

Journal of the Geological Society

Crustal melting and melt extraction, ascent and emplacement in orogens: mechanisms and consequences

Michael Brown

Journal of the Geological Society 2007, v.164; p709-730.
doi: 10.1144/0016-76492006-171

Email alerting service	click here to receive free e-mail alerts when new articles cite this article
Permission request	click here to seek permission to re-use all or part of this article
Subscribe	click here to subscribe to Journal of the Geological Society or the Lyell Collection

Notes

Bicentennial Review

Crustal melting and melt extraction, ascent and emplacement in orogens: mechanisms and consequences

MICHAEL BROWN

*Laboratory for Crustal Petrology, Department of Geology, University of Maryland, College Park, MD 20742-4211, USA
(e-mail: mbrown@geol.umd.edu)*

Abstract: Significant volume of wet melting requires an influx of H₂O-rich volatile phase. In hydrate-breakdown melting, initial melt accumulation is diffusion-controlled and melt accumulates around peritectic phases in low-pressure sites. As the melt-bearing rock weakens, it becomes porous at a few volume per cent melt, initiating an advective flow regime; as melt volume reaches the melt connectivity transition, melt may be lost from the system in the first of several melt-build-up–melt-loss events. Using mineral equilibria modelling, major and accessory phase controls on melt chemistry are evaluated. In residual migmatites and granulites, microstructures indicate the former presence of melt whereas leucosome networks record melt extraction pathways. Sites of initial melting nucleate shear instabilities; strain and anisotropy of permeability control the form of millimetre- to centimetre-scale inferred melt channels and strong anisotropy promotes high fluid focusing. Focused melt flow occurs by dilatant shear failure of low melt volume rocks, which leads to the formation of melt flow networks allowing accumulation and storage of melt, and forming the link for melt flow from grain boundaries to ascent conduits. Melt ascent is via ductile fractures, which may propagate from dilation or shear bands. Fractures are characterized by blunt tips; also they may exhibit zigzag geometry close to the tips and petrographic continuity between leucosome in the host and granite in the dyke. Horizontal tabular and wedge-shaped intrusions commonly are associated with the brittle–ductile transition zone. Vertical lozenge-shaped intrusions represent congelation of magma in ascent conduits and blobby plutons record lateral expansion localized by instability; these intrusion types characterize emplacement at deeper levels in the crust.

Melt-bearing rocks are present in a variety of tectonic settings in the Earth's crust and mantle. They occur beneath mid-ocean spreading centres and hotspots, underneath volcanic arcs and back-arcs at intraoceanic subduction zones (e.g. the Izu–Bonin–Mariana arc system), and under orogenic plateaux associated with continental arcs (e.g. the Andes) and subduction-to-collision orogens (e.g. the Himalayas). Melt-bearing rocks play an important role in continental tectonics.

Melt-bearing rocks are the source of magmas that create crust at mid-ocean spreading centres, generate calc-alkaline volcanic and plutonic complexes in oceanic and continental arcs, and form leucogranites in the later stages of subduction-to-collision orogenesis. At mid-ocean spreading centres the creation and lateral translation of oceanic lithosphere is the principal mechanism by which the Earth loses heat to space, whereas oceanic and continental arcs represent the principal sites of processing of mass additions from the mantle into future continental crust. In the terminal stage of subduction-to-collision orogenesis, melt extraction and magma ascent and emplacement represent a significant heat and mass transfer from lower to upper continental crust.

Melt-bearing horizons in the crust form mechanically weak zones, because the presence of even a low volume of melt significantly lowers rock strength (e.g. Rosenberg & Handy 2005). Within the last decade, we have gained new insights from geophysical data and modelling on the role of melting in the development of orogenic belts (e.g. Nelson *et al.* 1996; Babeyko

et al. 2002). The geophysical data have enabled us to quantify the amount of melt present under orogenic plateaux (Schilling & Partzsch 2001; Unsworth *et al.* 2005), which is thought to be the weakening mechanism that enables the crust to flow in response to gravity or tectonic stress (Clark & Royden 2000).

We know from geophysical surveys that granites are not distributed throughout the full thickness of the continental crust, but are concentrated in the upper part. From a geochemical perspective, extraction of melt from lower crust and its transfer to upper crust is the principal process by which continents became differentiated into a more mafic, minimally hydrated and residual lower portion and a more felsic, more hydrated and incompatible element-enriched upper portion (Brown & Rushmer 2006). This process stabilizes continental crust, as the more felsic upper crust is less dense than the more residual lower crust.

This Bicentennial Review is concerned with the mechanisms that allow and control the extraction of melt from anatectic lower continental crust, and the processes of ascent and emplacement of magma into the upper continental crust, and the consequences for crustal differentiation and rheology. It is the fourth in a series of reviews dating back to the Volume 150 Celebration Paper by Atherton (1993), which took a petrological–geochemical view of granite magmatism and included a historical perspective, and the review of crustal melting by Brown (1994), which took a dynamic view of melt extraction and stressed the role of deformation in melt extraction and magma emplacement; it

complements the views expressed in the paper by Petford *et al.* (2000).

What are the main issues?

There has been a tendency to separate melt generation and extraction from magma ascent and emplacement. This separation is artificial: a simple succession of events is unlikely and feedback relations are to be expected (Brown & Solar 1998a, 1999; Brown 2001a, b, 2005). However, linking granites with their sources is a challenge because open-system processes conspire to obscure relationships. Although we may think of sources in terms of average bulk composition and granites as uncomplicated with respect to major oxides, we know that these general features represent an oversimplification (e.g. Hogan & Sinha 1991; Brown & Pressley 1999; Zeng *et al.* 2005a). Therefore, we must look through the apparent simplicity to quantify the heterogeneity of lower crustal sources and upper crustal intrusions, to characterize the mesoscale spatial and temporal patterns of melt segregation and accumulation, and to understand the mechanisms of melt storage and extraction and, at the crustal scale, the processes of magma ascent and granite emplacement.

Melting occurs at distributed multiphase grain boundaries in a source (c. mm³), whereas a granite intrusion represents a large volume of melt (c. 10³–10⁴ km³) accumulated at a sink and crystallized within tens of thousand years to several million years (Solar *et al.* 1998; Coleman *et al.* 2004; Matzel *et al.* 2005, 2006). Melt extraction-to-emplacement is a process that spans >21 orders of magnitude by volume and that occurs on a time scale of c. 10¹²–10¹⁴ s (Brown 2004, 2006). Melt extraction from the source resolves to linkages between flow regimes at different scales; namely, pervasive grain-scale flow, porous flow in veins and channel flow in ascent conduits.

The height from the anatexis front (the upper surface of the melt-bearing source, a dynamic feature during orogenesis (Brown & Solar 1999)) to the level of emplacement varies according to the degree of orogenic thickening and amount of syncollision extension or orogenic collapse, but is rarely going to be more than 30 km and commonly will be (much) less (Fig. 1). The thickness of the source is likely to be <20 km. Because the number of feeders to an individual intrusion varies from one to several (Cruden 2006), and the footprint of the source that is drained for any single intrusion is unlikely to be smaller than the lateral extent of the granite (Fig. 2), the issue becomes one of focusing the melt. The spacing of intrusions in continental arcs and subduction-to-collision orogens (Bons & Elburg 2001; Cruden 2006) confirms that ascent is spatially focused.

Focusing melt flow to ascent conduits

There are several ways by which focused flow of melt to ascent conduits may be achieved. In the case of strong intrinsic or extrinsic anisotropy of permeability, melt will flow along the anisotropy. In the case where thickened lower crust is flowing outward as a result of differences in gravitational potential energy, the fabric and the anisotropy of permeability in the source are likely to be subhorizontal and melt flow will be down gradients in pressure to an ascent conduit that propagates out of the source or to a crustal-scale tectonic discontinuity such as a shear zone and then upwards to the emplacement level (e.g. Brown 2004, 2005, 2006; Fig. 1). Focusing also may be achieved in the case of deformation partitioning during transpressional thickening where the fabrics and the anisotropy of permeability

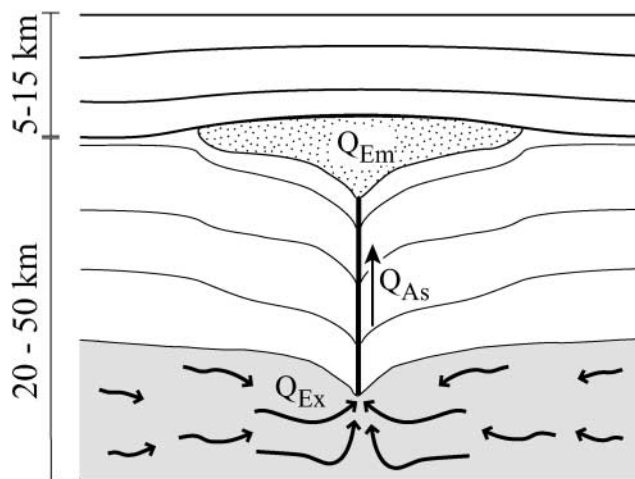


Fig. 1. A simple model of melt extraction, ascent and emplacement, in which the rates of melt production, extraction, ascent and emplacement ultimately are balanced at the crustal scale. The rate of melt production (Q_{Pr}) is controlled by heat flow, and is not shown explicitly in the figure. Volume fluxes for melt extraction (Q_{Ex}), ascent (Q_{As}) and emplacement (Q_{Em}) are shown schematically (Brown 2006). The depths shown on the left-hand side of the figure are intended as a general indication of the likely range in orogens.

will be steep rather than subhorizontal and ascent conduits will be concordant rather than discordant, as demonstrated by Solar & Brown (2001b; see also Brown & Solar 1998a, 1999) for the Acadian belt in Maine, northern Appalachians.

In the case where the anisotropy of permeability is weak, focusing may occur only after melt has accumulated at the top of the source against an impermeable barrier. Melt may flow laterally along the boundary layer to a localized area of dyke propagation and then upwards to the emplacement level in a manner similar to one model originally proposed for focusing basalt melt to mid-ocean spreading centres (Sparks & Parmentier 1991) and implied in the model for melt generation and segregation via buoyancy-driven compaction by Jackson *et al.* (2003). However, a model of anastomosing steeply oriented channels connecting porous flow to a single ascent conduit (Fig. 3a) is inappropriate at the crustal scale given the degree of horizontal flow required by the geometry of extraction to emplacement (Figs 1 and 2). A network similar to that shown in the lower half of Figure 3a (up to the level of the sill) may be an appropriate analogue for the model of Jackson *et al.* (2003).

Extraction to emplacement

In exposed lower crustal sources, leucosome is inferred to record remnants of the melt flow network (Brown 1994, 2001a, b, 2004, 2006; Collins & Sawyer 1996; Oliver & Barr 1997; Brown *et al.* 1999; Sawyer 2001; Guernina & Sawyer 2003; Marchildon & Brown 2003; White *et al.* 2004), and interconnection of leucosome has been demonstrated from grain to vein scale in multiple studies (Brown 1994, 2004, 2006; Oliver & Barr 1997; Marchildon & Brown 2001, 2002, 2003; Sawyer 2001; Guernina & Sawyer 2003; Fig. 3b and c). However, how these networks connect to magma ascent conduits remains enigmatic, although theoretical and phenomenological models have been proposed (Petford & Koenders 1998; Bons *et al.* 2001, 2004). I have postulated elsewhere that melt accumulates in deformation band networks in suprasolidus lower crustal rocks and is driven down

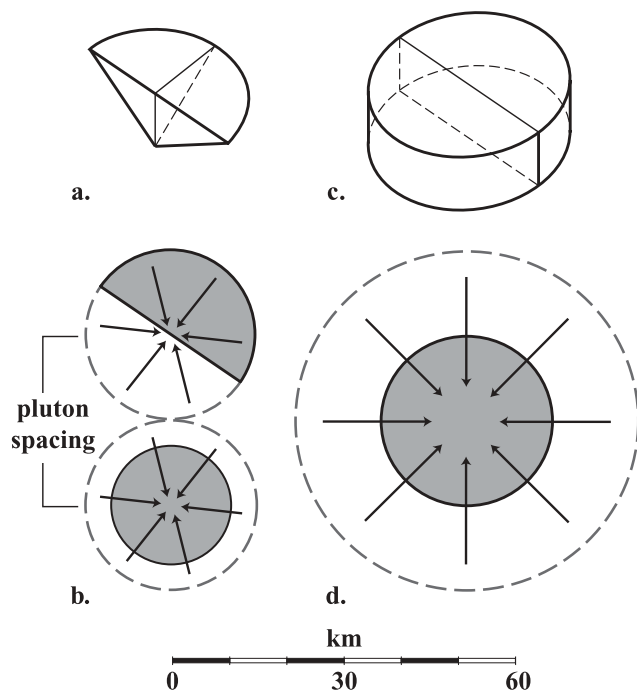


Fig. 2. Pluton-source relations in orogens. (a) Perspective view of horizontal semi-circular half-cone-shaped pluton of diameter about 30 km and half-height about 5 km. (b) Plan view of a horizontal semi-circular half-cone-shaped pluton of diameter about 30 km and a horizontal circular cone-shaped pluton of diameter about 21 km in relation to the footprint for a source of sufficient volume to fill the pluton (for a source about 30 km in diameter, based on an assumed thickness of 15 km and 20 vol.% melting with 20 vol.% fractional crystallization of the melt during transport through the source). One implication of this model for the assumptions made is that plutons will be spaced at least 30 km apart. (c) Perspective view of a horizontal tabular intrusion of diameter about 30 km and thickness about 10 km, which is four times greater volume than for the plutons discussed in (b). (d) Plan view of a horizontal tabular intrusion of diameter about 30 km in relation to the footprint for a source of sufficient volume to fill the intrusion (for a source about 60 km in diameter, based on an assumed thickness of 15 km and 20 vol.% melting with 20 vol.% fractional crystallization of the melt during transport through the source). One implication of this model for the assumptions made is that intrusions will be spaced at least 60 km apart. A more sophisticated model of pluton-source shape and volume, and melt extraction and accommodation, has been given by Cruden & McCaffrey (2001) and Cruden (2006).

gradients in pressure to dykes that are initiated by ductile fracture (Brown 2004, 2006).

Lower crustal melting is driven by heat and is a continuous process, although the rate of melt production, as a function of T and t , may vary with the enthalpy demands of melting in relation to the heat input to the system and the specific melting reaction(s) due to thermal buffering. In contrast, melt extraction generally is considered to be a discontinuous process that is cyclic (e.g. Brown & Solar 1998a; Handy *et al.* 2001; Rabinowicz & Vigneresse 2004). This cyclic behaviour is confirmed by the cryptic internal structure of many intrusions, identified by chronology and isotope chemistry (e.g. Brown & Pressley 1999; Pressley & Brown 1999; Coleman *et al.* 2004; Matzel *et al.* 2005, 2006), which requires that intrusions are composite bodies constructed from multiple batches of magma. The switch from a continuous to a discontinuous process is fundamental and has

been discussed before (e.g. Brown & Solar 1998a; Handy *et al.* 2001); it is a consequence of the nonlinear feedback relations that develop during melting, melt extraction and magma ascent.

In the upper continental crust, we are interested in why buoyancy-driven ascent changes to subhorizontal magma emplacement to form horizontal tabular or wedge-shaped intrusions or simply decreases to choke the ascent column, forming vertical lozenge-shaped intrusions or blobby plutons by lateral expansion (Brown 2001a, b). At long time scales there is a balance between mass gain in the upper continental crust and mass loss from the middle and lower continental crust that accommodates the crustal-scale issue of space for magma emplacement (Fig. 1). Commonly, local mechanisms of final emplacement and adjustment, particularly downward transport of host rock by magmatic stopping or downward flow by ductile strain, obscure the overall syntectonic nature of the process (e.g. Brown & Solar 1999; Brown 2001a, b; Cruden 2006).

Self-organized criticality

A critical point in a system is where a change occurs in behaviour or structure; in standard critical phenomena, such as melting, there is a control parameter (temperature in the case of melting) that may be varied to cause the change. In contrast, self-organized criticality is a property of slowly driven non-equilibrium systems with extended degrees of freedom and strong non-linearity; such systems exhibit spatial and/or temporal scale-invariance and reach a critical state as a result of their intrinsic dynamics, independent of the precise values of control parameters (Turcotte 1999, 2001). Based on results from numerical and analogue modelling (Vigneresse & Burg 2000; Bons & van Milligen 2001), melt extraction from anatectic continental crust has been argued to be a self-organized critical phenomenon similar to that proposed for magma transport in Hawaii (based on an analysis of tremors; Shaw & Chouet 1991) and basalt melt flow through the shallow mantle beneath ocean spreading centres (as recorded by dunite-filled conduits; Braun & Kelemen 2002).

In the case of natural crustal melting, we might expect the distribution of melt batch sizes and/or melt flow channels to be scale-invariant. Stromatic leucosomes, which are commonly controlled by the strain field, do not appear to be scale-invariant (Marchildon & Brown 2003); in contrast, the limited data for ascent conduits (e.g. Tanner 1999; Bons *et al.* 2004; Brown 2005) and intrusions (McCaffrey & Petford 1997; Bons & Elburg 2001; Koukouvelas *et al.* 2006) suggest scale-invariance. Thus, granite ascent and emplacement may be a self-organized critical phenomenon. The challenge we face is to develop an internally consistent model that integrates heat transfer and melting with melt extraction from a permeable and deformable matrix involving chemical interactions during transport, and to link these to ascent and emplacement mechanisms that may both predict and explain the nature and cause of this self-organization (Cruden 2006).

Why study residual migmatites and granulites?

Residual migmatites and granulites represent former melt-bearing crust and the sources of magma for upper crustal intrusions; such rocks represent both zones of melt generation and zones of melt transfer (e.g. Brown 2004; Olsen *et al.* 2004; Slagstad *et al.* 2005). One consequence of this flux of melt is that leucosomes are rarely primary melt compositions. Commonly, leucosomes exhibit either strong positive Eu anomalies and low Zr, features that are expected of leucosomes formed during the early stage of segregation and extraction and composed of predominantly early

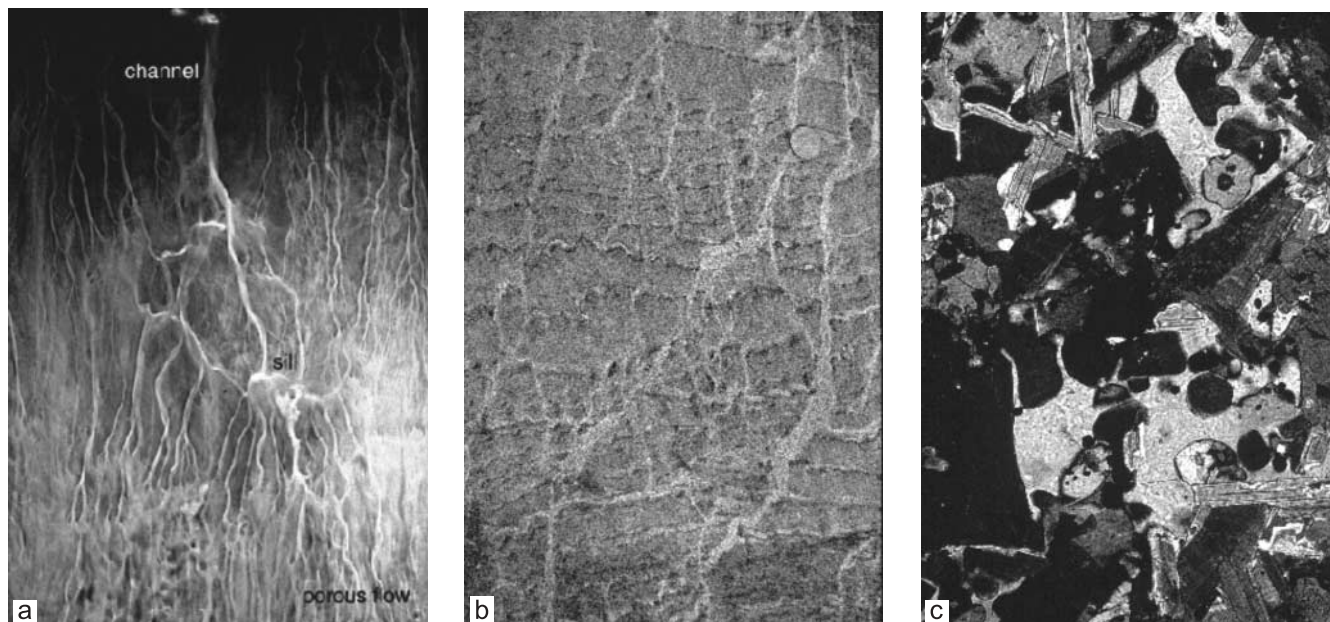


Fig. 3. (a) Air bubbling up through a fine powder constrained between two glass plates tilted at a 45° angle. The air bubbles create small upward-anastomosing channels; subsequent air bubbles follow the paths of least resistance, further deepening and elaborating the drainage network. (b, c) Contact migmatites from the middle-crustal Onawa contact aureole, Maine, USA. (b) Folded layer-parallel granite leucosomes are flanked asymmetrically by melanosomes and disrupted and crosscut by a network of discordant but petrographically continuous (microstructure, mineralogy and mode) granite leucosomes. Coin diameter is 19 mm. (c) Mesosome between the granite leucosomes in (b) shows evidence of having been partially melted, such as the K-feldspar pockets (light grey) with included faceted quartz grains shown here; the melt apparently did not segregate into mesoscopic connected pockets or veins (long dimension of image is c. 300 µm).

crystallized feldspars and quartz (Sawyer 1987; Solar & Brown 2001a, b; Johnson *et al.* 2003b), or compositions with strong negative Eu anomalies and high Zr, features that are expected of leucosomes formed from the last dregs of percolating fractionated melt trapped during cooling of the terrane (Sawyer 1987, 1998; Solar & Brown 2001a, b; Hinchey & Carr 2006). The rationale behind this contrast in leucosome compositions is as follows.

As melt segregates and migrates through channels it encounters host rock that may be cooler than the liquidus for the melt composition, which causes crystallization of liquidus phases (probably feldspar and quartz) on the channel walls. There is a continuous process of crystallization and liquid fractionation during extraction and ascent until the evolved melt encounters host rocks cold enough to freeze it at the margins of the conduit, generally as the magma crosses the anatexis front, after which the evolved liquid ascends to the upper crustal intrusion where it crystallizes. It is for this reason that leucosomes in residual migmatites and granulites are inferred to record the melt flow network (Brown *et al.* 1995; Collins & Sawyer 1996; Brown & Rushmer 1997; Brown 2001a, b, 2004; Sawyer 2001; Guernina & Sawyer 2003); this is also the evidence that residual source rocks are inferred to have been both zones of melt generation and zones of melt transfer (Brown 2004, 2006; Olsen *et al.* 2004).

A feature of some leucosomes is the presence of a peritectic phase, such as garnet, within the leucosome (not just at the edge or protruding from the melanosome or mesosome); for example, as imaged in the leucosomes of a stromatic migmatite from Southern Brittany using high-resolution computed X-ray tomography (Brown *et al.* 1999). This supports an inference derived from a comparison of experimental glass compositions with S-type granites of the Cape Granite Suite in South Africa that

the ferromagnesian contents of the granites are too high to represent melts and may result from selective entrainment of peritectic garnet and ilmenite (Stevens *et al.* 2007).

Water is partitioned into hydrous silicate melt. In a closed system, cooling potentially will lead to retrogression by reaction between melt and residue (Spear *et al.* 1999; Brown 2002; White & Powell 2002), unless segregation and melt migration has separated them; this raises the issue of equilibration domains, local equilibrium and effective bulk compositions (Brown 1983, 2002; Stüwe 1997; White *et al.* 2004). However, the common occurrence of pristine or weakly retrogressed residual migmatites and granulites suggests loss of melt from the lower crust (Waters 1988; Powell & Downes 1990; White & Powell 2002), which is consistent with the presence of leucogranites in the upper crust (e.g. Brown 2001b, 2005; Guernina & Sawyer 2003) and geochemical evidence that links leucogranites with lower crustal sources (e.g. Pressley & Brown 1999; Solar & Brown 2001a; Johnson *et al.* 2003b; Hinchey & Carr 2006).

Studies of residual migmatites and granulites allow multi-scale characterization of the (minimum) permeability of melt-bearing horizons in lower crust, as recorded by microstructures and hand-to outcrop-scale features in former melt-bearing rocks (Tanner 1999; Sawyer 2001; Marchildon & Brown 2002, 2003; Guernina & Sawyer 2003; Brown 2004; Bons *et al.* 2004). These studies provide constraints on realistic models of melt extraction and ascent.

What do we know about melting?

Earth's continental crust is composed of a diversity of rocks that neither melt equally nor behave uniformly in response to stress. Common crustal rocks may melt at temperatures as low as 650 °C, if H₂O-rich fluid is present (wet melting (high aH₂O

melting)). However, most crust-derived granite magmas (*sensu lato*) are formed by 'fluid-absent' melting reactions (hydrate-breakdown melting (moderate to low $a\text{H}_2\text{O}$ melting)), so that significant melt production occurs only at temperatures above about 750–800 °C. The amount of H_2O -undersaturated melt produced varies according to protolith composition and mineralogy, particularly the hydrous mineral species and mode, the depth of melting (pressure) and heat input (temperature), and the time above the solidus. Common crustal protoliths yield 10–50 vol.% of H_2O -undersaturated leucogranite, trondhjemite, granodiorite or tonalite melt at temperatures attainable during orogenesis (e.g. see the review by Clemens 2006).

Heat

The mechanism that provides the enhanced thermal flux necessary to drive crustal melting remains elusive in most circumstances. Radiogenic heating is important and sometimes is sufficient (Andreoli *et al.* 2006; McLaren *et al.* 2006). Hyndman *et al.* (2005; see also Currie & Hyndman 2005) have argued that mountain belts and orogenic plateaux occur in former subduction zone back-arcs, which are characterized by high-surface heat flow, $>70 \text{ mW m}^{-2}$ for continental crust with average radiogenic heat production, and thin weak lithosphere. Complementary processes associated with convergence and orogenic thickening (erosion, synconvergence extension and orogenic collapse) drive advective decompression that may lead to or enhance melting (e.g. Jones & Brown 1990; Harris & Massey 1994; Brown & Dallmeyer 1996; Whittington & Treloar 2002; Johnson & Brown 2004); exhumation of continental crust subducted to high- or ultrahigh-pressure metamorphic conditions also leads to melting during decompression (Labrousse *et al.* 2002; Auzanneau *et al.* 2006; Lang & Gilotti 2007).

Underplating by basaltic magma commonly is implicated to provide heat (e.g. Dewey *et al.* 2006), but the temporal relations between extension, crustal melting and basaltic underplating sometimes are unclear (e.g. Barboza *et al.* 1999). Advective heating by multiple intraplate of sills may be effective at melting the lower crust in continental arcs (e.g. Jackson *et al.* 2003; Annen *et al.* 2006a). The role of viscous dissipation or shear heating during deformation remains controversial but may be a contributory factor in the early stages of melting in subduction-to-collision orogens driven by fast rates of convergence (e.g. Leloup & Kienast 1993; Kincaid & Silver 1996; Harrison *et al.* 1998; Stüwe 1998; Burg & Gerya 2005). Finally, heat advection within the crust by pervasive melt flow or by ascent and emplacement of leucogranite magma may be important in some circumstances (Brown & Solar 1999; Weinberg 1999; Leitch & Weinberg 2002; Annen *et al.* 2006b).

The beginning of melting

Under equilibrium conditions in an isotropic crust and in the absence of an H_2O -rich volatile phase, melting begins at multiphase grain junctions that include quartz and feldspar and a hydrate phase. However, the Earth's crust is anisotropic and in a state of stress; variations in bulk composition and grain size influence sites at which melting begins, and differential stress varies within any volume of crust to generate pressure gradients.

Melting may begin at sites of lower or higher pressure, once the initial thermal overstep is close to that required to overcome the activation energy for the particular melting reaction (Brown & Solar 1998a). For wet melting, small increases in pressure may promote the start of reaction (because the wet solidus has a

negative slope in P – T space), whereas for hydrate-breakdown melting, small decreases in pressure may promote the start of reaction (because hydrate-breakdown melting reactions generally have positive slopes in P – T space at crustal pressures). Initially developed melt pockets may be over- or under-pressured, and may contract or expand, modifying dihedral angles dynamically, and lose or gain melt, according to whether pockets are interconnected or not, and over-pressured pockets may fail by ductile fracture.

Initial sites of melting along appropriately oriented multiphase grain boundaries weaken as melting progresses. Shear instabilities may develop at these sites or the intrinsic anisotropy of permeability may facilitate nucleation and lateral growth of melt pores after some critical volume of melt is achieved. These selective sites of strain accommodation grow to form melt-rich layers: melt is concentrated in these layers driven by the difference in normal stress, further enhancing their ability to accommodate strain. Feedback between deformation and melting creates a dynamic rheological environment (Brown & Solar 1998a; Rabinowicz & Vigneresse 2004).

Hydrate-breakdown melting

Major crustal melting is controlled by successive breakdown of muscovite, biotite and hornblende in compositions that vary from 'granodioritic' or 'pelitic' to 'dioritic' or 'basaltic' forming lower crustal gneiss, amphibolite or granulite. As a result, the products and the proportions of melt and residue vary according to protolith composition and the P – T regime. In hydrate-breakdown melting, the solid products of the reaction may have difficulty nucleating, and once melting is established at a particular site, it is energetically favourable for diffusion-controlled melting to continue at that site and for the solid products to grow at that site (Waters 1988; Powell & Downes 1990; Brown 1994, 2004; Brown & Dallmeyer 1996; White *et al.* 2004).

Melting occurs over a range of temperatures, beginning with muscovite breakdown under upper amphibolite-facies conditions and extending through the granulite facies with biotite breakdown and hornblende breakdown in metasedimentary and metabasaltic protoliths, respectively. Water is important, but the lower continental crust may be essentially 'dry' (Yardley & Valley 1997; Wannamaker 2000), and thus melts will be H_2O -undersaturated (Clemens 2006) and the residue ultimately will be (nominally) anhydrous (Xia *et al.* 2006). Heat generally will be the principal rate control at the crustal scale, whereas deformation as a result of regional stress will be the enabling mechanism. Probably, >90 vol.% of melt generated (or of melt left after partial crystallization) is extractable, with <10 vol.% remaining on grain boundaries. Hydrate-breakdown melting has been investigated by experiments on natural and synthetic samples (e.g. summary by Clemens 2006) and by phase equilibria modelling (e.g. White & Powell 2002).

Information from Experiments

In a comprehensive review, Clemens (2006) summarized the results of 25 years of melting experiments on a range of common crustal rock types at pressures up to 1.5 GPa. In metapelites, generalized melting reactions observed in experiments are (where mineral abbreviations are from Kretz (1983), plus L for H_2O -undersaturated melt that varies from leucogranite to trondhjemite or granodiorite to tonalite in composition) $\text{Ms} \pm \text{Bt} + \text{Pl} + \text{Qtz} \rightarrow \text{Als} \pm \text{Grt} \pm \text{Kfs} + \text{L}$ (amphibolite facies) and $\text{Bt} + \text{Pl} + \text{Als} + \text{Qtz} \rightarrow \text{Grt} \pm \text{Crd} \pm \text{Kfs} + \text{L}$ (upper amphibolite to granulite

facies). In alumina-poor metagreywackes and quartz-feldspathic rocks, generalized melting reactions observed in experiments are $\text{Bt} + \text{Pl} + \text{Qtz} \rightarrow \text{Hbl} + \text{Cpx} \pm \text{Kfs} + \text{L}$ (upper amphibolite to granulite facies) and $\text{Bt} \pm \text{Hbl} + \text{Pl} + \text{Qtz} \rightarrow \text{Opx} + \text{Cpx} \pm \text{Grt} \pm \text{Kfs} + \text{L}$ (granulite facies). In peraluminous metagreywackes, generalized melting reactions observed in experiments are $\text{Bt} + \text{Pl} + \text{Qtz} \rightarrow \text{Opx} \pm \text{Grt} \pm \text{Crd} \pm \text{Kfs} + \text{L}$ (granulite facies; a similar reaction may occur in some sub-aluminous metapelites to produce Opx, but with Grt as a reactant not a product). In metabasaltic to meta-andesitic amphibolites, generalized melting reactions observed in experiments are $\text{Hbl} + \text{Pl} \rightarrow \text{Opx} + \text{Cpx} \pm \text{Grt} + \text{L}$, or $\text{Hbl} \rightarrow \text{Opx} + \text{Cpx} \pm \text{Pl} \pm \text{Grt} \pm \text{Qtz}$ or $\text{Ol} + \text{L}$ (granulite facies; note that the experiments are dominated by SiO_2 -undersaturated starting compositions). As a result of reactions of this type (for details see Clemens 2006), common protoliths may yield 10–50 vol.% of H_2O -undersaturated melt at temperatures attainable during orogenesis.

Phase equilibria modelling

An alternative to experimental investigation of melting is to use a phase equilibria approach (Holland & Powell 2001; White *et al.* 2001). This approach is in its infancy but, with continuing improvement and extension of thermodynamic models for appropriate phases, including a wider range of melt compositions, holds great potential for investigating melting in specific compositions.

Crustal rocks are complex multicomponent systems, in which the melting reactions are multivariant. The system components may be represented by the oxides SiO_2 , TiO_2 , Al_2O_3 , Fe_2O_3 , MnO , MgO , FeO , CaO , Na_2O , K_2O , P_2O_5 , and H_2O , and important trace elements such as the large ion lithophile elements, REE, high field strength elements, etc. Within this system, the following are the principal phases: in addition to Qtz, Kfs and Pl, in metapelites Ms, Bt, Grt, Crd, Opx, Als and Hc, and in metabasalts Px, Hbl, (Bt) and Ep; the accessory phases Mnz, Xen, Zrn, Ttn, Rt, Ilm, Mag and Ap; and fluid (melt (L) or H_2O -rich volatile phase (H_2O)).

The phase relations and chemography of melting for a range of bulk compositions may be investigated using calculated suprasolidus equilibria in a simplified model system, such as (Mn)NCKFMASH(TO), for specific compositions (e.g. White & Powell 2002; Johnson *et al.* 2003a; Johnson & Brown 2004; White *et al.* 2004, 2005). Phase topologies, and molar proportions and volumes are calculated using an internally consistent thermodynamic dataset (Holland & Powell 1998), a thermodynamic model for silicate melt (Holland & Powell (2001; extended by White *et al.* 2001), and the software THERMOCALC (Powell *et al.* 1998); equilibria are visualized using P - T , T - X and P - X pseudosections or pseudosections that use conjugate thermodynamic variables as appropriate (e.g. White & Powell 2002; Johnson *et al.* 2004; Powell *et al.* 2005).

Modelling the melting process and the consequences for melt compositions: a general example

A P - T pseudosection calculated for an average sub-aluminous metapelite composition at suprasolidus conditions is shown in Figure 4a; it shows phase relations for predicted stable assemblages in the NCKFMASH system. Expanding the system to include Mn, Ti and O will increase the number of phases present and change slightly the absolute stability of, for example, biotite (to higher T if Ti is present). Generalized significant melting reactions in this system are $\text{Ms} + \text{Qtz} (+ \text{Pl}) \rightarrow \text{Kfs} + \text{Sil/Ky} +$

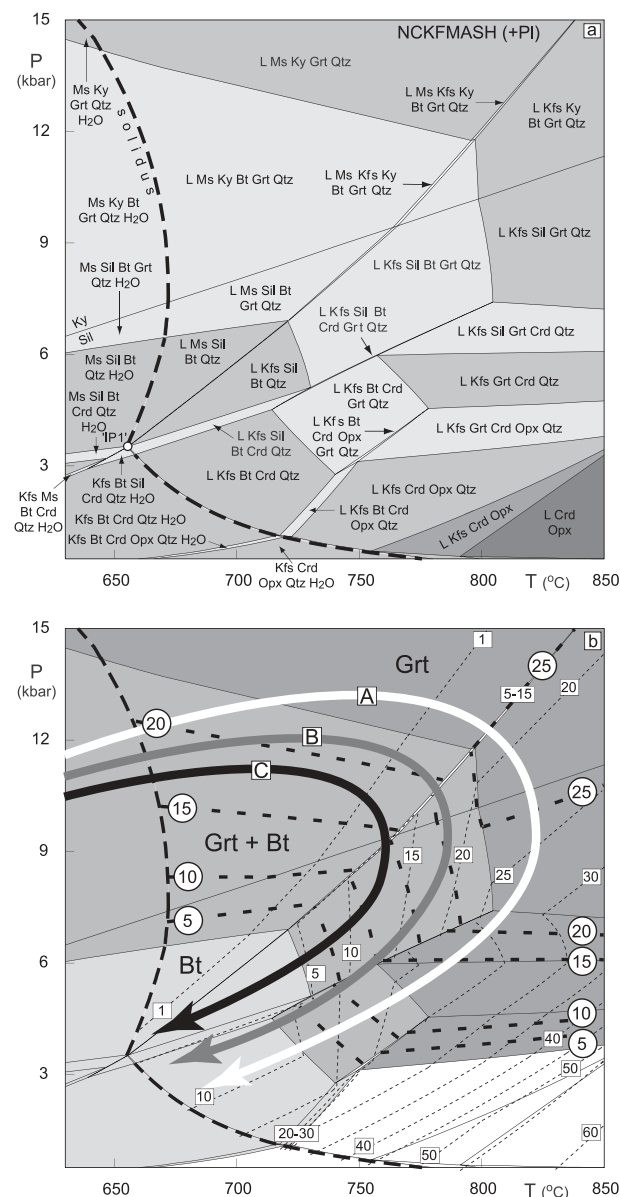


Fig. 4. (a) P – T pseudosection for the model system Na_2O – CaO – K_2O – FeO – MgO – Al_2O_3 – SiO_2 – H_2O for an average subaluminous metapelite composition (calculated using THERMOCALC v. 3.1 (information about THERMOCALC is available at <http://www.earthsci.unimelb.edu.au/tpg/thermocalc/>) with the Holland & Powell (1998) dataset, updated 1999). The darkest fields are sextariant ($F = 6$) and lighter shading indicates an increase in variance with the lowest variance assemblages (shown in bold) occurring as narrow divariant fields. Mineral abbreviations follow Kretz (1983). This diagram is appropriate only for consideration of the prograde phase equilibria, and if melt loss is cyclic it may be inappropriate even for phase relations close to the thermal peak (see text). **(b)** The same P – T pseudosection with the stability of the ferromagnesian phases (biotite (Bt), garnet (Grt), and biotite and garnet) emphasized by shading. Isopleths of mol% melt (squares) and garnet (circles) are shown as fine dashed and bold dashed lines, respectively (uncertainty on the position of the melt isopleths is of the order of $\pm 20^\circ\text{C}$). Three alternative clockwise P – T paths are shown; the petrological consequences of each of these evolutionary paths in P – T space are discussed in the text. Further details about the interpretation of suprasolidus pseudosections and melting relations, including the consequences of melt loss, have been given by White & Powell (2002) and Johnson & Brown (2004).

L, and Bt + Sil + Qtz (+ Pl) $\rightarrow$ Grt + Crd + Kfs + L or Bt + Grt + Qtz (+ Pl) $\rightarrow$ Crd + Opx + Kfs + L ($\pm$ Ilm, Rt, Mnz, Zrn, etc.). Figure 4b illustrates several aspects of melting and melt chemistry in relation to these phase relations.

Isopleths of melt mol% illustrate the principle that assemblages of lower variance produce larger increments of melt for a particular increase in T , with the highest rates of melt production occurring across narrow divariant fields corresponding to hydrate-breakdown melting reactions, such as the important muscovite-breakdown melting reaction (L Ms Kfs Ky Bt Grt Qtz (+ Pl)) in Figure 4a. These narrow low-variance fields correspond to univariant hydrate-breakdown melting reactions in smaller systems, such as (Na)KFMASH (e.g. Spear *et al.* 1999), and represent boundaries to the main indicator assemblages of upper amphibolite- and granulite-facies P – T conditions in metapelites. Melting may occur as a result of increasing temperature or decreasing pressure or both; it should be noted that the composition of all solid-solution phases, including melt, changes as P – T evolves.

The range of stability for ferromagnesian phases under suprasolidus conditions is clear. Biotite is stable at low to moderate pressures and high temperatures, although Ti (and F) will tend to stabilize it to ultrahigh temperatures in nature; because biotite may include accessory phases such as zircon and monazite, the stability of biotite may exert a strong control on trace element chemistry of the melt as well as the degree of inheritance in, for example, zircon. Garnet is stable at moderate to high pressures and temperatures, and the development of garnet as a peritectic product of biotite-breakdown melting exerts a strong control on the REE chemistry of the melt and phases that may replace it during P – T evolution (e.g. Clarke *et al.* 2007; note that in this example the P – T evolution is counter-clockwise). At moderate to low pressure at high temperature the ferromagnesian phases for the composition modelled are cordierite and orthopyroxene.

With reference to Figure 4b, we may assess what might be the consequences of a clockwise P – T evolution (e.g. in subduction-to-collision orogenesis) to illustrate the power and potential of the mineral equilibria approach. For clarity the pseudosection in Figure 4b has been shaded to show the suprasolidus stability of biotite before the appearance of garnet, the stability of biotite with garnet, and the stability of garnet beyond the stability of biotite. I do not discuss a counter-clockwise evolution as there are several well-documented examples in the literature (e.g. White & Powell 2002; White *et al.* 2004; Clarke *et al.* 2007), and it is an easy task to add into the published descriptions the trace element consequences in relation to the breakdown of major phases and the location of accessory phases along grain boundaries and as inclusions in major phases.

Consider path A: along the prograde segment of the P – T evolution, biotite is predicted to react out completely, creating the potential for dissolution of any sequestered zircon and/or monazite, with garnet as the peritectic product. This has consequences for the trace element and isotope chemistry of the melt, as garnet and monazite are the main hosts for the REE (e.g. Kelly *et al.* 2005; Zeng *et al.* 2005a–c). Significant melting is predicted to occur as a result of muscovite breakdown, producing up to 15 mol% melt, which has important consequences for Rb and Sr concentrations, and Sr isotope chemistry (e.g. Harris & Inger 1992; Zeng *et al.* 2005a–c). Melting continues during heating and decompression to peak T , at which point about 25 mol% melt is predicted together with slightly less than 25 mol% garnet.

During this evolution, at least two and possibly three or four

cycles of melt build-up and melt loss are predicted, based on an expectation that melt extraction occurs soon after the melt connectivity transition is reached at around 7 mol% melt (Rosenberg & Handy 2005; see also Handy *et al.* 2001). If melt has been lost, the bulk composition is likely to have been changed (even allowing for melt flux through the system; see Brown 2004, 2006), two, three or possibly four times during the evolution to peak T , in which case the pseudosection shown in Figure 4 will no longer be applicable to the final stages of the prograde or any part of the retrograde evolution (e.g. White & Powell 2002; Johnson & Brown 2004; White *et al.* 2004). Furthermore, because the composition of solid solutions changes with the P – T evolution and any change in the composition of the system, successive melt batches will have (slightly) different compositions in terms of major oxides. However, they may have marked differences in trace element and isotope chemistry.

The P – T evolution followed by path B does not exceed the stability of biotite, which may allow some zircon and monazite to remain sequestered in the residue, with implications for the trace element chemistry of the melt. The predicted maximum melt production is lower at about 18 mol% and less peritectic garnet is predicted at peak T (20 mol%). Significant melting is predicted as a result of muscovite breakdown, and a small increase in the amount of melt during (slight) heating and decompression to peak T . Along the retrograde segment, path B crosses the low-variance cordierite-producing biotite-breakdown melting reaction, which is predicted to produce an additional 5–6 mol% melt (e.g. Brown & Dallmeyer 1996; Johnson & Brown 2004). Along path C, even less biotite consumption is predicted, limiting the availability of sequestered zircon and monazite to dissolve in the melt, with implications for melt chemistry. The principal melt-producing reaction (muscovite breakdown) will be crossed during decompression, yielding about 15 mol% melt.

Evolution along a P – T path involves changes in the major phase assemblages and compositions as they register the changing P – T conditions. In petrology we take an inverse approach whereby the petrography, microstructural relationships among the solid phases and mineral chemistry of these phases are used to infer a P – T and melting or melt loss history for natural samples. Net melt loss changes the bulk composition and requires calculation of a new pseudosection for the modified composition or the use of different variables and an alternative pseudosection (such as a measure of melt loss with T at fixed P) to investigate further the P – T evolution of a residual granulite (e.g. Johnson & Brown 2004) or the consequences of melt loss in nature (e.g. White *et al.* 2004).

Role of accessory phases during melting

Many trace elements are essential structural constituents of accessory phases. However, we have only a limited knowledge of the distribution of the REE and other trace elements during melting, and it is unclear whether the breakdown of accessory phases necessarily leads to trace element depletion in the residue (Reid 1990; Fraser *et al.* 1997; Villaseca *et al.* 2003, 2007). This is an important issue given the concentration of trace elements in accessory phases in granites (e.g. Wark & Miller 1993) and the widespread use of Zr and light REE (LREE) contents to determine melt temperatures based on zircon and monazite solubility in melt (e.g. Miller *et al.* 2003). Rock composition (Rubatto *et al.* 2001), melt chemistry (Watson & Harrison 1983) and the stability of major phases (Bea *et al.* 2006) affect the Zr budget and whether zircon may dissolve or precipitate, but these

factors appear to have less influence on the dissolution and formation of monazite (Rubatto *et al.* 2001).

In hydrate-breakdown melting, initial melt pockets may be isolated along the hydrate grain boundaries. As a result, the melt may be able to reach equilibrium for trace elements concentrated in accessory phases located along these grain boundaries or sequestered in the hydrate but made available to the melt as it dissolves. In contrast, in wet melting, initial melt pockets form at quartz–feldspar grain junctions. As a result, the melt may not have the opportunity to equilibrate with trace elements housed in accessory phases associated with grain boundaries involving ferromagnesian phases or sequestered in hydrate phases.

Reactions that involve major phases (e.g. Pl, Ms, Bt, Hbl as reactants and Kfs, Grt, Crd, Opx as peritectic products) and accessory phases (e.g. Ap, Mnz, Xen and Zrn) potentially provide important controls on the trace element and isotope chemistry of anatectic melt, during both the prograde and the retrograde evolution. In one study of progressive metamorphism leading to melting, fine-grained discontinuously zoned monazite is present in the rock under subsolidus conditions, and prograde monazite growth and consumption are related to isograd-forming reactions (Williams 2001). Under suprasolidus conditions, monazite coarsens and commonly occurs as inclusions in peritectic phases; it displays complex zoning recording multiple growth and consumption events (involving peritectic phases and melt), and melt crystallization (Williams 2001).

At suprasolidus conditions, there is a significant enrichment in the major phases of the residue in trace elements released by breakdown of accessory phases. For example, an increase in Y and heavy REE (HREE) in garnet is related to dissolution of phosphates in melt (e.g. xenotime), with partitioning of highly compatible Y and HREE into garnet (e.g. Pyle & Spear 1999; Pyle *et al.* 2001), whereas the LREE–Eu budget is controlled by feldspar (e.g. Villaseca *et al.* 2007). Up to one-third of the Zr may be hosted in rutile and garnet; during decompression and suprasolidus garnet breakdown to cordierite, Zr is released, and because Zr cannot be incorporated into the cordierite structure, growth of new zircon occurs (e.g. Fraser *et al.* 1997; Degeling *et al.* 2001). This has implications for melt composition; low-Zr granites may reflect sources that follow a *P–T* evolution similar to path A in Figure 4b rather than low-*T* melts as might be inferred from their low Zr concentrations.

Assuming accessory phase(s) are available to the melt, the flux of trace element(s) into the melt during dissolution is a function of radial diffusion-controlled dissolution rate and surface area, which are correlated with grain size, and the degree of under-saturation of the melt with respect to the element(s) concerned (Watson 1996). Monazite and zircon populations commonly have different average grain size in any one sample and grain sizes that vary from sample to sample within a source terrane (Nemchin & Bodorkos 2000). These features may generate relative differences in LREE and Zr concentrations between small locally derived melt batches. Also, synanatectic deformation is important. For example, granular flow (diffusion-accommodated grain-boundary sliding) allows melt migration along grain boundaries, which allows better interaction between accessory mineral grains and the interstitial melt, and diffusion creep by dissolution–precipitation favours equilibration of grain surface compositions with the interstitial melt.

The initial Pb isotopic composition of individual batches of melt is sensitive to the age(s) and abundance of zircon in the source and the amount of radiogenic Pb that may be incorporated into the melt through dissolution of zircon (Hogan & Sinha 1991). Similarly, Sm/Nd ratios and Nd isotopic compositions of

individual batches of melt depend on the amount of apatite and monazite dissolved into the melt (Zeng *et al.* 2005a, b). In contrast, the Rb content of melt and the Rb/Sr ratio is controlled by the mica and feldspar, and fractionation during non-modal melting will yield liquid with a distinct isotope signature (Zeng *et al.* 2005a–c). Thus, the major phases and the grain size and modal proportions of accessory phases play important roles in determining the isotope composition of the melt.

Although mixing during accumulation and extraction may homogenize melt batches, heterogeneity in the initial isotopic composition of granites is inherited from the source and preserved as a result of incremental construction from multiple batches of melt (e.g. Deniel *et al.* 1987; Hogan & Sinha 1991; Pressley & Brown 1999; Glazner *et al.* 2004; Tomascak *et al.* 2005; Matzel *et al.* 2006). For incrementally constructed granites, fractional crystallization will be constrained to operate within the boundaries of individual batches of melt, and magmatic processes such as gravitational settling, sidewall accumulation and filter pressing produce only local geochemical variations within limited volumes of magma. In general, any *a priori* assumption that magmatic processes occur within a large-scale magma chamber to produce intrusion-wide variation as a result of fractional crystallization is not consistent with the geochemical data.

What evidence do we have from rocks about melt extraction?

Feedback relations between deformation and melting concentrate strain in anatectic domains and enhance contrasts in viscosity, facilitating melt extraction and partitioning deformation preferentially into zones in which melting has occurred (Brown & Solar 1998a; Rabinowicz & Vigneresse 2004). Melt flow and extraction occurs in actively deforming segments of orogens; as active deformation migrates through an orogen so does the locus of melt flow and extraction.

For flow, melt must form an interconnected network and the host must be permeable. For melt extraction, the vein or conduit topology infilled with leucosome or granite must show evidence of active flow and melt loss. Flow and extraction are inferred if bulk chemistry reveals residual compositions for anatectic rocks in comparison with protolith compositions, and the difference is consistent with net melt loss (e.g. Solar & Brown 2001a). However, net melt loss estimated by this method is a minimum, as residual migmatites and granulites preserve an integrated record that includes melt flux through the anatectic zone (Brown 2004; Olsen *et al.* 2004).

Microstructural criteria indicative of former melt

Microstructures in natural anatectic systems provide evidence of the former presence of melt in grain-scale pores (e.g. Brown *et al.* 1999; Sawyer 1999; Brown 2001a, b; Albertz *et al.* 2005; Fig. 3c). Melting may be inferred using the following criteria: mineral pseudomorphs after grain-boundary melt films and pockets (in contact migmatites (e.g. Rosenberg & Riller 2000; Marchildon & Brown 2001, 2002), in regional migmatites (e.g. Marchildon & Brown 2003) and in granulites (e.g. Sawyer 1999, 2001; Brown 2002; 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)) or magmatic microstructures in leucosomes (e.g. Vernon & Collins 1988; Brown 2002); spatial distribution of like and unlike phases (i.e. mineral phase distribution (Ashworth & McLellan 1985)); subparallel intragranular fractures (e.g. feld-

spar-filled fractures in quartz; Rosenberg & Riller 2000); and annealed microfractures (e.g. in quartz; Watt *et al.* 2000; Marchildon & Brown 2001, 2002).

Mesoscale structures

Under suprasolidus conditions, melt flow driven by gradients in pressure commonly appears to have been controlled by compositional heterogeneities and/or fabrics, and by the inhomogeneous patterns of strain that result from the compositional heterogeneities under applied differential stress (e.g. D'Lemos *et al.* 1992; Brown 1995; Collins & Sawyer 1996; Brown & Solar 1998a; Brown *et al.* 1999; Sawyer 2001; Guernina & Sawyer 2003; Marchildon & Brown 2003; White *et al.* 2004). Fabric-parallel and interboudin stromata form by mass transfer down gradients in pressure (van der Molen 1985; Maaløe 1992; Brown *et al.* 1995; Sawyer 2001; Guernina & Sawyer 2003), and melt becomes progressively concentrated into a network of deforma-

tion bands. The networks form by linking among shear bands, compaction bands and dilation bands. Melt may be expressed from compaction (generally fabric-parallel) bands into shear (oblique to fabric) and dilation (generally fabric-normal) bands (Brown & Rushmer 1997; Brown 2004; Fig. 5).

Incongruent melting involving biotite or hornblende breakdown results in melt concentration around the peritectic product mineral (Jones & Brown 1990; Powell & Downes 1990; Brown 2004; White *et al.* 2004; Fig. 6a and b); with increasing melt fraction progressive weakening may lead to formation of a network of deformation bands nucleated on the local concentrations of melt (see Grujic & Mancktelow 1998). The preferential occurrence of peritectic minerals in deformation bands, in spite of a homogeneous distribution of reactant phases in the matrix, supports the inference that deformation was localized by the concentration of melt at these sites (Brown 2004).

Ultimately, melt drains from the source via discrete planar or cylindrical conduits to form upper crustal intrusions (e.g.

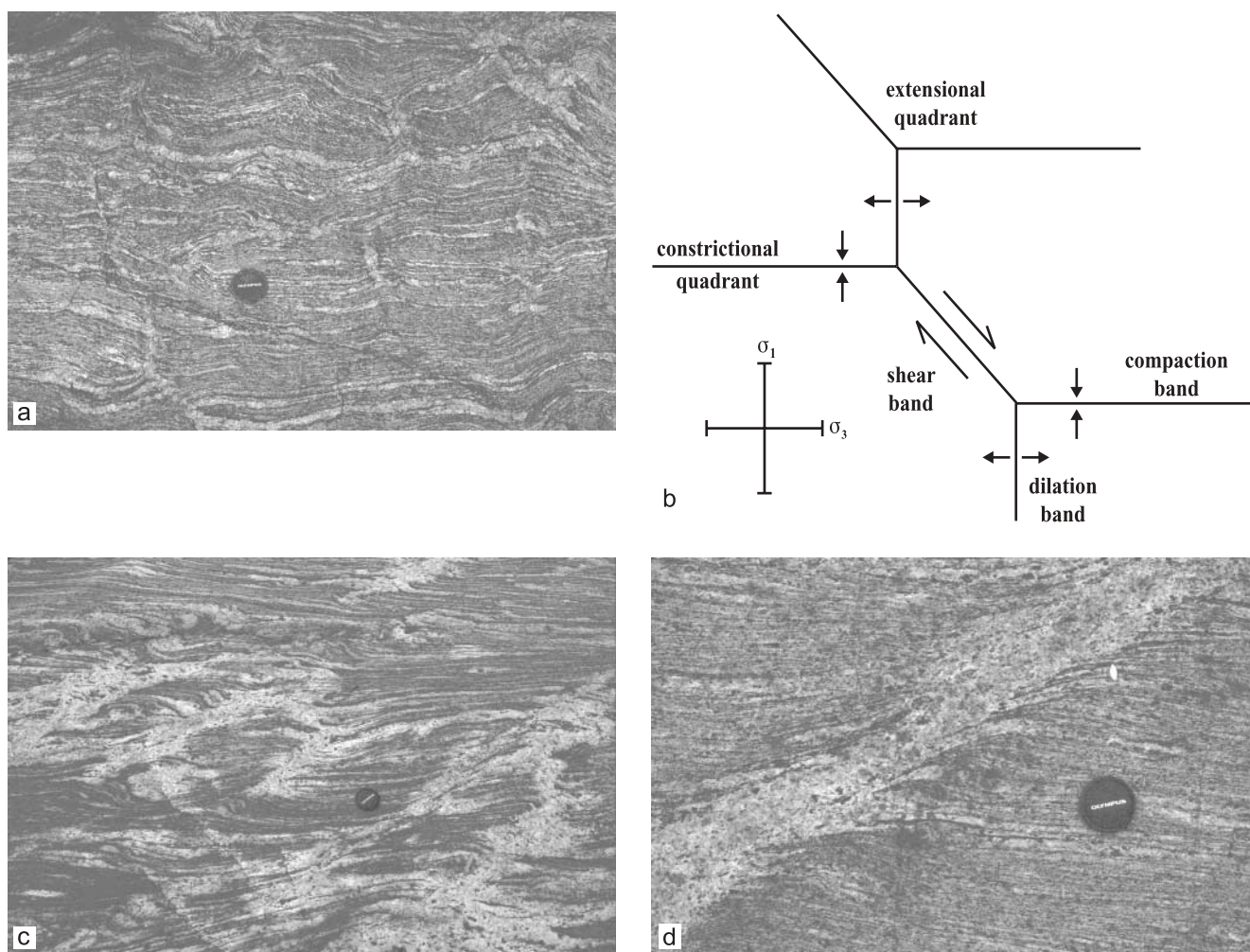


Fig. 5. Features associated with melting and inferred melt-bearing structures in granulite-facies anatectic migmatites. (a) Stromatic migmatite with layer-parallel leucosome stromata (equivalent to compaction bands) and leucosome in dilation and shear bands (Petit Mont, Morbihan, France). Diameter of lens cap is 55 mm. (b) Idealized diagram (modified after Du Bernard *et al.* 2002) to show the coexistence of shear, compaction and dilation bands (σ_1 and σ_3 are greatest and least far-field principal stresses, respectively). (c) Example of shear band network in which leucosome is concentrated in dilatant shear bands and is in petrographic continuity (microstructure, mineralogy and mode) with foliation-parallel leucosome stromata, recording a former melt-rich reservoir (Tolstik Peninsular, Karelia, Russia; diameter of lens cap is 55 mm). (d) Example of leucosome concentration in a dilatant shear surface, consistent with melt flow down gradients in pressure, where the leucosome in the shear surface is in petrographic continuity (microstructure, mineralogy and mode) with the foliation-parallel leucosome (Tolstik Peninsular, Karelia, Russia; diameter of lens cap is 55 mm).

D'Lemos *et al.* 1992; Brown 1994, 2004, 2005; Collins & Sawyer 1996; Brown & Solar 1999; Solar & Brown 2001a; Marchildon & Brown 2003; Fig. 7). Commonly these draining conduits are marked by cumulate granite, as seen in the Bruno's Gorge outcrops of the Wuluma Hills (Fig. 6c), in the Arunta Inlier of central Australia, where mostly pristine peritectic garnet preserved in K-feldspar-rich granite suggests loss of a fractionated liquid (now inferred to be represented by the Wuluma

granite; see Clarke *et al.* 2007). Multiple studies have provided general descriptions of the structural control over pervasive melt flow and the relationship between mesoscale structures and ascent conduits (e.g. D'Lemos *et al.* 1992; Brown 1994, 2004, 2006; Collins & Sawyer 1996; Brown & Solar 1998a, b, 1999; Solar & Brown 2001a; Marchildon & Brown 2003; Slagstad *et al.* 2005).

In the Acadian belt in Maine, northern Appalachians, the form of melt ascent conduits mimics the apparent strain ellipsoid recorded by the host rock fabrics (Brown & Solar 1999; Solar & Brown 2001a). Concordant tabular granites occur in zones of apparent flattening strain ($S > L$; Fig. 7a and b), whereas

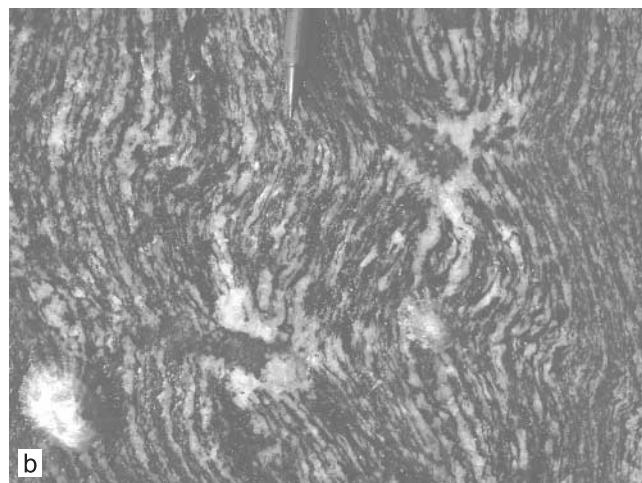
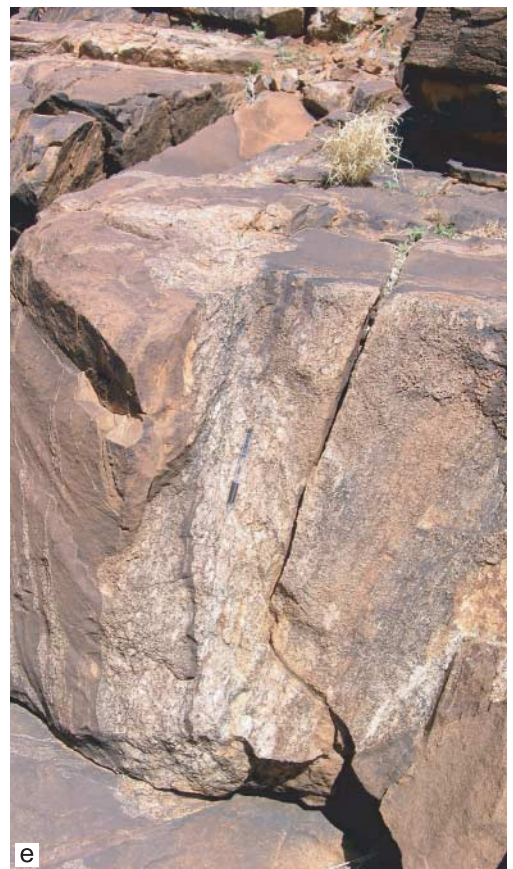


Fig. 6. Leucosome features associated with biotite-breakdown melting and inferred melt-bearing structures in granulite-facies anatectic migmatites. (a) Peritectic garnet poikiloblasts in medium-grained K-feldspar-quartz-rich leucosome contained within strongly foliated cordierite-sillimanite-biotite-rich host, Round Hill, Broken Hill, southeastern Australia (see White *et al.* 2004, for detailed discussion). (b) Stromatic migmatite with layer-parallel leucosome stromata showing periodic foliation boudinage with peritectic cordierite in minimal transverse leucosome that is petrographically continuous (microstructure, mineralogy and mode) infilling dilation bands (Port Navalo Plage, Morbihan, France; tip of pen for scale). (c) Cumulate quartz-K-feldspar-peritectic garnet sheets record melt transfer through stromatic migmatite, but limited retrogression of garnet demonstrates melt loss from this terrain, which is recorded as shallower levels by the Wuluma granite. The margin of the sheet (on the right-hand side) is diffuse because of petrographic continuity (microstructure, mineralogy and mode) between leucosome in the host and granite in the sheet. This example is from Bruno's Gorge in the Wuluma Hills of the Strangways Metamorphic Complex, eastern Arunta Inlier, central Australia.

Fig. 7. (a) View to the SE across the subhorizontal top surface of an outcrop of steeply east-dipping foliation-concordant dykes of granite in residual stromatic migmatite within a regional-scale zone of apparent flattening strain in the Acadian orogen (Swift River, Roxbury, west-central Maine, USA). The more irregular 'pinch-and-swell' structure of the sheets to the west side of the outcrop (relatively older?) in comparison with those to the east (relatively younger?) should be noted. (b) View to the NNE of a steep surface approximately perpendicular to the surface in (a) and to both foliation and the regional mineral elongation lineation. It should be noted that the 'pinch' of the dykes is more extreme in the subvertical section, with the axes of the 'pinch' subparallel to the regional mineral elongation lineation. Also, the centimetre-scale composite layering in the thicker sheet on the right is typical and records multiple use of this fracture channel by successive batches of magma. (c) Concordant, broadly cylindrical-shaped body of granite in diatexite from a regional-scale zone of apparent constrictional strain in the Acadian orogen (Swift River, south of Roxbury, west-central Maine, USA). At this locality, leucosome and the granite plunge parallel to the regional mineral elongation lineation. Further details concerning the outcrops shown in (a)–(c) have been given by Brown & Solar (1998a, 1999) and Solar & Brown (2001a). (d) Foliation-parallel stromatic leucosomes composed of peritectic garnet poikiloblasts in a medium-grained K-feldspar-quartz-rich matrix contained within strongly foliated cordierite-sillimanite-biotite-rich host record melt loss pathways through the residual migmatites, Round Hill, Broken Hill, southeastern Australia (see White *et al.* 2004, for detailed discussion). (e) Granite in cylindrical body that is inferred to record magma migration through Palaeoproterozoic lower crust via structurally controlled pathways parallel to the moderately to steeply plunging regional elongation direction, Mt. Hay, eastern Arunta Inlier, central Australia (for further details, see Collins & Sawyer 1996).



concordant rod-like granites occur in zones of apparent constrictional strain ($L > S$; Fig. 7c). In Mesoproterozoic aluminous metapelites from Broken Hill, Australia, White *et al.* (2004) described spatially focused melt formation where the resulting pathways for melt escape, as recorded by leucosome, are parallel to foliation defined by highly depleted melanosomes and intervening mesosomes (Fig. 7d; see also White *et al.* 2004, figs 3–5). In contrast, in the Mt. Hay area of the eastern Arunta Inlier, central Australia, magma is inferred to have migrated through Palaeoproterozoic lower crust via a network of narrow, structurally controlled pathways parallel to the moderately to steeply plunging regional elongation direction, defined by leucosome associated with coaxial folds and a strong mineral-elongation lineation (Collins & Sawyer 1996; Fig. 7e).

The rheology of melt-bearing crust and melt segregation mechanics

Understanding the rheology of melt-bearing crust is important because of the role it plays in the tectonic evolution of orogenic belts. However, the rheology of materials composed of two distinct but interacting phases, a stiffer solid and a much weaker liquid, is a complex problem, and evaluating the rheology of anatectic crust is particularly difficult because the dramatic change in strength that occurs with melting depends on both melt volume and grain-scale distribution.

The principal forces acting upon melt and matrix in melt-bearing crust are (Rutter 1997; Rutter & Mecklenburgh 2006): gravitationally induced forces, caused primarily by the density difference between melt and the solid residue; tectonically induced forces that impart momentum to the melt as a result of non-hydrostatic stress gradients; the forces owing to $\Delta V_{\text{reaction}}$ (Rushmer 2001); and the forces owing to surface tension. Actively deforming melt-bearing crust is a highly dynamic nonlinear system, with history- and time-dependent behaviour, in which deformation leads to melt redistribution and, potentially, changes in deformation mechanism; for example, from melt accommodated granular flow, wherein solid crystals slide over one another, to intracrystalline plasticity (Vigneresse 2004; Vigneresse & Burg 2004; Rutter & Mecklenburgh 2006; Rutter *et al.* 2006). Weakening may lead to localization but melt extraction is possible only if the solid residue may compact.

The grain-scale distribution of melt is of primary importance in determining the physical properties of melt-bearing continental crust. The permeability, the viscosity, which is sensitive to the presence and distribution of melt, and the effective diffusivity, which may be increased by orders of magnitude by the presence of interconnected melt, are strongly affected. Melt is likely to be interconnected to some degree in all common crustal protoliths (Laporte *et al.* 1997; Lupulescu & Watson 1999), but it is unlikely to be uniformly distributed, because of variable mineralogy or grain size (Watson 1999; Laporte & Provost 2000); melt tends to concentrate around minerals that have low crystal–fluid interfacial energies or small grains. Horizons dominated by minerals that wet or those with a smaller grain size may weaken as a result of increased melt abundance leading to complex mechanical behaviour.

In deformation experiments on analogue materials, dynamic wetting of grain boundaries enhances early interconnection of melt and promotes intergranular fracture and grain-boundary sliding (Rosenberg & Handy 2000, 2001). In deformation experiments on rocks, increasing melt volume induces an increase in the proportion of melt-bearing grain boundaries, which results in marked bulk weakening (Rutter 1997; Rosenberg & Handy 2005).

Rosenberg & Handy (2005) suggested there is a transition in behaviour from viscous flow en masse, induced by weakening as a result of interconnected grain-boundary melt, to localization of deformation into melt-filled deformation bands at around 10 vol.% melt. Increasing the volume of melt in the network of deformation bands further decreases the bulk-rock strength, but at a less dramatic rate than initial melting, and reduces permeability. It is likely that melt will be lost from the system at a melt volume around 7–10 vol.%, so that a simple increase in melt volume may not occur unless there are extrinsic constraints that prevent melt loss from a particular source horizon.

The mechanical behaviour of a melt-bearing source may be understood using an approach based on critical state soil mechanics for a porous solid saturated with fluid (Rutter 1997; Rutter & Mecklenburgh 2006; Rutter *et al.* 2006). The behaviour of porous rock changes from localized dilatant shear failure to distributed shear-enhanced compaction with increasing porosity (melt volume in the anatectic environment), grain size and effective pressure. Shear-enhanced compaction results in permeability reduction, whereas dilatant shear failure results in permeability enhancement if the melt volume is low. In melt-bearing lower crust, focused melt flow is likely as a result of dilatant shear failure of low melt fraction rocks, consistent with the leucosome-filled deformation band networks that record melt flow through a residual source.

Rabinowicz & Vigneresse (2004) presented a model of melt segregation in an isotropic mushy crust subject to both compaction and shear, which may be important in the development of leucosomes in homogeneous isotropic protoliths such as granite or orthogranulite. Compaction leads to the development of melt pockets whereas shear channelling, a faster melt segregation process, drives melt into veins. The size of the pockets and the distance between the veins approximate the compaction length, which is controlled by the viscosity ratio between the matrix and the melt. During an extraction cycle, melt is driven from the volume between veins to the veins, and successive batches of extracted melt, as well as the residual solid matrix, become increasingly more dehydrated. As a result, both melt and matrix progressively stiffen; although neither viscosity contrast nor compaction length change, the rate of melt segregation is reduced.

Melting and melt retention or loss has significant implications for crustal-scale rheology and tectonics. The deformation of low melt fraction source rocks probably is dominated by granular flow. The limited importance of intracrystalline plasticity, deduced from the appearance of naturally deformed partially molten rocks and from modelling based on extrapolation of laboratory-determined constitutive flow laws, indicates that melt-bearing crust must be extremely weak, able to flow at geological strain rates at differential stresses of the order of 1 MPa or less, according to the temperature (Brown 2001a, b; Albertz *et al.* 2005; Rutter & Mecklenburgh 2006; Rutter *et al.* 2006). Vertical intervals in the lower crust that are partially molten must therefore be extremely effective tectonic detachment horizons. However, melt loss leaves behind a stronger residue, leading to long-term strengthening the lower crust (Brown & Solar 2000; Brown 2004, 2006). In a complementary way, granites emplaced and crystallized around the brittle–ductile transition zone contribute significantly to the overall strength of the upper crust.

Proposed mechanisms of magma ascent

In the deep crust, there is abundant evidence, recorded by leucosome, of melt flow networks in residual migmatites and granulites, from which I infer that melt segregates and accumu-

lates in these networks until a drainage event occurs. Within the storage networks, mesoscale pods of leucosome or diatexite may occur that record local ponding of melt (e.g. southern Brittany, France, Jones & Brown 1990; Antarctica, Allibone & Norris 1992; Western Australia, Oliver & Barr 1997; Opatika Sub-province, Canada, Sawyer 1998; Arunta Inlier, central Australia, Sawyer *et al.* 1999; Cooma Complex, Australia, Vernon *et al.* 2003). It is these storage networks, including ponded diatexites, that feed drainage channels, recorded by granite, to transport melt from the source to shallower levels in the crust (Brown 1994, 2004, 2006; Sawyer 1998). In contrast, evidence of melt aggregating in large-scale magma chambers to feed dykes appears to be generally absent.

The mechanism by which melt is drained from storage networks is poorly understood. The main melt–magma extraction–ascent mechanisms suggested are: flow in dykes; stepwise accumulation of melt and ascent in mobile hydrofractures; pervasive migration by mesoscale porous melt flow, allowed by syn-ascent advection of heat; ascent by compaction-generated flow instabilities (porosity waves and related structures); and en-masse flow as diapirs.

Dykes

Dykes are melt-filled, opening mode fractures that propagate normal to the direction of minimum compressive stress ahead of the dyke tip. The slow rate of migration of silicic melt through pores in the source demands that melt drained in a dyke propagation event must have been previously extracted and reside in a storage network, and there must be sufficient hydraulic head to drive ascent. However, it is unclear whether dyking by brittle fracture is a viable mechanism for extracting melt from melt-bearing sources (e.g. Sleep 1988; Rubin 1995, 1998), although significant progress has been made in understanding dyke initiation from a magma chamber (e.g. Spence & Turcotte 1985; Chen & Jin 2006).

Dyking by ductile fracture avoids problems inherent in the brittle fracture model and is supported by observations from residual migmatites and granulites (e.g. Brown 2004, 2005, 2006), such as blunt fracture tips, zigzag geometry close to the fracture tips and petrographic continuity between leucosome in host and granite in dykes (Fig. 8). In this model, viscous flow of melt is inferred to be through self-generated melt-induced deformation band networks (a reservoir for melt storage) to ductile opening-mode fractures that propagate from shear bands as crack-like opening-mode ductile fractures (Eichhubl *et al.* 2001; Simakin & Talbot 2001; Du Bernard *et al.* 2002; Eichhubl & Aydin 2003; Brown 2004, 2005). The strong intrinsic anisotropy of permeability of melt-bearing crust allows strong melt focusing and flow down gradients in pressure to ascent conduits. However, to extract melt from the source requires sustained transient near-lithostatic to supra-lithostatic melt pore pressure in the melt flow network to prevent drawdown, decreased transport rates and possible collapse.

Hydrofractures

Hydrofractures form in response to internal fluid pressure rather than to externally applied stress (Bons 2001). The model assumes an isotropic protolith, pervasive interconnected melt-filled porosity, initial fracturing as a result of melt-enhanced embrittlement, and gradients in fluid pressure sufficient to drive melt flow from interconnected pores to the fractures. Steep fluid-filled fractures may become unstable if they exceed a critical

crack length (Secor & Pollard 1975; Spence *et al.* 1987; Takada 1990); for melt-enhanced embrittlement, it is assumed the critical length is not too long to allow spontaneous initiation, but this remains to be demonstrated. The instability arises from gradients in effective stress acting on the fracture, which are generated by the contrast in density between melt and host and the vertical extent of the fracture (Secor & Pollard 1975).

Once instability is triggered, a hydrofracture becomes mobile by propagation at one end and sealing at the other, so that it moves together with its content (Dahm 2000; Bons 2001; Bons *et al.* 2001). After instability causes fractures to propagate, transport of melt occurs via stepwise and discontinuous aggregation of mobile hydrofractures to increase the accumulated melt volume and promote ascent. Migration of melt changes the system by opening new fractures and closing existing pathways as melt drains from some parts and accumulates elsewhere; transport and accumulation are linked phenomena (Bons *et al.* 2004). The effect is a highly dynamic self-organized system in which intermittent, local events of fracture propagation cause avalanches of instabilities and merging of hydrofractures; however, the zipper effect of migrating mobile hydrofractures may not leave much evidence of melt passage through the crust.

Mesoscale porous melt flow

Exhumed anatectic crust exhibits a range of grain- to outcrop-scale structures that support an inference of mesoscale porous flow of melt (e.g. Brown 1994; Collins & Sawyer 1996; Brown & Solar 1998a; Sawyer *et al.* 1999; Weinberg 1999). Weinberg (1999) has argued that pervasive migration gives rise to magma ‘sheets’ preferentially emplaced parallel to high-permeability zones such as foliation or lithological contacts, similar to the strain-controlled tabular and rod-shaped granites in zones of apparent flattening and apparent constriction strain, respectively, in melt-depleted anatectic migmatites from the Acadian orogen of Maine, USA (Brown & Solar 1998a, 1999; Solar & Brown 2001a, b).

Although pervasive migration of melt must be restricted to suprasolidus crust to avoid freezing, advection of heat with migrating melt may expand the suprasolidus domain, allowing pervasive migration of melt to shallower depths (Brown & Solar 1999; Weinberg 1999). The feedback relation between pervasive migration of melt and heating of the country rock, which allows younger batches of melt to reach increasingly shallower levels, has been modelled numerically by Leitch & Weinberg (2002; see also Jackson *et al.* 2003). They suggested that pervasive migration of melt is an efficient way of heating the upper part of the lower crust, in effect raising the level of the viscous anatectic flow to viscous subsolidus flow transition zone by advection; those workers concluded that the process may result in an injection complex several kilometres thick, consisting of about half magma and half original crust. The Qôrqu granite complex of southern West Greenland (Brown *et al.* 1981), the Wuluma Granite of the Arunta Inlier, central Australia (Lafrance *et al.* 1995) and the Pangong injection complex of the Indian Karakoram (Weinberg & Searle 1998) represent examples of such injection complexes that are of the order of several kilometres thick.

Compaction-generated flow instabilities

The anatectic zone, that portion of continental crust above the solidus, is an over-pressured system with supra-magmatic melt pressure gradients and spatial and temporal variations in permeability. Permeability has a power law dependence on melt

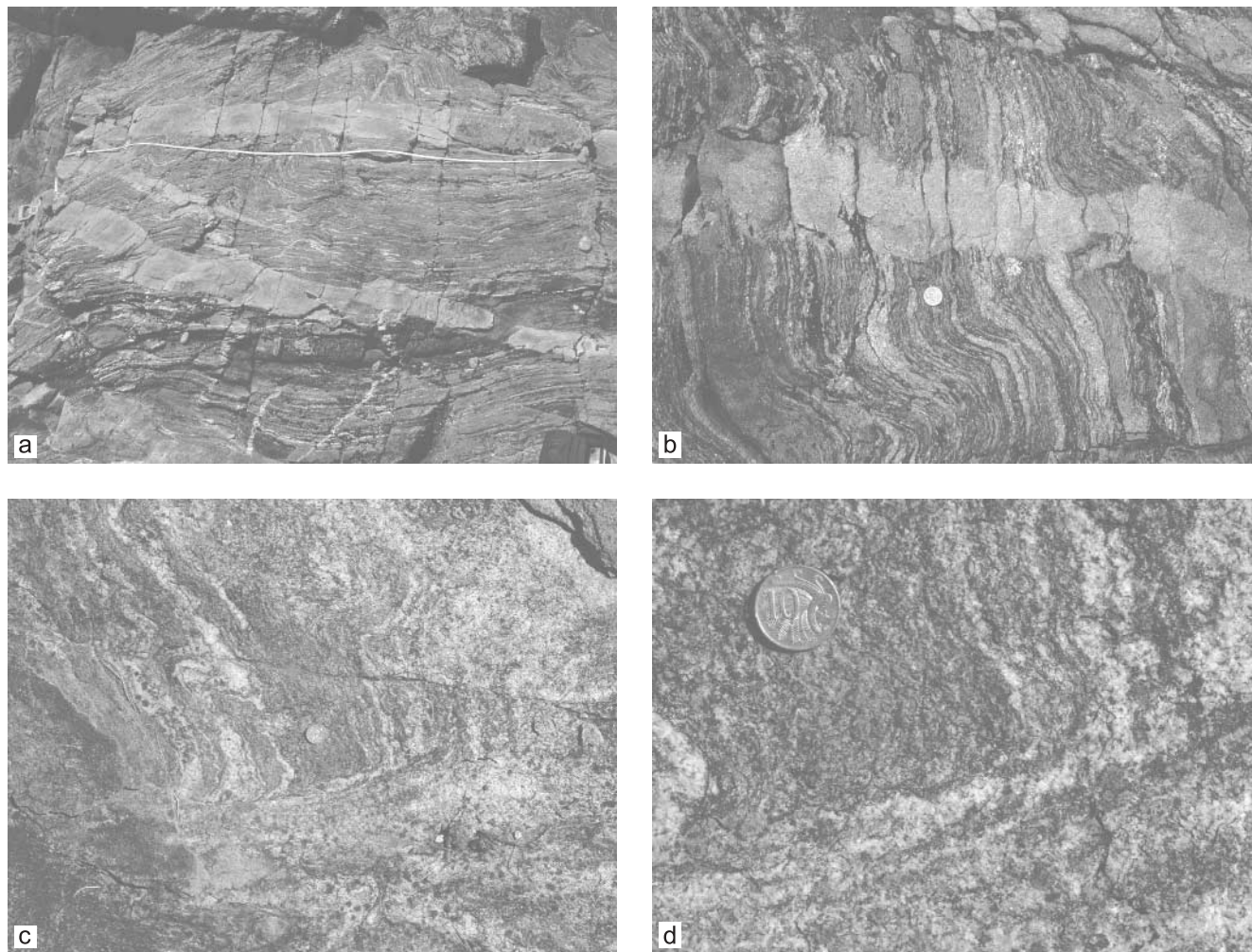


Fig. 8. Features associated with melting and inferred melt-bearing structures in granulite-facies anatectic migmatites. (a) View from cliff-top of subhorizontal wave-cut platform to show several dykes (tape measure for scale; Le Petit Mont, Arzon, Morbihan, France). The thinner dyke crossing diagonally between the two thicker dykes, and particularly the blunt nature of the dyke tip in the centre of the image, should be noted. (b) Leucosome in stromatic migmatite is in petrographic continuity (microstructure, mineralogy and mode) with granite in the highly discordant, apparently cross-cutting dyke (Le Petit Mont, Arzon, Morbihan, France). Further details about these outcrops have been given by Brown (2004, 2006). (c) Discrete leucosomes concordant with the host foliation are deformed by small folds and bounded at the bottom by a dyke of diatexite in petrographic continuity with these early leucosomes (Diatexite Hill, Broken Hill, southeastern Australia). The transition between the stromatic migmatite and the diatexite varies from sharp (on the left) to gradational (on the right). (d) Detail of the petrographic continuity (microstructure, mineralogy and mode) along the gradational contact shown in (c). Further details about this outcrop have been given by White *et al.* (2005).

fraction (Wark & Watson 1998), which may lead to instabilities in the form of porosity waves (Connolly & Podladchikov 1998; Rabinowicz *et al.* 2001), which represent an intermediate stage between pervasive and channelled flow. Flow is enhanced where porosity is high, leading to an influx of melt into these zones, which enhances flow and induces additional influx of melt. Coupling between flow and porosity leads to accumulation and transport in the form of upward-migrating high-porosity zones (Podladchikov & Connolly 2001); these may develop into channel flow in the core (Richardson *et al.* 1996) to transport melt through the anatectic zone under suprasolidus conditions.

At the peak of the thermal evolution, the solidus for granite melt (approximately an isothermal surface) will be elevated in the core of an orogenic belt (i.e. at its shallowest depth); it will be marked by the anatectic front, which is the upper depth limit of viscous anatectic flow (Brown & Solar 1999). The mechanism of melt transfer into subsolidus crust may involve melt-enhanced

embrittlement, as the solidus represents the boundary to the anatectic zone with interconnected melt-filled porosity. Compaction within this zone is likely to generate strong variations in pressure near the solidus, which may allow deformation at the anatectic front to be essentially elastic in character. Under this condition, melt may be transported into the subsolidus (zero-porosity) crust by an elastic shock. The physical expression of this phenomenon may be dykes, but amplification of flow instabilities across the solidus, which represents an unstable interface, may be an alternative mechanism of melt ascent in some circumstances (Connolly & Podladchikov 1998).

Diapirism

Diapirism may occur in the lower crust and in the formation of gneiss domes (Teyssier & Whitney 2002; Whitney *et al.* 2004), and it may be a mechanism by which (some) granites are

emplaced in the deep crust in continental arcs (e.g. Miller & Paterson 1999). However, whether diapirism is a viable ascent process for upper crustal intrusions remains a matter about which opinions differ (Emerman & Marrett 1990; Clemens & Mawer 1992; Rubin 1993; Weinberg & Podladchikov 1994; Burov *et al.* 2003). A structural analogy is sometimes argued between granites and salt diapirs, but it is clear that regional extension is the trigger for salt diapirism (e.g. Jackson & Vendeville 1994) and the same has been argued for formation of gneiss domes (e.g. Teyssier & Whitney 2002; Brown 2005). Diapirism involves melt plus residue, implies partial convective overturn and is limited by the brittle–viscous transition zone; it is argued to be an important process in the Archaean (e.g. Chardon *et al.* 1998; Collins *et al.* 1998; Peschler *et al.* 2004; Sandiford *et al.* 2004). That diapirism is a viable process in the right circumstances is clearly demonstrated by the formation of country rock diapirs that have ascended into the upper levels of the layered mafic–ultramafic rocks of the Bushveld complex (e.g. Uken & Watkeys 1997; Johnson *et al.* 2004).

What quantitative data do we have about melt flow networks and ascent conduits?

There have been several attempts 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.

An example from the Variscan orogen

I summarize the work of Marchildon & Brown (2003; see also Brown 2004). This study utilized well-exposed coastal outcrops of anatectic rocks south of the South Armorican Shear Zone, in the Variscan orogen of France. Using quantitative data on the 1D and 2D (and 3D) distribution of inferred melt-bearing structures, Marchildon & Brown investigated the transitions from fabric-parallel to network to conduit flow. Structures include millimetre- to centimetre-scale foliation-parallel and foliation-discordant leucosome stromata (reflecting compositional layering, interboudin partitions and extensional shear surfaces) and centimetre- to metre-scale foliation-discordant dykes of granite (inferred ascent conduits). Petrographic continuity (mineralogy, mode and microstructure) from structure to structure implies that the infilling material crystallized from a continuous melt-bearing network. Peritectic phases commonly are located in deformation bands, particularly interboudin partitions, which connote an important contribution by diffusive mass transfer in the local concentration of melt. Fugitive melt from the anatectic rocks accumulated as leucogranites at shallower levels of the crust (Gapais *et al.* 1993; Brown & Dallmeyer 1996; Marchildon & Brown 2003; Brown 2005).

Information about thickness and spacing distributions of fabric-parallel leucosome stromata is obtained from 1D traverses. Most thicknesses fall in the range 1–10 mm (upper limit 20–30 mm). The number of thicker stromata decreases abruptly with increasing thickness, inconsistent with scale-invariance; this points to short-range melt movement along grain boundaries to sites where melt-rich stromata develop. Spacing distributions also are not scale-invariant; however, the volume of leucosome (40–60%) suggests that spacing distributions may be controlled in part by impingement.

Observation of structures in mutually perpendicular 2D surfaces reveals anisotropy of the leucosome network related to a

subhorizontal lineation associated with transcurrent movement along the South Armorican Shear Zone. Analysis of leucosome distributions using the box-counting method corroborates the observed anisotropy, but indicates that leucosome morphology is not scale-invariant; melt flow is inferred to have been primarily subhorizontal. This network fed steeply dipping discordant dykes, some of which appear to nucleate from shear bands initially with zigzag form. The thickness of dykes >10 cm wide shows a power law distribution with an exponent of 1.11, comparable with values of 1.1 reported by Bons *et al.* (2004; who also reported an exponent of 1.9 for one suite of veins). This width–distribution relation suggests that dykes may be scale-invariant, although the dataset is small (87 dykes) and the range of observations is only two orders of magnitude (see Bons *et al.* 2004). The largest dykes along the traverse were 3 m and 5.5 m wide, within the range of critical dyke widths for flowing magma to advect heat faster than conduction through the walls and avoid freezing close to the source (Clemens 1998). Dykes of this size each may transport >1000 km³ of magma in *c.* 1200 years, assuming a horizontal length of 1 km (Clemens 1998).

What do we infer from these results?

Petrographic continuity indicates that conduits represent either structures analogous to large-scale dilation bands formed by opening-mode failure along zones of localized porosity increase or ductile opening-mode fractures formed by pore growth and coalescence (Eichhubl *et al.* 2001; Simakin & Talbot 2001; Du Bernard *et al.* 2002). Zigzag propagation paths at fracture tips point to ductile fracture as the mode of formation of the conduits (Eichhubl & Aydin 2003; Brown 2004). Draining of melt from deformation bands and capture of small conduits by large conduits will be favoured by lower melt pressure in the larger conduits (see Sleep 1988). The formation and orientation of conduits apparently was controlled by far-field stresses associated with dextral displacement along the South Armorican Shear Zone; spacing of conduits was controlled by the strong intrinsic permeability anisotropy in some layers and the weak extrinsic permeability anisotropy between layers, and consequent variable fluid focusing that results in a large collection zone for each conduit (Eichhubl & Boles 2000).

Magma emplacement

Limiting factors on the size of intrusions include the thickness of the source, the amount of melt that may be withdrawn from the source and the spacing between conduits that channel magma vertically to supply the intrusion. Tectonic setting is important in determining the spatial distribution of intrusions. In continental magmatic arcs, where a high degree of melting is focused in a narrow belt, intrusions are closely spaced and probably tabular in form. Conversely, in subduction-to-collision orogens, where low degrees of partial melting occur throughout aerially extensive sources, intrusions are widely spaced and commonly wedge-shaped. Also, in compression or extension variable shortening across the strike of an orogen will tend to drive melt along the strike of the belt, which may, in part, explain the elongation of synorogenic intrusions along strike.

To understand emplacement mechanisms we must realize the coupled nature of orogenic systems, with feedback relations between stress, deformation, thermal evolution, strength, strain rate, melt flow, etc. (Brown & Solar 1998a, b). Crustal-scale magma systems are probably strongly self-organized from the

bottom up (Cruden 2006), and a feedback relation between cyclic melt extraction and intrusion inflation is reflected in the sheeted nature of many granites (e.g. Brown *et al.* 1981; Brown & McClelland 2000; de Saint-Blanquat *et al.* 2007).

The approach adopted reconciles into one general model contradictory views put forth by proponents of the principal models for granite emplacement: dyke to horizontal tabular intrusion; viscous flow in structurally controlled channels to composite intrusion with root and horizontal tabular head; and ascent and emplacement as a viscoelastic magmatic diapir. This model also accommodates ascent and emplacement related to faults and shear zones (e.g. Strong & Hanmer 1981; Brown & Solar 1998a, b; Mahan *et al.* 2003; Rosenberg 2004; Weinberg *et al.* 2004).

The common perception that emplacement occurs at a 'level of neutral buoyancy' is inconsistent with the density of granite liquid (*c.* 2.44 kg m^{-3} at 1 GPa, 800 °C), negative gravity anomalies over many granites and the range of emplacement depths. Also, the generalization that emplacement is controlled by structural interactions between ascending magma and anisotropies in the upper crust is only part of the story. Although intrusions may be formed by magma expansion into an evolving structural trap (e.g. Hutton 1988; Grocott *et al.* 1994) or by multiple material transfer processes acting locally (Paterson & Fowler 1993; Paterson & Vernon 1995), the systematic variation in 3D shape with depth, from horizontal tabular (laccoliths to lopoliths) to blob-like to vertical lozenge, suggests that emplacement occurs by different mechanisms with increasing depth in the crust.

In the brittle regime, magma may be arrested by a structure, instability or thermal death (Clemens & Mawer 1992; Brown & Solar 1998b). Emplacement occurs when mainly vertical flow switches to predominantly horizontal flow and vertical inflation. Depression of the floor and/or lifting of the roof allow inflation (Brown & Solar 1998b; Cruden 2006), together with a component of ductile strain in the aureole (e.g. Wagner *et al.* 2006). Stopping and assimilation are inferred to be volumetrically insignificant in the creation of space for intrusion and represent mechanisms for local accommodation of magma (Glazner & Bartley 2006; Saito *et al.* 2007). In the viscous regime, emplacement occurs by lateral expansion (swelling out like a balloon, accommodation mechanism(s) undefined) localized by some instability in the magma-wall-rock system or back-freezing in the ascent column.

In either case, the intrusion may be constructed incrementally, although the cryptic nature of internal contacts in granites suggests that successive melt batches are emplaced into mushy magma chambers that are not fully crystallized; such mushy chambers are natural traps for ascending batches of magma. Growth of an intrusion is potentially very fast (thousands to hundreds of thousands of years) if magma is delivered continuously or in pulses separated by short time intervals (days to years). However, intrusions may take several million years to form if the interval between periodic additions of new pulses of magma is close to the time for crystallization of each addition (e.g. Solar *et al.* 1998; Coleman *et al.* 2004; Matzel *et al.* 2005, 2006; Tomascak *et al.* 2005). The time taken for construction and whether pulse additions are periodic, related to periods of high magma influx, or distributed throughout the intrusive history may vary with depth of emplacement (e.g. Matzel *et al.* 2006) and the size of melt reservoirs.

For subhorizontal, tabular or wedge-shaped intrusions, thickness, *T*, and width, *L*, data appear to define an S-shaped distribution when plotted in log *T* vs. log *L* space (Fig. 9a),

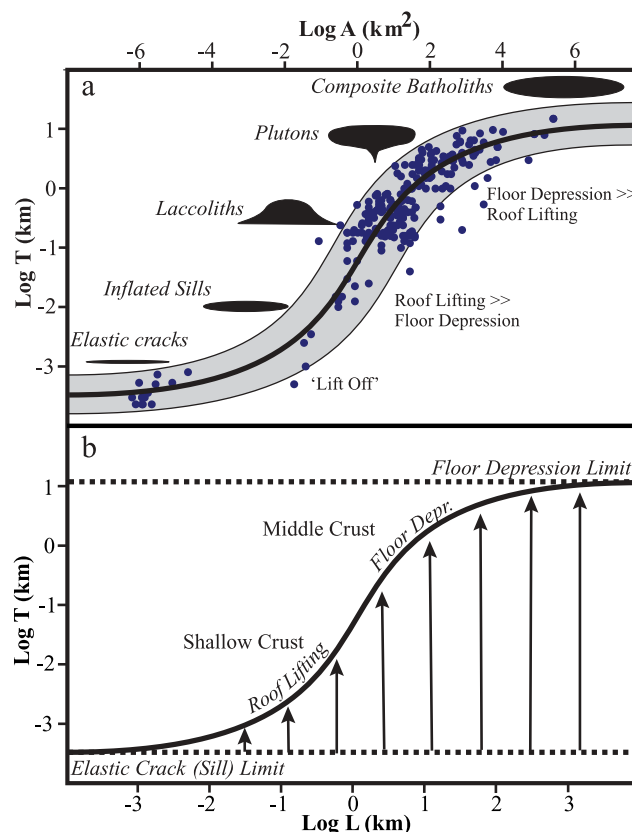


Fig. 9. (a) Compilation of intrusion thickness (*T*), width (*L*) and horizontal area (*A*) data (McCaffrey & Cruden 2002). The solid S-curve is drawn through the data by eye and the shaded area represents the approximate limits of the data about the curve. Representative intrusion styles are shown adjacent to appropriate parts of the curve. (b) An interpretation of the S-curve in terms of minimum and maximum growth limit, dominant emplacement mechanism and depth of emplacement in the crust (Cruden & McCaffrey 2002). Arrows show vertical growth trajectories along a possible end-member growth curve with a slope $a = \infty$.

which implies vertical limits for tabular intrusions of *T* *c.* 1 m when *L* < 10 m and *T* *c.* 10 km when *L* > 500 km (Fig. 9b). Between these limits *T* increases with increasing *L*, with a maximum slope, *a* *c.* 1.5, at around *L* = 1 km (McCaffrey & Cruden 2002; Cruden & McCaffrey 2002). Earlier empirical power law scaling relationships for laccoliths ($a = 0.8$) and plutons ($a = 0.6$) (McCaffrey & Petford 1997; Cruden & McCaffrey 2001) probably define tangents to the S-curve in Figure 9, rather than representing unique scaling relationships for different classes of intrusions. Limiting factors on vertical growth of intrusions include the host rock mechanical properties and the depth of emplacement. To lift the roof, P_{magma} must overcome lithostatic load and tectonic overpressure. At shallow levels, intrusions that grow by vertical inflation from a larger horizontal area are better able to lift the roof, but with increasing depth, vertical growth of large intrusions occurs dominantly by floor depression, which process is limited by the thickness of the source and the degree of melting in that zone (Cruden 2006).

Mechanical theories for the origination and growth of sills and laccoliths are reasonably well established (Cruden & McCaffrey 2002). One key parameter in the analysis of such intrusions is

the geometric ratio of the intrusion width to its emplacement depth, L/h . Given sufficient magma pressure, the sill–laccolith transition occurs where $L/h > 2\text{--}3$, depending on the geometry of the sill and the magma rheology. This reflects shallow emplacement ($h = 0.5\text{--}5\text{ km}$) for intrusions where $L = 1\text{--}10\text{ km}$, and corresponding to the lower half of the S-curve (Fig. 9), with the change of slope at $L = 0.1\text{--}1\text{ km}$ recording the increasing ability of shallow intrusions to lift their roofs as their horizontal area increases. The steepest slope probably represents a limiting curve for laccoliths as they expand vertically from an initial sill along a growth line with $a \gg 1$ (Fig. 9b; McCaffrey & Cruden 2002).

Vertical growth for moderate to large plutons occurs dominantly by floor depression (Brown & Solar 1998b; Cruden & McCaffrey 2001, 2002; Cruden 2006). Floor depression may occur by elastic down-bending of the pluton floor if L/h is small (deep intrusion) and if the elastic strength of the material beneath the intrusion is significantly less than that above it, as might be expected close to the brittle–ductile transition zone. An alternative mechanism for floor depression may be subsidence accommodated by the compacting source as melt is extracted, although the relationship is unlikely to be simple as the melt volume in the intrusion is from a much larger source volume. Kinematic models of floor depression involving piston and cantilever mechanisms indicate that emplacement and growth is a geologically rapid process, with typical intrusions forming over thousands to hundreds of thousands of years, at geologically reasonable strain rates (Cruden 2006). A number of different structural arrangements of phases within intrusions are predicted depending upon the growth mode (tabular vs. wedge) and nature of magma addition to the intrusion (continuous vs. pulsed; top-down vs. base-up). The vertical growth limit is now imposed by the thickness of the source and degree of melting, and the efficiency of melt extraction.

Composite batholiths with $L > 100\text{ km}$ lie on the uppermost part of the S-curve, where $a \ll 1$ (Fig. 9). Cruden & McCaffrey (2002) proposed that this represents an upper limit for space-creation by multiple stacking of intrusions in the middle crust. This limit is probably imposed by the maximum amount of floor subsidence that is possible.

In the viscous regime, melt flow is focused into actively deforming segments of orogens by localization of deformation. Within these segments melt flow is strongly domainal and is focused into zones of high permeability and/or melt pressure gradients, and generation and amplification of naturally occurring instabilities in the system probably cause the switch from ascent to emplacement. Instabilities may be internal to the ascent column, such as fluctuations in permeability or magma flow rate (or changes in cross-sectional shape), or external to the ascent column, such as variations in strength or strain rate or state of stress. These instabilities are not mutually exclusive, and feedback relations are likely to produce similar results whatever the initial instability.

Let us consider an ascent column with spatial and temporal variations in permeability; zones of higher permeability will be conduits of higher magma flux. However, higher magma flux may lead to increased permeability and heating, and consequent weakening of the rock surrounding these conduits, which in turn increases the strain rate, leading to higher magma fluxes and enhanced thermal weakening in zones of higher strain; all of these interactions lead to even higher magma fluxes and may lead to concentration of magma flow into a small number of larger ascent conduits. Magma in the preferred ascent conduits will exploit the weakening and may expand into the host rock,

switching ascent to emplacement as faster flow rates lead to expansion of the ascent conduit to form a vertical lozenge pluton or slower flow rates simply lead to contraction of the ascent conduit and crystallization *in situ*. Intrusions of this general type commonly are sheeted (e.g. Brown & Solar 1998a; Miller & Paterson 2001; Mahan *et al.* 2003), and the diapiric intrusions of Miller & Paterson (1999) may have been emplaced by such a mechanism.

A similar phenomenon occurs if the instability is due to differences in the strength of the host rock or the stress field around the ascent column. Magma exploits the weaker or lower stress sectors, swelling or intruding out of the ascent column into those sites (accommodation mechanism(s) will vary from the viscous to the brittle regime), heating and weakening the host rocks, which allows further lateral expansion, forming a blob-like intrusion. This is similar to the lateral expansion model proposed by Lagarde *et al.* (1990); the nested diapirs of Paterson & Vernon (1995) may have been emplaced by such a mechanism involving sequential arrival of successive batches of magma.

Above the anatectic zone, differences in flow rate or cross-sectional shape of the ascent conduit may lead to fluctuations around the critical width for flow without freezing. If freezing occurs in the slower (or narrower) parts of the conduit, flow will focus in the faster (or wider) parts. Heating of the host rock is likely to cause weakening that will facilitate swelling of the conduit. In the upper crust, at high strains, rock strength below the initial brittle–viscous transition depth-interval may decrease to less than that of the over- and underlying crust as a result of transitions from cataclasis (just above the initial brittle–viscous transition) or grain-size insensitive (dislocation) creep (just below the initial brittle–viscous transition) to grain-size sensitive creep (diffusion creep, diffusion- or reaction-accommodated granular flow) in fine-grained, polymineralic aggregates (Handy 1989; Handy *et al.* 2001). A fuller analysis of the energy feedbacks on continental strength, and the complex structural evolution related to dynamic changes in strength associated with the middle crust elastic core has been presented by Regenauer-Lieb *et al.* (2006).

A strain-dependent strength minimum just below the brittle–viscous transition has important implications for arresting ascent of magma and allowing lateral emplacement. At this crustal level, the switch to emplacement may lead to formation of an intrusion with a horizontal tabular head, formed by inflation of a subhorizontal magma fracture, and a hemi-ellipsoidal root that passes down into a migmatite zone through which magma was transferred (Brown & Solar 1999). In such a composite intrusion, the hemi-ellipsoidal root is thought to have formed by back filling and freezing as the thermal structure decayed as a result of declining magma flow rate. The decline in magma flow rate could be a consequence of the ‘end’ of magma emplacement or the ‘end’ of melt extraction. In effect, the rate of magma inflow to the intrusion is insufficient to maintain continued inflation and the resultant feedback ultimately terminates magma ascent and extraction from the source. This is consistent with the expected decline in extraction rate with time from a finite volume source.

Concluding summary

Melt concentration occurs by diffusive and advective process (Powell & Downes 1990; Brown *et al.* 1995; Brown 2004; White *et al.* 2004). Both processes may be inferred to have occurred in any one suprasolidus episode, based on evidence in individual outcrops of anatectic rocks, and commonly these processes act in concert. Permeability owing to an intergranular network of connected pores, and/or to the development of a network of

deformation bands, and/or to compositional layering, means that melt distribution commonly is heterogeneous at various length scales. The sequence of structures observed in nature records a transition from storage to drainage, and melt is inferred to have been extracted by upward movement along steeply dipping conduits.

Melt pathways are determined primarily by melt pressure gradients between source and sink at all scales; that is, from grain boundary to deformation band to conduit to intrusion (see Sleep 1988). Feedback relations allow an interconnected network of deformation bands to evolve to an optimum structure, probably controlled by rates of melt production and strain accumulation. The dynamic evolution of the network is illustrated by inference from concordant leucosome stromata. Although these sites initially accumulate melt, as the high-permeability network forms they come to represent structures analogous to compaction bands. Initially, for layer-parallel extension, shear-enhanced compaction drives melt from grain boundary pores in weaker layers to stronger layers (Brown *et al.* 1995) and to inter-boudin partitions. As an interconnected network of deformation bands develops and melt is drained from the network, a mix of residual and early crystallized phases with only minimal melt is retained as leucosome stromata because compactive deformation has expressed the melt from the sites of initial concentration into dilation and shear bands.

A critical point may be reached at some combination of melt fraction and distribution that allows formation of conduits of a size that allows melt to flow out of the source and upward through the system. For this to occur, it is likely that the melt volume will be in the range 7–10 vol.% and that a melt-bearing network of deformation bands will be interconnected. Such a conclusion is consistent with estimates of the melt fraction in the crust beneath the Central Andes and the Tibetan Plateau (Schilling & Partzsch 2001).

Whether each ascent conduit represents a single melt-draining event from its collection zone or whether there is 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. However, there is increasing evidence of common ascent conduits for multiple batches of melt (e.g. Brown & Solar 1999; Solar & Brown 2001a; Sawyer & Bonnay 2003), and some sheeted intrusions may represent staging levels in the transfer of melt to shallower levels, evidence of which is now eroded away (e.g. the McDoole pluton, Sierra Nevada; Mahan *et al.* 2003).

From the study of exhumed melt-depleted anatectic migmatite and granulite terranes, we are left with an impression that melt may be lost in batches. Such a conclusion is supported by detailed isotope geochemistry from granites where data from representative samples are interpreted to show that plutons commonly are constructed from multiple batches of isotopically distinct melt (Deniel *et al.* 1987; Hogan & Sinha 1991; Pressley & Brown 1999; Glazner *et al.* 2004).

We have come far in our understanding of melt generation and extraction from anatectic lower crustal sources, and magma ascent and emplacement into the upper crust of large hot orogens, but much remains to be done. I hope that this Bicentennial Review represents another step on the path to a better understanding. I have tried to represent the literature fairly within the constraints of reasonable length, and I acknowledge that my understanding, such as it is, owes as much to the many discussions, both indoors and in the field, that I have had with colleagues and students as to my own investigations and contributions. None the less I am responsible for any misrepresentations or infelicities that may persist. I

thank R. Vernon for his supportive review and R. White for detailed comments that materially improved the paper, T. Johnson (pseudosections) and B. Reno (figures) for help, and the US National Science Foundation and the University of Maryland for supporting my research into melt extraction from anatectic rocks.

References

- ALBERTZ, M., PATERSON, S.R. & OKAYA, D. 2005. Fast strain rates during pluton emplacement: magmatically folded leucocratic dikes in aureoles of the Mont Stuart Batholith, Washington, and the Tuolumne Intrusive Suite, California. *Geological Society of America Bulletin*, **117**, 450–465.
- ALLIBONE, A.H. & NORRIS, R.J. 1992. Segregation of leucogranite microplutons during syn-anatectic deformation—an example from the Taylor Valley, Antarctica. *Journal of Metamorphic Geology*, **10**, 589–600.
- ANDREOLI, M.A., HART, R.J., ASHWAL, L.D. & COETZEE, H. 2006. Correlations between U, Th content and metamorphic grade in the western Namaqualand Belt, South Africa, with implications for radioactive heating of the crust. *Journal of Petrology*, **47**, 1095–1118.
- ANNEN, C., BLUNDY, J.D. & SPARKS, R.S.J. 2006a. The genesis of intermediate and silicic magmas in deep crustal hot zones. *Journal of Petrology*, **47**, 71–95.
- ANNEN, C., SCAILLET, B. & SPARKS, R.S.J. 2006b. Thermal constraints on the emplacement rate of a large intrusive complex: the Manaslu leucogranite, Nepal Himalaya. *Journal of Petrology*, **47**, 71–95.
- ASHWORTH, J.R. & McLELLAN, E.L. 1985. Textures. In: ASHWORTH, J.R. (ed.) *Migmatites*. Blackie, Glasgow, 180–203.
- ATHERTON, M.P. 1993. Granite magmatism. *Journal of the Geological Society, London*, **150**, 1009–1023.
- AUZANNEAU, E., VIELZEUF, D. & SCHMIDT, M.W. 2006. Experimental evidence of decompression melting during exhumation of subducted continental crust. *Contributions to Mineralogy and Petrology*, **152**, 125–148.
- BABEYKO, A.Y., SOBOLEV, S.V., TRUMBULL, R.B., ONCKEN, O. & LAVIER, L.L. 2002. Numerical models of crustal-scale convection and partial melting beneath the Altiplano–Puna plateau. *Earth and Planetary Science Letters*, **199**, 373–388.
- BARBOZA, S.A., BERGANTZ, G.W. & BROWN, M. 1999. Regional granulite facies metamorphism in the Ivrea zone: is the Mafic Complex the smoking gun or a red herring? *Geology*, **27**, 447–450.
- BEA, F., MONTERO, P. & ORTEGA, M. 2006. A LA-ICP-MS evaluation of Zr reservoirs in common crustal rocks: implications for Zr and Hf geochemistry, and zircon-forming processes. *Canadian Mineralogist*, **44**, 693–714.
- BONS, P.D. 2001. The formation of large quartz veins by rapid ascent of fluids in mobile hydrofractures. *Tectonophysics*, **336**, 1–17.
- BONS, P.D. & ELBURG, M.A. 2001. Fractal size distribution of plutons: an example from the Lachlan Fold Belt, Australia. In: CHAPPELL, B.W. & FLEMING, P.D. (eds) *S-type Granites and Related Rocks*. Australian Geological Survey Organization Record, **2001/2**, 21–22.
- BONS, P.D. & VAN MILLIGEN, B.P. 2001. New experiment to model self-organized critical transport and accumulation of melt and hydrocarbons from their source rocks. *Geology*, **29**, 919–922.
- BONS, P.D., DOUGHERTY-PAGE, J. & ELBURG, M.A. 2001. Stepwise accumulation and ascent of magmas. *Journal of Metamorphic Geology*, **19**, 627–633.
- BONS, P.D., ARNOLD, J., ELBURG, M.A., CALDER, J. & SOESOO, A. 2004. Melt extraction and accumulation from partially molten rocks. *Lithos*, **78**, 25–42.
- BRAUN, M.G. & KELEMEN, P.B. 2002. Dunite distribution in the Oman ophiolite: implications for melt flux through porous dunite conduits. *Geochemistry, Geophysics, Geosystems*, **3**, article number 8603.
- BROWN, E.H. & McCLELLAND, W.C. 2000. Pluton emplacement by sheeting and vertical ballooning in part of the southeast Coast Plutonic Complex, British Columbia. *Geological Society of America Bulletin*, **112**, 708–719.
- BROWN, M. 1983. The petrogenesis of some migmatites from the Presqu'île de Rhuys, southern Brittany, France. In: ATHERTON, M.P. & GRIBBLE, C.D. (eds) *Migmatites, Melting and Metamorphism*. Shiva, Nantwich, 174–200.
- BROWN, M. 1994. The generation, segregation, ascent and emplacement of granite magma: the migmatite-to-crustally-derived granite connection in thickened orogens. *Earth-Science Reviews*, **36**, 83–130.
- BROWN, M. 1995. The late-Precambrian geodynamic evolution of the Armorican segment of the Cadomian belt (France): distortion of an active continental margin during south-west directed convergence and subduction of a bathymetric high. *Géologie de la France*, **3**, 3–22.
- BROWN, M. 2001a. Crustal melting and granite magmatism: key issues. *Physics and Chemistry of the Earth (A)*, **26**, 201–212.
- BROWN, M. 2001b. Orogeny, migmatites and leucogranites: a review. *Proceedings of the Indian Academy of Science*, **110**, 313–336.
- BROWN, M. 2002. Prograde and retrograde processes in migmatites revisited. *Journal of Metamorphic Geology*, **20**, 25–40.

- BROWN, M. 2004. Melt extraction from lower continental crust. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **95**, 35–48.
- BROWN, M. 2005. Synergistic effects of melting and deformation: an example from the Variscan Belt, western France. In: GAPAIS, D., BRUN, J.P. & COBBOLD, P.R. (eds) *Deformation Mechanism, Rheology and Tectonics: from Minerals to the Lithosphere*. Geological Society, London, Special Publications, **243**, 205–226.
- BROWN, M. 2006. Melt extraction from the lower continental crust of orogens: the field evidence. In: BROWN, M. & RUSHMER, T. (eds) *Evolution and Differentiation of the Continental Crust*. Cambridge University Press, Cambridge, 332–384.
- BROWN, M. & DALLMEYER, R.D. 1996. Rapid Variscan exhumation and role of magma in core complex formation: Southern Brittany metamorphic belt, France. *Journal of Metamorphic Geology*, **14**, 361–379.
- BROWN, M. & PRESSLEY, R.A. 1999. Crustal melting in nature: prosecuting source processes. *Physics and Chemistry of the Earth (A)*, **24**, 305–316.
- BROWN, M. & RUSHMER, T. 1997. The role of deformation in the movement of granite melt: views from the laboratory and the field. In: HOLNESS, M.B. (ed.) *Deformation-enhanced Fluid Transport in the Earth's Crust and Mantle*. Mineralogical Society Series, **8**, 111–144.
- BROWN, M. & RUSHMER, T. (eds) 2006. *Evolution and Differentiation of the Continental Crust*. Cambridge University Press, Cambridge.
- BROWN, M. & SOLAR, G.S. 1998a. Shear zone systems and melts: feedback relations and self-organization in orogenic belts. *Journal of Structural Geology*, **20**, 211–227.
- BROWN, M. & SOLAR, G.S. 1998b. Granite ascent and emplacement during contractional deformation in convergent orogens. *Journal of Structural Geology*, **20**, 1365–1393.
- BROWN, M. & SOLAR, G.S. 1999. The mechanism of ascent and emplacement of granite magma during transpression: a syntectonic granite paradigm. *Tectonophysics*, **312**, 1–33.
- BROWN, M. & SOLAR, G.S. 2000. Feedback relations between deformation and melt, the evolution from weakening to hardening in transpressive orogens. *Journal of the Czech Geological Society*, **45**, 215–216.
- BROWN, M., FRIEND, C.R.L., MCGREGOR, V.R. & PERKINS, W.T. 1981. The late Archaean Qorqut granite complex of southern West Greenland. *Journal of Geophysical Research*, **86**, 10617–10632.
- BROWN, M., AVERKIN, Y.A., McLELLAN, E.L. & SAWYER, E.W. 1995. Melt segregation in migmatites. *Journal of Geophysical Research*, **100**, 15655–15679.
- BROWN, M.A., BROWN, M., CARLSON, W.D. & DENISON, C. 1999. Topology of syntectonic melt flow networks in the deep crust: inferences from three-dimensional images of leucosome geometry in migmatites. *American Mineralogist*, **84**, 1793–1818.
- BURG, J.-P. & GERYA, T.V. 2005. The role of viscous heating in Barrovian metamorphism of collisional orogens: thermomechanical models and application to the Lepontine Dome in the Central Alps. *Journal of Metamorphic Geology*, **23**, 75–95.
- BUROV, E., JAUPART, C. & GUILLLOU-FROTTIER, L. 2003. Ascent and emplacement of buoyant magma bodies in brittle–ductile upper crust. *Journal of Geophysical Research*, **108**, 2177, doi:10.1029/2002JB001904.
- CHARDON, D., CHOUKROUNE, P. & JAYANANDA, M. 1998. Sinking of the Dharwar Basin (South India): implications for Archaean tectonics. *Precambrian Research*, **91**, 15–39.
- CHEN, Z. & JIN, Z.-H. 2006. Magma-driven subcritical crack growth and implications for dike initiation from a magma chamber. *Geophysical Research Letters*, **33**, doi:10.1029/2006GL026979.
- CLARK, M.K. & ROYDEN, L.H. 2000. Topographic ooze: building the eastern margin of Tibet by lower crustal flow. *Geology*, **28**, 703–706.
- CLARKE, G.L., WHITE, R.W., LUI, S., FITZHERBERT, J.A. & PEARSON, N.J. 2007. Contrasting behaviour of rare earth and major elements during partial melting in granulite facies migmatites, Wuluma Hills, Arunta Block, central Australia. *Journal of Metamorphic Geology*, **25**, 1–18.
- CLEMENS, J.D. 1998. Observations on the origins and ascent mechanisms of granitic magmas. *Journal of the Geological Society, London*, **155**, 843–851.
- CLEMENS, J.D. 2006. Melting of the continental crust: fluid regimes, melting reactions, and source-rock fertility. In: BROWN, M. & RUSHMER, T. (eds) *Evolution and Differentiation of the Continental Crust*. Cambridge University Press, Cambridge, 297–331.
- CLEMENS, J.D. & MAWER, C.K. 1992. Granitic magma transport by fracture propagation. *Tectonophysics*, **204**, 339–360.
- COLEMAN, D.S., GRAY, W. & GLAZNER, A.F. 2004. Rethinking the emplacement and evolution of zoned plutons: geochronologic evidence for incremental assembly of the Tuolumne Intrusive Suite, California. *Geology*, **32**, 433–436.
- COLLINS, W.J. & SAWYER, E.W. 1996. Pervasive magma transfer through the lower–middle crust during non-coaxial compressional deformation: an alternative to diking. *Journal of Metamorphic Geology*, **14**, 565–579.
- COLLINS, W.J., VAN KRANENDONK, M.J. & TEYSSIER, C. 1998. Partial convective overturn of Archaean crust in the east Pilbara Craton, Western Australia: driving mechanisms and tectonic implications. *Journal of Structural Geology*, **20**, 1405–1424.
- CONNOLLY, J.A.D. & PODLADCHIKOV, Y.Y. 1998. Compaction-driven fluid flow in viscoelastic rock. *Geodinamica Acta*, **11**, 55–84.
- CRUDEN, A.R. 2006. Emplacement and growth of plutons: implications for rates of melting and mass transfer in continental crust. In: BROWN, M. & RUSHMER, T. (eds) *Evolution and Differentiation of the Continental Crust*. Cambridge University Press, Cambridge, 455–519.
- CRUDEN, A.R. & McCAFFREY, K.J.W. 2001. Growth of plutons by floor subsidence: implications for rates of emplacement, intrusion spacing and melt-extraction mechanisms. *Physics and Chemistry of the Earth, A*, **26**, 303–315.
- CRUDEN, A.R. & McCAFFREY, K.J.W. 2002. Different scaling laws for sills, laccoliths and plutons: mechanical thresholds on roof lifting and floor depression. In: BREITKREUZ, C., MOCK, A. & PETFORD, N. (eds) *First International Workshop: Physical Geology of Subvolcanic Systems—Laccoliths, Sills and Dykes (LASI)*. Wissenschaftliche Mitteilung Institut für Geologie, Technische Universität Bergakademie Freiberg, **20/2002**, 15–17.
- CURRIE, C.A. & HYNDMAN, R.D. 2005. The thermal structure of subduction zones: the case for a hot backarc. *EOS Transactions, American Geophysical Union*, **86**, Fall Meeting Supplement, Abstract T11B-0366.
- DAHM, T. 2000. On the shape and velocity of fluid-filled fractures in the Earth. *Geophysical Journal International*, **142**, 181–192.
- DEGELING, H., EGGINS, S. & ELLIS, D.J. 2001. Zr budgets for metamorphic reactions, and the formation of zircon from garnet breakdown. *Mineralogical Magazine*, **65**, 749–758.
- DENIEL, C., VIDAL, P., FERNANDEZ, A., LEFORT, P. & PEUCAT, J.J. 1987. Isotopic study of the Manaslu Granite (Himalaya, Nepal)—inferences on the age and source of Himalayan leucogranites. *Contributions to Mineralogy and Petrology*, **96**, 78–92.
- DE SAINT-BLANQUAT, M., HABERT, G., HORSMAN, E., MORGAN, S.S., TIKOFF, B., LAUNEAU, P. & GLEIZES, G. 2007. Mechanisms and duration of non-tectonically assisted magma emplacement in the upper crust: the Black Mesa pluton, Henry Mountains, Utah. *Tectonophysics*, **428**, 1–31.
- DEWEY, J.F., ROBB, L. & VAN SCHALKWYK, L. 2006. Did Bushmanland extensionally unroof Namaqualand? *Precambrian Research*, **150**, 173–182.
- D'LEMOIS, R.S., BROWN, M. & STRACHAN, R.A. 1992. Granite magma generation, ascent and emplacement within a transpressional orogen. *Journal of the Geological Society, London*, **149**, 487–490.
- DU BERNARD, X., EICHHUBL, P. & AYDIN, A. 2002. Dilation bands: a new form of localized failure in granular media. *Geophysical Research Letters*, **29**, doi:10.1029/2002GL015966.
- EICHHUBL, P. & AYDIN, A. 2003. Ductile opening-mode fracture by pore growth and coalescence during combustion alteration of siliceous mudstone. *Journal of Structural Geology*, **25**, 121–134.
- EICHHUBL, P. & BOLES, J.R. 2000. Focused fluid flow along faults in the Monterey Formation, Coastal California. *Geological Society of America Bulletin*, **112**, 1667–1679.
- EICHHUBL, P., AYDIN, A. & LORE, J. 2001. Opening-mode fracture in siliceous mudstone at high homologous temperature—effective of surface forces. *Geophysical Research Letters*, **28**, 1299–1302.
- EMERMAN, S.H. & MARRETT, R. 1990. Why dykes? *Geology*, **18**, 231–233.
- FRASER, G., ELLIS, D. & EGGINS, S. 1997. Zirconium abundance in granulite-facies minerals, with implications for zircon geochronology in high-grade rocks. *Geology*, **25**, 607–610.
- GAPAIS, D., LAGARDE, J.-L., LE CORRE, C., AUDREN, C., JÉGOUZO, P., CASAS SAINZ, A. & VAN DEN DRIESCH, J. 1993. La zone de cisaillement de Quiberon: témoin d'extension de la chaîne varisque en Bretagne méridionale au Carbonifère. *Comptes Rendus de l'Académie des Sciences*, **316**, 1123–1129.
- GLAZNER, A.F. & BARTLEY, J.M. 2006. Is stopping a volumetrically significant pluton emplacement process? *Geological Society of America Bulletin*, **118**, 1185–1195.
- GLAZNER, A.F., BARTLEY, J.M., COLEMAN, D.S., GRAY, W. & TAYLOR, R.Z. 2004. Are plutons assembled over millions of years by amalgamation from small magma chambers? *GSA Today*, **14**, 4–11.
- GROCCOTT, J., BROWN, M., DALLMEYER, R.D., TAYLOR, G.K. & TRELOAR, P.J. 1994. Mechanisms of continental growth in extensional arcs: an example from the Andean plate boundary zone. *Geology*, **22**, 391–394.
- GRUJIC, D. & MANCKTELLOW, N.S. 1998. Melt-bearing shear zones: analogue experiments and comparison with examples from southern Madagascar. *Journal of Structural Geology*, **20**, 673–680.
- GUERNIN, S. & SAWYER, E.W. 2003. Large-scale melt-depletion in granulite terranes: an example from the Archaean Ashuanipi Subprovince of Quebec. *Journal of Metamorphic Geology*, **21**, 181–201.
- HANDY, M.R. 1989. Deformation regimes and the rheological evolution of fault zones in the lithosphere: the effects of pressure, temperature, grain size and time. *Tectonophysics*, **163**, 119–152.

- HANDY, M.R., MULCH, A., ROSENAU, N. & ROSENBERG, C.L. 2001. The role of fault zones and melts as agents of weakening, hardening and differentiation of the continental crust: a synthesis. In: HOLDSWORTH, R.E., MAGLOUGHLIN, J., KNIFE, R.J., STRACHAN, R.A. & SEARLE, R.C. (eds) *The Nature and Tectonic Significance of Fault Zone Weakening*. Geological Society, London, Special Publications, **186**, 305–332.
- HARRIS, N. & MASSEY, J. 1994. Decompression and anatexis of Himalayan metapelites. *Tectonics*, **13**, 1537–1546.
- HARRIS, N.B.W. & INGER, S. 1992. Trace-element modelling of pelite-derived granites. *Contributions to Mineralogy and Petrology*, **110**, 46–56.
- HARRISON, T.M., GROVE, M., LOVERA, O.M. & CATLOS, E.J. 1998. A model for the origin of Himalayan anatexis and inverted metamorphism. *Journal of Geophysical Research—Solid Earth*, **103**, 27017–27032.
- HINCHEY, A.M. & CARR, S.D. 2006. The S-type Ladybird leucogranite suite of southeastern British Columbia: geochemical and isotopic evidence for a genetic link with migmatite formation in the North American basement gneisses of the Monashee complex. *Lithos*, **90**, 223–248.
- HOGAN, J.P. & SINHA, A.K. 1991. The effect of accessory minerals on the redistribution of lead isotopes during crustal anatexis: a model. *Geochimica et Cosmochimica Acta*, **55**, 335–348.
- HOLLAND, T.J.B. & POWELL, R. 1998. An internally consistent thermodynamic data set for phases of petrological interest. *Journal of Metamorphic Geology*, **16**, 309–343.
- HOLLAND, T.J.B. & POWELL, R. 2001. Calculation of phase relations involving haplogranitic melts using an internally consistent thermodynamic data set. *Journal of Petrology*, **42**, 673–683.
- HUTTON, D.H.W. 1988. Granite emplacement mechanisms and tectonic controls: inferences from deformation studies. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **79**, 245–255.
- HYNDMAN, R.D., CURRIE, C.A. & MAZZOTTI, S.P. 2005. Subduction zone backarcs, mobile belts, and orogenic heat. *GSA Today*, **15**, 4–10.
- JACKSON, M.D., CHEADLE, M.J. & ATHERTON, M.P. 2003. Quantitative modeling of granitic melt generation and segregation in the continental crust. *Journal of Geophysical Research—Solid Earth*, **108**, article number 2332.
- JACKSON, M.P.A. & VENDEVILLE, B.C. 1994. Regional extension as a geologic trigger for diapirism. *Geological Society of America Bulletin*, **106**, 57–73.
- JOHNSON, T. & BROWN, M. 2004. Quantitative constraints on metamorphism in the Variscides of Southern Brittany—a complementary pseudosection approach. *Journal of Petrology*, **45**, 1237–1259; doi:10.1093/petrology/egh012.
- JOHNSON, T.E., BROWN, M. & SOLAR, G.S. 2003a. Low-pressure subsolidus and suprasolidus phase equilibria in the MnNCKFMASH system: constraints on conditions of regional metamorphism in western Maine, northern Appalachians. *American Mineralogist*, **88**, 624–638.
- JOHNSON, T.E., HUDSON, N.F.C. & DROOP, G.T.R. 2003b. Evidence for a genetic granite–migmatite link in the Dalradian of NE Scotland. *Journal of the Geological Society, London*, **160**, 447–457.
- JOHNSON, T., BROWN, M., GIBSON, R. & WING, B. 2004. Spinel–cordierite symplectites replacing andalusite: evidence for melt-assisted diapirism in the Bushveld Complex, South Africa. *Journal of Metamorphic Geology*, **22**, 529–545; doi:10.1111/j.1525-1314.2004.00531.x.
- JONES, K.A. & BROWN, M. 1990. High temperature ‘clockwise’ P–T paths and melting in the development of regional migmatites: an example from southern Brittany, France. *Journal of Metamorphic Geology*, **8**, 551–578.
- KELLY, N.M., CLARKE, G.L. & HARLEY, S.L. 2005. Monazite behaviour and age significance in poly-metamorphic high-grade terrains: a case study from the western Musgrave Block, central Australia. *Lithos*, **88**, 100–134.
- KINCAID, C. & SILVER, P. 1996. The role of viscous dissipation in the orogenic process. *Earth and Planetary Science Letters*, **142**, 271–288.
- KOUKOUELAS, I.K., PE-PIPER, G. & PIPER, D.J.W. 2006. The relationship between length and width of plutons within the crustal-scale Cobeguid Shear Zone, northern Appalachians, Canada. *International Journal of Earth Sciences*, **95**, 963–976.
- KRETZ, R. 1983. Symbols for rock-forming minerals. *American Mineralogist*, **68**, 277–279.
- LABROUSSE, L., JOLIVET, L., AGARD, P., HEBERT, R. & ANDERSEN, T.B. 2002. Crustal-scale boudinage and migmatization of gneiss during their exhumation in the UHP province of western Norway. *Terra Nova*, **14**, 263–270.
- LAFRANCE, B., CLARKE, G.L., COLLINS, W.J. & WILLIAMS, I.S. 1995. The emplacement of the Wuluma granite: melt generation and migration along steeply dipping extensional fractures at the close of the Late Strangways orogenic event, Arunta Block, central Australia. *Precambrian Research*, **72**, 43–67.
- LAGARDE, J.L., BRUN, J.P. & GAPAIS, D. 1990. Formation of epizonal granitic plutons by in situ assemblage of laterally expanding magma. *Comptes Rendus de l’Académie des Sciences, Série II*, **310**, 1109–1114.
- LANG, H.M. & GILLOTTI, J.A. 2007. Partial melting of metapelites at ultrahigh-pressure conditions, Greenland Caledonides. *Journal of Metamorphic Geology*, **26**, 129–147.
- LAPORTE, D. & PROVOST, A. 2000. Equilibrium geometry of a fluid phase in a polycrystalline aggregate with anisotropic surface energies: dry grain boundaries. *Journal of Geophysical Research*, **105**, 25937–25953.
- LAPORTE, D., RAPAILLE, C. & PROVOST, A. 1997. Wetting angles, equilibrium melt geometry, and the permeability threshold of partially molten crustal protoliths. In: BOUCHEZ, J.L., HUTTON, D.H.W. & STEPHENS, W.E. (eds) *Granite: From Segregation of Melt to Emplacement Fabrics*. Kluwer, Dordrecht, 31–54.
- LEITCH, A.M. & WEINBERG, R.F. 2002. Modelling granite migration by mesoscale pervasive flow. *Earth and Planetary Science Letters*, **200**, 131–146.
- LELOUP, P.H. & KIENAST, J.R. 1993. High-temperature metamorphism in a major strike-slip shear zone—the Ailao Shan–Red River, People’s Republic of China. *Earth and Planetary Science Letters*, **118**, 213–234.
- LUPULESCU, A. & WATSON, E.B. 1999. Low melt fraction connectivity of granitic and tonalitic melts in a mafic crustal rock at 800 °C and 1 GPa. *Contributions to Mineralogy and Petrology*, **134**, 202–216.
- MAALOE, S. 1992. Melting and diffusion processes in closed-system migmatization. *Journal of Metamorphic Geology*, **10**, 503–516.
- MAHAN, K.H., BARTLEY, J.M., COLEMAN, D.S., GLAZNER, A.F. & CARL, B.S. 2003. Sheeted intrusion of the synkinematic McDoyle pluton, Sierra Nevada, California. *Geological Society of America Bulletin*, **115**, 1570–1582.
- MARCHILDON, N. & BROWN, M. 2001. Melt segregation in late syn-tectonic anatectic migmatites: an example from the Onawa Contact Aureole, Maine, U.S.A. *Physics and Chemistry of the Earth (A)*, **26**, 225–229.
- MARCHILDON, N. & BROWN, M. 2002. Grain-scale melt distribution in two contact aureole rocks: implication for controls on melt localization and deformation. *Journal of Metamorphic Geology*, **20**, 381–396.
- MARCHILDON, N. & BROWN, M. 2003. Spatial distribution of melt-bearing structures in anatectic rocks from Southern Brittany: implications for melt-transfer at grain- to orogen-scale. *Tectonophysics*, **364**, 215–235.
- MATZEL, J., MUNDIL, R., PATERSON, S., RENNE, P. & NOMADE, S. 2005. Evaluating pluton growth models using high resolution geochronology: Tuolumne intrusive suite, Sierra Nevada, CA. *Geological Society of America, Abstracts with Programs*, **37**, 131.
- MATZEL, J.E.P., BOWRING, S.A. & MILLER, R.B. 2006. Time scales of pluton construction at differing crustal levels: examples from the Mount Stuart and Tenpeak intrusions, North Cascades, Washington. *Geological Society of America Bulletin*, **118**, 1412–1430.
- MCCAFFREY, K.J.W. & CRUDEN, A.R. 2002. Dimensional data and growth models for intrusions. In: BREITKREUZ, C., MOCK, A. & PETFORD, N. (eds) *First International Workshop: Physical Geology of Subvolcanic Systems—Laccoliths, Sills and Dykes (LASI)*. Wissenschaftliche Mitteilung Institut für Geologie, Technische Universität Bergakademie Freiberg, **20/2002**, 37–39.
- MCCAFFREY, K.J.W. & PETFORD, N. 1997. Are granitic intrusions scale invariant? *Journal of the Geological Society, London*, **154**, 1–4.
- MCCLAREN, S., SANDIFORD, M., POWELL, R., NEUMANN, N. & WOODHEAD, J. 2006. Palaeozoic intraplate crustal anatexis in the Mount Painter province, South Australia: timing, thermal budgets and the role of crustal heat production. *Journal of Petrology*, **47**, 2281–2302.
- MILLER, C.F., McDOWELL, S.M. & MAPES, R.W. 2003. Hot and cold granites? Implications of zircon saturation temperatures and preservation of inheritance. *Geology*, **31**, 529–532.
- MILLER, R.B. & PATERSON, S.R. 1999. In defense of magmatic diapirs. *Journal of Structural Geology*, **21**, 1161–1173.
- MILLER, R.B. & PATERSON, S.R. 2001. Construction of mid-crustal sheeted plutons: examples from the north Cascades, Washington. *Geological Society of America Bulletin*, **113**, 1423–1442.
- NELSON, K.D., ZHAO, W. & BROWN, L.D. ET AL. 1996. Partially molten middle crust beneath southern Tibet: synthesis of Project INDEPTH results. *Science*, **274**, 1684–1688.
- NEMCHIN, A. & BODORKOS, S. 2000. Zr and LREE concentrations in anatectic melt as a function of crystal size distributions of zircon and monazite in the source region. *Geological Society of America Annual Meeting, Program and Abstracts*, Abstract No. 52286.
- OLIVER, N.H.S. & BARR, T.D. 1997. The geometry and evolution of magma pathways through migmatites of the Halls Creek Orogen, Western Australia. *Mineralogical Magazine*, **61**, 3–14.
- OLSEN, S.N., MARSH, B.D. & BAUMGARTNER, L.P. 2004. Modelling mid-crustal migmatite terrains as feeder zones for granite plutons: the competing dynamics of melt transfer by bulk versus porous flow. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **95**, 49–58.
- PATERSON, S.R. & FOWLER, T.K. JR 1993. Re-examining pluton emplacement processes. *Journal of Structural Geology*, **15**, 191–206.
- PATERSON, S.R. & VERNON, R.H. 1995. Bursting the bubble of ballooning plutons: a return to nested diapirs emplaced by multiple processes. *Geological Society of America Bulletin*, **107**, 1356–1380.
- PESCHLER, A.P., BENN, K. & ROEST, W.R. 2004. Insights on Archean continental geodynamics from gravity modelling of granite–greenstone terranes. *Journal*

- of *Geodynamics*, **38**, 185–207.
- PETFORD, N. & KOENDERS, M.A. 1998. Self-organization and fracture connectivity in rapidly heated continental crust. *Journal of Structural Geology*, **20**, 1425–1434.
- PETFORD, N., CRUDEN, A.R., McCAFFREY, K.J.W. & VIGNERESSE, J.L. 2000. Granite magma formation, transport and emplacement in the Earth's crust. *Nature*, **408**, 669–673.
- PODLADCHIKOV, YU.YU. & CONNOLLY, J.A.D. 2001. The transition from pervasive to segregated melt flow in ductile rock. *EUG XI, Symposium VPPS, Journal of Conference Abstracts*, 814.
- POWELL, R. & DOWNES, J. 1990. Garnet porphyroblast-bearing leucosomes in metapelites: mechanisms, phase diagrams and an example from Broken Hill. In: ASHWORTH, J.R. & BROWN, M. (eds) *High Temperature Metamorphism and Crustal Anatexis*. Unwin Hyman, London, 105–123.
- POWELL, R., HOLLAND, T. & WORLEY, B. 1998. Calculating phase diagrams involving solid solutions via non-linear equations, with examples using THERMOCALC. *Journal of Metamorphic Geology*, **16**, 577–588.
- POWELL, R., GUIRAUD, M. & WHITE, R.W. 2005. Truth and beauty in metamorphic phase equilibria: conjugate variables and phase diagrams. *Canadian Mineralogist*, **43**, 21–33.
- PRESSLEY, R.A. & BROWN, M. 1999. The Phillips Pluton, Maine, USA: evidence of heterogeneous crustal sources, and implications for granite ascent and emplacement mechanisms in convergent orogens. *Lithos*, **46**, 335–366.
- PYLE, J.M. & SPEAR, F.S. 1999. Yttrium zoning in garnet; coupling of major and accessory phases during metamorphic reactions. *Geological Materials Research*, **1**, 6.
- PYLE, J.M., SPEAR, F.S., RUDNICK, R.L. & McDONOUGH, W.F. 2001. Monazite–xenotime–garnet equilibrium in metapelites and the new monazite–garnet thermometer. *Journal of Petrology*, **42**, 2083–2107.
- RABINOWICZ, M. & VIGNERESSE, J.L. 2004. Melt segregation under compaction and shear channeling: application to granitic magma segregation in a continental crust. *Journal of Geophysical Research—Solid Earth*, **109**, article number B04407.
- RABINOWICZ, M., KENTON, P., CEULENEER, G. & HILLAIRET, M. 2001. Compaction in the mantle mush with high concentrations and the generation of magma chambers. *Earth and Planetary Science Letters*, **188**, 313–328.
- REGENAUER-LIEB, K., WEINBERG, R.F. & ROSENBAUM, G. 2006. The effect of energy feedbacks on continental strength. *Nature*, **442**, 67–70.
- REID, M.R. 1990. Ionprobe investigation of rare earth element distributions and partial melting of metasedimentary granulite. In: VIELZEUF, D. & VIDAL, P. (eds) *Granulites and Crustal Evolution*. NATO ASI Series. Series C: Mathematical and Physical Sciences, **311**, 507–522.
- RICHARDSON, C.N., LISTER, J.R. & MCKENZIE, D. 1996. Melt conduits in a viscous porous matrix. *Journal of Geophysical Research*, **101**, 20423–20432.
- ROSENBERG, C.L. 2004. Shear zones and magma ascent: a model based on a review of the Tertiary magmatism in the Alps. *Tectonics*, **23**, article number TC3002.
- ROSENBERG, C.L. & HANDY, M.R. 2000. Syntectonic melt pathways during simple shearing of a partially molten rock analogue (Norcamphor–Benzamide). *Journal of Geophysical Research*, **105**, 3135–3149.
- ROSENBERG, C.L. & HANDY, M.R. 2001. Mechanisms and orientation of melt segregation paths during pure shearing of a partially molten rock analog (norcamphor–benzamide). *Journal of Structural Geology*, **23**, 1917–1932.
- ROSENBERG, C.L. & HANDY, M.R. 2005. Experimental deformation of partially melted granite revisited: implications for the continental crust. *Journal of Metamorphic Geology*, **23**, 19–28.
- ROSENBERG, C.L. & RILLER, U. 2000. Partial-melt topology in statically and dynamically recrystallized granite. *Geology*, **28**, 7–10.
- RUBATTO, D., WILLIAMS, I.S. & BUICK, I.S. 2001. Zircon and monazite response to prograde metamorphism in the Reynolds Range, central Australia. *Contributions to Mineralogy and Petrology*, **140**, 458–468.
- RUBIN, A.M. 1993. Dykes vs. diapirs in visco-elastic rock. *Earth and Planetary Science Letters*, **117**, 653–670.
- RUBIN, A.M. 1995. Getting granite dikes out of the source region. *Journal of Geophysical Research—Solid Earth*, **100**, 5911–5929.
- RUBIN, A.M. 1998. Dyke ascent in partially molten rock. *Journal of Geophysical Research*, **103**, 20901–20919.
- RUSHMER, T. 2001. Volume change during partial melting reactions: implications for melt extraction, melt geochemistry and crustal rheology. *Tectonophysics*, **342**, 389–405.
- RUTTER, E.H. 1997. The influence of deformation on the extraction of crustal melts: a consideration of the role of melt-assisted granular flow. In: HOLNESS, M.B. (ed.) *Deformation-enhanced Fluid Transport in the Earth's Crust and Mantle*. Chapman & Hall, London, 82–110.
- RUTTER, E.H. & MECKLENBURGH, J. 2006. The extraction of melt from crustal protoliths and the flow behavior of partially molten crustal rocks: an experimental perspective. In: BROWN, M. & RUSHMER, T. (eds) *Evolution and Differentiation of the Continental Crust*. Cambridge University Press, Cambridge, 386–429.
- RUTTER, E.H., BRODIE, K.H. & IRVING, D.H. 2006. Flow of synthetic, wet, partially molten 'granite' under undrained conditions: an experimental study. *Journal of Geophysical Research*, **111**, doi:10.1029/B004257.
- SAITO, S., ARIMA, M. & NAKAJIMA, T. 2007. Hybridization of a shallow 'I-type' granitoid pluton and its host migmatite by magma-chamber wall collapse: the Tokuku pluton, central Japan. *Journal of Petrology*, **48**, 79–111.
- SANDIFORD, M., VAN KRANENDONK, M.J. & BODORKOS, S. 2004. Conductive incubation and the origin of dome-and-keel structure in Archean granite–greenstone terrains: a model based on the eastern Pilbara Craton, Western Australia. *Tectonics*, **23**, article number TC1009.
- SAWYER, E.W. 1987. The role of partial melting and fractional crystallization in determining discordant migmatite leucosome compositions. *Journal of Petrology*, **28**, 445–473.
- SAWYER, E.W. 1998. Formation and evolution of granite magmas during crustal reworking: the significance of diatexites. *Journal of Petrology*, **39**, 1147–1167.
- SAWYER, E.W. 1999. Criteria for the recognition of partial melting. *Physics and Chemistry of the Earth, (A)*, **24**, 269–279.
- SAWYER, E.W. 2001. Melt segregation in the continental crust: distribution and movement of melt in anatectic rocks. *Journal of Metamorphic Geology*, **18**, 291–309.
- SAWYER, E.W. & BONNAY, M. 2003. Melt segregation and magma movement in the crust. *Geophysical Research Abstracts*, **5**, abstract 02458.
- SAWYER, E.W., DOMBROWSKI, C. & COLLINS, W.J. 1999. Movement of melt during synchronous regional deformation and granulite-facies anatexis, an example from the Wuluma Hills, central Australia. In: CASTRO, A., FERNANDEZ, C. & VIGNERESSE, J.-C. (eds) *Understanding Granites: Integrating New and Classical Techniques*. Geological Society, London, Special Publications, **168**, 221–237.
- SCHILLING, F.R. & PARTZSCH, G.N. 2001. Quantifying partial melt fraction in the crust beneath the central Andes and the Tibetan Plateau. *Physics and Chemistry of the Earth, A*, **26**, 239–246.
- SECOR, D.T. & POLLARD, D.D. 1975. On the stability of open hydraulic fractures in the Earth's crust. *Geophysical Research Letters*, **2**, 510–513.
- SHAW, H.R. & CHOUET, B.A. 1991. Fractal hierarchies of magma transport in Hawaii and critical self-organization of tremor. *Journal of Geophysical Research*, **96**, 10191–10207.
- SIMAKIN, A. & TALBOT, C. 2001. Transfer of melt between microscopic pores and macroscopic veins in migmatites. *Physics and Chemistry of the Earth, A*, **26**, 363–367.
- SLAGSTAD, T., JAMIESON, R.A. & CULSHAW, N.G. 2005. Formation, crystallization, and migration of melt in the mid-orogenic crust: Muskoka domain migmatites, Grenville Province, Ontario. *Journal of Petrology*, **46**, 893–919.
- SLEEP, N.H. 1988. Tapping of melt by veins and dykes. *Journal of Geophysical Research*, **93**, 10255–10272.
- SOLAR, G.S. & BROWN, M. 2001a. Petrogenesis of migmatites in Maine, USA: possible source of peraluminous leucogranite in plutons. *Journal of Petrology*, **42**, 789–823.
- SOLAR, G.S. & BROWN, M. 2001b. Deformation partitioning during transpression in response to Early Devonian oblique convergence, northern Appalachian orogen, USA. *Journal of Structural Geology*, **23**, 1043–1065.
- SOLAR, G.S., PRESSLEY, R.A., BROWN, M. & TUCKER, R.D. 1998. Granite ascent in convergent orogenic belts; testing a model. *Geology*, **26**, 711–714.
- SPARKS, D.W. & PARMENTIER, E.M. 1991. Melt extraction from the mantle beneath spreading centers. *Earth and Planetary Science Letters*, **105**, 368–377.
- SPEAR, F.S., KOHN, M.J. & CHENEY, J.T. 1999. *P–T* paths from anatectic pelites. *Contributions to Mineralogy and Petrology*, **134**, 17–32.
- SPENCE, D.A. & TURCOTTE, D.L. 1985. Magma-driven propagation of cracks. *Journal of Geophysical Research*, **90**, 575–580.
- SPENCE, D.A., SHARP, P. & TURCOTTE, D.L. 1987. Buoyancy-driven crack propagation: a mechanism for magma migration. *Journal of Fluid Mechanics*, **17**, 135–153.
- STEVENS, G., VILLAROS, A. & MOYEN, J.-F. 2007. Selective peritectic garnet entrainment as the origin of geochemical diversity in S-type granites. *Geology*, **35**, 9–12.
- STRONG, D.F. & HANMER, S.K. 1981. Peraluminous granites. *Canadian Mineralogist*, **19**, 163–176.
- STÜWE, K. 1997. Effective bulk composition changes due to cooling: a model predicting complexities in retrograde reaction textures. *Contributions to Mineralogy and Petrology*, **129**, 43–52.
- STÜWE, K. 1998. Heat sources of Cretaceous metamorphism in the Eastern Alps—a discussion. *Tectonophysics*, **287**, 251–269.
- TAKADA, A. 1990. Experimental study on propagation of liquid-filled crack in gelatin: shape and velocity in hydrostatic stress conditions. *Journal of Geophysical Research*, **95**, 8471–8481.
- TANNER, D.C. 1999. The scale-invariant nature of migmatites from the Oberpfalz, NE Bavaria and its significance for melt transport. *Tectonophysics*, **302**, 297–305.

- TEYSSIER, C. & WHITNEY, D.L. 2002. Gneiss domes and orogeny. *Geology*, **30**, 1139–1142.
- TOMASCAK, P.B., BROWN, M., SOLAR, G.S., BECKER, H.J., CENTORBI, T.L. & TIAN, J. 2005. Source contributions to Devonian granite magmatism near the Laurentian Border, New Hampshire and Western Maine, USA. *Lithos*, **80**, 75–99.
- TURCOTTE, D.L. 1999. Self-organized criticality. *Reports on Progress in Physics*, **62**, 1377–1429.
- TURCOTTE, D.L. 2001. Self-organized criticality: does it have anything to do with criticality and is it useful? *Nonlinear Processes in Geophysics*, **8**, 193–196.
- UKEN, R. & WATKEYS, M.K. 1997. Diapirism initiated by the Bushveld Complex, South Africa. *Geology*, **25**, 723–726.
- UNSWORTH, M.J., JONES, A.G., WEI, W., MARQUIS, G., GOKARN, S.G. & SPRATT, J.E. 2005. Crustal rheology of the Himalaya and Southern Tibet inferred from magnetotelluric data. *Nature*, **438**, 78–81.
- VAN DER MOLEN, I. 1985. Interlayer material transport during layer-normal shortening, II, boudinage, pinch-and-swell and migmatite at Søndre Strømfjord Airport, west Greenland. *Tectonophysics*, **115**, 275–295.
- VERNON, R.H. & COLLINS, W.J. 1988. Igneous microstructures in migmatites. *Geology*, **16**, 1126–1129.
- VERNON, R.H., COLLINS, W.J. & RICHARDS, S.W. 2003. Contrasting magmas in metapelitic and metapsammitic migmatites in the Cooma Complex, Australia. *Visual Geoscience*, **8**, 45–54.
- VIGNERESSE, J.-L. 2004. Rheology of a two-phase material with applications to partially molten rocks, plastic deformation and saturated soils. In: ALSOP, G.I., HOLDSWORTH, R.E., MCCAFFREY, K.J.W. & HAND, M. (eds) *Flow Processes in Faults and Shear Zones*. Geological Society, London, Special Publications, **224**, 79–94.
- VIGNERESSE, J.L. & BURG, J.P. 2000. Continuous vs. discontinuous melt segregation in migmatites: insights from a cellular automaton model. *Terra Nova*, **12**, 188–192.
- VIGNERESSE, J.L. & BURG, J.P. 2004. Strain-rate-dependent rheology of partially molten rocks. In: GROCOTT, J., TIKOFF, B., MCCAFFREY, K.J.W. & TAYLOR, G. (eds) *Vertical Coupling and Decoupling in the Lithosphere*. Geological Society, London, Special Publications, **227**, 327–336.
- VILLASECA, C., MARTÍN ROMERA, C., DE LA ROSA, J. & BARBERO, L. 2003. Residence and redistribution of REE, Y, Zr, Th and U during granulite-facies metamorphism: behaviour of accessory and major phases in peraluminous granulites of central Spain. *Chemical Geology*, **200**, 293–323.
- VILLASECA, C., OREJANA, D. & PATERSON, B.A. 2007. Zr–LREE rich minerals in residual peraluminous granulites, another factor in the origin of low Zr–LREE granitic melts? *Lithos*, doi: 10.1016/j.lithos.2006.11.002.
- WAGNER, R., ROSENBERG, C., HANDY, M.R., MÖBUS, C. & ALBERTZ, M. 2006. Fracture-driven intrusion and upwelling of a mid-crustal pluton fed from a transpressive shear zone—the Rieserferner Pluton (Eastern Alps). *Geological Society of America Bulletin*, **118**, 219–237.
- WANNAMAKER, P.E. 2000. Comment on ‘The petrologic case for a dry lower crust’ by Bruce W. D. Yardley and John W. Valley. *Journal of Geophysical Research—Solid Earth*, **105**, 6057–6064.
- WARK, D.A. & MILLER, C.F. 1993. Accessory mineral behavior during differentiation of a granite suite: monazite, xenotime and zircon in the Sweetwater Wash pluton, southeastern California, USA. *Chemical Geology*, **110**, 49–67.
- WARK, D.A. & WATSON, E.B. 1998. Grain-scale permeabilities of texturally equilibrated, monomineralic rocks. *Earth and Planetary Science Letters*, **164**, 591–605.
- WATERS, D.J. 1988. Partial melting and the formation of granulite facies assemblages in Namaqualand, South Africa. *Journal of Metamorphic Geology*, **6**, 387–404.
- WATSON, E.B. 1996. Dissolution, growth and survival of zircons during crustal fusion: kinetic principles, geological models and implications for isotopic inheritance. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **87**, 43–56.
- WATSON, E.B. 1999. Lithologic partitioning of fluids and melts. *American Mineralogist*, **84**, 1693–1710.
- WATSON, E.B. & HARRISON, T.M. 1983. Zircon saturation revisited—temperature and composition effects in a variety of crustal magma types. *Earth and Planetary Science Letters*, **64**, 295–304.
- WATT, G.R., OLIVER, N.H.S. & GRIFFIN, B.J. 2000. Evidence for reaction-induced microfracturing in granulite facies pelitic migmatites. *Geology*, **28**, 331–334.
- WEINBERG, R.F. 1999. Mesoscale pervasive felsic magma migration: alternatives to dyking. *Lithos*, **46**, 393–410.
- WEINBERG, R.F. & PODLADCHIKOV, Y. 1994. Diapiric ascent of magmas through power-law crust and mantle. *Journal of Geophysical Research—Solid Earth*, **99**, 9543–9559.
- WEINBERG, R.F. & SEARLE, M.P. 1998. The Pangong injection complex, Indian Karakoram: a case of pervasive granite flow through hot viscous crust. *Journal of the Geological Society, London*, **155**, 883–891.
- WEINBERG, R.F., SIAL, A. & MARIANO, G. 2004. Close spatial relationship between plutons and shear zones. *Geology*, **32**, 377–380.
- WHITE, R.W. & POWELL, R. 2002. Melt loss and the preservation of granulite facies mineral assemblages. *Journal of Metamorphic Geology*, **20**, 621–632.
- WHITE, R.W., POWELL, R. & HOLLAND, T.J.B. 2001. Calculation of partial melting equilibria in the system Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O (NCKFMASH). *Journal of Metamorphic Geology*, **19**, 139–153.
- WHITE, R.W., POWELL, R. & HALPIN, J.A. 2004. Spatially-focused melt formation in aluminous metapelites from Broken Hill, Australia. *Journal of Metamorphic Geology*, **22**, 825–845.
- WHITE, R.W., POMROY, N.E. & POWELL, R. 2005. An *in situ* metatexite–diatexite transition in upper amphibolite facies rocks from Broken Hill, Australia. *Journal of Metamorphic Geology*, **23**, 579–602.
- WHITNEY, D.L., TEYSSIER, C. & VANDERHAEGHE, O. 2004. Gneiss domes and crustal flow. In: WHITNEY, D.L., TEYSSIER, C.T. & SIDDOWAY, C. (eds) *Gneiss Domes in Orogeny*. Geological Society of America, Special Papers, **380**, 15–33.
- WHITTINGTON, A.G. & TRELOAR, P.J. 2002. Crustal anatexis and its relation to the exhumation of collisional orogenic belts, with particular reference to the Himalaya. *Mineralogical Magazine*, **66**, 53–91.
- WILLIAMS, I.S. 2001. Response of detrital zircon and monazite, and their U–Pb isotopic systems, to regional metamorphism and host-rock partial melting, Cooma Complex, southeastern Australia. *Australian Journal of Earth Sciences*, **48**, 557–580.
- XIA, Q.-K., YANG, X.-Z., DELOULE, E., SHENG, Y.-M. & HAO, Y.-T. 2006. Water in the lower crustal granulite xenoliths from Nushan, eastern China. *Journal of Geophysical Research*, **111**, B11202, doi:10.1029/2006JB004296.
- YARDLEY, B.W.D. & VALLEY, J.W. 1997. The petrologic case for a dry lower crust. *Journal of Geophysical Research*, **102**, 12173–12185.
- ZENG, L., ASIMOW, P.D. & SALEEBY, J.B. 2005a. Coupling of anatectic reactions and dissolution of accessory phases and the Sr and Nd isotope systematics of anatectic melts from a metasedimentary source. *Geochimica et Cosmochimica Acta*, **69**, 3671–3682.
- ZENG, L., SALEEBY, J.B. & ASIMOW, P. 2005b. Nd isotope disequilibrium during crustal anatexis; a record from the Goat Ranch migmatite complex, southern Sierra Nevada Batholith, California. *Geology*, **33**, 53–56.
- ZENG, L., SALEEBY, J.B. & DUCEA, M. 2005c. Geochemical characteristics of crustal anatexis during the formation of migmatite at the Southern Sierra Nevada, California. *Contributions to Mineralogy and Petrology*, **150**, 386–402, doi:10.1007/s00410-005-0010-2.

Received 8 November 2006; revised typescript accepted 9 January 2007.

Scientific editing by Rob Strachan