A matter of time: The importance of the duration of UHT metamorphism

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The duration of granulite (G) and ultra-high temperature (UHT) conditions in regional metamorphism is critical to arguments regarding the tectonic settings of granulites and their relationships to Supercontinent assembly. Analysis of zircon geochronology integrated with trace element (REE) constraints on the timing of zircon growth or modification and evidence for metamorphic temperatures from Ti-in-zircon reveals that zircon can form episodically at or near the metamorphic peak in long-lived UHT granulites. This reflects the segregation, transfer and stagnation or trapping of melts which leads to local melt-rock interactions that promotes zircon crystallisation. In contrast, although rutile can retain UHT information in short duration ('fast') G-UHT terrains it is afflicted by Zr loss in long duration ('slow') terrains and yields temperatures significantly lower than those preserved by zircons formed in the same G-UHT events. An assessment of the age-duration evidence from several well-documented G-UHT terrains reveals that most are 'slow', having durations of UHT, Δt_{900} , in the range 30–100 Myr. Some UHT terrains previously considered to be short-lived ($\Delta t_{900} < 10$ Myr) have longer durations of UHT in the light of recent geochronology and hence are also classed as 'slow'. Of the models proposed to account for UHT metamorphism, the 'large hot orogen' (LHO) model for collisional orogeny provides the best setting for the formation of such 'slow' UHT granulites. LHO models can account not only for the P - T paths, which can range from UHT with near-isothermal decompression (UHT-ITD) through to decompression-cooling and UHT followed by near-isobaric cooling (UHT-IBC), but also for residence times under UHT conditions. The Napier Complex, the Earth's premier UHT terrain, probably formed as trapped deep crust in the hot underbelly of a late-Archaean LHO. Shorter duration UHT and near-UHT granulites also exist, mostly with Δt_{900} less than 10 Myr. A number of these are likely to have formed as a consequence of severe lithospheric thinning and crustal extension accompanied by voluminous magmatism, which could occur in arc and back-arc settings affected by subduction roll-back or, as advocated by previous workers, continental arcs undergoing extension. However, attaining long-lived UHT conditions in these settings is unlikely unless the crust has inherently high radioactive heat production in addition to the transient heat added during extension and magmatism.

Keywords: Ultrahigh temperature (UHT) metamorphism, U-Pb geochronology, Zircon, Monazite, Large Hot Orogen, Napier Complex

INTRODUCTION

It is now widely recognised that ultrahigh temperature (UHT) metamorphism is an important variant of crustal behaviour that has occurred throughout much of Earth history. Largely as a result of the developments encapsulated in the reviews by Harley and co-workers (Harley 1998a, 1998b, 2004, 2008; Clark et al., 2011) and Kelsey and co-workers (Kelsey, 2008; Kelsey and Hand, 2015) it

is now accepted that granulite, and in particular UHT, metamorphism plays a fundamental role in the development and stabilisation of continents through accretionary and collisional orogenesis, as well as being of importance in extensional environments. For example, with over 60 UHT areas or localities now recognised globally (Kelsey and Hand, 2015), which formed at various times from the mid Archaean through to the Cenozoic (Brown, 2006), it is clear that understanding this type of extreme metamorphism is critical to constraining the large-scale tectonic processes affecting the deep crust and lithosphere throughout Earth history. Brown (2006, 2009) and Sizova

et al. (2010, 2014) have further argued that granulite and UHT metamorphism, or high (dT/dz) metamorphism more generally, are intrinsically linked to major episodes of Supercontinent amalgamation at least since the late Archaean — their preservation being facilitated by isolation within collisional belts internal to the Supercontinents of Gondwana, Rodinia, Columbia or Nuna and perhaps Superia and Scelavia.

The application of calculated phase diagrams to quantify the P - T conditions of granulites and UHT rocks (G-UHT metamorphism) has been described and evaluated in some detail in recent reviews (Kelsey, 2008; Kelsey and Hand, 2015), and it is now largely accepted that this approach (i.e., the pseudosection methodology based on the pioneering theoretical work of B.J. Hensen) — when used carefully with due regard for system complexity, melt production and transfer and domainal equilibrium — provides important insights into granulite and UHT P - T conditions and histories for many rock compositions. Major mineral thermobarometry (e.g., ternary feldspars and pyroxenes, Al in orthopyroxene) and trace element thermometry of accessory (e.g., zircon, rutile) and major (e.g., quartz, garnet) minerals, described and evaluated in general terms by Harley (2008), Kelsey (2008) and Kelsey and Hand (2015), are also shown to be valuable recorders of the P - T conditions of G-UHT rocks, once again providing due consideration is given to the effects of post-peak diffusional exchange, exsolution or decomposition, and the timing of mineral growth — particularly for the accessory minerals in relation to the major mineral assemblages. Aspects of the latter issue in the characterisation of G-UHT, for example the crystallisation of zircon in partially-melted rocks and its potential inhibition by the stability of Zr-rich rutile or sequestration of Zr in major phases such as garnet, have been considered by Kelsey et al. (2008) and Kelsey and Powell (2011).

These approaches highlight the key issue of the age and duration of G-UHT metamorphism: whilst we may be able to quite accurately constrain a P - T condition for peak metamorphism or indeed a series of P - T conditions reflecting mineral reactions preserved along the G-UHT P - T path, defining the ages of those events and hence the timescale or duration of G-UHT in any one terrain, or even rock, remains problematic. The duration of G-UHT, expressed in terms of its T - t curve, is an important parameter that potentially allows discrimination of viable tectonic models for the generation and context of the G-UHT event from those that require timescales different to those recorded.

In a recent review of UHT metamorphism Kelsey and Hand (2015) tabled and described the durations of

Figure 1. Schematic Temperature-duration (Δt) plot illustrating six types of thermal excursion into the UHT regime. Time is relative, with broader peaks signifying longer durations at G-UHT conditions. See text for discussion. Color version is available online from <http://doi.org/10.2465/jmps.160128>.

G-UHT meta morphism for a number of terrains. These durations ranged, in their synthesis, from 80–100 Myr down to a few (5–10) Myr, with most terrains having durations of G-UHT of less than 30 Myr. They inferred a broad continuum in terms of duration, although the data could be interpreted equally in terms of a duality of typical timescales: those UHT belts with durations greater than 20 Myr (most of them) and a few UHT belts with durations significantly less than 20 Myr. The purpose of this paper is to elucidate some of the issues surrounding the definition of durations of UHT metamorphism and re-examine the duration data for some key UHT regions to address the central question: are most G-UHT (and hence UHT) terranes ‘slow’ or ‘fast’?

Figure 1 illustrates a variety of potential T -time (T - t) evolutions for G-UHT terrains. The time axis is relative, with the longer timescales being 30 to 100 Myr. Case ① represents a simple increasing T path to peak, followed by decreasing T on a similar cooling rate to the heating rate, with no long-lived thermal plateau at or near the peak G-UHT condition. Case ② differs from ① in having a long-lived UHT thermal plateau at peak and near-peak T conditions, with this plateau being terminated by rapid cooling to below 800 °C. Case ③ is similar to ② in having a long-lived UHT thermal plateau, but this plateau is followed by slow cooling through the granulite facies. Case ④ is a short-lived G-UHT excursion, involving rapid heating and/or cooling. Case ⑤ represents long-lived granulite metamorphism (G) with a short-lived thermal excursion into UHT conditions. Polymetamorphism [case ⑥], with two or more G-UHT events separated by significant episodes of cooling, could appear

similar to cases ①, ② and ③ in the absence of definitive evidence for intervening lower- T events, for example the deposition of new sediments after the first event, or the chilling of dyke swarms intruded between events.

In each case we may characterise the thermal history, whatever the depth or dP/dt record, in terms of the duration of metamorphism above two thresholds: major granulite metamorphism at $\sim 800^\circ\text{C}$ and UHT at 900°C . The duration of UHT metamorphism is then defined as Δt_{900} (or Δt_{UHT}), and that of granulite conditions as Δt_{800} . The rate of cooling accompanying the termination of the G-UHT event is also significant, as for peak metamorphism at the equivalent of lower- to mid-crustal depths ($P = 7\text{--}12$ kbar) rapid cooling would require significant decompression and exhumation (Harley, 1989, 1998a, 1998b; Clark et al., 2011; Jamieson and Beaumont, 2013). For a tectonic model of granulite metamorphism, and especially UHT, to be viable it must meet at least two necessary conditions: 1) the setting and model achieves peak conditions at appropriate pressures as recorded by the G-UHT rocks, and 2) the duration of G-UHT, Δt_{900} and Δt_{800} , are commensurate with geochronological T -time records for the terrain. A third condition which could be applied is that the model P - T path (or P - T - t path) matches the observed P - T (t) path, but as it is seldom possible to uniquely reconstruct the whole prograde to peak and to post-peak P - T path for a terrain this condition may be less restrictive and diagnostic (e.g., Harley, 1989 in reference to IBC granulites; Kelsey and Hand, 2015 in reference to UHT in general).

MODELS FOR UHT METAMORPHISM AND ITS DURATION

There is considerable debate as to the tectonic settings of granulite metamorphism, particularly with respect to the importance of granulite formation related to collisional orogeny versus formation within arc-backarc accretionary settings where lithospheric thinning and magmatic underplating may add significantly to the thermal budget (Ellis, 1980, 1987; Harley, 1989; Sandiford and Powell, 1990; Bohlen, 1991; Collins and Vernon, 1991; Warren and Ellis, 1996). This short-duration high- T metamorphism of recently-formed protoliths potentially contrasts with the generally longer-duration G-UHT metamorphism of varied lithologies of different ages that might be expected in 'hot' collisional orogens (Jamieson et al., 2004; Beaumont et al., 2006; Jamieson et al., 2010; Jamieson and Beaumont, 2011).

Models of G-UHT metamorphism in orogenic belts must allow the middle to deep crust to remain at $T > 800^\circ\text{C}$, and even $>900^\circ\text{C}$ in the case of UHT, for timescales

of the order of 5 Myr to >60 Myr (Harley, 1998a, 2008; Kelsey, 2008; Clark et al., 2011; Kelsey and Hand, 2015). In the following section some important aspects of these potential settings are explored in more detail, with an emphasis on the durations of G-UHT metamorphism, Δt_{900} and Δt_{800} , generated in these settings.

Arc, back-arc and extensional-accretionary settings for G-UHT metamorphism

At one end of the spectrum of thermotectonic models, magmatic underplating involving the emplacement of voluminous crustal and mantle-derived or hybrid magmas over timescales from 1 Myr to 10's Myr can give rise to contemporaneous regional scale granulite facies metamorphism. Wells (1980) generated granulite metamorphism at 7–10 kbar with timescales of greater than 20 Myr in his pioneering 1-dimensional thermal modelling of magmatic underaccretion. Such magmatic loading, both thermal and baric, was preferred as the principal cause of granulite metamorphism by several workers in the 1980's and more recently, and in particular was strongly advocated by Bohlen (1991) who preferred arc settings for most granulite genesis.

The role of magmatic underaccretion and heat advection from the mantle is important in extensional-accretionary settings, such as back arc basins, for which recent modelling indicates residency of deep crust at 800°C for periods of several Myr (e.g., Sizova et al., 2010; Clark et al., 2011). Recent studies have focused on G-UHT metamorphism as a result of post-extensional thickening and inversion of back-arc basins, with associated arc accretion or arc-arc collision (Figure 2: Collins, 2002; Currie and Hyndman, 2006; Kemp et al., 2007). The favourable thermal features of the back-arc setting for generating G-UHT are the high initial thermal gradients and heat flows that result from the extension of mainly tectonically young crust above a shallow asthenosphere, and the concomitant advection of significant heat through the intrusion and eruption of syn-extensional mainly juvenile high- T magmas (e.g., Sizova et al., 2010). This accretionary tectonic setting, and especially the closure and thickening of continental back arcs, has in more recent syntheses been advocated as central to granulite formation and potentially supercontinent development through time (e.g., Brown, 2006, 2007).

A key characteristic of the arc-back arc setting for G-UHT metamorphism is the strong temporal link between magmatism, accretionary crustal growth and granulite development. Even in scenarios where older continental crust is present there is still a high proportion of juvenile crustal material produced and added in order to attain the

Figure 2. Modified after Collins (2002), with an indicative 900 °C isotherm superimposed based on the thermal models of Sizova et al. (2010). Color version is available online from <http://doi.org/10.2465/jmps.160128>.

thermal conditions appropriate for granulite development. In these settings most sedimentation, magmatism and metamorphism occurs within a relatively short geological time interval (5–20 Myr). Collins (2002) applied an arc extension / accretion model to explain the origin of the Palaeozoic low-*P* high-grade metamorphic record of eastern Australia, suggesting that granulite conditions would occur at depth and be enhanced in duration through the post-extensional accretion to collision episode. Similar arguments have been presented by Hyndman et al. (2005) and Currie and Hyndman (2006), who highlight the high heat flows of areas such as the Basin and Range to advocate for the role of continental arc extension in producing low-*P* high-grade and granulite metamorphism. Figure 2, based on and modified from Collins (2002), illustrates the tectonic scenario and presents two time snapshots. An initial phase of slab roll-back leads to high-*T* at shallow depths in the back-arc area (Fig. 2a), accommodating magmatic underplating and advective transfer of heat. Later flat subduction in this model leads to arc and back-arc compression and the burial of the pre-heated crust to *P-T* conditions appropriate for granulite formation. Added to the cartoons are indicative 900 °C isotherms based on the modelling by Sizova et al. (2014)

of a crust-mantle system with an elevated potential temperature (present T_p plus 150 °C). The latter modelling demonstrates that short-lived, or ‘fast’, UHT can be generated on regional scales in the lowermost parts of the extended crust during extension and magmatic underplating or intraplate. However, the short timescales involved in their models (<5 Myr) suggests that the high temperatures associated with back-arc extension are dissipated too rapidly to enable the formation of UHT granulites during the subsequent compressional phase even in a hotter Earth, unless the thickening occurs very soon after extension or the crustal rocks have anomalously high inherent radiogenic heat production.

The results of Sizova et al. (2014) are in general terms consistent with the simpler 1-dimensional modelling by Clark et al. (2011), who examined the potential for generation of UHT metamorphism in back arc settings in which initial extension is followed by basin closure and crustal thickening. Their results indicated that temperatures of 800 °C could be attained at 40 km post-thickening depths but only after a time-lag of 30–35 Myr dependent principally on the internal heat production of the crust, rather than the initial syn-extensional high heat flow. UHT conditions could not be attained for any reasonable

combination of thermal, erosional and lithospheric parameters in their modelling, indicating that the thickening of back-arc basins is not likely to be an effective means of forming long-lived UHT metamorphic terrains.

In summary, the evidence from thermal modelling is that the arc and back-arc setting is unlikely to generate long-duration UHT conditions over reasonable thicknesses of the deep continental crust and at pressures of greater than 7 kbar (~ 21–24 kms) during the extensional phase, and later arc compression and collision does not by itself enable UHT conditions to be sustained. Nevertheless, there are a number of relatively ‘fast’, or short-lived, G-UHT regions in which Δt_{800} and/or Δt_{900} are less than 5–10 Myr. Some examples include the young (<25 Myr) G-UHT rocks of Hidaka, Hokkaido (Kemp et al., 2007), Seram, Indonesia (Pownall et al., 2014) and the Betics of southern Spain (Platt et al., 1998). These, and granulites from somewhat older orogens in which UHT may have been approached or attained for short periods of time, are discussed in a subsequent section.

Large hot orogens (LHO) and G-UHT metamorphism

At the other end of the spectrum of potential thermotectonic models, prolonged timescales for G-UHT metamorphism are indicated for the interiors of large hot orogens (LHO), exemplified by the channel flow model for the Himalayas (Beaumont et al., 2001, 2004; Jamieson et al., 2004; Beaumont et al., 2006), and incipient channel and indentation or nappe variations applied to the Grenville Orogen (e.g., Beaumont et al., 2010; Jamieson et al., 2010), as illustrated in Figure 3. In such large-scale collisional belts, and in the absence of lateral escape processes, deep crust trapped in an orogen interior could reside at >800 °C and even UHT conditions for periods of 30–100 Myr (Jamieson and Beaumont, 2011, 2013; Fig. 3)

In a LHO the input of relatively high heat production crust into the collision zone over time is accommodated not only by crustal thickening but eventually by lateral widening of the orogen to form a topographic plateau (Jamieson and Beaumont, 2011, 2013). Broadening of the orogenic plateau is facilitated by the lateral flow of thermally weakened and generally melt-bearing middle to deep crust under the influence of gravity. For such lateral flow to initiate there needs to be a thermal incubation period during which the deeper crust undergoes radiogenic heating; the time required depending on the magnitude of crustal heat production and hence the ages, compositions and volumes of incoming crust (Clark et al., 2011; Jamieson and Beaumont, 2013).

In the seminal Himalayan LHO model (HT1: Beaumont et al., 2004) early-accreted mid-crustal material is

translated laterally into the orogen during the thermal incubation period but once it attains temperatures of >700 °C it is incorporated into a melt-bearing low-viscosity zone, or channel. This low-viscosity channel may then flow outwards and be juxtaposed against later-arriving cooler material at the leading edge of the orogen, or conceivably flow longitudinally towards more distal orogen flanks. The longer orogenic residence times enabled by this lateral transport leads to higher temperatures in the deep crust if that crust has moderate to elevated heat production. In the HT1 model, for example, much of the crust underlying the orogenic plateau at depths greater than 40 km records $T > 800$ °C some 32 Myr after initial collision (e.g., Beaumont et al., 2004). This is illustrated in Figure 5 by the ‘channel flow model HT1’ P - T - t path modelled for an initially mid-crustal rock package incorporated into the low-viscosity channel and then expelled outwards to the edge of the orogen (Jamieson et al., 2004). This particular rock package is selected here to indicate the temperatures attained in the HT1 model by rocks that undergo near isothermal decompression (ITD) through the pressure interval 12–6 kbar by 54 Myr after collision. This pressure interval is typical of the ITD paths of many granulites and G-UHT terrains (Harley, 1998a, 1998b, 2008; Fig. 4).

Depending on a range of factors, including topographic gradients driving lateral extrusion, upper crustal extension and convergence rates, the high temperature rocks in the leading edges of the low-viscosity channel may be extruded laterally or vertically. P - T paths in these cases are clockwise, involving significant ITD from the kyanite stability field (12 kbar) into the sillimanite stability field (10–7 kbar) at 600–800 °C, followed by cooling at <6 kbar. Those rocks that remain trapped deep within a large hot orogen core at the end of collision, whether in a low-viscosity channel or beneath it, will eventually undergo near-isobaric cooling (IBC) unless additional processes intervene to facilitate their exhumation. The total time that these granulites spend near their peak T conditions (>800 °C) will be of the order of 40 Myr. Post-collisional gravitational spreading may remove the upper part of the thickened crust, the orogenic lid, or cause this lid to break apart, enabling the upward extrusion of deeper-level rocks which will undergo high- T decompression and potentially form mantled migmatitic core complexes (Vanderhaeghe et al., 1999; Rey and Houseman, 2006; Vanderhaeghe, 2009). Alternatively, the structural architecture of the orogenic lid and ductile deeper crust may be reorganised in response to changes in tectonic boundary conditions, such as arrival of a strong crustal indenter or switch from orthogonal to oblique convergence (e.g., Schulmann et al., 2008).

Slow Convergence, incipient channel flow and heterogeneous stacking:
GO Model (Jamieson et al., 2010; Jamieson and Beaumont, 2011)

Figure 3. Large Hot Orogen crustal sections as modelled for the Grenville Province of Ontario (Jamieson et al., 2010; Jamieson and Beaumont, 2011, 2013; model GO-ST87). Upper-, middle- and lower-crust rock packages marginal to the Superior Province are shaded in increasing grey intensities, and Superior Province cold and strong lower crust in the darkest grey shade. The cross-sections are snapshots at two time slices. The 700 °C isotherm is shown for both the syn-convergence and post-convergence cases. Three *P-T* paths are shown for rock packages that have experienced the syn-convergence thickening followed by post-convergence lateral spreading. The rock packages are from the late-thrusted frontal zone of the orogen (path 1), within a stalled nappe (path 2), and in a deep-crustal nappe thrust forward over a deep rigid indenter (path 3). Simplified and modified from Jamieson and Beaumont (2011). Filled circles show the *P-T* conditions for the selected rock package at times, in Myr, after the onset of collision in the model. Color version is available online from <http://doi.org/10.2465/jmps.160128>.

The LHO model has been applied to the Grenville Province of eastern Canada (Beaumont et al., 2006; Jamieson et al., 2010; Jamieson and Beaumont, 2011), with important modifications from the Himalayan models introduced to take the imaged structure, long-lived high-*T* history, and record of two distinct tectonic episodes in

the Grenville into account (Rivers, 2009; Jamieson and Beaumont, 2011). A convergence rate of 2 cm/year in the resultant Grenville Orogen ('GO') suite of models leads to slower orogen growth and longer thermal insulation / maturation periods. Temperatures of >700 °C are attained in the deeper crust about 40 Myr after the onset

Figure 4. (a) P - T diagram showing P - T paths modelled for a mid-crustal rock incorporated deep within a ductile channel in the HT1 model of Beaumont et al. (2004) and Jamieson et al. (2004) and that modelled for the rock package (3) (an orogen core nappe thrust over a rigid indenter) in the Grenville Orogen model GO-ST87 shown in Figure 3 (Jamieson et al., 2010; Jamieson and Beaumont, 2011). In the HT1 case the rock package crosses into the sillimanite field and achieves granulite facies P - T conditions, Δt_{800} , for 16 Myr, but fail to attain UHT. The GO model rock package attains UHT conditions, and undergoes UHT-ITD for 20–25 Myr ($\Delta t_{900} = 20$ –25 Myr). Filled circles on the P - T paths represent 10 Ma intervals after initiation of the model. Mineral equilibria positions and facies or sub-facies fields are from Brown (2006) and Kelsey and Hand (2015). (b) P - T diagram showing the modelled GO P - T path of (a) in comparison to selected UHT granulites with clockwise UHT-ITD paths. The selected P - T paths are: Mat – Mather Paragneisses, Rauer Islands, east Antarctica (590–530 Ma) (Harley, 1998b, 2004; Kelsey et al., 2005; Harley, 2008); Ouz – In Ouzal, Algeria (Ouzegane and Boumaza, 1996; Adjerid et al., 2013); Pal – Palni Hills, southern India (545 Ma) (Raith et al., 1997; Brandt et al., 2011). (c) P - T diagram showing the clockwise UHT-ITD path of the Mather Paragneisses (Mat) in comparison to the UHT-IBC P - T path of the Archaean Napier Complex of Antarctica (Ellis et al., 1980; Harley, 1985, 1989, 1998a). The brown-coloured field is that of the UHT mineral assemblage orthopyroxene + sillimanite + K-feldspar + quartz + melt, and the blue field that of the key UHT assemblage sapphirine + quartz, based on Kelsey et al. (2004). Color version is available online from <http://doi.org/10.2465/jmps.160128>.

of collision (Fig. 3). Continuous channel flow is inhibited by low rates of melt production resulting from the infertility of the incoming old and dry crust, but radiogenic heating still leads to eventual thermal weakening. An orogenic plateau forms, characterised by a brittle lid underlain by deeper units of varying strength that are translated as nappes by ductile thrusting. The arrival of strong crustal indentors may lead to upwards ramping, and lateral expulsion of weaker units already resident within the orogen. The lower crustal rocks forced upwards and outwards over the indentors would experience some decompression with cooling if near the flank of the orogen. By the end of collision in the ‘GO’ models the 700 °C isotherm occurs at 15–20 km below the orogenic plateau surface (Fig. 3; Jamieson and Beaumont, 2011). Hence, much of the orogen core beneath 30 km would be within the granulite facies and would remain so for a considerable period of time post-convergence. Continued radioactive self heating for some 25–30 Myr may weaken these granulites enough for them to flow outwards over the adjacent craton, resulting in post-collisional back-thrusting (Fig. 3). Slow cooling with decompression may ensue with post-collisional extension up to 50 Myr after the end of collision itself (Jamieson and Beaumont, 2011).

The GO model provides some very useful insights into the attainment of UHT conditions and possible mechanisms for the eventual exhumation of the long-lived G–UHT deep crust of orogen interiors. Figure 4a presents the P - T - t path for rock package 3 as shown in Figure 3, from a model unit trapped in the deep crust until approached and underthrust by the indenter. This activates decompression, and the P - T path traverses from the kyanite field into the sillimanite field at >900 °C, to traverse through the UHT P - T field at 980–900 °C across the pressure interval 14 kbar to 7 kbar. This model P - T excursion involving significant ITD takes 25–30 Myr; the total time spent at $T > 900$ °C (Δt_{900}) is in the range 30–50 Myr.

The P - T path derived from the Grenville model is compared in Fig. 4b with a selection of P - T paths synthesised from three UHT areas characterised by ‘clockwise’ overall P - T evolutions terminated by extensive near-isothermal decompression (ITD) or decompression with cooling. These UHT-ITD granulites (Mather Paragneisses, Antarctica (Mat): Harley, 1998b, 2008; Palni, India (Pal): Raith et al., 1997; Brandt et al., 2011; In Ouzal, Algeria (Ouz): Ouzegane and Boumaza, 1996; Adjerid et al., 2013) experienced considerable decompression at $T > 900$ °C, involving initial steep ITD ($dP/$

$dT = 40\text{--}80$ bars/ $^{\circ}\text{C}$) followed by cooling ($dP/dT = 4\text{--}8$ bars/ $^{\circ}\text{C}$). These paths traverse forward and back across the G-UHT metamorphic domain at P - T conditions comparable to those of the GO model path. The timescales of UHT-ITD in these cases are also comparable to those obtained in the chosen GO model lower-crustal package (Jamieson and Beaumont 2011, 2013), confirming the viability of LHO settings for the generation of UHT granulites with appropriate P - T - t records.

Weak lithospheres, ‘ultrahot orogens’ and G-UHT metamorphism

‘Ultrahot orogens’ (UHO) have been proposed to develop on hot, weak lithospheres (Chardon et al., 2009; Gapais et al., 2009). Higher T_p , Moho temperatures and heat productions in the crust-lithosphere system in the Archaean are considered to be favourable to the development of weak lithospheres (Marshak, 1999; Rey and Houseman, 2006) and hence facilitate the formation of UHO (Chardon et al., 2009). As a consequence, UHO are thought to be characterised by rapid accretionary crustal growth, juvenile magmatism and high geothermal gradients (Gapais et al., 2009). They are visualised as having plateau elevations of only 1.5–2 km as crustal accretion is accommodated by combined thickening and lateral spreading *during* convergence. Limited thickening involving sagduction, diapirism and gravitational overturn or redistribution are expected (Chardon et al., 2009). In the lower crust composite fabrics, sharing a sub-horizontal stretching lineation, are proposed to characterise lenses between steep transpressive high-strain zones, formed contemporaneously with partial melting.

In the UHO model proposed by Chardon et al. (2009) upper and lower crustal deformation regimes are connected across an attachment zone in the middle crust (12–20 km depths: Fig. 5). The attachment zone partitions strain between the upper and lower crustal domains but is not permanent. For very weak and buoyant deep crust the structures above the attachment zone may not pass through it, in which case the deep crust will be dominated by sub-horizontal planar and linear fabrics. If the deep crust is not quite so weak it is argued that some structures may transiently pass through the attachment and so partition lower crust deformation into cells bounded by zones of steep orogen-parallel flow, in which internal vertical mass transfer can occur independently of that in other cells.

The UHO are proposed to be typified by large areas of low- P / high- T metamorphism, and potentially UHT, which may be long-lived or involve groups of shorter-lived events within a long-lived overall evolution (e.g.,

Figure 5. Conceptual block diagram of structural patterns and kinematics in an accretionary ultra-hot orogen (UHO) as envisaged by Chardon et al. (2009). In this model deformation in the upper crust is linked to that in the lower crust via the sub-vertical high-strain zones, though this linkage may not be present in cases with even weaker lower crust and mantle. Infracrustal sagduction may be localised along shear zones or occur more broadly as upper crustal sagging into weak lower crust. Lower crustal regions that are not penetrated by sub-vertical shears are dominated by sub-horizontal foliations and mineral elongation lineations. Weak, melt-rich mantle is located immediately beneath the weak lower crust. There is negligible strength contrast between the lower crust and melt-bearing mantle for Moho temperatures in excess of 900 $^{\circ}\text{C}$. Figure modified from Chardon et al. (2009). Color version is available online from <http://doi.org/10.2465/jmps.160128>.

80–100 Myr). Metamorphic P - T paths in UHO will vary depending on the evolution of the deformation cell, for example whether the rock package stays in the deep crust or extruded into the attachment zone and vertically upwards. If the rocks stay deep they will eventually undergo IBC. If they enter the attachment then IBC and decompression-cooling will occur (depending on extrusion, or incorporation into shear zone walls). Looped P - T paths characterised by small changes in P through at most a few kbars and by excursions to high- T and UHT would result in this case. Gapais et al. (2009) argue that rocks staying within their crustal panel or cell will track out post-peak P - T paths sub-parallel to geotherms, or decompression-cooling paths.

This UHO concept has not, as yet, been directly investigated using coupled thermo-mechanical numerical modelling. UHO bear some similarities with models for Archaean hot orogens that invoke and model a major role for gravitational instability and overturn (Gerya et al., 2001, 2002), but in those models there is no mid-crustal attachment zone that acts as a limit on the amounts and extents of sagduction and vertical extrusion. Recent modelling of the craton-bound (intracratonic) Mesoproterozoic Musgrave orogen in central Australia (Tucker et al., 2015) by Gorczyk et al. (2015) may provide insights into the thermal evolution of a broad accretionary UHO. In that model, the thermal-mechanical evolution of a broad (1200 km wide) belt of hot, thinned lithosphere with an

initial narrower zone of further weakness is examined. As the thinned lithosphere is pinned between two stronger cratons with much thicker lithosphere it evolves through cycles of lithosphere and deep-crustal delamination, with re-lamination or lithosphere accretion impeded by the lateral gradients towards the strong cratons. This enables UHT to occur below about 30–35 km for considerable durations, up to 70 Myr in one model case. However, in these model situations the crust is only 32–37 km thick over that time interval, so UHT characterized by clockwise ITD P – T paths only affects a few kilometres thickness of crust at most. Whilst the models developed by Gorczyk et al. (2015) are insightful, as they start with a wide, hot crust–lithosphere zone already established they do not address the more fundamental question of how a UHO is initiated and how long it may take to develop G–UHT conditions at 6–12 kbar. In the light of this, investigation of UHO settings for the generation of UHT granulites remains a fertile area for further work, especially integrating petrological P – T – t data with modelling that considers both the orogen–normal and orogen–parallel components of deformation that are highlighted by Chardon et al. (2011).

CONSTRAINING TEMPERATURE AND TIME IN UHT METAMORPHISM

The preceding section on models for G–UHT, and especially UHT, metamorphism demonstrates that knowledge of the duration of UHT is crucial. In general, however, the high temperatures attained in G–UHT metamorphism render most lower- T chronometers of limited use in constraining the ages and durations of peak and near-peak metamorphism and the post-peak P – T evolutions of many terrains. Exceptions to this would include rapidly-evolved ‘fast’ G–UHT terrains in which the lower- T chronometers (e.g., Ar–Ar mica chronology) yield ages close to those obtained from accessory minerals formed at or near peak conditions. Even in those cases, however, the definition of the age of the metamorphic peak may not be straightforward because of the complex behaviour of the key high- T accessory minerals, in particular zircon and monazite, and their propensity to form at different times and respond to a variety of reaction-, melt- and fluid-related processes in G–UHT rocks (e.g., Roberts and Finger, 1997; Rubatto et al., 2001; Harley et al., 2007; Kelsey et al., 2008; Taylor et al., 2014).

Zircon and monazite have very low diffusivities for Pb and so, in the absence of later chemical reaction or deformation, can preserve G–UHT U–Pb ages. Zircon is particularly important as it may also provide information on the temperatures of its formation through Ti in zircon

thermometry (Watson et al., 2006; Ferry and Watson, 2007) and the context of its formation or modification in relation to major mineral phases such as garnet through REE partitioning data (Harley et al., 2001; Rubatto, 2002; Rubatto and Hermann, 2007; Taylor et al., 2015). In order to use such information and interpret zircon age data it is critical to understand the processes that govern zircon behaviour in HT/UHT environments, and influence what events zircon may record in its preserved age. The range of processes, summarised by Harley et al. (2007) may include, but are not limited to, ostwald ripening during prograde melting, crystallisation from partial melts locally and transiently saturated in Zr, coupled dissolution–precipitation in the presence of both melts and fluids, often driven by mineral and mineral–melt–fluid reactions (e.g., garnet corona formation; fluid expulsion from late-stage melts), and zircon precipitation that is decoupled from in-situ dissolution but instead promoted by the release of Zr from other phases (e.g., Zr exsolution from high- T rutile). To add to this potential complexity, as elegantly summarised by Kelsey and Hand (2015), the ages retained by zircon may be disturbed or modified (updated) by a variety of post-crystallisation effects including crystal–plastic deformation, metamictisation and Pb–redistribution, fluid-mediated recrystallisation and even diffusion itself if $T > 1000$ °C for prolonged periods (e.g., Ashwal et al., 1999). Likewise, monazite may be retentive of peak or near-peak ages but may also be strongly affected by late-stage dissolution and modification, as manifested in the formation of Th-rich rinds and rims on monazites in many UHT terrains (e.g., Watt and Harley, 1993; Braun et al., 1998; Rubatto et al., 2001; Harley and Nandakumar, 2014; Taylor et al., 2014).

Harley et al. (2007) presented a process-oriented framework for the formation and modification of zircon in G–UHT metamorphism that emphasised the role of high- T but nevertheless post-peak crystallisation of zircon from melts with different H₂O contents, along with post-peak reactions involving the breakdown of Zr-bearing major and minor minerals such as garnet and rutile. This analysis also suggested that peak- T growth of zircon is negligible in G–UHT melt-bearing terranes. Kelsey et al. (2008) and Kelsey and Powell (2011) developed a complementary thermodynamic phase equilibrium approach to evaluate the crystallisation of zircon in granulites and concluded that zircon would undergo dissolution on prograde heating into G–UHT conditions, and only form during melt crystallisation on the post-peak cooling part of the G–UHT P – T path. Clearly, in this scenario if substantial decompression (ITD) occurred prior to cooling the zircon might only record a rather late stage of the P – T history of the terrain (e.g., Roberts and Finger,

1997) that could post-date peak UHT by a considerable time period.

Both the Harley et al. (2007) T - t framework and Kelsey et al. (2008) modelling approach fail to address significant issues that bear on the behaviour and crystallisation of zircon under UHT conditions. Firstly, the equilibrium thermodynamic modelling approach only provides an end member for behaviour as the dissolution rates of zircon and effects of mineral shielding and other disequilibrium factors are not able to be considered in the equilibrium modelling. Secondly, and more importantly, neither study takes the dynamic nature of melt production, segregation, mobilisation, transfer, stagnation and final entrapment into account in the assessment of the timing of zircon growth, or recognises that these physical processes, which will involve or stimulate local chemical interactions and transient gradients in chemical potential, may occur multiple times in different sites in the one G-UHT region or even locality.

The problem with the approaches noted above is that temperature is taken as the principal or only control on zircon crystallisation in the melt-bearing interval. Whilst it is a first-order control, it is plausible that in the transiently open systems of the deforming melt-bearing deep crust chemical potential gradients, particularly in H_2O , may drive melt crystallisation even at fixed temperature (White and Powell, 2010). When melts which have segregated from one site move elsewhere in the migmatite and stagnate or pond at that new site they may undergo wall-rock interaction that promotes fractional crystallisation (Harley and Nandakumar, 2014), or lose H_2O to their new host rock if that was refractory and already melt-depleted. This dynamic scenario means that anatectic zircon could in principle form in or from melt at any time in the G-UHT evolution as the melt-bearing chemical system changes with both host environment and T . Hence, one would expect multiple zircon ages or clusters of ages within the overall G-UHT evolution at $T > 800$ °C.

The evidence for the recording by zircon of multiple melt-related ‘events’ within longer G-UHT evolutions can be seen in migmatitic granulites from many terrains (Ashwal et al., 1999; Harley et al., 2001; Korhonen et al., 2013, 2014; Clark et al., 2015; Johnson et al., 2015). Two examples of this behaviour are illustrated in Figure 6. A compilation of zircon, and monazite, ages for G-UHT metamorphism in the Trivandrum Block of southern India (Whitehouse et al., 2014; Harley and Nandakumar, 2014; Taylor et al., 2014; Kröner et al., 2015) shows not only that the metamorphism was long-lived, or ‘slow’, but also that the accessory minerals grew over several time periods in the overall 580–515 Ma age range. Anatectic zircons are reported to have formed at 575, 560, 545, 535, and

Figure 6. Age diagram for zircon and monazite growth or recrystallization events recorded in the Trivandrum Block of Southern India (Whitehouse et al., 2014; Harley and Nandakumar, 2014; Taylor et al., 2014; Kröner et al., 2015), and the Napier Complex of East Antarctica (Harley et al., 2001; Carson et al., 2002; Hokada et al., 2004; Kelly and Harley, 2005; Horie et al., 2012). Color version is available online from <http://doi.org/10.2465/jmps.160128>.

even 520 Ma, with some localities preserving more than one age. The simplest interpretation of this data is that zircon crystallisation from melt depended on the local chemical environment rather than fluctuations in temperature. Similarly, zircon U-Pb age data compiled for anatectic leucosomes and migmatitic gneisses from the Napier Complex (Fig. 6: Carson et al., 2002; Hokada et al., 2004; Hokada and Harley, 2004; Kelly and Harley, 2005; Horie et al., 2012) shows at least four crystallisation ‘events’ from perhaps 2585 Ma through to 2510 Ma and even 2490 Ma. Again, rather than invoking T fluctuations it is conceivable that local melt / wall rock interactions governed zircon crystallisation under UHT conditions.

Whatever the processes responsible for zircon crystallisation and the U-Pb age recorded, it is critical, if possible, to be able to assign a temperature to the age in order to evaluate the T - t history. The development of thermometry utilising the Ti content of zircon (Watson et al., 2006; Ferry and Watson, 2007) has led to a boom in the use of zircon to characterise the ages and temperatures of granulite metamorphism (e.g., Baldwin et al., 2007; Baldwin and Brown, 2008). However, the Ti-in-zircon data have in several instances been applied in the absence of appropriate mineral assemblage constraints, that is zircon in equilibrium with rutile and quartz, so that only minimum temperatures can be inferred (Harley and Nandakumar, 2014).

Kelsey and Hand (2015) have summarised some of the issues surrounding the application of zircon Ti thermometry, noting that despite its very low diffusivities for Ti (Cherniak and Watson, 2007) zircon often yield tem-

Figure 7. (a) Zr contents in rutiles analysed by high-beam current (200 nA) EPMA for 4 samples of the UHT Mather Paragneiss (samples reported in Harley, 1998b) and 5 samples from 3 localities in the Napier Complex. (b) Ferry and Watson (2007) Zr-in-rutile temperatures calculated, at 10 kbar, from the Mather and Napier datasets. The Mather rutiles approach but do not overlap with the UHT temperatures inferred from silicate mineral assemblages (980–1030 °C: Harley, 1998b, Kelsey et al., 2004). The Napier rutiles are all severely reset to temperatures within the upper amphibolite to lowermost granulite facies, well below UHT conditions required by the stability of sapphirine + quartz and garnet + orthopyroxene + sillimanite + mesoperthite + quartz in the selected samples (Fig. 4c). Color version is available online from <http://doi.org/10.2465/jmps.160128>.

temperatures less than those corresponding to UHT. This observation was also used as an argument for the late-stage crystallisation of zircon, alluded to above. Kelsey and Hand (2015) also compared the Ti-in-zircon results with Zr-in-rutile thermometry (Tomkins et al., 2007; Ferry and Watson, 2007) for several UHT terrains, concluding that rutile could retain its UHT Zr contents because of chemical isolation from zircon, and hence that Zr-in-rutile is a more robust thermometer in application to UHT metamorphism, even if post-peak cooling is slow.

The retentivity of UHT temperatures by rutile and zircon are examined here for two case studies, presented in Figures 7 and 8, underpinned by EMPA Zr and additional trace element data for rutile and SIMS Ti data for zircon in Supplementary Tables S1, S2, and S3 (Tables S1, S2, and S3 are available online from <http://doi.org/10.2465/jmps.160128>). Figure 7 shows the rutile Zr content and resultant 10 kbar temperature data for rutiles analysed from 4 samples from the Neoproterozoic UHT Mather Paragneiss (980–1030 °C, 10–12 kbar) in the Rauer Islands of East Antarctica (Harley, 1998b, 2004, Kelsey et al., 2004, 2005; Harley, 2008) and 5 UHT gneisses from late Archaean Napier Complex (990–1050 °C, 8–11 kbar: Ellis et al., 1980; Ellis, 1980; Harley, 1985, 1987; Hokada

et al., 1999; Harley and Motoyoshi, 2000; Hokada, 2001; Kawasaki et al., 2002; Hollis and Harley, 2002; Shimizu et al., 2013). The Mather Paragneisses are quartz-bearing garnet + sillimanite and orthopyroxene + sillimanite gneisses containing rutile within the main sillimanite + quartz foliation and included in garnet porphyroblasts. The Napier Complex samples include two sapphirine + orthopyroxene + quartz gneisses, a sapphirine + garnet + orthopyroxene granulites, and two garnet + sillimanite + orthopyroxene + quartz gneisses. Rutile is present both within garnet porphyroblasts and matrices of these gneisses.

The thermometry results for individual samples are summarised in Table 1. The Rauer rutiles preserve high Zr contents (233 analyses), but these vary considerably from 1500–3300 ppm and result in temperatures that range from 890–790 °C when the rutiles included in garnet are excluded. Two peaks in the histogram are evident, but neither correspond to UHT (peak 1: 860 °C; peak 2: 820 °C). Strikingly, the Napier rutiles (316 EMPA analyses, 26 SIMS analyses) only contain 1100–400 ppm Zr and return temperatures of 740–660 °C with a broad peak at ~ 710 °C. In neither the Mather or Napier Complex do the rutile preserve the UHT peak temperature, let alone

Figure 8. Comparison of rutile Zr thermometry for the Napier Complex (Fig. 7b) with Ti-in-zircon thermometry (Ferry and Watson, 2007) applied to zircons in two of the same thin section samples. Ti contents in zircon were measured in-situ in thin section on the Edinburgh ims4f ion microprobe (SIMS). Sample 49643 was also analysed by SIMS for Zr in the rutiles adjacent to the analysed zircons. The SIMS rutile results for 49643 are consistent with the EMPA results, and confirm that rutile records temperatures some 140–200 °C lower than those retained in adjacent UHT zircons / zircon rims illustrated in Figure 9. Color version is available online from <http://doi.org/10.2465/jmps.160128>.

UHT conditions. Rutiles in the Mather Paragneiss, which preserves a ~ 20–30 Myr UHT-ITD *P-T* history (Harley, 1998b; Hokada et al., 2016), record temperatures following the end of the UHT-ITD and subsequent decompression-cooling path (Fig. 4c). In the slowly-cooled Napier Complex the rutiles are wholly reset to record amphibolite to lowermost granulite facies temperatures and have no retention of the UHT conditions required by the observed mineral assemblages (Fig. 4c).

The Napier Complex rutile temperature data is compared with Ti in zircon temperatures from two of the same samples in Figure 8. 22 SIMS analyses of convolute and planar-sector zircons from samples 49643 (McIntyre Island) and 49796 (Tonagh Island) yield Ti contents of 36 ± 6 and 43 ± 10 ppm respectively, which translate to 10 kbar temperatures of 886 ± 36 °C and 901 ± 30 °C. In contrast to the rutiles, all of the zircon analyses lie in the G-UHT temperature field and many straddle the UHT lower-*T* boundary. Hence, for at least the long-lived and slowly-cooled Napier Complex, the classic UHT terrain, the zircon Ti thermometry is far more robust and informative as to the UHT conditions than the rutile Zr thermometry, a conclusion that is the opposite of that presented by Kelsey and Hand (2015) for these systems but which in the current analysis is based on directly comparable zircon and rutile data from the same thin sections.

In the light of the issues noted above concerning modelling of UHT and providing constraints on the age and duration of UHT and indeed granulite conditions, the subsequent sections of this review describe and highlight examples of G-UHT metamorphic belts or regions that can be demonstrated to be truly short-lived (‘fast’; $\Delta t_{900} < 5$ Myr), rapidly-terminated (short duration with $\Delta t_{900} < 10$ Myr at the end of a longer orogenic event) and long-lived (‘slow’: $\Delta t_{900} > 20$ Myr).

FAST GRANULITES AND SHORT LIVED UHT

Very young and fast: Kobipoto Mountains, Seram, Indonesia

Pownall et al. (2014) have described very young, 16 Ma, garnet + cordierite + spinel + quartz ($\pm$ magnetite, $\pm$ ilmenite) migmatitic granulites from the Kobipoto Mountains area of the island of Seram. Phase equilibrium modelling has been used to infer UHT ($P = 7.5$ – 8.0 kbar, $T = 950$ – 975 °C) conditions for these complex and oxidised granulites, which occur sporadically within a region strongly dissected by arrays of strike-slip faults. The faulting is considered to be associated with subduction-rollback driven oblique extension. In the light of this, Pownall et al. (2014) have proposed an extensional setting for G-UHT metamorphism in which the deeper crust, originally no thicker than ~ 50 km, is extended to ~ 25–30 km thick and in which UHT is promoted by the presence of mantle asthenosphere at only 3–35 km depths. That is, the UHT occurs in crust that is only some 5–10 km above 1300 °C mantle, trapped above a lithosphere detachment fault. Slab rollback is interpreted to have occurred from before 16 Ma, generating the extension in the Banda Arc and at Seram in particular. The ‘hot crust’ core complex setting proposed for Kobipoto is clearly a rapid or ‘fast’ G-UHT setting, and this is substantiated by the $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages of the granulites, which are within error of the peak / near-peak ages obtained on migmatite zircon (zircons 16 ± 0.5 Ma; Ar-Ar biotite 16.3 Ma at 290 °C, 14.8 Ma at 230 °C). The cooling rate for the complex is at least 40 °C/Myr, and the overall Δt_{800} certainly less than 10 Myr and probably only a few million years: the granulites are now exposed to the Earth’s surface.

With respect to the UHT conditions of >950 °C deduced for the Kobipoto granulites, it must be noted that this result strongly depends on the treatment of spinel (and hence spinel + quartz) in the calculated phase diagrams, and is also subject to some uncertainty given the disequilibrium character of the spinel textural settings. Nevertheless, even if the peak *T* was less than UHT, the very short Δt_{800} duration still stands: this is a fast

Table 1. Napier Complex and Mather Paragneiss rutile and zircon temperatures calculated at 10 kbar using the thermometers of Ferry and Watson (2007)

a. Rutiles [Zr in rutile thermometer Ferry and Watson (2007)]										
Sample	Locality	Assemblage	n	Si	V	Cr	Fe	Nb	Zr	T (°C)
Napier Complex, Enderby Land, Antarctica										
49643-a	McIntyre Island	Grt+Opx+Sil+Qtz	92	57	2422	1634	1435	3059	691	720
				49	1266	75	483	487	114	15
49643 (SIMS)	McIntyre Island	Grt+Opx+Sil+Qtz	27	80	5255	3550	1970	5906	824	736
				190	313	272	533	1441	162	19
49638	McIntyre Island	Grt+Opx+Sil+Spr+Qtz	70	87	3674	2390	3762	18157	664	716
				107	1264	522	1133	4065	128	20
49377	Mt Riiser-Larsen	Spr+Opx+Qtz	34	159	8706	2678	3615	16659	578	703
				97	888	724	987	4717	151	23
49356	Mt Hardy	Spr+Opx+Qtz	77	146	7822	2384	2982	14370	558	701
				55	304	186	614	1525	86	14
49796	Tonagh Island	Grt+Opx+Sil+Spr	43	166	9100	2938	3860	18446	590	706
				80	877	788	1085	4935	118	20
Mather UHT Paragneiss, Rauer Islands, Antarctica										
SH-91-90	Mather Peninsula	Grt+Sil+Qtz±Crd	35	42	5790	2539	884	1870	2607	858
				30	1226	314	314	458	308	14
SH-91-95	Mather Peninsula	Opx+Sil+Qtz	33	29	2854	2520	546	1908	2303	843
				18	1824	1204	257	269	366	20
SH-88-195	Mather Peninsula	Grt+Sil+Qtz	74	89	3578	2778	271	2033	2638	859
				116	2119	399	183	120	417	19
SH-88-197	Mather Peninsula	Grt (inclusions)	9	69	n.d	831	4042	3900	1115	767
				49		68	566	254	42	4
		Sil (inclusions)	9	35	n.d	1411	2963	3567	2450	852
				15		56	1003	135	18	1
		Grt+Sil+Qtz±Crd	73	110	1503	1894	755	1927	1991	826
				105	607	380	804	1996	401	21
b. Zircons [Ti in zircon thermometer Ferry and Watson (2007)]										
Sample	Locality	Assemblage	n	P	Y	Ti	T (°C)			
Napier Complex, Enderby Land, Antarctica										
49643 (SIMS)	McIntyre Island	Grt+Opx+Sil+Qtz	14	3044	1440	38.5				
				3512	575	12.6				
49796 (SIMS)	Tonagh Island	Grt+Opx+Sil+Spr	7	655	404	42.8	901			

granulite that was rapidly exhumed and cooled, exposed within a few million years of G-UHT metamorphism.

Significantly older examples of comparably short-lived G-UHT metamorphism associated with the rapid transfer of heat promoted by extension and magmatic underplating may include the Palaeozoic (272 ± 4 Ma) Gruf Complex, within the Alps (Galli et al., 2011), which underwent UHT for less than 5 Myr, and, in the light of the most recent geochronology, the Palaeoproterozoic (1900 ± 6 Ma) South Harris Complex of Scotland (Baba et al., 2012).

Young, arc-related G-UHT: Hidaka Complex, Hokkaido, Japan

The Cenozoic granulites exposed in the Hidaka HT/UHT

metamorphic complex of Hokkaido, Japan, have been cited as the classic example of granulite formation related to arc and back-arc settings (Kemp et al., 2007). These ~ 19 Ma granulites have been proposed to have formed through the metamorphism of 41–37 Ma arc-related gabbros and tonalities and 80–50 Ma accretionary complex sediments. In this case metamorphism at P - T conditions of 8 kbar and ~ 900 °C is attributed to heating through back-arc extension, magmatic under-accretion and associated lithospheric thinning (Kemp et al., 2007). This thermal pulse was short-lived, being terminated by subsequent arc-arc collision which captured the hot crust and promoted the rapid exhumation of the granulites along an ITD P - T path to 3.5 kbar at $T > 800$ °C, followed by cooling through 500 °C by 15 Ma. The clockwise P - T path produced in this extension-accretion setting is sim-

ilar in form to those deduced for many granulites from orogenic belts, but the HT residence time of the Hidaka granulites, Δt_{800} , is very short (2–6 Ma) in comparison to most examples cited by Kelsey and Hand (2015), Kelsey (2008) and Harley (2008).

High-*T* termination of orogeny: Betic Cordillera — Alboran Sea, Spain

The Betic Cordillera of southern Spain shows the rapid attainment of granulite, but not UHT, *P-T* conditions (5–7 kbar and 850 °C) followed by evolution on steep ITD paths over a short time interval of perhaps 15 Myr (Platt et al., 1998). Peak granulite metamorphism and ITD occurred at 22–18 Ma, some 25 Myr after crustal thickening as recorded by HP metamorphism in high-level nappes (Platt et al., 1998; Whitehouse and Platt, 2003). The thermal excursion and contemporaneous decompression is considered to have been initiated by the rapid convective removal of the lithospheric mantle root at ~ 70 km beneath the orogen from ~ 27–28 Ma. The advection of heat from the underlying mantle facilitated the rapid post-collisional formation of the granulites, with ITD promoted by the extensional spreading of the crust as it heated up in response to the increased basal heat flux (Platt et al., 1998).

More generally, such gravitational spreading is a natural progression in orogeny when collision ceases, providing a large mass of topographically elevated crust remains that is hot (>700 °C) and hence able to flow at depth (Harley, 1989; Sandiford, 1989; Sandiford and Powell, 1990; Platt and England, 1993; Rey et al., 2001; Rey and Houseman, 2006; Jamieson and Beaumont, 2013). The numerical modelling indicates that heterogeneous domal flow of the hot middle and deep crust will occur contemporaneously with extension in the upper crust (Vanderhaeghe, 2009; Jamieson and Beaumont, 2011, 2013). Variably focussed upper crustal extension is implicated in facilitating the vertical extrusion and exhumation of the domed migmatitic core complexes observed in the UAS-Canadian Cordillera Core and parts Variscan of Europe (Vanderhaeghe et al., 1999; Rey et al., 2009). For example, in the Thor Odin Dome of the Shuswap Metamorphic Core Complex (Vanderhaeghe et al., 1999; Rey et al., 2009) migmatites were generated from 60 Ma at initial *P-T* conditions of 10.5 bar and 760–800 °C, evolving to lower-*P* on an ITD path to about 4 bar over a time period of some 8 Ma. Isobaric cooling to Ar–Ar closure temperatures of 300 °C was completed by 48 Ma, giving a 12 Myr timescale for the granulite metamorphic evolution related to post-collisional extensional collapse, similar to the duration of G-ITD in the Betics. Again, it is noteworthy that these hot regimes at

the end of orogeny did not yield UHT conditions or *P-T* paths in the rocks now exposed.

Rapid buoyancy-driven vertical extrusion of High-Pressure Granulites into near-UHT conditions: the Bohemian Massif

The Variscan orogen in the Bohemian Massif is typified by gneisses of upper crustal derivation, metamorphosed at 16–22 kbar and 850–1000 °C at 340 Ma to produce garnet + kyanite leucogranulites (O'Brien, 2008), classic examples of High Pressure granulites (HPG) (O'Brien and Rötzler, 2003; Brown, 2006). The HPG conditions are suggested to have been produced by continental subduction terminated by delamination of the lowermost subducted crust and its lithospheric mantle (O'Brien, 2008), leading to a transition from a longer-lived colder orogenic regime to that of a hot orogen dominated by felsic crust with relatively high heat productions at deep levels in the thickened orogenic architecture. In the southern Bohemian Massif the HPG gneisses underwent post-peak ITD through near-UHT to granulite *P-T* conditions of 12–7 kbar at 880–800 °C within 5–10 Ma of the 340 Ma metamorphic peak. Schulmann et al. (2008) have proposed that this decompression and exhumation reflects vertical extrusion of the hot and buoyant HPG felsic gneisses in concert with horizontal compression that accommodated the growth of domes and subsidence of intervening higher-level rocks. In some parts of the orogen lateral flow ensued at 7–10 kbar to produce strongly recrystallised granulites transported in hot nappes over a rigid continental indentor, all prior to ~ 330 Ma and so within 10 Myr of the peak-*T* conditions. Although true UHT conditions (>900 °C at 7–12 kbar) were not attained in this example, within the relevant pressure range Δt_{800} lasted only 5 Myr: a case of rapid termination to orogeny.

Fast or slow: the Anápolis-Itaçu Complex, Brazil

UHT metamorphism in the Anápolis-Itaçu Complex of Brazil has been well-documented on the basis of extensive petrological, phase-equilibrium and accessory mineral thermometric studies (Moraes et al., 2002; Baldwin and Brown, 2008; Baldwin et al., 2007). The UHT event has been proposed to have been of short duration, terminated by UHT-ITD and cooling over a 15 Myr time interval from 649 to 634 Ma (Baldwin et al., 2007; Baldwin and Brown, 2008). As a 'fast' UHT region, the rapid evolution has been interpreted to reflect thickening of hot back-arc crust as a consequence of arc-continent collision (Baldwin and Brown, 2008). More recent dating of metamorphic zircon rims on grains in layered mafic-ultramafic

complexes that also experienced the UHT conditions (Della Guistina et al., 2011) yields a spread of concordant U-Pb zircon ages (LA-MC-ICPMS, SHRIMP and ID-TIMS) within an age interval of ~ 80 from ~ 680 Ma. Della Guistina et al. (2011) interpret the spread of concordant U-Pb ages to reflect coupled dissolution–reprecipitation promoted by the high grade metamorphism. G-UHT metamorphism is considered to have either been pulsed or rather more long-lived than reported by Baldwin and Brown (2008), being 680–630 Ma from peak to post-peak but still near-UHT, and potentially more directly associated with magmatic accretion over that prolonged time period.

SLOW GRANULITES AND LONG LIVED UHT

Whilst many UHT granulites are typified by ITD and decompression-cooling paths, a smaller but significant number may be characterised by near-isobaric cooling (IBC) paths (Harley, 1989, 1998a, 2004, Brown, 2007; Harley, 2008; Brown, 2009; Korhonen et al., 2013; Kelsey and Hand, 2015). These typically record long time-scales of cooling (e.g., Ashwal et al., 1999), but like the UHT-ITD examples noted in the section on Large Hot Orogens, the duration of UHT itself, Δt_{900} , is often elusive or poorly-constrained. Nevertheless, there are several examples where, on the basis of the collective age information obtained on zircons and monazites coupled with thermometry on the zircons and coexisting minerals, it is now possible to demonstrate long-lived UHT with Δt_{900} durations of 30–100 Myr (Kelsey and Hand, 2015). For example, long-lived orogen residency under UHT or near-UHT conditions is inferred for central Eastern Ghats of India (Korhonen et al., 2013, 2014), which record high- T metamorphic events from 1000 Ma through to 930–920 Ma and may have resided at >900 °C for some 80 Myr. In this final section of this review two further examples of long-lived UHT, most likely developed within LHO settings and promoted by the burial of high heat-production crust, are presented. The first is the Neoproterozoic–Cambrian of southernmost India and Madagascar, and the final example is Earth’s premier UHT terrain: the latest-Archaeon Napier Complex of east Antarctica.

Long duration Neoproterozoic–Cambrian UHT in Southern India and Southern Madagascar

The timescales of UHT in the regional granulite belts of southern India and southern Madagascar are of considerable interest with respect to their implications for processes involved in the formation of the supercontinent

of Gondwana from its constituent parts (Brown, 2009; Santosh et al., 2009; Santosh and Kusky, 2010; Collins et al., 2014; Clark et al., 2015), and also because in the compilation of Kelsey and Hand (2015) both regions are reported as having short duration UHT events (i.e. ‘fast’ granulites; 5–10 Myr: Kelsey and Hand, 2015, Fig. 12). This inference, however, is not consistent with an extensive range of zircon and monazite isotopic and chemical data that instead indicate that both areas are characterised by ‘slow’ granulites and long-lived UHT.

The duration of HT/UHT metamorphism in the Trivandrum Block of Southern India can be defined on the basis of a number of recent zircon U-Pb and monazite U-Pb isotopic or Th-U-Pb chemical dating studies in which the accessory minerals have been examined in their petrographic contexts. Temperatures in this belt are provided by phase equilibrium modelling of formerly melt-bearing assemblages, as the absence of rutile from all studied assemblages means that Ti in zircon thermometry can only give a lower bound on the prevailing temperature conditions. In several cases, however, the modelled P - T conditions of ~ 6 –8 kbar and 850–940 °C can be tied to zircon ages through in-situ REE microanalysis of dated zircons and garnets that provides zircon-garnet REE partitioning evidence for the petrological context of accessory mineral growth and modification. Harley and Nandakumar (2014), for example, have documented two-stage anatectic zircons featuring early skeletal-planar zoned cores that contain inclusions of micro- and nanogranite, overgrown by sector-planar zircon. A dramatic change in the HREE patterns of the zircons from strongly HREE enriched to flat HREE is correlated with the textural progression. Harley and Nandakumar (2014) interpreted this to reflect the dynamics of localised melt transfer and stagnation, with zircon crystallising during the HT-UHT metamorphism as a result of melt interaction with its host rock on entrapment. The zircon cores were produced through melt crystallisation under open-system conditions, triggered by initial melt stagnation, at ~ 545 Ma. Further precipitation of the planar zircon occurred under closed-system conditions (melt entrapment) in REE equilibrium with garnet from ~ 542 to 535 Ma, the age of weakly zoned lower-Th monazite.

The 10 Myr duration of melt migration, stagnation and residence inferred by Harley and Nandakumar (2014) represents but a single snapshot from the longer-lived HT-UHT evolution of the Trivandrum Block. The compilation (Fig. 6) of the recent zircon and monazite age information from localities within the Trivandrum Block (Braun and Bröcker, 2004; Whitehouse et al., 2014; Taylor et al., 2014; S.L. Harley and V. Nandakumar, unpublished data, 2016; Kroner et al., 2015) indicates likely

anatectic zircon growth at several times within the period 580–520 Ma (575 Ma, 567 Ma, 545 Ma, 540 Ma, and 523 Ma) and monazite crystallisation at broadly similar times (585 Ma, 570–565 Ma, 545 Ma, and 535 Ma). Chemical modification of zircon and monazite occurred from ~515 Ma, at which time the temperatures were still at least 780 °C based on Ti in zircon thermometry (Harley and Nandakumar, 2014). The consensus, therefore, is that the Trivandrum Block of Southern India experienced HT-UHT conditions for a period of 40 Myr. The duration spent under HT conditions, Δt_{800} , was possibly as long as 60 Myr. These timescales are similar to those reported from the neighbouring Nagercoil Block (Johnson et al., 2015: 570–535 Ma suprasolidus conditions) and to the duration of HT-UHT inferred by Clark et al. (2015) for the Madurai Block (>40 Myr: 575–540 Ma based on monazite and to ~520 Ma based on zircon rims). Clark et al. (2015) interpreted the HT-UHT metamorphism to be the result of conductive self-heating and thermal maturation of high heat-production thickened crust trapped within a long-lived orogenic plateau which formed following continent-microcontinent collision (East Africa-Azania) at 610 Ma. The duration of the HT-UHT metamorphism coupled with the prevalence of dominantly decompressional (ITD) *P-T* paths accompanied by cooling at crustal levels <6 kbar across all three crustal blocks is consistent with the Large Hot Orogen (LHO) setting for UHT metamorphism.

The HT-UHT granulites of southern Madagascar can be similarly viewed as the products of long-lived conductive self-heating and thermal maturation of the deeper levels of high heat-production thickened crust following final Gondwanan (580 Ma) assembly of previously accreted arc and collided Palaeoproterozoic crustal domains. Horton et al. (2016) have demonstrated that UHT conditions prevailed in the deep crust of the Anosyen domain from 580 Ma through to exhumation and crustal thinning at 525–510 Ma. Measured U-Th-K contents of the preserved gneisses lead to calculated high heat productions at 580 Ma ($>3.5 \mu\text{Wm}^{-3}$ in the UHT region) that are modelled to produce UHT conditions at 30–35 km depths after 40–60 Myr. These UHT conditions could persist for some 40 Myr provided the thickened crust resides under an orogenic plateau and hence undergoes negligible exhumation on this timescale, consistent with the general model for orogenic self-heating outlined by Clark et al. (2011).

Slow archaean UHT: The Napier Complex, east Antarctica

The UHT Napier Complex of East Antarctica fulfils many

of the critical characteristics of a long-lived or ‘slow’ granulite terrane developed and potentially left ‘trapped’ within the lower parts of a large hot to ultrahot orogen. It is late Archaean in metamorphic age, containing abundant felsic rocks whose heat productions at ~2.8–2.5 Ga would be ~2 times present-day values. It is dominated by protoliths that are older by >200 Myr than its UHT metamorphism, which can be shown to have occurred from ~2585 to 2500 Ma based principally on zircon U-Pb and trace element data (Harley et al., 2001; Grew et al., 2001; Crowe et al., 2002; Hokada et al., 2004; Suzuki et al., 2006). Whilst orthogneisses as young as 2620 Ma have been reported (Carson et al., 2002), the dominant orthogneiss precursors are older than ~2840 Ma and include components that are 3000, 3300 and 3600–3850 Ma in age.

Regional-scale UHT metamorphism in the Napier Complex has been demonstrated to have progressed at >950–1050 °C and 7–11 kbar by several early studies (1980s–1990s) that relied on conventional Fe-Mg-Al thermobarometry, experimentally derived sapphirine-quartz, osumilite and orthopyroxene-sillimanite equilibria, and additional mineral composition constraints from metamorphic pigeonite and ternary feldspar (e.g., Ellis, 1980; Ellis et al., 1980; Harley, 1985, 1987; Hokada et al., 1999). More recent studies employing phase diagram (pseudosection) analysis based on best-fit internally consistent thermodynamic datasets (Shimizu et al., 2013) improved models for ternary feldspar solvi thermometry (Hokada, 2001; Hokada and Suzuki, 2006), sapphirine-orthopyroxene-quartz equilibrium (Harley and Motoyoshi, 2000), and trace element thermometry applied to quartz and garnet (Kawasaki and Motoyoshi, 2000; Kawasaki et al., 2002) generally support and lend precision to the earlier studies, defining UHT at >1000–1050 °C and 7–8 kbar in the Tula Mountains and Amundsen Bay area and at 8–12 kbar further south in the Fyfe-Hills Khmara Bay area and in the NW Scott Mountains. Pseudosection analysis (Shimizu et al., 2013) can be interpreted as supportive of the interpretation of early near-peak to peak ITD preceding the dominant post-peak near-isobaric cooling (IBC), as first proposed by Hollis and Harley (2002) for the UHT *P-T* path in the Amundsen Bay area. The key metamorphic evidence remains that the Napier Complex underwent UHT at or above 1000 °C and that this UHT event affected a significant thickness of deep crust. UHT gneisses with peak metamorphic pressures of between 7–8 kbar and 11–12 kbar all underwent post-peak IBC, suggesting that 12–15 km thickness of crust attained >950–1000 °C.

Estimates of the duration of UHT metamorphism in the Napier Complex are principally reliant on U-Pb age data obtained from zircon. All published studies of the region have yielded monazite ages that range up to

~ 2480–2500 Ma but are usually much younger than the UHT event itself (e.g., <2400 Ma: Suzuki et al., 2006). Concordant U–Pb ages obtained from zircon inferred to have grown or been modified in the UHT metamorphic episode are commonly in the range 2530–2485 Ma (Grew et al., 2001; Carson et al., 2002; Crowe et al., 2002; Hokada et al., 2004; Kelly and Harley 2005; Grew et al., 2006; Horie et al., 2012), but ages of 2585 and 2555–2550 Ma have also been recorded from zircons inferred on textural grounds to be of UHT origin (Harley et al., 2001; Carson et al., 2002). The duration of UHT can also be bracketted by age data on the youngest pre-metamorphic intrusives and detrital zircons in metasediments and cooling ages provided by complementary isotopic systems such as garnet–rock Sm–Nd and Lu–Hf mineral isochrons. The youngest protolith age for an in-situ orthogneiss from the Napier Complex, 2626 ± 28 Ma (Carson et al., 2002), constrains UHT to be <2620 Ma. Horie et al. (2012) report ~ 2580 Ma ages for igneous zoned zircon cores in one quartzofeldspathic gneiss and two metaquartzites from Mt Cronus, and from these results infer a maximum possible age of UHT of ~ 2580 Ma. Sm–Nd and Lu–Hf mineral isochrons indicate cooling through 800–650 °C by ~ 2450–2350 Ma (Suzuki et al., 2006; Choi et al., 2006). From these data it can be inferred that HT conditions could have prevailed for as long as 130 Myr and hence that the duration of UHT, Δt_{900} , would be <130 Myr.

In order to evaluate the duration of UHT more closely it is necessary to couple the zircon age information with zircon Ti thermometry carried out on the same zircon grains, in-situ and in textures that can be linked with the UHT mineral assemblage in the host rock (Kelly and Harley, 2005; Harley et al., 2007). Such an integrated age–temperature evaluation, complemented by zircon–garnet REE distribution analysis, has been carried out on two UHT leucosomes from one island in the Napier Complex (McIntyre Island) by Harley et al. (2001). Both leucosomes contain the simple major phase mineral assemblage garnet + mesoperthite + quartz, with additional ilmenite or, in one case, rutile. Both contain mesoperthites that indicate minimum temperatures of >920–940 °C based on ternary feldspar thermometry using the calibration of Fuhrman and Lindsley (1988). Ovoid, weakly sector-planar zoned and chemically homogeneous zircons in a patch leucosome yield a spread in U–Pb ages along Concordia, from 2590 to 2480 Ma, but preserve a robust $^{206}\text{Pb}/^{238}\text{U}$ age of 2586 ± 8 Ma for the oldest cluster of analyses, an age that overlaps with the minimum depositional age for metaquartzite protoliths as inferred by Horie et al. (2012) based on zircon population analysis (Fig. 6). In-situ SIMS analysis of the zircons in the

~ 2585 Ma leucosome yields 47 ± 10 ppm Ti ($n = 40$), which translates to minimum temperatures of 914 ± 28 °C as the coexisting Ti-phase is ilmenite in this instance. Two-phase zircons occur in a second leucosome, a stringer or vein type leucosome sampled from the limb of a F2 fold only 200 metres away from the first site. The moderately CL-luminescent euhedral zircon rim overgrowths in this case yield a $^{206}\text{Pb}/^{238}\text{U}$ age of 2557 ± 5 Ma. These rims, which contain rutile inclusions, have Ti contents of 59 ± 9 ppm ($n = 33$) that translate to temperatures of 945 ± 20 °C. Collectively these two leucosomes strongly indicate that UHT conditions prevailed at or by 2585 Ma and, in the simplest interpretation, continued through to at least 2555 Ma at McIntyre Island. Hence, the *minimum* duration of UHT, Δt_{900} , was 30 Myr (Fig. 10).

Hokada and Harley (2004) carried out a similarly integrated in-situ zircon–garnet REE study of a feldspathic leucosome that, from field and textural relationships, was interpreted to have invaded and interacted with a garnet-rich quartz metaquartzites host rock. This interaction resulted in a change in zircon chemistry from HREE-enriched cores to flat-HREE outer zones and rims in close REE equilibrium with adjacent garnet rims based on the experimental distribution data of Taylor et al. (2015). Zircon U–Pb ages of 2496–2470 Ma for the inner cores indicate that melt entrapment and subsequent reaction occurred later in the UHT history of the Napier Complex than the generation and crystallisation of the McIntyre Island leucosomes. Unfortunately, zircon was not analysed for Ti to provide a direct estimate for the temperature of crystallisation, but the high Zr contents in the adjacent garnet rims (115 ppm) are compatible with UHT conditions accompanying melt interaction (Kelsey et al., 2008; Kelsey and Powell, 2011), and mesoperthites in the feldspathic domains yield minimum temperatures of ~ 950 °C (Hokada et al., 2004). Coupled with the data from McIntyre Island, this either implies a duration of UHT, Δt_{900} , of 90 Myr (Fig. 10), or distinct ages for UHT at the two localities which are separated by some 80 km geographically.

The zircon U–Pb and Ti content data from rutile-bearing pelitic granulites described earlier in this paper add to this record of long lived UHT in the Napier Complex and confirm the regional extent of this duration. Modified zircon rims and anatectic zircon overgrowths and domains in both pelites formed at 891 ± 35 °C. In particular, pelite 49643 from McIntyre Island, the same island as the leucosomes referred to in the preceding paragraphs, preserves internally complex but texturally equilibrated zircon that records Ti in zircon temperatures of ~ 870–890 °C at 2558 Ma ($^{207}\text{Pb}/^{206}\text{Pb}$ age; Figs. 9a and 9b) and <860 °C in the texturally youngest age do-

Figure 9. Cathodoluminescence [CL: (a), (c), (e), (g), and (h)] and Back-scattered electron [BSE: (b), (d), and (f)] images of selected zircon grains from Napier Complex UHT samples 49643 [McIntyre Island; (a)–(d)] and 49796 [Tonagh Island; (e)–(h)]. Two-figure numbers on all images are the Ti contents (ppm), measured via SIMS. Three digit numbers in italics are the calculated 10 kbar temperatures based on Ti in zircon (Ferry and Watson, 2007). Four digit numbers in yellow are $^{207}\text{Pb}/^{206}\text{Pb}$ ages; these have typical uncertainties of ± 5 –8 Ma for the individual analyses. The ‘snail’ zircon [images (c) and (d)] shows complex core or main phase UHT zircon with marked variation in CL and BSE response (reflecting variations in U), similar to the triplet of zircons in images (a) and (b). The strongly zoned zircon is replaced by a weakly planar zoned zircon with lower Ti content and younger ages. This phase of the zircon is adjacent to rutile that preserves only 500–600 ppm Zr up to 100 micron from the zircon. The zircon in (f) partially preserves an inherited core with modified and smeared-out former oscillatory zoning, replaced by weakly planar-zoned rims. These, and the other rims in zircons from 49796 (Tonagh Island) all record UHT to near-UHT Ti-in-zircon temperatures. Color version is available online from <http://doi.org/10.2465/jmps.160128>.

main (Figs. 9c and 9d). Whilst the temporal correspondence between preserved Ti-temperature and zircon age cannot be assured in such complexly zoned grains, it is apparent that considerable zircon modification, including dissolution and re-precipitation or recrystallisation, occurred under UHT and near-UHT conditions in the pelites from ~ 2560 to <2500 Ma (Fig. 10).

These zircon data indicate that UHT metamorphism in the Napier Complex was long lived, with Δt_{900} being of the order of 90 Myr. The published Sm–Nd (2380–2360 Ma) and Lu–Hf (2403 ± 43 Ma) mineral isochrons indicate that UHT was followed by further slow cooling over a time interval of 40–100 Ma at 2–3 $^{\circ}\text{C}/\text{Ma}$ (Kelly and Harley, 2005; Harley, 2008).

In principle, extreme metamorphic conditions, intense deformation features, and long-lived deep crustal residency of the Napier Complex are reconcilable with the UHO conceptual model of Chardon et al. (2009). However, the Napier Complex lacks the long-lived juvenile magmatism and marked acceleration in crustal growth that is regarded as characteristic of UHO and hence does not correspond to an accretionary UHO in the sense proposed by Chardon et al. (2009). Instead, given the minimal crustal growth that appears to be associated with its UHT metamorphism, the Napier Complex is interpreted here to represent the oldest, and hottest, example of a collisional ‘large hot orogen’ (LHO; Jamieson and Beaumont, 2011, 2013), in which deep crust formed from

Figure 10. Temperature–Time path or record for the Napier Complex based on the zircon age–Ti data obtained from leucosomes at McIntyre Island (606 and 639; Harley et al., 2001, submitted) and Mt Riiser-Larsen (R23; Hokada et al., 2004; Hokada and Harley, 2004) and individual concordant zircon age–Ti data from the pelite 49643, described in this study (orange diamonds). The grey curve represents a T–time that suggests the persistence of UHT for 80–90 Myr based on the leucosome data. Color version is available online from <http://doi.org/10.2465/jmps.160128>.

considerably older continental materials is left trapped for long periods under an orogenic plateau to undergo prolonged self-heating (Clark et al., 2011) and then cooling following the cessation of orogeny.

CONCLUSIONS

1. Of the models proposed to account for long-lived UHT metamorphism, the 'large hot orogen', LHO, model for collisional orogeny provides the best setting. The LHO model, which is not simply conceptual but has been tested numerically (Jamieson and Beaumont, 2013) accounts not only for the P - T paths, which can range from UHT-ITD through to decompression-cooling and UHT-IBC, but also for the evidence for the longevity of many well-studied examples of UHT granulites, with durations of UHT, Δt_{900} , in the range 30–100 Myr.
2. Shorter duration UHT and near-UHT granulites also exist, mostly with Δt_{900} less than 10 Myr. A number of these, such as Seram, are likely to have formed as a consequence of severe lithospheric thinning and crustal extension accompanied by voluminous magmatism, which could occur in arc and back-arc settings affected by subduction roll-back or, as advocated by previous workers, continental arcs undergoing extension. However, attaining long-lived UHT conditions in these settings is unlikely unless the crust has inherently high radioactive heat production in addition to the transient heat added during extension and magmatism, as concluded by Clark et al. (2011).
3. UHO setting for the generation of UHT metamorphism remain enigmatic. Whilst recent modelling by Gorczyk et al. (2015) demonstrates the possibility of maintaining UHT metamorphic conditions in the deepest levels of an intracratonic UHO for considerable timespans, mechanisms for initial development of such wide, low orogens on thinned lithosphere are not yet addressed. Any assessment of the relevance of the UHO setting for UHT metamorphism in a particular terrain requires extensive geochronology demonstrating the addition of voluminous magmas broadly contemporaneous with metamorphism.
4. Contrary to popular belief, zircon can and does form at various times at and near the peak of UHT metamorphism and is not confined to only form from melts on post-peak cooling. This arises because the segregation, transfer and stagnation or trapping of melts in G-UHT terrains leads to dynamic local melt-rock interactions that promote zircon crystalli-

sation. This contention is supported by extensive zircon age and Ti data from a number of long-lived, or 'slow' G-UHT regions.

5. Whilst rutile can retain UHT information in 'fast' G-UHT terrains, it is strongly afflicted by Zr re-equilibration and loss in 'slow' terrains, such as the Napier Complex, and yields temperature significantly lower than those preserved by zircons formed or modified in the same UHT episode.
6. The Napier Complex, the Earth's premier UHT terrain, most likely formed as trapped deep crust in the hot underbelly of a late-Archaeon LHO. UHT at ~ 1000 °C occurred throughout 12–15 km thickness of that deep crust, over an area of some 20000 km², for up to 90 million years. This was followed by prolonged near-isobaric cooling that most likely was a simple consequence of the waning effects of radiogenic heat production.

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SUPPLEMENTARY MATERIALS

Supplementary Tables S1–S3 and color version of all Figures are available online from <http://doi.org/10.2465/jmps.160128>.

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