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Mesoscale pervasive felsic magma migration: alternatives to dyking

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Abstract

This paper reviews the literature on dyking as a mechanism of felsic magma extraction from a source and transport to shallower crustal levels, and review the recent literature suggesting a range of alternative mechanisms of magma migration in hot crustal zones which produce mesoscale pervasive granite sheet intrusions. Recent papers have strongly favoured dyking as the main mechanism controlling magma migration. However, the initiation of dykes from a felsic magma source is fraught with difficulties, even when magma is immediately available for transportation, as in magma chambers. Within a partially molten source, magma may reside in a range of structures with a wide range of shapes, sizes and degrees of connectivity. Whereas the growth of individual dykes within a partially molten zone, and the self-propagation of large dykes into subsolidus crust, have both been studied in some detail, little attention has been given to the crucial intermediate step of the growth of a dyke network capable of producing wide crustal scale dykes. The rarity of granite dyke swarms suggests that, if dyking is the preferred mechanism of magma transport, felsic magma sources produce only few major transporting dykes during their lifetime. Alternatively, dyking is not an important mechanism. The parameters controlling the volume of the catchment drained by one such dyke, as well as other basic geometrical parameters controlling the structure of the dyke network within the source, are unknown. The ability of dyking to drain a partially molten source depends crucially on these variables and particularly on the horizontal permeability of the source. The slow velocity of viscous felsic magmas traveling in rock pores implies that magma drained during dyking is mostly that previously extracted from the pores, and resident in irregular magma bodies or dyke networks. The observation that large volumes of buoyant magma are commonly present in migmatite zones, and that dyking in these zones plays a secondary role, suggest that dyking is inefficient and is able to extract only a fraction of the total melt available in the source. In support of this conclusion, recent detailed studies of exhumed hot crustal zones have revealed a range of alternative migration mechanisms characterized by mesoscale pervasive magma flow (outcrop scale as opposed to porous flow). Pervasive migration gives rise to magma sheets preferentially emplaced parallel to high-permeability zones such as foliation or bedding planes. Apart from local dyking, three alternative mechanisms have been proposed to account for pervasive migration of magmas, namely tectonic pumping; magma wedging into low-viscosity rocks; and volatile-driven intrusion. Because of their unfocused, pervasive character, these mechanisms are restricted to hot crust where magmas are not exposed to rapid freezing. A model is proposed whereby heat advected with

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the magma pushes crustal geotherms upwards, allowing pervasive magma migration to shallower depths.

Keywords: Magma extraction; Tectonic pumping; Magma wedging; Pervasive magma migration; Dyking; Felsic magma

1. Introduction

Whereas granitoid intrusions are generally found as low aspect ratio bodies (plutons), mafic intrusions are typically found as large aspect ratio bodies (e.g., dykes, sills and laccoliths). Dyke swarms are also commonly associated with regions of mafic magmatism, but are rare in regions of felsic magmatism. These differences have been ascribed to the much higher viscosity of felsic magmas compared to that of mafic magmas. In the past, felsic magmas were thought to be incapable of travelling long distances in dykes because, being stiff, they travel slowly and freeze by rapid heat loss to cold country rock. A popular alternative then was the rise of large buoyant bodies, by viscously deforming the surroundings (diapirs). The large volume and compact shape of diapirs protect magma from rapid freezing, and diapirs can directly account for the common shape of plutons and their associated structures (e.g., Weinberg and Podladchikov, 1994, 1995).

In the late 1980s, early 1990s, there was renewed interest in the physical processes that lead to dyking. Some studies concentrated on the mechanisms of dyke growth within partially molten rock (e.g., Ribe, 1986, 1987; Wickham, 1987; Sleep, 1988; Stevenson, 1989), and some concentrated on the conditions required for self-propagation of dykes through subsolidus crust (e.g., Bruce and Huppert, 1989, 1990; Emerman and Marrett, 1990; Lister and Kerr, 1991; Rubin, 1993a,b, 1995). These new studies led to papers reassessing the ability of felsic dykes to propagate through cold crust (e.g., Clemens and Mawer, 1992; Petford et al., 1993, 1994). These found that, for likely values of magma viscosity and pressure gradient, felsic magmas could easily give rise to self-propagating dykes after achieving a given critical width, w_c, normally of the order of a few meters. Within such dykes, magma velocity is such that a single dyke may propagate through the crust and feed a voluminous pluton in a very short time (days or months, e.g., Petford et al., 1993).

For the proponents of dyking of felsic magmas, the low aspect ratio shape of plutons results from the focusing of flow in dykes to vertical cylinders, associated with the ballooning of a magma chamber at an appropriate crustal level, producing the shortening structures commonly observed around plutons. It is further argued that feeder dykes are seldom described because they are squeezed and narrowed after melt stops flowing, resulting in modest dykes underneath plutons, seldom seen and often disregarded (feeder dykes have been described in the Chemehuevi Mountains (John, 1988), the Bergel pluton (Rosenberg and Heller, 1997), and in the Gangotri pluton (Scaillet et al., 1996)).

The apparent efficiency of dykes in transporting felsic magma led to a surge of papers in the 1990s proposing that dykes account for magma transport. However, very little attention has been given to a crucial step in the evolution of dykes: the growth of a connected network of tributary dykes within the source, necessary to drain the magma and feed a transporting dyke. The arguments for dyking of felsic magmas overlook two important processes: (a) felsic dykes may freeze during early growth stages before attaining critical width (Rubin, 1993a; Weinberg, 1996); and (b) the network of dykes and magma pools in the source may be unable to provide the flow rate required to maintain the critical width of dykes.

Counterbalancing this tendency, a number of more recent papers describe in situ and injection migmatites where melt migration took place through a complex network of channels by a range of alternative mechanisms where dyking played only a minor role. In these papers, the term *pervasive migration* is used to describe regions where numerous magma sheets and pools intrude country rock on outcrop scale. In this paper, I aim to review present knowledge of dyking of felsic magmas, and to review alternative mechanisms of magma migration which, like dykes, give rise to magma sheets. Finally, I will propose a mechanism whereby mesoscale pervasive magma migration may heat up the crust allowing migration to increasingly shallower crustal levels.

2. Dyking

The term *dyking* refers to the specific process of elastic cracking of country rock by tensional stresses concentrated at the tips of magma-filled fractures (e.g., Lister and Kerr, 1991) that gives rise to dykes. Not all described magma sheets result from dyking; many sheet-like bodies may arguably result from other processes. Here, the term 'transporting dyke' is used for large dykes able to transport large volumes of magma to shallow crustal levels. The term 'tributary dykes' is used here for dykelets, veins and sheets in the magma source that will feed into the transporting dyke. Two distinct steps control magma extraction from the source by dykes: the first is the extraction from rock pores into tributary dykes, and the second is drainage of the tributary dykes by the successful departure of a transporting dyke away from the source.

The variables controlling the structure of the dyke network in the source (its geometrical arrangement) are the size distribution of dykes, their width and aspect ratio, the average spacing between dykes of a similar size, connectivity of the network, the critical width of transporting dykes, the volume and aspect ratio (shape) of the source, whether the system is self-similar and, if so, the order of the system and its length and bifurcation ratios, (see Fig. 1 and Horton, 1945, for definition of these ratios) and the *diameter exponent* Δ, defined as the downstream widening of the channel network (see below; Karlinger et al., 1994).

In microscopic scale, the factors controlling the spacing, length and width of magma-filled cracks are a function of several parameters including volume change of the melting reaction, rate of melting, and magma viscosity (Rushmer, 1996). In outcrop scale, spacing between dykes may be controlled by complex interaction between the compaction length (Stevenson, 1989), tectonic stresses and rock anisotropy. Unfortunately, the structure of the dyke network in the source is poorly known, and field evidence of microscopic fracturing during melting is generally erased by later crystallization. Moreover,

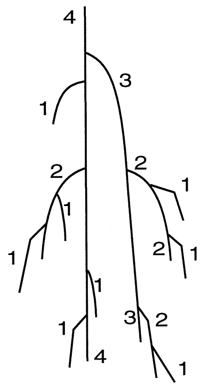

Fig. 1. A schematic drainage system, where the streams are numbered according to their orders. Stream orders are defined on the basis of their branching or bifurcation, and the smallest unbranched stream is given order 1; second order streams receive only 1st order ones, but third order streams may receive 1st order streams directly, but must receive at least one 2nd order stream, and so on (Horton, 1945). A stream is of the same order throughout its length. Horton (1945) defined the bifurcation ratio, r_b, as the ratio between the number of bifurcations of a stream of a given order, to that of the next lower order. The stream length ratio, r_l, is the ratio of the average length of streams of a given order to that of streams of the next lower order. He found that for river drainage systems, these ratios are constant, defining a self-similar (fractal) system.

preserved dyke width is only indirectly related to the width during magma transport. Narrow frozen dykes may have been important transporting dykes, drained at the end of magma transport.

3. Dyke growth from a chamber

The volume flow rate of magma through a dyke is determined by its width and magma velocity, a function of magma viscosity and driving pressure gradient (e.g., Lister and Kerr, 1991). Studies determining the critical width of dykes, conveniently assume that the source is able to provide whatever magma flow rate the dyke requires. This implicitly assumes that large volumes of magma are readily available for transportation.

Rubin (1993a), in contrast to workers before him, asked how dykes grow to the size at which they become successful in propagating through cold surroundings. He studied in detail the physical processes involved in generating a dyke from an initial crack at the top of a magma chamber. For the discussion here, his main result was to show that successful dyke propagation, avoiding early freezing, is restricted to a relatively special range of conditions, requiring a combination of low temperature gradient away from the magma chamber/source, high magma pressure and low magma viscosity. Weinberg (1996) applied Rubin's results to investigate dyke propagation from the top of a rising diapir and found that low-viscosity diapirs will be easily drained by the propagation of dykes from their top, whereas high-viscosity diapirs are unable to give rise to such dykes. Intermediate viscosity magmas will initially rise as diapirs, but will be drained by dykes during ascent.

Rubin (1993a) also suggested that dyke propagation is led by the exsolution of volatiles into the low-pressure zone that characterizes the tip of a propagating dyke, where viscous magma is unable to reach in short time scales. As the volatiles fill the tip, pressure increases, leading to further cracking, crack widening and forward propagation of the dyke. He envisaged that where preserved, the typical tip of granitoid dykes would be formed by pegmatites, and the main dyke body would retain a narrow film of pegmatites resulting from the early exsolution of volatiles (Fig. 2).

The important point here is that even in the simple case where magma is immediately available for transportation, such as in a magma chamber, dyking requires a somewhat special set of physical conditions to grow without freezing. Low melt-fraction sources impose additional difficulties on the growth of transporting dykes.

4. Dyke growth from a partially molten source

Many granitoid bodies are believed to have resulted from a low degree of partial melting of the source rock ($<35\%$ melting, e.g., Himalayan leucogranites, Harris et al., 1995; see also Clemens and Vielzeuf, 1987), or from the gradual extraction of melt from the source (fractional melting). In these cases, the source behaves as a solid mass with interstitial liquid, and the magma flow rate into propagating transporting dykes depend on the geometry and dynamics of magma segregation and flow in the source (e.g., Petford, 1995).

There are two important phases characterizing the evolution of the magma network in the source: the growth of the network before propagation of a transporting dyke; and the drainage of the network during its propagation. The former is characterized by in-

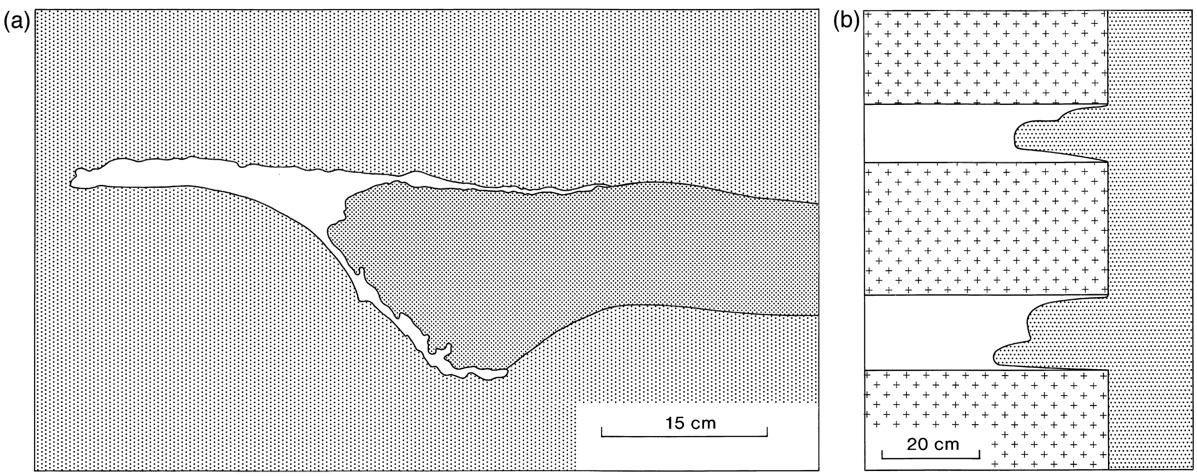

Fig. 2. (a) Frozen tonalite dyke tip filled by pegmatite. The pegmatite tip is interpreted as the low-pressure leading wedge of the dyke during propagation, into which volatile-rich fluids have been exsolved. The tonalite dyke intrudes lighter tonalite in Bingie–Bingie point, NSW, Australia. (b) Dykes in granite (crosses) of the Ladakh Batholith, Indian Himalayas. Two pegmatite dykes (white) originate from the granodiorite dyke on the right-hand side (stippled). Granodiorite magma is dragged into the base of the pegmatite dykes forming bulges, but were unable to flow into.

creasing numbers of dykes, accompanied by their increasing size, width and connectivity (Gueguen and Dienes, 1989). This phase determines the size of the connected network and the total amount of magma available for extraction by a transporting dyke. The second phase starts when the system matures to produce a transporting dyke; magma is then drained out of the system. The volume and duration of this extraction event depends on the structure of the network, more specifically, on its ability to provide the transporting dyke with the magma flow it requires to maintain the critical width.

I reviewed above the difficulties that a dyke may have when propagating from a magma chamber, where magma is immediately available for transportation. I am now going to investigate the factors limiting dyke growth within a partially molten source. There are two important limitations, both related to the high viscosity of felsic magmas: (a) the dyke has a limited ability to extract magma from its surroundings because of the short compaction length (e.g., McKenzie, 1984; Wickham, 1987); and (b) slow migration of pore melts leads to slow dyke infilling rates (Wickham, 1987). As a result, a transporting dyke requires a large network of connected dykes in place in the source. (For an extensive review of segregation mechanisms of granitic melt from rock pores, see the review by Brown (1994) and references therein.)

4.1. Compaction length

The low pressure within dykes and veins imposes a pressure gradient on the surrounding rocks that causes the melt to flow towards the dyke (Ribe, 1986; Sleep, 1988; Stevenson, 1989). This pressure gradient is in the main dissipated within a distance equivalent to the compaction length, δ_c (McKenzie, 1984, 1987):

$$\delta_c^2 = k(\lambda_2 + 2\eta_2)/\eta_1, \qquad (1)$$

where η_1 is the melt viscosity, and λ_2 and η_2 are the Lamé coefficient of viscosity and the shear viscosity of the grains, respectively. One of several definitions of rock permeability, k, and the one assumed here is that $k = Ca^2 f^3$ (from Sleep, 1988), where the constant $C = 10^{-3}$ (from McKenzie, 1984), a is the grain size and f is the melt fraction. For example, for $a = 1$ mm, $\eta_1 = 8 \times 10^5$ Pa s (a typical value for granites, e.g., Petford et al., 1993), $\lambda_2 + 2\eta_2 = 10^{18}$ Pa s (following Sleep, 1988), and melt fraction, $f = 0.2$ permeability is $k = 8 \times 10^{-12}$ m^2 and a compaction length is $\delta_c = 3$ m (or 9 m for $f = 0.4$). The small value of the compaction length for high-viscosity felsic magmas limits the ability of dykes to drain melt from its surrounding (e.g., Ribe, 1986; Wickham, 1987; Sleep, 1988; Stevenson, 1989). A more comprehensive study of the compaction length for granitic systems may be found in Petford (1995), however, he used a larger value of the constant C (~ 0.022), resulting in larger compaction lengths than determined above.

4.2. Slow porous flow

As noted by Wickham (1987), porous flow of felsic magmas is extremely slow due to their high viscosity. The typical growth rate of a dyke by porous flow of interstitial magma, U, may be determined for a given pressure reduction imposed by the dyke (Sleep, 1988):

$$U = \frac{\nabla P k}{\eta_1}, \qquad (2)$$

where ∇P is the pressure gradient in the immediate surroundings of the dyke. Here, I assume arbitrarily that ∇P is one order of magnitude higher than the pressure gradient created by magma buoyancy:

$$\nabla P = c_1 \Delta\rho g = 10 \Delta\rho g.$$

This assumption ensures that melt flows towards the dyke rather than being controlled by magma buoyancy. For the values of k, density difference between melt and matrix $\Delta\rho$, and η_1 given in Table 1, and gravity acceleration $g = 10$ m/s^2, the rate of dyke infilling is $U \sim 2 \times 10^{-13}$ m/s (or ca. 1 m/Ma). Wickham (1987) found much higher infilling rates during extensional fracturing of the source rocks by using a much higher $\nabla P = 10$ MPa/m, suggested by Etheridge et al. (1984). Petford (1995) found using similar equations that dyke infilling may be as fast as 1 m/a in extremely favourable conditions. More

Table 1
Notation

Symbol	Meaning	Values	Dimensions
a	mean grain size	10^{-3}	m
A	average aspect ratio of cracks in Eq. (5)		none
A_t	total surface area of dyke network		m (in 2D)
C	constant in permeability equation	10^{-3}	none
c	crack radius in Eq. (5)		m
c_1	constant	10	none
f	porosity	0.2 (or 0.4)	none
f_1	fraction of connected cracks in Eq. (5)		none
g	gravity acceleration	10	m/s^2
k	permeability	8×10^{-12}	m^2
l	average crack distance in Eq. (5)		m
n	total number of dykes		none
p	probability that two cracks intersect, Eq. (7)	0 to 1	none
Q_i	volume flow rate through dykes of order i		m^2/s (in 2D)
Q_p	volume flow rate from pores to dyke network		m^2/s (in 2D)
Q_{td}	volume flow rate of magma in transporting dyke	0.06	m^2/s (in 2D)
t	time		s
$\overline{U}$	average magma velocity in dyke	0.01	m/s
U	rate of dyke infilling	2×10^{-13}	m/s
V	magma volume	1.2×10^4	m^2 (in 2D)
w_c	critical dyke width	6	m
w	dyke width		m
W	width of melt source	2000	m
δ_c	compaction length	9	m
Δ	diameter exponent (Karlinger et al., 1994)	3	none
$\Delta\rho$	density difference melt − matrix	200	kg/m^3
∇P	pressure gradient		Pa/m
$(\lambda_2+2\eta_2)$	matrix viscosity	10^{18}	Pa s
η_1	magma viscosity	8×10^5	Pa s

typically, however, U is slow, and the widening of dykes is constrained by either short compaction length or slow porous flow. Despite these limitations, dykes may still widen upwards as a result of the upward flow of magma from connected dykes.

4.3. Tributary dyke network

In this section, the relation between the width and magma flow rates in transporting dykes, and those in the tributary dyke network will be investigated. This relation is dictated by the need to match the magma flow rate within transporting dykes to the flow rate provided by the source network. Here, the source will be treated as a two-dimensional system.

The average velocity, $\overline{U}$, and flow rate of magma in transporting dykes are (Petford et al., 1993):

$$\overline{U}=\frac{\Delta\rho g w^2}{12\eta_1} \tag{3}$$

$$Q=\overline{U}w, \tag{4}$$

where $\Delta\rho g$ is the pressure gradient that drives melt flow. A typical value of the critical dyke width of a transporting dyke, w_c, and its flow rate, Q_{td}, may be found by using the values for η_1 and $\Delta\rho$ in Table 1 and Fig. 2 in Petford et al. (1993). For these, $w_c=6$ m, $\overline{U}=0.01$ m/s and $Q_{td}=0.06$ m^2/s.

When two dykes meet, the width of the dyke downstream from the junction is determined by:

$$Q = Q_1 + Q_2$$

because the flow rate Q in a dyke depends on the cube of its width:

$$w^3 = w_1^3 + w_2^3.$$

The exponent 3 in this equation corresponds exactly to the *diameter exponent* Δ, used by Karlinger et al. (1994) for river networks to define the stream widening process. In dyke networks, the diameter exponent $\Delta = 3$ may be increased somewhat by flow from pores into the dykes, and may be decreased by tectonic pressures opposing dyke widening (Δ will be larger in terranes undergoing extension than those undergoing shortening). Assuming a value for $\Delta = 3$, it is possible to determine the number of dykes (n) of a certain width necessary to feed a transporting dyke:

$$n = (w_c/w)^{\Delta} = (w_c/w)^3.$$

For example, the number of 1-m wide dykes necessary to feed the 6-m wide transporting dyke with its required volume flow is $6^3 = 216$.

In the hypothetical case that this transporting dyke is fed only by these 1-m wide tributary dykes, they would need to contain sufficient magma to sustain flow throughout propagation of the transporting dyke. For example, the 6-m wide dyke will take 2×10^6 s to propagate across 20 km of crust, and the 1-m wide dykes would need to be at least 600 m long to sustain its propagation. Many more and much longer tributary dykes are required if the transporting dyke feeds a shallow-level pluton.

Clearly, this oversimplified network does not portray the complexity of natural tributary networks. These can be more realistically compared with a river drainage basin, formed by an array of connected streams (possibly starting at grain-size scale) coming together to feed the transporting dyke. In this case, the most important variables controlling the ability of the network to produce transporting dykes are: (a) the dimensions of the basin drained by a single transporting dyke (not necessarily the dimension of the source); (b) the average width and length of the smallest subset of dykelets able to feed the system with a flow rate compatible with that of a transporting dyke (a function of the number of dykelets, and of magma velocity within them); and (c) the structure of the source including whether the system is self-similar. For self-similar systems, (c) includes parameters such as dyke density distribution, bifurcation ratio r_b, dyke length ratio r_l, diameter exponent Δ, and maximum dyke order (these are directly analogous with parameters determined for river basins; Horton, 1945 and Karlinger et al., 1994). At present, none of these variables have been determined in crustal migmatites.

4.4. Permeability and connectivity

As shown by Gueguen and Dienes (1989) and stressed by Petford and Koenders (1998), it is the *connectivity* and resulting *permeability* of the dyke network that control the ability of dykes to drain the source. In a study of the ability of rocks to transport fluids through microcracks, Gueguen and Dienes (1989) showed that the permeability of a cracked rock depends on three independent variables: average radius of disc-shaped cracks c, average crack aperture w, and average crack spacing l. These results may be applied to a dyke network, where a dyke may be considered simply as a magma-filled crack. Permeability, k, of a crack (dyke) system may be defined as (simplified from Gueguen and Dienes, 1989):

$$k = \frac{4\pi}{15}\frac{A^3}{l^3}c^5 f_1, \tag{5}$$

where

$$A = w/c \tag{6}$$

is the average crack aspect ratio and f_1 is the fraction of connected cracks. f_1 is a function of p, the probability that two randomly oriented cracks intersect (see Gueguen and Dienes, 1989, for further details on the relation between f_1 and p):

$$p = \frac{\pi^2}{4}\frac{c^3}{l^3}. \tag{7}$$

Thus, as a melt source evolves and the number (average spacing) and length of dykes increase, the probability of them connecting increases (Eq. (7)). For $p=1$, all cracks are connected to an effectively infinite network ($f_1=1$), and the permeability reaches a maximum. Fig. 3 shows that permeability is zero below a critical value of p (p_c) called the percolation threshold, and that there is a rapid increase in k over small changes in p at around 0.5, tending asymptotically to a maximum value as p approaches 1.

The difficulty in directly applying these equations to magma sources lies on the wide variation in size and average spacing of dykes (from grain size to kilometer scale). A step towards solving this problem is to organize the dyke network into dyke orders (Fig. 1 and Horton, 1945) and apply the above equations for individual orders (comprised of dykes of similar dimensions). However, without knowledge of the relationship between different orders (the bifurcation and length ratios mentioned above), there is no direct way of calculating the permeability and volume flow through the system to the transporting dyke. For example, if permeability and volume flow through an extensively connected network of dykelets of one order, feed a few poorly connected dykes of higher order, the system will stagnate.

Nevertheless, the above equations may be used to determine the minimum crack size (c and w) and separation (l) necessary to feed a transporting dyke. This can be done by assuming a network of similarly sized dykes (dykes of a given order), and that they have reached maximum permeability ($p=f_1=1$).

The minimum crack size may be found by determining the conditions for which the flow rate in the

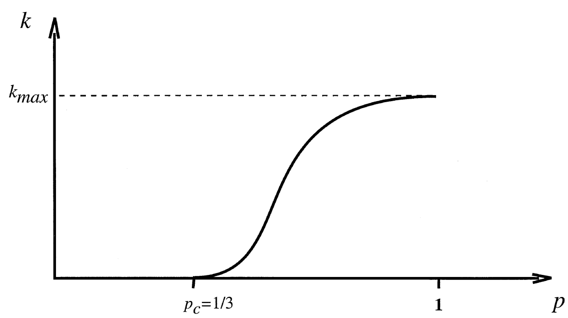

Fig. 3. Plot of the permeability, k as a function of the probability, p, that two cracks are connected. p_c is the percolation threshold below which crack connectivity is unable to produce permeability (from Gueguen and Dienes (1989)).

transporting dyke, Q_{td}, equals that within the tributary dykes of a given size or order Q_i:

$$Q' = \frac{Q_{td}}{Q_i} = 1, \tag{8}$$

where Q_i in 2D is of the order of:

$$Q_i \sim kWg\Delta\rho/\eta_1,$$

where W is the width of the source zone. Rewriting Eq. (8)

$$Q' \sim \frac{\bar{U}w_c}{kWg\Delta\rho/\eta_1}. \tag{9}$$

Making $p=f_1=1$ in Eq. (5), Eqs. (3), (5) and (7) may be used to rewrite Eq. (9) as:

$$Q' \sim \frac{1}{4}\frac{w_c^3 c}{Ww^3} = 1. \tag{10}$$

For the dimensional values previously used in this paper, and for an arbitrary $W=2000$ m and $A=0.1$, the minimum dimensions of a fully connected tributary dyke network capable of producing a 6-m wide transporting dyke is of the order of: $c \sim 5$ m and $w \sim 0.5$ m. The distance between dykes, $l \sim 7$ m, can be found for $p=1$ using Eq. (7).

These values must be taken with caution because of the assumptions underlying the equations used. The first assumption is that the dykes are isotropically distributed, which will tend to increase the probability of dyke intersection, and therefore increase permeability. The second and more important assumption is that the dimensions and spacing of the cracks are given parameters of the system, produced independently of the evolution of the system itself. This assumption most likely does not represent the dynamic evolution of a dyke network where the dimensions and spacing of dykes are a function of the evolution of the network. This dependency may lead to a self-organized system, where drainage is maximized and energy expenditure optimized (Rinaldo et al., 1993) leading to higher permeability. The example of Section 4.3 of a few hundred 1-m wide dykes feeding a transporting dyke, shows how organized dykes that are allowed to widen as they link up (to the width of the transporting dyke), may allow faster flow. Conversely, poorly connected networks will be unable to give rise to large dykes, in which case, magma will remain within the source or its immediate surroundings and migrate slowly and locally, driven by magma buoyancy and tectonic pressure gradients.

4.5. Flow rate

Another way of approaching the problem of melt extraction by dykes is by studying the relation between the flow rate from the pores into the dyke network, and the flow rate out of the network. These two rates may be used to define another dimensionless magma flow rate:

$$Q' = \frac{Q_{td}}{Q_p} = \frac{\overline{U}w_c}{UA_t} = \frac{1}{12}\frac{w_c^3}{c_1 kA_t}. \tag{11}$$

where A_t is the total surface area of the connected dyke network (it excludes the area of permeable pore network). A_t varies by several orders of magnitude, and depends on a range of assumptions about the structure of the system. As above, Q' is independent of the magma properties (for a given w_c): it depends only on the structure of the network. For $Q'=1$, the flow into the system equals the flow out of the system. The transporting dyke will tap melt from the pores, via the network of tributary dykes, and the system is in steady state, as long as U remains constant. When $Q'=1$, the tributary network plays only the role of magma pathway. However, because the ratio $\overline{U}/U$ is generally very large, $Q'=1$ requires an extremely large A_t combined with large permeability. For the values of w_c, c_1 and k in Table 1, $A_t = 2.5\times10^{11}$ m, indeed a spectacular number (if such a network was on average a narrow 0.1 mm, it would store ca. 12 km^2 of melt, in 2D, and would have drained at least 60 km^2 of source).

Q' values larger than one imply that more magma is extracted by the transporting dyke than is fed to the tributary network. Such systems rely on magma stored in the tributary network and for large Q' values ($Q' > e$), flow from the pores into the system may be disregarded. The requirement of a very extensive connected network, suggests that in most cases, transporting dykes rely on pre-extracted magma, resident in dykes.

Because A_t controls the time for infilling of the tributary network, it controls also the time evolution of the system and thus the chemistry of magmas. Two variables control A_t, the density distribution of dykelets and the total size of the drained reservoir. Systems with large A_t values, will be rapidly filled by magma, and assuming rapid connection of the network, will rapidly produce transporting dykes.

The value of A_t may only be guessed at present. However, I will show below that, if magma drainage systems are self-similar, accurate estimates can be made by extrapolation from outcrop scale observations.

4.6. Size of the reservoir

It is obvious from the discussion above that the problem of dyking is intrinsically related to the size of the reservoir drained by each transporting dyke. However, there are virtually no available constraints on this variable. It depends not only on the size and shape of the source zone, but also on the ability of the network to focus magma flow towards transporting dykes, which in turn depends on the connectivity and structure of the dyke network.

The volume of the reservoirs drained by each transporting dykes can be estimated for plutons where the number of feeder dykes (n), the melt fraction at the source (f), and the total volume of the source (V_s), or alternatively the total magma volume (V_m), are known. Taking the dyke swarm feeding the Gangotri pluton in the Himalayas as an example (Scaillet et al., 1996), and using the likely estimates of $n = 10^2$–10^3, $f = 0.2$ and $V_m = 150$ km^3, each dyke must have transported 0.15 to 1.5 km^3 of magma and drained 0.75 to 7.5 km^3 of source. The assembly time of the pluton cannot be found by estimating magma velocity and volume flow rate within individual dykes (10–50 m wide), but requires knowledge of the number of dykes leaving the source per unit time, an unconstrained variable.

4.7. Magma focusing and dyke swarms

In Section 4.4 the dimensions and average spacing of isotropically distributed dykes necessary to

Table 2
Magma extraction from a low melt-fraction source by dykes

(1)	Slow pore flow and short compaction length limits drainage of the country rock around a single felsic dyke.
(2)	A large number of connected tributary dykes is therefore necessary to feed transporting dykes.
(3)	The volume of magma extracted from pores during dyking is likely to be negligible. Transporting dykes drain mostly pre-extracted magma, either from chambers or from well-developed, mature tributary networks.
(4)	Fundamental variables controlling dyking are unknown. These are: the volume of the region drained by a transporting dyke and the structure of the drainage system. Is this system self-similar and if so, what are its bifurcation and length ratios, and the total surface area of the dyke network?
(5)	Few transporting dykes are produced by felsic sources (as implied by the rarity of felsic dyke swarms). Unless there is significant source-scale horizontal migration, dykes will extract only a small percentage of the magma in the source.

produce a fully connected network capable of sustaining the flow rate through a transporting dyke have been determined. In this case, a single dyke is capable of tapping magma from the entire system.

However, because dykes tend to orient themselves parallel with the direction of maximum shortening (e.g., Stevenson, 1989), a system of semi-parallel dykes will develop. If we consider an extensional environment, more conducive to dyking, dykes will tend to grow in the vertical direction. Vertical dykes will propagate upwards, driven by magma buoyancy, and connect with dykes growing further up within the source zone. If the source layer is sufficiently deep, the vertical drainage of a narrow region may be able to produce transporting dykes.

Source zones tend to follow the shape of geotherms, tending to be more extensive in the horizontal than in the vertical direction. Preferential orientation of dykes gives rise to strong permeability anisotropy. Vertical dyke systems in the source will lead to poor horizontal permeability, breaking up the drainage of the source into vertical slices which could potentially lead to numerous transporting dykes. However, felsic dyke swarms (i.e., numerous, similarly oriented, planar dykes) away from source zones are relatively rare. This is somewhat striking particularly because during a prolonged melting event, the source could potentially undergo several cycles of magma extraction.

The relative rarity of felsic dyke swarms is in stark contrast with common swarms occurrence in regions of basaltic magmatism. The compaction length in basaltic systems may be up to a few orders of magnitude higher than that of granitic systems. This implies that more dykes are required to drain a granite source than a basalt source. The rarity of felsic dyke swarms must therefore imply that, despite the need for a large number, few dykes are able to leave the source. These observations together suggest that felsic dykes are inhibited and that a well-connected system of tributary channelways must be in place before transporting dykes are able to develop.

Those arguing for dyking state that the rarity of felsic swarms results from the ability of a single dyke to rapidly drain the source and transport the entire volume of a typical pluton across the crust in a matter of days. However, it is equally possible that these rare dykes are only able to tap a small fraction of the source magma. It all depends on the undetermined structure of the source.

To summarize Section 4, the limited ability of felsic magmas to produce dykes is clearly exemplified by the comparison to basaltic systems. The significant consequences of the short compaction length and slow porous flow of felsic systems, on the development of the source and magma drainage, and are summarized in Table 2. Whereas single dykes may be able to transport large volumes of felsic magma through the crust if they tap large magma bodies of high melt fraction ($>$ 40–50% melt fraction), the difficulties encountered by dykes in draining melt from a low melt-fraction source cannot be overlooked.

5. Pervasive magma migration in hot country rocks

Studies of crustal magma sources, summarized in Brown and Solar (1998), suggest that magma migration in in situ migmatites is not concentrated to planar, dyke-like, connected sheets feeding transporting dykes. Rather, migmatites evolve through a network of channels and magma bodies of complex shapes and a range in sizes. The same is true for injection migmatites where the country rock is hot (Weinberg and Searle, 1998, Weinberg and Searle, in press). Four different migration mechanisms have been proposed to account for magmatic structures in in situ and injected migmatites: dyking, tectonic pumping, pervasive flow into low-viscosity supersolidus country rocks, and volatile-driven intrusion (Fig. 4). These can be seen as end members of migration processes that more often than not may be active simultaneously in hot crust, enhancing magma migration. Dyking has already been discussed in detail above and elsewhere; in this section, the three other mechanisms will be summarized.

5.1. Tectonic pumping

Field evidence led a few workers to propose an alternative mechanism to dyking and diapirism, where magmas rise by exploring crustal weaknesses (e.g., Hutton et al., 1990; D'Lemos et al., 1992; Brown, 1994, 1995; Grocott et al., 1994; Brown and Rushmer, 1997; Collins and Sawyer, 1996). These authors concluded that magma is driven by buoyancy assisted to different degrees by contemporaneous tectonic deformation (the 'deformation enhanced ascent' of Brown, 1994). Brown (1995) and Collins and Sawyer (1996) described migmatite terrains in which magma intruded pervasively parallel to anisotropies such as foliation planes, fold hinges, layering, mineral lineation and boudin necks. Because magma is distributed widely in small pockets (meter scale), its migration requires hot surroundings (temperatures above magma solidus temperature). Furthermore, because the magma bodies are relatively small, their buoyancy stress are low, resulting in migration rates compatible with the rates of tectonic deformation. The duration of the process is controlled by a combination of thermal and tectonic evolution of terrane and is likely to be of the order of 10^6 a.

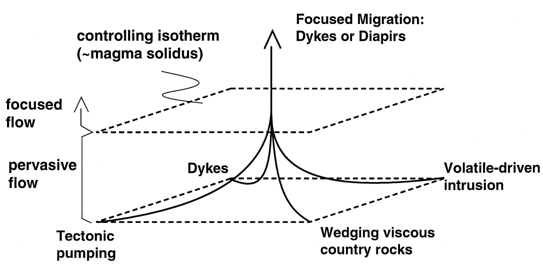

Fig. 4. Four processes control magma migration in in situ and injection migmatites. Except for dyking, the pervasive nature of these processes require country rock temperatures close to magma solidus. Migration through cold crustal rocks requires focused transport mechanism such as dyking or diapirism. Heat advected by pervasive magma migration may effectively push the isotherm limiting pervasive migration to shallower levels.

More recently, two new mechanisms of magma migration have been proposed based on the structures found in injection complexes of leucogranites produced by the continental collision between India and Eurasia. In both cases, the leucogranite intruded into hot country rocks and gave rise to sheets mostly parallel to the country rock foliation: the Pangong Injection Complex in the Karakoram Range of Ladakh in NW India (Weinberg and Searle, 1998), and the Imja Khola granites of the Khumbu area (Nuptse–Lhotse mountains) in Nepal (Weinberg and Searle, in press).

5.2. Magma wedging into low-viscosity country rocks: Pangong Injection Complex (PIC)

In the Karakoram Range of NW India, Weinberg and Searle (1998) described the Pangong Injection Complex where numerous leucogranite sheets intruded a sequence of partially molten amphibolites and gneisses, and minor pelites and calc-silicates (Fig. 5). At the time of leucogranite intrusion, the country rocks were hot, possibly above solidus, as inferred from similar crystallization ages of the intrusive leucogranite and in situ partial melting of the country rock migmatites. The high temperature of the country rocks is reflected in the low-ductility structures developed in response to leucogranite intrusion. The hot surroundings freed the leucogranite melts from freezing and allowed its pervasive flow in mesoscale through the country rock, giving rise to leucogranite sheets. The viscous rocks also inhibited dyke propagation by blunting of the dyke tips. Lo-

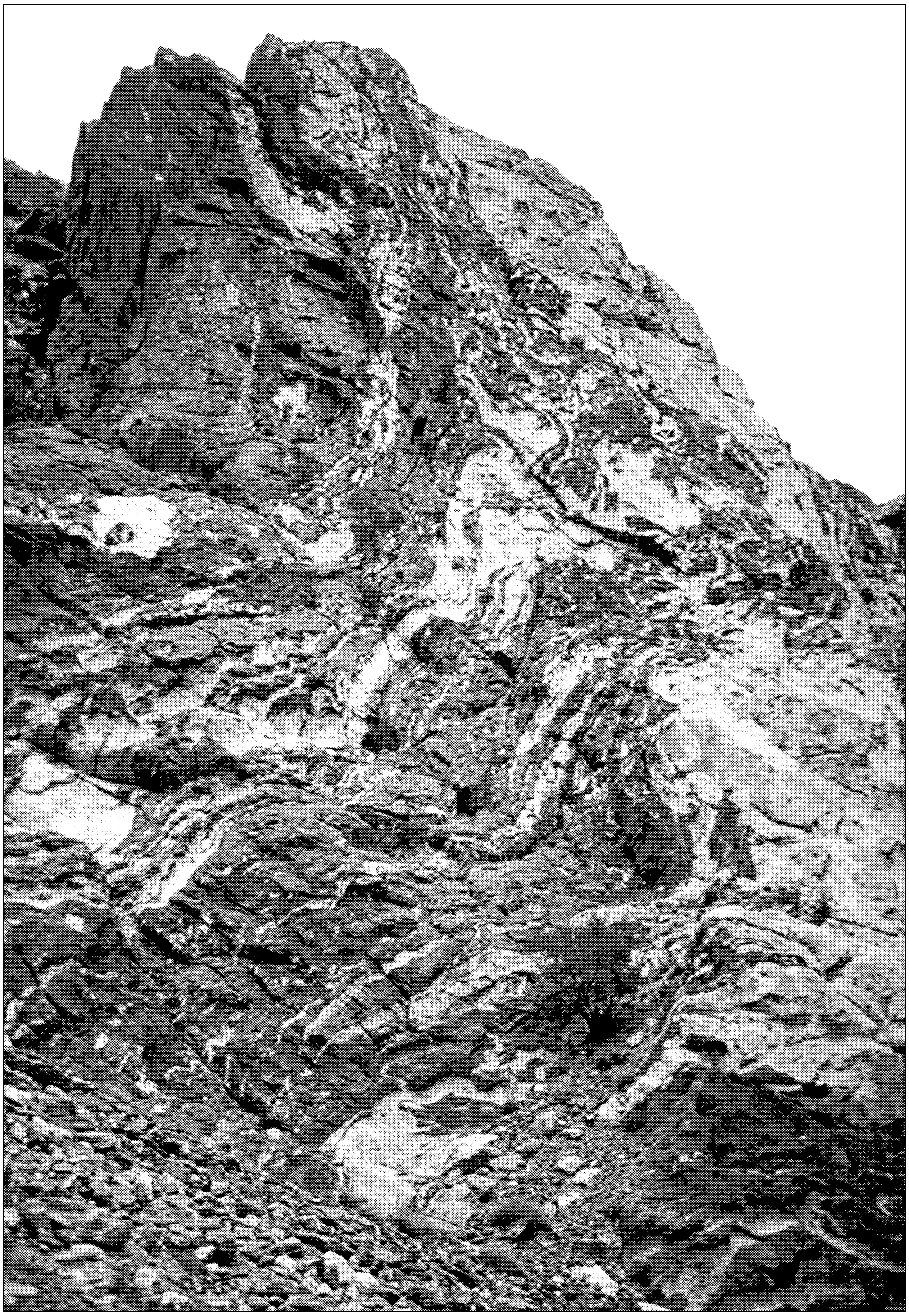

Fig. 5. Pangong Injection Complex. Leucogranite sheets intruding and folding hot country rocks. The low viscosity of the country rock led to its convolute folding and to the development of irregular, nonplanar granite sheets with decimeter to meter-scale blisters (balloons). Height of the rock wall is approximately 100 m (from Weinberg and Searle, 1998).

cally, these sheets coalesced to produce plutons where narrow country rocks screens are preserved. Whereas there is clear evidence for local, planar magma sheets, cross-cutting the foliation, most likely resulting from dyking, most magma sheets resulted from the slow wedging of weakness planes.

The Pangong Injection Complex was interpreted as representing the pathways of leucogranites towards the structurally higher Karakoram batholith emplaced, presumably, into colder country rocks. An important question which remains to be answered is how magma migration evolved from this pervasive style, permitted by the hot country rock, to a more focused migration, required by colder, subsolidus country rocks. Or does the large batholith overlying these pathways represent the accumulation of magmas at the upper end of the hot crustal zone of pervasive magma migration?

5.3. Volatile-driven intrusion

Walker and Mathias (1947) described an injection migmatite zone around the granite in Sea Point, Cape Town (South Africa), in which truly granitic material intruded only along larger ruptures in contact zones. Volatile-rich aplitic phases were capable of intimate penetration of the country rocks by working along schistosity planes. Aplitic phases intruded more easily into micaceous layers than more quartzose beds which are commonly free from veins.

Similar structures were found in the Imja Khola (Khumbu Himal, Nepal, Weinberg and Searle, in press), where large leucogranite sheets intrude the gently dipping regional foliation, to form sills of up to a few hundred meters thick. In this area, there are large volumes of pegmatites and aplites suggesting volatile-rich magmas, and there are relatively few dykes cross cutting foliation. Aplitic–pegmatitic phases pervasively intrude the regional schist as veins down to millimeter-scale, and disrupt and brecciate country rock layering (Fig. 6). In contrast, granites tend to be restricted to large bodies, seldom in direct contact with the country rock.

The regional temperature during intrusion was below anatexis of biotite-bearing rocks, but within the sillimanite–cordierite–muscovite facies, as implied by the mineral assemblage found in the schists and by the contemporaneity of magma intrusion and peak temperatures and regional deformation. The temperatures may therefore be constrained to a broad range between 500–700°C, probably somewhat below magma solidus, but still sufficiently warm to allow mesoscale pervasive flow of the low-viscosity, volatile-rich phases.

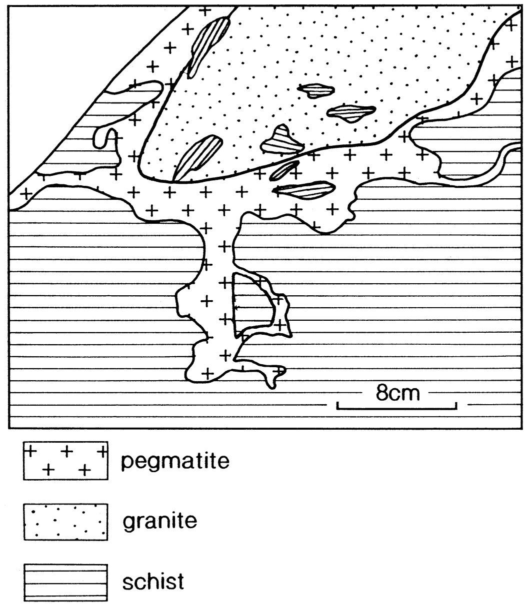

Fig. 6. Typical example of the relation between pegmatites, granites and schists in the Nuptse area of the Khumbu Himalayas. Volatile-rich fluid phases which produced pegmatites (crosses) intruded, disrupted and locally altered the schist (horizontal lines), down to millimeter-scale. Granite (dots), due to its much higher viscosity, was unable to interact with the schist in such an intimate way, and is not in direct contact with the schist (redrawn from Weinberg and Searle, in press). This and other structures in this area suggest that, similar to the processes in dyke tips, but not restricted to these tips, volatile-rich phases were exsolved and intruded into low-pressure, high-permeability zones in the country rocks.

The ease with which the volatile-rich phases intruded the warm schists in Imja Khola, evidenced by swarms of millimeter-sized aplitic veins, in contrast with the more passive role of the magmas, led Weinberg and Searle (in press) to suggest that magma migration was driven by the opening of channelways in the country rocks by the migration of mobile volatile-rich fluids. Exsolution of volatiles would have occurred early in the intrusive process, and would have been caused by a combination of local variations in pressure and magma crystallization. The exsolved volatiles rapidly flowed through the country rock, moving along local pressure gradients and using high-permeability pathways. These fluids heated the country rocks and opened pathways for the stiffer magmas.

6. Discussion

A number of authors have spent considerable energy studying basaltic magma migration in mid-ocean ridges. In this environment, the combined effects of magma buoyancy and material flow lines imposed by plate divergence, focus magma flow towards a narrow zone in the central region of the ridge. This system is controlled by magma porous flow and relatively well-known dynamics. This is in stark contrast with present knowledge of felsic magma migration in continental settings.

The evidence from hot crustal zones is that large magma volumes have been unable to rise to their neutral buoyancy levels. Magma migration in hot crust is freed from the constraints of freezing, and is generally pervasive in mesoscale, showing a range of intrusion mechanisms most of which are limited by the crustal isotherm corresponding to the magma solidus. These different mechanisms may develop in parallel, and may include the intermittent departure of transporting dykes. Whereas pervasive migration may dominate magma transport up to shallow crustal levels in hot crust (high T/low P metamorphism), it is unable to account for plutons emplaced into cold rocks. Migration into cold rocks requires other, more focused mechanisms, such as diapirism or dyking (Fig. 7 and Weinberg and Searle, 1998, for detailed discussion).

However, geotherms may be pushed to shallower levels by heat advection accompanying magma migration. I propose a simple mechanism that would allow pervasive migration to expand its domain by heating the country rocks. If magmas had the same solidus temperature as their source rock, pervasive magma migration would be almost entirely restricted to the melting zone, with little effect on the regional geotherms (Fig. 7a). However, because segregated melt generally has lower solidus temperature than its source, melt may leave the source and intrude pervasively the region above it, free from freezing. The difference between the solidus temperature of the source and melt may be of the order of 100°C, allowing a magma to migrate upwards ca. 3–5 km before encountering rocks at temperatures below its solidus (Fig. 7b). In this way, heat is transferred from the melting zone to shallower levels, pushing the geotherms upwards, allowing further melt migration. The final depth of the controlling geotherm results from the interaction between the speed of magma production, extraction and transport to that

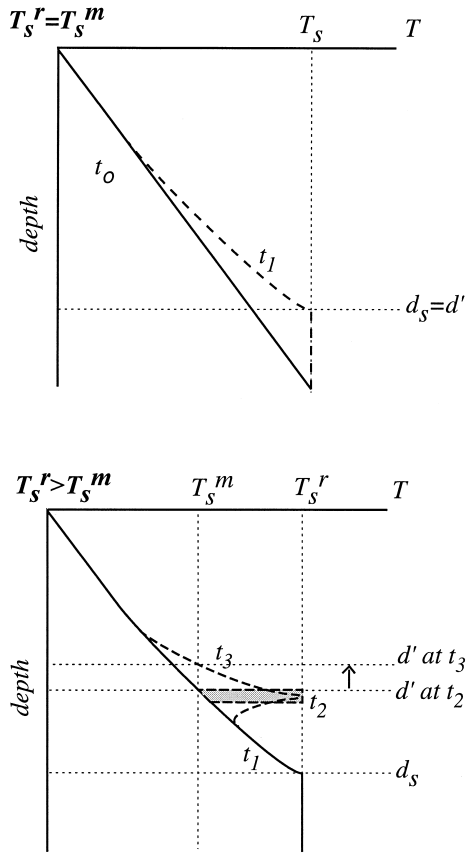

Fig. 7. (a) Schematic temperature–depth plot of a system where the magma solidus temperature (T_s^m) equals that of the source rock (T_s^r). Melting is driven by heating the base of the system, where the initial steady-state temperature at time t_0 is equal to T_s^r. At t_1, the depth controlling the upper limit of pervasive magma migration (d') equals the upper limit of the region undergoing anatexis. d' migrates slowly upwards, driven by the heating of the crust. Magma migration within the source has only a modest effect on crustal geotherms. (b) When $T_s^r > T_s^m$, melt leaves the source and migrates pervasively to the controlling depth d' (the geotherm corresponding to the melt solidus temperature), a few kilometers above. Migration starts sometime after source melting (t_2), depending on the rate of melt segregation from the pores into magma sheets and pools capable of faster migration. Magma ponds at d', and heats the surroundings, driving d' upwards (t_3). This leads to a positive feedback between magma migration and the migration of d'. Melting is assumed for simplicity to be eutectic (constant temperature), and the vertical dotted lines correspond to the particular case where solidi temperatures do not vary with depth.

depth, and the temperature and latent heat of the magma (i.e., the rate of heat advection), the regional steady state geotherm, the difference in temperature between source solidus and magma solidus (a function of rock composition and melt fraction), and the solidus curve of the magma (solidus temperature as a function of pressure). This process may be capable of expanding the domain of pervasive migration by several kilometers given favourable conditions.

A particularly interesting case is that of magmas invading hot shear zones. In these zones, magma would be able to reach shallower levels than in the surroundings enhancing the temperature contrast to the surroundings. This would allow increased magma migration and further weakening and widening of the shear zone, giving rise to a positive feedback mechanism between shearing and magma migration (e.g., Brown and Solar, 1998).

6.1. Field constraints: characterization of channelway geometry

Detailed field studies of the geometry of magma channel networks in migmatite terranes may help us constrain and understand magma extraction from the source. Field observations may help us in finding ways to extrapolate outcrop scale observations to source scale behaviour and to predict the ability of the system to produce transporting dykes. In this section, I outline a few first steps that may be used by field geologists with that aim.

The geometry of magma networks depends on a range of variables such as rock strength and strength anisotropy, permeability anisotropy, melt fraction, volume expansion during melting and rate of melting, melt viscosity, and tectonic deformation (e.g., extensional environments may lead to faster channel widening upstream, larger Δ).

I suggest that a first approach is to directly apply the systematics developed for river drainage systems by Horton (1945) and in more recent work (e.g., Rinaldo et al., 1993 and Karlinger et al., 1994). Can we gainfully order magma channels as river networks? The combined effects of syn-magmatic deformation and late-stage deflation of the channel network, may heavily blur the initial geometry. Thus, as a first approach, this systematics should be tried out in areas little affected by syn-magmatic deformation and ideally in systems where magma has in the main remained within the source.

Once channels are ordered, the average catchment volume (or area) of channels of increasing orders may be determined. Can we predict the catchment volume of high-order channels (major dykes) by the study of low-order catchments? This would enable us to predict how many transporting dykes are necessary to drain the source and how much magma a transporting dyke may have transported. Many migmatite terranes seem to develop cylindrical channelways. If this is so, we need to determine their interspacing, their length and whether several of them drain into larger channelways.

For ordered channel networks, it is possible to determine whether the system is self-similar with constant values of bifurcation and length ratios (r_b and r_l, defined in the caption to Fig. 1). (To test self-similarity, channels of lengths varying by up to two orders of magnitude need to be analyzed.) For self-similar magmatic systems, a new parameter is required to characterize the preserved widening of the channels. This could be either the diameter exponent, Δ, or a width ratio, r_w, analog to r_l.

The properties of self-similar (fractal) networks can be used to determine the system's geometry at scales beyond those measured. For example, by extrapolating the results from mesoscale field observations, down to grain scale, it is possible to more accurately estimate the total surface area of the channel network and the importance of magma flow from pores during dyke propagation.

The main variable controlling the ability of a source to produce transporting dykes is the permeability of the interconnected network. The large range in size, width and spacing of dykes make determining the permeability a difficult task. An interesting step forward would be to determine the permeability of dykes (channels) of a certain order. Do these channels form an infinite interconnected network ($p = 1$)? Does the permeability and p decrease or increase with channel orders? Are there major bottlenecks in the system?

It may well be that magma channelways are not self-similar and that deformation and drainage of the system partly destroy the original geometry of the network. Nevertheless, efforts to characterize source geometry is the most direct means to understand magma extraction and migration from partially molten zones.

7. Conclusions

This paper reviewed the literature on dyking as a possible transport mechanism of felsic magmas, and revealed that whereas the growth of a single dyke in the source, and the requirements for dyke propagation from the source, have both been the subject of several studies, the crucial step linking small individual dykes to major transporting dykes has seldom been studied (see Petford and Koenders, 1998). In order to produce a transporting dyke, the source must evolve to a stage of maturity in which an extensive tributary dyke network is capable of maintaining the high flow rates in the transporting dyke. The structure and size of this network is of fundamental importance, because it controls the entire process of magma extraction by dykes, from pores to pluton.

The main conclusions regarding felsic dyking are summarized in Table 2. The rarity of dyke swarms in regions of felsic magmatism suggests that magma sources may only be able to produce few transporting dykes during its lifetime. The question is whether these are able to drain most of the magma resident in the source or only a small fraction of it. To answer this question, knowledge on the structure of the magma source is required. First and foremost, it is necessary to determine the factors controlling the size of the reservoir drained by a transporting dyke; whether the drainage system is self-similar; and to determine simple geometrical parameters controlling the structure of the network.

Evidence from in situ and injection migmatites suggest that dyking plays a secondary role in magma migration, and that other processes may be more important (Fig. 4). Three alternative processes have been identified: tectonic pumping, pervasive flow into low-viscosity supersolidus country rocks, and volatile-driven intrusion (Table 3). They all lead to pervasive magma flow, but are restricted to hot crustal zones (temperatures close to magma solidus), and therefore are unable to directly account for plutons emplaced in cold crust. However, despite the strong control of temperature on pervasive magma migration, heat advected by the magma may induce

Table 3
Magma migration in hot country rocks

(1)	Tectonic pumping
	–slow process controlled by magma buoyancy, pressure gradients resulting from tectonic forces and rock strength anisotropy
	–duration controlled by thermal and straining evolution of the zone (order of million years; e.g., Collins and Sawyer, 1996; Brown and Solar, 1998)
(2)	Magma wedging of low-viscosity country rocks
	–magma pressure leads magma wedges into low-viscosity country rock where dyking is inhibited
	–structures controlled by country rock weakness zones
	–slow process led by magma buoyancy
	–type locality: Pangong Injection Complex (Weinberg and Searle, 1998)
(3)	Volatile-driven intrusion
	–fast process in local, outcrop scale, but may last millions of years in macroscale, controlled by the thermal and tectonic evolution of the area
	–the low viscosity of volatile-rich fluids allows highly penetrative intrusion, forming small, millimeter-scale sheets (unlike magmas), mostly parallel to high-permeability planes such as schistosity
	–pressurized volatile-rich phases leads to breaking up the country rock and to cross-cutting structures
	–magmas follow the pathways opened by the volatile-rich phases at a later stage
	–type localities: Imja Khola, Nepal (Weinberg and Searle, in press), Cape Town granite (Walker and Mathias, 1947)
(4)	Dyking
	–very rapid process
	–requires high magma pressure and elastic response of the country rock
	–a rare event during the evolution of a felsic magma source region, potentially able to drain large volumes of magma
	–occasionally gives rise to dyke swarms (e.g., Gangotri pluton, Scaillet et al., 1996)

crustal heating expanding the domain where pervasive migration prevails.

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