Orogeny, migmatites and leucogranites: A review

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The type of P-T-t path and availability of fluid (H_2O-rich metamorphic volatile phase or melt) are important variables in metamorphism. Collisional orogens are characterized by clockwise P-T evolution, which means that in the core, where temperatures exceed the wet solidus for common crustal rocks, melt may be present throughout a significant portion of the evolution. Field observations of eroded orogens show that lower crust is migmatitic, and geophysical observations have been interpreted to suggest the presence of melt in active orogens. A consequence of these results is that orogenic collapse in mature orogens may be controlled by a partially-molten layer that decouples weak crust from subducting lithosphere, and such a weak layer may enable exhumation of deeply buried crust. Migmatites provide a record of melt segregation in partially molten crustal materials and syn-anatectic deformation under natural conditions. Grain boundary flow and intra- and inter-grain fracture flow are the principal grain scale melt flow mechanisms. Field observations of migmatites in ancient orogens show that leucosomes occur oriented in the metamorphic fabrics or are located in dilational sites. These observations are interpreted to suggest that melt segregation and extraction are syntectonic processes, and that melt migration pathways commonly relate to rock fabrics and structures. Thus, leucosomes in depleted migmatites record the remnant permeability network, but evolution of permeability networks and amplification of anomalies are poorly understood. Deformation of partially molten rocks is accommodated by melt-enhanced granular flow, and volumetric strain is accommodated by melt loss. Melt segregation and extraction may be cyclic or continuous, depending on the level of applied differential stress and rate of melt pressure buildup. During clockwise P-T evolution, H_2O is transferred from protolith to melt as rocks cross dehydration melting reactions, and H_2O may be evolved above the solidus at low P by crossing supra-solidus decompression-dehydration reactions if micas are still present in the depleted protolith. H_2O dissolved in melt is transported through the crust to be exsolved on crystallization. This recycled H_2O may promote wet melting at supra-solidus conditions and retrogression at subsolidus conditions. The common growth of 'late' muscovite over sillimanite in migmatite may be the result of this process, and influx of exogenous H_2O may not be necessary. However, in general, metasomatism in the evolution of the crust remains a contentious issue. Processes in the lowermost crust may be inferred from studies of xenolith suites brought to the surface in lavas. Based on geochemical data, we can use statistical methods and modeling to evaluate whether migmatites are sources or feeder zones for granites, or simply segregated melt that was stagnant in residue, and to compare xenoliths of inferred lower crust with exposed deep crust. Upper-crustal granites are a necessary complement to melt-depleted granulites common in the lower crust, but the role of mafic magma in crustal melting remains uncertain. Plutons occur at various depths above and below the brittle-to-viscous transition in the crust and have a variety of 3-D shapes that may vary systematically with depth. The switch from ascent to emplacement may be caused by amplification of instabilities within (permeability, magma flow rate) or surrounding (strength or state of stress)

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the ascent column, or by the ascending magma intersecting some discontinuity in the crust that enables horizontal magma emplacement followed by thickening during pluton inflation. Feedback relations between rates of pluton filling, magma ascent and melt extraction maintain compatibility among these processes.

1. Introduction

Much of Earth's crust comprises deformed metamorphic rock formed and modified during orogenesis. Orogens are open systems in which self-organized behavior enables efficient dissipation of energy, manifest in feedback relations among deformation, metamorphism and anatexis (Brown and Solar 1998a; Hodges 1998). How feedback is recorded relates to properties of rocks, particularly the rheology. As suggested by Haugerud and Zen (1991), metamorphic rocks are flight recorders recovered from the wreckage of an orogen that can tell us something of the history of the passage of material through the orogen. On the flight through *P-T* space, what happens to a particular rock is determined by the reaction lines/fields crossed along the prograde (increasing T) stage, whether any of these reaction lines/fields are recrossed along the retrograde (decreasing T) stage, and whether any fluid generated along the prograde segment of the path (H_2O-rich metamorphic volatile phase or melt) is retained to allow reaction reversal along the retrograde segment. The critical step for nucleation of product phases and the rate of dissolution of refractory phases are additional factors. Field relations, rock fabrics and microstructures, mineral assemblages, reaction textures, mineral compositional variation and isotope data provide a record of the rock history (e.g. Brown 1993; Brown and O'Brien 1997; Brown 2001). Our ability to read the evidence preserved in metamorphic rocks is critical to understanding the history they record, which is essential to constraining alternative models of orogenesis. This is a prerequisite to improving our understanding of orogenic processes.

Although granites are found in a variety of plate tectonic settings (Atherton and Tarney 1979; Pitcher 1993), the principal environments of mountain building and crustal melting are continental arcs and collision zones (Brown *et al.* 1995a). Evidence of crustal melting in orogens is present at all scales, from microfracturing in residual quartz and inferred melt pseudomorphs in grain boundary pores in protolith materials (e.g. Connolly *et al.* 1997; Holness and Clemens 1999; Rosenberg and Riller 2000; Watt *et al.* 2000; Marchildon and Brown 2001) to plutonic belts the length of lithosphere plate boundaries (Kay and Rapela 1990; Harmon and Rapela 1992) and major collisional mountain belts (LeFort 1986; Harrison *et al.* 1997). Continental collision is occurring today in the subduction of India beneath the Eurasian tectonic plate to form the Himalayas and the Tibetan Plateau (Hodges 2000), and is widely inferred for many major Phanerozoic orogens. Subduction precedes collision, so that the degree of preheating during ocean closure may be an important variable in the thermal evolution of collisional orogens (Jamieson *et al.* 1998). The process of melt generation and segregation, and magma extraction, ascent and emplacement leads to differentiation of the continental crust into a depleted lower crust and an enriched upper crust (e.g. Brown *et al.* 1995a). However, in continental arcs there is a significant involvement of mantle-derived melts in the generation of calc-alkaline granites (e.g. Patiño Douce 1999), which means that granite magmatism in continental arcs results in crustal growth. The role of lithosphere delamination, slab detachment and mantle magmatism in the evolution of collision zones requires further study.

2. An overview of orogens and melting

Geodynamics is driven by heat released from Earth's core and radioactive decay. Heat transport, by convection, conduction or advection, determines the T attained at depth in the lithosphere and, therefore, the type of crustal metamorphism, whether melting occurs, and the rheological behavior. During orogeny, tectonic and magmatic processes advect heat to shallower levels in the deforming lithosphere, and melting and melt transport redistribute radioactive elements. These processes control the evolution and differentiation of continental crust, and create depleted granulites typical of the lower crust and upper-crustal granites. Orogens are systems far from equilibrium that exhibit a coherent arrangement of features in space and time—they are dissipative structures that provide mechanisms for dispersal of the internal energy of Earth (Brown and Solar 1998a; Hodges 1998).

Orogenic evolution in collision zones comprises a period of crustal thickening, a period during which thickening and exhumation commonly are balanced, and a period of collapse (Dewey 1988; Vanderhaeghe and Teyssier 2000), although there are many variables in the evolution of orogens (e.g. Ellis and Beaumont 1999). Since strain in orogenic belts does not appear to be linked in a simple way to plate kinematics, other factors contribute to deformation of the continental crust, particularly gravity-driven flow. Thermomechanical numerical modeling of the dynamic evolution of convergent orogens driven by subduction provides constraints on the physical conditions required to generate temperatures high enough for crustal anatexis, and allows investigation of the rheological structure leading to mechanical decoupling between the weakened crust and the subducting plate (Royden 1993; Thompson *et al.* 1997; Huerta *et al.* 1998; Jamieson *et al.* 1998).

Geological studies in ancient orogens (e.g. Brown 1994; Brown *et al.* 1995a) and geophysical studies in active orogens (e.g., Nelson *et al.* 1996; Partzsch *et al.* 2000) have been interpreted to indicate that the part of the crust is in a partially molten state during orogenesis. The presence of melt may weaken the crust sufficiently to decouple it from the subducting lithosphere plate (e.g. Burg *et al.* 1994; Vanderhaeghe *et al.* 1999; Vanderhaeghe and Teyssier 2001), and may enable exhumation of deeply buried crust (e.g. Hollister 1993; Brown and Dallmeyer 1996). Thus, it is important to understand the rheology of partially molten crust, to determine the mechanical effects of melt in the crust and to evaluate the consequences of these effects during orogenic evolution (Davidson *et al.* 1994; Handy *et al.* 2001). There are feedback relations between heat and melting and melt transport and deformation (Brown and Solar 1998a). Thus, crustal melting and melt segregation, extraction, ascent and emplacement are inevitably linked with the developing tectonic structure as a principal energy dissipation mechanism in the lithosphere.

One spatial feature of orogens that requires explanation is the variation in volume of exposed granite. Examples of granite-rich orogens include the Appalachian Mountain Belt, the European Variscan belt and the Lachlan fold belt of SE Australia, whereas granite-poor orogens include the Scandinavian Caledonides and the Central-Western Alps. One explanation put forth for the variation in amount of granite is the contrasting fertility of the major rock types involved in the different orogens (Vielzeuf *et al.* 1990). Where orogenic processes mainly rework an older basement, composed of differentiated crust, significant granite magmatism is unlikely (e.g. Sawyer 1998). In contrast, orogens that incorporate large quantities of fertile sedimentary rocks offer the possibility of voluminous granite production.

The premier analog for many ancient, deeply eroded collisional orogens is the Himalayan-Tibetan system. Combined seismic reflection profiling, wide-angle reflection, broadband teleseismic studies, and magnetotelluric surveys yield coincident observations of reflection bright spots, a low velocity zone, and a low resistivity zone. These features coincide with an area of elevated heat flow. This remarkable collection of data has been interpreted to suggest a molten layer capped at 15–18 km depth by a semi-continuous horizon of granite plutons within the crust of southern Tibet (Nelson *et al.* 1996; Alsdorf *et al.* 1998; Alsdorf and Nelson 1999; Partzsch *et al.* 2000). This interpretation is supported by analogy with the exposed geology. It appears that the crust in southern Tibet may act like a 'water bed', whereby the elevated topography of the Tibetan Plateau is maintained above a viscous middle crust under hydraulic pressure, principally as a result of the 'extensive presence of melt'. In this model, the upper crust is supported on a lower viscosity (melt-bearing) layer that flows laterally when loaded, and orogen-parallel flow of the Tibetan Plateau is one manifestation of this phenomenon. Application of the 'water bed' model to other orogens may yield insight into previously intractable phenomena such as strain partitioning into contractional structures at depth and coeval extensional structures at a higher level in the crust during overall convergent orogenic deformation. There is little erosion at the surface across the Tibetan Plateau, so that if collision slows to a stop, the likely response may be extension and preservation of the middle crust as a migmatite belt. Thus, Tibet might be an appropriate analog for some high-T − low-P metamorphic belts particularly those commonly associated with voluminous granite plutons (Gerdes *et al.* 2000; Ledru *et al.* 2000).

It is an open question whether significant melt-producing perturbations arise from 'tectonics as usual' or from special mantle events (cf. Patiño Douce 1999; Thompson 1999). There is, however, clear evidence for melting during collapse late in the evolution of many orogens formed by collisional processes, popularly related to lithosphere delamination or slab detachment (von Blanckenburg and Davis 1995; Davis and von Blanckenburg 1995; Barboza *et al.* 1999; Handy *et al.* 1999). Collapse is recorded by metamorphic mineral parageneses that evidence decompression (e.g. Brown 1993) and the close age span of mineral geochronometers that cover a range of closure temperatures (Baldwin *et al.* 1993). Implied rapid cooling, inferred to reflect fast exhumation, is common in Phanerozoic orogens (e.g. Brown and Dallmeyer 1996; Zeck 1996), but may be less common in ancient orogens.

One model for the large ion lithophile element depletion of crustal rocks during granulite facies metamorphism is that of melt loss (e.g. Vielzeuf *et al.* 1990). In this model, granulite-facies metamorphism and crustal anatexis are coupled processes that result in chemical differentiation of continental crust. For example, on the southern margin of Variscan Europe, the Permian granites of the Strona-Ceneri zone commonly are regarded as the fugitive melt that complements residual stronalites of the adjacent Ivrea zone (e.g. Schnetger 1994). Additional information about processes in the lowermost crust may be deduced from studies of xenoliths of crustal materials brought to the surface in lavas (Zeck 1970; Cesare *et al.* 1997), and partial-melting processes recorded in xenoliths can be compared with those in exposed exhumed crust (e.g. Braun and Kriegsman 2001). Geochemical methods enable comparison between migmatites and plutonic granites (Brown and D'Lemos 1991; Williamson *et al.* 1997; Pressley and Brown 1999; Brown and Pressley 1999; Solar and Brown 2001b), and between exposed crust and lower crust brought up in xenolith suites in volcanic rocks (Rudnick and Presper 1990; Rudnick 1992; Rudnick and Fountain 1995), although such suites may be biased in favor of better preserved mafic rock types. Further, migmatite terranes may be compared with plutonic granites to evaluate the question of whether migmatites are feeder zones for granites (Brown and Solar 1999; Del Moro *et al.* 1999; Solar and Brown 2001b). There is also the issue of whether metasomatism may play a role in the chemical and rheological evolution of the crust (Franz and Harlov 1998; Harlov *et al.* 1998).

During clockwise *P-T* evolution of orogenic belts, H_2O is recycled as rocks cross dehydration melting reactions, with the H_2O being transported in melt (Holk and Taylor 1997), and suprasolidus decompression-dehydration reactions at low *P*, where H_2O is released that may promote new melting in adjacent rocks under appropriate circumstances (Thompson 2001a, 2001b). During retrogression above the wet solidus, back reaction between melt and residue may happen if the cooling path approximates the heating path, which may cause modification of mineral assemblage, chemistry and texture (Kriegsman and Hensen 1998; Kriegsman 2001a). It is likely that an H_2O-fluxed zone straddles the anatectic front, reflecting exsolution of H_2O during crystallization of stagnant melt arrested close to the anatectic front. This exsolved H_2O causes retrogression typical of migmatite terrains in collisional zones, particularly the late growth of muscovite at the expense of sillimanite (e.g. Solar and Brown 2001b).

Magma ascent and emplacement has been a controversial topic for the past decade. Pluton emplacement mechanics are investigated in the field and by modeling (e.g. Brun *et al.* 1990; Brown and Solar 1998b; Benn *et al.* 1998; Cruden 1998). However, the question of what causes the switch from magma ascent to emplacement is an issue that needs attention (e.g. Brown and Solar 1998b; Miller and Paterson 1999). Within plutons, we are beginning to understand better flow processes in magma (e.g. Petford and Koenders 1998a, compositional and thermal convection (e.g. Weinberg *et al.* 2000) and the role of deformation in the extraction of residual melts, both by shearing and by compaction (e.g. Park and Means 1996). Finally, rates of melt extraction from partially molten crust, magma ascent and pluton filling all are much faster than previously thought (Clemens and Mawer 1992; Clemens 1998; Cruden 1998; Tanner 1999; Thompson 1999; Harris *et al.* 2000; Petford *et al.* 2000), and linked by feedback relations that moderate the rates to maintain overall compatibility.

3. Thermal structure of orogens

The thermal structure of post-Archaean orogens reflects variations among internal heat production, advection, diffusion and basal heat flux. The normal orogenic evolution of subduction, followed by collision that leads to increased crustal thickening and decreased convergence rate, allows self-heating of the thickened orogen; clockwise *P-T* paths are characteristic. The following processes are important: material transfer advecting hot crust to shallow depths, and heat conduction to the surface by syntectonic erosion (Jamieson *et al.* 1998); increased radioactive heat generation due to thickening, important if high heat-producing units are present (Chamberlain and Sonder 1990; Gerdes *et al.* 2000); dissipation of mechanical energy generated during deformation (Scholz 1980; Barr and Dahlen 1989; Molnar and England 1990); and, increased strain rates due to weakening in the presence of melt (Rutter 1997). The initial distribution of radioactive heat production with depth, including the location and proportion of any anomalous heat-producing layers, is an important variable (Jamieson *et al.* 1998). Heat advected with mantle-derived magmas is important in the development of continental arcs, like the Andes, but the apparent smaller volume of mafic plutonic rocks exposed in ancient collisional belts means the role of mantle-derived magma in the thermal evolution of these belts is not well constrained. A combination of these processes strongly perturbs the geothermal gradient by displacing isotherms toward the surface, creating a thermal antiform with a near-isothermal core (Royden 1993; Thompson *et al.* 1997; Huerta *et al.* 1998; Jamieson *et al.* 1998), and generating temperatures exceeding the wet solidus and the stability of mica and amphibole in common crustal rocks (Brown and Solar 1999).

4. Melting, strain localization and pluton-driven metamorphism

Experimental melting of natural rock samples has demonstrated the importance of reactions involving breakdown of mica and amphibole (dehydration melting; Thompson 1999). Such reactions, which have steeply positive dP/dT slopes, are crossed during the prograde part of the P-T evolution (figure 1). One consequence of regional melting along clockwise P-T paths is that part of the orogenic crust 'must' be molten for a substantial portion of the evolution, perhaps for periods of millions of years, which is likely to have significant geodynamic implications (Vanderhaeghe and Teyssier 2000). During regional melting, transport of melt out of the source region takes with it H_2O, which has consequences for the evolution of the lower crust and hydration reactions in the upper crust once the transported H_2O is exsolved on crystallization of the melt.

Melting proceeds in a stepwise fashion (figure 1), beginning at the wet solidus, which is crossed at middle to lower-crustal depths in collisional orogens. Melt production at the wet solidus is limited by the amount of H_2O-rich metamorphic volatile phase present in the limited porosity remaining under amphibolite facies conditions. Muscovite dehydration (figure 1) is likely to be the major melt-producing step in the evolution of many orogens (Patiño Douce 1999), and is particularly important in the metapelitic component of the supracrustal succession in orogens (Patiño Douce and Harris 1998). Without an ingress of a H_2O-rich metamorphic volatile phase, the amount of melt produced generally is limited by the amount of muscovite in the protolith. Although the muscovite-biotite schist studied by Patiño Douce and Harris (1998) produced $\sim$ 20 vol.% melt within 25° C of the solidus at P of 6 kbar, the muscovite-biotite gneiss studied by Castro *et al.* (2000) produced melt at a slower but constant rate with increasing temperature to yield $\sim$ 15 vol. % melt by 900° C at 6 kbar. High (and ultra-high) temperature metamorphism may involve biotite dehydration (figure 1), which is particularly important for the greywacke component of supracrustal successions. Thus, temperature is buffered sequentially by the latent heat of melting required at the wet solidus, and during muscovite dehydration melting and biotite dehydration melting.

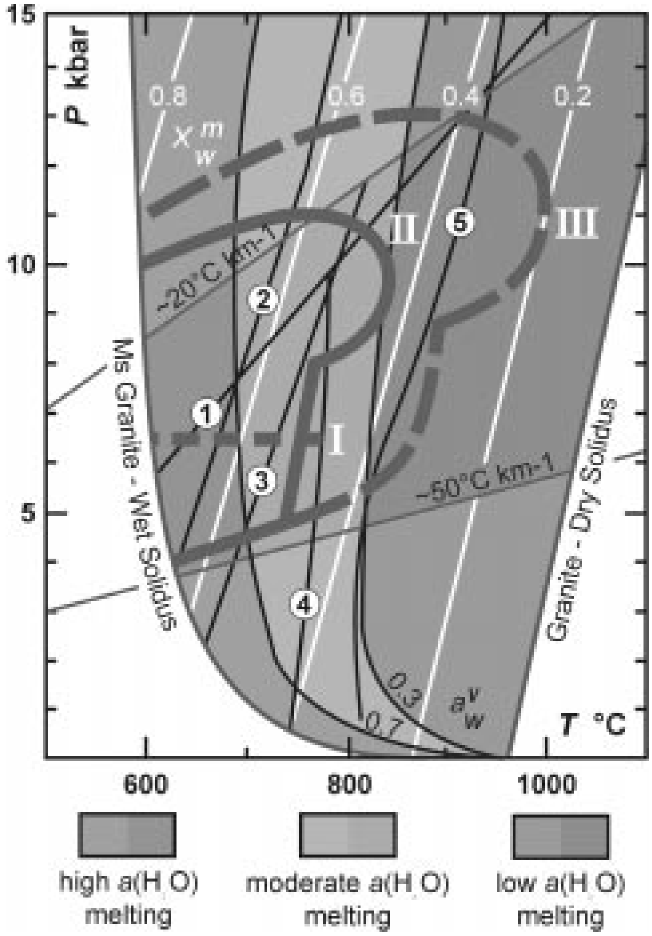

Figure 1. P-T diagram of the anatectic zone, that region in P-T space above the wet solidus for crustal rocks of granitic composition in which melt may be present. Reaction **(1)** is Ky → Sil, the Ms granite – wet solidus is from Huang and Wyllie (1981); reaction **(2)** Ms + Ab + Qtz → Kfs + Als + L is from Petö (1976) and marks the lower T limit of plagioclase reactions (Thompson and Tracy 1979); reaction **(3)** Ms + Pl + Qtz → Kfs + Sil + Bt + L is from Patiño Douce and Harris (1998); reaction **(4)** Bt + Pl + Als + Qtz → Grt + Kfs + L is from LeBreton and Thompson (1988); and, reaction **(5)** Bt + Pl + Qtz → Opx + Grt + Kfs + L is from Vielzeuf and Montel (1994). The symbol X_w^m is used to denote the mole fraction of H_2O in the melt and is considered to equal the activity of H_2O in the melt (Burnham 1979); isopleths of X_w^m are from Thompson (1996). The symbol X_w^v is used to denote activity of H_2O in the volatile phase, whereas a(H_2O) refers to water activity in the environment; isopleths of X_w^v are from Thompson (1996). Schematic P-T paths: I. Isobaric heating – cooling path characteristic of deep contact metamorphism; II. Stepped clockwise path (based on Brown and Dallmeyer 1996); and, III. Stepped clockwise path at ultrahigh T (based on Brown and Raith 1996; Raith *et al.* 1997).

Recent modeling (Harrison *et al.* 1998; Leloup *et al.* 1999; Nabelek and Liu 1999) suggests shear heating may be more important than realized previously (Brun and Cobbold 1980; Fleitout and Froidevaux 1980) as a contributor to the heat budget and thermal structure of orogens. Although shear heating in lithosphere-scale fault zones can induce a modest increase in temperature, results of modeling continental strike-slip shear zones suggest that shear heating in the upper mantle may induce lower crustal melting (Leloup *et al.* 1999). The possibility of such melting depends on upper mantle rheology and on the fertility of the lower crust, but the modeling predicts crustal melting at slip rates of 10-30 mm yr^{-1}. Syn-slip ascent of the melt will enhance the rising temperature in shallower parts of the fault zone. Further, shear heating should induce strain localization in the deeper parts of fault zones, consistent with observations from deeply eroded ancient orogens. This raises the issue of the relationship between localization of strain and accumulation of melt (e.g. Brown and Solar 1998a). At high fluid pressure, deformation is dilatant (i.e. pore space is created and permeability is higher). Thus, zones of localization should attract melt and become the principal ascent conduits in the viscous crust. However, shear zones are weaker than wall rock, a contrast that increases with decreasing depth in the crust. Thus, shear zones might have a tendency to expel melt as the contrast in strength exceeds some critical value. In shallower crust, the location of plutons between faults (Paterson and Schmidt 1999) is consistent with the expectation that where faults overlap the intra-jog region is a mean stress low and potentially a site of melt accumulation and pluton construction (cf. Connolly and Cosgrove 1999).

Temperatures at shallower structural levels in rocks below the solidus may be buffered by crystallization of melt that has escaped from the melting zone at lower structural levels. As a consequence, a popular paradigm for high-T – low-P metamorphism has been advection of heat with migrating magma, the magma having been generated by crustal anatexis in regions of thickened continental crust (DeYoreo *et al.* 1989). The term 'pluton-driven metamorphism' can be used to describe the metamorphic product of such a process; it refers to upper-crustal metamorphism in which the higher-grade metamorphic zones appear to mimic pluton contacts, the implication being that the metamorphism is pluton-related (DeYoreo *et al.* 1991). Such a term is more appropriate than "regional contact metamorphism" (Kays 1970) since magmatic underplating, which is a common paradigm for granulite-facies metamorphism, also can be thought of as regional contact metamorphism. There is no doubt that plutons can be an important contributor to the driving force for high-T – low-P metamorphism, particularly in the upper crust where commonly there is a spatial relationship between plutons and isograds. The maximum thermal effect of pluton-driven metamorphism in the upper crust occurs when intrusions are contemporaneous, but significant heating will occur even when intrusions are temporally separate (Barton and Hanson 1989). If the granites are crustally derived, however, the first order problem is what causes crustal melting, and we should avoid following casually an implied cause and effect (Brown and Solar 1999; Gerdes *et al.* 2000).

For illustrative purposes only, I take as an example the Devonian Acadian metamorphism of the Northern Appalachians, USA (see the papers by Brown and Solar 1998b, 1999; Brown and Pressley 1999; Pressley and Brown 1999; Solar and Brown 1999, 2001a, 2001b). The Acadian metamorphic belt is characterized by elevated modern-day heat flow ($\sim 65\,\mathrm{m\ Wm^{-2}}$) and high heat production ($\sim 3.5 \times 10^{-6}\,\mathrm{Wm^{-3}}$). Metamorphic field gradients suggest high-to-moderate rates of temperature change during metamorphism, but reveal only small variations in pressure. The stratigraphic sequence includes formations with high heat production, a consequence of high U and Th contents fixed in strongly reduced sediments of the precursor anoxic basin (Chamberlain and Sonder 1990). Oblique translation during contractional deformation thickened the stratigraphic sequence and displaced isotherms toward the surface, to create the thermal structure imaged by the 'migmatite front', essentially an isothermal surface given the high $\mathrm{d}P/\mathrm{d}T$ slope of the beginning of melting in most crustal protoliths (Brown and Solar 1999). Melt transport to progressively shallower crustal levels by differential stress-induced processes and buoyancy helps propagate the thermal corridor upward by advecting of heat into the upper crust (Brown and Solar 1999).

Once melt is trapped and crystallized in a pluton, its source can be traced using isotopic fingerprinting, so that we can assess the volume of melt derived from different sources in the crust. This information can be used to determine if the thermal perturbation associated with high-temperature metamorphism extended to the Moho, reflected in voluminous granite magmatism from lower-crustal or mixed (crust and mantle) sources, or was damped in the lower crust.

Again referring to the Acadian metamorphic belt of the northern Appalachians, in western Maine — eastern New Hampshire the absence of a significant volume of granite with a geochemical signature indicating derivation from basement inferred to underlie the Central Maine belt (e.g. Lathrop *et al.* 1996; Pressley and Brown 1999; Brown and Pressley 1999) is consistent with calculated intermediate-to-low reduced heat flow from the lower crust and mantle (Chamberlain and Sonder 1990). This implies low thermal gradients in the lower crust under assumed granulite facies conditions. Thus, Acadian orogenesis involved redistribution of energy and mass within the crust, rather than addition of energy and mass by mantle processes. However, in circumstances where granites image lower-crustal sources, particularly older basement, the heat required for melting implies some mantle involvement, although this may not be a simple underplate (Brown and Dallmeyer

1996; Barboza *et al.* 1999; Handy *et al.* 1999; Leloup *et al.* 1999). Indeed, according to the modeling of Leloup *et al.* (1999), partial melting of the upper mantle during strike-slip displacement along a lithospheric fault zone could occur in the presence of small amounts of fluid, with ascent of the resultant melts along the fault zone contributing to the upward displacement of isotherms and causing crustal melting.

At the implied high geothermal gradients characteristic of collisional orogens, the distribution of strength in the crust is variable according to the magnitude of the geothermal gradient and lithological make-up. Thus, large contrasts in strength may occur across rheologically defined and lithological boundaries (Ord and Hobbs 1989). Lateral variations in geothermal gradient implied by the thermal antiform produce shallow-dipping interfaces between units of highly contrasting strength that abut a zone of plastic yield. Further, anatexis is expected to have a profound effect on the rheology of crust, and hence the deformation history of orogens. The thermal antiform characteristic of regional metamorphic belts would inevitably cause strain localization in the deeper parts of collisional orogens. Pervasive melt flow is possible within the thermal antiform outlined by the 'migmatite front', because of the difference in temperature between melt-producing reactions and the wet solidus (Brown and Solar 1999; Weinberg 1999). Heat advected with migrating melt may promote amplification of this thermal antiform in a feedback relation extending the zone of plastic deformation and melt migration to shallower levels (e.g. Brown and Solar 1999), propagating upward the weakening front and promoting further amplification of the thermal antiform. Culminations in the thermal antiform may be sites of melt accumulation, creating perturbations from which magma may escape to form plutons. Although melting is a consequence of thermal evolution driven ultimately by tectonics, the resultant plutons may effect high-T metamorphism in the upper crust (e.g. Brown and Solar 1999).

In some orogens, anatectic melt has a role in enabling both uplift and exhumation of deep crust (Hollister 1993; Brown and Dallmeyer 1996). If the presence of melt in the middle crust is the facilitator, then the question of what vol.% melt is necessary to cause these effects must be addressed. One area of current interest concerns the use of geophysical observations to quantify the vol.% melt in the crust of active orogens (Partzsch *et al.* 2000), with interpretations calibrated on the basis of laboratory experiments that characterize the petrophysical properties of partially-molten rocks. The inferred minimum melt fraction to satisfy the data in the Central Andes, part of a continental arc, is $\sim$ 15 vol.%, which is four times greater than in either the Himalayan or Pyrenean collision zones, where $\sim$4 vol.% melt satisfies the data. Rushmer (2001) has suggested that the low volume of melt produced in the early stages of biotite dehydration may mean that it becomes trapped along grain boundaries and remains distributed at the grain scale until sufficient volume of melt to allow extraction is generated. This trapped melt may be a mechanism that weakens the crust during orogeny.

5. Criteria used to infer that leucosomes are products of partially molten systems

A primarily anatectic origin for many migmatites is intimated by their macrostructure (e.g. Brown 1973, 1994; Sawyer 1999), revealed by their microstructure (e.g. Cuney and Barbey 1982; Vernon and Collins 1988; McLellan 1989; Harte *et al.* 1991; Brown 1998; Sawyer 1999; Vernon 1999), suggested by geochemical data (e.g., Dougan 1979, 1981; Weber *et al.* 1985; Sawyer and Barnes 1988; Sawyer 1998), permitted by estimates of P-T conditions (e.g. Ashworth 1985; Ashworth and Brown 1990; Vielzeuf and Vidal 1990), evidenced by textures that record dehydration melting (Waters 1988), and indicated by accumulation of quartzofeldspathic material in dilatant sites formed during syn-anatectic deformation of the protolith (e.g. Brown 1994, 1997; Sawyer 1994; Brown *et al.* 1995b; Brown and Rushmer 1997; Snoke *et al.* 1999; Vernon and Paterson 2000).

Two examples of dehydration melting are shown in figures 2 and 3. In the first example, the peritectic product of the biotite dehydration melting reaction occurs in leucosome patches, which suggests some melt retention, although there is no back reaction, whereas in the second example the absence of leucosome suggests most of the melt has been extracted, at least from the volume exposed by the plane of section. Two examples of leucosome accumulation in dilatant sites are shown in figures 4 and 5, in which inter-boudin partitions and dilatant shear surfaces are inferred to have been sites of lower pressure into which melt accumulated.

Leucosome geometry in these migmatites is inferred to record evidence of active melt flow, rather than stagnant melt, if:

- bulk rock compositions are consistent with depletion in felsic components, implying that leucosome was not simply a product of *in situ* segregation during partial melting; and
- leucosomes preserve textures resulting from the presence of melt, such as euhedral phenocrysts/peritectic phases, textures that mimic solid-melt relations (quartz/feldspar films along grain boundaries, interstitial-xenomorphic

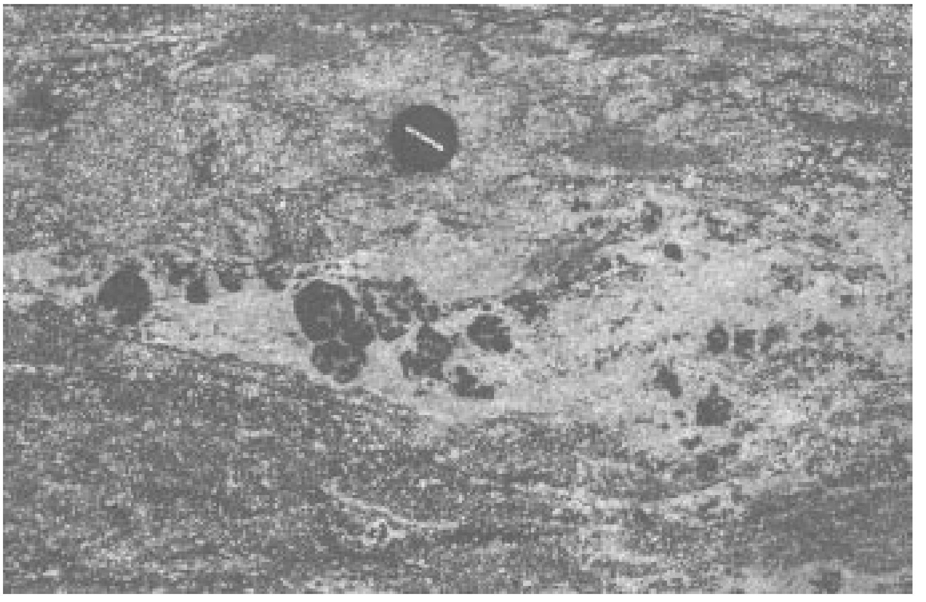

Figure 2. Dehydration melting in migmatitic granulite facies paragneisses. In this example, the peritectic product of the biotite dehydration melting reaction, euhedral garnet, occurs in leucosome patches, although there is no back reaction, which suggests accumulation of cumulate minerals and melt loss. Enseada Azul Beach, Guarapari, Espirito Santo, Brazil.

Figure 3. Dehydration melting in migmatitic granulite facies paragneisses. In this example, the absence of discrete leucosome suggests the melt was only locally segregated into patches, but segregation was sufficient that back reaction, although evidenced by the thin biotite rind on garnet, was minimal. Enseada Azul Beach, Guarapari, Espirito Santo, Brazil.

quartz, residual quartz in feldspar), melt-solid reaction textures and fractured residual grains (e.g. Cuney and Barbey 1982; Vernon and Collins 1988; Powell and Downes 1990; Harte *et al.* 1991; Ellis and Obata 1992; Schnetger 1994; Nyman *et al.* 1995; Brown and Dallmeyer 1996; Hartel and Pattison 1996; Carson *et al.* 1997; Brown 1998; Sawyer 1999, 2001; Vernon 1999; Watt *et al.* 2000; Marchildon and Brown 2001; and Brown 2001).

The simple presence of euhedral crystals alone clearly is not sufficient, since euhedral crystals can form in sub-solidus conditions as well as in the presence of a melt phase, and textural analysis in migmatites remains an uncertain science in which care must be exercised to distinguish textures related to melting from those produced by subsolidus processes alone (e.g. Dallain *et al.* 1999; Vernon 1999).

Figure 6 shows a hand sample of a stromatic migmatite in which smaller leucosomes appear to be crystallized from melt derived locally, whereas the larger leucosome on the left-hand side likely records evidence of melt moving through the

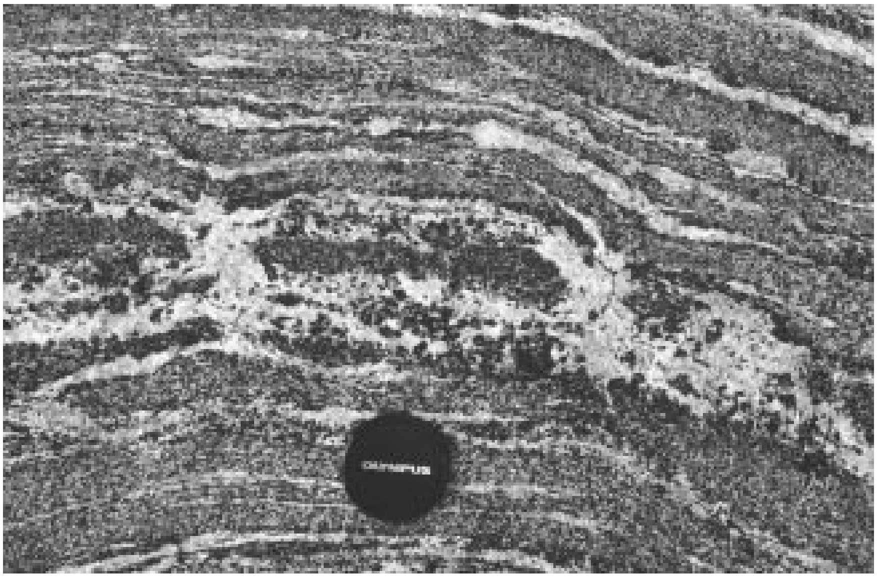

Figure 4. An example of leucosome accumulation in inter-boudin partitions, sites that are inferred to have been of lower pressure and in which melt therefore accumulated. Tolstik Peninsular, Karelia, Russia.

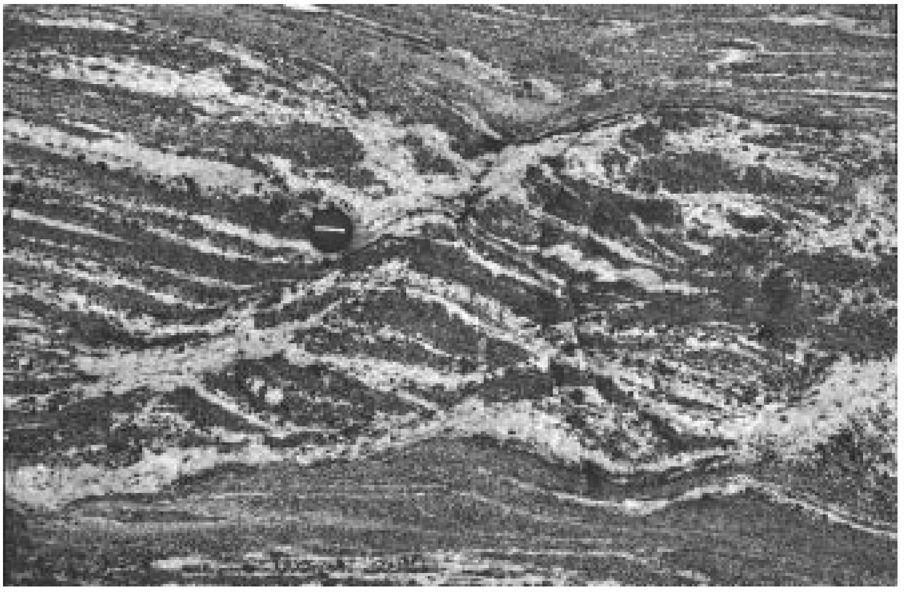

Figure 5. An example of leucosome accumulation in dilatant shear surfaces, sites that are inferred to have been of lower pressure and in which melt therefore accumulated. Tolstik Peninsular, Karelia, Russia.

system; a representative view of the microstructure of the larger leucosome is shown in figure 7. Figure 8 shows contact migmatites from a middle crustal metamorphic aureole; the mesosome between the leucosome veins shows evidence of having been partially melted, such as the K-feldspar patches shown in figure 9, although it is likely that the melt did not segregate into mesoscopic patches or veins.

Some migmatites exhibit evidence of melt-enhanced embrittlement (e.g. Davidson *et al.* 1994) and leucosomes commonly show microstructures that suggest flow in the magmatic state (e.g. Blumenfeld and Bouchez 1988; Sawyer 1996; Brown and Solar 1998a, 1999). However, leucosome modes and/or chemical compositions commonly are interpreted to reflect dominance of cumulate and/or residual solid phases after melt escape, rather than melt compositions (e.g. Cuney and Barbey 1982; Powell and Downes 1990; Ellis and Obata 1992; Carson *et al.* 1997; Solar and Brown 2001b). Examples of deformed migmatite terrains in which the mineral assemblages and geochemical data suggest that melt loss has occurred are common (Weber *et al.* 1985; Powell and Downes 1990; Ellis and Obata 1992; Nyman *et al.* 1995; Hartel and Pattison 1996; Kriegsman 2001a, 2001b; Solar and Brown 2001b). In one example, Hartel and Pattison (1996) estimate 5–30 vol.% melt loss from garnet amphibolites by comparison between observed

Figure 6. A hand sample of a stromatic migmatite in which smaller leucosomes on the right-hand side appear to be crystallized from melt derived locally, whereas the larger leucosome on the left hand side likely records evidence of melt moving through the system. Stromatic migmatite from South Brittany, France.

volume of leucosome and volume of melt expected based on mass balance calculations. This estimate is similar to those derived from experimental petrology for formation of tonalite-trondhjemite-granodiorite melts (e.g. Rapp and Watson 1995, 20–40 vol.%) and those based on REE patterns of tonalite-trondhjemite-granodiorite suites in some Archean terranes (e.g. Luais and Hawkesworth 1994, ~ 25 vol.%). Thus, migmatitic garnet amphibolites may be representative of partially depleted source rocks for tonalite-trondhjemite-granodiorite suites. The critical point, however, is that although melt may have been lost from the system investigated by Hartel and Pattison (1996), not all of the melt generated was expelled; the partially depleted garnet amphibolites exhibit continuous concordant and discordant leucosomes that preserve the crustal plumbing that allowed melt loss to occur.

6. Leucosome as evidence of a remnant permeability network

Melt segregation, formation of permeability networks and melt extraction from crust in orogens occurs synchronously with deformation, as evidenced by field examples (Brown 1994; Brown and Rushmer 1997; Marchildon and Brown 2001). These processes have been investigated using analog models (Rosenberg and Handy 2000a, 2000b). The modeling emphasizes the importance of dilatant shear surfaces as sinks for melt and as important escape channels. It has been suggested (Brown and Solar 1998a, 1999; Brown *et al.* 1999) that

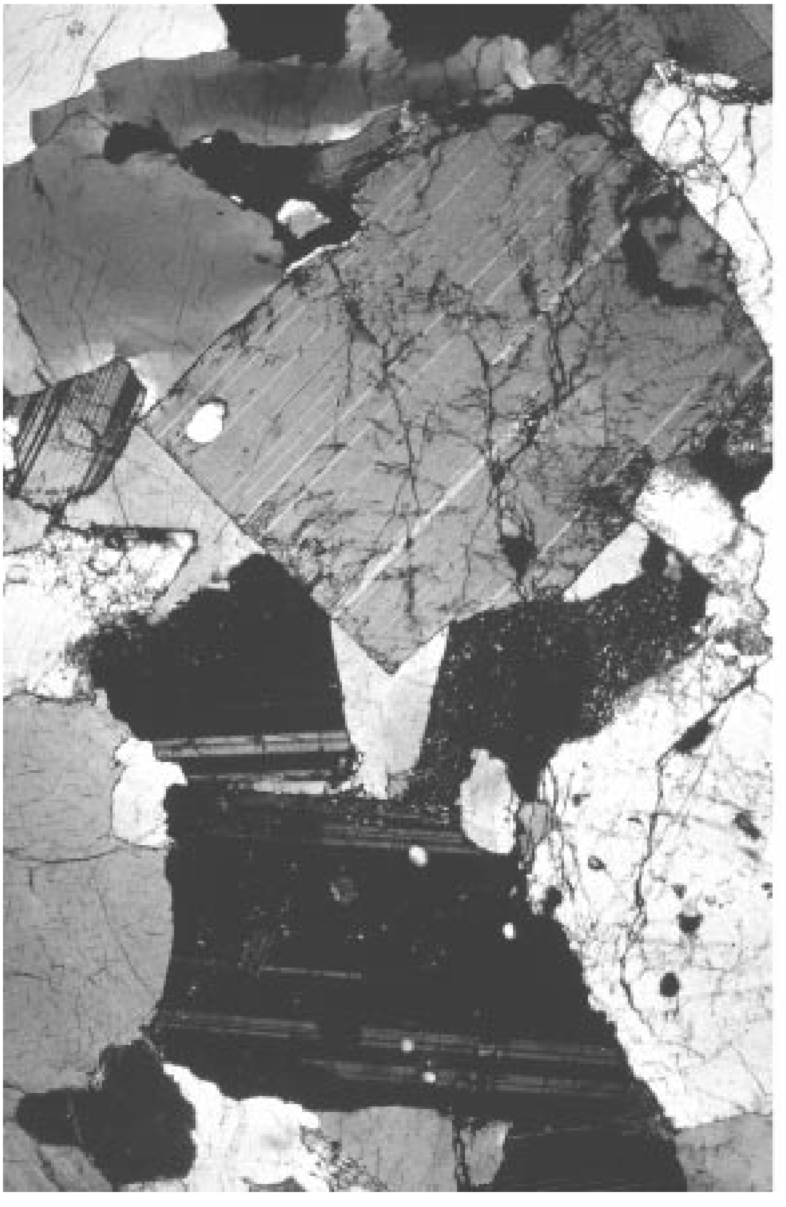

Figure 7. Microstructure of the larger leucosome shown in figure 6. Note the euhedral shape of the plagioclase phenocrysts against intersertal quartz.

melt migration pathways in migmatites relate to rock fabrics, although this may not be the case in all syntectonic vein networks (Handy *et al.* 2001). Some migmatites are depleted in a component equivalent to leucogranite, and these migmatite terrains may represent sources of leucogranite (Sawyer 1998; Milord *et al.* 2001; Solar and Brown 2001a).

In a recent study of the 3-dimensional topology of leucosome in a stromatic migmatite from the migmatitic core of the southern Brittany metamorphic belt (Brown M A *et al.* 1999), garnet crystals in leucosome are only locally in contact with the melanosome or mesosome; apparently they are never in contact with melanosome or mesosome on both sides of the leucosome. Jones and Brown (1990) suggested that one reaction responsible for the production of melt was Bt + Als + Qtz (+ Pl) $\rightarrow$ Grt $\pm$ Kfs + L. The garnet observed in the leucosomes is likely to be the product of this moderate-to-low $a(H_2O)$, incongruent, biotite-dehydration melt-producing reaction, which suggests that its present geometry represents suspension in melt. It could be argued that the planarity of the leucosomes is not a primary feature of the melt flow conduits but a secondary feature due to subsolidus deformation. This is unlikely to be the case because the microstructure of the leucosomes is largely magmatic, with some plagioclase crystals that have euhedral facies against quartz and some quartz that occurs as grain boundary films between plagioclase crystals; quartz shows only slight undulose extinction. Based on these observations, we infer that any post-crystalline defor-

Figure 8. Contact migmatite from a middle crustal metamorphic aureole, Onawa, Maine. Note lit-par-lit leucosomes parallel to compositional layering and cross-cutting leucosomes discordant to composition layering.

mation has not modified the leucosome structure. However, there is no back-reaction between garnet and leucosome, which suggests that the leucosome composition does not represent the putative melt composition produced by the biotite dehydration melting reaction.

Based on the work of Brown M A *et al.* (1999), we may regard the leucosome retained in migmatites as recording the remnant permeability network, but how permeability networks evolve and how anomalies in permeability amplify during collisional orogenesis are poorly understood. However, for the specific case of crustal underplating by mafic melts, Petford and Koenders (1998b) have shown how isolated fractures that form during rapid thermal perturbation in a source can combine to form a single, interconnected structure with high permeability by self-organization. Further, Miller and Nur (2000) have pointed out that local permeability can change instantaneously from one extreme to the other as a consequence of the fluid pressure evolution of the system leading to pore pressure increases sufficient to induce hydrofracture. Miller and Nur have used a cellular automaton model driven by an internal fluid source that incorporates a toggle-switch permeability to investigate how such a system self-organizes and develops to the critical state. Modeling ultimately will enable us to understand how permeability networks evolve (e.g. Petford and Koenders 1998b; and Miller and Nur 2000).

7. Cyclic or continuous melt loss

Most silicate melts wet the grain boundaries of the solid grains under static and dynamic condi-

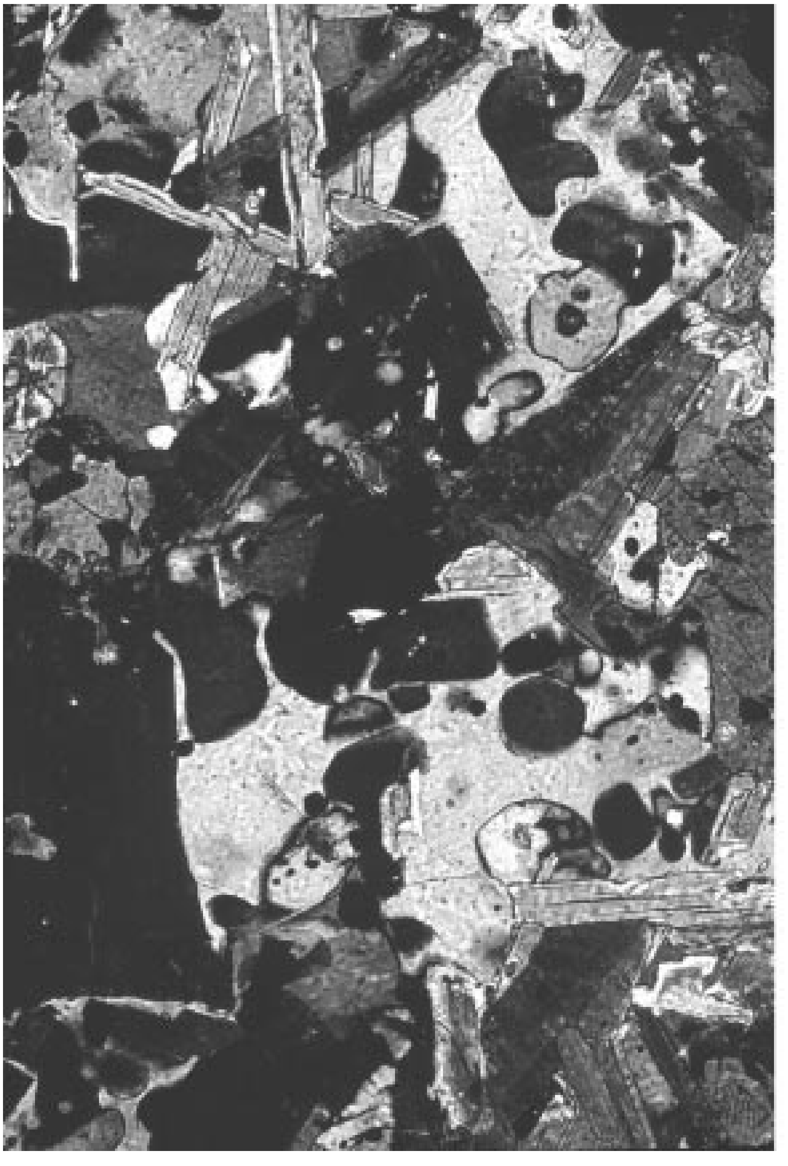

Figure 9. The mesosome between the leucosome veins in the migmatite of figure 8 shows evidence of having been partially melted, such as the K-feldspar patches (light gray) shown here, although the melt apparently did not segregate into mesoscopic patches or veins.

tions (Laporte *et al.* 1997; Brown and Rushmer 1997; Rosenberg and Riller 2000). Although the melt enhances diffusive mass transfer and accommodates granular flow of the solid grains (Paterson 2001), the melting rock initially behaves as a closed system and no melt is extracted. As melting progresses, during syntectonic anatexis, the protolith weakens and the solid material may deform by melt-assisted creep if the melt is retained. At an appropriate strain rate, at some strain determined by properties such as grain size, rate of melting and melt volume, buildup of melt pressure leads to formation of dilatant structures into which melt will flow down gradients in melt pressure (Brown and Rushmer 1997; Vanderhaeghe 1999). For example, nucleation of dilatant shear surfaces on melt pores localizes the strain, and these developing shear zones suck in melt from the surrounding matrix to further weaken the system. If such shear zones interconnect, then melt flows through the system, it becomes open and melt loss from the system causes strain hardening (Handy *et al.* 2001). This cycle of melt accumulation and melt loss repeats itself, and the system alternates between closed and open (Brown and Rushmer 1997; Brown and Solar 1999; Miller and Nur 2000).

In order to understand the migration of melt and the strength of partially molten rocks in orogens, it is necessary to understand the hydraulic transport properties of these materials and how those properties vary with space and time (e.g. Renner *et al.* 2000). Melt migration in the anatectic zone by pervasive flow (Weinberg 1999; Brown and Solar 1999; Leitch and Weinberg 2000) is not problematic, as Tanner (1999) has shown, although pervasive flow is limited by congelation when intruding cold country rock. At much lower strain rates, melt is extracted from the system fast enough that melt pressure does not build up sufficiently to allow magma fracture (Rutter 1997; Handy *et al.* 2001). Thus, low strain rates favor preservation of compositional layering, and dilatant structures, which fill with melt and disrupt layering, are unlikely to form. In this case, a melt-depleted granulite is the likely residue.

8. The metamorphic volatile phase

The presence or absence, nature and composition of the metamorphic volatile phase are important controls on what happens along both the prograde and retrograde portions of the *P*-*T* path. Below the wet solidus, devolatilization reactions contribute products to a sporadic, spatially heterogeneous volatile phase. It is likely that the activities of components in the volatile phase are lithologically controlled, and flow either is lithologically or is structurally controlled. In contrast, a decrease in $a(H_2O)$, as well as an increase in T, and the onset of dehydration melting reactions characterize the transition from the amphibolite to the granulite facies. This precludes the general presence of a ubiquitous, pervasive metamorphic volatile phase (Yardley and Valley 1997), a conclusion that is supported by the H_2O-undersaturated nature of high-T crustal melts (Clemens and Watkins 2001). The lowered $a(H_2O)$ could be achieved even in the presence of a metamorphic volatile phase as long as the quantity of this phase is limited, such that the system is rock-dominated rather than fluid-dominated, or in the presence of an unlimited quantity of a volatile phase that is dominated by components other than H_2O, although both of these scenarios preclude melting. Melting is inferred to have occurred in most granulite facies terrains, whether the rocks have a migmatitic appearance or not (Brown 1994). Following this logic, Clemens and Watkins (2000) have argued that the correlation between melt T and initial H_2O content requires a rock-dominated, reaction buffered system. In this system, the volatile components are dissolved in the granite melt and transported with it to be exsolved upon crystallization. Thus, although segregation of melt from residue by distances greater than the equilibration volume may prevent back reaction from occurring, crystallization of the melt may promote retrogression in the surrounding host.

One feature common to many high-grade metamorphic terrains is the occurrence of leucosomes throughout the deformation history (e.g. Brown 1978; Fleming and White 1984; Jones and Brown 1990; Mogk 1990, 1992; Stevens and Van Reenen 1992; Brown and Dallmeyer 1996; Collins and Sawyer 1996). This ubiquitous association of anatectic leucosome with each phase of deformation may have implications concerning recycling H_2O internally during orogenesis in a melting/movement/congelation positive feedback relation driven by deformation. Thus, melting may be initiated, for example by localization of deformation or migration of H_2O into a shear zone, and dissolved H_2O may migrate with the melt to be exsolved as the magma congeals in a structurally controlled site somewhere else. This exsolved H_2O may initiate a new cycle of melt generation and migration, effectively recycling the H_2O, and the process may be repeated several times (e.g., Mogk 1990, 1992). Recycling of the metamorphic volatile phase in this manner has been suggested for several terrains, including the Thor-Odin metamorphic core complex of British Columbia (Holk and Taylor 1997), the Limpopo Belt of South Africa (Stevens 1997) and the Central Maine belt of New England, USA (Solar and Brown 2001b).

9. Leucogranites in migmatite terrains

If migmatites represent a crustal record of regional melting events, we must consider whether there is a relationship between migmatite leucosomes and residual host material, and granite plutons. Depending upon the composition of the protolith, whether there is any influx of a H_2O-rich metamorphic volatile phase, the rate of heating, the type of deformation and rate of strain and the maximum temperature achieved, as rocks cross the solidus they evolve sequentially from low melt fraction metatexites to high melt fraction diatexites, and to *in situ* granites if the melt is not extracted. Once the melt is extracted igneous processes may affect the magma leading to evolved granites typical of the upper crust. Factors that influence the composition of the melt include the composition of the source, the *P*-*T* of melting and the melt-producing reaction, the distribution of accessory phases, whether they are on grain boundaries or within grains, and the lack of equilibrium during the melting process.

The occurrence of leucosome in metamorphic fabrics and in dilational sites within plastically deformed host rocks suggests it records syntectonic melt flow through deforming crust (Brown 1994; Brown and Rushmer 1997; Brown and Solar 1999), although leucosome compositions do not preserve melt compositions. Thus, leucosome in migmatite provides a record of the melt flow paths, and the leucosome structure in migmatite may provide a good analog for the melt flow network in the source

(Brown M A *et al.* 1999), particularly if the process is scale invariant (Tanner 1999). In some cases, the form of magma ascent conduits apparently was deformation-controlled (Brown and Solar 1999), and was governed by the contemporaneous strain-partitioning characteristic of most orogens (Solar and Brown 2001a). Although migmatite terrains commonly have geochemical compositions consistent with melt loss (Brown *et al.* 1995a; Brown and Rushmer 1997; Solar and Brown 2001b), some schlieric migmatites reflect melt redistribution and accumulation (Sawyer 1996, 1998; Milord *et al.* 2001). Further, the cyclic behavior of melt segregation and escape leads to a pulsed flux of melt through and out of the anatectic zone (Brown and Solar 1999; Solar and Brown 2001b), consistent with the internal structure and geochemistry of many granite plutons (e.g. Brown *et al.* 1981; Brown and Pressley 1999; Pressley and Brown 1999).

Based on a variety of geochemical evidence, the leucosomes in many anatectic migmatites likely are the result of accumulation of crystals, both residual and newly crystallized (cumulate), during escape of residual melt (Marchildon and Brown 2001; Solar and Brown 2001a). As a direct complement, many large-volume upper-crustal leucogranites apparently represent evolved melt compositions. Thus, some migmatite terrains may represent pluton sources, and what we see at any particular crustal level depends on what gets stuck on its way through the system and how evolved this magma had become at the point of arrest (e.g. Sawyer 1998; Solar and Brown 2001b). What we do not see, however, are magma chambers in the source.

This process can be quantified by estimation of the volume of leucosome in residual migmatite terrains, knowledge of the bulk chemistry of the migmatites and the putative protolith, and estimation of the volume of evolved granite at shallower structural levels. For example, in the Acadian metamorphic belt of the northern Appalachians, comparison of the bulk chemical composition of the protolith metasedimentary rock succession and residual migmatites suggests an average melt loss of $\sim$ 20 vol.%. In these migmatites, leucosomes vary from probable melt compositions to compositions that imply a dominantly cumulate mineralogy, whereas small-volume granites vary from melt compositions to compositions that imply a cumulate mineralogy with entrained residue (Solar and Brown 2001a). Common leucogranites in one larger-volume granite studied in detail (Pressley and Brown 1999; Brown and Pressley 1999) have a range of compositions consistent with being derived by up to $\sim$ 20 vol.% fractional crystallization of an anatectic melt derived from a source similar to the enclosing metasedimentary rocks. This suggests a relationship between crustal melting, leucosomes and smaller-volume granites that vary from melt to cumulate compositions, and larger-volume granites that vary from primary to evolved melt compositions.

Based on the example above, I calculate that a pluton of volume $\sim 1750\,\mathrm{km}^3$ crystallized from a liquid with an evolved composition, derived by ~ 20 vol.% fractional crystallization of ~ 20 vol.% anatectic melt, requires a source volume of $\sim 11,000\,\mathrm{km}^3$. Put another way, a horizontal semi-circular half-cone-shaped pluton of diameter ~ 30 km and half height ~ 5 km (figure 10a) can be derived by $\sim$ 20 vol.% fractional crystallization of $\sim$ 20 vol.% anatectic melt segregated from a ~ 15 km thick source of horizontal diameter ~ 30 km (figure 10b). One implication of such a model, for the assumptions made, is that plutons derived in this manner might be spaced $\sim$ 30 km apart. A symmetric pluton (e.g. horizontal circular cone-shaped) of the same volume has a horizontal diameter of ~ 21 km, but requires the same source volume and spacing (figure 10b). For a pluton of four times greater volume (e.g. a horizontal tabular pluton of diameter ~ 30 km and thickness $\sim$ 10 km, figure 10(c), or horizontal circular cone-shaped pluton with diameter 42 km and half height ~ 5 km), the horizontal diameter of the source, for the same assumptions, including a source thickness of $\sim$ 15 km, becomes $\sim$ 60 km (figure 10d), which doubles the projected spacing of plutons. One implication of these geometric relationships between pluton and source is that flow through the storage porosity in the source is predominantly horizontal (or shallowly upward) rather than vertical, and flow is towards the ascent column, or radially to the center of the source for the simple cylinder geometry used above. A more sophisticated analysis of pluton-source shape and volume, and melt extraction and accommodation is given in Cruden and McCaffrey (2001).

10. The switch from magma ascent to emplacement during orogeny

In collisional orogens, although regional stress and buoyancy drive melt flow, ascent of magma commonly is controlled by the structures, particularly those that localize strain such as dilational shear zones. Melt flow networks require sustained transient near-lithostatic to supra-lithostatic melt pore pressure during deformation to prevent draw down, decreased transport rates and possible collapse. Assuming availability of melt in the storage porosity provided by the anatectic zone, limited by the granite wet solidus, and melt flow out of this source by pervasive and channeled mechanisms

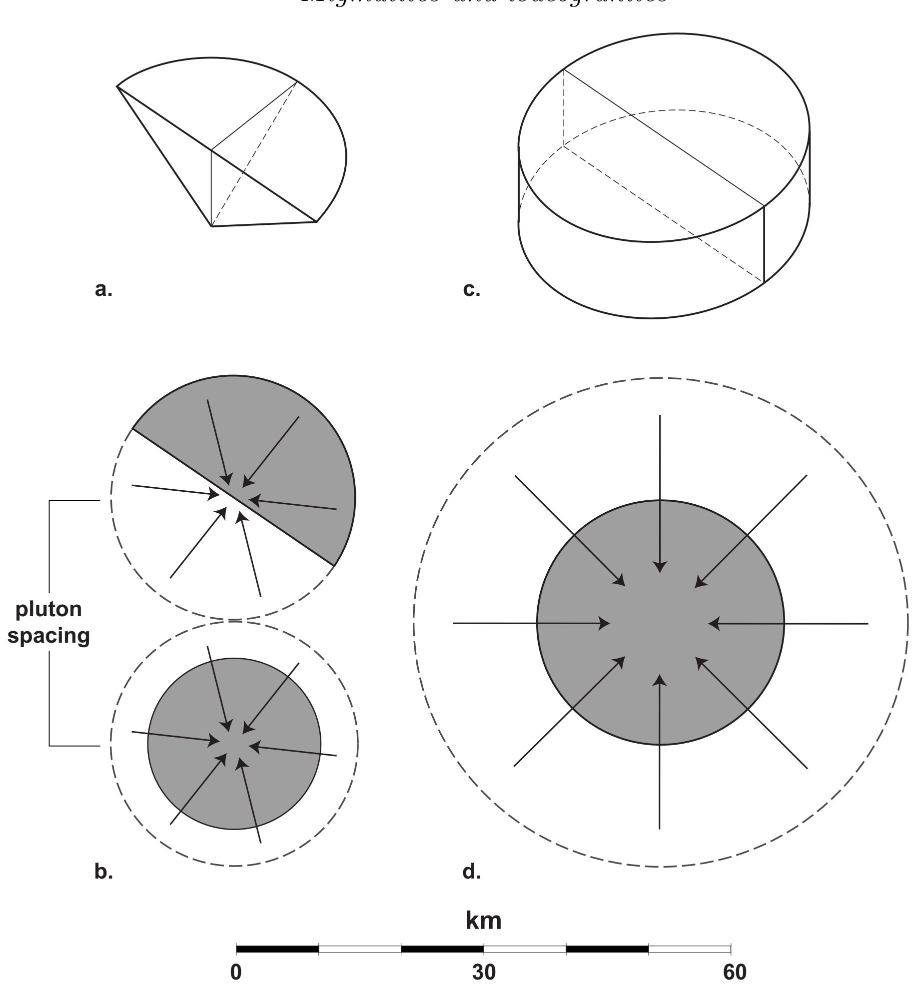

Figure 10. Pluton-source relations. **(a)** Perspective view of horizontal semi-circular half-cone-shaped pluton of diameter ~ 30 km and half-height ~ 5 km. **(b)** Plan view of horizontal semi-circular half-cone-shaped pluton of diameter ~ 30 km and horizontal circular cone-shaped pluton of diameter ~ 21 km in relation to a source of sufficient volume to fill the pluton (~ 30 km diameter and ~ 15 km thick). **(c)** Perspective view of horizontal tabular pluton of diameter ~ 30 km and thickness ~ 10 km. **(d)** Plan view of horizontal tabular pluton of diameter ~ 30 km in relation to a source of sufficient volume to fill the pluton (diameter ~ 60 km and ~ 15 km thick).

through an ascent column, we may ask: What controls the switch from magma ascent to emplacement, and how does this happen? To answer this we must realize the coupled nature of orogenic systems, with feedback relations among stress, deformation, thermal evolution, strength, strain rate, melt flow, etc. These systems are deterministic at large length scales, i.e., at the scale of the orogen itself, controlled by the external forces acting on them, but stochastic at small length scales, i.e., at the scale of part of the crust or the pluton, governed by the laws of probability. In such non-linear dynamical systems, large effects are generated from small fluctuations. The approach adopted here reconciles into one general model contradictory views put forth by proponents of the dike – horizontal tabular pluton, viscous flow in structurally-controlled channels – composite pluton with root and horizontal tabular head, and visco-elastic magmatic diapir ascent and emplacement models.

The perception that emplacement occurs at a 'level of neutral buoyancy' is inconsistent with the density of granite liquid (~ 2.44 kg m^3 at 1 GPa, 800°C) and the range of emplacement depths. Further, the generalization that emplacement is controlled primarily by structural interactions between ascending melt and anisotropies in upper crust is only part of the story. Although plutons 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 and Fowler 1993; Paterson and Vernon 1995; Paterson *et al.* 1996), the systematic variation in 3-D shape with depth, from horizontal tabular to blob-like to vertical lozenge, suggests that most emplacement occurs by two principal mechanisms, according to host rock behavior (a function of thermal gradient, strain rate, etc.). These are: in the brittle regime, by vertical inflation (lifting the roof/depressing the floor, accommodation mechanism(s) unspecified) after horizontal flow in a fracture or along a pre-existing anisotropy; and, in the viscous regime, by lateral expansion (swelling out like a balloon, accommodation mechanism(s) undefined) localized by some instability in the magma - wall-rock system.

In the brittle regime, magma may be arrested by a structure or 'crack stopper', some instability or thermal death (Clemens and Mawer 1992; Brown and Solar 1998b). Emplacement occurs when mainly vertical flow switches to predominantly horizontal flow. Analysis of the length *vs.* thickness of horizontal tabular plutons suggests they inflate according to a power law relationship, interpreted to mean that vertical thickening only occurs after magma has traveled horizontally some critical minimum distance (McCaffrey and Petford 1997; Cruden 1998; Petford *et al.* 2000). Depression of the floor and/or lifting of the roof allow inflation to be assimilated (Cruden 1998; Brown and Solar 1998b). Although sagging of the floor can be accommodated because magma has been extracted at depth, no simple relationship exists since the volume of melt in the pluton is from a much larger source volume (although see Cruden 1998; Cruden and McCaffrey 2001). To lift the roof, P_{melt} must overcome lithostatic load and tectonic overpressure.

In the viscous regime, amplification of naturally occurring 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 the host rock. These instabilities are not mutually exclusive, and feedback relations are likely to produce similar results whatever the initial instability. Consider an ascent column with zones of higher permeability; these will be conduits of higher magma flux. However, higher flux may lead to increased permeability, and increased heating and consequent weakening of the host rock surrounding these conduits, which in turn increases the strain rate; all of these interactions lead to even higher magma flux, and so on. If a feedback relation of this kind indeed exists in nature, it 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 and forming a vertical lozenge pluton. The 'diapiric' intrusions of Miller and 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/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 enables further lateral expansion, forming a blob-like pluton. The nested 'diapirs' of Paterson and Vernon (1995) may have been emplaced by such a mechanism.

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/narrower parts of the conduit, flow will focus in the faster/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-to-viscous transition depth-interval may decrease to less than that of the over- and underlying crust (Handy 1989; Handy *et al.* 2001). This strain weakening is attributed to transitions from cataclasis (just above the initial brittle to viscous transition) or grain-size insensitive (dislocation) creep (just below the initial brittle-to-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). This strain-dependent strength minimum just below the brittle-to-viscous transition has important implications for arresting the ascent of magma and enabling lateral emplacement (Handy *et al.* 2001). At this crustal level, the switch to emplacement may lead to formation of a pluton with a horizontal tabular head, formed by inflation of a sub-horizontal magma fracture, and a hemi-ellipsoidal root that passes down into a migmatite zone through which magma was transferred (Brown and Solar 1999). In such a composite pluton, the hemi-ellipsoidal root is thought to have formed by back filling and freezing as the thermal structure decayed due to 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 pluton is insufficient to maintain continued inflation and the resultant feedback ultimately terminated magma ascent and extraction from the source. This is consistent with the expected decline in extraction rate with time from a finite volume source.

11. Rates of processes

We may separate the processes of melt generation and melt segregation, and magma extraction, ascent and emplacement. However, this is not meant to imply a simple succession of events, and in any case feedback relations are expected among these processes and with the applied differential stress. For example, in principle, the relative rates of melt generation and segregation will determine whether the melt is able to separate from the residue above some percolation threshold. However, in collision zones these processes are associated with deformation, and the rate of melt pressure buildup may determine by what mechanism and at what rate the melt segregates. Further, we may consider the relative rates of melt segregation and extraction to be the control on whether migmatites or depleted granulites form, or whether sufficient melt escapes to form plutonic bodies of granite. However, the cumulate mineralogy of many migmatite leucosomes suggests that crystallization will begin before melt escape from the anatectic zone is achieved. Also, the rates of melt segregation and extraction will be influenced by the rates of melt generation and magma ascent. There is a feedback relation between the rate of magma emplacement and the rate of ascent, with fast ascent potentially driving fast birth and inflation of plutons (McCaffrey and Petford 1997; Cruden 1998). Conversely, at the limit of maximum pluton inflation, i.e. when the pluton cannot grow further for whatever reason, feedback must slow down the ascent rate. The rate-limiting step in pluton formation is unlikely to be emplacement rate, ascent rate or even extraction rate, but is most likely to be related to the rates of generation and segregation, which will be controlled by the thermal evolution and the deformation.

What are these rates (see also Thompson 1999)? Based on detailed regional studies of crystallization ages in migmatites and granites, we know that the whole process, from the peak of metamorphism to pluton emplacement, can be accomplished on a time scale of less than 1 Ma (e.g. Solar *et al.* 1998). The rate of melt generation will be limited by heat flow, but heat flow is only likely to be a rate-limiting step at short time scales (Rutter 1997). Further, calculation of volume change associated with melting in experiments on muscovite-, muscovite + biotite-, and biotite-bearing rocks suggests that muscovite-bearing assemblages, with or without biotite, have a high enough positive volume change to enable magma fracture, whereas biotite-bearing assemblages have negligible volume change (Rushmer 2001). Muscovite dehydration melting reactions occur over a narrow temperature range. In the experiments on protolith schists from the Himalayas, Patiño Douce and Harris (1998) suggest that $\sim$ 20 vol.% melt is generated within 25°C rise in T above the solidus, with a higher vol.% melt generated if H_2O is added to the system. This estimate is consistent with estimates of melt loss based on the depletion in some migmatite terrains (e.g. Solar and Brown 2001b). Muscovite dehydration melting produces a significant pulse of melt that will increase melt pore pressure rapidly. Thus, lower temperature crustal melting involving muscovite dehydration may be associated with fast melt segregation and magma extraction, whereas higher temperature crustal melting involving biotite dehydration may lead to melt retention at lower vol.%, unless deformation increases the rate of melt pore pressure buildup (cf. Rushmer 2001).

There are many variables that contribute to an evaluation of the rate of segregation of melt from partially-molten crust (Rutter 1997), including protolith grain size, H_2O content and applied differential stress. At lower differential stresses and higher T, shear-enhanced compaction dominates the segregation process, whereas at higher differential stresses and lower T, extraction occurs more efficiently via vein networks. Nonetheless, to extract melt in a geologically reasonable timescale a combination of grain-scale flow, driven by gradients in melt pressure, to veins and flow through veins, driven by buoyancy forces, is required. For an appropriate grain size (5 mm), segregation of 10 vol.% melt from a segment of crust by porous flow to veins by shear-enhanced compaction and flow of melt through veins likely takes place on the order of $10^3 - 10^5$ years, implying bulk strain rates in the range $10^{-12} - 10^{-14}$ s^{-1}. This estimate is consistent with timescales of $< 10^3$ based on accessory phase dissolution (Harris *et al.* 2000).

Once melt is segregated into veins it represents a source to support extraction. The rate of melt extraction potentially may be significantly faster that segregation, supported by the melt stored in veins, particularly if migmatites have a fractal structure similar to the example studied by Tanner (1999). In this example, the vein network has a fractal dimension similar to the Menger Sponge. In a Menger Sponge, melt must travel a distance twice the width of the vein to reach a vein of three times the size. For an initial vein of several centimeters width and an appropriate melt viscosity, it is clear that melt accelerates through the anatectic zone as increasing channel widths allow faster rate of flow. The Menger Sponge model is an analog to the pervasive flow mechanism of Brown and Solar (1999), Vanderhaeghe (1999) and Weinberg (1999).

There are many variables also that control the rate of ascent. Typical ascent rates for crystal-free magma in a dike of several meters width are on the order of 0.1 m s^{-2} (Clemens 1998), which means that a pluton of volume $\sim 1{,}000\,\mathrm{km}^3$ can be filled in a time as short as $\sim 1{,}000$ years (if both the extraction rate and the inflation rate can keep pace). Cruden (1998) estimates that a ~ 3 km thick horizontal tabular pluton of diameter

10–100 km will fill on a time scale of 100 years to 1 million years, with an average host strain rate to accommodate the emplacement in the range 10^{-10} – 10^{-15} s^{-1} (see also Petford *et al.* 2000; Cruden and McCaffrey 2001). Harris *et al.* (2000) argue that melt ascent will be extremely fast, occurring on a timescale possibly as short as ~ 10 yr. According to Clemens (1998), such a ~ 3 km thick pluton in the upper crust will cool below the solidus at its center in a time of $\sim 30,000$ years.

What are the implications of fast timescales for melt segregation, and magma extraction, ascent and emplacement? First, the overall rate-determining step in crustal melting and the formation of granite plutons will be heat flow. Second, the rapid timescales imply that this equilibrium between granite magma and its source will be preserved (e.g. Hammouda *et al.* 1996; Davies and Tommasini 2000) and reflected in the heterogeneous isotope compositions of plutons (e.g. Deniel *et al.* 1987; Pressley and Brown 1999). It follows, that one popular assumption underlying many geochemical studies of granites, that the magmas image the source region, may not be reasonable.

12. Conclusions

Granite magmatism represents a major crustal recycling mechanism in orogens and the principal mechanism by which the geochemical differentiation of Earth's crust has happened since the Archaean. The intimate link between orogeny and crustal melting requires an interdisciplinary approach in which field observations, experimental investigations and theory from diverse disciplines including geophysics, geochemistry, petrology, tectonics and modeling contribute. This interdisciplinary research in orogenesis and crustal melting is leading to significant advances in understanding.

Examples of such advances are:

- The improvement in geophysical techniques and the development of the petrophysical basis for the interpretation of geophysical data, which have enabled us to infer the presence of melt in the middle crust of active orogens and to place limits on the amount of melt present in the crust.
- The revelation that melt generation and segregation, and magma extraction, ascent and emplacement generally are syntectonic processes involving feedback relations, and that rates and timescales likely are much shorter than once believed.
- A better understanding of the feedback relations among the various processes involved, particularly the thermal structure and deformation. Compatibility is maintained by these feedback relations, exemplified by the fact that rates of melt segregation fall within the range of strain rates for host rocks to accommodate pluton emplacement.
- The use of more sophisticated analog and numerical models in understanding the interrelationships between deformation and melt migration.
- Understanding the switch from ascent to emplacement as a process controlled by various interrelated factors, but in which the style of this switch and the accommodation mechanisms of emplacement relate to the depth of the viscous-to-brittle transition in the crust.
- A better understanding of the rheology of partially molten crust, and its role in orogenic collapse.
- The use of geochemistry as a discriminator between competing hypotheses; for example, concerning whether or not there is a relationship between migmatites and granites.

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