The mechanism of melt extraction from lower continental crust of orogens

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ABSTRACT: Melt extraction is a process with a length scale that spans many orders of magnitude. Studies of residual migmatites and granulites suggest that melt has migrated from grain boundaries to networks of leucosome-filled structures to steeply inclined cylindrical or tabular granites inferred to have infilled ascent conduits. For example, in anatectic rocks from southern Brittany, France, during decompression-induced biotite-breakdown melting, melt is inferred to have been expressed from foliation-parallel structures analogous to compaction bands to dilation and shear bands, based on location of residual leucosome, and from this network of structures to ascent conduits, preserved as dykes of granite. The leucosome-filled deformation band network is elongated parallel to a sub-horizontal lineation, suggesting that mesoscale melt flow was focused primarily in the plane of the foliation along the lineation to developing dilatant transverse structures. The leucosome network connects with petrographic continuity to granite in dykes; however, the orientation of dykes discordant to fabric anisotropy suggests that their formation was controlled by stress, which indicates that the process is a fracture phenomenon. Blunt fracture tips and zigzag propagation paths indicate that the dykes represent ductile opening-mode fractures; these are postulated to have formed by coalescence of melt pockets. The structures record a transition from accumulation to draining; quantitative volume fluxes are calculated and presented for the generalised extraction process. The anatectic system may have converged to a critical state at some combination of melt fraction and melt distribution that enabled formation of ductile opening-mode fractures, but fractal distribution of inferred mesoscale melt-filled structures has not been demonstrated; this may reflect the inherent anisotropy and/or residual nature of the drained source. Melt extraction has been modelled as a self-organised critical phenomenon, but the mechanism of extraction is not described and the relationship between these models and the spatial and temporal granularity of lower continental crust is not addressed. Self-organised critical phenomena are driven systems involving 'avalanches' with a fractal frequency-size distribution; thus, the distribution of melt batch sizes might be expected to be fractal, but this has not yet been demonstrated in nature.

KEY WORDS: crustal melting, deformation bands, ductile opening-mode fractures, magma ascent, melt flow networks, melt fluxes, spatial patterning.

Melt extraction, magma ascent and pluton emplacement represent a significant and irreversible heat and mass transfer from lower to upper continental crust (e.g. Brown *et al.* 1995a). In many cases, the size, shape and spatial distribution of upper crustal plutons probably reflect aspects of melting in the lower crust, and extraction and emplacement are presumed complementary processes between which there must be feedback modulated by the ascent mechanism (e.g. Brown & Solar 1999). Superimposed on this intrinsic control are conduit-forming and space-making processes induced by regional deformation (Brown 2001a, b). However, in the extreme case where magma ascent is strongly controlled by a major steeply inclined crustal-scale shear zone, such as the Peri-Adriatic Fault System, a linear alignment of plutons in the upper crust may not necessarily reflect a linear source region in the partially molten lower crust (e.g. Rosenberg 2004).

Melting occurs at distributed multiphase grain boundaries in a source (mm^3), whereas a typical pluton represents a large volume of melt accumulated at a sink (10^3–10^4 km^3); plutons commonly crystallise within 10 s to 100 s of ky, and batholiths within several my (cf. Glazner *et al.* 2004). Assuming 10–20 vol.% melt extraction, source volumes are 5×10^3–10^5 km^3. Thus, melt extraction, from segregation to emplacement, is a process with a length scale that spans at least seven orders of magnitude, occurring on a time scale of $<10^5$–$>10^6$ y. In the exposed lower continental crust of deeply eroded orogens, where evidence of melting is preserved in residual migmatites and granulites, magma chambers of any kind related to the inferred melting event appear to be absent. Therefore, either melt must accumulate in a network that feeds drainage channels (e.g. Brown & Solar 1999), or there must be some other mechanism by which melt batches aggregate to increase in size and ascend through the crust (e.g. Bons *et al.* 2001, 2004), or the source must ascend *en masse* (Marsh 1982; Olsen *et al.* 2004).

To a first approximation, assuming no mass addition from the mantle, the volume of granite emplaced in the upper crust will be equal to the volume of melt extracted from the lower crust, and rates of melt production, extraction, ascent and emplacement are likely to have been balanced at the crustal scale. Volume fluxes [Fig. 1; Q_{Pr}, melt production (implicit in source); Q_{Ex}, melt extraction; Q_{As}, melt ascent; and Q_{Em}, melt emplacement] over distinct time intervals have been considered qualitatively by, for example, Thompson (1999) and Brown (2001a, b), although the role of deformation in enhancing rates was not considered explicitly by either author. These authors argued that relationships between pairs of these quantities can be used to deduce the relative efficiencies and magnitudes of

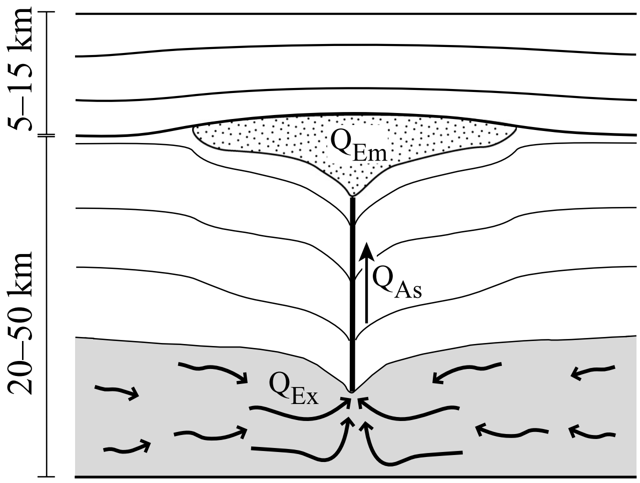

Figure 1 A simple model of melt extraction, ascent and emplacement in which the rates of melt production, extraction, ascent and emplacement are likely to have been balanced at the crustal scale. The rate of melt production (Q_{Pr}) is controlled by heat flow and is not shown explicitly in the figure. Volume fluxes for melt extraction (Q_{Ex}), melt ascent (Q_{As}) and melt emplacement (Q_{Em}) are shown schematically. The depths shown on the left-hand side of the figure are intended as a general indication of the likely range in orogenic environments. From Cruden 1998, modified by permission of the Geological Society.

melt-flow mechanisms, and although this analysis is simplistic in relation to orogens where feedback occurs between thermally induced weakening, melt-induced weakening and deformation, a simple qualitative approach nonetheless provides a starting point.

If melt production, extraction, ascent and emplacement are equal (i.e. $Q_{Pr}=Q_{Ex}$, $Q_{Ex}=Q_{As}$ and $Q_{As}=Q_{Em}$, then $Q_{Pr}=Q_{Em}$), the system may be described by percolative flow (cf. Olsen *et al.* 2004). These rates vary with time, but if the rate of melt production is faster than the rate of extraction at any time, the source will stagnate, melt fraction will increase and the volume change related to the melting reaction will increase the fluid pressure in the source. Thus, when $Q_{Pr}>Q_{Ex}$, melt-induced dilatancy potentially occurs in the source, and Q_{Ex} is increased by channelled flow. On the other hand, if melt emplacement is faster than melt extraction ($Q_{Em}>Q_{Ex}$), episodic extraction is more likely (cf. Handy *et al.* 2001) or the melt flux is rapidly exhausted. Finally, if the rate of melt production is less than the rate of melt extraction ($Q_{Pr}<Q_{Ex}$), the source will be drained slowly. Ultimately, since Q_{Pr} is controlled by the heat flow, it must exert control over the other fluxes.

To improve our knowledge of melt extraction from the continental crust, we need to: develop realistic models of how and at what rate melt segregates from residual material; establish the spatial (pervasive or channelled) and temporal (continuous or transient) characteristics of melt extraction; and pursue realistic models of magma ascent. To this end, the present author briefly summarises postulated mechanisms for magma ascent, and briefly reviews the microstructures and the mesoscale spatial patterning of former melt-bearing rocks, such as residual migmatites and granulites. Then, he uses one particular example from southern Brittany, France, related to extensional exhumation of lower crust to form a core complex, to illustrate in more detail one mechanism of melt extraction. The author presents quantitative volume fluxes related to the extraction process calculated for the general case, and in the final section, develops a general model for melt extraction from the lower continental crust of orogens.

1. Mechanisms proposed for magma ascent in orogenic belts

Dyking by brittle fracture is not a viable mechanism for extracting melt from partially molten crustal sources (e.g. Rubin 1998), although brittle fracture is a viable mechanism for second-stage ascent of melt from an upper crustal magma chamber into sub-solidus crust (e.g. Weinberg 1996). Alternatives for initial melt ascent through suprasolidus crust include mobile melt-filled hydrofractures (Dahm 2000; Bons *et al.* 2001, 2004) and various mechanisms to bridge grain-boundary flow to channelled flow in ductile crust (e.g. Brown 1994; Collins & Sawyer 1996; Brown & Solar 1998a; Sawyer *et al.* 1999; Weinberg 1999; Leitch & Weinberg 2002; Olsen *et al.* 2004).

The slow rates of migration of viscous silicic melt through pores and shear-enhanced compaction of residue probably require that melt drained in a dyke propagation event must have been previously extracted from the pores and reside an accumulation network; additionally, there must be sufficient hydraulic head to drive ascent (Brown & Solar 1998b). In general, tabular intrusions are favoured in nature (Emerman & Marrett 1990; Rubin 1995; Weinberg 1999), and initiating dykes by ductile fracture avoids the problems and assumptions inherent in brittle fracture. By ductile, the present author means fracturing associated with distributed inelastic deformation at the fracture tip (cf. Rutter 1986), but no particular stress-strain relationship or deformation mechanism is implied (cf. Eichhubl 2004).

Suprasolidus crust is an over-pressured system with supra-magmastatic fluid pressure gradients and spatial and temporal variations in permeability. Permeability has a power law dependence on melt fraction (Wark & Watson 1998), which may lead to compaction-generated flow instabilities, manifested as self-propagating high-porosity structures (Connolly & Podladchikov 1998; Rabinowicz & Vigneresse 2004). These high-porosity structures may grow by influx of melt, which may lead to fully-segregated melt flow (Richardson *et al.* 1996; Podladchikov & Connolly 2001). Thus, melt flow in high porosity structures may represent an intermediate mechanism bridging porous flow to channelled flow.

Diapirism involves ascent of melt plus residue; it is not generally considered to be a viable process for emplacement of upper crustal plutons (Emerman & Marrett 1990; Clemens & Mawer 1992; Rubin 1993), although subsequent modelling contradicts this view (Weinberg & Podladchikov 1994; Burov *et al.* 2003). Diapirism may occur in the lower crust (Barbey *et al.* 1999; Teyssier & Whitney 2002), and it is a mechanism by which (some) plutons are inferred to have been emplaced in the deep crust in continental arcs (e.g. Miller & Patterson 1999).

2. Petrological and field-based arguments and observations

2.1. Microstructural scale

Under equilibrium conditions in isotropic crust, melting begins at multiphase grain junctions which include quartz and feldspar, and a hydrate phase. However, Earth's crust is anisotropic and in a state of stress, and variations in bulk composition and grain size influence the sites at which melting begins, and orientation of hydrate phases in a metamorphic fabric may influence the spatial development of melt pockets (Sawyer 2001). Differential stress varies within any given volume of crust, and these variations give rise to pressure gradients throughout the volume of crust. Melting may begin at sites of lower or higher P, once the initial thermal overstep is close to that required to overcome the activation energy for the melting reaction. For high $a(H_2O)$ melting, where the dP/dT of reaction is negative, small increases in P may promote the start of reaction, whereas, for moderate-to-low $a(H_2O)$ hydrate-breakdown melting, where the dP/dT of reaction is positive, small decreases in P may promote the start of reaction. Furthermore, for moderate-to-low $a(H_2O)$ melting, the solid products of the reaction may have difficulty nucleating (Powell & Downes 1990). Thus, once melting is established at a particular site, it is energetically favourable for melting to continue at that site and for the solid products to grow at that site (Fig. 2) by a process of diffusional mass transfer (Powell & Downes 1990; Hand & Dirks 1992).

The grain-scale distribution and connectivity of silicic melt is of primary importance in determining the physical properties [e.g. permeability, rheology (which is sensitive to the presence and distribution of melt) and effective diffusivity (which may be increased by orders of magnitude by the presence of interconnected melt)] of melt-bearing lower continental crust. Permeability is especially important because it, melt viscosity and matrix viscosity are the main parameters which control the dynamics of melt segregation. Under conditions of mechanical equilibrium, silicic melt wets matrix minerals in crustal rocks (Laporte & Watson 1995; Laporte & Provost 2000); thus, in melt-bearing crust, melt is expected to be interconnected at very low melt fraction (e.g. Laporte *et al.* 1997; Lupulescu & Watson 1999; Wark *et al.* 2003). Also, deformation experiments on analogue materials show evidence of dynamic wetting of grain boundaries, which enhances early interconnection of melt, and inter-granular fracture and grain boundary sliding (Rosenberg & Handy 2000, 2001). Thus, melt is likely to be connected in all common crustal protoliths (Laporte *et al.* 1997).

This conclusion is supported by observations of microstructures in natural anatectic systems, which provide evidence of the former presence of melt in grain-scale pores (e.g. Brown *et al.* 1999; Brown 2001a, b). Melting and crystallisation from melt may be inferred using the following criteria: the igneous microstructure of leucosomes in migmatites and granulites (e.g. Cuney & Barbey 1982; Vernon & Collins 1988); mineral pseudomorphs after grain-boundary melt films and pockets [in contact migmatites (e.g. Holness & Clemens 1999; Marchildon & Brown 2001, 2002), in regional migmatites (e.g. Marchildon & Brown 2003) and in granulites (e.g. Sawyer 1999, 2001; Guernina & Sawyer 2003)]; magmatic rims on subsolidus cores of grains [e.g. rational faces, overgrowths of different compositions (e.g. Marchildon & Brown 2001, 2002; Sawyer 2001)]; spatial distribution of like and unlike phases [i.e. mineral phase distribution (Ashworth & McLellan 1985)]; and, annealed microfractures (Watt *et al.* 2000; Marchildon & Brown 2001, 2002).

2.2. Mesoscale structures

First, the present author evaluates whether features preserved in lower crust represent what intuition leads us to infer – *viz.* melt accumulation networks and magma ascent conduits (Brown 1994). For flow, melt must form an interconnected network and the host must be permeable and for extraction, there must be a driving force and a sink. It is important to ascertain whether the conduit topology infilled by remnant leucosome and the structures infilled by granite represent evidence of active crustal-scale melt transfer from lower to upper crust, or the preserved record of a non-productive system that failed to support crustal-scale melt flow. Active crustal-scale melt flow is inferred if the bulk rock chemical composition is residual for anatectic rocks (migmatites and granulites) in comparison with protolith compositions, and the difference is consistent with net melt loss (e.g. Sawyer *et al.* 1999; Solar & Brown 2001; Guernina & Sawyer 2003); the net difference represents a minimum estimate of melt loss attributable to melt flux through any level of observation and partial crystallisation of migrating melt. Also, melt loss is implied if peritectic phases survive without significant retrogression (Powell & Downes 1990; Brown 2002; White & Powell 2002). For the purpose of this paper, the present author assumes melt loss to be the common situation based on published examples (Sawyer *et al.* 1999; Solar & Brown 2001; Guernina & Sawyer 2003), but emphasises that net melt loss must be evaluated as part of any specific case study.

In the case of isotropic protoliths with randomly oriented pores of melt, analytical (Sleep 1988) and numerical (Simakin & Talbot 2001a) solutions, and finite element modelling (Simakin & Talbot 2001b), indicate that melt-filled veins will form parallel to σ_1 and perpendicular to σ_3 by closure of pores perpendicular to σ_1 and expansion of pores perpendicular to σ_3. Melt segregation into vein networks may lead to tabular bodies of melt of sufficient size that buoyant ascent becomes possible. Alternatively, in appropriate circumstances, bottom-to-top directionality (vertical fracture alignment) may emerge by non-linear evolution of an isotropic fracture network formed by thermal expansion (Petford & Koenders 1998). Such models may apply where lower crustal protoliths are isotropic and heating is rapid, which might be the case in the lower parts of arcs during the subduction period of the orogenic cycle, although Rubin (1993, 1995) has argued that melt-filled fractures are unlikely to propagate out of a melt-bearing source, unless the source is already pre-warmed by dyking from deeper levels.

In anisotropic protoliths, the scale of compositional layering and the strain field generally control sites of early melt concentration; foliation-parallel and transverse stromata form by mass transfer down gradients in pressure (van der Molen 1985a, b; Maaløe 1992; Brown *et al.* 1995b). With increasing melt fraction (MF), melt becomes progressively concentrated into a developing network of leucosome-filled structures (cf. Brown & Rushmer 1997), including interboudin partitions, shear surfaces, and axial surfaces and hinges of folds; the exact form of the network being controlled by the style of deformation (Brown *et al.* 1995b; Brown *et al.* 1999; Sawyer *et al.* 1999; Guernina & Sawyer 2003; Marchildon & Brown 2003). The role of anisotropy in controlling crustal-scale melt transport has been demonstrated by several studies (e.g. D'Lemos *et al.* 1992; Collins & Sawyer 1996; Brown & Solar 1998a).

Incongruent melting that involves biotite or hornblende breakdown commonly results in melt concentration around the peritectic product minerals (e.g. Powell & Downes 1990), so that nucleation and growth of peritectic phases also controls sites of early melt concentration (Hand & Dirks 1992; Johnson & Brown 2004; Fig. 2a–d). Increasing MF leads to weakening and formation of a network of deformation bands, most probably nucleated on the local concentrations of melt (e.g. as modelled by Grujic & Mancktelow 1998). The preferential occurrence of peritectic minerals in the deformation bands, in spite of a homogeneous distribution of reactant phases in the matrix, supports the inference that, in general, deformation was localised by the concentration of melt (higher porosity) at these sites. This contrasts with the study by Hand & Dirks (1992), where strain associated with pre-existing crenulation structures apparently controlled the sites of initial incongruent melting; in general, this seems to be less common.

2.2.1. Specific examples. At Mt Hay, central Australia, based on sites of leucosome entrapment, Collins & Sawyer (1996) inferred that melt migrated parallel to the steep regional

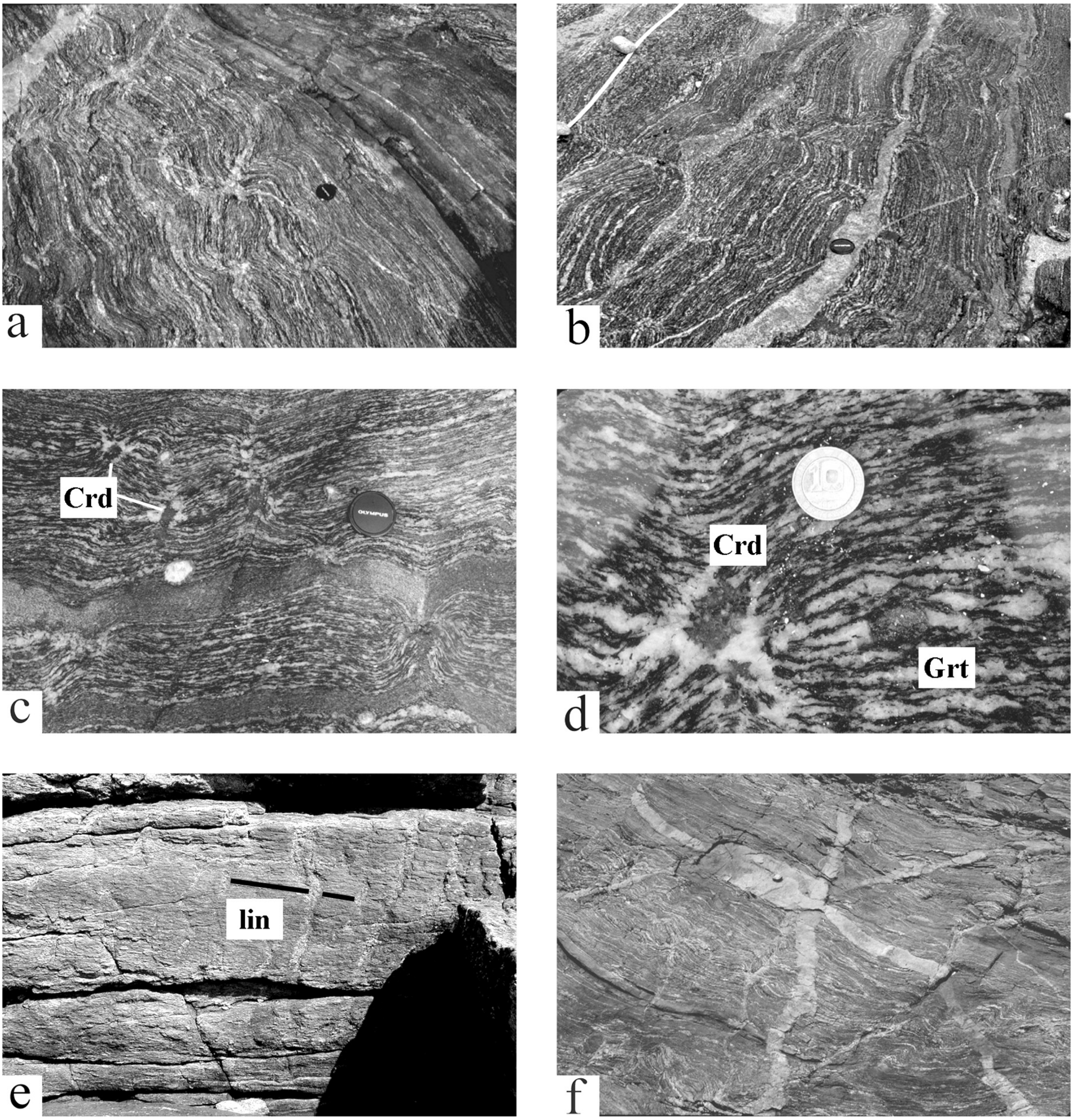

Figure 2 Features associated with melting and inferred melt-bearing structures in granulite facies anatectic migmatites from southern Brittany, France: (a) Stromatic migmatite with layer-parallel leucosome stromata and leucosome in shear bands (Port Navalo Plage, Morbihan); (b) Stromatic migmatite with layer-parallel leucosome stromata showing periodic foliation boudinage (Port Navalo Plage, Morbihan); (c) Stromatic migmatite with layer-parallel leucosome stromata showing periodic foliation boudinage; note cordierite (Crd) in interboudin partitions (Port Navalo Plage, Morbihan); (d) Close-up of the petrographic details in (c); note partially reacted garnet (Grt, in foliation below and right from coin) and cordierite in interboudin partition (below and left of coin); (e) Steeply oriented surface parallel to foliation (note subhorizontal elongation lineation) to show vertical extent of leucosome in transverse structures (Petit Mont, Arzon, Morbihan); (f) Sub-horizontal surface through stromatic migmatite to show granite dykes which intersect in a common steeply inclined cylindrical channel structure; these dykes do not clearly cross-cut each other, but rather merge where they meet. Although the dykes are discordant at outcrop scale, there is only limited evidence for cross-cutting relationships at the vein-to-grain scale, suggesting that material now in the dykes was molten at the same time as the material in the host rock included a melt phase (Petit Mont, Arzon, Morbihan).

elongation direction defined by a strong mineral lineation and coaxial folds via a network of diverse structurally controlled conduits (fold hinges and axial surfaces, interboudin partitions and shear zones) or created new channels via magma fracture, commonly parallel to foliation. Furthermore, these authors argued that porous flow of melt through the interconnected network of mesoscale structures fed larger conduits which transported melt to the upper crust. Similarly, at Wuluma Hills, central Australia, in residual granulites, Sawyer *et al.* (1999) concluded that syn-anatectic deformation enabled melt migration from grain boundaries through a network of veins to hinge zones of moderately plunging folds, and subsequently, into mostly foliation-concordant veins in shear zones which parallel the fold axial surfaces.

Based on detailed studies of granulites in the Ashuanipi subprovince, Quebec, Guernina & Sawyer (2003) suggested two alternative short-range strategies for melt migration. First, in melt-depleted layers without leucosome stromata within layers, but where layer-parallel leucosomes occur at the interfaces between layers, melt movement is inferred to have been primarily orthogonal to the layering, and melt extraction is inferred to have taken place by flow along the layer interfaces. Secondly, in melt-depleted layers with leucosome stromata within the layering as well as in the interfaces between layers, the leucosome forms networks composed of wide transverse leucosome-filled structures which are inferred to have drained individual layers and linking layer-parallel leucosomes between the layers which apparently record pathways of melt extraction out of the source. In this second case, melt movement is inferred to have been primarily within the plane of the layer towards the developing dilatant transverse structures which are now infilled by leucosome.

In the Acadian belt of western Maine, northern Appalachians, USA, in residual migmatites, the form of melt ascent conduits mimics the apparent strain ellipsoid recorded by steeply inclined host rock fabrics. Thus, steeply inclined tabular granites occur in zones of apparent flattening strain, whereas moderate-to-steeply plunging cylindrical granites occur in zones of apparent constrictional strain (Brown & Solar 1999; Solar & Brown 2001). Other studies have provided general descriptions of the relationship between mesoscale structures and ascent conduits, which vary from shear zones (e.g. D'Lemos *et al.* 1992; Brown 1994; Rosenberg 2004) to dykes (Marchildon & Brown 2003). The study by Marchildon & Brown (2003) is considered further below.

One implication derived from all of these examples is that large magma chambers do not form in lower continental crust during syn-anatectic deformation. Ultimately, melt drains from the source via discrete tabular or cylindrical conduits to form plutons. How such conduits form is controversial, but deformation-enhanced variations in pressure gradients and variations in porosity are thought to be the key.

2.3. Quantitative data about melt flow networks from southern Brittany

There have been several attempts recently to obtain quantitative data from anatectic rocks. Tanner (1999) and Bons *et al.* (2004) argued that leucosome distributions are scale invariant, but Marchildon & Brown (2003) cautioned that such a conclusion may not be supported by careful analysis of the data.

Marchildon & Brown (2003) utilised well-exposed coastal outcrops of granulite facies migmatites from southern Brittany, in the Variscides of France, to obtain descriptive and quantitative data on the one- and two-dimensional distribution of inferred melt-bearing structures, and the transitions from foliation-parallel to network to conduit flow in an attempt to understand the evolution of the architecture (connectivity and permeability) of melt-bearing systems. The structural relationships between leucosome network morphology and granite in dykes investigated by Marchildon & Brown (2003) are a late-stage feature of the evolution of the Variscan belt in southern Brittany which relate to decompression and biotite-breakdown melting producing peritectic cordierite; the earlier history is not discussed in detail in the present paper. Fugitive melt from the anatectic rocks is inferred to have accumulated as leucogranite plutons at shallower levels of the crust (Gapais *et al.* 1993; Brown & Dallmeyer 1996; Marchildon & Brown 2003).

In this example, leucosome and granite-filled structures include (Fig. 2) millimetre to centimetre scale foliation-parallel and transverse leucosome stromata (reflecting compositional layering, interboudin partitions and extensional shear surfaces) and centimetre- to metre-scale foliation-discordant steeply inclined tabular granites (representing inferred ascent conduits). Petrographic continuity (mineralogy, mode and microstructure) from structure to structure implies that the infilling material crystallised from a continuous melt-bearing network (Fig. 3). Peritectic phases are commonly located in deformation bands, particularly interboudin partitions (Fig. 2), which connotes an important contribution by diffusive mass transfer in the local concentration of melt, particularly at low melt fractions early during biotite-breakdown melting. Using the equation for dimensionless melt transport (MT) number of Olsen *et al.* (2004) to evaluate porous flow versus bulk flow for melt fractions appropriate to biotite-breakdown melting producing peritectic cordierite yields values for MT <1 which are consistent with macroscopic porous flow.

Marchildon & Brown (2003) obtained information about thickness and spacing distributions of foliation-parallel leucosome stromata from one-dimensional traverses (Figs 4 & 5). Cumulative leucosome thickness versus distance along the traverse, and log-log plots of leucosome cumulative frequency versus thickness or spacing are useful in discriminating between different possible types of thickness or spacing distributions (e.g. Gillespie *et al.* 1993, 1999; Loriga 1999). For instance, based on an analysis of synthetic datasets for different vein distribution types, Gillespie *et al.* (1999) argued that fractal distributions will yield a stepwise plot of cumulative thickness versus distance because of inherent clustering. Fractal distributions also define linear trends (with slope related to the fractal dimension, D) in log cumulative frequency versus log thickness or spacing plots. On the other hand, random vein distributions define smooth and regular cumulative thickness versus distance plots, and concave downward patterns of log cumulative frequency versus log thickness and spacing (Gillespie *et al.* 1999).

In Figure 4 the plots of cumulative leucosome thickness versus distance along the traverse are essentially linear, whereas, in Figure 5, the downward concavity of arrays on the plots of log cumulative frequency versus log thickness or spacing indicates a limited range of leucosome thicknesses [generally between 1 mm (the limit of resolution) and 10 mm (upper limit 20–30 mm)], and shows that frequency decreases rapidly and non-linearly with increasing thickness. These features are inconsistent with scale invariance. In these migmatites, the volume of leucosome (~ 50 vol.%) far exceeds that expected at the calculated *P-T* conditions (~ 25 vol.%; Johnson & Brown 2004). In general, leucosomes do not have melt compositions, but represent some combination of residual reactants and products in the deformation bands, together with any early crystallised products trapped during syn-transport crystallisation of the migrating melt as the exhumed lower crust initially cooled under suprasolidus conditions. The amount of melt present at final solidification of the leucosome is unknown, but if the minimum is that required for the leucosome to be grain-supported, the proportion of melt remaining is likely to have been small (cf. Powell & Downes 1990).

Observation of structures in mutually perpendicular two-dimensional surfaces suggests that the leucosome network is elongated in relation to a sub-horizontal lineation. Analysis of leucosome distributions using the box-counting method

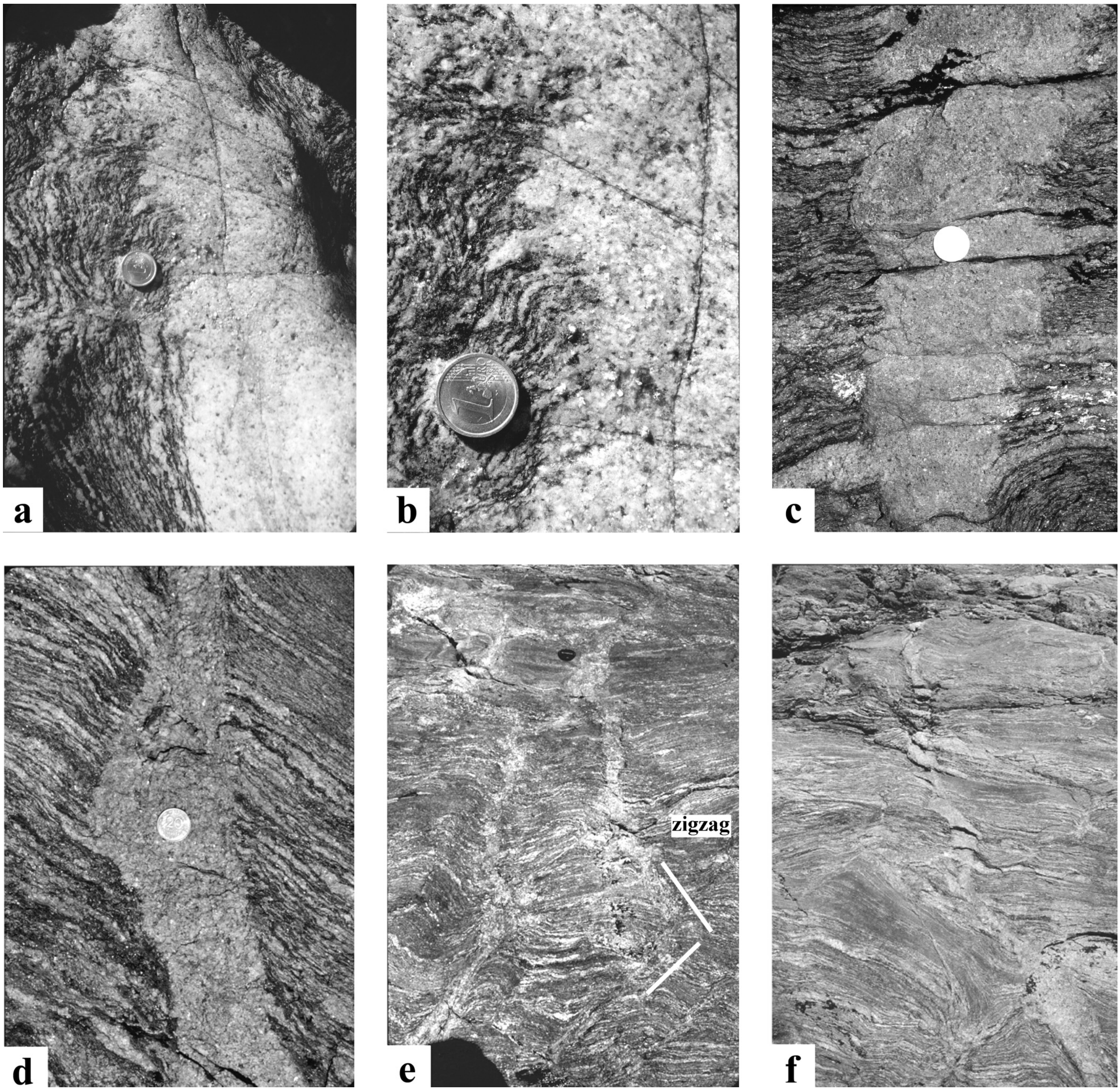

Figure 3 Leucosome in petrographic continuity (microstructure, mineralogy and mode) with granite in dykes in migmatite from southern Brittany, France. (a) Irregular interface between migmatite and granite in ductile opening-mode fracture (Port Navalo Plage, Morbihan); (b) Close-up of (a) to show petrographic continuity between leucosome and granite; (c) Highly discordant granite dyke showing petrographic continuity with the layer-parallel leucosomes it cross-cuts (Le Petit Mont, Arzon, Morbihan); (d) Highly discordant granite dyke showing petrographic continuity with the layer-parallel leucosomes it cross-cuts (Le Petit Mont, Arzon, Morbihan); (e) Thin, discordant sheet-like granite with zigzag form at tip (bottom of image; Le Petit Mont, Arzon, Morbihan); (f) Thin, discordant sheet-like granite with zigzag form (Le Petit Mont, Arzon, Morbihan).

corroborates the suggested anisotropy, but indicates that leucosome morphology is not scale invariant. Melt flow is inferred to have been primarily in the plane of the foliation and along the lineation to developing transverse structures. This network fed steeply dipping discordant planar conduits now filled by dykes of granite.

Draining of melt from deformation bands and capture of small conduits by large conduits will be favoured by the lower melt pressure in the larger conduits (cf. Sleep 1988). In southern Brittany, the formation and orientation of dykes was apparently controlled by far-field stresses associated with dextral transtension (Marchildon & Brown 2003; Cagnard *et al.* 2004). However, spacing of conduits is inferred to have been controlled by the strong intrinsic permeability anisotropy in layers with well-developed stromatic layering and the weak implied extrinsic permeability anisotropy between layers. This leads to variable focusing of melt that requires a larger collection zone for each conduit than would be the case for an isotropic protolith (cf. Eichhubl & Boles 2000).

3. Fluxes

In this section, the present author considers quantitatively fluxes associated with the processes of melt extraction from lower continental crust. Given a pluton of volume of 10^3–10^4 km^3 crystallised within a period of 10^4–10^6 y, and assuming that crystallisation times correspond to transport times,

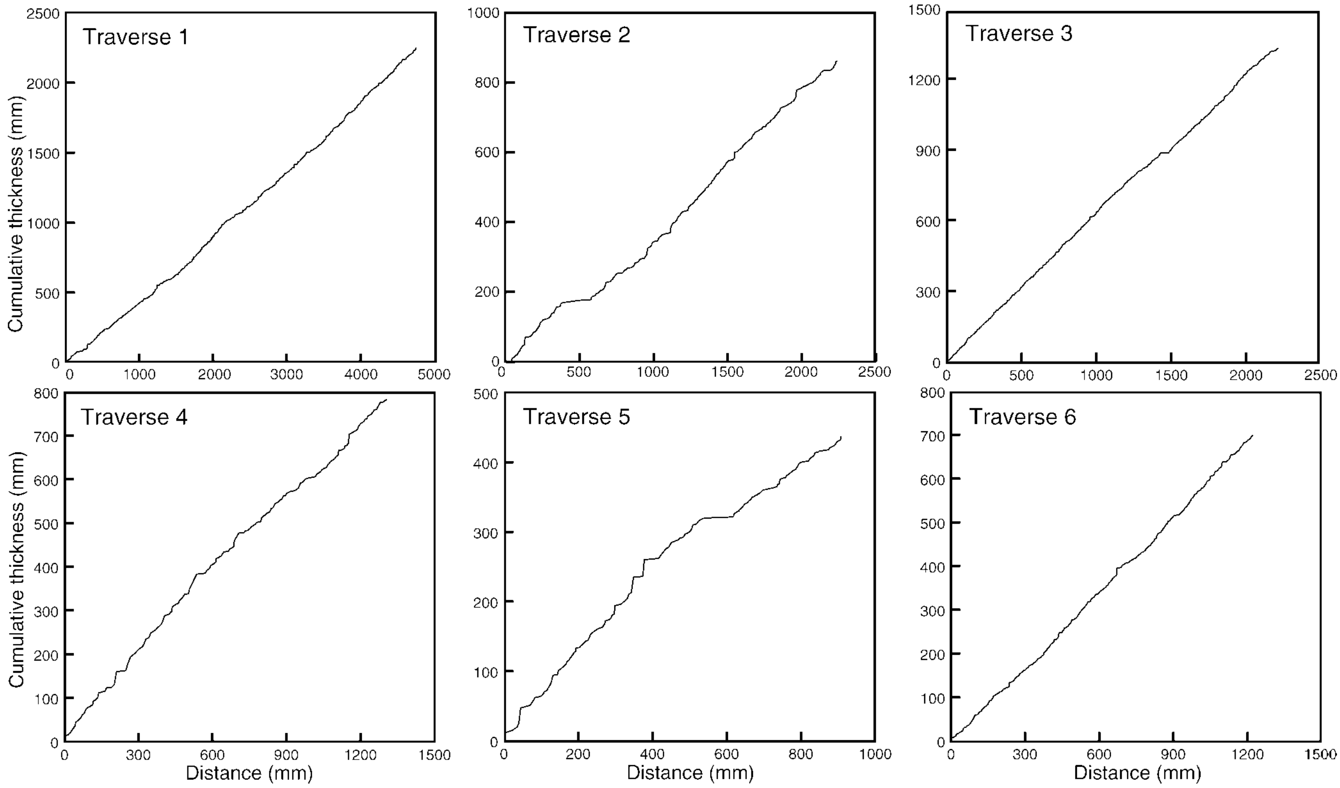

Figure 4 Information about thickness of foliation-parallel leucosome stromata from one-dimensional traverses (after Marchildon & Brown 2003) shown as plots of cumulative leucosome thickness versus distance along the traverse.

then, if a pluton is formed in a single filling event, the crystallisation time gives a maximum time for transport, and all transport estimates will be lower as a consequence. The lower limit for the crystallisation period is consistent with the minimum time of *ca.* 50 ky to extract, via a vein network, 10 vol.% of melt to the top of 10 km thickness of protolith with 0·5–1 wt.% H_2O (assuming approximately isothermal conditions of $\sim$850°C and applied differential stress of <10 but >1 MPa, calculated based on data in Rutter & Neumann 1995). The upper limit for the crystallisation period is consistent with melt extraction from a protolith with similar H_2O content and under similar applied differential stress, but at a lower temperature of $\sim$750°C.

Consider a source of area of 10^3–10^4 km^2 beneath which an average of 1 km^3 per km^2 of melt is extracted from 10 km depth of suprasolidus continental crust (e.g. Zen 1992). For a pluton that is 1 km thick, the melting footprint, the region of melt throughput and the pluton area are identical. This geometry allows calculation of reference melt extraction rates.

To form a pluton of 1 km thickness with a region of melt throughput to melting footprint ratio of 1 requires an integrated melt extraction flux of 1 km^3 melt per km^2 crust, which yields an average melt extraction flux of 10^{-6} to 10^{-4} km^3 melt per km^2 crust y^{-1}. This is equivalent to the situation where $Q_{Pr}=Q_{Em}$, as discussed in the introduction. These fluxes should be contrasted with the reference melt extraction velocity, which is set by the depth of melting. In order to get the last drop of melt through the source area requires transport over at least 10 km, which yields an average melt velocity of 10^{-5}–10^{-3} km melt y^{-1}. The melt velocity is faster than the melt flux because most of the crust is solid; thus, we may think of the melt velocity in units of volume of melt per area of melt per unit time and the melt flux in units of volume of melt per area of crust per unit time.

Given the distinction between melt flux and velocity, the ratio of melt footprint to the region of melt throughput is useful. Consider the extreme case of a melt-filled conduit. With reference to the average pluton volumes (above), a melt-filled conduit of area 10^{-2}–10^{-1} km^2 requires an integrated melt velocity of 10^4 to 10^6 km^3 melt km^{-2} melt, which yields an average melt velocity in the conduit of 10^{-2}–100 km melt y^{-1}. This is equivalent to a single dyke ascent model (Clemens 1998). In this limiting case, the 'crust' through which the melt flows (i.e. the conduit) is all melt, so the melt flux in the region of melt throughput will be the same. However, the melt flux through the melting footprint is the same as estimated above because the pluton is being filled from the same volume of melted crust. These estimates bracket the actual melt extraction rate.

A related issue is the transport of melt within the melting volume to the melt-filled conduit. Whether the melt-filled conduit extends to the base or to the top of the melting volume [10^3 to 10^4 km^2 by 10 km], the distance over which melt travels to the conduit is 10^1–10^2 km. This estimate yields an average melt velocity of 10^{-5}–10^{-2} km melt y^{-1}. A rough estimate of the melt extraction flux to the conduit can be made by assuming that the melt to crust volume fraction is 0·1 [1 km^3 of melt per 1 km^2 by 10 km of crust]. The average melt extraction flux to the conduit is given by the product of the melt fraction with the melt velocity to yield 10^{-6}–10^{-3} km^3 melt per km^2 crust y^{-1}.

Convergent fluxes increase approaching a sink. The melt flux just before it reaches the conduit will be greater than the average melt flux through the melting volume. If the conduit reaches to the bottom of the melting volume, all the melt generated within the melting volume will be converging on a conduit with a surface area of 1–10^{-1} km^2. The integrated melt flux converging on the conduit will be 10^3–10^5 km^3 melt per km^2 crust, which yields melt fluxes across the boundary of

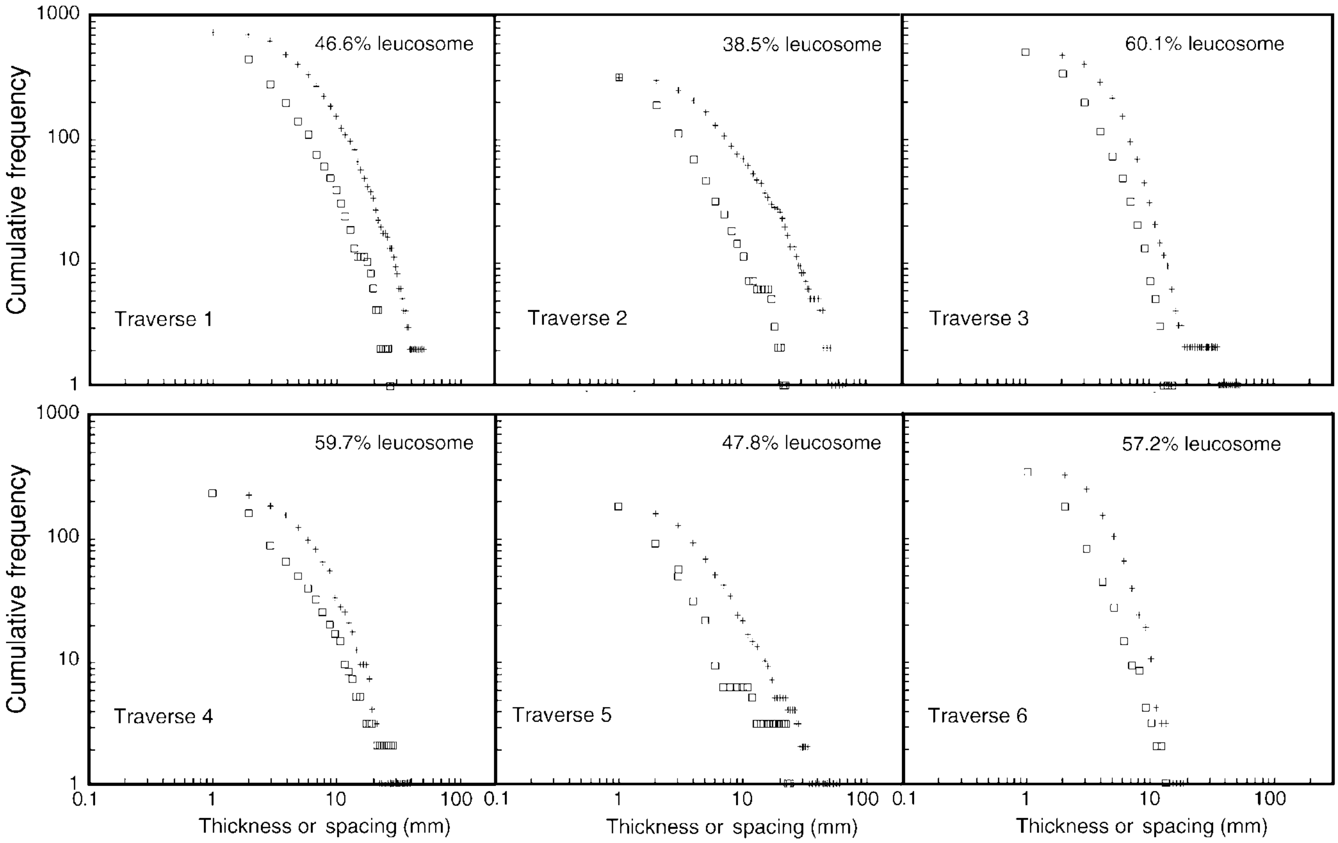

Figure 5 Log-log plots of foliation-parallel leucosome cumulative frequency versus thickness (open squares) or spacing (upright crosses) of leucosome stromata from one-dimensional traverses (after Marchildon & Brown 2003).

the conduit of 10^{-3}–10 km^3 melt per km^2 crust y^{-1}. These estimates are misleading because the melt flux falls away as distance squared from a cylindrical conduit and as distance cubed from a point sink. The mismatch between estimated transport rates in the conduit and rates at the boundary of the conduit also suggests that the MF probably increases as the conduit is approached. It requires a more rigorous analysis outside the scope of the present paper to take into account the spatial variation in MF.

The present author notes that the range of estimates for melt extraction calculated here is similar to the range of estimates derived by Harris *et al.* (2000) of 10^4–10^2 y, based on accessory phase thermometry. Furthermore, these authors calculated that melt may ascend through dykes from lower to upper crust in $<10^{-2}$ y. These rates are fast in comparison with estimates of the time scale for prograde metamorphism to the metamorphic peak, which is on the order of millions of years. This suggests a mechanism for the punctuated nature of upper crustal magmatism during orogenesis. The melt production rate in the lower crust is determined by heat flow, which means that the time scale to achieve interconnection of melt in the source determines the beginning of a melt extraction window. Once interconnection of melt has been achieved, the time scale for extraction and ascent is short in relation to the time scale associated with orogenesis, or prograde and retrograde metamorphic evolution (cf. Harris *et al.* 2000).

4. A paradigm for melt extraction from lower continental crust of orogens

Field studies have characterised the spatial and temporal granularity, and petrological studies have characterised the thermal evolution of migmatites in an attempt to understand the physical process of melt extraction from anisotropic lower crustal rocks. The sequence of mesoscale leucosome- and granite-filled structures described from lower continental crust records a transition from accumulation to draining, and melt is inferred to have been extracted by gravity-driven ascent in steeply oriented conduits. This is the outcome of multiple studies (e.g. Collins & Sawyer 1996; Brown & Solar 1999; Sawyer *et al.* 1999; Marchildon & Brown 2003), and although the specific details vary from example to example, the generic conclusion is robust. The inherent complexity of residual migmatites and granulites has led to several attempts to characterise the spatial patterning by fractal dimensions (Marchildon & Brown 2003) or to determine leucosome connectivity using probability (Petford & Koenders 1998; Weinberg 1999). However, in the present study, the author considers the dynamic evolution of melt-bearing deformation band networks and melt extraction from the source via ductile fracturing; this is a new approach that takes into account the ductile behaviour of lower crustal rocks, particularly under suprasolidus conditions.

4.1. Grain-to-vein scale

Mesoscale concentration of melt may occur in one of two ways (Powell & Downes 1990; Hand & Dirks 1992; Maaløe 1992; Brown *et al.* 1995b; Watson 2001). First, mass transfer may be by diffusional flow – chemical migration of components through a matrix by a dissolution-diffusion-crystallisation process in the matrix through which it moves (e.g. in metasedimentary protoliths with compositional anisotropy, formation of leucosome stromata with melanocratic selvedges in which the grain size is coarser than that for the same mineral in the matrix; and for hydrate dissolution under suprasolidus conditions, concentration of leucosome around a peritectic phase).

Secondly, mass transfer may be by advective flow – physical migration of melt through the matrix without chemical interactions (e.g. preferential movement of melt through structures with high melt fraction to draining conduits).

Average rates of diffusive mass transfer yield characteristic distances which may approach one metre on time scales around one million years, but there are many variables which affect rates and efficiencies of diffusion (e.g. Costa *et al.* 2003; Dohmen & Chakraborty 2003). In contrast, advective fluid flow is several orders of magnitude faster (e.g. Harris *et al.* 2000). Both processes may be inferred to have occurred in any one suprasolidus episode, based on evidence in individual outcrops of anatectic rocks, and these processes commonly act in concert. Coupling between mechanical and physico-chemical aspects of rock behaviour is significant in the anatectic environment, but the complex interplay among melt fraction, melt segregation, bulk strength contrasts, mechanical properties of different minerals and the melt phase, and time means that the rheology is difficult to describe by a single flow law (e.g. Berger & Kalt 1999).

Gradients in melt fraction will affect density and buoyancy forces, and may necessitate separated flow of solids and melt rather than a bulk response. However, what happens to such a mixture of solids and melt, in which the potential range of viscosities may approach 14 orders of magnitude and each end member has a different rheology (e.g. power law with $n=3$ for the solid and Newtonian with $n=1$ for the liquid), will depend upon the level of applied stress, the differential stress, the effective stress and the strain rate, the melt fraction and changes in the proportion of melt to solid with time, and whether localisation occurs. At high strain rates, the contrast in effective viscosity is reduced, and the bulk response most likely will be coupled flow (e.g. dominated by the matrix or at high melt fraction behavior as a pseudo-fluid), whereas, at low strain rates, the rheology may be dominated by one of the phases, and phase separation (strain rate compatibility) or strain localisation (stress compatibility) may occur (Vigneresse & Burg 2000; Vigneresse 2004).

It is clear that the key to understanding melt migration is an understanding of the nature of strain localisation in melt-bearing crust, and how localisation controls melt extraction. Discordant veins of leucosome in residual migmatite and granulite have commonly been interpreted as resulting from brittle fracture induced by embrittlement as a result of high melt pore pressure (e.g. van der Molen & Paterson 1979). However, fracture toughness analysis of crystalline solids indicates that there is a change in failure from brittle to ductile modes with increasing temperature (e.g. Hirth & Tullis 1994; Hirsch & Roberts 1997; Stockhert *et al.* 1999). At middle and lower crustal depths, below the brittle-to-viscous transition zone, lower strain rates, higher temperatures and confining pressures, and the presence of a volatile or melt phase attributable to dehydration reactions or hydrate-breakdown melting reactions promote ductile processes.

For layers or elongate bands of differing rheology in extension, mean stress is higher in the weaker material, but in shortening, it is lower. Thus, in extension, the resulting difference in mean stress drives melt from layers of greater melting, which are weaker, down gradients in melt pressure to stronger layers (cf. Brown *et al.* 1995b). Also, for an individual melt-bearing elongate shear band that is stretching and thinning, mean stress is higher in the band and melt will be expelled (cf. Mancktelow 2002). Conversely, for an individual melt-bearing elongate shear band that is shortening and thickening, mean stress is lower in the band and melt is sucked into it. Also, fluid pressure enables melt to exploit planes of weakness in anisotropic protoliths, for example by dilating foliation and/or interfaces between layers of contrasting composition, in a fashion similar to that described by Barraud *et al.* (2004) from analogue experiments.

The structural control on sites of leucosome concentration is consistent with gradients in mean stress as the principal driving force for matrix compaction, melt segregation and metre-scale melt flow (Rutter 1997). Thus, melt may be pumped in cyclic fashion along shallowly oriented fabrics by a dilation–contraction mechanism, moving through foliation-parallel segments via connecting transverse structures. Also, melt may be pumped through shear systems from thinning to thickening shear bands, and as shear bands rotate into the field of finite extension, melt increasingly is expelled from the source.

The rate of volumetric strain by melt expulsion cannot exceed the rate at which the solid matrix deforms by shear-enhanced compaction, and therefore, the rate of melt expulsion is controlled by the rate of matrix compaction or the rate of volumetric strain is limited by the rate of melt expulsion. If the solid matrix is able to compact at a higher rate than the rate at which melt may be expelled, then a higher distortional strain will be the result, requiring melt-bearing shear bands to develop or more distributed flow to occur (Rutter 1997). Overall, the positive feedback relation enables interconnected networks of deformation bands to evolve to an optimum structure, controlled by the rates of strain accumulation (and melt production) in relation to rates of matrix deformation and melt expulsion.

4.2. Mesoscale-to-crustal scale

A critical point is reached at some combination of MF and distribution, probably contemporaneously with the developing ductile deformation band network reaching the percolation threshold, where conduits of a size that allows melt to flow out of the local source volume and upward through the crust are formed. For this to occur, it is likely that the MF will be in the range 0·05–0·2; such a conclusion is consistent with estimates of the MF in the crust beneath the Central Andes and the Tibetan Plateau (Schilling & Partzsch 2001).

The leucosome networks described from the southern Brittany migmatites are composed of foliation-parallel and transverse stromata (interboudin partitions and shear surfaces) which represent remnants of melt flow through a linked network of deformation bands. The present author suggests that leucosome networks in residual migmatites and granulites, such as those in southern Brittany, are analogous to the ideal deformation band networks described by Du Bernard *et al.* (2002). Ideal deformation band networks are composed of three kinds of structure (Fig. 6): shear bands, characterised by a dominant shear displacement gradient that is accompanied by porosity reduction or compaction, or porosity increase or dilation (Antonellini *et al.* 1994); compaction bands, characterised by a localised porosity reduction or compaction, and lacking macroscopic shear offset (Mollema & Antonellini 1996); and dilation bands, characterised by localised porosity increase and opening-mode with respect to their boundaries (Du Bernard *et al.* 2002).

The dynamic evolution of these networks is illustrated by foliation-parallel leucosome stromata in residual migmatites and granulites. Assuming layer-parallel extension, foliation planes (or stronger layers) initially dilate to accumulate melt, but as transverse structures form, so the foliation (or melt-rich layer) contracts as melt flows into the developing transverse structures down gradients in fluid pressure. The present author infers that foliation-parallel leucosome stromata in melt-depleted crustal rocks are structures analogous to compaction bands. As an interconnected network of deformation bands develops, the system will approach the percolation threshold,

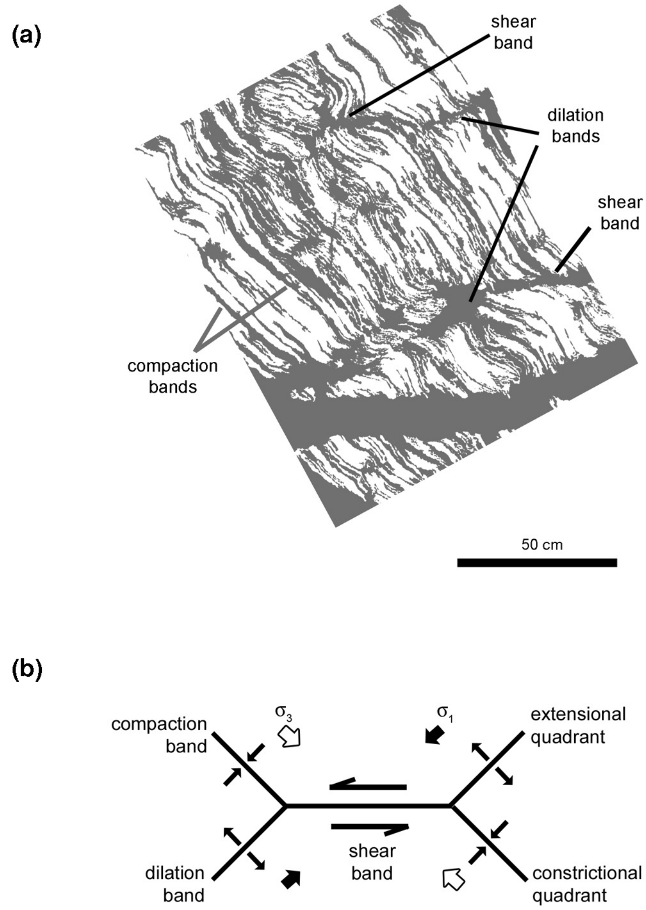

Figure 6 (a) Binary image of (sub-horizontal) surface perpendicular to foliation and parallel to lineation to show patterning of leucosomes (after Marchildon & Brown 2003; Le Petit Mont, Arzon, Morbihan). (b) Idealised diagram to show the coexistence of shear, compaction and dilation bands (σ_1 and σ_3 are greatest and least far field principal stresses, respectively; note that locally the effective stress parallel to σ_3 will be tensile because of melt pressure). Modified from Du Bernard 2002 by permission of the American Geophysical Union. © 2002 American Geophysical Union.

the permeability will change rapidly with concomitant increase in melt flow rates, and melt will drain from the system if appropriate conduits for ascent are available or conditions are appropriate for these to form. In the residue, a mix of residual and early crystallised (cumulate) phases with only minimal melt is retained as leucosome stromata in the deformation bands because shear and compactive deformation has expelled most of the melt from these sites into ductile opening-mode fractures and out of the local source volume as dykes (cf. Solar & Brown 2001; Guernina & Sawyer 2003).

The common view that melt extraction is a brittle fracturing phenomenon from the source to the upper crust is based on linear elastic fracture mechanics in which inelastic deformation in a small process zone ahead of the fracture tip, with or without chemical weakening (subcritical brittle fracture or brittle fracture, respectively), is driven by the magnified stresses at the macroscopic fracture tip (e.g. Rubin 1995; Eichhubl 2004). However, the application of brittle fracture theory to the suprasolidus environment is not intuitively obvious since it neglects thermally-activated flow processes. These thermally-activated processes overtake brittle processes with increasing depth to dominate in the crust below the brittle-to-viscous transition zone, which leads to extensive inelastic deformation, the blunting of crack tips and a change in fracture mode. Processes similar to those in creep failure of ceramics at high homologous temperature under low rates of loading, such as nucleation, growth and coalescence of pores by predominantly diffusive deformation mechanisms and side-lobe damage, are likely to be particularly relevant to failure of suprasolidus crust, leading to ductile fracture (cf. Eichhubl 2004).

As a hypothesis, the present author suggests that brittle theory is unlikely to be applicable to depths beyond 10–15 km and at elevated temperatures, and that ductile fracturing dominates in the lower crust. There are a number of geological features which characterise the nature of the dominant fracturing mechanism in the lower crust. These features include the nature of the tips of melt-filled fractures (which are blunt in ductile fracture but sharp in brittle fracture), the geometry of coalescing melt-filled fractures (which zigzag in ductile fracture but form *en echelon* arrays in brittle fracture), and the nature of the margins of the fracture, and between the host and the fracture-filling material (irregular margins indicate ductile fracture whereas sharp margins indicate brittle fracture).

In suprasolidus lower crust, the author suggests that fracture propagation takes place by development and coalescence of melt-filled pores ahead of the fracture tip, with fracture opening involving extensive inelastic deformation and diffusive mass transfer (cf. Regenauer-Lieb 1999; Eichhubl *et al.* 2001; Eichhubl & Aydin 2003; Eichhubl 2004). This coupling between fracture formation and mass transfer is significant for both opening- and shearing-mode fracturing in a suprasolidus crustal environment, where a large amount of plastic strain may accumulate before fracture. In this environment, the differential stresses are low, the effective mean stress is low, the cohesive strength is low (relative to the strength of the grains) at melt fractions greater than $\sim$7–8 vol.%, and the rapid exhumation of the lower crust means that the loading rate is fast, which are the conditions necessary for ductile failure (P. Eichhubl, pers. comm., 2003).

In southern Brittany, dyke emplacement occurred along a preferred direction independent of anisotropy, suggesting that the orientation is controlled by stress, which with the macroscopic fracture-like discontinuity infilled by the granite, characterises the process of dyke formation as a fracture phenomenon. The petrographic continuity between leucosome in the deformation band networks and granite in dykes (Fig. 3) indicates that the ascent conduits represent either structures analogous to large-scale dilation bands formed by opening-mode failure along zones of localised porosity increase or ductile opening-mode fractures formed by pore growth and coalescence of melt pockets (e.g. Regenauer-Lieb 1999; Eichhubl *et al.* 2001; Du Bernard *et al.* 2002; Eichhubl & Aydin 2003). Zigzag propagation paths at tips of dykes (Fig. 3e, f), which commonly propagate from transverse structures, point to ductile fracture (Fig. 7) as the mode of formation of these conduits (cf. Eichhubl & Aydin 2003).

Whether each ascent conduit was the product of a single melt-draining event from its collection zone or whether there is a build-up of melt pressure before periodic activation of a draining conduit in a cyclic fashion (e.g. Handy *et al.* 2001) must be determined on a case-by-case basis (e.g. Solar & Brown 2001; Marchildon & Brown 2003). However, there is increasing evidence of common ascent conduits for different batches of melt (e.g. Brown & Solar 1999; Solar & Brown 2001; Sawyer & Bonnay 2003). Thus, studies of exhumed melt-depleted anatectic migmatite and granulite terrains suggest that melt may be lost in batches. This view is supported by detailed isotope geochemistry from some plutons (e.g. the Phillips pluton in Maine, northern Appalachians, USA) where data from representative samples are interpreted to show that the pluton was constructed from multiple batches of isotopically distinct melt (Pressley & Brown 1999). Furthermore, this

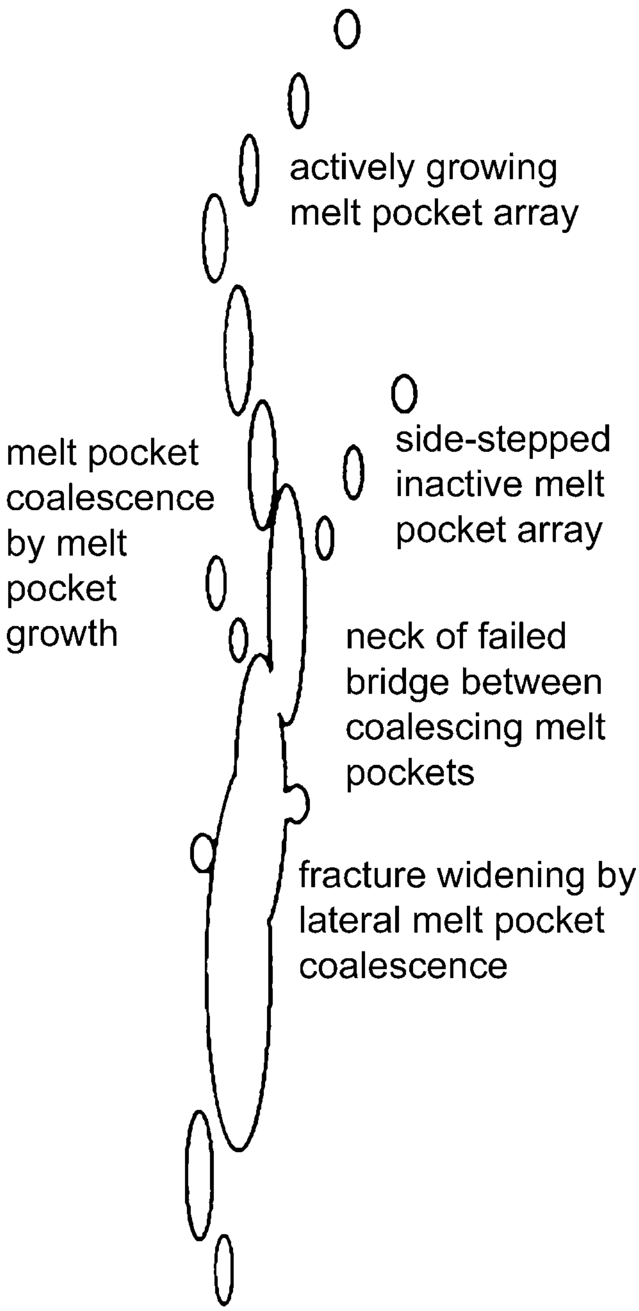

Figure 7 Sketch to illustrate the mechanism of formation of ductile opening-mode fractures by pore growth and melt pocket coalescence. Switching between left- and right-stepping direction leads to zigzag propagation paths. Modified from Eichhubets & Aydin 2003 by permission of Elsevier B.V. © 2003 Elsevier B.V.

conclusion is consistent with the occurrence of clearly discordant, but petrographically continuous dykes at the same structural level in the source as developing ductile opening-mode fractures (Figs 2 & 4); thus, at any particular structural level in the source, newly forming conduits and conduits transporting melt through that level will occur together.

4.3. Relationship to other postulated mechanisms of ascent

The model proposed in this paper combines elements of pervasive migration with dyking by ductile fracture; there is no *a priori* reason why these dykes cannot be the physical expression of mobile hydrofractures frozen in ascent. It is also possible that such a model also may represent the physical expression of successive porosity waves passing through the anatectic zone.

5. Self-organisation

Self-organisation is a fundamental property of natural systems, including orogens, driven far from equilibrium by a constant energy flux (e.g. Hodges 2000). The segregation, extraction and ascent of magma in orogens have been postulated to be self-organised, resulting from non-linear interactions and feedback relations among deformation and melting (Brown & Solar 1998a, b). Sources of non-linearity include spatial and temporal variation in melting, melt production and melt distribution, and the combination of linear and non-linear rheologies in the melting environment. Thus, deformation during anatexis induces strain partitioning at all scales, increasing the vorticity within the weaker phase, which promotes melt redistribution and bulk softening (shear thinning), and further concentrates strain in a positive feedback relationship that induces increasing non-coaxial deformation and accelerates melt displacement. As a result, the large-scale interaction between melting and deformation, and granite magma ascent and emplacement has been proposed (e.g. Brown & Dallmeyer 1996; Brown & Solar 1998a, b), although such an intimate feedback relation may not always be the case (e.g. Fitzsimons 1996). However, the studies to date have provided only qualitative descriptions of the interactions and feedback relations, and a quantitative analysis of how these non-linear effects interact in the differentiation of heterogeneous, anisotropic lower continental crust during orogenesis has not been undertaken.

One feature of self-organised phenomena is that these are driven systems which may achieve a critical state, which may lead to 'avalanches' with a fractal (power-law) frequency-size distribution (Turcotte 1999, 2001). Thus, if melt extraction reflects a critical state in melt-bearing lower crust, we might expect the distribution of melt batch sizes and/or melt flow channels to be fractal. Whereas the system as a whole may converge to a critical state that allows melt escape, fractal (scale invariant) distribution of melt batches (or leucosome-filled structures) has not yet been demonstrated satisfactorily (see Tanner 1999; Marchildon & Brown 2003; Bons *et al.* 2004). This may reflect the residual nature of the drained source in depleted anatectic migmatite and granulite terrains. In a recent example of modelling melt segregation from a homogeneous isotropic mush subject to both compactive and shear deformation, Rabinowicz & Vigneresse (2004) demonstrated that instabilities will generate parallel melt-filled veins in the σ_1–σ_2 plane, with size and spacing close to the compaction length, which is expected to be metric or sub-metric. However, the above authors do not address how this modelling relates to observations from field examples of exposed lower continental crust, where the tectonically induced metamorphic fabric clearly exerts the primary control over leucosome orientation and spacing.

6. Conclusions

Melt extraction from the lower continental crust of orogens involves the migration of melt down gradients in pressure over distances of millimetres by flow along grain boundaries to form layer or foliation-parallel leucosome stromata. With increasing melt fraction, melt moves distances of tens of centimetres along these concordant structures to developing dilatant transverse structures, such as ductile extension and shear bands, which become interconnected. These ductile deformation band networks, and similar networks of dilatant structures in examples described in the literature, represent accumulation networks in melt-bearing lower continental crust. Melt is extracted from the accumulation networks by porous flow down-gradients in pressure to ductile opening-mode fractures and shear zones, where ascent is buoyancy-driven. Extraction occurs at some critical combination of melt fraction and distribution, most probably as the developing accumulation network reaches the percolation threshold. The partially molten lower continental crust of orogens is driven far from thermal and mechanical equilibrium in relation to the adjacent subsolidus crust, so that melt extraction is via emergent structures which enable dissipation of excess energy in the system.

7. Acknowledgements

I thank Boswell A. Wing for assistance with the flux calculations, Margaret A. Baker and Barry Reno for help in the preparation of figures, and Jeanne Martin for help in the preparation of the text. This paper was improved as a consequence of review comments by Keith Benn, Jean-Louis Vigneresse and Roberto Weinberg; however, any remaining infelicities are the responsibility of the author alone.

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MS received 29 September 2003. Accepted for publication 18 June 2004.