Introduction to a *virtual special issue* on crustal melting

This virtual special issue represents a collection of papers concerning **Crustal Melting** selected by the Editor from those published on various aspects of this theme in the *JMG* since 1982. The papers are grouped into sequences that address topics that have been prominent in the *JMG* during the last 30 years concerning the origin, evolution and tectonic role of migmatites and migmatitic granulites in crustal evolution. These topics are: Open- and closed-system processes in the formation of migmatites and migmatitic granulites; thermobarometry, *P–T* paths, phase equilibria modelling and retrograde processes in formerly melt-bearing rocks; geochronology in partially melted rocks; and, microstructures, deformation and tectonics of melt bearing rocks.

About one-third of the papers published in the *JMG* since its inception concern the origin, evolution and tectonic significance of migmatites and migmatitic granulites in crustal evolution, including the first special issue published by the *JMG* concerning 'Studies in the genesis and deformation of migmatites' edited by Tracy & Day (1988; Volume 6, Issue 4, Pages 385–543). Three subsequent Special Issues of the *JMG* include papers relevant to the theme of this virtual special issue; they are 'Metamorphic processes: a celebration of the career contribution of Ron Vernon', 'Processes in granulite metamorphism' and 'Granulites, partial melting and the rheology of the lower crust', edited by Brown & Clarke (2002; Volume 20, Issue 1, Pages 1–213), Brown & White (2008; Volume 26, Issue 2, Pages 121–299) and Brown *et al.* (2011; Volume 29, Issue 1, Pages 1–166), respectively.

The selection of papers in this virtual special issue is by no means comprehensive, but it is intended as a representative selection of what has been published in the *JMG* over 30 years to give the reader a broad overview of crustal melting. Furthermore, although many papers address more than one topic, each is included only once and has been placed within the most appropriate section.

INTRODUCTION

Migmatites are rocks found in high-grade metamorphic areas that are commonly heterogeneous at the microscopic to macroscopic scale. They generally consist of two, or more, petrographically different parts that result from partial melting and melt segregation and migration. Many granulites are migmatitic with a heterogeneous macroscopic structure. Examples are shown in Fig. 1. Migmatites occur in the inner parts of contact metamorphic aureoles, in the high-grade parts of many low-pressure–high-temperature metamorphic belts and metamorphic core complexes, and in large areas of exposed deep continental crust. In some cases granites associated with areas of migmatite have been shown to be petrogenetically related to each other.

The genesis of migmatites, granulites and granites and their role in the evolution, rheology and stabilization of continental crust are issues that continue to excite discussion among a broad spectrum of geologists. Migmatites and granulites are widely distributed in the exposed high-grade continental crust, whereas granites dominate in shallowly eroded orogenic belts; commonly, these deep crustal migmatites and granulites represent the residues of partial melting of undifferentiated accreted and magmatic precursors, whereas the granites represent melt extracted from these precursors during orogenesis. Crustal melting and transfer of melt from deeper to shallower crust is the process by which the precursor crust becomes differentiated into a more evolved upper and a more residual lower continental crust, although the process by which this happens during orogenesis may have changed over time (Hacker *et al.*, 2011).

Crustal melting produces striking rocks—migmatites and granulites—many of which have complex morphologies that result from syn-anatectic deformation during their formation (Fig. 1a–c). Most also have eye-catching microstructures. These microstructures may be due to the preservation of coarse-grained peritectic minerals relict from prograde melting reactions (Fig. 1d) and other more exotic minerals, such as sapphirine, or they may result from the effects of suprasolidus retrograde reaction between these minerals and residual melt during exhumation and cooling—commonly producing spectacular intergrowths and coronae.

In the crust, metamorphic rocks at upper amphibolite facies conditions—but below their wet solidus—have very low porosity and so contain only a small amount of aqueous fluid in pores and along grain boundaries, which limits the amount of melt produced at the wet solidus to ~1 vol.%. Consequently, production of the large volumes of melt required to differentiate the precursor crust during subduction and collision occurs either by fluid-absent melting of assemblages including one or more hydrous minerals, such as mica or amphibole, or it requires a large volume of aqueous fluid to be introduced into upper amphibolite facies crust during deformation, as, for
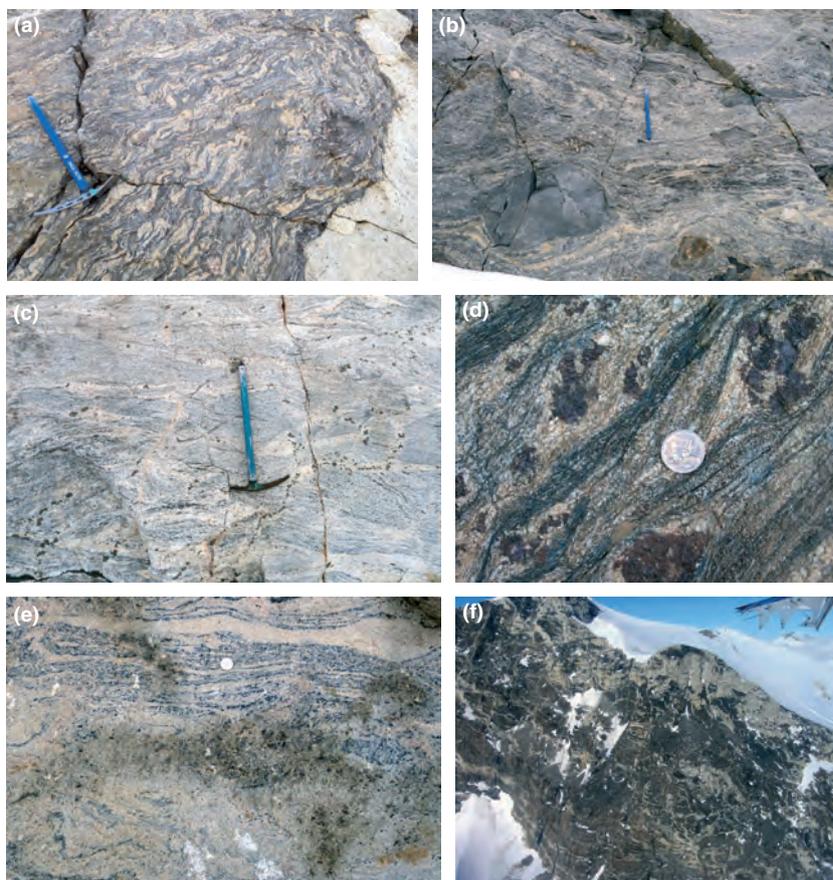
Fig. 1. Examples of migmatites and related phenomena (terminology after Sawyer, 2008). (a) Refolded isoclines in granulite facies garnet-bearing stromatic metatexite migmatite derived from a metasedimentary protolith, eastern Mt. Avers, Fosdick Mountains, West Antarctica. (b) Inhomogeneous diatexite migmatite with resisters of garnet-bearing orthogneiss, south-eastern Mt. Iphigene, Fosdick Mountains, West Antarctica. (c) Granulite facies cordierite- (± garnet) bearing veined metatexite migmatite derived from an igneous protolith, south-eastern Marujupu Peak, Fosdick Mountains, West Antarctica. (d) Peritectic garnet poikiloblasts in K-feldspar–quartz-rich leucosomes separated by sillimanite–biotite-rich melanosomes, Round Hill, Broken Hill Block, Australia. (e) Migmatite and granite variably darkened with a charnockite overprint at the southern termination of the Closepet granite, Kabbaldurga quarry, southern Karnataka, India. (f) Sill and dyke network in stromatic metatexite migmatite at Maigetter Peak (height 480 m), Fosdick Mountains, West Antarctica. Close examination in outcrops reveals that intersecting granites do not appear to truncate or displace each other, and although the granites crosscut foliation they may be continuous with or discordant to leucosomes in the migmatite.

example, might occur in association with the formation of major shear zones.

An origin by partial melting has not always been the preferred mechanism for the formation of migmatites (see the historical review in Sawyer, 2008), and fluid infiltration remains a controversial alternative for the formation of some granulites, where large-scale movement of high-salinity aqueous fluids and, to a lesser extent, carbonic fluids, is argued to be responsible for the reduction in a_{H2O} recorded across the amphibolite facies to granulite facies transition (e.g. Touret & Huizenga, 2011). For example, patchy charnockite (Fig. 1e) is sometimes explained by such a mechanism. Although this virtual special issue concerns crustal melting, the early volumes of the *JMG* include several important papers on fluid inclusion studies and fluid infiltration that are included herein,

together with later papers debating and testing such models, and extending to the infiltration of melts.

Returning to partial melting, the overall complexity of outcrops of migmatite and granulite representative of the deep crust is caused by the interaction between several factors. First, the bulk chemical composition of the protoliths determines the *P–T* conditions that must be achieved for fluid-absent melting to begin and the amount of melt produced in the crust as a whole for any given *P–T* conditions. Second, the peak *P–T* conditions, since higher peak temperatures allow a wider range of bulk compositions to be involved in melting and enable a larger amount of melt to be produced in the crust as a whole. These two factors together determine the extent of melt-induced weakening of the crust. Third, the level of applied differential stress, which drives synanatectic deformation

and melt segregation. Melt migration influences rheology in different parts of the orogenic crust at different times at all scales, particularly to produce the characteristic complexity of structures in the residual crust.

Pelitic rocks commonly contain 30–50 vol.% mica at the onset of partial melting and will produce melt progressively via mica-breakdown melting reactions with increasing temperature along a suprasolidus P–T path (Sawyer et al., 2011); as long as quartz and feldspar are present, any infiltration of aqueous fluid at this stage of the P–T evolution simply increases the volume of melt produced at any given temperature. Other protoliths such as metagreywackes, meta-andesites and meta-basalts also produce melt progressively along a suprasolidus prograde P–T path by fluid-absent melting of mica- and amphibole-bearing mineral assemblages, but the melt composition varies with protolith composition, being tonalitic, for example, if derived from a meta-basalt (Clemens, 2006). Fluid-absent melting of mica- and amphibole-bearing mineral assemblages is incongruent, producing residues with typical granulite facies mineral assemblages, but these assemblages are only preserved if melt is lost, preventing suprasolidus retrograde reaction with melt from replacing the peritectic products of the prograde melting reactions (White & Powell, 2002). These topics are all represented in the 30 volumes of the *JMG* and in this virtual special issue.

With the advent of comprehensive thermodynamic datasets and advanced computer protocols (Powell & Holland, 1988; Connolly, 1990; Holland & Powell, 1998; Powell et al., 1998; de Capitani & Petrakakis, 2010), phase equilibria modelling has become accessible to everyone in their pursuit of P–T conditions and P–T–t paths, and in understanding the development and evolution of mineral assemblages. Phase equilibria modelling of crustal melting is possible for peraluminous protolith compositions (White et al., 2001, 2007), such as pelites and granites, and has led to advances in understanding crustal melting and the evolution of residues—migmatites and granulites—after melt loss.

Comparisons between experimental studies of partial melting and model calculations using equivalent bulk compositions in either the Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O–TiO_2–O_2 (NCKFMASHTO) system (Grant, 2009) or the Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O (NCKFMASH) system (White et al., 2011) show similarities between what can be calculated and what may be interpreted from experiments for crustal melting, particularly for the topologies of key melting equilibria. Although there remain significant differences between the two approaches (White et al., 2011), overall these comparative studies suggest that phase equilibria modelling, within appropriate limits, does provide an acceptable way to extend the limited number of experimental studies of crustal melting using particular rock compositions within a limited range of P–T

conditions to a wider range of protolith compositions and P–T–a_{H_2O} conditions in nature.

The extent of post thermal peak retrograde reaction is influenced by the shape of the P–T path, for example the magnitude of the pressure drop between the prograde and the retrograde paths, the rate of cooling and the availability of melts and/or fluids to allow melt- and fluid-consuming reactions to proceed (Brown, 2002). Microstructures such as corona and symplectites are commonly used to infer retrograde reactions, which are then used to track a path for a particular sample through a generalized P–T grid or a sample-specific P–T pseudosection constructed for suprasolidus conditions, although care must be taken to ensure the correct interpretation (e.g. Baldwin et al., 2005; White & Powell, 2011). Retrograde processes have important implications for how ages determined from accessory phases in high-grade metamorphic rocks should be interpreted. Retrograde processes are well represented in the 30 volumes of the *JMG* and in this virtual special issue.

The mechanisms by which melt in suprasolidus regions of the crust is segregated and extracted to ascend through the crust remain controversial. The connection between small-scale features, such as networks of leucosomes that can be mapped in an outcrop of migmatite or granulite, to medium-scale features, such as metric granites (Fig. 1f), that transport melt longer distances may enable distinction among the several mechanisms proposed for the large-scale movement of anatectic melts through the continental crust (Brown, 2010). Papers concerning melt extraction and ascent have been published frequently in the *JMG*, particularly during the past decade, and a selection is included in this virtual special issue.

As Editor of this virtual special issue, it is my hope that having this selection of papers from the *JMG* in one place, together with a brief introduction to the papers in each section, will promote further advances in our understanding of crustal melting and the origin of migmatites and granulites. I acknowledge constructive review comments on the first draft of this virtual special issue by R.W. White and D.L. Whitney.

In each section below, the papers are listed in the order discussed in the associated introductory text; insofar as scientific coherence allows, the papers are listed approximately in chronological order.

- Open- and closed-system processes in the formation of migmatites and migmatitic granulites
- Thermobarometry, P–T paths, phase equilibria modelling and retrograde processes in formerly melt-bearing rocks
- Geochronology in partially melted rocks
- Microstructures, deformation and tectonics of melt-bearing rocks

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Open- and closed-system processes in the formation of migmatites and migmatitic granulites

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Migmatites are widely interpreted to have been formed by partial melting, particularly by hydrate-breakdown melting reactions, and processes in melt-bearing rocks feature prominently in subsequent sections. However, this view of migmatite formation was not always as widely held as at the present. For example, in the Quetico Metasedimentary Belt layer-parallel leucosomes were interpreted to have formed by a sub-solidus, stress-induced mass transfer process that enabled mobilization of the quartz and feldspar from the host rocks to form leucosomes by pressure solution (Sawyer & Barnes, 1988). In contrast, discordant leucosomes in these rocks were interpreted to represent injected, variably fractionated melts generated during exhumation.

Whether the formation of a particular migmatite or granulite complex was an open- or closed-system process has been a long-standing question in the history of petrogenetic hypotheses for the origin of these rocks. One approach to resolving this issue is through the study of fluids captured as inclusions in the various components that make up a migmatite or granulite at outcrop. These studies do not necessarily provide unique explanations (e.g. Hansen *et al.*, 1984; Hollister, 1988). However, if, for example, CO_2-rich or CO_2–H_2O fluid inclusions are captured only in the minerals comprising the leucosome and not in the minerals of the melanosome in a migmatite outcrop, and if the trapped fluids preserve compositions and densities consistent with the retrieved P–T–a_{H_2O} conditions of migmatization, then an origin by partial melting driven by infiltration of an externally derived fluid, interacting with either graphite-bearing or graphite-free rocks, respectively, is likely (e.g. Whitney, 1992; Giorgetti *et al.*, 1996).

Hypotheses of migmatization driven by infiltration of fluids that are not based in fluid inclusion studies form a parallel theme in a series of contemporary studies discussing the formation of migmatites in diverse settings. Interpretations relate to topics, such as the transfer of subduction-derived components from the slab to the mantle wedge (Sorensen, 1988), relations between rock composition, deformation, transport of H_2O-rich fluid, partial melting and the formation of layered migmatites (Johannes, 1988) and why different parts of an aureole exhibit very different amounts of inferred melting, from locally developed to widespread, and dramatic differences in structure (Pattison & Harte, 1988). In the latter case, the hypothesis of fluid infiltration was subsequently tested using oxygen isotopes and supported (Linklater *et al.*, 1994).

Fluid infiltration is commonly proposed in the case of partial melting of felsic igneous rocks, such as occur in the northern Gallatin Range of Montana, where leucosomes form narrow envelopes adjacent to ductile shear zones that are inferred to be a response to infiltration of H_2O-rich fluids (Mogk, 1992). Mogk proposed a feedback relation between ductile shearing and migmatization through which the felsic igneous rocks were chemically and mechanically reworked by cyclical fluid infiltration, partial melting and melt-enhanced deformation. Similarly, for the Inzie Head gneisses of northeast Scotland, it has been suggested that an influx of H_2O-rich fluid channelled through a regionally extensive shear zone network was necessary to produce early, low volume cordierite-bearing leucosomes in the gneisses (Johnson *et al.*, 2001). After this initial influx of H_2O-rich fluid, Johnson *et al.* argued that the system became fluid-absent and melting reactions buffered a_{H_2O} to lower values with increasing temperature.

In contrast, Maaløe (1992) argued against migmatization by infiltration of fluids and used variations in mineral concentrations across leucosomes to suggest they were generated during anatexis by a melt-mediated diffusive exchange between the leucosomes and the melanosome, and not by the migration of melt from the melanosome to the leucosomes. As the leucosomes increase in width with progressive anatexis, refractory phases, such as biotite stick together and become concentrated at the margins of the developing leucosomes forming selvedges by a process analogous to liquid phase sintering.

Returning to open- *versus* closed-system processes in migmatization, another way to address this question is using mass balance. Bea (1989) introduced a new 'anatexis mixing model,' with appropriate caveats, to characterize quantitatively the behaviour of minerals during partial melting. Others using a mass balance procedure have generally concluded that the system is open during migmatization, either requiring injection of externally derived melt and local metasomatism (Olsen & Grant, 1991) or infiltration of an externally derived fluid, such as a low a_{H_2O} fluid with dissolved solutes that may change in chemistry during the process of migmatization (Pattison, 1991) or alternatively a locally derived H_2O-rich fluid that must be recycled to achieve the melt volumes inferred from the mass balance (Jung *et al.*, 1998).

In contrast, Waters (1988) and Hartel & Pattison (1996) have argued that the granulite-forming process involved hydrate-breakdown melting reactions without fluid infiltration. In an internally buffered system,

hydrate-breakdown melting reactions buffer a_{H_2O} and as the reaction progresses with increasing temperature, so a_{H_2O} is generally driven to lower values. In the case of Namaqualand (Waters, 1988), the occurrence of largely anhydrous neosomes within hosts that preserve hydrous mineral assemblages implies gradients in a_{H_2O}, and the limited occurrence of retrograde hydrous minerals in neosomes suggests that reaction between hydrous melt and anhydrous residue in the neosomes did not occur, leading to the conclusion that H_2O dissolved in the melt was lost from the system on crystallization. In the Kapuskasing migmatitic mafic granulites studied by Hartel & Pattison (1996), during progressive hydrate-breakdown melting of an amphibolite protolith at 1.1 GPa and $\sim$850 °C, quartz was always a reactant suggesting that the abundance of quartz in the protolith was the primary control on the variable extent of reaction progress recorded in these rocks.

Indeed, absence of fluid may be the general case during granulite metamorphism, as suggested by the work of Harley et al. (2002) using the $H_2O + CO_2$ contents of cordierite in equilibrium with melts, although some caution is warranted. In most cases, the volatile contents of cordierite yield calculated activities for $H_2O + CO_2$ that are significantly less than those required for fluid saturation at the P–T conditions of cordierite formation. Whether this reflects absence of fluid, dilution of fluid by other components or leakage of fluid from cordierite is an issue that must be evaluated on a case-by-case basis.

Controversy may arise, as occurred in the discussion by Blattner (2005) of the study by Daczko et al. (2001) and the reply by Clarke et al. (2005). The issue at hand was not open- versus closed-system behaviour, all agree the system was locally open, but whether the infiltrating medium was a metasomatizing, sodium-rich fluid promoting dehydration during passage through a hornblende-bearing granulite converting it to garnet-bearing granulite or a locally derived, volatile scavenging trondhjemitic melt that caused the conversion.

Although high-volume melting at upper amphibolite facies conditions demands influx of H_2O-rich fluid, a source is not commonly identified; but sometimes it is (White et al., 2005; Lancaster et al., 2009). To explain a heterogeneous transition at all scales from metatexite to schlieren and massive diatexite, White et al. (2005) used geochemistry and phase equilibria modelling to support a proposed influx of H_2O-rich fluid at temperatures just above the diatexite solidus. Furthermore, these authors proposed that the fluid was derived from adjacent metapelitic rocks where muscovite breakdown occurred under sub-solidus conditions due to a slightly higher temperature solidus. Lancaster et al. (2009) also used geochemistry, in this case oxygen isotopes and rare earth elements, to test competing models of melt injection v. partial melting for the origin of migmatites in amphibolite-to-granulite facies metasedimentary rocks of the Adirondack Mountains, New York. The high $\delta^{18}O$ obtained from zircon clearly shows that leucosomes are not injected melts derived from local igneous protoliths, but indicates that they and their associated melanosomes were derived by partial melting of a metasedimentary protolith. In addition, rare earth element disequilibrium between zircon rims and garnet in these migmatites demonstrates that ages obtained from zircon may not date the time when the P–T conditions derived using garnet-bearing assemblages were set.

Thermobarometry, *P–T* paths, phase equilibria modelling and retrograde processes in formerly melt-bearing rocks

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An important task in any petrological study is to quantify the *P–T–X* conditions of metamorphism. The first group of papers discussed in this section concerns determination of the *P–T* conditions and definition of the *P–T* paths from metamorphic rocks using combinations of results from thermobarometry and reactions formulated from interpretations of microstructural relations linked to generalized petrogenetic grids. Although evidence of the prograde path is commonly lacking, sometimes inclusion assemblages in garnet enable this to be constrained and allow for a more complete *P–T* path to be constructed (Jones & Brown, 1990).

In various migmatite belts (e.g. Jones & Brown, 1990; Montel *et al.*, 1992; Brown, 1998; Greenfield *et al.*, 1998; Kawakami, 2001), estimates of peak temperatures are broadly similar, within 50° of 800 °C, although pressures vary from moderate (1.0–0.5 GPa) to low (< 0.5 GPa). A high-*T* decompression segment in the *P–T* path may occur in the higher-*P* belts, but the lower-*P* belts commonly have hairpin-type clockwise *P–T* paths or more-or-less isobaric heating and cooling paths. In one example, the time-integrated features of a sequence of anatectic granites demonstrate the progressive dehydration of a deep crustal segment over tens of millions of years (Montel *et al.*, 1992). In another example, near the tourmaline-out isograd, large tourmaline crystals occur in leucosome in interboudin partitions, whereas in the host rocks reaction microstructures involving the breakdown of tourmaline are preserved. These observations imply that tourmaline breakdown is related to a melting reaction. However, at higher grade within the migmatite zone, the almost complete absence of any borosilicate minerals points to the effective removal of boron by loss of melt and/or fluid from the migmatites (Kawakami, 2001).

Much of what can be retrieved from formerly suprasolidus migmatites and granulites is information related to retrograde reaction. The extent of retrograde reaction, including the formation of reaction selvedges between leucosome and melanosome, is a function of the proportion of melt produced that has been lost from the system and the rate of change of size of the equilibration volumes as temperature declines (Brown, 2002; see also White *et al.*, 2001; White & Powell, 2002). Because melt loss generally has occurred, rarely is retrograde reaction more than partial. However, it is this very feature that allows a *P–T* path to be constrained (Raith *et al.*, 1997; Cenki *et al.*, 2002).

Melting may occur under very fast rates of heating. For example, mid-crustal pelitic granulites in the Vredefort Dome record evidence of a near-instantaneous, short-lived low–*P* granulite facies event due to a meteorite impact (Gibson, 2002). Maximum temperatures recorded within a few kilometres of the centre of the dome were > 900 °C, locally > 1350 °C, which led to comprehensive melting of the precursor granulites and segregation of biotite leucogranites in the centre of the dome. However, the pre-impact features of the granulites are largely preserved further away even though melting occurred, and coronae and symplectites partially replace precursor minerals allowing the retrograde evolution to be constrained.

At the most extreme *P–T* conditions recorded by crustal rocks, evidence of the prograde path commonly has been completely overprinted. Nonetheless, these rocks generally preserve evidence in the form of a sequence of suprasolidus reaction microstructures to enable the retrograde *P–T* path to be constrained. For example, regional ultrahigh temperature (UHT) metamorphism in southern Peru produced stromatic migmatites with high Al contents in orthopyroxene (~10 wt% Al_2O_3) and the local occurrence of sapphirine–quartz assemblages, consistent with $T > 900$ °C at $P > 1.0$ GPa (Martignole & Martelat, 2003). A plethora of reaction microstructures in these rocks enables construction of a stepped decompression–cooling–decompression suprasolidus retrograde *P–T* path. In northeast Greenland, where coesite eclogites record ultrahigh-pressure (UHP) metamorphism reaching 970 °C and 3.6 GPa, associated metapelites display abundant layers and lenses composed of quartz, antiperthitic plagioclase and K-feldspar with minor garnet and kyanite and host garnet with polyphase inclusions consistent with melting via phengite + coesite $\Rightarrow$ kyanite + K-feldspar + rutile + melt (Lang & Gilotti, 2007). For the peak UHP phase assemblage of phengite–coesite–garnet–kyanite–rutile–fluid $\pm$ omphacite $\pm$ epidote, phase diagrams indicate that phengite-breakdown melting would have occurred after initial decompression from peak-*P* but still within the coesite stability field.

The extension of a published thermodynamic model for haplogranitic melts in the $Na_2O–CaO–K_2O–Al_2O_3–SiO_2–H_2O$ system to include end-members for iron and magnesium allowed modelling of melting reactions to be undertaken in the more complete $Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O$ system (White *et al.*, 2001). Using this system allowed investigation of crustal melting in a more quantitative way for a variety of metasedimentary protolith compositions, including the ability to address questions relating to the open-system behaviour

of these rocks at suprasolidus conditions. As a result, a quantitative explanation for the preservation of granulite facies mineral assemblages was developed (White & Powell, 2002). P–T paths derived from migmatites and granulites became quantitative (Harris *et al.*, 2004), processes such as diapiric rise could be tracked by quantitative interpretation of microstructures and the implied reaction relations in rocks (Johnson *et al.*, 2004), the generation and preservation of garnet-bearing leucosomes in migmatitic granulites could be better explained (White *et al.*, 2004), and differences in the rare-earth element concentrations of orthopyroxene in garnet-bearing versus garnet-free hosts and in associated leucosomes could be related to decoupling between the behaviour of the major elements and the rare-earth elements (Clarke *et al.*, 2007).

Subsequently, White *et al.* (2007) introduced improved activity–composition relationships for biotite, garnet and silicate liquid to enable calculations in the Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O–TiO_2–O_2 system, which is the most comprehensive system currently available for modelling of this type. As a result, much has been learnt about processes operating in suprasolidus crust from generalized modelling of metapelitic (White *et al.*, 2007) and metagreywacke compositions (e.g. Johnson *et al.*, 2008). Grant (2009) compared the results obtained from

melting experiments on natural rock samples at low-P with pseudosections constructed for the same initial compositions and demonstrated that the experimental results and the calculated model are in good agreement. Overall, we can be confident that during the last decade we have moved closer towards an understanding of what has happened in crustal melting in nature.

Using a recently published thermodynamic model for sapphirine, Baldwin *et al.* (2005) undertook quantitative studies of sapphirine-bearing UHT granulites to determine peak metamorphic conditions and to investigate double crossing of the reaction orthopyroxene + sillimanite $\Longleftrightarrow$ garnet + sapphirine + quartz, which constrains a two-stage isothermal decompression–isobaric cooling path. Additionally, these authors showed that the presence or absence of cordierite in the post-peak mineral assemblage might be explained by local variation in a_{H2O}. We are familiar with crystallization of melt due to cooling and also by pressure quenching, but crystallization also may occur by loss of H_2O through diffusion to the host (White & Powell, 2010). This diffusive process may result in the development of anhydrous leucosomes in migmatites and granulites with only minor retrogression of the melanosome, and the extent of diffusion of H_2O may determine whether or not retrograde selvedges are developed between the leucosomes and the melanosome.

Geochronology in partially melted rocks

EDITED BY MICHAEL BROWN

An important goal in studies of crustal melting is to link *P–T* determinations with time, particularly to understand rates of heating, lengths of melting episodes and rates of cooling and exhumation. Although accessory minerals with a high closure temperature for diffusion of Pb, for example monazite and zircon, are most suitable for dating close to peak *P–T* conditions for partially melted rocks, linking to a *P–T* point or range along the *P–T* path presents challenges, as these minerals may survive from the precursors, recrystallize around peak conditions or exhibit new growth as melt crystallizes along the retrograde path. Nowadays these alternatives potentially may be discriminated using microstructural characterization, micro-beam chemical analysis, accessory phase thermometry and trace element distributions between accessory minerals and associated rock-forming minerals to provide constraints on whether grains are relict from the precursor or record the timing of recrystallization and/or new growth in relation to the *P–T* evolution.

In a pair of early papers on zircon in migmatites, Gupta & Johannes (1985) and Dirks & Hand (1991) documented systematic changes in detrital zircon populations related to the beginning of partial melting and the onset of dissolution of zircon, leading to the progressive loss of zircon by dissolution with increasing melt volume. Dirks & Hand (1991) also suggested that melt loss may limit new zircon growth during cooling and crystallization of residual melt.

Other studies have used a variety of different isotope systems to date a partial melting event. Suzuki & Adachi (1998) used the chemical Th–U–total Pb isochron method to date monazite in migmatites and granites from the western and eastern parts of the Ryoke metamorphic belt in Japan. For the migmatites, the monazite ages are between *c.* 102 and *c.* 98 Ma from both the western and eastern parts of the belt, and as the ages are interpreted to date crystallization they disprove the hypothesis of a diachronous metamorphic event related to a migrating trench–ridge–trench triple junction. Jung & Mezger (2001) used TIMS Sm–Nd garnet and U–Pb monazite ages from migmatites in the Damara orogen, Namibia, to constrain the timing of migmatization to the interval *c.* 535 to 520 Ma. Late veins emplaced at *c.* 505 Ma and younger Sm–Nd garnet ages of *c.* 500 Ma are interpreted to record a late reheating event, whereas variably younger Rb–Sr biotite ages from the different components of the migmatites are interpreted to reflect a later disturbance of the Rb–Sr isotopic system on the sub-sample scale.

Four papers published in the JMG during the past decade, two from the same region, report detailed studies of complex age systematics in zircon, and in one case data for monazite are also reported. Moller

et al. (2002) reported SHRIMP U–Pb zircon ages from the high-grade aureole of the anorthosite–norite intrusive complex, in Rogaland, Norway. For samples from outside the hypersthene-in isograd, ages of 1020–1007 Ma were retrieved from metamorphic rims on magmatic zircon cores, which correspond to the regional Sveconorwegian granulite facies metamorphism. In contrast, samples from inside the pigeonite-in isograd exhibit multiple ages, two groups of magmatic cores and two groups of metamorphic overgrowths, the first corresponding to the Sveconorwegian event and the second corresponding to the contact metamorphism giving an age of 922 ± 14 Ma. What this study demonstrates is the extraordinary survival of magmatic and metamorphic zircon relicts in rocks that experienced a thermal overprint of ~950 °C for at least 1 Ma, leading to the conclusion that resetting of zircon ages by diffusion in dry granulite facies rocks of the aureole was negligible.

Tomkins *et al.* (2005) reported data for the regionally distributed migmatites from Rogaland, where some of the zircon is related to the retrograde reaction garnet + sillimanite + quartz → cordierite + zircon. This zircon is texturally, chemically and isotopically distinct from an older generation of zircon (1035 ± 9 Ma) in other textural settings in the matrix interpreted to record growth during prograde partial melting of the rocks at ~700 °C at 0.6–0.8 GPa. Thus, the SHRIMP U–Pb age of 955 ± 8 Ma obtained from zircon in cordierite may be linked to a point on the retrograde *P–T* path, estimated from garnet–cordierite equilibria to be 710 °C at 0.56 GPa. Zircon growth occurred during both the prograde and retrograde segments of the Sveconorwegian event, which appears to have lasted at least 70 Ma based on these data, although this range is shorter, if based on the data reported by Moller *et al.* (2002; 1017–992 Ma).

In the third example, Hermann & Rubatto (2003) identified three metamorphic overgrowths on inherited zircon from granulites exposed in Val Malenco in northern Italy. These authors use mineral inclusions in zircon and garnet and rare earth element distributions to relate the ages to prograde mica-breakdown melting (281 ± 2 Ma), growth of zircon from melts depleted in heavy rare earth elements by crystallization of garnet at 269 ± 3 Ma and a second stage of partial melting documented by growth of new garnet at 258 ± 4 Ma. Monazite yields similar ages of 279 ± 5, 270 ± 5 and 257 ± 4 Ma, and although it displays less clear relationships between growth zones and mineral inclusions than zircon, differences in the trace element patterns of the two older generations of monazite allow them to be distinguished. Overall, the granulite facies partial melting event lasted for *c.* 20 Ma. This study reports trace element partitioning between coeval garnet,

zircon and monazite that may be used in future to evaluate equilibrium between these minerals.

Following the advances by Hermann & Rubatto (2003) in linking ages to specific steps in the *P–T* evolution and subsequent expansion of this approach, Baldwin & Brown (2008) reported an integrated study in which the petrology and chemistry of zircon, garnet and rutile from melt-depleted ultrahigh temperature granulites in the Anápolis–Itauçu Complex in Brazil are used to constrain the significance of zircon ID-TIMS U–Pb ages. These authors also use Ti-in-zircon thermometry to argue that zircon growth relates to both the prograde and retrograde stages but not to the peak temperature stage. ID-TIMS U–Pb data from two of the four samples investigated, define upper intercept ages of 641.3 $\pm$ 8.4 Ma and 638.8 $\pm$ 2.5 Ma that correlate with periods of zircon growth along the prograde segment of the *P–T* path. Individual zircon U–Pb dates retrieved from all samples range from 649 to 634 Ma, indicating a maximum duration of *c.* 15 Ma for the suprasolidus ultrahigh temperature event.

In an important step forward, Kelsey *et al.* (2008) integrated experimental data on zircon and monazite dissolution and growth with phase equilibria by expressing the growth and dissolution results as contours on *P–T* pseudosections. Using this information, these authors proposed an explanation for why zircon ages are generally older than monazite ages in aluminous metapelites, suggested that the growth rate of accessory minerals is non-linear, and argued that the difference in zircon and monazite ages from the same rock may be ascribed to differences in the temperature of growth rather than differences in closure temperature systematics. Although this study provided a predictive framework for the dissolution and growth of zircon and monazite as a function of *P–T–X* conditions because Zr and LREE were not explicitly included in thermodynamic models used to calculate the phase diagram on which the dissolution contours were superimposed, it was not a fully quantitative approach. A quantitative approach followed in Kelsey & Powell (2011), where *P–T* pseudosections that explicitly include ZrO_2 in the bulk composition are used to predict the dissolution and growth of zircon within the context of the evolution of the rock-forming mineral assemblages in *P–T–X* space for a metapelite. This represents another important step forward in understanding the controls on the stability of zircon and in linking zircon U–Pb ages and Ti-in-zircon thermometry to the evolution of rock-forming minerals in a wide range of rocks.

Microstructures, deformation and tectonics of melt-bearing rocks

EDITED BY MICHAEL BROWN

The time-scale of a metamorphic event and whether penetrative deformation occurred during partial melting are important variables. Partial melting in shallow contact metamorphic environments, where the time-scale may be very brief indeed, commonly occurs in the absence of significant applied stress. As a result the features produced during partial melting in the inner parts of contact metamorphic aureoles and the microstructures of the melted rocks tend to be well preserved in comparison with those formed in deeper regional metamorphic environments, where deformation is an important element controlling the macroscopic features of migmatites and the microstructures commonly have been modified during slow cooling.

This section leads off with a study by Holness *et al.* (2005), which is included to illustrate the microstructures that are produced by melting in the case of very short-lived contact metamorphism. In the aureole of the Glenmore plug, a decametric Tertiary dolerite intrusion in Ardnamurchan, Scotland, melting initially occurred along quartz–feldspar grain boundaries producing films of melt, now glass, the distribution and thickness of which were controlled by the distribution of limited amounts of H_2O on the grain boundaries. In the inner aureole, within 0.6 m of the contact where temperatures were high enough for fluid-absent melting in the Qtz–Ab–Or system, layer-parallel segregations and cross-cutting veins of glass up to 1 cm thick that form branching arrays demonstrate local melt movement. This was most likely due to internally generated overpressure, as the melt volume increased to more than 40% of the rock.

In the next study, McLellan (1983) argued that differences at the grain scale in the spatial distributions of minerals in migmatites could be used to distinguish those formed by solid-state differentiation—characterized by aggregate distributions—from those formed by partial melting—characterized by random distributions—during a regional metamorphic event. A transition from sub-solidus to anatectic processes is to be anticipated, and such a transition is described by Dallain *et al.* (1999); within the transition some minerals within aggregates exhibit growth related to the presence of melt (e.g. K-feldspar in plagioclase-dominated aggregates), whereas in others aggregates remain melt-free (e.g. K-feldspar-dominated aggregates).

The relationship between crustal melting and deformation has been studied at all scales confirming that the strain field influences melt-flow patterns and that applied differential stress is a critical driver of melt segregation and extraction from partially molten crust. In more slowly cooled melt-bearing rocks than the aureole at Glenmore, feldspar (or quartz) commonly mimics the grain-scale distribution of former melt pockets, where quartz (or feldspar) is the major constituent of the solid matrix. Using this microstructural feature of anatectic rocks, Marchildon & Brown (2002) were able to relate the grain-scale spatial arrangement of feldspar pseudomorphs after melt in layer-parallel leucosomes in rocks from two different contact aureoles to different controls. In one example, feldspar inferred to replace melt is aligned sub-parallel to the shape-preferred orientation of quartz, indicating that pre- or syn-anatectic strain controlled the grain-scale distribution of melt. In the other example, the preferred orientation of feldspar inferred to replace melt is different from the orientations of all other mesoscopic or microscopic structures in the rock, indicating that differential stress controlled grain-scale melt localization; the differential stress in this case was inferred to have been higher than in the first case.

Hasalová *et al.* (2008a, b) described a gradual transition from high-grade solid-state layered orthogneisses through stromatic and schlieren migmatites to foliation-parallel bodies of nebulitic migmatite in the eastern part of the Gföhl Unit in the Bohemian Massif. Based on a microstructural study, these authors argued that syn-anatectic deformation occurred by low volume, melt-enhanced granular flow possibly associated with melt flow through the rocks. Although these rocks have a common mineral assemblage, modes and mineral compositions exhibit systematic changes across the transition from layered orthogneiss to nebulite that are consistent with a decrease in *P–T* conditions and modification of whole-rock compositions, features that suggest open-system behaviour. Phase equilibria modelling allows an interpretation that a large volume of melt passed through these rocks and interacted with them en route.

The role of melt in the deformation of crustal rocks and the effects of different melting reactions on the mechanisms of melt extraction have been evaluated using rock deformation experiments. The effect of melt on deformation depends on the amount and distribution of the melt, the grain size of the matrix and the strain rate. Dell'Angelo & Tullis (1988) reported the results of experiments on samples of leucogranite of various grain sizes, with the addition of different amounts of water to vary the amount of melt produced, at two different strain rates. In extending their results to natural strain rates, these authors argued that melt-enhanced diffusion creep might be expected to dominate at larger grain sizes, and generally there should be little or no cataclasis if an equilibrium distribution of melt exists and if the melt can flow as fast as the imposed strain rate. Holyoke & Rushmer (2002) investigated the role of muscovite-breakdown and biotite-breakdown melting in the development of

permeability in a muscovite–biotite metapelitic schist and a biotite gneiss, respectively, under lower crustal conditions. It was argued that muscovite-breakdown melting is likely to develop melt migration pathways close to the beginning of melting due to the positive ΔV of reaction and the large volume of melt produced close to the solidus, which together generate high pore pressures, potentially enabling fast melt extraction. In contrast, the ΔV of reaction for biotite breakdown is not as large and melting progresses over a wider range of temperature; as a result the melt volume increases more gradually, and melt migration pathways are more likely to be controlled by structural anisotropies in the protolith. A review and reinterpretation of previous experimental data on the deformation of melt-bearing crustal rocks by Rosenberg & Handy (2005) reveals that the relationship between aggregate strength and melt fraction is non-linear, with a dramatic drop in strength between the solidus and a melt fraction of 0.07, which coincides with melt along ~80% of the grain boundaries. These authors suggested that it is the increase of melt interconnectivity within the first few volume per cent of melt generation that causes the dramatic drop in strength observed, which will lead to crustal-scale localization and facilitate melt extraction.

Around the inlet of Boca de Quadra in southeast Alaska, McLellan (1988) identified two types of anatectic migmatites, patch migmatites—representing melting and *in situ* crystallization—and diktyonitic migmatites—representing segregation of melt into a discontinuous vein network surrounding residual boudins of host rock. It is argued that these differences are not related to protolith composition, but are more likely due to differences in the amount of applied differential stress during anatexis. Continuing the topic of deformation and melt migration, Allibone & Norris (1992) described a suite of migmatites in which the morphology is directly related to their position relative to the hinge of a km-scale antiform. Stromatic migmatites on the fold limbs are cut by leucosome-filled extensional shear bands and fractures, whereas veins and dykes of leucosome become more common and thicker from the limb to the hinge region of the antiform, where major parasitic folds have trapped migrating melt in microplutons up to 50 m across, beneath refractory layers of the host.

The structural control on melt migration is exemplified well by the study of Collins & Sawyer (1996), who described a diverse network composed of narrow structurally controlled channels in migmatitic granulites comprising pre-existing folds, interboudin partitions and shear bands and newly created channels formed by melt fracture parallel to layering and in irregular arrays; melt ascended obliquely, parallel to the plunging regional elongation direction. These observations suggest that melts from the deep crust are capable of pervasive migration through the middle crust to accumulate as plutons in the shallow crust during deformation.

Sawyer (2001) presented grain- and outcrop-scale maps of leucosome distributions in regional and contact metamorphic environments. At the grain scale, although most melt was located parallel to the principal fabric in the rock, former melt-bearing grain boundaries connect melt-depleted parts of the matrix to diffuse zones of melt accumulation. At the outcrop scale residual rocks show where melt has been removed, whereas leucosomes preserve channels through which melt passed or represent sites where it pooled or structures through which it escaped. In another study, Guernina & Sawyer (2003) determined the grain- and outcrop-scale melt-flow paths during melt segregation and estimate how much melt was formed and extracted from a large (90 000 km^2), mostly metagreywacke granulite facies terrane. The metagreywackes produced an average of 31 vol.% melt, and mass balance indicates that ~640 000 km^3 of melt was almost completely extracted from this residual granulite terrane during deformation.

To explain high-T–low-P metamorphic belts, Sandiford & Powell (1991) proposed that advection of significant heat through the orogenic crust is required, for example by the extraction and ascent of anatectic melts. As an example, consider the Higo Metamorphic Complex in Japan (Miyazaki, 2004). Reconstruction of the geological and metamorphic structure of the Complex reveals a metamorphic field P–T array, with a lower dP/dT slope at higher structural levels and a higher dP/dT slope at lower structural levels. Thermal modelling shows that the composite thermal structure may be formed by a change of the dominant heat transfer from an advective regime with melt migration and solidification to a conductive regime with decreasing depth.

Many migmatite terranes are characterized by decompression at high-temperature. Carson *et al.* (1992) demonstrated that the high-grade retrograde evolution of anatectic migmatites from the Larsemann Hills in East Antarctica was dominated by decompression that generated extensive partial melting; syntectonic pegmatites and extensive granite plutons represent the fugitive melt. Similarly, Brown & Dallmeyer (1996) showed that the high-grade migmatitic core to the southern Brittany metamorphic belt records petrological features that indicate high-temperature decompression. These authors use extant geochronology and new $^{40}Ar/^{39}Ar$ data to suggest a minimum time-integrated average cooling rate for the temperature interval 725–125 °C of ~14 ± 4 °C Ma^{-1}, although below 600 °C the data permit an infinitely fast rate of cooling. Mineral assemblages and reaction microstructures in diatexite migmatites suggest ~0.4 GPa decompression at 800–750 °C, which must have pre-dated the rapid cooling. The high-grade migmatitic core of the southern Brittany metamorphic belt is interpreted as a type of metamorphic core complex formed during orogen-parallel extensional unroofing and regional-scale ductile flow.