The spatial and temporal patterning of the deep crust and implications for the process of melt extraction

BY MICHAEL BROWN*

Laboratory for Crustal Petrology, Department of Geology, University of Maryland, College Park, MD 20742-4211, USA

Volumetrically significant melt production requires crustal temperatures above approximately 800°C. At the grain scale, the former presence of melt may be inferred based on various microstructures, particularly pseudomorphs of melt pores and grain-boundary melt films. In residual migmatites and granulites, evidence of melt-extraction pathways at outcrop scale is recorded by crystallized products of melt (leucosome) and residual material from which melt has drained (melanosome). These features form networks or arrays that potentially demonstrate the temporal and spatial relations between deformation and melting. As melt volume increases at sites of initial melting, the feedback between deformation and melting creates a dynamic rheological environment owing to localization and strain-rate weakening. With increasing temperature, melt volume increases to the melt connectivity transition, in the range of 2–7 vol% melt, at which point melt may escape in the first of several melt-loss events, where each event represents a batch of melt that left the source and ascended higher in the crust. Each contributing process has characteristic length and time scales, and it is the nonlinear interactions and feedback relations among them that give rise to the dissipative structures and episodicity of melt-extraction events that are recorded as variations in the spatial and temporal patterning of the crust. Focused melt flow occurs by dilatant shear failure of low-melt fraction rocks creating melt-flow networks that allow accumulation and storage of melt, and form the link for melt flow from grain boundaries to veins allowing drainage to crustal-scale ascent conduits. Preliminary indications suggest that anatectic systems are strongly self-organized from the bottom up, becoming more ordered by decreasing the number and increasing the width of ascent conduits from the anatectic zone through the overlying subsolidus crust to the ductile-to-brittle transition zone, where the melt accumulates in plutons.

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*mbrown@umd.edu

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1. Introduction

Earth is a self-organized system with an irreversible evolution dissipating energy to space. Mantle convection and differentiation of the planet are emergent features of this secular evolution (e.g. O’Neill et al. 2007; Sizova et al. in press). Early differentiation of Earth included generation of Hadean crust that subsequently has been largely recycled back into the mantle (Armstrong 1991). Since the Hadean eon, major additions to the continental crust from the mantle have occurred at ca 3.3 Ga and ca 1.9 Ga (Hawkesworth & Kemp 2006), otherwise the generation and recycling of continental crust at convergent plate margins have been approximately in balance since the Neoarchean Era (Scholl & von Huene 2007). Thus, since the Neoarchean Era, the composition and structure of much of the continental crust has evolved mostly by: (i) magmatic additions and subduction-related losses at accretionary convergent plate margins (e.g. Hawkesworth & Kemp 2006; Scholl & von Huene 2007; Niu & O’Hara 2009) and (ii) tectonometamorphic reworking involving deformation and intracrustal differentiation by melting at accretionary convergent plate margins and those involving collisions between arcs and continents (e.g. Brown & Rushmer 2006; Brown 2007b).

Geophysical surveys confirm the view from geological mapping that granites are concentrated in the upper part of the continental crust, whereas petrological and geochemical studies of outcrops and xenoliths demonstrate that the source for these granites was residual supracrustal rocks, orthogneisses, amphibolites and granulites in the lower part of the continental crust. Thus, extraction of melt from the lower crust and its transfer to the upper crust is the principal process by which continents have become 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 irreversible process stabilizes continental crust, because the heat-producing elements become concentrated in the less dense, granite-dominated upper portion, whereas the lower portion becomes depleted of a granite component and the heat-producing elements, making it stronger and denser (Sandiford & McLaren 2006).

The formation of upper crustal plutons requires that melt separate from solid residue within lower crustal melt-bearing sources and become focused into high-permeability ascent conduits that feed the roots of the plutons. The dissipation of energy during the focusing process is because of mechanical, thermal and chemical processes (Hobbs & Ord 2010). There are differences in the importance of coupling among these processes at different length scales. Chemical–mechanical processes dominate at the grain scale, whereas thermal–mechanical processes dominate at the crustal scale. Nonetheless, these different coupled processes are argued to be responsible for the development of similar structures at all scales, leading to the scale-invariance observed in the structure of the continental crust (Hobbs et al. 2010; Regenauer-Lieb et al. 2009).

In orogenic belts, however, melt flow at the grain-to-vein scale is commonly controlled by the developing anisotropy owing to accumulating strain (e.g. Brown 2004). In transpressional orogens characterized by steep anisotropies, melt extraction, ascent and emplacement nonetheless have been argued to be scale-invariant (e.g. Brown & Solar 1998a, b, 1999; Solar & Brown 2001a, b). In contrast, in other orogenic belts, for example those undergoing orogenic collapse,
it is necessary for the melt to be drained along shallowly oriented flow paths to reach discordant high-permeability ascent conduits; in these examples, the ascent conduits commonly exhibit scale-invariance (Brown 2005), but this may not apply at the scale of the anisotropy (e.g. Marchildon & Brown 2003).

My aim in this contribution is to characterize and interpret the spatial and temporal patterning imprinted on the residual deep (middle-to-lower) crust by the process of melt extraction. I will accomplish this by bringing together observations and quantitative data from grain to crustal scale, and by considering different points of view about how we interpret these data. This synthesis will inform future field, experimental and modelling studies to assist us in achieving our ambition to understand how anatectic crust organizes to enable the extraction of large volumes of melt. I will begin this contribution by briefly considering the view from the top, as represented by granites in the shallow (middle-to-upper) crust. In the main sections of this contribution, I will evaluate the process of melting and its consequences, and describe and interpret the spatial and temporal patterning of crystallized products of melt (leucosome) and residual material from which melt has drained (melanosome), features that form networks or arrays that are preserved in the residual middle-to-lower crust and that record evidence of melt flow and melt loss from the residual middle-to-lower crust. I will finish this contribution by considering pervasive melt flow through the anatectic zone and the episodicity of melt-extraction events.

2. The view from the top down: granites in the middle-to-upper crust and implications for source processes

The size, shape and spatial distribution of plutons in the middle-to-upper crust reflect melting processes in the middle-to-lower crust, and the size–frequency distributions and spacing of plutons in several studies show power-law distributions that have been used to argue that magmatic systems are self-organized from the bottom up (Bons & Elburg 2001; Cruden & McCaffrey 2001; Cruden 2006; Koukouvelas et al. 2006). This view is consistent with studies of relict anatectic systems (e.g. Brown & Solar 1998a; Brown 2005) and results from modelling (Petford & Koenders 1998; Vigneresse & Burg 2000; Bons & van Milligen 2001; Ablay et al. 2008; Hobbs & Ord 2010) that suggest melt extraction may be a self-organized critical phenomenon.

(a) Characteristics of granite plutons

There is a systematic variation in the three-dimensional shape of granite plutons with increasing depth, from horizontal wedge or tabular to blobby to vertical lozenge (e.g. Brown 2001a,b; 2007b), which suggests that emplacement varies according to changing host-rock behaviour with depth (a function of temperature, deformation mechanisms, strain rate, etc.). Melt ascent generally slows at a crustal level, corresponding to the ductile-to-brittle transition zone (e.g. Brown & Solar 1998b; Vigneresse 1999). Here, emplacement occurs when mainly vertical flow changes to predominantly horizontal flow, which is a function of the buoyancy of the magma and the balance between the internal magma pressure, the lithostatic pressure owing to the overlying crust and the horizontal stresses. Analysis of the length versus thickness of horizontal tabular plutons suggests
that they inflate according to a power-law relationship, interpreted to mean that vertical thickening only occurs after magma has travelled horizontally some critical minimum distance (McCaffrey & Petford 1997; Cruden & McCaffrey 2001, 2002; McCaffrey & Cruden 2002, 2003). Inflation is accommodated by depression of the floor and/or lifting of the roof (Cruden 1998, 2006).

In the ductile regime, instabilities in the system likely causes 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 state of stress in host rock (Brown 2001a,b, 2007b). These instabilities are not mutually exclusive. Consider an ascent conduit with zones of higher permeability; these zones have higher magma flux, which heats and weakens the host rock adjacent to these zones, which in turn increases the strain rate marginal to these zones and so on. Magma in the preferred zones of the ascent conduit will exploit the weakening and may expand into the host rock, switching ascent to emplacement and forming a vertical lozenge pluton. A similar phenomenon occurs if the instability is because of differences in the strength of the host rock or to variations in the stress field along the margins of the ascent conduit. Magma exploits the weaker or lower stress sectors, swelling out of the ascent column into those sites, heating and weakening the host rocks, which enables further lateral expansion, forming a blobby pluton (e.g. Brown 2001a,b; Weinberg et al. 2004).

At crustal levels above the zone of anatexis, differences in the rate of flow of the melt or shape of the ascent conduit may affect the critical width for flow to occur without freezing. If freezing occurs in the slower/narrower parts of the conduit, melt flow will focus in the faster/wider parts. Heating of the host rock adjacent to the faster/wider parts will cause weakening that will facilitate swelling of the conduit and slow the rate of flow. At the ductile-to-brittle transition zone, the switch to emplacement may lead to the formation of a pluton with a horizontal tabular head, formed by inflation of a subhorizontal magma fracture, and a hemi-ellipsoidal root or feeder representing the top of the ascent conduit that passes down into a migmatite zone through which magma was transferred from the source (e.g. Brown & Solar 1998b, 1999; Vigneresse 1999).

(b) The length scales and time scales of melt flow, and the volumes of melt involved in inflating plutons

Melting occurs along grain boundaries in a fertile source (less than a cubic millimetre) during a metamorphic cycle that may take several millions to several tens of millions of years (10^6–10^7 years; e.g. Hermann & Rubatto 2003; Kylander-Clark et al. 2007; Reno et al. 2009). In contrast, a pluton represents a large volume of magma (10^3–10^4 km^3 or more) aggregated from many batches of melt (each perhaps 10^{-1} to 10^2 km^3) that crystallized during tens of thousands to several millions of years (10^4–10^7 years; Brown 2001b; Coleman et al. 2004; Matzel et al. 2006; Miller 2008). Thus, melt extraction from segregation to emplacement is a process with a length scale that spans more than seven orders of magnitude, or a volume concentration factor that exceeds 10^{21}.

Anatectic melts are undersaturated in light rare-earth elements and zirconium, consistent with fast rates of deformation-assisted segregation, perhaps in the order of 10–10^4 years for a single batch of melt (Sawyer 1991; Harris et al. 2000). Ascent
times for melt in a dyke, likely $\ll 1$ year for an ascent rate of $10^{-2}$ to $10^{-1}$ m s$^{-1}$ (Petford et al. 1993; Clemens 1998), suggest that a small dyke 1 km in length and 3 m wide could transfer $10^3$ km$^3$ of melt from source to sink in ca. $10^3$ years. Thus, melt extraction from segregation to emplacement is a process with a time scale that spans more than seven orders of magnitude (Brown 2004).

There are several studies that have made estimates of the volumes of melt generated from source terranes in different tectonic settings. One study from an area of $1.5 \times 10^5$ km$^2$ in the Lachlan Fold Belt of southeastern Australia—an accretionary orogenic belt that developed above a subducting ocean plate at a convergent plate margin (e.g. Collins & Richards 2008)—estimates that about $10^3$ plutons with a volume greater than 1 km$^3$ were emplaced during Siluro-Devonian orogenic events, giving a total volume of melt transferred to the middle-to-upper crust of $1.5 \times 10^5$ km$^3$ (Bons & Elbureg 2001). This distribution is consistent with an average melt loss of approximately 10 vol% from a source of a similar horizontal area or footprint that was approximately 10 km thick (e.g. Zen 1992); such an estimate ignores any mass input owing to melting of asthenosphere in the mantle wedge, which not only will reduce the proportion of lower crustal melt required, but also is the likely driver of crustal differentiation in arc-related settings (e.g. Kemp et al. 2007).

In another study from three successive highly focused, linear plutonic suites in an area of about $10^4$ km$^2$ in the Cretaceous Central Sierra Nevada batholith in California, which also developed above a subducting ocean plate at a convergent plate margin, yields magma volumes per unit area that are up to four times larger than those in the Lachlan Fold Belt (Cruden 2006). The difference may reflect a higher proportion of mass additions from the mantle and/or a higher degree of crustal melting in the formation of the Sierra Nevada batholith, which in turn may relate to plate kinematics and/or rates of subduction.

In contrast to these simple accretionary orogenic systems at convergent plate margins along the continental edge, the Neoarchean Superior Province in Canada comprises two elements: (i) the western Superior Province, which is a transpressive accretionary orogen involving the amalgamation of continental fragments, arcs and oceanic terranes at a long-lived convergent plate margin (Percival et al. 2006) and (ii) the Ashuanipi Subprovince, which is an exposed orogenic hinterland perhaps similar to deeper levels of the crust under the Tibet Plateau. In the Ashuanipi Subprovince, over an area of $9 \times 10^4$ km$^2$, approximately 30 vol% melt is inferred to have been extracted on average from residual granulites derived from greywacke protoliths at temperatures up to 900°C at depths equivalent to pressures of 0.6–0.7 GPa (Guernina & Sawyer 2003). Extrapolation to the full extent of the Ashuanipi Subprovince suggests that approximately $6.5 \times 10^5$ km$^3$ of melt could have been extracted from granulate facies middle-to-lower crust; this would represent an equivalent ‘layer’ of granite some 7 km thick (Guernina & Sawyer 2003).

An alternative calculation for the whole of the Superior Province, including the western Superior Province, based on the map distribution of plutons that form the late Neoarchean pan-Superior granodiorite–granite suite, yields an average estimate of $2–6 \times 10^6$ km$^3$ of melt generated across the Province, which would represent a ‘layer’ of granite ‘only’ 1–3 km thick (Cruden 2006). There is a larger area (proxy for volume) of granodiorite–granite suite plutons exposed in the Ashuanipi Subprovince in comparison with the Superior Province as a whole,
which may indicate generation of a larger amount of melt by higher degrees of melting in the source under that Subprovince. Such an interpretation is consistent with the strongly depleted nature of the associated residual granulites exposed at the surface in the Ashuanipi Subprovince (Guernina & Sawyer 2003). These melt volumes are large and reflect massive differentiation of the continental crust during Neoarchean cratonization.

The focused nature of magmatic events in the Central Sierra Nevada batholith and the predominantly tabular form of the plutons suggests higher degrees of melting and larger magma fluxes here than in the Superior Province (Cruden 2006). In contrast, the defocused nature of the granodiorite–granite magmatism in the Superior Province and the predominantly wedge-shaped form of the plutons are argued to be more consistent with generally moderate degrees of melting (Cruden 2006). These differences may reflect the long-lived continental margin setting of the Central Sierra Nevada batholith. In this setting, mantle wedge-derived magmas advect heat into the lower crust, promoting overall higher degrees of melting (Cruden 2006) and hybridization with lower crustal melts (Kemp et al. 2007), particularly during high-flux pulses of magmatism when the arc was in a ‘flare-up’ state (Ducea & Barton 2007).

In comparison, the evolution of the western Superior Province and the Ashuanipi Subprovince as a transpressive accretionary orogen system and orogenic hinterland, respectively, may be responsible for the defocused nature of the magmatism throughout much of the Superior Province. The higher volume of melt lost from the Ashuanipi Subprovince and the uniform pressure and similarly depleted nature of the residual granulites (Guernina & Sawyer 2003) are consistent with an interpretation that this region represents the exposed middle-to-lower crust from under a Neoarchean orogenic plateau, rather than an event related to ridge subduction, as has been proposed previously (Percival et al. 2003).

The problem we are trying to understand is defined by the range of length scales of anatectic systems, the short time scales for extraction events and the large volumes of melt lost from deep crustal sources and emplaced in shallow crustal plutons. It resolves into two simple questions: how is melt extraction from the source focused and why does melt ascent switch to emplacement? The issues of scale are further compounded by the multi-phase (solid plus melt) nature of the source material, the multiple scales of flow implicit in moving melt from grain boundaries to crustal scale ascent conduits, and the three-dimensional nature of the deep crustal sources. It is argued, in this contribution, that evidence for the focusing mechanisms that enable melt extraction is preserved in exposed residual deep crustal sources, and the remainder of this contribution will address this evidence and its interpretation. The switch from melt ascent to emplacement is largely outside the scope of this contribution.

3. Melting

Crustal melting, which is fundamentally controlled by the available heat, occurs via a sequence of three types of multi-variant reactions (figure 1; Brown & Korhonen 2009): fluid-present (‘wet’) melting at the solidus (the anatectic front); hydrate mineral-breakdown melting at moderate-to-high temperatures; and anhydrous melting at ultrahigh temperatures. As a multi-variant process,
crustal melting occurs over an interval of temperature, and because it is strongly endothermic, the rate of temperature increase during heating is buffered by the melting and may be self-limiting.

(a) The melting process and its consequences

Common crustal rocks may melt at temperatures as low as 650°C if H₂O-rich fluid is available by generalized reactions such as quartz + plagioclase + K-feldspar + H₂O ↔ melt and quartz + plagioclase + biotite + H₂O ↔ melt + amphibole + titanite (‘wet’ melting at the solidus; figure 1). However, ‘wet’ melting is generally limited by the small amount of aqueous phase present immediately below the solidus (<1 vol%, unless there is influx of H₂O-rich fluid), which is immediately dissolved in the incipient melt. As a result of this limitation, it is fluid-absent hydrate mineral-breakdown melting at moderate-to-high temperatures that is responsible for the production of most crustally derived granites (figure 1; Brown & Korhonen 2009; see also Clemens 2006). Generalized examples of these reactions are quartz + plagioclase + muscovite ↔ melt + K-feldspar + aluminosilicate + biotite, quartz + plagioclase + biotite + aluminosilicate ↔ melt + K-feldspar + garnet and quartz + plagioclase + biotite ↔ melt + orthopyroxene + ilmenite.

Fluid infiltration to promote extensive wet melting may be recognized by an inconsistency between the observed volumes of (apparently) locally derived melt now crystallized as leucosome and the maximum expected for the calculated peak P–T conditions assuming no melt loss. Local introduction of water may occur in the inner zone of contact aureoles around granites and mafic intrusions (e.g. Pattison & Harte 1988; Symmes & Ferry 1995; Johnson et al. 2003), and water may infiltrate from adjacent units that have a higher temperature solidus, or along pervasive dilatant fractures, or in shear zones, or perhaps regionally where there is generally strong focusing into specific layers (e.g. Sibson 1986; Wickham & Taylor 1987; McCaig et al. 1990; Upton et al. 1995; Selbekk et al. 2000; Johnson et al. 2001; Connolly & Podladchikov 2004; Slagstad et al. 2005; White et al. 2005; Berger et al. 2008; Florian et al. 2008; Ward et al. 2008; Rubatto et al. in press). Given the limited porosity of high-grade subsolidus crust (Yardley 2009), deformation-enhanced permeability is likely required to facilitate fluid infiltration (e.g. Upton et al. 1995; Connolly & Podladchikov 2004), and the same limitation applies to pervasive melt migration, which likely only occurs in association with strong deformation (Hasalová et al. 2008a,b). Zones of regional-scale transpressive deformation may be particularly appropriate to promote fluid-enhanced melting and melt flow; examples include the St Malo migmatite belt, France (Brown 1995; Milord et al. 2001) and the Tumbledown and Weld anatectic domains in Maine, USA (Brown & Solar 1999; Solar & Brown 2001a,b).

In addition, in convergent tectonic settings in general, relaxation of the yield stress developed in the brittle portion of the crust may result in a depth interval below the brittle-to-ductile transition zone characterized by an inverted pressure gradient (e.g. Stüwe et al. 1993; Stüwe & Sandiford 1994; Petrini & Podladchikov 2000). At depths where this inverted pressure gradient is less than the hydrostatic gradient of an interstitial fluid, and if the fluid is subject to rock confining pressure, then the fluid will migrate downward and stagnate where the rock pressure gradient becomes identical to the hydrostatic fluid gradient.
(Connolly & Podladchikov 2004). This condition defines a depth of tectonically induced neutral buoyancy that also acts as a barrier to upward fluid flow. In combination with dynamic downward propagation of the brittle-to-ductile transition zone during orogenic thickening, this phenomenon provides a mechanism to sweep upper crustal fluids into the lower crust to promote melting as prograde heating evolves to peak temperatures.

In the absence of fluid infiltration, the amount of H$_2$O-undersaturated melt produced by hydrate-breakdown melting (referred to as the fertility) varies according to protolith composition and mineralogy, particularly the hydrate species and mode (vol% of the hydrate species present), the depth of melting (pressure) and heat input (temperature), the length of time at temperatures above the solidus and the maximum temperature achieved. The continental crust comprises around 65 vol% igneous rock compositions, 27 vol% metasedimentary rock compositions and 8 vol% sedimentary rock compositions. The fertility of common crustal compositions varies from pelite and greywacke, with 50–20 vol% muscovite and biotite yielding up to 65 mol% melt, to granite and leucogranite,
Figure 1. (Opposite.) Melting relations for pelite and greywacke calculated in the chemical system Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O–TiO₂–Fe₂O₃ (see Brown & Korhonen (2009) for details of the calculation). (a) P–T pseudosection (isochemical phase diagram) calculated for an average amphibolite facies pelite composition (see Brown & Korhonen (2009) for details and the full pseudosection). This diagram is drawn for a fixed composition, which means that processes such as melt loss require calculation of a new pseudosection for the residual composition after an imposed melt-loss event. A melt-loss event is imposed along a prograde P–T path that crosses the biotite-out line at 0.9 GPa (black circle labelled ‘ML’); since complete melt loss does not occur in nature, 1 mol% melt is retained. The pseudosection is contoured for melt mol% up to the biotite-out line for a rock saturated in H₂O at 0.9 GPa at the solidus. Owing to the different amount of H₂O required to saturate the rock in a H₂O fluid at different pressures along the solidus, the amount of fluid present at the low-P segment of the solidus is an overestimation, and consequently the amount of melt produced at low pressures is overestimated. The dashed lines represent contours of mol% melt. (b) P–T pseudosection calculated for the residual composition derived from the average amphibolite facies pelite composition in (a) after a melt-loss event at the biotite-out line at 0.9 GPa. The pseudosection is contoured for melt mol% at supersolidus conditions, and demonstrates that the residual composition is less fertile than the protolith composition. The dashed lines represent contours of mol% melt. For further discussion, see Brown & Korhonen (2009). (c) P–T pseudosection calculated for a greywacke composition (see Brown & Korhonen (2009) for details and the full pseudosection). This diagram is drawn for a fixed composition, which means that processes such as melt loss require calculation of a new pseudosection for the residual composition after an imposed melt-loss event. A melt-loss event is imposed along a prograde P–T path that crosses the biotite-out line at 0.9 GPa (black circle labelled ‘ML’); since complete melt loss does not occur in nature, 1 mol% melt is retained. The pseudosection is contoured for melt mol% up to the biotite-out line for a rock saturated in H₂O at 0.9 GPa at the wet solidus. Because of the different amount of H₂O required to saturate the rock in a H₂O fluid at different pressures along the wet solidus, the amount of fluid present at the low-P segment of the solidus is an overestimation, and consequently the amount of melt produced at low pressures is overestimated. The dashed lines represent contours of mol% melt. (d) P–T pseudosection calculated for the residual composition derived from the greywacke composition after a melt-loss event at the biotite-out line at 0.9 GPa. The pseudosection is contoured for melt mol% at supersolidus conditions, and demonstrates that the residual composition is less fertile than the protolith composition. The dashed lines represent contours of mol% melt. For further discussion, see Brown & Korhonen (2009).

with 25–5 vol% muscovite and biotite yielding less than 35 mol% melt. Once the hydrate phase is exhausted, assuming temperature continues to rise, any melt that was not extracted by this point becomes progressively drier as a high proportion of the anhydrous phases dissolve with increasing temperature (anhydrous melting at ultrahigh temperatures; see figure 1 and Brown & Korhonen 2009).

The anatectic zone lies beneath the anatectic front at the top, defined by the H₂O-saturated solidus (separating melt-bearing crust below from sub-solidus crust above; figure 2)—a spatially and temporally mobile boundary owing to the dynamic nature of orogenesis and feedback relations between deformation, the geotherm and advection of melt—and extends to a transition at the bottom as the rate of melt production drops across the hydrate-out surface, assuming melt loss from the zone (figure 1). In general, melting affects metasedimentary and granitic protolith compositions at lower temperature and intermediate and mafic protoliths at higher temperatures. For this reason, with increasing temperature in an evolving orogen, the highest grade crust progressively encounters muscovite and biotite breakdown in metasedimentary and (leuco-) granite protoliths,
Figure 2. Schematic diagrams to illustrate the thermal structure and migmatite–pluton relations in a dextral transpressive orogenic system based on the western Maine area in the Acadian orogen of the northern Appalachians (Brown & Solar 1998b, 1999; Solar & Brown 2001a,b; Tomascak et al. 2005). (a) Schematic structure section drawn perpendicular to foliation. The anatectic front is recorded in orogenic crust by the first appearance of migmatite—corresponding to the final solidus. The anatectic front, which tracks the solidus, was progressively extended into shallower parts of the orogenic system by advection of material during contractional thickening, including the sequential ascent of granite melt (e.g. Brown & Solar 1999). The three-dimensional form of the final solidus surface projected onto the section illustrates the thermal structure at the peak of orogenesis (continuous line labelled ‘solidus’). A H$_2$O-rich volatile phase is exsolved at the solidus as granite crystallizes. This aqueous fluid is postulated to be responsible for the widespread generation of retrograde muscovite in migmatites. Dashed lines are boundaries between structural domains (ACZ, zones of apparent constrictional strain in between zones of apparent flattening strain, identified by dashed ornament); Ms, muscovite; Bt, biotite; BHB rocks, Bronson Hill belt rocks; CMB rocks, central Maine belt rocks; TAD and WAD, Tumbledown and Weld anatectic domains, respectively; P, the Phillips pluton. This figure is modified from Solar & Brown (2001a) and is used in accordance with the publications rights policies of Oxford University Press. (b) Schematic north-northwest–south-southeast model section to show the immediately post-thermal peak stage of evolution in a transpressive system, based on the model for the structural evolution of the western Maine area by Solar & Brown (2001b); the form of the granites, which are projected east-northeast onto the plane of the section, is based on information in Brown & Solar (1998b, 1999). Notice that the granite plutons are rooted in the migmatites, and melt flow is interpreted to be upward along the fabrics (Brown & Solar 1998a). The level of horizontal expansion and emplacement of the Lexington (L) and Kingsman (Ki) plutons, as exposed, is interpreted to correspond approximately to the contemporary brittle–ductile transition zone. At a deeper level, the Lexington pluton has a similar form to the Phillips (P) pluton (see (a)), which is interpreted to be the root to a formerly more extensive pluton that has been eroded.
biotite with or without hornblende breakdown in quartzo-feldspathic (e.g. calc-alkaline) plutonic rocks and hornblende breakdown in amphibolites (e.g. Clemens 2006; Johnson et al. 2008; White 2008; Brown & Korhonen 2009).

In summary, sufficient magma to produce the inferred volume of middle-to-upper crustal granite in orogens requires source temperatures above approximately $800^\circ C$. Common crustal protoliths may yield 10–50 vol% of $H_2O$-undersaturated leucogranite, trondhjemite, granodiorite or tonalite melt at temperatures of up to approximately $1000^\circ C$ (Clemens 2006) that are now considered attainable during orogenesis (Brown 2007a).

(b) From the initiation of melting to the melt connectivity transition

In simple melting experiments in the laboratory, as temperature is increased under conditions of hydrostatic stress, the volume of melt increases and melt pores enlarge and coalesce to form an interconnected grain-boundary network (Laporte et al. 1997). In nature, under equilibrium conditions in an isotropic crust, melting will begin at multi-phase grain junctions that include quartz and feldspar, and commonly a hydrate mineral. However, the Earth’s crust is anisotropic and in a state of non-hydrostatic stress. There are variations in bulk composition, mineralogy and mode, and grain size, and the differential stress varies within any particular volume of crust. As a result, once the initial thermal overstep is close to that required to overcome the activation energy for the particular melting reaction, melting may begin at locations where the heat generated by viscous dissipation is higher or along grain boundaries experiencing lower or higher resolved normal stress.

For wet melting, where the solidus has a neutral to negative $dP/dT$ (figure 1), the start of reaction may occur at grain boundaries experiencing higher resolved normal stress, particularly at lower pressures where the solidus has a strong negative slope, whereas for hydrate-breakdown melting, where the solidus has a positive $dP/dT$ (figure 1), the start of reaction is more likely at grain boundaries experiencing lower resolved normal stress (Brown & Solar 1998a). In addition, rock-forming minerals show some degree of anisotropy of surface energies, especially those hydrate phases that are likely to dominate melting and melt geometry, which may constrain initial melt pores to be aligned with the hydrate grains parallel to foliation and/or lineation (cf. Sawyer 2001). At low melt fractions, intergranular pores also may develop by a cavitation-driven dilation process during melt-enhanced grain-boundary sliding that is not fully compensated by changes in grain shape owing to diffusion and/or dislocation creep; because the viscosity of the silicate melt derived from typical crustal protoliths is sufficiently high to prevent relaxation of the local stresses, this process generates local under-pressure that attracts melt from over-pressured sites (Závada et al. 2007).

Initial melt pores may be over- or under-pressured, and may contract or expand, dynamically modifying melt–solid dihedral angles, and lose or gain melt, according to whether or not individual pores fail by loss of cohesion, leading to coalescence of pores and crack propagation by ductile fracture (Brown 2004; Eichhubl 2004). Within the first few per cent of melt production, the melt reaches the melt connectivity transition in all common crustal protoliths (Laporte et al. 1997; Rosenberg & Handy 2005), although not necessarily at a common temperature (for a constant pressure or depth) and, therefore, not at the same
time. The development of permeability enables advective melt flow from grain boundaries to veins, assuming dilatancy collapse by shear-enhanced compaction of the matrix (Rutter & Mecklenburgh 2006); this may occur earlier in some horizons than in others.

Melt segregation is a consequence of the relative motion between melt and solids. Although the density difference between melt and residue provides a gravitational potential for the separation of melt by porous flow, the viscosity of the silicate melt may be too high to allow gravity-driven homogeneous porous flow to segregate melt in geologically reasonable periods of time (Wickham 1987; Brown et al. 1995; Rutter & Mecklenburgh 2006), except possibly for basaltic protoliths (Jackson et al. 2003, 2005). However, local gradients in differential stress generate gradients in pore fluid pressure that may be hundreds to thousands of times greater than the gradients owing to the gravitational potential. These forces are sufficient to drive melt from grain-boundary pores to a dynamic network of deformation bands, the scale of which is controlled by compositional layering, fabric and the compaction length (Brown et al. 1995; Brown 2004, 2008; Rabinowicz & Vigneresse 2004; Rutter & Mecklenburgh 2006).

For isotropic crust, the combination of compaction and shear channelling instabilities may segregate melt into parallel veins spaced according to the compaction length that, for an appropriate viscosity contrast, is likely to be decimetric (Rabinowicz & Vigneresse 2004). In contrast, in metamorphic tectonites, the intrinsic anisotropy of permeability due to fabrics facilitates foliation- and/or lineation-parallel growth of melt concentrations. These structures grow to form melt-rich layers or rods (Brown et al. 1999)—melt flows down gradients in pore fluid pressure to be concentrated in these sites—enhancing the ability of melt-rich structures to accommodate strain, and creating a dynamic feedback relation between melting and deformation that enables melt flow (Brown & Solar 1998a).

Commonly, the solid products of hydrate-breakdown reactions (e.g. orthopyroxene, garnet and cordierite in metasedimentary protoliths, and pyroxene in basalt protoliths) have difficulty nucleating. Thus, it is energetically favourable for melting to continue at initially established sites and for the solid products to continue to grow at these sites by a diffusive melting process (Waters 1988; Jones & Brown 1990; Powell & Downes 1990; Brown & Dallmeyer 1996; Brown 2004; White et al. 2004) until the volume of melt generated is sufficient to enable advective flow. In a closed system, cooling potentially will lead to retrogression by reaction between melt and/or exsolved water and residue, unless segregation has separated them sufficiently (Brown 2002). The common occurrence of pristine or only weakly retrogressed residual granulites suggests that melt has been lost from these rocks (e.g. White & Powell 2002; Guernina & Sawyer 2003), taking with it the dissolved H2O, consistent with the occurrence of granites in the upper crust.

(c) Self-organized criticality

The point where a change occurs in the behaviour or structure of a system is called a critical point. In standard critical phenomena such as melting, there is a control parameter, generally temperature in the case of melting, which may be varied to cause the change; for example, increasing temperature across the

Phil. Trans. R. Soc. A (2010)
solidus causes melting. In contrast, self-organized criticality is a property of slowly driven non-equilibrium systems with extended degrees of freedom and strong nonlinearity; such systems exhibit spatial and temporal scale-invariance, and reach a critical state owing to 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. For comparison, self-similar behaviour in space and time has been documented for hot-spot volcanism (Shaw & Chouet 1991), for basalt eruptions and basalt fields in the North American Cordillera (Pelletier 1999), and for melt flow through the shallow mantle beneath ocean spreading centres, as recorded by dunite-filled conduits in mantle sections of ophiolites (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; Hobbs & Ord 2010).

\((d)\) Feedback relations

Feedback relations lead to acceleration or deceleration of coupled processes. For example, at the scale of an orogen, higher thermal gradients, conductive heating and regional metamorphism are related to crustal thickening, but heating leads to thermal weakening and crustal thinning (for appropriate boundary conditions). Metamorphism may culminate in melting, which further weakens the crust and may lead to localization and accelerated strain rates, but localization and accelerated strain rates drive melt loss from the source and may lead to strengthening of the residual crust. Consequently, melting and melt loss have a profound effect on crustal rheology. In reality, the processes involved are many, and the properties of the crust are sufficiently varied that the feedback relations are complex, as exemplified by the relations shown in figure 3.

Processes during orogenesis and melting are linked by nonlinear functional relations (figure 3), so there are generally no simple links between cause and effect. For example, strain rate is an exponential function of temperature but a cubic function of stress, whereas gravitational potential energy (GPE in figure 3) is a quadratic function of thickness (Stüwe 2007). Hydrate-breakdown and anhydrous melting are nonlinear processes, with a tendency for a larger vol% melt production per unit increment in temperature as the hydrate-out line or the liquidus is approached (figure 1; Brown & Korhonen 2009). With respect to rheology, residual material in the anatectic zone deforms according to a power-law
Figure 3. Interactions and feedback relations between processes during orogenesis and crustal melting (cf. Stüwe 2007). Note that many of the functional relations between processes are nonlinear, which may lead to emergent behaviour and self-organization of these complex systems. In the figure, ‘melt connect. transition’ is an abbreviation of melt connectivity transition (Rosenberg & Handy 2005).

relation (assuming dislocation creep), whereas melt behaves as a Newtonian fluid, so the viscosity contrast between them is strain-rate dependent, which may lead to the development of instabilities in rheology (Vigneresse et al. 2008; Hobbs & Ord 2010). Indeed, the presence of melt in the crust at scales from grain-boundary films to plutons will lead to localization at various scales in the crust (Brown & Solar 1998a; Walte et al. 2005; Rutter & Mecklenburgh 2006). As a result of these nonlinear functional relations, anatectic systems and orogenic systems in general have the potential to change or reorganize in unpredictable ways, for example, in anatectic systems, there is a switch from melt segregation and storage to melt extraction and drainage. Many of these processes interact dynamically at multiple scales, leading to emergent behaviour and self-organization (Brown & Solar 1998a).

(e) Implications of melting for the rheology and tectonics of the continental crust

Continental crust is composed of a range of rock types and compositions that do not melt equally for a particular $P-T$ condition or temperature interval (figure 1a,c). As a result, melt volume may vary from one rock unit
to another, so that some units generate and lose more melt than others, and the residual mineralogy may vary from one unit to another according to composition (Brown & Korhonen 2009). Theory and rock mechanics experiments (Rabinowicz & Vigneresse 2004; Rosenberg & Handy 2005; Rutter et al. 2006) show that the presence of melt significantly lowers rock strength, leading to a profound change in the behaviour of the crust as the bulk viscosity is reduced and the system changes from monophase (solid only) to two-phase (solid and melt). This may lead to increases in strain rate and ultimately the rate of heat transfer by advection as orogenesis and melting progress and melt is transported from the source to the upper crust. Locally, differences in melt volume from one unit to another lead to differences in bulk viscosity, so that each unit may respond differently to stress, and melt will migrate in response to the inevitable gradients in fluid pressure that are generated to accumulate in dilatant sites prior to extraction (e.g. Weinberg et al. 2009).

Melt-bearing horizons in the crust will be mechanically weak zones that are likely to play an important role in continental tectonics. For example, constrained by geophysical data, the amount of melt present under the Andes–Altiplano Plateau and the Himalayas–Tibet Plateau has been modelled as 5–15 vol% (e.g. Schilling & Partzsch 2001; Babeyko et al. 2002; Unsworth et al. 2005), and the presence of this melt is inferred to weaken the middle-to-lower crust allowing it to flow in response to GPE and far-field tectonic stress (e.g. Beaumont et al. 2006; Royden et al. 2008). The weakening, most probably, is because of a change in deformation mechanism from dislocation creep to granular flow (Rosenberg & Handy 2000, 2001; Rosenberg & Berger 2001; Walte et al. 2005; Rutter et al. 2006).

4. Inferences derived from the patterning of residual middle-to-lower crust

What are the challenges in interrogating a relict anatectic system exposed at the Earth’s surface, what information may we expect to obtain and how may we use this information?

(a) Challenges in the interrogation of relict anatectic systems

In the deep crust, there is abundant evidence of melt-bearing networks in residual migmatites and granulites, as recorded by the products crystallized from melt, referred to as leucosome, from which I infer that melt segregates and accumulates in these networks until a drainage event occurs. Within these networks, mesoscale pods of leucosome or leucogranite 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; Opatica Subprovince, Canada—Sawyer 1998; Arunta Inlier, central Australia—Sawyer et al. 1999; Cooma Complex, Australia—Vernon et al. 2003; Ladakh, northwestern India—Weinberg et al. 2009). It is these storage networks, including the locally ponded melt, that fed ascent conduits, as recorded by mesoscale cylindrical and tabular granites. In turn, the ascent conduits moved 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.
This patterning of the lower crust we see at outcrop represents the integrated record of relative motion during two-phase processes and overprinting of multiple melt-extraction events. The spatial and temporal patterning of melt-related structures marked out by products crystallized from melt, referred to as leucosome, residual material that identifies where melt has been lost, referred to as melanosome, and outcrop scale bodies of leucogranite is widely interpreted as registering the movement of melt from sites of incipient melting along grain boundaries to dynamic deformation band networks (melt flow or drainage networks) controlled by the compositional and fabric anisotropy, and from these to steeply inclined fabric concordant or discordant cylindrical and tabular granites that infill inferred ascent conduits (e.g. Brown 1994, 2001a,b, 2004, 2008; Sawyer 1994, 2001; Brown et al. 1995; Brown & Solar 1998a,b; Sawyer et al. 1999; Marchildon & Brown 2002, 2003; Guernina & Sawyer 2003). This interpretation is supported by results from experiments on analogue materials (Rosenberg & Handy 2000, 2001; Barraud et al. 2001, 2004; Walte et al. 2005). Based on these interpretations and experimental results, melt extraction is inferred to be a multi-scale phenomenon with linkages between flow regimes at different scales, viz. pervasive grain-scale flow, porous flow in veins and channel flow in ascent conduits. Furthermore, the patterning implies multiple couplings and feedback relations between chemical and mechanical processes (Brown et al. 1995; Rushmer & Jackson 2008).

Each contributing process has characteristic length and time scales, and it is the nonlinear interactions and feedback relations among them that give rise to the dissipative structures and episodicity of melt-extraction events that are recorded as variations in spatial and temporal patterning of the crust. Neither the spatial nor the temporal record need be complete at any single locality, and we have to distinguish between what was active at a given time and the integrated record. The final patterning and complexity comprise the ‘memory’ of a rock body. The spatial distribution of ‘memory’ is unlikely to be linearly related to time (Bergantz & Barboza 2006), especially where rheological transitions relate to melt volume (Vigneresse et al. 2008), which is controlled by heat input and change in enthalpy.

This is a three-dimensional problem involving open system processes. To illustrate this point, consider migmatite terranes, which are levels of the crust composed of rocks that show evidence of melting, such as the presence of leucosomes commonly with melanosomes. All migmatite terranes are likely to be polygenetic and polychronic because they represent a level in the crust where there is potential for both local losses and additions of melt, as well as for interaction between melt generated in situ with melt from below migrating through the level of exposed crust. As a consequence, the leucosomes in a migmatite, for example, may have liquid compositions, they may have cumulate compositions, or they may have fractionated compositions, and any of these types may occur more than once in any two-dimensional exposure through the terrane. These terranes commonly separate deeper-level, residual granulites that may only preserve minimal leucosome from shallower-level granite-dominated terranes where the melt was emplaced in subsolidus crust.

Examples of these relationships are provided by multiple studies of the continental crust from around the world, and a full inventory is provided in Brown (2008, pp. 124–127). The upper and intermediate crustal levels of such
granite–migmatite–granulite systems are exemplified by the Carboniferous and Cretaceous polyphase granites and migmatites of the Fosdick Mountains in Marie Byrd Land, west Antarctica (Korhonen et al. 2007, 2009; Saito et al. 2007) and the Acadian migmatites and granites of the northern Appalachians of New Hampshire and Maine, USA (Brown & Solar 1998a,b; Solar et al. 1998; Brown & Pressley 1999; Brown et al. 1999; Pressley & Brown 1999; Solar & Brown 2001a; Tomascak et al. 2005), whereas the intermediate and lower levels of these systems are exemplified by the Superior Province, Canada (Cruden 2006), particularly the Opatica (Sawyer 1998) and Ashuanipi (Guernina & Sawyer 2003) Subprovinces.

In migmatite terranes and deeper level residual granulites, the intrinsic hierarchy of size, scale or levels of detail of a geological dataset comprises its granularity. However, the spatial granularity, such as the size, composition and distribution of leucosomes in a migmatite or residual granulite may be different from the temporal granularity, for example, due to overprinting of multiple melt-extraction events, and therefore the definition of proximity in space and time may be different (Bergantz & Barboza 2006). This is implicit in the recognition that intracrustal differentiation is an open system process.

The challenge is to determine whether suites of leucosomes and granites in the relict anatectic system were active at the same time (whether continuously or intermittently) and whether their size reflects some measure of conditions at the time of putative melt transport. That is, we want to know whether the spatial elements of the granularity differ from the temporal elements of the granularity, and whether information in the outcrop is representative of the processes involved. A leucosome, for example, may represent the cumulative record of intermittent melt extraction or even remelting so that the final composition may not have any simple relationship to the protolith composition (e.g. Solar & Brown 2001a,b; Rubatto et al. 2009) and the thickness of a dike while active, for example, may be greater than the measured thickness at outcrop so that our natural spatial data tend to be minima. Granite in ascent conduits and leucosome in drainage networks may be linked by petrographic continuity (mineralogy, grain size and microstructure), implying temporal contemporaneity, or the granite and leucosome may be petrographically distinct, implying a temporal sequence. Given these constraints, what information about the two-phase processes involved in intracrustal differentiation may be derived from relict anatectic systems?

(b) What data do we have and how may we use them?

One challenge is how to recognize the former presence of melt at the grain scale, particularly in regionally metamorphosed rocks. In the absence of overprinting, various characteristic microstructures have been used to infer the former presence of melt at the grain scale in residual migmatites and granulites (e.g. Brown et al. 1999; Sawyer 1999, 2001; Brown 2001a,b; Albertz et al. 2005; Holness & Sawyer 2008).

Melting may be inferred using the following criteria: (i) the presence of mineral pseudomorphs after grain-boundary melt films, particularly those associated with hydrate minerals (e.g. muscovite and biotite), and/or cuspate volumes of quartz, K-feldspar or sodic plagioclase, inferred to be pools of crystallized melt, surrounded by embayed grains or grains with more regular outlines owing to crystal growth from the melt (in contact migmatites (e.g. Harte et al. 1991;
Rosenberg & Riller 2000; Marchildon & Brown 2001, 2002; Holness et al. 2005), in regional migmatites (e.g. Marchildon & Brown 2003) and in granulites (e.g. Sawyer 1999, 2001; Brown 2002; Guernina & Sawyer 2003; Holness & Sawyer 2008), (ii) measured dihedral angles of $\leq 60^\circ$ formed where a grain of feldspar and/or quartz meets two grains of inferred residual minerals (Holness & Clemens 1999; Rosenberg & Riller 2000; Holness & Sawyer 2008), (iii) peritectic products of the melting reaction (e.g. garnet in pelites or orthopyroxene in greywackes) with crystal faces against the inferred melt (Powell & Downes 1990; Brown & Dallmeyer 1996; Sawyer 2001; White et al. 2004), (iv) the presence of magmatic rims on subsolidus cores of grains (e.g. rational faces, overgrowths of different compositions; Marchildon & Brown 2001, 2002; Sawyer 2001) or magmatic microstructures in leucosomes (e.g. Vernon & Collins 1988; Brown 2002), (v) rounded and corroded reactant minerals within or surrounded by minerals or quartzo-feldspathic aggregates inferred to be crystallized from melt (Brown 1998; Sawyer 2001), and (vi) the occurrence of subparallel intragranular fractures (e.g. feldspar-filled fractures in quartz; Rosenberg & Riller 2000) and/or evidence of annealed microfractures (e.g. in quartz; Watt et al. 2000; Marchildon & Brown 2001, 2002).

In deformed anatectic rocks, former melt-bearing microstructures may also be used to evaluate deformation mechanisms and to reconstruct grain-scale melt migration paths because deformation and melt extraction are coupled (e.g. Brown 1994, 2004; Collins & Sawyer 1996; Brown & Rushmer 1997; Marchildon & Brown 2002). In materials that are only weakly anisotropic, melt is distributed along grain boundaries oriented approximately parallel to the shortening direction and is inferred to have been squeezed from grain boundaries approximately parallel to the elongation direction (e.g. Dell’Angelo & Tullis 1988; Gleason et al. 1999; Rosenberg & Handy 2001; Rosenberg & Berger 2001; Sawyer 2001; Walt et al. 2005), compatible with dilation of a granular aggregate owing to melt pressure exceeding the confining pressure (cf. Hubbert & Rubey 1959). In contrast, in more strongly anisotropic rocks, melt is inferred to have been distributed in seams preferentially along grain boundaries parallel to the foliation and/or lineation (e.g. Sawyer 2001), which is compatible with loss of cohesion between grains during the combined operation of grain-boundary sliding and crystal-plasticity at very low melt fractions (Schulmann et al. 2008).

Only one study has attempted to map in detail the distribution of residual melt at grain scale, as inferred from microstructures of the kind described above (Sawyer 2001; although see also Rosenberg & Riller 2000; Marchildon & Brown 2002), so there is very little information on the grain-scale distribution of former melt in residual source rocks. The sample mapped by Sawyer is a residual granulite, interpreted to be derived from a greywacke protolith, collected from a low-strain domain. This sample is composed of plagioclase, quartz, orthopyroxene (approx. 7 vol%), biotite (<5 vol%), accessory minerals and inferred former melt (approx. 2 vol%). Based on an appropriate melting reaction, the presence of approximately 7 vol% orthopyroxene indicates that about 18 vol% melt was produced by biotite breakdown melting. Only approximately 2 vol% melt, or about one-ninth of that produced, remains in the residual granulite crystallized along grain boundaries (figure 4a), which indicates that approximately 16 vol% melt, or about eight-ninths of that produced, was able to escape from the host. This suggests that, at high temperatures, the effects of

Phil. Trans. R. Soc. A (2010)
Figure 4. The distribution of melt remnants in granulite. (a) Thin-section map showing the distribution of melt remnants in granulite facies metagreywacke from the Ashuanipi Subprovince, Québec (thin section is cut perpendicular to the foliation and parallel to the regional lineation trend, i.e. it is approximately the XZ plane of the strain ellipsoid). The distribution of melt is uniform and occurs along grain boundaries; there is no alignment of through-going melt-filled structures that might indicate a fracture-controlled distribution on the thin-section scale (adapted from Sawyer (2001) with permission from Wiley-Blackwell.). Scale bar, 5 mm; ‘fol’ is the strike of the foliation. (b) Outcrop map (vertical face, parallel to the stretching lineation) to show the distribution of leucosome in a strongly melt-depleted granulite facies metagreywacke. Thin, continuous leucosomes are oriented parallel to the layering and wider discordant leucosomes occur in shear bands (labelled SZ in the right-hand half of the figure) and apparent interboudin partitions (labelled IB in the right-hand half of the figure) to form a net-like array of interconnected former melt channels in the outcrop. A significant feature of the leucosome array is that no point in the outcrop is more than a few tens of centimetres from a former melt channel. However, the leucosomes in these granulites do not have melanosomes, but the host is residual in composition with orthopyroxene contents that suggest loss of 30–40 wt% melt. Hence, the leucosome network is inferred to be the remains of a melt-flow network that drained the host (adapted from Guernina & Sawyer (2003) with permission from Wiley-Blackwell.). Scale bar, 50 cm.

Contemporaneous deformation and associated pressure gradients lead to a melt connectivity transition that may be significantly lower than the 7 vol% estimated by Rosenberg & Handy (2005) and lower than the upper limit for the percolation threshold of 3.4 vol% suggested by Laporte et al. (1997). As figure 4a shows, the distribution of melt is relatively uniform with no tendency to form veins, although
most of the former melt occurs as connected arrays up to five grain diameters long located along grain boundaries that lie within 20° of the foliation, whereas those grain-boundary melt pseudomorphs at high angles to the foliation rarely extend for more than one grain diameter.

Evidence used to infer the form of former melt networks in the anatectic zone comprises the outcrop-scale arrays and networks of leucosomes and associated melanosomes. These features are inferred to be related to segregation and loss of melt, where the leucosomes represent accumulations of early formed crystals or unfractonated melt compositions or late-crystallized residual melt left in the drainage network, and the melanosomes represent the local residues left after the melt has been extracted (Brown et al. 1995; Sawyer 2001). Leucosomes may not be associated with melanosomes, and melanosomes may not always occur adjacent to leucosomes; in the former case, migrating melt has become trapped, whereas in the latter case, the residue has lost most of the melt generated at that site, and melt remaining on grain boundaries most likely crystallized as overgrowths on pre-existing grains in the residue. The form of some of the structures filled by relict leucosome or marked by melanosome are clearly collapse structures (Kriegsman 2001; Brown 2005; Bons et al. 2008).

At the hand-sample to outcrop scale, there are only a few studies of: (i) the spatial relations among in-source leucosomes and melanosomes, and between these and leucocratic veins (melt that has migrated out of source to form sills and/or dykes) and/or irregular small bodies of ponded anatectic granite and (ii) the geometric relationship of leucosomes, leuco-granite veins and irregular bodies of granite to the associated tectonic structures and/or the apparent finite state of strain. Most of these studies are in one or two dimensions (e.g. Oliver & Barr 1997; Sawyer 2001; Guernina & Sawyer 2003; Bons et al. 2004; Soesoo et al. 2004), or they use approximately perpendicular two-dimensional surfaces to infer the three-dimensional form of the leucosome networks (e.g. Collins & Sawyer 1996; Tanner 1999; Solar & Brown 2001a; Marchildon & Brown 2003; Brown 2005); one study is in three dimensions (Brown et al. 1999), and one study investigated the original form of collapse structures formed by draining of melt (Bons et al. 2008). There is only one study in which melanosome distributions were systematically mapped in several outcrops (Sawyer 2001). A small number of these studies has evaluated whether in-source leucosomes and leucogranite veins (sills and/or dykes) are scale-invariant (e.g. Tanner 1999; Marchildon & Brown 2003; Brown 2005), although only one study appears to have clearly separated the in-source leucosomes from the leucogranite veins (Marchildon & Brown 2003; Brown 2005). Most of these studies were undertaken in migmatite outcrops derived from metasedimentary protoliths. Two studies show leucosome arrays in migmatitic mafic gneiss or amphibolite (Hartel & Pattison 1996; Sawyer 2001), both of which display features such as fabric parallel, transverse and oblique leucosomes similar to those seen in migmatitic pelites and greywackes.

One approach to interpreting leucosome arrays or networks and their relations to granite in veins or dykes is to map their two-dimensional distribution at outcrop scale (e.g. Oliver & Barr 1997; Sawyer et al. 1999; Sawyer 2001). Other approaches include obtaining the three-dimensional geometry of leucosome at hand-sample scale, either by reconstructing the morphology from serial sections or by using high-resolution computed X-ray tomography (e.g. Brown et al. 1999),
Figure 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). (b) Steeply-oriented surface parallel to foliation (note subhorizontal elongation lineation) to show vertical extent of leucosome in transverse structures (Petit Mont, Arzon, Morbihan).

and mapping mutually perpendicular two-dimensional surfaces to infer the three-dimensional form and distribution of leucosome for small outcrops (e.g. Tanner 1999; Marchildon & Brown 2003). Alternatively, aspects of the leucosome and/or granite vein or dyke distribution may be quantified by measuring their widths and/or spacing along a linear section perpendicular to the network or set and presenting these data as cumulative frequency plots of the number of, for example, veins or vein intervals wider than a certain dimension against that dimension (e.g. Marchildon & Brown 2003; Soesoo et al. 2004).

In a study of leucosome distribution in net-veined migmatites (figure 5) from the southern Brittany migmatite belt of western France in the European Variscides, Marchildon & Brown (2003) measured thickness and spacing of foliation-parallel (stromatic) leucosomes along one-dimensional line traverses perpendicular to foliation in close-to-horizontal, lineation parallel outcrop surfaces. Leucosome thicknesses mostly fall in the range of 1–10 mm, with an upper limit between 20 and 30 mm. The number of thicker layers decreased
Figure 6. Features associated with melt extraction in migmatites from Petit Mont, Morbihan. (a) View roughly northwest of 0.5 m-wide dykes of granite in stromatic migmatite. (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). For more details about these outcrops see Brown (2004, 2006).

abruptly with increasing thickness, which is inconsistent with scale invariance. This result led Marchildon & Brown (2003) to infer that leucosome formation was controlled by short-range melt movement along grain boundaries to form melt-rich layers constrained by the tectonic fabric and pre-existing compositional layering.

In the same area, centimetric- to metric-scale granite dykes are abundant at outcrop (figure 6a). Although, the volumetric importance of the dykes varies in space, they may represent up to 20 per cent of the area of an outcrop. At map scale in the wider region, the dykes may be as large as several hundreds of metres in width. In the study area, dykes with thicknesses greater than 10 cm wide show a power-law distribution with an exponent of 1.11 (Brown 2005), suggesting that they may be scale-invariant, although the dataset is small (87 dykes), and the range of observations is only two orders of magnitude. The largest dykes measured were 3 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).
Figure 7. Subhorizontal surface through stromatic migmatite to show granite dykes that intersect in a common steeply-inclined cylindrical channel structure; these dykes do not clearly cross-cut each other, but rather merge where they meet. Although the dykes are discordant at outcrop scale, there is only limited evidence for cross-cutting relationships at the vein-to-grain scale, suggesting that material now in the dykes was molten at the same time as the material in the host rock included a melt phase (Petit Mont, Arzon, Morbihan).

Although the dykes appear to cross-cut structures in the migmatites (figure 6a), the modal mineralogy, grain size and microstructure of granite in the dykes is indistinguishable from those of leucosome in the migmatites (figure 6b). Marchildon & Brown (2003) interpreted this petrographic continuity to mean that these structures once hosted a continuous melt-bearing network, and to indicate that material in leucosomes and in dykes underwent final crystallization at the same time. This inference does not mean that leucosomes (or necessarily granite in dykes) have liquid compositions. Individual dykes may intersect one another, but where this occurs, the individual dykes tend to lose continuity through the intersection, indicating contemporaneity of emplacement and synchronicity of final crystallization, and suggesting that these intersections may form backbone structures for faster ascent of melt (figure 7).

This petrographic continuity between leucosome in the drainage networks and granite in dykes suggests that the ascent 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 of melt pockets (e.g. Regenauer-Lieb 1999; Eichhubl et al. 2001; Du Bernard et al. 2002; Eichhubl & Aydin 2003). Zigzag propagation paths and blunt tips of dykes point to ductile fracture as the mode of formation of these conduits (Brown 2004), consistent with modelling the behaviour of fluid-bearing ductile rock materials (Regenauer-Lieb & Yuen 2000; Regenauer-Lieb et al. 2006). If dyking is a general ascent mechanism, it is likely that a ductile-to-brittle fracture process is the mechanism by which the dykes initiate and propagate (Brown 2008). Thus, dykes born as ductile fractures will become brittle fractures as the dyke traverses from the suprasolidus anatectic zone to cross the subsolidus interval of crust beneath a pluton. Dyke interaction during ascent may focus melt flow into a smaller number of more widely spaced dykes (e.g. Ito & Martel 2002).
Two other studies by Soesoo et al. (2004) and Bons et al. (2004) report widths of magmatic veins in a drill core from the Palaeoproterozoic Estonian basement and from a stromatic migmatite north of Turku, in the Palaeoproterozoic migmatite–granite belt of southern Finland. In the Estonian drill core, which is oriented at 70° to the stromatic foliation, leucosome and veins make up 24 per cent of the 40 m long section. Here, fine-grained leucosomes are distinguished from coarser granite veins, each of which shows a power-law distribution with exponents of 1.9 and 1.1 for thicknesses greater than 5 and 10 mm, respectively. In a second drill core, granite veins form 26 per cent of the 94 m section; they show a power-law distribution with an exponent of 1.15. In southern Finland, veins make up 31 per cent of a 5 m long section measured perpendicular to stromatic foliation; a best fit through all veins thicker than 10 mm gives a distribution exponent of 1.1. Although the width-distribution exponent along a linear section cannot be simply equated with a volume-distribution exponent, these examples show that power-law distributions of melt batches are indeed found in nature. For comparison, a theoretical model of the distribution of melt batch sizes for different values of the volume-distribution exponent is given in Bons et al. (2004).

It has been argued on the basis of two studies, one at outcrop to map scale and the other at grain to hand-sample scale, that in the transition from grain-scale melt flow to channellized flow in veins and ascent conduits, the direction of melt flow will be controlled by elements of the metamorphic fabric, particularly the foliation and lineation (Brown & Solar 1998a; Brown et al. 1999). In their study, Marchildon & Brown (2003) observed that structures in mutually perpendicular two-dimensional surfaces suggest an anisotropy of the leucosome network related to a subhorizontal lineation, which they confirmed using the box-counting method to analyse the structure of the outlines of the leucosome network in surfaces parallel to $X–Z$ and $Y–Z$ sections of the finite-strain ellipsoid (where $X$ is parallel to the lineation, $Z$ is perpendicular to foliation and the $Y–Z$ plane is perpendicular to lineation). The leucosome morphology exhibits only limited scale invariance, with a fractal dimension around 1.4 in the $X–Z$ plane and around 1.6 in the $Y–Z$ plane (Marchildon & Brown 2003). The higher fractal dimension of the leucosome morphology in the $Y–Z$ plane in comparison with the $X–Z$ plane is consistent with the observed greater apparent complexity and roughness of the leucosome morphology viewed in the $Y–Z$ plane compared with that in the $X–Z$ plane. This is consistent with an inference that melt flow was primarily in the plane of the foliation along the lineation to feed dykes propagating through this level of the anatectic zone, although the final leucosome morphology must record a late stage in the process as melt flow declined.

These leucosome networks are inferred to represent former sites of melt accumulation, forming links for melt drainage from grain boundaries to ascent conduits, and acting as reservoirs for melt during build-up of melt volume between extraction events (Handy et al. 2001). Similarly, in the examples described by Oliver & Barr (1997) and Guernina & Sawyer (2003), melt is inferred to flow in the plane of the foliation to developing transverse dilatant structures or to interfaces between layers of different composition, then laterally to dilatant sites and onto ascent conduits to escape the source. Recently, Weinberg & Mark (2008) have interpreted layer-parallel leucosomes linked with axial plane leucosomes that disrupt and transpose fold hinges to

Phil. Trans. R. Soc. A (2010)
record melt flow parallel to layering down gradients in melt pressure into the
axial planar foliation to facilitate melt extraction and fold transposition in
the process.

In a study of predominantly net-veined migmatites from the Moldanubian Zone
of western Bohemia in the European Variscides, Tanner (1999) used the box-
counting method to analyse the structure of the leucosome network in mutually
perpendicular surfaces parallel to the \( X-Z \) and \( Y-Z \) sections of the finite-strain
ellipsoid (where \( X \) is parallel to the lineation, \( Z \) is perpendicular to foliation and
the \( Y-Z \) plane is perpendicular to lineation). Tanner interpreted the results to
indicate that the leucosome morphology exhibits scale invariance, with a fractal
dimension of 1.80 in the \( X-Z \) plane and 1.81 in the \( Y-Z \) plane. He further
conjectured that the closeness of these values in both the \( X-Z \) and \( Y-Z \) planes
to 1.89, which is the characteristic fractal dimension of the Sierpinski Carpet,
suggests the Menger Sponge an appropriate model for the three-dimensional scale-
andinvariant geometry of the leucosome network in the net-veined migmatites. Such
a model implies a scale-invariant hierarchy with a single backbone structure with
the largest dimension of the set transferring melt to the upper crust.

(c) Pervasive melt migration through the anatectic zone

As we have seen in the previous section, efficient migration of melt requires the
development of an extensive permeable network of channels during deformation.
Where the tectono-metamorphic rock fabrics are steep, such as may be developed
in transpressive orogens, melt ascent may occur via conduits concordant with
these fabrics (D’Lemos et al. 1992; Brown 1994). In two parallel studies, Brown &
Solar (1998a, 1999; see also Solar & Brown 2001a,b) and Weinberg (1999; see also
Weinberg & Searle 1998; Weinberg & Mark 2008) proposed models for pervasive
fabric-parallel melt ascent through the anatectic zone in which the heat advected
with melt displaces isotherms upwards, allowing melt migration to shallower
depths.

In the Acadian transpressive belt in the northern Appalachians, Brown & Solar
(1999; see also Brown & Solar 1998a; Solar & Brown 2001a,b) demonstrated that
the form of leucosomes and granites in ascent conduits in host migmatites mimics
the apparent strain ellipsoid recorded by the host rock fabrics. Concordant rod-
like leucosomes and decametric granites occur in zones of apparent constrictional
strain, whereas stromatic migmatites and concordant metric to rarely kilometric
and polyphase tabular granites occur in zones of apparent flattening strain
(figures 2 and 8). The granites in the zones of apparent flattening strain
are similar to the ‘magma sheets’ of Weinberg (1999), who argued that melt
migrates parallel to high-permeability zones such as foliation or bedding planes
(see also Guernina & Sawyer 2003). Larger polyphase granites and granite
injection complexes that are common in anatectic terranes may be formed by
amalgamation of many such ‘magma sheets’ as the thermal structure evolves
and the anatectic front migrates to shallower levels in the crust (figure 2). This
mechanism has been explored by Leitch & Weinberg (2002) who numerically
modelled the feedback relation between melt migration and heating of the host
rock (cf. Hobbs & Ord 2010). These authors concluded that pervasive migration
is an efficient way of advectively heating the lower-to-middle crust, and further
that this mechanism could produce an injection complex several kilometres thick,
consisting of about half injected melt and half original crust.

Phil. Trans. R. Soc. A (2010)
A similar process is inferred for leucosomes in Mesoproterozoic aluminous metapelites from Broken Hill, Australia. Here, White et al. (2004) describe spatially focused melt formation that resulted in pathways for melt escape, as recorded by leucosome, that are parallel to foliation defined by highly depleted melanosomes. In contrast, in the Mount 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 moderate-to-steeply plunging regional elongation direction, defined by leucosome associated with coaxial folds and a strong mineral-elongation lineation (Collins & Sawyer 1996).

In a related process involving pervasive melt flow, but at the grain scale in a zone of high strain, Hasalová et al. (2008a,b) have demonstrated a gradual transition from high-grade solid-state orthogneisses through stromatic and schlieren migmatite into irregular bodies of granite conformable with a
foliation in the eastern part of the Gföhl Unit in the Moldanubian domain of the Bohemian Massif. This interpretation is consistent with the view that the Gföhl Unit records melt-assisted crustal-scale channel flow (Schulmann et al. 2008). The orthogneiss-to-granite sequence is characterized by progressive changes in mineralogy, mineral chemistry and microstructure consistent with increasing grain-scale equilibration between melt infiltrating from a deeper source and the host orthogneisses during exhumation from 790°C at 0.9–0.6 GPa to 690°C at 0.5–0.4 GPa. Overall, the chemical variations require a large volume of silicate melt to have passed through and interacted with the orthogneisses, contrary to an intuitive view that such a process will not be efficient and might be unlikely to occur in nature.

Jackson et al. (2003, 2005) present a quantitative numerical model of melt segregation by buoyancy-driven flow along grain edges coupled with compaction of the residue. The spatial distribution and composition of the melt depends upon the relative upward transport rates of heat and melt because the melt thermodynamically equilibrates with the host rock at progressively lower temperatures during migration. As a result, the chemical composition of the melt varies spatially and temporally, and is controlled not only by the composition and mineralogy of the source rock, the depth of melting and the melting reactions, but also by the physical processes by which the melt migrates and segregates from its residue. In this model, petrologic diversity is predicted, even for melts derived from a homogenous protolith (Rushmer & Jackson 2008).

The accumulating evidence suggests that the three main modes of melt flow through the anatectic zone—intergranular porous flow, channel flow in vein networks and channel flow in ascent conduits—are able to occur synchronously to enable melt extraction. Synchronicity of these flow modes may be explained by granular behaviour, where there is a melt film present between grains and between aggregates of grains in melt-bearing rock that reduces the large cohesive forces that characterize a solid medium (cf. Nicolas & Poliakov 2001). Melt migration through the anatectic zone will depend on the size of the aggregates, particularly the lozenges of melt-depleted or low-melt fraction rock between vein networks in many migmatites (Brown 2004; Olsen et al. 2004; Rutter & Mecklenburgh 2006). Melt flow evolves in time and space from intergranular porous flow to channel flow in vein networks to channel flow in larger conduits, as the dimension of the aggregates increases (cf. Olsen et al. 2004). Nicolas & Poliakov (2001) use a physical experiment on pressurized air circulation through a granular medium to illustrate these synchronous flow modes (figure 9), but the rates of flow for crustal systems will be somewhat slower than those they calculate for mafic systems owing to the higher viscosity of the melt (one to three orders of magnitude higher; according to composition, water content and temperature; see Clemens & Petford 1999).

5. Episodicity of melt-extraction events

Episodicity is an important feature of melt extraction, where the time scale of a melt-extraction event, defined as a batch of melt that leaves the source and reaches a pluton, is short in comparison with the build-up, and the source is drained of melt by multiple events, consistent with evidence from plutons for
construction from multiple additions of melt (e.g. Brown et al. 1981; Pressley & Brown 1999; Mahan et al. 2003; Coleman et al. 2004; Bartley et al. 2006; Matzel et al. 2006; Miller et al. 2007; Walker et al. 2007; Miller 2008). Melt-extraction events are inferred to be triggered because a threshold is exceeded during the continuous thermal evolution. The threshold for melt extraction is reached at some combination of melt volume and distribution within a subvolume of the source (the catchment) that enables episodic draining via a common ascent conduit. Since we know from geophysical data that plutons are not fed by more than one, or at most several, ascent conduits (Vigneresse 1988, 1990), these common ascent conduits are inferred to be spaced in a manner consistent with the spacing of plutons (Cruden 2006).

In order to predict a melt-extraction event, we require knowledge of the relation between the melt fraction that must be reached and the length scale of the catchment over which it must be exceeded, which is likely to be different for distinct source geometries and stress conditions. Rosenberg & Handy (2005) have argued that the threshold volume for melt extraction is approximately 7 vol% melt
Figure 10. Schematic diagram to illustrate melt extraction, ascent and emplacement along an active continental margin (based on a model in Cruden 1998). (a) Melt in the source, shown schematically in black, is located in fabric-parallel stromata and thin sills prior to formation of dikes to allow ascent. (b) After magma ascent, a pluton has been formed at the level of the brittle–ductile transition zone, with space being made by a combination of lifting the roof and depression of the floor, accommodated by volume loss in the source owing to excavation of the melt. The volume flux rate of melt extraction from the source ($Q_{Ex}$), the volume flux rate of melt ascent ($Q_{As}$) and the volume flux rate of melt into the pluton ($Q_{Em}$) are assumed to equal the volume expansion rate of the pluton and to be balanced at the crustal scale by the rate of relaxation of the deep crust. The depths shown on the right-hand side of the figure are intended as a general indication of the likely range along an active continental margin.

(cf. Handy et al. 2001). The threshold length scale of the catchment is unknown, but it must be big enough to generate a batch of melt of sufficient size to reach the site of pluton emplacement. For example, for a system similar to the one shown schematically in figure 10, an extraction event comprising approximately 10 km$^3$ of melt at a melt volume threshold of approximately 7 vol% over an area or footprint of approximately 700 km$^2$ involves significant horizontal permeability in a catchment of approximately 200 m thick, which implies a kilometric length scale for horizontal melt flow to the ascent conduit.
Episodic melt extraction from the source is consistent with increasing evidence of ascent conduits that record evidence of passage of multiple batches of melt (e.g. Brown & Solar 1999; Solar & Brown 2001a; Sawyer & Bonnay 2003). Limiting melt in the source to an amount below a threshold value of approximately 7 vol% is consistent with the lower end of the range of estimates of the melt volume in the crust beneath the central Andes and the Tibetan Plateau at the present day (Schilling & Partzsch 2001; Babeyko et al. 2002; Unsworth et al. 2005). To a first approximation, the volume of magma emplaced in the middle-to-upper crust equals the volume of melt extracted from the middle-to-lower crust, and there is no space problem at the crustal scale (figure 10). Furthermore, if the interval between melt-extraction events is in the order of 1000 years, then for reasonable values of bulk viscosity and elastic shear modulus, the relaxation time of the crust is of the same order (Vigneresse 2006; Ablay et al. 2008), and extraction and emplacement are complementary processes between which there is a feedback relation modulated by processes in the ascent conduit (Brown 2001a,b, 2007b; Hobbs & Ord 2010). In this fashion, a pluton of volume $10^4$ km$^3$, composed of $10^3$ individual batches of melt each of $10$ km$^3$, will be constructed in 1 Ma, through a combination of initial elastic deformation followed by viscous flow and stress relaxation, and decreasing melt pressure in the pluton chamber that allows a new batch of melt to exploit the opportunity for emplacement.

In circumstances where plutons are constructed from multiple batches of magma intruded at a similar level in the crust, an injection complex or sheeted granite may be the result according to the depth at which magma is emplaced. The Pangong injection complex of the Indian Karakorom may represent a staging level in the transfer of melt to shallower levels (Weinberg et al. 2009). The Wuluma Granite of the Arunta Inlier in central Australia (Lafrance et al. 1995), the Qørqut granite complex of southern west Greenland (Brown et al. 1981) and the Fosdick migmatite–granite complex in west Antarctica (Korhonen et al. in press) represent granite injection complexes located close to the source. In contrast, the McDoogile pluton in the Sierra Nevada (Mahan et al. 2003) and the Galway granite in western Ireland (Leake 2006) on the one hand, and the batholiths of the Coast Plutonic Complex of British Columbia (Brown & McClelland 2000) on the other hand, may represent intermediate depth examples of subvertically sheeted and subhorizontally sheeted plutons, respectively. These injection complexes and sheeted granites record pluton growth by multiple emplacements of many small magma batches (perhaps $10^3$–$10^5$ batches).

6. Concluding remarks

In this contribution, I have presented a summary of the melting process and its consequences for the behaviour and differentiation of the crust. Crustal rocks undergo melting via a sequence of reactions beginning with limited melt production at the wet solidus (generally less than 1 vol% melt) and continuing viahydrate mineral-breakdown melting reactions. Major melt production is related to hydrate mineral-breakdown melting, with the possibility of greater than 50 vol% melt being generated according to the fertility of the protolith composition and the intensive variables. Initial melting is commonly diffusion controlled, but melt-bearing rocks become porous at a few vol% melt, initiating an advective flow.
regime driven by gradients in fluid pressure. As the melt volume approaches and exceeds the melt-connectivity transition (approx. 7 vol% melt), melt may be lost from the system in the first of several melt-loss events.

In addition, I have presented a synthesis of the small number of studies of the spatial and temporal patterning imprinted on the residual middle-to-lower crust by the process of melt extraction. This synthesis shows that compositional layering and tectono-metamorphic fabrics exercise a strong control on the form of melt-flow channels at all scales, and that melt distribution in the crust commonly will be heterogeneous at various length scales. Arrays or networks of leucosome and melanosome observed in nature provide a record of melt-flow pathways that were operative during the transition from storage to drainage. Melt flow is driven by gradients in fluid pressure at all scales, that is from grain boundary to vein to ascent conduit.

In strongly anisotropic protoliths, I have proposed that feedback relations enable an interconnected network of melt-filled deformation bands to evolve to an optimum structure, probably controlled by rates of melt production and strain accumulation. A threshold or critical point may be reached at some combination of melt fraction and distribution within a subvolume in the source that enables formation or activation of conduits of a size that allows melt to flow out of the source and upward through the crust in steeply dipping ascent conduits that may be concordant with steeply oriented fabrics or discordant dykes. There is increasing evidence of common ascent conduits for multiple batches of melt, and some intrusions emplaced below the ductile-to-brittle transition zone may represent staging levels in the transfer of melt to shallower levels. Also, I have argued that melt extraction is a self-organized critical phenomenon, which is supported by the scale-invariance of granite dykes and plutons from a limited number of studies.

Finally, I suggest that, at the scale of the crust, melt extraction, ascent and emplacement may constitute a closed, internally driven system—an inherent consequence of melt withdrawal at depth and the buoyant rise of magma to higher levels in the crust. The observed relationships between pluton emplacement and the development of regional structures is a natural consequence of multi-scale interactions between vertical material transfer (melt up and host rocks down) and tectonic deformation during orogenesis. Preliminary indications are that anatectic systems are strongly self-organized from the bottom up as they become more ordered to move melt from grain boundaries in the anatectic zone via veins and dykes to a limited number of larger ascent conduits that transport melt through the subsolidus crust to the ductile-to-brittle transition zone where it is emplaced to form plutons of granite. This is consistent with the expectation that ordered systems are more efficient at dissipating potentials and produce entropy faster than disordered systems (Hobbs & Ord 2010).

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Review. Patterning of the deep crust


Phil. Trans. R. Soc. A (2010)


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