

# Retrograde processes in migmatites and granulites revisited

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**ABSTRACT** Many migmatites and granulites preserve evidence of a clockwise  $P$ – $T$  evolution involving decompression (decrease in  $P$ ) while close to the thermal peak. The extent of post-thermal peak reaction is influenced by several factors, including: (1) the  $P$ – $T$  path in relation to invariants in the system and the Clapeyron slopes of the equilibria; (2) the rate of cooling; and (3) the availability of fluid ( $H_2O$ -rich volatile phase or melt) for fluid-consuming reactions. Reaction may occur between products of a prograde (increasing  $T$ ) fluid-generating reaction as the same equilibrium is re-crossed in the retrograde (decreasing  $T$ ) sense. In general, reaction reversal or ‘back reaction’ requires the  $P$ – $T$  path to approximate isobaric heating and cooling, without significant decompression, and evolved fluid to remain within the equilibration volume. The larger the decompression segment in the  $P$ – $T$  evolution, the more chance there is of crossing different reactions along the retrograde segment from those crossed along the prograde segment. For common pelite compositions, we may generalize by considering three pressure regimes separated by the [Spl, Ms,  $H_2O$ ] invariant in KFMASH (approximately 9 kbar) and the intersection of muscovite breakdown with the  $H_2O$ -rich volatile phase-saturated solidus (approximately 4 kbar). Reaction reversal cannot occur along  $P$ – $T$  paths that traverse around one of these points, but may occur along  $P$ – $T$  paths confined to one of the three regimes in between. Additionally, above the solidus, melt segregation and loss potentially change the composition of the equilibration volume; and, the size of the equilibration volume shrinks with decreasing  $T$ . Since the proportion of melt to residue in the equilibration volume may change with decreasing size, the composition of the equilibration volume may change throughout the supra-solidus part of the retrograde segment of the  $P$ – $T$  evolution. If melt has been lost from the equilibration volume, reaction reversal may not be possible or may be only partial; indeed, the common preservation of close-to-peak mineral assemblages in migmatite and granulite demonstrates that extensive reaction with melt is uncommon, which implies melt isolation or loss prior to crossing potential melt-consuming reactions. Water dissolved in melt is transported through the crust to be exsolved on crystallization at the solidus appropriate to the intrinsic  $a(H_2O)$ . This recycled water causes retrogression at subsolidus conditions. Consideration of the evidence for supra-solidus decompression-dehydration reactions, and review of microstructures that have proven controversial, such as corona and related microstructures, selvage microstructures and ‘late’ muscovite, leads to the conclusion that there is more than one way for these microstructures to form and reminds us that we should always consider multiple working hypotheses!

**Key words:** granulite; microstructure; migmatite;  $P$ – $T$  path; reaction reversal; retrograde processes.

## INTRODUCTION

Metamorphic rocks record information about material transport through orogens. Discerning the history of burial, residence at depth and exhumation of rock samples from orogens, and unravelling what processes have affected them in the course of their passage through an orogen, are essential elements in constraining alternative models of orogenesis and understanding crustal differentiation. Our ability to interpret correctly the evidence preserved in these rock samples is critical to these endeavours. Field relations, rock fabrics and porphyroblast-fabric relations, mineral assemblages and microstructures inferred to record reaction relations, and elemental and isotope compositional variations in minerals are evidence of this history (Brown,

1993; 2001a; Brown & O’Brien, 1997). In any consideration of microstructures inferred to record reactions there are kinetic factors to consider, such as the critical step for nucleation of product phases and the rate of dissolution of reactant phases. However, which reactions are crossed along the prograde (increasing  $T$ ) stage, whether these reactions are re-crossed along the retrograde (decreasing  $T$ ) stage, and whether fluid generated along the prograde segment of the path is retained to allow reaction reversal along the retrograde segment determines what happens to a particular rock as it is transported through an orogen.

In this paper, I review retrograde processes above the metapelite solidus and the interpretation of microstructures in migmatites and granulites. Rock

microstructure comprises the shape, size, orientation and distribution of grains, and the relationship of these grains with each other, which records important information about processes that have affected the rock. Microstructural interpretation is an important part of petrology. It is an area in which seminal contributions have been made by Ron Vernon (e.g., 1968; 1972; 1977; 1978; 1982; 1988, 1996; 1999; 2000; Flood & Vernon, 1988; Vernon & Collins, 1988; Collins & Vernon, 1991, 1993; Johnson & Vernon, 1995), to whose inspiring influence this paper is dedicated.

### *P–T–t* PATH AND FLUIDS

If the metamorphic history can be shown to result from a single metamorphic cycle, tectonostratigraphic relationships, microstructural interpretation and thermobarometry, and chronology may be used to resolve the depth–temperature–time evolution of orogens. Early studies on metamorphic rocks were concerned with reliable estimation of the *P–T* conditions of formation of a mineral assemblage inferred to have once been at chemical and microstructural equilibrium. However, to understand the evolution of *P–T* conditions as a rock is transported through an orogen, it is necessary to use samples that show apparent disequilibrium features in the microstructure and mineral chemistry. In fact, in using these samples we are applying the principles of an equilibrium approach, but at the domainal scale now permitted by advances in microscopy, compositional mapping and *in situ* analysis.

Various methods are used to unravel the history of individual rocks and their constituent minerals. Evidence includes microstructural information, such as mineral and fluid inclusions within porphyroblasts and replacement relations among minerals, and chemical information, such as compositional zoning in mineral grains and the make up of trapped fluids. Using these data with petrogenetic grids and/or thermobarometry, *P–T* conditions can be quantified for points and/or segments along the evolutionary path in *P–T* space. Improved calibration of mineral equilibria, development of new techniques to retrieve close-to-peak *P–T* conditions, the increasing use of pseudosections and the availability of *in situ* techniques to date specific petrological events have allowed refinement of the particular *P–T–t* path followed by individual rocks. Nonetheless, much of what we do in petrology is based on interpretation of microstructures, and therefore, much of what we do is open to re-evaluation. The interpretation of microstructures used to infer reactions is one example that focuses this discourse, particularly those in migmatites and granulites (e.g. Brown, 1983; Jones & Brown, 1990; Collins & Vernon, 1991, 1993; Hand *et al.*, 1992; Brown, 1993; Brown & Dallmeyer, 1996; Vernon, 1996; Kriegsman & Hensen, 1998; White *et al.*, 2002).

### Types of *P–T–t* path

Different stages of the evolution may be recorded by mineral inclusions in a porphyroblast phase, by compositional zoning and by microstructural relations that imply partial reaction to new assemblages. Using information of this kind, we have the potential to identify multiple points along the *P–T* path, and following the law of parsimony, the simplest trajectory is inferred in connecting these points. Critical to testing this interpretation is the determination *in situ* of ages that can be unambiguously linked to the metamorphic history recorded by the mineral assemblage in a particular microstructure. Natural *P–T–t* paths followed by individual rock volumes within orogens are unlikely to have been as smooth as those generated by models (e.g. Ruppel & Hodges, 1994; and references therein). For example, intrusion of magma may generate high-*T* spikes during exhumation (e.g. De Yoreo *et al.*, 1989; Brown & Solar, 1999), rock strength is strongly temperature dependant but under greenschist and lower amphibolite facies conditions, metamorphic pressures may be influenced by deviatoric stresses (e.g. Sandiford *et al.*, 1991; Stüwe & Sandiford, 1994; Handy *et al.*, 2001), and rates of exhumation may vary during the orogenic cycle (e.g. Hames *et al.*, 1989; Koons, 1995; Brown, 2001a).

Orogens commonly record evidence of a clockwise evolution in *P–T* space. These *P–T* paths range from tight hairpins close to isobaric heating and cooling paths to clockwise loops of variable  $\Delta P$  and  $\Delta T$  with decompression segments outside the errors inherent in geobarometry (e.g. Albarède, 1976; Hollister, 1982; Brown, 1983, 1998; Brown & Earle, 1983; Jones & Brown, 1990; Williams & Karlstrom, 1996). Harley (1989) summarized information about peak metamorphic conditions and postpeak evolution from high-grade metamorphic rocks worldwide. He used these data to argue that postpeak *P–T* paths commonly are close to one of two end-members, near-isobaric cooling or near-isothermal decompression. An outstanding petrological problem concerns whether microstructures developed along near-isothermal decompression path segments are distinguishable from retrograde features produced by near-isobaric cooling.

The retrograde evolution may be composite and stepped, being composed of decompression and cooling segments (Harley *et al.*, 1990; Brown & Dallmeyer, 1996; Brown & Raith, 1996; Raith *et al.*, 1997). Such stepped retrograde paths are likely to cross fluid-consuming reactions on close-to-isobaric cooling segments and fluid-producing reactions on close-to-isothermal decompression segments (see, e.g. White *et al.*, 2001). At temperatures above the solidus, melt in melt-bearing protoliths will crystallize along close-to-isobaric cooling segments of the retrograde evolution, and the volume of the system may decrease, whereas solids will produce additional melt along close-to-isothermal decompression segments of the retrograde

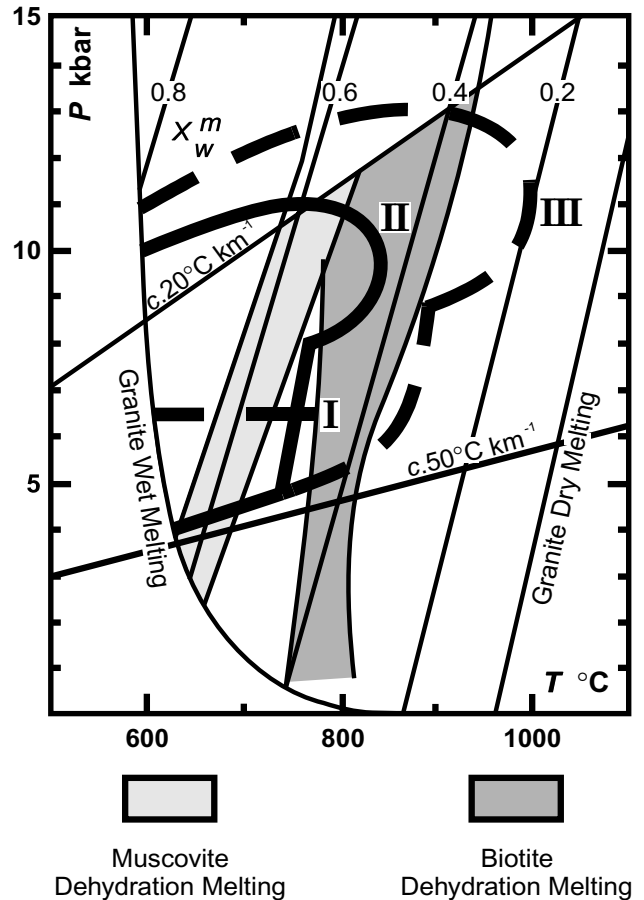
evolution, and the volume of the system will increase (Brown & Dallmeyer, 1996; Clemens & Droop, 1998; Spear *et al.*, 1999; White *et al.*, 2001). These effects have thermal, rheological and deformational implications.

### The volatile phase

The presence or absence, nature and composition of a volatile phase are important controls on what happens along both the prograde and retrograde portions of the  $P$ - $T$  path. High-grade metamorphic rocks in stable lower crust are unlikely to have a free, connected volatile phase, although this apparently unavoidable conclusion from petrology remains controversial (e.g. Yardley & Valley, 1997; Wannamaker, 2000; Yardley & Valley, 2000). However, in the dynamic environment of an active mountain belt, at temperatures below the pelite wet solidus, devolatilization reactions will contribute products to a sporadic, spatially heterogeneous volatile phase (Guiraud *et al.*, 2001). In this environment, it is likely that lithology controls the activities of components in the volatile phase, that is the system is rock-dominated, and lithology or structure controls volatile phase flow. In contrast, the transition from the amphibolite to the granulite facies is characterized by a decrease in the activity of water, as well as an increase in temperature, and the onset of dehydration melting reactions, which precludes the general presence of a ubiquitous, pervasive volatile phase. Even in the presence of a volatile phase a lowered activity of water could be achieved as long as the quantity of this phase is limited, such that the system is rock-dominated rather than fluid-dominated, although this precludes melting (Clemens & Watkins, 2001). Melting is inferred to have occurred in most granulite facies terranes (Fyfe, 1973), whether the rocks have a migmatitic appearance or not (Brown, 1994). Following this logic, Clemens & Watkins (2001) argued that the correlation between melt temperature and initial water content of granite magmas 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.

### Melting regimes

Melting may occur along an infinite number of  $P$ - $T$ - $t$  paths with different prograde and retrograde histories, from a clockwise path with a substantial decompression segment to a close-to-isobaric clockwise hairpin loop. Melting regimes in the crust are related to hydrate stability and the two main types of melting environment (Fig. 1): contact melting associated with plutons, which is characterized by essentially isobaric heating and cooling  $P$ - $T$  paths on relatively short timescales; and, regional melting during orogeny, which generally occurs along clockwise  $P$ - $T$  paths of longer duration leading to regionally developed migmatite and granulite terranes. The phase relations for a particular bulk



**Figure 1.**  $P$ - $T$  diagram of the anatectic zone, that region in  $P$ - $T$  space above the wet granite melting curve in which melt may be present in many common crustal rocks. For muscovite dehydration melting, the position of the equilibrium curve varies according to the amount of phengite substitution in the muscovite. For biotite dehydration melting, the position of the equilibrium curve varies according to bulk composition, being at lower temperature in common pelites and at higher temperature in graywackes. The symbol  $X$  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 (Thompson, 1996). Schematic  $P$ - $T$  paths: I. Isobaric heating – cooling path characteristic of deep contact metamorphism (e.g. around deep granites of the Cascades of Washington state, USA); II. Stepped clockwise path, characteristic of collisional metamorphism (e.g. the Himalayas or the ancient Variscan belt of Europe); and, III. Stepped clockwise path at UHT (e.g. ancient rocks in Peninsula India).

composition determine the number of melting regimes. For example, melting of a metasedimentary protolith may occur in one of three melting regimes ( $H_2O$ -rich volatile phase saturated melting, muscovite dehydration melting, and biotite dehydration melting) each involving different melting reactions with different thermal requirements and melt production. With increasing temperature across each of these regimes, the  $a(H_2O)$  progressively decreases, and in a closed system melt volume increases by dissolution of solid phases as the melt becomes drier (Spear *et al.*, 1999).

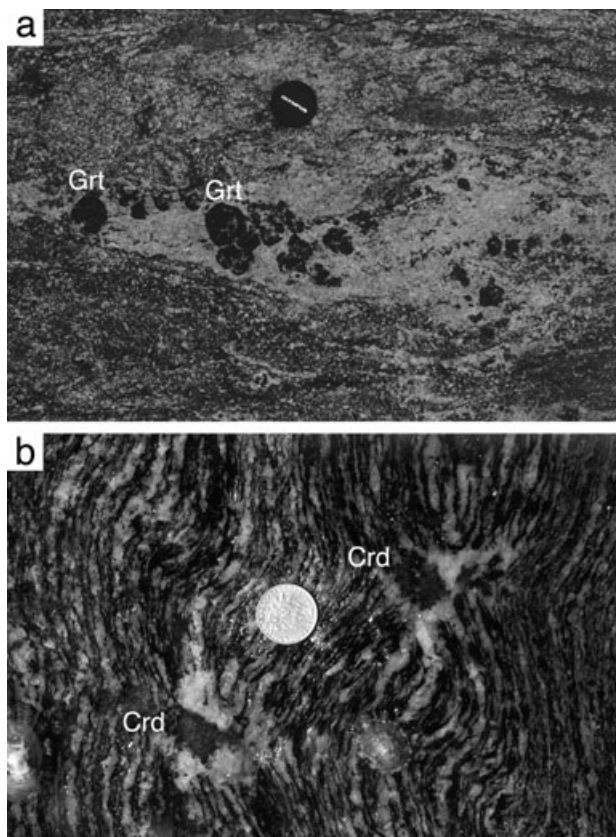


One feature common to many high-grade metamorphic terranes is the occurrence of leucosomes throughout the deformation history (e.g. Brown, 1979; Fleming & White, 1984; Jones & Brown, 1990; Mogk, 1990, 1992; Stevens & Van Reenen, 1992; Brown & Dallmeyer, 1996; Collins & Sawyer, 1996). Commonly, these leucosomes preserve microstructural features consistent with the presence of melt and lack evidence of significant intracrystalline plasticity, i.e. any deformation of the leucosome was largely accommodated by a melt-enhanced grain boundary process. This ubiquitous association of leucosome with each phase of deformation may have implications concerning recycling water internally during orogenesis in a melting/movement/congelation positive feedback loop driven by deformation (Holk & Taylor, 1997; Stevens, 1997).

#### Leucosomes and melt Loss

Although subsolidus processes may allow effective segregation of quartzo-feldspathic material, particularly at temperatures close to the solidus (e.g. Sawyer & Robin, 1986; Holness, 1995), most studies on migmatites and granulites regard leucosome as having an anatectic origin (e.g. Brown, 1994, 2001b). However, this should not be taken to mean that leucosome simply represents melt congealed *in situ*. Leucosomes may represent enhanced pre-existing layering due to addition of melt in more quartzose layers, or they may represent layers composed of early crystallized phases from which residual melt has been lost (e.g. Brown *et al.*, 1995; Marchildon & Brown, 2001, 2002; Solar & Brown, 2001). In general, melanosome may be evidence of where melt formed and leucosome most probably is evidence of where it collected (Sawyer, 1999). Products of biotite dehydration melting, such as garnet or cordierite with feldspar and quartz (Fig. 2), which have features of both melanosome and leucosome, may represent sites of melt generation (Powell & Downes, 1990; White *et al.*, 2001).

Microstructural analysis in migmatites and granulites requires care to distinguish features related to melting from those produced by subsolidus processes alone (e.g. Dallain *et al.*, 1999; Vernon, 1999). For example, Harlov & Wirth (2000) used multiple lines of evidence to suggest that K-feldspar microveins in metabasite layers in the upper Val Strona, Ivrea Zone, formed during granulite facies metamorphism as the consequence of localised dehydration reactions involving the breakdown of amphibole in the presence of quartz to orthopyroxene, clinopyroxene, plagioclase, K-feldspar and H<sub>2</sub>O. These authors argued for a subsolidus origin initiated by the presence of a low  $a(\text{H}_2\text{O})$  activity fluid (most likely a NaCl-KCl supercritical brine), related to the magmatic underplating event responsible for the Mafic Complex in the Ivrea Zone, and that this dehydration event did not involve partial melting.



**Figure 2.** Mesoscopic structures inferred to record evidence of dehydration melting in migmatitic granulite facies paragneisses. In both examples, the peritectic product of a biotite dehydration melting reaction, euhedral garnet in (a) and euhedral cordierite in (b), occurs in leucosome patches. However, there is no significant reaction reversal in either case, which suggests accumulation of peritectic and early crystallized minerals and melt loss in both cases. (a) Enxada Azul Beach, Guarapari, Espírito Santo, Brazil (b) Plage de Port Navalo, Morbihan, southern Brittany metamorphic belt, France.

A primarily anatectic origin for many migmatites is intimated by their macrostructure (e.g. Brown, 1973, 1994; Davidson *et al.*, 1994; Sawyer, 1999), revealed by their microstructure (e.g. Vernon & Collins, 1988; McLellan, 1989; Harte *et al.*, 1991; Brown, 1998; Sawyer, 1999, 2001; Marchildon & Brown, 2001, 2002; Vernon, 1999), suggested by geochemical data (e.g. Dougan, 1979, 1981; Weber *et al.*, 1985; Sawyer & Barnes, 1988; Sawyer, 1998; Milord *et al.*, 2001; Solar & Brown, 2001), permitted by estimates of  $P$ - $T$  conditions (e.g. Ashworth, 1985; Jones & Brown, 1989; Ashworth & Brown, 1990; Vielzeuf & Vidal, 1990), evidenced by microstructures inferred to record dehydration melting reactions (Waters, 1988; Sawyer, 2001), and indicated by accumulation of quartzo-feldspathic material in dilatant sites formed during syn-anatectic deformation of the protolith (e.g. Brown, 1994; Sawyer, 1994; Brown *et al.*, 1995; Brown & Rushmer, 1997; Snoko *et al.*, 1999; Tanner, 1999; Vernon & Paterson, 2001).

In putative anatectic rocks, including granulites, microstructures such as euhedral phenocrysts or peritectic phases in leucosomes, 'pseudomorphs' that mimic melt relations with solid grains (e.g. quartz or feldspar films along grain boundaries, interstitial-xenomorphous quartz or feldspar, residual quartz in feldspar), and fractured residual grains are interpreted to reflect the presence of melt (Fig. 3a,b & c) (e.g. Cuney & Barbey, 1982; Vernon & Collins, 1988; Harte *et al.*, 1991; Brown & Dallmeyer, 1996; Brown, 1998; Sawyer, 1999, 2001; Vernon, 1999; Watt *et al.*, 2000; Marchildon & Brown, 2001, 2002). Furthermore, the bulk rock composition of many migmatite and granulite terranes is consistent with depletion in felsic components (e.g. Weber *et al.*, 1985; Ellis & Obata, 1992; Schnetger, 1994; Nyman *et al.*, 1995; Hartel & Pattison, 1996; Solar & Brown, 2001). This implies that melt was not simply stagnant after *in situ* segregation and that melt has been extracted from these terranes. This further suggests that the leucosome geometry in these rocks records a once-active flow network through which melt has been lost from the system. In these examples, leucosome modes and/or chemical compositions usually reflect dominance of cumulate and/or residual solid phases after escape of some (fractionated) melt rather than melt compositions, even though migmatites may exhibit evidence of flow in the magmatic state (e.g. Blumenfeld & Bouchez, 1988; Powell & Downes, 1990; Sawyer, 1996, 2001; Carson *et al.*, 1997; Brown & Solar, 1998a, 1998b; Watt *et al.*, 2000; Solar & Brown, 2001).

In one example, Hartel & Pattison (1996) estimated 5–30 vol.% melt loss from garnet amphibolites by comparison between observed volume of leucosome and volume of melt expected based on mass balance calculations. This estimate is reasonable given data from experimental petrology that suggest 20–40 vol.% melting is required for formation of tonalite-trondhjemite-granodiorite melts (e.g. Rapp & Watson, 1995), and an estimate of 25 vol.% melting, based on REE patterns, for formation of the tonalite-trondhjemite-granodiorite suite in Archean terranes (e.g. Luais & Hawkesworth, 1994). Thus, migmatitic garnet amphibolites like those studied by Hartel & Pattison (1996) may be representative of the partially depleted source rocks for the widespread tonalite-trondhjemite-granodiorite suite. Although melt was lost from the system investigated by Hartel & 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. A critical observation, however, is that melt-consuming retrograde reaction between residual stagnant melt in the flow network and the residual matrix did not occur.

#### CONTROLS ON RETROGRADE REACTION

In this section I consider factors that may affect whether retrograde reaction is possible, how it may be

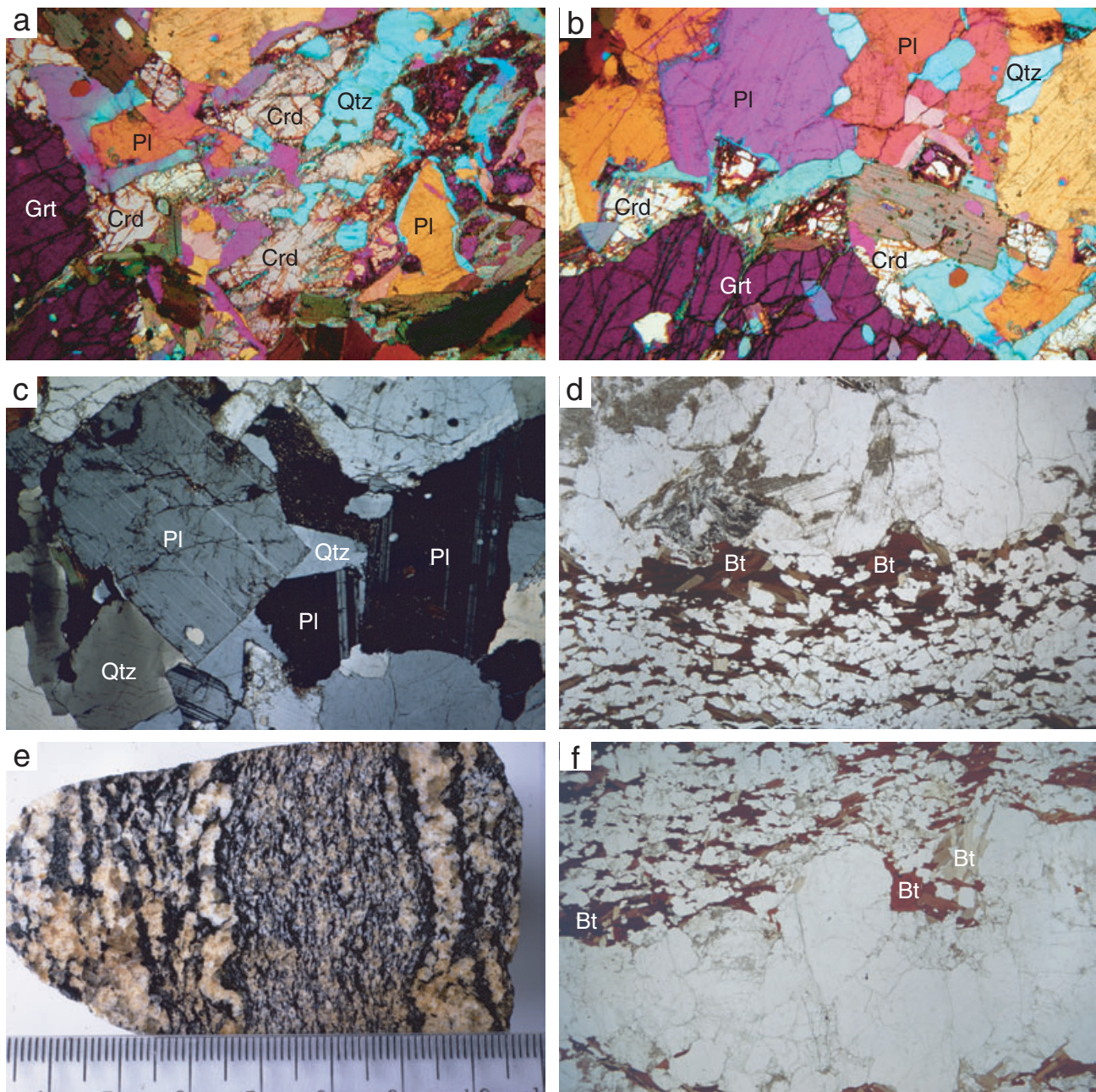
recorded in rocks and the extent to which it occurs in nature. Post-thermal peak reaction is influenced by: (1) the  $dP/dT$  slope of the  $P$ – $T$  path segments and the amount of decompression, in relation to invariants in the system under consideration and the Clapeyron slopes of the various equilibria; (2) the rate of cooling; and (3) the availability of fluid ( $H_2O$ -rich volatile phase or melt) for fluid-consuming reaction.

Along approximately isobaric  $P$ – $T$  paths, reactions crossed along the prograde evolution are recrossed during the retrograde evolution. Assuming no fluid migration within the system, fluid-producing reactions that progressed forward during the prograde evolution may be reversed during the retrograde evolution. However, in the dynamic environment of an active mountain belt where the  $P$ – $T$  path is likely to be clockwise with a decompression segment close to the thermal peak, the assumption that the fluid is static or that the length-scale of fluid migration is less than the radius of the equilibration volume is unlikely to be correct. This will be true whether the fluid is volatile phase or melt (Stüwe & Powell, 1989; Powell & Downes, 1990), which means that the retrograde history is unlikely ever to be the reverse of the prograde history.

In melt-bearing rocks, the radius of the equilibration volume will vary with structure, being controlled by the fabric in the same manner as melt connectivity (Brown *et al.*, 1999). Thus, the radius of the equilibration volume is likely to be longer in the plane of foliation parallel to the lineation and shorter in all other directions (Hand & Dirks, 1992). The size of the equilibration volume is also controlled by factors that affect diffusion, such as the temperature-time dependence (Dodson, 1976), the grain size (Stüwe, 1997), and the presence of a fluid and the grain boundary structure (Spear & Daniel, 2001), because diffusion is likely to be more rapid along triple-grain intersections than along double-grain boundaries. Since the proportion of melt to reside in the equilibration volume may change with decreasing size of this volume, the bulk composition of the equilibration volume potentially changes during cooling. Melt segregation and melt loss also potentially changes the bulk composition of the equilibration volume (Powell & Downes, 1990; White *et al.*, 2001). Melt loss is to be expected from melt bearing rocks, the most compelling evidence for which is the residual bulk composition (e.g. Schnetger, 1994; Solar & Brown, 2001) and largely preserved granulite facies mineral assemblages (e.g. Waters, 1988, 2001; Powell & Downes, 1990; Sawyer, 2001) of many migmatites and granulites.

For clockwise  $P$ – $T$  paths, the larger the decompression segment in the  $P$ – $T$  evolution the more likely it is that reactions different to those crossed along the prograde path will be encountered along the retrograde path. For example, for common pelite compositions (e.g. White *et al.*, 2001; see also Spear *et al.*, 1999), the anatectic zone may be divided into three pressure regimes: a high-pressure regime, at pressures





**Figure 3.** Microstructures in migmatites. (a) Photomicrograph of residual diatexite, Crouesty, Morbihan, southern Brittany metamorphic belt, France, to show quartz inferred to mimic former melt (in green, along edge of plagioclase, which is orange, in centre-right of field of view), and optically continuous peritectic cordierite (in grey, intergrown with quartz (magenta and green)); long dimension of field of view is 10 mm (b) Photomicrograph of residual diatexite, Crouesty, Morbihan, southern Brittany metamorphic belt, France, to show euhedral feldspar (magenta/purple) in interstitial quartz (green) inferred to mimic melt pocket (left-centre of field of view), and microstructural evidence of the Grt→Crd reaction (lower half of field of view); long dimension of field of view is 10 mm (c) Photomicrograph of euhedral plagioclase and intersertal quartz inferred to record crystallization from a melt in leucosome of stromatic migmatite, Crouesty, Morbihan, southern Brittany metamorphic belt, France; long dimension of field of view is 7 mm (d) Photomicrograph of discontinuous coarse biotite selvage at interface between leucosome and host; note reaction between leucosome and biotite selvage (left-hand side of field of view). Stromatic migmatite, Crouesty, Morbihan, southern Brittany metamorphic belt, France; long dimension of field of view is 17 mm (e) Mesoscopic structure of stromatic migmatite, note that biotite selvages vary from continuous to discontinuous, Crouesty, Morbihan, southern Brittany metamorphic belt, France. (f) Photomicrograph of discontinuous coarse biotite selvage against leucosome in stromatic migmatite shown in Fig. 3(e); long dimension of field of view is 17 mm.

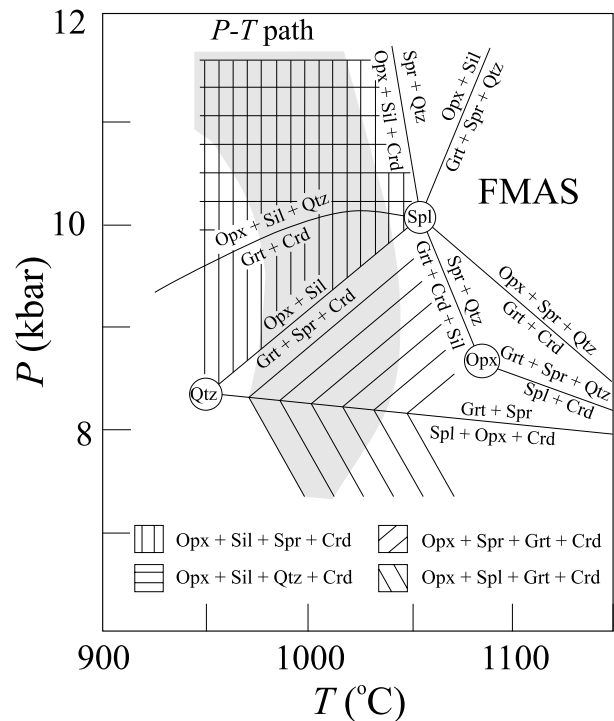
above the [Spl, Ms, H<sub>2</sub>O] invariant in KFMASH (approximately 9 kbar); a low-pressure regime, at pressures below the intersection of muscovite breakdown with the H<sub>2</sub>O-rich volatile phase-present solidus (approximately 4 kbar); and, a medium-pressure regime in between. Reaction reversal or ‘back reaction’ cannot occur along  $P$ – $T$  paths that traverse around one of these points, but may occur along  $P$ – $T$  paths confined to one of the three pressure regimes. However, the composition and modal coefficients of reactions change along the length of the equilibria in  $P$ – $T$  space. Thus, even within a single regime, perfect reaction reversal requires the equilibrium to be re-crossed at the same  $P$ – $T$  as occurred on the prograde path. For this reason, Kriegsman (2001) distinguished between ‘back reaction’ *sensu stricto* (*s.s.*), defined as ‘those retrograde reactions with *in situ* crystallizing melt that are the exact reversal of prograde reactions that produced the melt phase’, and ‘back reaction’ *sensu lato* (*s.l.*), defined as ‘any retrograde reaction with *in situ* crystallizing melt’ (reaction reversal of Jones & Brown, 1990). In ‘back reaction’ *s.l.*, the retrograde path passes through a different part of  $P$ – $T$  space in comparison with the prograde path, and commonly a decompression segment connects these two path segments.

The decompression segment of the  $P$ – $T$  evolution may cross reactions with shallow  $dP/dT$  slopes that are located around an invariant (e.g. Jones & Brown, 1990; Spear *et al.*, 1999) or a small multivariant field (e.g. White *et al.*, 2001). Normally these reactions are not recrossed during the same orogenic cycle. This issue of a rock sequentially crossing discontinuous reactions and intervening multivariant continuous reaction fields around an invariant or a small multivariant field during a clockwise  $P$ – $T$  evolution is particularly relevant to rocks at UHT metamorphic conditions (e.g. Harley *et al.*, 1990; Brown & Raith, 1996; Raith *et al.*, 1997; Moraes *et al.*, unpublished data). At these  $P$ – $T$  conditions, the mineral assemblages in rocks of metapelite or greywacke protolith composition commonly have low variance when considered in simple model systems. The number of equilibria and their topology in these simple model systems then allow us to constrain the trajectory of the  $P$ – $T$  path with a high level of confidence using the sequence of reactions inferred based on microstructural relations and mineral compositional data (e.g. Fig. 4).

Although local controls may apply to the selection of which mineral grain will react or where in a rock volume reaction will occur, mineral growth is primarily related to changes in  $P$ – $T$ – $X$  conditions. Within this context, some assemblages may form from a continuous reaction and the phases may change in mode and/or composition over a wide range of  $P$ – $T$ , whereas other assemblages may appear and become unstable over a very restricted range of  $P$ – $T$ . Minerals may grow in multiple generations within different assemblages from different reactions, whether continuous or discontinuous. For example, as Spear *et al.*, (1990, 1999)

have pointed out, garnet growth and resorption generate a complex reaction history in which the processes of growth and resorption may be switched several times during a clockwise  $P$ – $T$  evolution. Some reactions lead to large changes in modal mineralogy over small changes in  $P$ – $T$  conditions (White *et al.*, 2002), whereas for other reactions the converse is the case. Thus, a particular microstructure inferred to record evidence of a reaction may represent only a small portion of the complete  $P$ – $T$  path, and any inferred  $P$ – $T$  vector based on this information alone must be interpreted with due caution (cf. Vernon, 1996).

If retrograde reaction is effective, the evidence may be largely microstructural and subtle, and a reaction front related to rehydration may be mistaken for a prograde fluid-generating isograd reaction (McGregor & Friend, 1997). However, if most fluid is lost from the equilibration volume and ingress of external fluid



**Figure 4.** The close to metamorphic peak  $P$ – $T$  path segment followed by high-MgAl granulites from Perumalmalai, Palni Hill Ranges, southern India, as deduced from the sequence of observed mineral assemblages and microstructures inferred to record metamorphic reactions (based on data in Brown & Raith, 1996; and Raith *et al.*, 1997). This partial petrogenetic grid for the FMAS system at low oxygen fugacity was constructed by R. Moraes using data in Hensen (1987), Hensen & Harley (1990) and Bertrand *et al.* (1991). The trajectory of the  $P$ – $T$  path segment is constrained to pass at  $T > 950$  °C ([Qtz] invariant) and  $< 1050$  °C ([Spl] invariant), connecting to a shallow negative  $dP/dT$  prograde segment at 12–11 kbar and a shallow positive  $dP/dT$  retrograde segment at 8–7 kbar, both of which segments also are based on observed mineral assemblages and microstructures inferred to record metamorphic reactions (see Brown & Raith, 1996; Raith *et al.*, 1997).



does not occur, retrograde reaction essentially may be prevented from happening (Guiraud *et al.*, 2001; White *et al.*, 2001). This is seen in the preservation of pristine granulite facies assemblages in many terranes, and when fluid ingress occurs, it is commonly controlled by structure (e.g. Boundy *et al.*, 1992). Nonetheless, some high-grade terranes show evidence of limited retrograde reaction between stagnant melt and residual host during cooling (e.g. Kohn *et al.*, 1997; Kriegsman & Hensen, 1998; Spear *et al.*, 1999; Indares & Dunning, 2001). Retrograde reaction normally is recognized because reaction is either patchy or incomplete. This implies local control over which grains of the reactant phase break down and the extent of reaction progress. Decreasing temperature and decreasing diffusion rates, limited availability of fluid, or stress and/or strain partitioning could be the local control. Where the retrograde segment of the  $P$ - $T$  path occurs at lower pressure than the prograde segment, the topology of the reaction equilibria may allow consumption of the fluid by reversal of a higher temperature reaction before the  $P$ - $T$  path intersects a lower temperature reaction. Such an explanation is preferred by Indares & Dunning (2001) to explain the lack of development of retrograde muscovite during the retrograde evolution of high- $P$  migmatitic metapelites.

Kriegsman & Hensen (1998) posit that melt extraction from migmatites is generally incomplete and retrograde reaction between crystallizing melt and the residue is inevitable upon cooling. Although melt extraction from migmatites may be incomplete, the melt left behind may not be located at the site of melting. Commonly, melt segregates from residue along a melt drainage network migrating into dilational sites of lower melt pressure that link to form a melt transfer network (Brown & Rushmer, 1997; Brown & Solar, 1999; Brown, M. A. *et al.*, 1999; Sawyer, 2001). This segregated melt may collect into masses large enough to exceed the equilibration volume, making it chemically isolated from the surrounding residue (except very close to the contact) over the time scale of the retrograde evolution. In this case, it is reasonable to conclude that retrograde reaction is largely prohibited rather than inevitable. Only melt segregation at a scale smaller than the equilibration volume will allow reaction between melt and residue during retrogression, and only if no melt escapes from this volume does reaction have the potential to be complete. Although segregation of melt from residue by distances greater than the equilibration volume may prevent retrograde reversal of a prograde reaction, crystallization of the melt may promote retrogression in the surrounding host.

Processes involving melt retention or melt loss may be modelled using  $P$ - $T$  and  $T$ - $X$  pseudosections calculated for particular bulk compositions with appropriate software (e.g. THERMOCALC, Powell *et al.*, 1998) using an internally consistent thermodynamic dataset (e.g. Holland & Powell, 1998, as extended for

silicate melts by Holland & Powell, 2001 and White *et al.*, 2001). A fundamental conclusion from this work is that melt loss from granulites is necessary for the preservation of the characteristic mineral assemblages of the granulite facies, and the absence of widespread retrogression (White *et al.*, 2001). However, modelling of this kind cannot account per se for the physical process of melt segregation, and therefore it cannot distinguish between melt loss from the system and melt simply being outside the equilibration volume for the  $P$ - $T$  being considered. Thus, we must always interpret information gained using pseudosections in the light of spatial information from hand samples and field outcrops. Nonetheless, we may distinguish between rock-dominated behaviour, in which retrogression is limited by the distribution or amount of residual melt, and melt-dominated behaviour, as exemplified by some S-type granites, where higher  $P$ - $T$  ferro-magnesian minerals may be replaced by lower  $P$ - $T$  phases during crystallization and mineral-melt reaction (White & Chappell, 1990; Brown, 1995; Clemens & Droop, 1998; Stephens, 2001; White *et al.*, 2001).

#### MICROSTRUCTURES AND THEIR INTERPRETATION

Interpretation of the putative products of retrograde reaction has sometimes proven controversial, but the disputes relate to generalizations that may be inappropriate given the complexity involved, the multiple interpretations that are possible and the influence of many different processes. For example, we might consider consistency between a reaction inferred from a particular microstructure and a model reaction in a simple model system to mean that the microstructure was produced by that reaction, but it may instead be a product of another (similar) reaction. This uncertainty is illustrated using a particular microstructure that could be explained as the result of several alternative reactions. In these circumstances, a preference among the alternative reactions might be established by using additional data. Also, I review corona microstructure, which is a feature of many rocks that are inferred to have decompressed along the  $P$ - $T$  evolution. However, this microstructure also can form along isobaric cooling segments of  $P$ - $T$  paths, which makes the common reference to this microstructure as a 'decompression texture' ill advised! Finally, I argue that selvage microstructures are unlikely to have been formed by only one process, and it is possible that they form along both prograde and retrograde segments of the  $P$ - $T$  evolution.

In the interpretation of microstructures it is clear we should follow the method of multiple working hypotheses proposed by Chamberlain in 1890. If only one hypothesis is considered, data may be collected to support only this hypothesis and conflicting data may be ignored. In essence Chamberlain (1890) argued that we should seek as many explanations or hypotheses as



possible for a given phenomenon. Each hypothesis should be tested against data, during which procedure some of the hypotheses will be falsified. Once again, the law of parsimony is followed, and a simple hypothesis is normally to be preferred over those that are more complex. Unfortunately, the absence of an alternative explanation is no assurance that the interpretation preferred is indeed correct! I now discuss the interpretation of some microstructures that are common in migmatites and granulites.

#### Supra-solidus decompression-dehydration reactions

Thompson (2001) has reminded us that there is a class of reactions, which he refers to as supra-solidus decompression-dehydration reactions, that have the form hydrous and anhydrous phases + hydrous melt  $\rightarrow$  anhydrous phases + volatile phase. In effect, the reaction freezes the melt to liberate the volatile phase from both the remaining hydrous phase(s) and the melt. It is not clear what microstructure might reflect such reactions, but I speculate that a reaction of the type  $Bt + Als + Qtz + L \rightarrow Crd + Grt + Kfs + V$  (Thompson, 2001) might produce a microstructure similar to that in Fig. 5(a). This microstructure, from a Ryoke Belt migmatite, was interpreted by Brown (1998) to reflect the reaction  $Bt + Sil + Qtz (+ Pl) \rightarrow Grt + Crd + Kfs + L$ . Given the  $P$ - $T$  path inferred for the Ryoke belt (Fig. 8 in Brown, 1998), either reaction potentially is possible. Thus, other evidence must be used to establish which reaction produced this particular microstructure. In the case of the supra-solidus decompression-dehydration reaction, the expelled  $H_2O$ -rich volatile phase may promote retrogression. In the absence of evidence of significant retrogression in rocks associated with the sample shown in Fig. 5(a), the original interpretation of this microstructure as reflecting operation of a melting reaction is preferred. However, the observation that a specific microstructure is consistent with operation of a particular reaction is not a conclusive proof that it was produced by that reaction!

#### Coronae and related microstructures

One group of post-thermal peak microstructures that has been discussed in the recent literature is coronae and related microstructures. Examples of this type of microstructure are shown in Fig. 5(b,c). In both examples, cordierite is a product of garnet breakdown. The partial replacement of garnet by cordierite may occur along a variety of retrograde  $P$ - $T$  paths, but early descriptions of this microstructure were from migmatitic metapelites that underwent close-to-isothermal decompression from lower crustal to upper crustal pressures (e.g. Albarède, 1976; Hollister, 1982; Brown, 1983; Brown & Earle, 1983; Brown & Phadke, 1983). In these studies, the large decompression recorded by the rocks was quantified using thermobarometry on the garnet- and cordierite-bearing

assemblages, which interpretation was supported by other specific features of the mineral assemblages, such as  $Ky \rightarrow Sil$ .

Garnet-to-cordierite reactions for common quartz-bearing metapelites have rather shallow  $dP/dT$  (e.g. in NCKFMASH; White *et al.*, 2001). Thus, a corona microstructure in which a cordierite moat surrounds garnet has become associated with decompression. In the case of the example shown here (Fig. 5b), formation along a decompression segment of the  $P$ - $T$  evolution remains the preferred interpretation. However, the microstructure itself may not form uniquely by decompression. For example, in the special circumstance in which the assumption of  $a(SiO_2) = 1$  does not apply such coronae may result from a heating-cooling cycle without substantial change in pressure (Kriegsman & Hensen, 1998; Spear *et al.*, 1999).

Many of these corona microstructures involve more than one phase, and these phases commonly are intergrown in a symplectic texture (e.g. Fig. 5c). The reactions responsible for these corona microstructures with symplectic texture generally have rather shallow  $dP/dT$ , and commonly this microstructure is produced by reaction along a decompression segment of a  $P$ - $T$  path. However, neither the microstructure nor the internal texture by itself is diagnostic of decompression, although decompression remains the preferred interpretation of the example in Fig. 5(c).

Finally, a corona-type moat structure may be the result of nucleation preference and mole proportions. For example, the quartz-absent FMAS continuous reaction  $Opx + Sil \rightarrow Spr + Crd$  produces a symplectitic intergrowth of sapphirine and cordierite that replaces sillimanite, which is commonly surrounded by a moat of cordierite that separates the symplectite from residual orthopyroxene (e.g. Harley *et al.*, 1990). This particular microstructure also is a product of decompression at UHT conditions. The lesson to take away from this discussion is that a  $P$ - $T$  vector cannot be inferred from either a particular microstructure or a particular reaction unless there is additional contextual information (cf. Vernon, 1996; Kohn, 1999).

#### Selvage microstructures

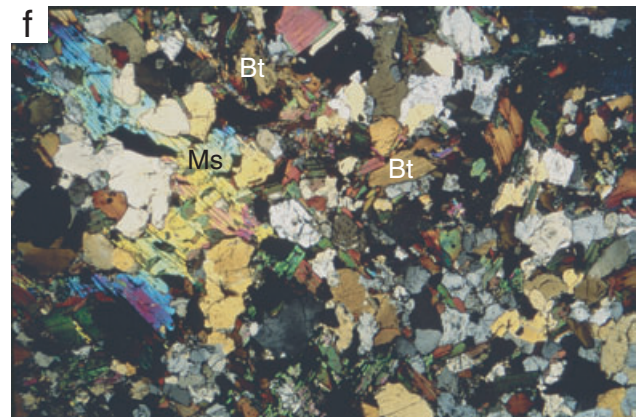
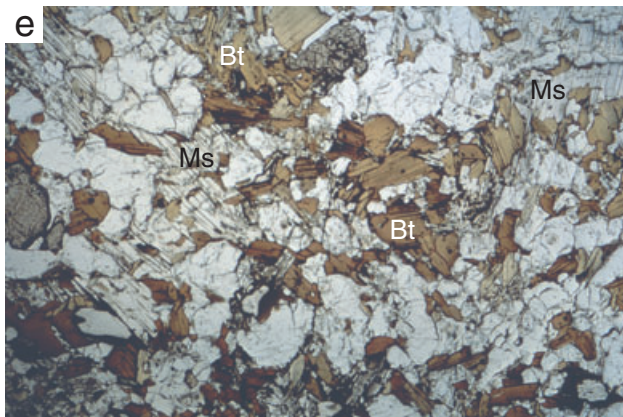
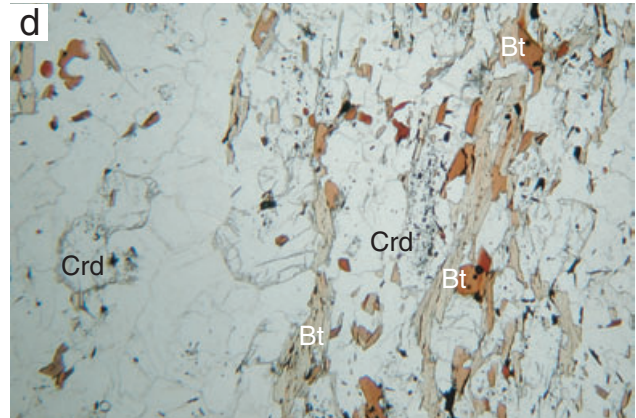
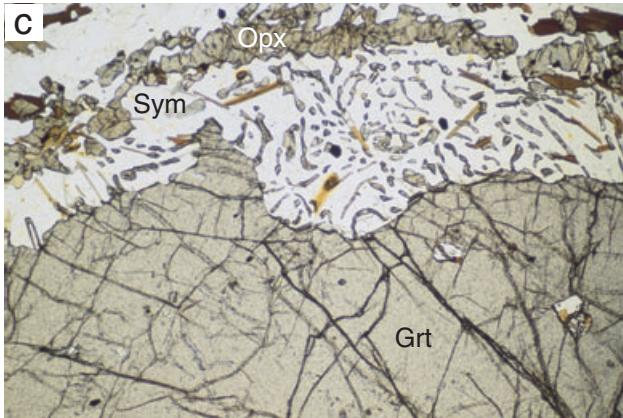
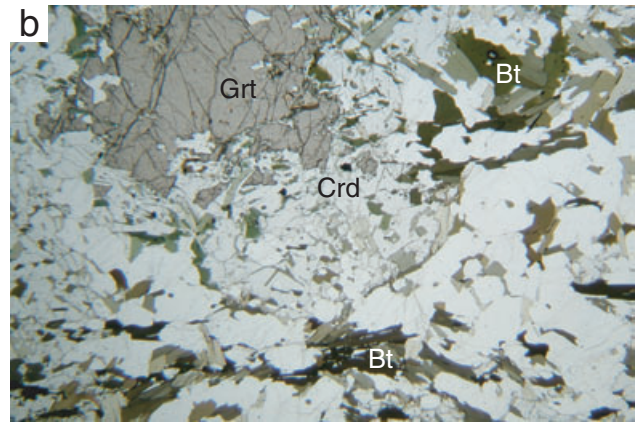
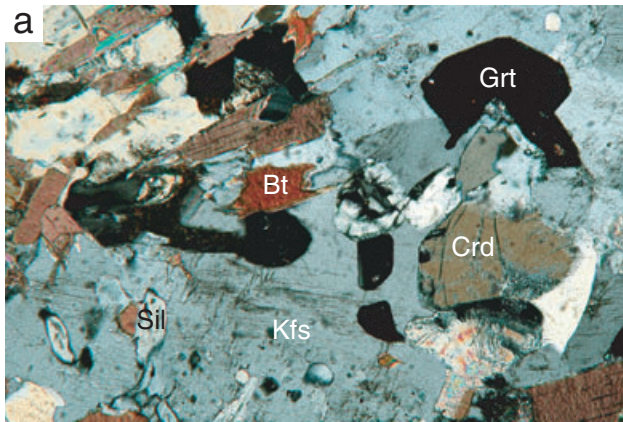
Biotite-rich melanosome and biotite rims between garnet and leucosome may result from reaction between melt and residue if other products of the postulated reaction are present (Kriegsman & Hensen, 1998). Stüwe (1997) argued that biotite may form at the expense of either cordierite or garnet during cooling from the granulite facies peak parageneses of garnet-cordierite-sillimanite, depending on grain size and equilibration volume, which together determine what Stüwe calls the 'effective bulk composition'. For the formation of biotite-rich selvages in stromatic migmatites formed by anatexis there are three possible end-member models: accumulation of phases residual from melting, particularly biotite; accumulation of

newly crystallized biotite; reaction between crystallizing melt and residue with biotite as the principal product. Which model applies is likely to relate to the melting regime, as discussed further below.

*H<sub>2</sub>O-rich volatile phase saturated melting*

H<sub>2</sub>O-rich volatile phase-saturated melting of amphibolite facies metasedimentary rocks commonly is modelled by a reaction of the form quartz + feldspar + volatile phase → melt ± residual phases. However, for significant melting, influx of a H<sub>2</sub>O-rich metamorphic

volatile phase is required, which likely limits this regime to contact metamorphic aureoles (e.g. Symmes & Ferry, 1995; Marchildon & Brown, 2001). Water-fluxed melting of muscovite-biotite schist has been investigated by Patiño Douce & Harris (1998), who ran some of their experiments with added H<sub>2</sub>O, resulting in the reaction: 9Ms + 15PL + 7Qtz + xH<sub>2</sub>O → 31melt, in which mica is involved as well as quartz and feldspar. The amount of melt is limited by protolith plagioclase content and the amount of added H<sub>2</sub>O, but generally is large. The composition of these melts varies from granite (lower pressure, water-fluxed melting) to





trondhjemite (higher pressure, water-fluxed melting). If the residue is dominated by biotite, as would be expected, then selvages may form by concentration of residual phases, particularly biotite, by a mechanism similar to the models in Brown *et al.*, (1995). In the contact metamorphic environment, lit par lit or stromatic leucosomes are formed by accumulation of melt in quartz-rich layers (cf. Brown *et al.*, 1995), leaving a concentration of residual phases in adjacent layers at edges against leucosomes, although melting apparently occurred throughout the protolith (Marchildon & Brown, 2001). Thus, a biotite-rich selvage may be developed, in which the grain size of the phases is similar to the grain size of the same minerals in the host.

#### *Muscovite dehydration melting*

Formation of stromatic migmatites by muscovite dehydration melting is widely suggested (e.g. Dougan, 1983; Milord *et al.*, 2001; Solar & Brown, 2001). Following Patiño Douce & Harris (1998), the metamorphic volatile phase-absent melting of muscovite-biotite schist below biotite dehydration can be modelled as follows:  $22Ms + 7PL + 8Qtz \rightarrow 25melt + 5Kfs + 5Sil + 2Bt$ . This reaction is widely considered appropriate for producing leucogranitic melt from metasedimentary protoliths of metapelitic composition in collisional orogens (e.g. Patiño Douce, 1999). Biotite is a peritectic product of this melt-producing reaction due to the incongruent breakdown of muscovite, and this product biotite may partly account for the larger biotite crystals forming selvages at the edges of leucosomes in stromatic migmatite (e.g. Fig. 3d,e & f). The amount of melt is limited by protolith muscovite content, and the composition of the melt is granite.

Muscovite dehydration melting involves a significant increase in volume due to the positive  $\Delta V_r$ , which will drive segregation (Brown *et al.*, 1995; Rushmer, 2001). Segregation of melt from residue should create concentrations of new (coarser?) biotite crystals (Fig. 3d),

which may not be continuously developed along the leucosome–host interface (Fig. 3f). In rocks that have developed a P (phyllosilicate-dominated) and Q (quartz-dominated) domainal fabric during subsolidus evolution, more melt is produced in the P domains, which are weaker, and melt flows from P to Q domains, to generate a stromatic structure, most likely with biotite-rich selvages that fingerprint the former P domains (cf. Brown *et al.*, 1995). Many such migmatites have apparently depleted bulk compositions (Solar & Brown, 2001), suggesting melt loss by flow from grain boundaries to leucosomes and along layering to planar veins related to shear surfaces or fractures that are oblique to and extend beyond the layering. In this latter case we might expect selvages to be essentially continuous.

#### **Biotite dehydration melting**

Common melting reactions in this regime are of the form  $Bt + Sil + Qtz \rightarrow$  peritectic phase (Grt/Crd/Spl) + melt (Al-rich pelite) or orthopyroxene also may be a product (common pelite or greywacke). Although K-feldspar may be expected as a product, under  $a(H_2O)$  conditions commonly prevailing in the middle crust during biotite dehydration melting, these reactions normally are K-feldspar absent (Carrington & Watt, 1995). Since biotite is consumed, selvages are not expected to form by biotite dehydration melting. Although it has been argued that biotite-sillimanite selvages will form by retrograde reaction reversal (Kriegsman & Hensen, 1998; Kriegsman, 2001; Spear *et al.*, 1999), it is common to find the products of, e.g. a garnet consuming reaction around the reactant phase (e.g. Brown & Dallmeyer, 1996; Fig. 7c).

Extensive biotite dehydration melting produces residual diatexites and granulites with distributed peritectic products (e.g. Brown, 1983; Stüwe & Powell, 1989; Sawyer, 2001), in contrast to stromatic migmatites, which appear to be more commonly formed under

**Figure 5.** Photomicrographs of microstructures in migmatites. (a) Small leucosome pod in migmatite, Ryoke Belt, Japan, cut perpendicular to lineation to show a microstructure inferred to represent either a supra-solidus decompression-dehydration reaction of the type  $Bt + Sil + Qtz + L \rightarrow Crd + Grt + Kfs + V$  or a dehydration melting reaction of the type  $Bt + Sil + Qtz (+ Pl) \rightarrow Grt + Crd + Kfs + L$ . Either interpretation is supported by the scalloped residual biotite (centre of field of view), residual sillimanite (bottom left-hand corner of field of view), prismatic cordierite (right-centre of field of view), and prismatic garnet (top right-hand corner) all of which are included within K-feldspar. Other evidence must be used to establish which reaction produced this particular microstructure. Long dimension of field of view is 3 mm (b) Residual diatexite, Ramakona area, Chindwara District, Madhya Pradesh, India, to show garnet (top left-hand sector of field of view) replaced by cordierite (centre of field of view) inferred to record the reaction  $Grt \rightarrow Crd$  during decompression (Brown & Phadke, 1983). Note that the biotite replacing garnet or cordierite is green ( $< 0.08$  Ti pfu, 22 O), whereas biotite in the host is brown (around 0.28 Ti pfu, 22 O) (Brown & Phadke, 1983); long-dimension of field of view is 10 mm (c) Garnet with a partial orthopyroxene necklace (plagioclase along inside edge) that outlines the original extent of the porphyroblast. Inside the orthopyroxene necklace, a symplectic intergrowth composed of cordierite and orthopyroxene (+euhedral biotite) has partially replaced the garnet porphyroblast. This microstructure and symplectic texture commonly is interpreted to represent decompression. Migmatitic metapelite, Sharyzhgay Complex, Lake Baikal, Russia; long dimension of field of view is 6 mm (d) Stromatic migmatite, Ryoke Belt, Japan, to show elongate inclusion-rich cordierite associated with biotite in mesosome (right-hand half of field of view), in contrast to equant subhedral-to-euhedral inclusion-free cordierite in leucosome (left-hand half of field of view); length of field of view is 5 mm (e & f) Plane polarized light and crossed polarized light views of 'late' muscovite in residual diatexite, Acadian metamorphic belt, west-central Maine, USA; long dimension of field of view is 14 mm. This microstructure is inferred to record a reaction of the type  $Kfs + Sil + H_2O$  (r)  $Ms + Qtz$ .



the water-fluxed and muscovite dehydration melting regimes. There are several characteristic microstructures associated with biotite dehydration melting, but the tendency of the reaction products, solid phases and melt, sometimes to form discordant coarse-grained segregations in which the felsic material surrounds the peritectic product is perhaps the most important (Fig. 2). The host rock commonly still carries hydrous phases, but the segregations essentially are anhydrous, commonly with compositions that do not match experimentally produced melts (Waters, 1988; Stüwe & Powell, 1989; Powell & Downes, 1990; Hand & Dirks, 1992). Chemical differentiation and melt separation are two processes that may induce modal proportions in these segregations that are inconsistent with the inferred reaction mode, which will inhibit retrograde reaction (e.g. Waters, 1988).

A principal assumption in the interpretation of processes in these rocks is that the segregations result from the presence of melt as the product of a dehydration melting reaction. It is only in the case of near-isochemical melting in which the segregations preserve the anhydrous peritectic products of the dehydration melting reaction together with the hydrous melt that crystallizes *in situ* that reaction reversal may occur. More common is the situation in which the peritectic phase shows no evidence of retrograde reaction (e.g. the cordierite shown in Fig. 5d), even in circumstances where the  $P$ - $T$  path was a tight clockwise hairpin loop with only limited decompression at peak temperature (Brown, 1998).

#### 'Late' muscovite

The common occurrence of 'late' muscovite in many migmatite terranes (e.g. St. Malo migmatite belt, see: Brown, 1979) is likely a response to  $H_2O$  exsolved as stagnant melt crystallizes (Solar & Brown, 2001). The common muscovite dehydration melting reaction  $Ms + Pl + Qtz \rightarrow melt + Sil + Bt \pm Kfs$  (Patiño Douce & Harris, 1998) produces sufficient melt such that it generally leads to melt loss (e.g. Solar & Brown, 2001). However, melt loss at one level implies melt flow through another level. As this melt flow network dies, crystallization of residual melt that becomes stagnant will exsolve  $H_2O$  and may lead to a reaction of the kind  $Kfs + Sil + H_2O \rightarrow Ms + Qtz$ , which may explain the common 'late' quartz-sieved muscovite characteristic of many migmatite leucosomes (Fig 5e,f). Indeed, chloritization of biotite in leucosomes may be produced in an analogous manner. Once again, local control may be important and muscovitization need not be uniformly developed. These hydration reactions do not constitute evidence that melt was retained in the sample or evidence for fluid infiltration. If melt commonly has been lost from regionally developed migmatites, the muscovite-producing reaction is unlikely to be the reverse of the melt-producing reaction, viz.  $melt + Sil + Bt + Kfs \rightarrow Ms + Pl + Qtz$  (cf. Kohn *et al.*, 1997),

although it may be the result of some other multivariant reaction. In circumstances where melt is retained in migmatites, and where the retrograde segment of the  $P$ - $T$  path passes at higher pressure than approximately 4 kbar, on encountering the solidus, the melt recrystallizes fully, releasing  $H_2O$  and producing muscovite (Spear *et al.*, 1999). Whether muscovite produced by these different mechanisms can be distinguished on a microstructural basis is arguable, but since melt loss removes  $H_2O$  from the system, the amount of muscovite produced in such cases should be smaller than in the examples where melt was retained.

Whatever the retrograde evolution, a chemical signature will be preserved in the distribution of trace elements between the major mineral phases. These distributions are controlled by residue-melt interactions during melting, armouring of accessory phases, melt loss and residue-melt interactions during crystallization of any remaining melt. The distributions may mimic solid-state equilibria (Nabelek, 1999), but they may also reflect disturbance of the systematics by the growth of 'late' muscovite, which may explain Rb-Sr-Ba relations in residual migmatites (Solar & Brown, 2001), and subsolidus re-equilibration (Fourcade *et al.*, 1992).

#### CONCLUSIONS

A preponderance of evidence indicates that many migmatites and granulites generally preserve close to metamorphic peak mineral assemblages. This is inevitable given melt loss from high-grade metamorphic terranes to feed upper crustal granite plutons. However, these migmatites and granulites do also preserve some microstructural and mineral chemical evidence of apparent disequilibrium inferred to record limited progress of metamorphic reactions during the  $P$ - $T$  evolution of these rocks, particularly along the retrograde segment of the  $P$ - $T$  path. The  $P$ - $T$  paths for individual or suites of rock samples are constructed from the succession of reactions inferred from this evidence, and it is the patchy and incomplete nature of these reactions that makes them useful for this purpose. In using samples that exhibit apparent disequilibrium, we are nonetheless following an equilibrium approach, but at the scale of a reaction domain. Such an approach has been made possible by advances in microscopy, compositional mapping and *in situ* analysis.

Segregation of melt from residue enables melt loss. Although limited retrograde reaction may occur at the interface between isolated pockets of segregated but stagnant melt and solid residue, melt segregation and melt loss are essentially irreversible processes, and as a result reaction reversal ('back reaction') is a process of limited volumetric importance in most migmatites and granulites. An integrated investigation of the effects of melt loss on retrograde evolution is overdue. This will involve studies of natural examples of

migmatites and granulites and comparison with predictions from  $P$ - $T$  and  $T$ - $X$  pseudosections calculated for bulk compositions of the natural examples using an internally consistent thermodynamic dataset.

Microstructures that record apparently arrested reactions, coronae and related microstructures and selvage microstructures may have multiple origins. So called 'late' muscovite most likely is a response to water exsolved from crystallizing melt that has become stagnant as the rate of melt flow declined and magma became stuck in the melt transfer network. However, the particular reaction by which muscovite is produced may be one of several alternatives according to circumstances, for example whether the system is open or closed to melt loss at the scale of the equilibration volume and whether reaction reversal is possible or not.

In any consideration of retrograde processes in migmatites and granulites we should use the method of multiple working hypotheses and follow the law of parsimony in selecting a preferred explanation for some particular phenomenon. In this respect, Ron Vernon has always set an example.

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#### REFERENCES

- Albarède, F., 1976. Thermal models of post-tectonic decompression as exemplified by the Haut-Allier granulites (Massif Central, France). *Bulletin de la Société Géologique de France*, **18**, 1023–1032.
- Ashworth, J. R., 1985. *Migmatites*. Blackie, Glasgow.
- Ashworth, J. R. & Brown, M., 1990. High-temperature Metamorphism and Crustal Anatexis. The Mineralogical Society Series, 2 Unwin-Hyman, London.
- Bertrand, P., Ellis, D. J. & Green, D. H., 1991. The stability of sapphirine-quartz and hypersthene-sillimanite-quartz assemblages: An experimental investigation in the system FeO–MgO–Al<sub>2</sub>O<sub>3</sub>–SiO<sub>2</sub> under H<sub>2</sub>O, H<sub>2</sub>O–CO<sub>2</sub> and CO<sub>2</sub> conditions. *Contributions to Mineralogy and Petrology*, **108**, 55–71.
- Blumenfeld, P. & Bouchez, J. L., 1988. Shear criteria in granite and migmatite deformed in the magmatic and solid states. *Journal of Structural Geology*, **10**, 361–372.
- Boudry, T. M., Fountain, D. M. & Austrheim, H., 1992. Structural development and petrofabrics of eclogite facies shear zones, Bergen Arcs, western Norway: implications for deep crustal deformation processes. *Journal of Metamorphic Geology*, **10**, 127–146.
- Brown, M., 1973. The Definition of Metatexis, Diatexis and Migmatite. *Proceedings of the Geologists' Association*, **84**, 371–382.
- Brown, M., 1979. The petrogenesis of the St. Malo migmatite belt, Armorican Massif, France, with particular reference to the diatexites. *Neues Jahrbuch Fur Mineralogie Abhandlungen*, **135**, 48–74.
- Brown, M., 1983. The petrogenesis of some migmatites from the presqu'île de Rhuys, southern Brittany, France. In: *Migmatites, Melting and Metamorphism* (eds Atherton, M. P. & Gribble, C. D.), pp. 174–200. Shiva Publishers Ltd., Nantwich.
- Brown, M., 1993.  $P$ - $T$ - $t$  evolution of orogenic belts and the causes of regional metamorphism. *Journal of the Geological Society, London*, **150**, 227–241.
- Brown, M., 1994. The generation, segregation, ascent and emplacement of granite magma: the migmatite-to-crustally-derived granite connection in thickened orogens. *Earth-Science Reviews*, **36**, 83–130.
- Brown, M., 1995. The late-Precambrian geodynamic evolution of the Armorican segment of the Cadomian belt (France): Distortion of an active continental margin during south-west directed convergence and subduction of a bathymetric high. *Géologie de la France*, **3**, 3–22.
- Brown, M., 1998. Unpairing metamorphic belts:  $P$ - $T$  paths and a tectonic model for the Ryoke Belt, southwest Japan. *Journal of Metamorphic Geology*, **16**, 3–22.
- Brown, M., 2001a. From microscope to mountain belts: 150 years of petrology and its contribution to understanding geodynamics, particularly the tectonics of orogens. *Journal of Geodynamics*, **32**, 115–164.
- Brown, M., 2001b. Crustal melting and granite magmatism: key issues. *Physics and Chemistry of the Earth (A)*, **26**, 201–212.
- Brown, M., Averkin, Y., McLellan, E. & Sawyer, E., 1995. Melt segregation in migmatites. *Journal of Geophysical Research*, **100**, 15,655–15,679.
- Brown, M. A., Brown, M., Carlson, W. D. & Denison, C., 1999. Topology of syntectonic melt flow networks in the deep crust: inferences from three-dimensional images of leucosome geometry in migmatites. *American Mineralogist*, **84**, 1793–1818.
- Brown, M. & Dallmeyer, R. D., 1996. Rapid Variscan exhumation and role of magma in core complex formation: Southern Brittany metamorphic belt, France. *Journal of Metamorphic Geology*, **14**, 361–379.
- Brown, M. & Earle, M. M., 1983. Cordierite-bearing schists and gneisses from Timor, eastern Indonesia:  $P$ - $T$  conditions of metamorphism and tectonic implications. *Journal of Metamorphic Geology*, **1**, 183–203.
- Brown, M. & O'Brien, P. J., 1997. Evolution of Metamorphic Belts: A Changing View. In: *Precambrian Geology and Metamorphic Petrology* (eds Xianglin, Q., Zhendong, Y. & Hall, H. C.), pp. 217–231. Proceedings of the 30th International Geological Congress, VSP, Zeist.
- Brown, M. & Phadke, A. V., 1983. High temperature reactions in pelitic gneiss from Precambrian Sausar metasediments of the Ramakona area, Chindwara District, Madhya Pradesh (India): definition of the exhumation  $P$ - $T$  path and the tectonic implications. In: *Prof. Kelkar Memorial Volume* (eds Phadke, A. V. & Pansalkar, V. G.), pp. 61–96. Indian Society of Earth Scientists, Poona.
- Brown, M. & Raith, M., 1996. First evidence of ultrahigh-temperature decompression from the granulite province of Southern India. *Journal of the Geological Society, London*, **153**, 819–822.

- Brown, M. & Rushmer, T., 1997. The role of deformation in the movement of granitic melt: Views from the laboratory and the field. In: *Deformation-Enhanced Melt Segregation and Metamorphic Fluid Transport* (ed Holness, M.), pp. 111–144. The Mineralogical Society Series: 8. Chapman & Hall, London.
- Brown, M. & Solar, G. S., 1998a. Shear zone systems and melts: Feedback relations and self-organization in orogenic belts. *Journal of Structural Geology*, **20**, 211–227.
- Brown, M. & Solar, G. S., 1998b. Granite ascent and emplacement during contractional deformation in convergent orogens. *Journal of Structural Geology*, **20**, 1365–1393.
- Brown, M. & Solar, G. S., 1999. The mechanism of ascent and emplacement of granite magma during transpression: a syntectonic granite paradigm. *Tectonophysics*, **312**, 1–33.
- Carrington, D. P. & Watt, G. R., 1995. A geochemical and experimental study of the role of K-feldspar during water-undersaturated melting of metapelites. *Chemical Geology*, **122**, 59–76.
- Carson, C. J., Powell, R., Wilson, C. J. L. & Dirks, P. H. H. M., 1997. Partial melting during tectonic exhumation of a granulite terrane: an example from the Larsemann Hills, East Antarctica. *Journal of Metamorphic Geology*, **15**, 105–126.
- Chamberlain, T. C., 1890. The method of multiple working hypotheses. *Science*, **15**, 92–96.
- Clemens, J. D. & Droop, G. R. T., 1998. Fluids, *P-T* paths and the fates of anatexitic melts in the Earth's crust. *Lithos*, **44**, 21–36.
- Clemens, J. D. & Watkins, J. M., 2001. The fluid regime of high-temperature metamorphism during granitoid magma genesis. *Contributions to Mineralogy and Petrology*, **140**, 600–606 (DOI 10.1007/s004100000205).
- Collins, W. J. & Sawyer, E. W., 1996. Pervasive magma transfer through the lower-middle crust during non-coaxial compressional deformation: An alternative to dyking. *Journal of Metamorphic Geology*, **14**, 565–579.
- Collins, W. J. & Vernon, R. H., 1991. Orogeny associated with anticlockwise *P-T-t* paths: evidence from low-*P*, high-*T* metamorphic terranes in the Arunta Inlier, central Australia. *Geology*, **19**, 835–838.
- Collins, W. J. & Vernon, R. H., 1993. Comment on Hand *et al.*: How well established is isobaric cooling in Proterozoic orogenic belts? An example from the Arunta Inlier, central Australia. *Geology*, **21**, 953–954.
- Cuney, M. & Barbey, P., 1982. Mise en évidence de phénomènes fractionnée dans les migmatites. *Compte Rendu de L'académie de France*, **295**, 37–42.
- Dallain, C., Schulmann, K. & Ledru, P., 1999. Textural evolution in the transition from subsolidus annealing to melting process, Velay Dome, French Massif Central. *Journal of Metamorphic Geology*, **17**, 61–74.
- Davidson, C., Schmid, S. M. & Hollister, L. S., 1994. Role of melt during deformation in the deep crust. *TERRA Nova*, **6**, 133–142.
- De Yoreo, J. J., Lux, D. R. & Guidotti, C. V., 1989. The role of crustal anatexis and magma migration in the thermal evolution of regions of thickened continental crust. In: *Evolution of Metamorphic Belts, Geological Society Special Publication*, **43** (eds Daly, J. S., Cliff, R. A. & Yardley, B. W. D.), 187–202, Geological Society, London.
- Dodson, M. H., 1976. Kinetic processes and thermal history of slowly cooled solids. *Nature*, **259**, 551–553.
- Dougan, T. W., 1979. Compositional and modal relationships and melting relationships in some migmatitic metapelites from New Hampshire and Maine. *American Journal of Science*, **279**, 897–935.
- Dougan, T. W., 1981. Melting reactions and trace element relationships in selected specimens of migmatitic pelites from New Hampshire and Maine. *Contributions to Mineralogy and Petrology*, **78**, 337–344.
- Dougan, T. W., 1983. Textural relations in melanosomes of selected specimens of migmatitic pelitic schists. Implications for leucosome generating processes. *Contributions to Mineralogy and Petrology*, **83**, 82–98.
- Ellis, D. J. & Obata, M., 1992. Migmatite and melt segregation at Cooma, New South Wales. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **83**, 95–106.
- Fleming, P. D. & White, A. J. R., 1984. Relationships between deformation and partial melting in the Palmer migmatites. *Australian Journal of Earth Sciences*, **31**, 351–360.
- Flood, R. H. & Vernon, R. H., 1988. Microstructural evidence of orders of crystallization in granitoid rocks. *Lithos*, **21**, 237–245.
- Fourcade, S., Martin, H. & De Bremond d'Ars, J., 1992. Chemical exchange in migmatites during cooling. *Lithos*, **28**, 43–53.
- Fyfe, W. S., 1973. The granulite facies, partial melting and the Archaean crust. *Philosophical Transactions of the Royal Society (London)*, **A273**, 457–461.
- Guiraud, M., Powell, R. & Rebay, G., 2001. H<sub>2</sub>O in metamorphism and unexpected behaviour in the preservation of metamorphic mineral assemblages. *Journal of Metamorphic Geology*, **19**, 445–454.
- Hames, W. E., Tracy, R. J. & Bodnar, R. J., 1989. Post metamorphic unroofing history deduced from petrology, fluid inclusions, thermochronometry, and thermal modeling: An example from southwestern New England. *Geology*, **17**, 727–730.
- Hand, M. & Dirks, P. H. G. M., 1992. The influence of deformation on the formation of axial-planar leucosomes and the segregation of small melt bodies within the migmatitic Napperby Gneiss, Central Australia. *Journal of Structural Geology*, **14**, 591–604.
- Hand, M., Dirks, P. H. G. M., Powell, R. & Buick, I. S., 1992. How well established is isobaric cooling in Proterozoic orogenic belts? An example from the Arunta Inlier, central Australia. *Geology*, **20**, 649–665.
- Handy, M. R., *et al.*, 2001. Rheology and geodynamic modelling: the next step forward. *International Journal of Earth Sciences*, **90**, 149–156, (DOI 10.1007/s005310000161).
- Harley, S. L., 1989. The origins of granulites: a metamorphic perspective. *Geological Magazine*, **126**, 215–247.
- Harley, S. L., Hensen, B. J. & Sheraton, J. W., 1990. Two-stage decompression in orthopyroxene-sillimanite granulites from Forefinger Point, Enderby Land, Antarctica: Implications for the evolution of the Archaean Napier Complex. *Journal of Metamorphic Geology*, **8**, 591–613.
- Harlov, D. E. & Wirth, R., 2000. K-feldspar-quartz and K-feldspar-plagioclase phase boundary interactions in garnet-orthopyroxene gneiss from the Val Strona di Omega, Ivrea-Verbano Zone, northern Italy. *Contributions to Mineralogy and Petrology*, **140**, 148–162.
- Harte, B., Pattison, D. R. M. & Linklater, C. M., 1991. Field relations and petrography of partially melted pelitic and semi-pelitic rocks. In: *Equilibrium and Kinetics in Contact Metamorphism, The Ballachulish igneous complex and its aureole* (eds Voll, G., Töpel, J., Pattison, D. R. M. & Seifert, F.), pp. 82–210. Springer-Verlag, Berlin and Heidelberg.
- Hartel, T. H. D. & Pattison, D. R. M., 1996. Genesis of the Kapuskasing (Ontario) migmatitic mafic granulites by dehydration melting of amphibole: The importance of quartz to reaction progress. *Journal of Metamorphic Geology*, **14**, 591–611.
- Hensen, B. J., 1987. *P-T* grids for silica-undersaturated granulites in the systems MAS (n+4) and (n+3) – tools for the derivation of *P-T* paths of metamorphism. *Journal of Metamorphic Geology*, **5**, 255–271.
- Hensen, B. J. & Harley, S. L., 1990. Graphical analysis of *P-T-X* relations in granulite facies metapelites. In: *High Temperature Metamorphism and Crustal Anatexis* (eds Ashworth, J. R. & Brown, M.), pp. 19–56. Unwin-Hyman, New York.
- Holk, G. J. & Taylor, H. P., 1997. <sup>18</sup>O/<sup>16</sup>O homogenization of the middle crust during anatexis: The Thor-Odin metamorphic core complex, British Columbia. *Geology*, **25**, 31–34.
- Holland, T. J. B. & Powell, R., 1998. An internally-consistent thermodynamic dataset for phases of petrological interest. *Journal of Metamorphic Geology*, **16**, 309–344.



- Holland, T. J. B. & Powell, R., 2001. Calculation of phase relations involving haplogranitic melts using an internally-consistent thermodynamic dataset. *Journal of Petrology*, **42**, 673–683.
- Hollister, L. S., 1982. Metamorphic evidence for rapid (>2 mm/yr) uplift of a portion of the Central Gneiss Complex, Coast Mountains, BC. *Canadian Mineralogist*, **20**, 319–332.
- Holness, M. B., 1995. The effect of feldspar on quartz-H<sub>2</sub>O-CO<sub>2</sub> dihedral angles at 4 kbar, with consequences for the behaviour of aqueous fluids in migmatites. *Contributions to Mineralogy and Petrology*, **118**, 356–364.
- Indares, A. & Dunning, G., 2001. Partial melting of high-*P*-*T* metapelites from the Tshenukush terrane (Grenville Province): petrography and U–Pb geochronology. *Journal of Petrology*, **42**, 1547–1565.
- Johnson, S. E. & Vernon, R. H., 1995. Stepping stones and pitfalls in the determination of an anticlockwise *P*-*T*-*t*-deformation path in the low-*P*, high-*T* Cooma Complex, Australia. *Journal of Metamorphic Geology*, **13**, 165–183.
- Jones, K. A. & Brown, M., 1989. The metamorphic evolution of the southern Brittany Migmatite Belt. In: *Evolution of Metamorphic Belts*, Geological Society Special Publication, 43 (eds Daly, J. S., Cliff, R. A. & Yardley, B. W. D.), 501–505, Geological Society, London.
- Jones, K. A. & Brown, M., 1990. High-temperature ‘clockwise’ *P*-*T* paths and melting in the development of regional migmatites: an example from southern Brittany, France. *Journal of Metamorphic Geology*, **8**, 551–578.
- Kohn, M. J., 1999. Metamorphic petrology. *Geotimes*, **July**, 22–23.
- Kohn, M. J., Spear, F. S. & Valley, J. W., 1997. Dehydration-melting and fluid recycling during metamorphism, Rangeley Formation, New Hampshire, USA. *Journal of Petrology*, **38**, 1255–1277.
- Koons, P. O., 1995. Modeling the Topographic Evolution of Collisional Belts. *Annual Review of Earth and Planetary Sciences*, **23**, 375–408.
- Kriegsman, L. M., 2001. Partial melting, partial melt extraction and partial back reaction in anatectic migmatites. *Lithos*, **56**, 75–96.
- Kriegsman, L. M. & Hensen, B. J., 1998. Back reaction between restite and melt: Implications for geothermobarometry and pressure-temperature paths. *Geology*, **26**, 1111–1114.
- Luais, B. & Hawkesworth, C. J., 1994. The generation of continental crust: An integrated study of crust-forming processes in the Archean of Zimbabwe. *Journal of Petrology*, **35**, 43–93.
- Marchildon, N. & Brown, M., 2001. Melt segregation in late syn-tectonic anatectic migmatites: an example from the Onawa Contact Aureole, Maine, USA. *Physics and Chemistry of the Earth (A)*, **26**, 225–229.
- Marchildon, N. & Brown, M., 2002. Grain-scale melt distribution in two contact aureole rocks: implications for controls on melt localization and deformation. *Journal of Metamorphic Geology*, **20**, in press.
- McGregor, V. R. & Friend, C. R. L., 1997. Field recognition of rocks totally retrogressed from granulite facies: An example from Archean rocks in the Paamiut region, south-west Greenland. *Precambrian Research*, **86**, 59–70.
- McLellan, E. L., 1989. Sequential formation of sub-solidus and anatectic migmatites in response to thermal evolution, eastern Scotland. *Journal of Geology*, **97**, 165–182.
- Milord, I., Sawyer, E. W. & Brown, M., 2001. Formation of diatexite migmatites and granite magma during anatexis of semi-pelitic metasedimentary rocks: An example from St. Malo, France. *Journal of Petrology*, **42**, 487–505.
- Mogk, D. W., 1990. A model for the granulite-migmatite association in the Archean basement of southwestern Montana. In: *Granulites and Crustal Evolution* (eds Vielzeuf, D. & Vidal, Ph.), pp. 133–155.
- Mogk, D. W., 1992. Ductile shearing and migmatization at mid-crustal level in an Archean high-grade gneiss belt, northern Gallatin Range, Montana, USA. *Journal of Metamorphic Geology*, **10**, 427–438.
- Moraes, R., Brown, M., Fuck, R. A., Camargo, M. A. & Lima, T. M., 2002. Role of melt in the evolution of UHT mineral assemblages: an example from central Brazil. *Journal of Petrology*, in review.
- Nabelek, P. I., 1999. Trace-element distribution among rock-forming minerals in Black Hills migmatites, South Dakota: a case for solid-state equilibrium. *American Mineralogist*, **84**, 1256–1269.
- Nyman, M. W., Pattison, D. R. M. & Ghent, E. D., 1995. Melt extraction during formation of K-feldspar + sillimanite migmatites, west of Revelstoke, British Columbia. *Journal of Petrology*, **36**, 351–372.
- Patiño Douce, A. E., 1999. What experiments tell us about the relative contributions of crust and mantle to the origin of granitic magmas. In: *Understanding Granites. Integrating New and Classical Techniques*, Geological Society Special Publication, 168 (eds Castro, A., Fernandez, C. & Vigneresse, J.-L.), pp. 55–75, Geological Society, London.
- Patiño Douce, A. E. & Harris, N., 1998. Experimental constraints on Himalayan anatexis. *Journal of Petrology*, **39**, 689–710.
- Powell, R. & Downes, J., 1990. Garnet porphyroblast-bearing leucosomes in metapelites: Mechanisms, phase diagrams, and an example from Broken Hill, Australia. In: *High-Temperature Metamorphism and Crustal Anatexis. The Mineralogical Society Series: 2* (eds Ashworth, J. R. & Brown, M.), pp. 105–123. Unwin-Hyman, London.
- Powell, R., Holland, T. J. B. & Worley, B., 1998. Calculating phase diagrams involving solid solutions via non-linear equations, with examples using THERMOCALC. *Journal of Metamorphic Geology*, **16**, 577–588.
- Raith, M., Karmakar, S. & Brown, M., 1997. Ultrahigh-temperature metamorphism and multi-stage decompressional evolution of sapphirine granulites from the Palni Hill Ranges, Southern India. *Journal of Metamorphic Geology*, **15**, 379–399.
- Rapp, R. P. & Watson, E. B., 1995. Dehydration melting of metabasalt at 8–32 kbar: Implications for continental growth and crust-mantle recycling. *Journal of Petrology*, **36**, 891–931.
- Ruppel, C. & Hodges, K. V., 1994. Pressure-temperature-time paths from two-dimensional thermal models: Prograde, retrograde, and inverted metamorphism. *Tectonics*, **13**, 17–44.
- Rushmer, T., 2001. Volume change during partial melting reactions: Implications for melt extraction, melt chemistry and crustal rheology. *Tectonophysics*, in press.
- Sandiford, M., Martin, N., Zhou, Z. & Frasier, G., 1991. Mechanical consequences of granite emplacement during high-*T*, low-*P* metamorphism and the origin of ‘anticlockwise’ *PT* paths. *Earth and Planetary Science Letters*, **107**, 164–172.
- Sawyer, E. W., 1994. Melt segregation in the continental crust. *Geology*, **22**, 1019–1022.
- Sawyer, E. W., 1996. Melt-segregation and magma flow in migmatites: Implications for the generation of granite magmas. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **87**, 85–94.
- Sawyer, E. W., 1998. Formation and evolution of granite magmas during crustal reworking: the significance of diatexites. *Journal of Petrology*, **39**, 1147–1167.
- Sawyer, E. W., 1999. Criteria for the recognition of partial melting. *Physics and Chemistry of the Earth*, **A24**, 269–279.
- Sawyer, E. W., 2001. Melt segregation in the continental crust: Distribution and movement of melt in anatectic rocks. *Journal of Metamorphic Geology*, **19**, 291–309.
- Sawyer, E. W. & Barnes, S.-J., 1988. Temporal and compositional differences between subsolidus and anatectic migmatite leucosomes from the Quetico metasedimentary belt. *Canada. Journal of Metamorphic Geology*, **6**, 437–450.
- Sawyer, E. W. & Robin, P.-Y. F., 1986. The subsolidus segregation of layer-parallel quartz-feldspar veins in greenschist

- to upper amphibolite facies metasediments. *Journal of Metamorphic Geology*, **4**, 237–260.
- Schnetger, B., 1994. Partial melting during the evolution of the amphibolite-to-granulite-facies gneisses of the Ivrea Zone, northern Italy. *Chemical Geology*, **112**, 71–101.
- Snoke, A. W., Kalakay, T. J., Quick, J. E. & Sinigoi, S., 1999. Deep-crustal shear zone as a result of mafic igneous intrusion in the lower crust, Ivrea-Verbanò Zone, Southern Alps, Italy. *Earth and Planetary Science Letters*, **166**, 31–45.
- Solar, G. S. & Brown, M., 2001. Petrogenesis of Migmatites in Maine, USA. Possible Source of Peraluminous Granite in Plutons. *Journal of Petrology*, **42**, 789–823.
- Spear, F. J. & Daniel, C. G., 2001. Diffusion control of garnet growth, Harpswell Neck, Maine, USA. *Journal of Metamorphic Geology*, **19**, 179–196.
- Spear, F. S., Kohn, M. J. & Cheney, J. T., 1999. *P-T* paths from anatectic pelites. *Contributions to Mineralogy and Petrology*, **134**, 17–32.
- Spear, F. J., Kohn, M. J., Florence, F. P. & Menard, T., 1990. A model for garnet and plagioclase growth in pelitic schists: implications for thermobarometry and *P-T* path determinations. *Journal of Metamorphic Geology*, **8**, 683–696.
- Stephens, W. E., 2001. Polycrystalline amphibole aggregates (clots) in granites as potential I-type restite: an ion probe microstudy of rare-earth distributions. *Australian Journal of Earth Sciences*, **48**, 591–602.
- Stevens, G., 1997. Melting, carbonic fluids and water recycling in the deep crust: An example from the Limpopo belt, South Africa. *Journal of Metamorphic Geology*, **15**, 141–154.
- Stevens, G. & Van Reenen, D., 1992. Partial melting and the origin of metapelitic granulites in the Southern Marginal Zone of the Limpopo Belt, South Africa. *Precambrian Research*, **55**, 303–319.
- Stüwe, K., 1997. Effective bulk composition changes due to cooling: a model predicting complexities in retrograde reaction + textures. *Contributions to Mineralogy and Petrology*, **129**, 43–52.
- Stüwe, K. & Powell, R., 1989. Metamorphic segregations associated with garnet and orthopyroxene porphyroblast growth: two examples from the Larsemann Hills, East Antarctica. *Contributions to Mineralogy and Petrology*, **103**, 523–530.
- Stüwe, K. & Sandiford, M., 1994. Contribution of deviatoric stresses to metamorphic *P-T* paths: An example appropriate to low-*P*, high-*T* metamorphism. *Journal of Metamorphic Geology*, **12**, 445–454.
- Symmes, G. H. & Ferry, J. M., 1995. Metamorphism, fluid flow and partial melting in pelitic rocks from the Onawa contact aureole, Central Maine, USA. *Journal of Petrology*, **36**, 587–612.
- Tanner, D. C., 1999. The scale-invariant nature of migmatite from the Oberpfalz, NE Bavaria and its significance for melt transport. *Tectonophysics*, **302**, 297–305.
- Thompson, A. B., 1996. Fertility of crustal rocks during anatexis. *Transactions of the Royal Society, Edinburgh: Earth Science*, **87**, 1–10.
- Thompson, A. B., 2001. Clockwise *P-T* paths for crustal melting and H<sub>2</sub>O recycling in granite source regions and migmatite terrains. *Lithos*, **56**, 33–45.
- Vernon, R. H., 1968. Microstructures of high-grade metamorphic rocks at Broken Hill, Australia. *Journal of Petrology*, **9**, 1–22.
- Vernon, R. H., 1972. Reactions involving hydration of cordierite and hypersthene. *Contributions to Mineralogy and Petrology*, **34**, 125–137.
- Vernon, R. H., 1977. Relationships between microstructures and metamorphic assemblages. *Tectonophysics*, **39**, 439–452.
- Vernon, R. H., 1978. Pseudomorphous replacement of cordierite by symplectic intergrowths of andalusite, biotite and quartz. *Lithos*, **11**, 283–289.
- Vernon, R. H., 1982. Isobaric cooling of two regional metamorphic complexes related to igneous intrusions in southeastern Australia. *Geology*, **10**, 76–81.
- Vernon, R. H., 1988. Sequential growth of cordierite and andalusite porphyroblasts, Cooma complex, Australia: microstructural evidence of a prograde reaction. *Journal of Metamorphic Geology*, **6**, 255–269.
- Vernon, R. H., 1996. Problems with inferring *P-T-t* paths in low-*P* granulite facies rocks. *Journal of Metamorphic Geology*, **14**, 143–153.
- Vernon, R. H., 1999. Quartz and feldspar microstructures in metamorphic rocks. *Canadian Mineralogist*, **37**, 513–524.
- Vernon, R. H., 2000. Review of Microstructural Evidence of Magmatic and Solid-State Flow. *Electronic Geosciences*, **5**, 2000.
- Vernon, R. H. & Collins, W. J., 1988. Igneous microstructures in migmatites. *Geology*, **16**, 1126–1129.
- Vernon, R. H. & Paterson, S. R., 2001. Axial-surface leucosomes in anatectic migmatites. *Tectonophysics*, **35**, 133–192.
- Vielzeuf, D. & Vidal, P., 1990. *Granulites and crustal evolution*. Kluwer Academic Publishers, Norwell.
- Wannamaker, P. E., 2000. Comment on ‘The petrologic case for a dry lower crust’ by Bruce W. D. Yardley and John W. Valley. *Journal of Geophysical Research*, **B105**, 6057–6064.
- Waters, D. J., 1988. Partial melting and the formation of granulite facies assemblages in Namaqualand, South Africa. *Journal of Metamorphic Geology*, **6**, 387–404.
- Waters, D. J., 2001. The significance of prograde and retrograde quartz-bearing intergrowth microstructures in partially-melted granulite-facies rocks. *Lithos*, **56**, 97–110.
- Watt, G. R., Oliver, N. H. S. & Griffin, B. J., 2000. Evidence for reaction-induced microfracturing in granulite facies pelitic migmatites. *Geology*, **28**, 331–334.
- Weber, C., Barbey, P., Cuney, M. & Martin, H., 1985. Trace element behavior during migmatization. Evidence for a complex melt–residuum–fluid interaction in the St. Malo Migmatitic Dome (France). *Contributions to Mineralogy and Petrology*, **90**, 52–62.
- White, A. J. R. & Chappell, B. W., 1990. Per migma ad magma downunder. *Geological Journal*, **25**, 221–225.
- White, R. W., Powell, R. & Clarke, G. L., 2002. The interpretation of reaction textures in Fe-rich metapelitic granulites of the Musgrave Block, central Australia: Constraints from mineral equilibria calculations in the system K<sub>2</sub>O–FeO–MgO–Al<sub>2</sub>O<sub>3</sub>–SiO<sub>2</sub>–H<sub>2</sub>O–TiO<sub>2</sub>–Fe<sub>2</sub>O<sub>3</sub>. *Journal of Metamorphic Geology*, **20**, This issue.
- White, R. W., Powell, R. & Holland, T. J. B., 2001. Calculation of partial melting equilibria in the system Na<sub>2</sub>O–CaO–K<sub>2</sub>O–FeO–MgO–Al<sub>2</sub>O<sub>3</sub>–SiO<sub>2</sub>–H<sub>2</sub>O (NCKFMASH). *Journal of Metamorphic Geology*, **19**, 139–153.
- Williams, M. L. & Karlstrom, K. E., 1996. Looping *P-T* paths and high-*T*, low-*P* middle crustal metamorphism: Proterozoic evolution of the southwestern United States. *Geology*, **24**, 1119–1122.
- Yardley, B. W. D. & Valley, J. W., 1997. The petrologic case for a dry lower crust. *Journal of Geophysical Research*, **B102**, 12173–12185.
- Yardley, B. W. D. & Valley, J. W., 2000. Comment on ‘The petrologic case for a dry lower crust’ by Yardley, B. W. D. & Valley, J. W. – Reply. *Journal of Geophysical Research*, **B105**, 6065–6068.

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