault/shear zones is discussed and it is suggested that this may result from shear zones focusing melt migration in the source and thus controlling the site for magmatic ascent either as diapirs or dykes. Then, I discuss the difficulties in finding evidence of the pathways of diapirs and the question of why felsic plutons balloon (Rubin 1995).

1. Controls on dyke initiation from a partially molten zone

We know that large wide dykes (two or more metres in the direction perpendicular to the dyke walls) may be able to crack the surrounding rocks and transport rapidly large volumes of felsic magmas through the crust. We also know that wide felsic dykes are not uncommon (Wada 1994). Very small and narrow dykes will take infinitely long to grow because their velocity tends to zero as their length or width tends to zero. If the propagation velocity is slow the magma in the dyke will freeze if surrounded by rocks at subsolidus temperature (e.g. Bruce & Huppert 1989; Rubin 1995). Bruce and Huppert (1989) introduced the concept of 'critical initial width', which is the minimum width of a dyke which allows it to propagate without becoming clogged by freezing magma. If dykes are able to grow within the supersolidus magma source to widths beyond critical, they will survive when they propagate into and across the subsolidus crust. The question then is how melt segregation processes in the source are likely to influence or control dyke initiation. Is the process of porous flow of melt into a network of veins that drain into large dykes capable of draining the source efficiently, and are dykes within the source large enough to survive the freezing temperatures outside the magma source (Sleep 1988)? If segregation is unable to give rise to such dykes, then doming of the buoyant source with or without concomitant segregation may lead to diapirs (Fig. 1). The limited ability of stiff magmas to initiate dykes may ultimately control which transport mechanism dominates ascent. In the next section I discuss how magma segregation, tectonic stresses and viscosity ratios between magmas and wall rocks may favour or inhibit the initiation of dykes.

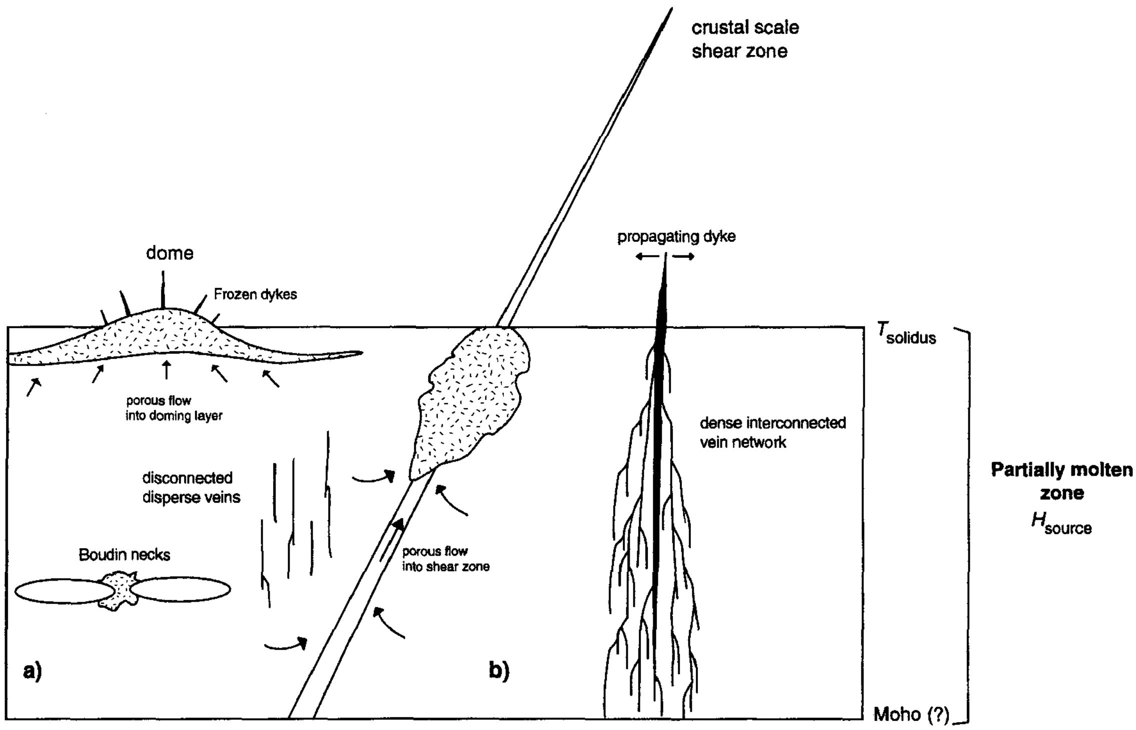

Figure 1 Schematic view of melt segregation, doming and dyking of a partially molten zone. Melt flows into low pressure sites such as boudin necks, low viscosity shear zones (active or inactive) and the hinge of a dome or regions previously enriched in melt to form veins (Stevenson 1989). Where dykes are unable to propagate, either the whole source may dome or, as depicted, melt accumulated close to the top of the source layer through porous flow (or in any low pressure site) may start to dome (a); where melt segregation gives rise to a dense network of interconnected veins, dykes may be able to propagate into colder surroundings without freezing (b). The shear zone cutting across the source may act as a magma sink and drain the surroundings. Independently of whether shear zones control the mechanism of ascent, by acting as a low pressure site, they may be an important control on the site of magma ascent.

1.1. Veins in the source

The starting point of this discussion is that creep is the natural response of the lower crust to applied stress and that granite veins only develop because of particular conditions prevailing in the molten zone. Dell'Angelo and Tullis (1988) found that in partially molten granite samples undergoing deformation in laboratory experiments, the presence of melt may result in the concurrent operation of rock cataclasis, diffusion creep and dislocation creep. These mechanisms are not normally operative for the same P–T conditions and their simultaneous occurrence may lead to incorrect interpretations regarding these conditions. Particularly relevant here is that their experiments showed that when the imposed strain rate is faster than the rate with which melt can flow laterally to low pressure sites (out of the system in their experiments), high pore pressures result in rock cataclasis with fractures parallel to the orientation of the maximum compressional axis (σ_1). Once cataclasis occurs, melt migrates to low pressure sites, the pore pressure drops and the rock resumes creep. The results of Dell'Angelo and Tullis (1988) suggest that cataclasis of the source would be enhanced by high strain rates or high magma viscosity. Owing to spatial variations in melt fractions and permeability, cataclasis may occur in some parts of the source while deformation is controlled by creep in others. As cataclasis at these high P–T conditions is triggered by the presence of melt, creep will still dominate the deformation of the surrounding solid rock. In other words, the fact that the viscous partially molten zone is able to crack does not imply that the stiffer subsolidus crust is able to do the same. Cataclasis is thus a brittle mechanism that may give rise to veins within the source zone when the strain rate is fast enough to increase the local pore pressure.

In a heterogeneous partially molten zone there are generally several natural low pressure sites that act as melt sinks, such as boudin necks (Brown 1994), fold hinge zones (Allibone & Norris 1992) or, less obviously, low pressure zones resulting from the layering of rocks of different viscosities (filter pressing; Robin 1979; Miller & McLellan 1986; Wickham 1987). Stevenson (1989) showed that a partial melt undergoing large-scale deformation is unstable, and melt migrates through pores in the direction of the minimum compressional axis σ_3 to low pressure regions slightly enriched in melt to give rise to veins parallel to the maximum compressional axis σ_1. This process may have a preferred length scale of the order of a metre (Stevenson 1989) and although it does not involve rock failure, it could give rise to a network of interconnected veins oriented parallel to σ_1. This mechanism is probably common to any source regions undergoing large-scale deformation and gives rise to veins without requiring the brittle failure of source rocks.

1.2. Melt segregation, tectonic stresses and magma stresses

Different segregation mechanisms and different stress orientations will influence the ability of dykes to propagate into subsolidus crust by controlling the magma pressure of initiating dykes. The pressure at the tip of a dyke is one of the main factors controlling the success of dyke propagation (Lister & Kerr 1991; Rubin 1993a, b). Magma pressure in the source depends on the dynamics of melt segregation and on dyke orientation, which is strongly controlled by tectonic stresses. If the segregation of magmas occurs by porous flow (McKenzie 1984) resulting in a magma layer close to the top of the partially molten zone (Fig. 1; Fountain *et al.* 1989), the maximum bouyancy stress (magma pressure, σ_m) on an ensuing crack will be

$$\sigma_m = \Delta\rho\ g\ H_{melt}$$

where $\Delta\rho$ is the density difference between the melt and its surroundings, g is the gravity acceleration and H_{melt} is the thickness of the molten layer ($H_{melt} < H_{source}$, where H_{source} is the source layer thickness). If, on the other hand, melt segregates to form veins, the maximum buoyancy stress results from an interconnected network of steeply dipping static (immobile melt) veins cutting across the entire source. In this ideal case

$$\sigma_m = \Delta\rho\ g\ H_{source}.$$

If magma is moving through this network (most likely), the pressure decreases significantly due to friction at the walls to

$$\sigma_m = \Delta\rho\ g\ L$$

where L is the compaction length (McKenzie 1987), characteristically much smaller than H_{source}.

Vein orientation is strongly controlled by crustal heterogeneities and the orientation of tectonic stresses. Although extensional regimes would tend to favour vertical or steeply dipping veins, shortening regimes would favour horizontal or shallow dipping veins. If the combination of crustal heterogeneities and stress results in horizontal or shallow dipping veins and dykes (e.g. compressional regimes), their orientation is perpendicular to the direction of the buoyancy force, which is distributed over a large area (low stress), and

$$\sigma_m = \Delta\rho\ g\ w_{dyke}$$

where w_{dyke} is the dyke width. (In this discussion, any overpressure caused by the exsolution of gases has been neglected, as the melt is assumed to be in equilibrium within the source).

In summary, for a given magma viscosity, the most favourable conditions for dyke propagation from the source arises in an extending crust in the case where segregation gives rise to an immobile interconnected network of steeply dipping veins where high pressure is concentrated at the dyke tips (Fig. 1b). Conversely, conditions least likely to favour dykes and most favourable to diapirs arise in shortening environments, when segregation is dominated by porous flow or when it gives rise to disconnected veins (Fig. 1a). The magma pressure in dyke tips required to enable propagation into cold subsolidus crust depends, among other parameters, on the temperature gradient in the subsolidus crust, magma viscosity and its excess temperature (i.e. the difference between magma temperature and solidus temperature, ΔT^*; see Rubin 1995 for detailed discussion). If segregation within the source gives rise to these high pressure dykes, we may have rapid melt extraction. On the other hand, if the source is unable to give rise to high pressure dykes or if this process is slow (e.g. high viscosity melt or low permeability), the buoyant source or those low pressure sites where magma converged during segregation may start to dome. Doming may in turn enhance melt segregation and focus melt flow towards the dome's hinges. Magma bodies thus formed may eventually detach from the source, forming a diapir and leaving restite-rich migmatites along its tail. The efficiency and velocity of this process depends on the efficiency of melt segregation, the magma buoyancy, and the viscosity of the wall rocks (a function of rock rheology, temperature and diapir buoyancy; see Weinberg & Podladchikov 1994).

1.3. Viscosity ratio between melt and solid rocks

Corriveau and Leblanc (1995) showed in the Grenville Province, Quebec how magmas that were ascending in dykes became trapped in a low viscosity marble-rich layer, through which they rose as diapirs. This change in ascent mechanism results from the faster viscous response of the marble to the applied magma stress compared with that of stiffer rocks. Two mechanisms were envisaged by Rubin (1993a) where the viscosity contrast between the magma and the surrounding rocks may prevent dykes propagating efficiently: (a) if the viscosity contrast is sufficiently small, the tip of an initiating dyke might become blunt in the time required for the dyke to inflate, so that the stress concentration at the tip (a function of tip sharpness) never becomes sufficient to further crack the rock; (b) in the case of magma invading a pre-existing fracture above a rising diapir, if the viscosity contrast is sufficiently small, the large-scale flow around the diapir widens the dyke faster than the dyke tip propagates. Whereas the diapiric flow of the wall rocks limits the dyke's ability to develop in the latter, it is the process of fracturing that limits dyking in the former. In the field example of Quebec, either of the two (or both) mechanisms may have prevented the dykes from propagating through the marble-rich layer.

Rubin's (1993a) study of the response of viscoelastic rocks to a propagating dyke showed that for expected magma pressures, elastic stiffness $G = (1-\nu)/\mu$ (where ν is the Poisson's ratio and μ the shear modulus) and magma viscosities, the wall rock viscosities would have to be extremely low to respond as an essentially viscous fluid. For example, using Rubin's (1993a) figure 4, for magma viscosity of 10^8 Pa s, the surrounding crust would need viscosities less than 10^{15} Pa s to behave as a purely viscous fluid. Although the viscosity of a very soft marble submitted to magma stresses could be that low, most crustal rocks are likely to be stiffer. For this more general case the crust would respond to a propagating dyke either as a purely elastic medium or as a combination of viscous and elastic. The viscoelastic response requires a low, but geologically reasonable, viscosity ratio, and results in a wide dyke with the tip propagating through an essentially elastic medium, whereas the more central parts of the dyke encounter a viscous response from the surroundings and widens.

2. Magma ascent: from diapirs to dykes

Rubin (1993a, 1995) assumed steep temperature gradients around magma chambers ($dT/dz = 0{\cdot}1$–$1°C/m$) that tend to cause early dyke freezing. In natural magma chambers the temperature gradient is likely to evolve in time from steep during and just after magma emplacement, to shallow as the magma heat diffuses into the surroundings. Dykes that would initially freeze in the cold surroundings might be able to propagate at a later stage through warmer wall rocks. Similarly, the fate of dykes initiating at the top of a rising diapir may change as the temperature gradient evolves. In low viscosity hot crust, fast diapirs impose a steep temperature gradient in the surroundings that causes early dyke freezing. As diapirs rise to stiffer and colder crust, the temperature gradient becomes shallower and dykes may successfully leave the diapir. In this section, we examine if, at some point, as a diapir progresses upwards from hot and soft to cold and stiff crust, magma-filled cracks initiating at the diapir's top are able to propagate without freezing and drain the diapir. This is carried out here by studying the evolution of rising diapirs of different sizes and different viscosity magmas using a modified version of the program Rise (Weinberg & Podladchikov 1994, 1995) that includes the results of Rubin (1993a).

2.1. Method

Rubin (1993a, b) showed that, for magmas at the solidus temperature, a single freezing parameter β predicts the ability of dykes leaving a chamber to propagate into subsolidus crust

$$\beta = \frac{2(3^{\frac{1}{2}}c|\mathrm{d}T/\mathrm{d}z|(\kappa\eta)^{\frac{1}{2}}}{\pi^{\frac{1}{2}}L(p/G)^2 p^{\frac{1}{2}}} < \approx 0{\cdot}15 \qquad (1)$$

where c is the heat capacity, $\mathrm{d}T/\mathrm{d}z$ is the temperature gradient away from the chamber walls, κ is the thermal diffusivity, η is the magma viscosity, L is the latent heat and G is the elastic stiffness. The magma pressure, p, at the dyke entrance is taken here to be the buoyancy stress of the diapir ($p \approx \Delta\rho g r$, where r is the diapir radius, Rubin 1993a). Rubin (1993b) showed that for $\beta < 0{\cdot}15$ dykes would be able to propagate faster than the speed with which freezing would shut them. This critical value of β is weakly dependent on the suction at the cavity at the tip of the dyke and may be rewritten for magmas at solidus temperature and $c = 1$ kJ kg^{-1}°C^{-1}, $L = 400$ kJ kg^{-1}, $\kappa = 10^{-6}$ m^2 s^{-1} and $G = 10$ GPa as

$$\beta^* = \frac{|\mathrm{d}T_0/\mathrm{d}z|^2\eta}{p^5} < 10^{-31}\ \mathrm{s°C\ m^{-2}\ Pa^{-4}} \qquad (2)$$

The computer code Rise calculates the velocity of spherical diapirs rising through crust of Westerly granite rheology (strain rate and temperature-dependent viscosity; Hansen & Carter 1982). As the diapir rises into colder and stiffer crust, the magma temperature and the temperature gradient above the diapir are calculated. The magma temperature is assumed to be homogeneous within the diapir and the magma viscosity and density difference to the surroundings is constant throughout the ascent. In this way β^* can be calculated at each step and compared with the critical value. When $\beta^* \leqslant \beta^*_{\mathrm{crit}} = 10^{-31}$, magma-filled fractures starting at the top of the diapir will be able to propagate into the surrounding crust. The value of β^*_{crit} used assumes that the magma is at its solidus temperature. Although magma, in the calculations, is generally above the solidus temperature, this assumption should not greatly influence the results because, as shown by Rubin (1995), the addition of an excess temperature (ΔT^*) should not greatly enhance the ability of magmas to penetrate the subsolidus crust as dykes. For example, using figure 3 in Rubin (1995), for $\Delta T^* = 100$°C, and $\mathrm{d}T/\mathrm{d}z$ of 0·1 or 1°C/m, when β exceeds the critical value by three times, a dyke would propagate only a distance of $2{\cdot}6 l_0$ (where $l_0 = \Delta T^*/\mathrm{d}T/\mathrm{d}z = 1000$ or 100 m, respectively). If β exceeds the critical value by more than three this propagating distance becomes even smaller. (It is important to note that the maximum ΔT^* coincides with the maximum β at the first step of the calculation, so that ΔT^* will not greatly enhance dyke penetration.)

The solutions of Rubin (1993a, 1995) are for Newtonian fluids. Solutions for power law fluids are unavailable, but the concentration of stresses at the tip of dykes might decrease the viscosity of power law rocks considerably and allow a prompter viscous response to tip stresses so that cracking may become more difficult than for Newtonian fluids. The approach of this paper is to calculate the velocity and the temperature gradient in front of diapirs rising through power law fluids, but to assume Newtonian behaviour during dyke propagation and use Rubin's results. This approach favours dyke propagation. In Equation (2), for a fixed magma chamber, $\mathrm{d}T/\mathrm{d}z$ does not depend on the magma pressure and $\beta \propto p^{-5}$. However, when considering diapirs, an increase in p causes an increase in diapir velocity proportional to p^n (where n is the power law exponent of crustal rocks; Weinberg & Podladchikov 1994). As $\mathrm{d}T/\mathrm{d}z \propto Nu$ (from Daly & Raefsky 1985) and $Nu \propto Pe^{\frac{1}{2}} \propto V^{\frac{1}{2}} \propto p^{n/2}$ (see Weinberg & Podladchikov 1994) then $\mathrm{d}T/\mathrm{d}z \propto p^{-n/2}$ (where Nu is the Nusselt number). Thus for diapirs $\beta \propto p^{n-5}$.

As a diapir rises through the crust, it warms up its surroundings. The temperature gradient around the diapir is controlled by its size, velocity and magma temperature, as well as the temperature and thermal diffusivity of the wall rocks. The temperature gradient also varies around the diapir, being steepest at the top and shallowest at the tail, as rocks at the tail spent a longer time in the vicinity of the diapir. The size and velocity of the diapir control the width of the thermal boundary layer through the Peclet number ($Pe = Vr/\kappa$, where V is the diapir's velocity). As the diapir's velocity continuously decreases as the diapir rises into colder and stiffer crust, Pe decreases and the magma heat propagates further into the surroundings, decreasing the temperature gradient ($\mathrm{d}T/\mathrm{d}z$) and increasing the survival chances of initiating dykes (decrease in β^*). In this way, dykes that would have rapidly frozen when leaving a fast rising lower crustal diapir (high β^*) may become successful in stiffer shallower crust (low β^*).

The temperature gradient in front of the diapir is calculated at every step by first determining, based on Pe, the distance from the diapir's surface in which the temperature decays to 1/e of the diapir's temperature (δ_{T} as defined by Daly & Raefsky 1985), following the methodology described in Weinberg & Podladchikov (1994). In the calculations the initial excess temperature (ΔT^*) equals the initial temperature difference between magma and surrounding solid rock ($\Delta T^* = \Delta T_0$). In this instance, the width of the thermal aureole is generally larger than, but of the same magnitude, as Rubin's (1995) l_0, the distance that a dyke would propagate before it reached subsolidus crust. Therefore, it is assumed here that the temperature gradient within this thermal aureole effectively controls dyke propagation. Because the influence of variation in $\mathrm{d}T/\mathrm{d}z$ in dyke propagation is unknown, $\mathrm{d}T/\mathrm{d}z$ is assumed to be arbitrarily linear within the thermal aureole (Fig. 2) and is

$$\frac{\mathrm{d}T}{\mathrm{d}z} = \frac{T_{\mathrm{m}} - T_{\mathrm{c}}}{\delta_{\mathrm{T}}}\left(1 + \frac{1}{\mathrm{e}}\right) \qquad (3)$$

where T_{m} is magma temperature, T_{c} is undisturbed crustal temperature at the crustal depth δ_{T} above the top of the diapir. As the diapir rises from step to step, the diapir and the undisturbed surroundings cool and a new temperature gradient is calculated (see Weinberg & Podladchikov 1994 for the cooling of the diapir).

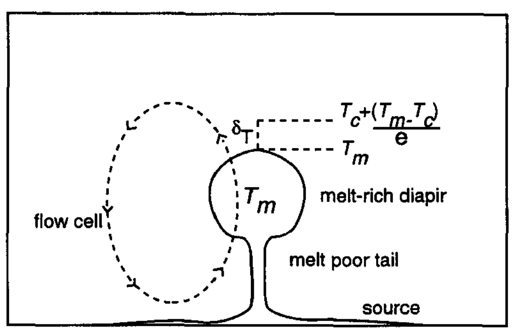

Figure 2 A melt-rich diapir leaves a melt-poor tail behind and causes convection of the surrounding crust. Convection feeds the source with new, fresh material from shallower crustal levels, that may warm up, melt and give rise to a new diapir. The temperature gradient above the diapir is found by determining δ_{T} (the distance from the top in which the magma temperature decays to 1/e) and assuming a linear gradient within that zone (see text for details).

2.2. Results

The calculation starts with a spherical body of magma at 880°C, 100°C warmer than the surroundings, which are at the magma solidus temperature ($\Delta T^* = \Delta T_0 = 100°C$, $T = 780°C$) and a starting depth of 50 km. The temperature of the crust decreases upwards linearly by 15·6°C/km. The magma pressure is assumed to be constant, which disregards any pressure changes related to magma solidification and volatile exsolution. The spherical diapir rises and β is calculated at each step. If at any point during ascent $\beta \leqslant \beta_{crit}$ the calculation is stopped, because the diapir will cease to exist as its magma is drained rapidly through propagating dykes. The same procedure was carried out for a series of diapirs of different sizes and viscosities (Fig. 3). As expected from Equation (2), increasing magma pressures (increasing radii in Fig. 3) allow more viscous magmas to crack the crust and rise in dykes. Highly viscous magmas rise as diapirs without ever being drained by dykes. Conversely, low viscosity magmas will immediately go from the initial geometry into dykes. There is, however, a zone of intermediate viscosity in which ascent starts as diapirs and swaps to dykes at some depth. Surprisingly, this intermediate zone occurs at a range of viscosities that corresponds broadly to that of magmas of intermediate composition. (Note that wet, silica-rich magmas may have viscosities as low as 10^5 Pa s, but the tendency of wet magmas to solidify when decompressed is likely to cause them to remain close to their source whether they rise as diapirs or dykes.)

The influence of the initial temperature difference between magma and solid crust, ΔT_0, was studied. When $\Delta T_0 = 0$ (the case of a hot crust surrounding the magma), the temperature gradient above the magma body is the geothermal gradient, generally a small value that would greatly favour dyke propagation. The dynamics of magma segregation within the source is likely to give rise, through the flow of magma and advection of heat, to magma pools within the source region that are warmer than their immediate melt–depleted surrounding rocks ($\Delta T > 0$; e.g. Fountain *et al.* 1989). It was found here that, above some critical value, changes in ΔT_0 do not greatly influence either the depth at which the swap occurs or the viscosity that limits the fields in Figure 3. This is because for ΔT_0 above critical, warm magma diapirs will cool faster than cool diapirs and their temperatures tend to converge rapidly as they rise. For the input parameters of Figure 3 this critical value was found to be between 5 and 10°C. For temperatures below the critical value, the temperature gradient becomes so small that dykes will propagate from the very beginning of the calculations even for high viscosity magmas. For values around the critical value, small temperature differences may cause significant changes in the distance diapirs are able to rise before they go on to dykes, especially for small diapirs.

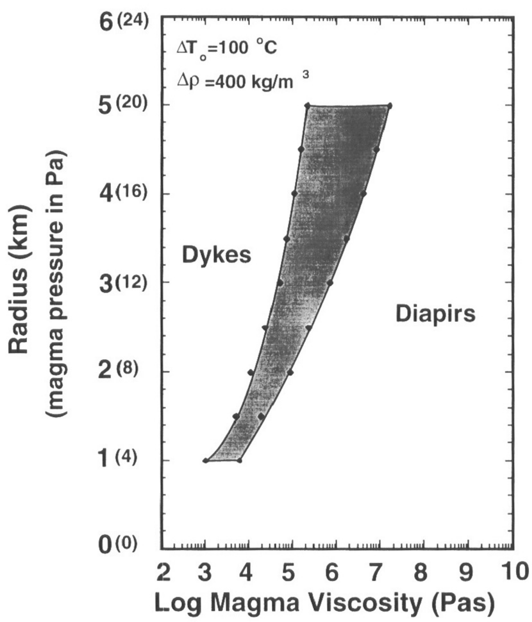

Figure 3 Results of calculations for spherical diapirs of different radii and magma viscosity rising through crust of Westerly granite rheology ($n = 1{\cdot}9$, $A = 2 \times 10^{-4}$ MPa^{-n}/s and $E = 1{\cdot}41 \times 10^5$ kJ/mol). 'Dykes' define conditions in which the magma leaves the spherical chamber at the initial set-up conditions; 'diapirs' defines conditions in which diapirs rise to their freezing depth without giving rise to successful dykes; and the intermediate field (shaded area) defines the conditions under which diapirs swap to dykes at some point during ascent. This field corresponds to viscosities of magmas of intermediate composition (10^3–10^6 Pa s, depending on magma pressure).

The rheology of Westerly granite used in these calculations is particularly suited for diapirs, as it yields a low viscosity when submitted to common magma buoyancy stresses. If stiffer rocks such as mafic granulites had been used, the domain in Figure 3 in which dykes dominate would have been larger, and perhaps only very viscous magmas would be unable to give rise to dykes. Similarly, the spherical initial geometry of the diapir gives rise to relatively fast initial velocities and steep temperature gradients that inhibit dyke propagation. If the initial geometry of the buoyant body was a layer or an oblate ellipsoid (of long horizontal dimensions), its initial velocity would be slow and the temperature gradient at the initial stages would be shallow and favour dykes. However, this may be partly counteracted by a decrease in magma pressure for these other geometries (remember $\beta^* \approx p^{n-5}$). Another unexplored possibility is the propagation of dykes from the sides of diapirs. These areas are warmer than the top (wider thermal boundary layer) and dykes could be favoured.

These results should be seen as a limited application of Rubin's results. Several important aspects have not been considered such as the increase in magma viscosity as the magma cools; the faster cooling of the magma close to the margins; the changes in the relation between excess magma temperature (ΔT^*) and temperature gradient (dT/dz) that controls the length of dykes as the diapir rises; changes in magma pressure due to solidification and gas exsolution; and the improved propagation of dykes by the transverse flow of magma within the dyke induced by rising bubbles or boundary roughness (Carrigan *et al.* 1992). Despite the simplifications the results corroborate what geological observations had already suggested, and what Rubin (1995) concluded: that the transport of felsic and mafic magmas may differ simply because of differences in magma viscosity; it also explains the often observed radial pattern of dykes around felsic, elliptical plutons (Paterson & Vernon 1995). These results also add that the ascent of magmas as diapirs may be suppressed at some point by the successful propagation of dykes.

3. Discussion

3.1. Shear zone control on emplacement and ascent location

The suggestion that faults control magma ascent (e.g. Hutton 1988; Hutton *et al.* 1990) is appealing because of the common association between regional-scale faults and plutons and because faults have the potential of creating, through extension, the necessary space for the emplacement of large magma volumes. However, there are still several aspects of this relation that remain unclear.

Before discussing unsolved problems, it is worth discussing recent developments regarding space for magma emplacement. Although numerous plutons have been shown to be a collection of sheets (e.g. Hutton 1992; McCaffrey 1992), the largest volume of felsic magmas resides in plutons that better fit models of steady-state magma chambers. If extension or the heat supply to emplacement levels is not fast enough, each magma batch will freeze before the next one arrives, resulting in a collection of frozen sheets (Sleep 1975). It was unknown until recently whether extension along faults in the continental crust is sufficient to open space rapidly enough to give rise to a steady-state felsic magma chamber in cold crust. Hanson and Glazner (1995) showed that this is possible and that it depends on several parameters such as the heat content of the magma (the difference between the initial temperature and solidus temperature, latent heat and heat capacity), the undisturbed temperature of the surroundings at emplacement level, heat diffusivity, the extension rate and the volume of the initial magma chamber.

Fracture resistance of elastic (and viscoelastic) rocks can be neglected for wide, self-propagating dykes (Lister & Kerr 1991; Rubin 1993a). Short initiating dykes, however, may greatly depend on pre-existing cracks to grow to the size where they become independent and self-propagating (Rubin 1993a). However, it seems unlikely that cracks would remain open in the lower crust, particularly in the neighbourhood of hot, partially molten zones, because the stress around the fracture would cause rocks to flow and close it. Faults in the lower crust are therefore unlikely to provide cracks that could be used by growing dykelets.

As rocks tend to soften when strained (Drury *et al.* 1991), active or inactive shear zones cutting across a partially molten zone provide potential low pressure sites that may act as melt sinks, focusing magma migration (Fig. 1). The extent to which such low pressure sites might influence melt migration is unknown, but a shear zone a few kilometres wide may be able to drain magma from a large volume of surrounding rocks. Whatever the preferred mechanism of magma ascent, the accumulation of magma in the shear zone within the source can control the site where magma ascends, explaining the often observed spatial closeness between plutons and faults. If the magma is unable rise as dykes (as discussed earlier), the low viscosity of the shear zone associated with a high melt concentration may lead to early doming and diapirism close to the shear zone.

Finally, and perhaps most importantly, the interaction between the melt and an active shear zone is poorly known. Would an active shear zone attract or repel melt? Does the high strain rate during shearing expel melt out of the shear zone, as in the experiments of Dell'Angelo and Tullis (1988)? Or would the permeability within the zone be so much higher than the surroundings that melt flows towards the zone and is flushed vertically? If movement along the fault is not constant, what happens to the melt during faulting and during repose intervals?

In summary, the opening of magma chambers by faults is plausible, but is likely to be a long drawn out process and to generate sheeted intrusions when extension or heat input are slow. Although pre-existing fractures may help short dykes to get started, large dykes are independent of any crustal weaknesses. The main point here regarding the influence of shear zones is that rather than controlling the ascent mechanism, low viscosity shear zones cutting across the magma source may control the site of magma ascent by focusing melt accumulation in the source and providing a favourable site for the initiation of diapirs or dykes. The close spatial relationship between plutons and shear zones does not necessarily imply that the latter controlled either magma ascent or emplacement.

3.2. Diapir pathways

Diapirs are expected to leave along their path characteristic structures that should be easily recognisable in the field. An argument that has been raised against diapirism is the lack of field examples that indicate their paths through the crust (see Clemens & Mawer 1992; Petford, this issue). I argue here that the lack of such field examples may be partly due to the difficulty for structural geologists to assign observed structures to diapirs that are no longer there, and partly due to the simplified picture of diapirism emerging from models, compared with the structural complexities arising from diapirs travelling through heterogeneous crust that may undergo syn- or post-emplacement deformation.

As pointed out earlier, it seems that diapirs require very special (unusual?) crustal conditions to rise to shallow crustal levels (< 10 km). Typically, diapirs would rise to the 10–20 km depth range, so that their pathways may only be exposed in deep crustal sections. Not only are these sections rarely exposed, but they very often present complex deformation patterns. The results of simplified numerical and laboratory models suggest that the structures at the tail of a diapir should be a vertical narrow cylinder of intensely sheared rocks with subvertical lineations, showing a diapir-up sense of shear and an increase in metamorphic temperature towards the centre where granitic rocks could be found. The cylinder diameter should be of the order of the diapir diameter, i.e. a few kilometres or less depending on the power law exponent of the wall rocks, and sometimes a rim syncline could have developed (see Weinberg & Podladchikov 1995). The *P–T–t* history of a rock sample in the aureole should be one of fast heating followed by slow cooling, accompanied by accelerating decompression (as the diapir approaches the sample), followed by decelerating decompression (as the diapir leaves it behind). If the crust is undergoing deformation during diapirism, the hot and sheared tail of diapirs might localise and rapidly become overprinted by tectonic strains. Even after the tail cools, strain softening due to earlier shearing in the tail might localise later strains. A possible result of combining the deformation caused by the passage of diapirs and contemporaneous or later deformation is the development of large kilometre-scale steeply plunging sheath folds. A possible example of a diapir tail is the Vrådal granite in Norway (Fig. 4; Sylvester 1964), where several of these features have been described.

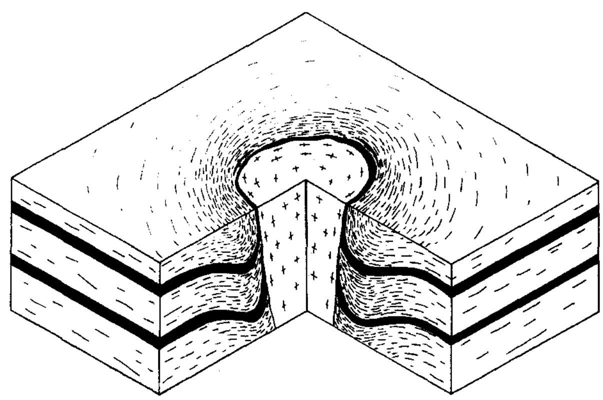

Figure 4 Vrådal pluton in Norway (redrawn from Sylvester 1964) showing structures typical of those expected inside and around diapirs or their tails, such as the circular shape, the concentric pattern of foliation increasing towards the contact and the rim synform in the wall rocks.

3.3. Why do felsic plutons balloon?

Apart from the main question as to how felsic dykes begin, two very puzzling questions remain to be answered by the proponents of felsic transport mainly through dyke. Firstly, as pointed out by Rubin (1995), why do granitic magmas balloon when many large basaltic magmas very often seem not to (e.g. Muscox, Canada and the Skaergaard, Greenland)? This is particularly puzzling because the heat content of basaltic magmas is much higher than that of granitic magmas, and they should therefore be able to soften and viscously deform their surroundings more easily than felsic magmas. A possible answer for that question suggested by R. Kerr (pers. comm.) is that the single magma batch that formed Skaergaard filled the chamber on too short a time-scale too allow viscous deformation of the surroundings. A second point, brought up by Paterson (pers. comm.) is why should granitic magmas be able to deform their surroundings at the initially cold emplacement level viscously, but be unable to deform the hot and soft lower crust viscously?

4. Conclusions

The main conclusion of this discussion is that the transport of felsic magmas through the crust is most likely to rely on a combination of mechanisms, which may simultaneously or sequentially control ascent and emplacement. A single mechanism seems to be unable to overcome all physical barriers as well as explaining the common features of plutons and batholiths (Paterson & Fowler 1993; Petford, this issue). A single diapir seems to be unable to reach shallow crustal levels and their pathways through the crust are yet to be clearly described. Felsic dykes, on the other hand, may be unable to get started because freezing may close the dykes faster than they propagate. The ideal conditions for dyke initiation occur in extensional environments, when melt segregation gives rise to a dense interconnected network of veins that drains within the source into a few large high pressure dykes. Dykes may be inhibited by shortening environments associated with segregation mechanisms that are unable to give rise to large dykes (e.g. porous flow or sparse disconnected veins). If dykes are unable to drain the source, or if the drainage is too slow compared with viscous deformation of the source and surroundings, then diapirs may dominate. The influence of shear zones in the initiation of dykes is unclear, but the low viscosity expected in those zones is likely to focus melt migration in the source and provide a favourable site for the initiation of both dykes and diapirs.

The results of calculations of the temperature evolution of diapirs and surroundings suggest that diapirs of intermediate composition may start their ascent as diapirs and swap to dykes as they slow down when reaching stiffer rocks. Large felsic diapirs may also undergo this swap, but more generally felsic magmas will tend to rise as diapirs whereas mafic magmas rise through dykes. Future work should concentrate on finding examples of diapir pathways by taking into consideration the possible interaction between diapirs and syn- or post-emplacement regional deformation. There is also a need to understand better the part that melt segregation may play in initialising dykes, and to understand the interaction of active and inactive shear zones with segregation and dyke initiation. Most importantly, we need to know more about the interplay between the several mechanisms that might enhance or inhibit the transport of felsic magma through the crust.

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