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The contribution of metamorphic petrology to understanding lithosphere evolution and geodynamics

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ABSTRACT

In the early 1980s, evidence that crustal rocks had reached temperatures \(>1000\,^\circ\text{C}\) at normal lower crustal pressures while others had followed low thermal gradients to record pressures characteristic of mantle conditions began to appear in the literature, and the importance of melting in the tectonic evolution of orogens and metamorphic metasomatic reworking of the lithospheric mantle was realized. In parallel, new developments in instrumentation, the expansion of in situ analysis of geological materials and increases in computing power opened up new fields of investigation. The robust quantification of pressure \((P)\), temperature \((T)\) and time \((t)\) that followed these advances has provided reliable data to benchmark geodynamic models and to investigate secular change in the thermal state of the lithosphere as registered by metamorphism through time. As a result, the last 30 years have seen significant progress in our understanding of lithospheric evolution, particularly as it relates to Precambrian geodynamics.

Eoarchean–Mesoarchean crust registers uniformly high \(T/P\) metamorphism that may reflect a stagnant lid regime. In contrast, two contrasting types of metamorphism, eclogite–high-pressure granulite metamorphism, with apparent thermal gradients of \(350–750\,^\circ\text{C/GPa}\), and granulite–ultrahigh temperature metamorphism, with apparent thermal gradients of \(750–1500\,^\circ\text{C/GPa}\), appeared in the Neoarchean rock record. The emergence of paired metamorphism is interpreted to register the onset of one-sided subduction, which introduced an asymmetric thermal structure at these developing convergent plate margins characterized by lower \(T/P\) in the subduction channel and higher \(T/P\) in the overriding plate. During the Paleoarchean to Paleoproterozoic the ambient mantle temperature was warmer than at present by \(\sim 300–150\,^\circ\text{C}\). Although the thermal history of Earth is only poorly constrained, it is likely that prior to ca. 3.0 Ga heating from radioactive decay would have exceeded surface heat loss, whereas since ca. 2.5 Ga secular cooling has dominated the thermal history of the Earth. The advent of paired metamorphism is consistent with other changes in the geological record during the Neoarchean that are best explained as the result of a transition from a stagnant lid to subduction and a global plate tectonics regime by ca. 2.5 Ga. This interpretation is supported by results from 2-D numerical experiments of oceanic subduction that demonstrate a change to one-sided subduction is plausible as upper mantle temperature declined to \(<200\,^\circ\text{C}\) warmer than at present during the late Neoarchean–Paleoproterozoic. This is the beginning of the Proterozoic plate tectonics regime.

At 1.0 Ga the ambient mantle temperature was still \(\sim 150–100\,^\circ\text{C}\) warmer than at present. Continued secular cooling caused a transition to cold subduction registered in the crustal record of metamorphism by the first appearance of blueschist and high to ultrahigh pressure metamorphism during the Neo-proterozoic. Results of 2-D numerical experiments of continental collision demonstrate a transition from
shallow to deep slab breakoff associated with stronger crust–mantle coupling that enabled continental subduction to mantle depths as upper mantle temperature declined to \(<100\, ^\circ\text{C}\) warmer than at present during the late Proterozoic. This is the beginning of the modern plate tectonics regime.

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1. Introduction

On contemporary Earth, regional metamorphism is associated with geodynamic settings of lower or higher than average heat flow, such as subduction zones and arc–back arc systems, or is related to tectonic processes that disrupt the steady-state thermal structure of the lithosphere by thickening or thinning, such as occurs in orogenic hinterlands and collisional mountain belts or during orogenic collapse and regional extension (Brown, 2008, 2009). In addition, the long-term cooling of the lithosphere drives garnet-producing metamorphic reactions, which decreases the buoyancy of crustal roots and enhances lithosphere stability (Fischer, 2002). To extend this understanding of the relationship between geodynamics and metamorphism back in time, we use the forensic methods and techniques of metamorphic petrology.

Thus, metamorphic petrology is concerned with decoding the mineralogical and microstructural record of burial/heating and exhumation/cooling imprinted on pre-existing sedimentary, igneous and metamorphic rocks by processes such as subduction, accretion, trench advance or retreat, collisional orogenesis and orogenic collapse. As petrologists, we address multiple scales of activity, from the pressure \((P)\)–temperature \((T)\)–time \((t)\) evolution of a single rock to the evolution of \(P–T\) in space and time. Both subduction/accretion and collisional orogenesis/orogenic collapse cause perturbations of steady-state geotherms on time scales and length scales constrained by the mechanical properties of the lithosphere, which vary with composition, temperature, fluid/melt presence/absence, and strain rate. Thus, by determining quantitative \(P–T–t\) paths petrologists make available information to parameterize subduction zone processes and collisional orogenesis. Inverting this \(P–T–t\) information, provides constraints on the geodynamic and metasomatic processes involved, and consequently advances understanding of lithosphere evolution and geodynamics.

Although mobile-lid plate tectonics (hereafter “plate tectonics”) has provided us with a context to understand metamorphism and its relationship to tectonics at least as far back as the dawn of the Phanerozoic (Brown, 2008, 2010b), during the Precambrian geodynamics may have been different (Sizova et al., 2010, 2014). For example, a strong case has been made for a stagnant-lid plate tectonics regime on the early Earth (Debaille et al., 2013; Griffin et al., 2013; O’Neill et al., 2013; hereafter “stagnant lid”). In addition, Earth’s mantle was hotter and Moho temperatures were higher in the past, and in the Archean conditions were probably appropriate for lithosphere foundering by Rayleigh–Taylor instabilities (Jull and Kelemen, 2001; Toussaint et al., 2004), which may have played a much more important role in lithosphere evolution on the early Earth than on contemporary Earth (Johnson et al., 2014).

In the first section of this paper I review examples of major discoveries in metamorphic petrology and advances in instrumentation and methods during the last 30 years. This provides a perspective for the second section in which I review evidence from the crustal record of metamorphism and recent results from geodynamic modeling to investigate the major geodynamic transitions in Earth evolution. The first of these concerns the transition from an Archean stagnant-lid regime to a Proterozoic plate tectonics regime, as evidenced by the appearance of paired metamorphism, whereas the second concerns the Cryogenian to Cambrian transition from the Proterozoic plate tectonics regime to the modern plate tectonics regime characterized by cold subduction, as evidenced by the appearance of blueschist and ultrahigh-pressure metamorphism in the geological record.

2. The state of the art in metamorphic petrology

2.1. Brave new world

It is now 30 years since the discovery that some lithospheric materials record evidence of metamorphism at pressures characteristic of mantle conditions (e.g. Chopin, 1984, 2003; Dobzhanskiy et al., 2011). Just a few years earlier, persuasive evidence had appeared in the literature demonstrating that some crustal rocks registered temperatures \(\geq1000\, ^\circ\text{C}\) over large areas at ‘normal’ lower crustal pressures (e.g. Ellis, 1980; Harley, 1998, 2008; Kelsey, 2008). Although both of these statements appear to be broadly correct, it is wise to be circumspect of extreme values in case the recorded pressure was not lithostatic (Mouls et al., 2013) or where the calculated temperature approaches or exceeds the liquidus for the continental crust (\(\sim1100\, ^\circ\text{C}\)). Also during the same period, the realization that melting and melt drainage and migration might play an important role in crustal deformation was becoming widespread (e.g. Hollister and Crawford, 1986; D’Lemos et al., 1992; Brown, 1994, 2010a). Furthermore, it became clear that much of the underlying mantle has undergone dynamic reworking (Mercier and Nicolas, 1975), including metasomatic modifications (Thompson, 1992). This revelation opened both the mantle (Trane et al., 2009; Holland et al., 2013), including the deep mantle (Ganguly et al., 2009), and subducted lithospheric slabs (Hacker et al., 2003a) to study using the methods of metamorphic petrology.

These fascinating developments during the past 30 years have not only changed our perception of the field of metamorphism but also of the role that metamorphic petrology can play in understanding orogenesis. But how have these developments helped us to understand the evolution of the lithosphere and geodynamics during the Precambrian in relation to secular change?

2.2. Metamorphism and tectonics

In addressing the issue of metamorphism and tectonics, both regional and global scales of interactions are of interest. At the regional scale, we are concerned with differences from terrane to terrane that enable us to distinguish the particular combination of tectonic events responsible for the metamorphism of each terrane, and which inform us about tectonic interactions at the terrane scale. At the global scale, we are concerned with similarities in styles of metamorphism and relationship to plate tectonic setting, which relationships are likely to reflect global geodynamic processes.

The asymmetry in the thermal structure characteristic of one-sided subduction creates contrasting thermal environments—lower \(T/P\) in the subduction zone and higher \(T/P\) in the arc–backarc/orogenic hinterland—that are registered in the rock
2.3. What information is provided by metamorphic petrology?

Metamorphic rocks record information about their $P$--$T$--$t$ evolution in the mineral assemblages, the chemistry and variation in composition of these minerals, and the microstructural relationships among them. It is the job of the metamorphic petrologist to invert this information to retrieve quantitative data about spatial variations in the $P$--$T$--$t$ evolution of ancient metamorphic belts and to use forward models to better understand the tectonic controls over these variations in evolution.

Substantial a posteriori evidence indicates that mineral assemblages in rocks undergoing metamorphism equilibrate continuously on some scale throughout the prograde evolution as fluid or melt is being generated, but undergo little or no change during the retrograde evolution once the rock becomes fluid absent or after any remaining melt has reacted with peak minerals while cooling to the solidus (Pollard et al., 2005). As a result, an equilibrium mineral assemblage is likely to be preserved once fluid is lost under subsolidus conditions (around the peak $T$ of metamorphism) or as the last dregs of melt crystallize in residual rocks at suprasolidus conditions, this assemblage reflecting the $P$--$T$ conditions of fluid loss or the solidus. Thus, the mineral assemblage of a metamorphic rock is the key to estimating the conditions prevailing when it formed. Furthermore, this is the basic inference underlying the concept of metamorphic facies (Fig. 1a), which are defined by sets of mineral assemblages developed in rocks of different bulk compositions that recur through space and time. Thus, metamorphic facies represent the overlapping stability fields in $P$--$T$ space for these mineral assemblages and their recurrence supports the notion of equilibrium at or soon after peak $T$. Notwithstanding the success of the facies concept, today it is preferable to think in terms of

![Figure 1](image-url)
apparent thermal gradients in the classification of types of metamorphism (Fig. 1b; Brown, 2006, 2007a,b, 2008; Stüwe, 2007). The rationale for the view will be explained below.

Although the rock record stretches back to ca. 4.0 Ga, the first billion years is represented by only one millionth of the exposed continental crust and retrieving quantitative P–T–t information is rarely possible. However, there is a reliable record of P–T–t information retrieved from metamorphic belts since ca. 3.0 Ga (Brown, 2006, 2007a,b), the likely beginning of a transition to one-sided subduction and steady-state plate tectonics on Earth (Brown, 2008; Sizova et al., 2010; Dhuime et al., 2012; van Hunen and Moyen, 2012; Johnson et al., 2014).

2.4. Advances in numerical tools, analytical instrumentation, geophysical imaging and geodynamic modeling

During the past 30 years the range of numerical tools and instrumentation available to the metamorphic petrologist has expanded beyond imagination. Examples include thermodynamic modeling and in situ analytical methods for U/Th–Pb isotope and trace element chemistry of minerals, and for mineral characterization.

The present level of sophistication in thermodynamic modeling of phase equilibria has its foundation in the 1980s with the introduction of appropriate internally-consistent thermodynamic datasets (Holland and Powell, 1985, 1998, 2011; Berman, 1988) and the development and expansion of suitable activity–composition models for phases of interest in parallel with the evolution of software for quantitative phase equilibria modeling (Powell and Holland, 1985, 1988; Powell et al., 1998; White et al., 2007, 2014, see also the THERMOCALC website at http://www.metamorph.geo.uni-mainz.de/thermocalc/index.html; Berman, 1988, 1991; Brown et al., 1988; Connolly, 1990, 2005, 2009; Connolly and Petrin, 2002; Connolly and Kerrick, 1987, see also the PerpleX website at http://www.perplex.ethz.ch/; de Capitani and Brown, 1987; de Capitani, 1994; de Capitani and Petrakis, 2010, see also the Theriaik website at http://titan.minpet.unibas.ch/minpet/theriaik/theriuser.html). Although the use of in situ U/Th–Pb geochronology (e.g. the SHRIMP on accessory minerals was developed during the 1980s and became accepted during the 1990s (Williams, 2013), it was only with the advent of routine in situ trace element analysis of minerals, particularly zircon, monazite and garnet (Harley and Kelly, 2007; Harley et al., 2007; Rubatto and Hermann, 2007), that it has become possible to link ages reliably to calculated P–T–t conditions. The recent introduction of laser ablation split-stream petrochronology (Kylander-Clark et al., 2012), in which U/Th–Pb isotope and trace element analyses are made simultaneously, provides a rapid method to obtain these data. This advance will not only enable ages to be linked to P–T–t conditions via rare earth element data from accessory minerals and garnet (e.g. Donaldson et al., 2013), but also will allow campaign-style geochronology that will provide datasets to improve our understanding of the spatial and temporal variations in P–T–t conditions in metamorphic belts (e.g. Spencer et al., 2013).

There are numerous new analytical methods that are now used routinely in metamorphic petrology—too many to review in detail here. Among spectroscopic methods employed in the characterization of geological materials (Beran and Libowitzky, 2004), Raman spectroscopy (Dubessy et al., 2012) is a highly flexible non-destructive technique used not only in the analysis of fluids, minerals and glasses but also in thermobarometry. For example, Raman spectroscopy is routinely used in the characterization of minerals such as diamond and coesite in UHPM rocks (Nasdala and Massonne, 2000; O’Brien and Ziemann, 2008), in the assessment of peak-metamorphic conditions using carbonaceous (graphitic) materials (Beya et al., 2002), and in the characterization of glass (melt) and nanogrante inclusions in migmatites and granulites (CESARE et al., 2011). In addition, given a calculated or assumed T or P, Raman spectroscopy of mineral inclusions in other minerals may be used to determine either P or T (Kohn, 2014). This technique utilizes the pressure-dependence of the Raman peak positions in combination with differences in compressibility (pressure-sensitive) and thermal expansivity (temperature-sensitive) between the inclusion and host mineral, although the residual pressure retrieved depends on the elastic parameters of the host mineral (Izrael et al., 1999; Parkinson and Katayama, 1999; Sobolev et al., 2000; Guiraud and Powell, 2006; Enami et al., 2007).

There have also been significant advances in geophysical imaging during the past 30 years, in part driven by large National programs, such as Lithoprobe (Canada), Earthscope (USA) and SinoProbe (China), with the overall aim of exploring the composition, structure and evolution of the continental lithosphere. These advances include: establishing differences between Archean and Proterozoic lithosphere (Artemieva and Mooney, 2001); identifying melt in the deep crust under contemporary orogens and quantifying its effect on crustal rheology (Nelson et al., 1996; Li et al., 2003; Unsworth et al., 2005), which aids in the interpretation of deeply eroded orogens; and, recognizing reaction fronts in orogenic crust (Mechie et al., 2004), identifying the water flux delivered to the deep Earth by subduction (Savage, 2012) and linking metamorphic reactions in actively subducting lithosphere to earthquake genesis (Hacker et al., 2003b; Nakajima et al., 2013).

The use of numerical experiments to investigate mantle convection (e.g. Doin et al., 1999; Twissley, 2000; Davies, 2004) and the thermo-mechanical evolution of the lithosphere during orogenesis (e.g. Royden, 1993a,b, 1996; Willett et al., 1993; Beaumont et al., 1994, 1996a,b; Pfeffer et al., 2000; Faccenda et al., 2008) and the supercontinent cycle (e.g. Yoshida and Santosh, 2011; Rolf et al., 2012) has expanded as computer power has increased to enable more sophisticated geodynamic modeling (e.g. Gerya and Yuen, 2003, 2007). Geodynamic modeling has been used to address important processes such as slab breakoff (Gerya et al., 2004; Dutert et al., 2011), shallow subduction (van Hunen et al., 2004; Currie and Beaumont, 2011), ridge subduction (Groome and Thorkelson, 2005), deep crustal flow (Royden et al., 1997; Clark and Royden, 2000; Beaumont et al., 2004), delamination (Gray and Pyskyewycz, 2012; Ueda et al., 2012), intraplate orogenesis (Nelson and Houseman, 1999; Pyskyewycz and Beaumont, 2004; Gorczyk et al., 2012, 2013; Gorczyk and Vogt, 2014), and exhumation of UHPM terranes (Gerya et al., 2002; Warren et al., 2008; Beaumont et al., 2009; Sizova et al., 2012). Geodynamic modeling has also been used to evaluate crustal growth at active continental margins (Vogt et al., 2012; Zhu et al., 2013) and to understand the links between metamorphism and tectonics (Jameson and Beaumont, 1988; Sandiford and Powell, 1990; Jameson et al., 2002, 2004; Gerya and Stockhert, 2006; Li et al., 2010). Of course, geodynamic modeling has also been applied to the problem of Precambrian geodynamics (van Thienen et al., 2004; van Hunen and van den Berg, 2008; Sizova et al., 2010, 2014; Gerya, 2014; Johnson et al., 2014) and tectonics (Perchuk and Gerya, 2011), and the formation of the cratonic roots composed of strongly depleted sub-continenental mantle lithosphere (Gray and Pyskyewycz, 2010; Griffin et al., 2013).

3. The evolution of the continental lithosphere

In this section I set out a frame of reference for the thermal evolution of the continental lithosphere based on my own work in metamorphic petrology and work done in collaboration with others that contributes to a better understanding of geodynamics. These
contributions are twofold. First, we have used a comprehensive dataset of PeTet information from Neoarchean to Cenozoic crust (Brown, 2006, 2007a,b) coupled with 2-d numerical geodynamic modeling (Sizova et al., 2010, 2012, 2014) to investigate secular change. This work suggests that there may have been three global-scale geodynamic regimes that were separated by two periods of change on Earth. The first of these is the initiation of subduction and the transition from a dominantly stagnant-lid regime to a plate

**Figure 2.** To show snapshots from a geodynamic model of delamination and recycling of crust caused by gravitational instabilities in a stagnant-lid plate tectonics regime. In the experiment shown, the initial primary crust was 45 km thick with a mantle potential temperature at the base of the lithosphere of 1600 °C. (a) To show the model configuration and initial conditions (dark blue, initial primary crust; light blue, negatively buoyant initial primary crust; dark pink brown, lithospheric mantle; light pink-brown, unmelted asthenosphere; yellow, melt-bearing asthenosphere). (b) To show local thickening of the initial primary crust by addition of new primary crust (green) and delamination of negatively buoyant initial primary crust. The positive buoyancy of the melt-bearing asthenosphere is the primary cause of the large-scale mantle flow; crustal drips induce smaller circulation patterns. (c) To show density (left) and viscosity (right) for the left-hand and right-hand halves of b, respectively. (d) To show local thickening and delamination of the negatively buoyant new primary crust. Figure reproduced from Johnson et al. (2014).
tectonics regime in the late Archean—early Proterozoic, whereas the second is the transition to cold subduction that occurred during the Ediacaran—Cambrian. Second, we have used phase equilibria modeling of Earth’s early lithosphere coupled with 2-D numerical geodynamic modeling of lithosphere stability to investigate early Archean geodynamics (Johnson et al., 2014). The results of this last study demonstrate the plausibility of a stagnant lid regime and have implications for the production of tonalite–trondhjemite–granodiorite suite magmas in the early Archean.

3.1. Early Archean (pre-Mesoarchean) lithosphere and geodynamics

3.1.1. Lithosphere

The evolution of the continental lithosphere is a response to geodynamics, which in turn is a response to the thermal evolution of Earth. There are differences in the geology and tectonic style between Archean and post-Archean crust. Gray gneisses and plutonic complexes of mainly tonalite–trondhjemite–granodiorite (TTG) suite affinity dominate the Archean geological record with volcano-sedimentary greenstone belts forming a minor component (Goodwin, 1991). Archean TTG gneisses typically have higher Na/K and Na/(Ca + Na + K) than post-Archean gneisses (e.g. Keller and Schoene, 2012), and high ratios of light to heavy rare-earth elements (L/HRREE, for example, La/Yb). They are inferred to have been derived from sources of broadly basaltic composition (e.g. Foley et al., 2002; Moyen, 2011; Ziaja et al., 2014) in which HREE depletion reflects sequestration by residual garnet. The sub-continental lithospheric mantle (SCLM) underlying Archean crust is also different from the post-Archean SCLM (e.g. Hamilton, 1998; Djomani et al., 2001; Griffin et al., 2013). Different geology may indicate different geodynamic behavior and if this is true the differences are probably related to Earth’s thermal evolution and mechanism of heat loss (Korenaga, 2006; Labrosse and Jaupart, 2007).

Volcanic rocks of Archean greenstone belts are dominantly subaqueous (Arndt, 1999). They comprise mostly primary high-Mg non-arc basalts with 18–24 wt.% MgO produced by 30–45 vol.% partial melting of the ambient mantle together with lesser amounts of plume-related komatiites (de Wit and Ashwal, 1997), both of which indicate that the mantle was considerable hotter in the Mesoarchean–Neoarchean than today (Herzberg et al., 2007, 2010). This is consistent with arguments based on the constancy of continental freeboard (Gale and Mezger, 1998; Flamant et al., 2008). The associated sedimentary rocks are mostly cherts and ironstones (Goodwin, 1991), and clastic sedimentary rocks only appear widely in the geological record after 3.2 Ga (Erikkson and Wilde, 2010). This is consistent with modeling that indicates the continents were mostly flooded until the Neoarchean (Flament et al., 2008, 2011).

Although the thermal history of Earth is only poorly constrained (Korenaga, 2006, 2008, 2011, 2013; Labrosse and Jaupart, 2007; Silver and Bohn, 2008; Davies, 2009; van Hunen and Moyen, 2012), it is likely that prior to ca 3.0 Ga heating from radioactive decay would have exceeded surface heat loss, whereas since ca 2.5 Ga secular cooling has dominated the thermal history of the Earth. Ambient upper mantle temperature controls the tectonic regime and consequently the mechanisms(s) for formation of primary crust and its reworking (Sizova et al., 2010, 2014; Johnson et al., 2014). Petrological data and generalizations from thermal history models suggest ambient upper-mantle temperatures in the Archean were significantly hotter than the present day, but with similar global variation in potential temperature of magma sources of \( \sim 120^\circ\text{C} \) (Herzberg et al., 2007, 2010; van Hunen and Moyen, 2012), leading to a high percent of melting and generating thick MgO-rich primary crust underlain by highly residual mantle (Herzberg and Rudnick, 2012). However, the preserved volume of this crust is low suggesting much of it was recycled. Furthermore, since TTGs must have been sourced from broadly basaltic protoliths, as discussed above, they cannot have been generated directly from this primary MgO-rich crust.

Given the dependence of rheology on \( T \) and \( \rho_{\text{in situ}} \), the geodynamic regime in the Hadean–Archean may have been different and models for the formation of Hadean–Archean crust based on uniformitarian principles may be misleading. Geodynamic regimes proposed for the Hadean–Archean include: uniformitarian/non-uniformitarian mobile-lid plate tectonics; stagnant-lid plate tectonics, without subduction; episodic subduction events in a dominantly stagnant-lid plate tectonics regime; or, variation in regime from place to place and time to time. Further numerical geodynamic modeling, such as that recently done by Griffin et al. (2013), O’Neill et al. (2013) and Johnson et al. (2014), is required to separate plausible from improbable geodynamic regimes for this period in Earth history.

3.1.2. Geodynamics

The stability of primary crust in the early Archean may be investigated by coupling calculated phase equilibria for hydrated and anhydrous low to high MgO crust compositions and their complementary mantle residues for a range of temperatures with 2-D numerical geodynamic models to evaluate potential lithospheric instability (Johnson et al., 2014). The study by Johnson et al. (2014) showed that the density of primary crust increases more than the density of residual mantle decreases with increasing ambient mantle potential temperature. As a result, the base of MgO-rich primary crust could have become gravitationally unstable even when fully hydrated. Furthermore, Johnson et al. (2014) demonstrated that magmatically over-thickened MgO-rich crust (i.e. over-thickened with respect to the initial thickness) could have delaminated by Rayleigh–Taylor instabilities at mantle potential temperatures >1500–1550°C. Delamination would generate a return flow of asthenospheric mantle and melting in both the dripping crust and the rising mantle. This process is illustrated in Fig. 2, which shows snapshots from a numerical experiment with 45-km-thick initial primary crust and mantle potential temperature of 1600°C. Melting at the base of the overthickened hydrated primary crust is also plausible at the predicted temperatures.

Since most Archean primary crust would have had MgO >18 wt.% TTG magmas in the Archean cannot have been produced directly from this crust (Johnson et al., 2014). Garnet-bearing hydrated basaltic crust with MgO <18 wt.% is required as the source of most TTGs (Johnson et al., 2014). Partial melting of the primary crust could have generated this lower MgO secondary crust; also, some lower MgO crust could have formed by crystal fractionation of sub-volcanic primary magmas. Such a scenario in which extensive reworking of the primary crust is a dominant process explains the preponderance of magmatic rocks of TTG affinity in the early Archean. Moreover, it provides an explanation for the fragmentary metamorphic record, since this early TTG crust was also subjected to extensive reworking (e.g. Friend and Nutman, 2005; Horie et al., 2010; Nutman et al., 2013). Additionally, this scenario is consistent with production of the SCLM in a manner similar to that suggested by Spengler et al. (2006, 2008) via extensive melting in the Archean and subsequent modification, although it may have been thickened subsequently (Gray and Pysklywec, 2010).

Unfortunately, modern quantitative metamorphic studies on crustal rocks older than Neoarchean are generally lacking. This gap in our knowledge must be filled in the future if we are to understand fully the evolution of the lithosphere and early Archean geodynamics. Using the limited available data, Brown (2009) argued that Paleoarchean–Mesoarchean crust generally registers
low-to-moderate-\(P\)–moderate-to-high-\(T\) conditions, implying high but uniform apparent thermal gradients of 850–1350 °C/GPa. Ultra-high temperature metamorphism appears to be absent from the geological record before the Neoarchean (Brown, 2009), although this may reflect a lack of recognition, and there is no imprint in the geological record of subduction of continental crust to mantle conditions prior to the Cryogenian (Brown, 2009), most likely due to decoupling of continental crust from the underlying mantle lithosphere during subduction and/or shallow slab breakoff on a hotter Earth (Sizova et al., 2014). This record is inconsistent with one-sided subduction and contemporary plate tectonics (Stern, 2005; Brown, 2006, 2010b), as discussed above. This raises the question: at what point in the geological record of metamorphism do we find evidence of one-sided subduction and plate tectonics?

3.2. Post-Paleoarchean lithosphere and geodynamics

In contrast to the early Archean, modern quantitative metamorphic studies are widely available for crustal rocks younger than Mesoarchean. Where these studies may be linked to reliable geochronology, it is possible to interrogate the geological record to investigate the thermal history of the lithosphere.

3.2.1. Methods used to interrogate the crustal record of metamorphism

One aim of my recent metamorphic studies has been to invert the record of crustal metamorphism as a proxy for earlier thermal environments, particularly to elucidate changes in the geodynamic history of Earth, and then to test the outcome using 2-D numerical geodynamic modeling. There are two principal caveats associated with an approach based on the geological record. First, metamorphic imprints in the crust record discrete events, for example a change in plate kinematics or the evolution from subduction to terminal collision, whereas subduction is thought to be globally continuous, at least on a timescale shorter than the supercontinent cycle. Second, the record of crustal metamorphism is a function of what is preserved—if the preservation potential of some types of metamorphic rock was poor earlier in Earth history evidence of their existence may not have been retained and the geological record may be biased. Indeed, we know that the crustal record of metamorphism is generally biased in favor of preservation during periods of amalgamation of the continental lithosphere into supercontinents (Brown, 2006, 2007a,b; Hawkesworth et al., 2009; Cawood et al., 2013).

This type of analysis relies on two important features of high-temperature and high-pressure metamorphic rocks. First, that the close-to-peak mineral assemblages are robust recorders of \(P\) and \(T\). This was discussed above in relation to metamorphic facies. In particular, at higher temperatures, prograde dehydration and melt loss produce nominally anhydrous mineral assemblages that are difficult to retrogress or overprint without fluid influx. For this reason, this study has been limited to rocks equilibrated under conditions of relatively high temperature, such as granulites and...
eclogites. Thus, for granulite facies metamorphism at $P < 1$ GPa a minimum temperature for inclusion was set at 700 °C, whereas for higher pressure metamorphism at $T < 700$ °C a minimum pressure of 1 GPa was applied to ensure a reasonable minimum temperature and the likelihood of an equilibrated peak mineral assemblage. Using these thresholds, the resulting $P-e$ data are believed to be robust. Second, that in rocks overprinted during exhumation, close-to-peak phase assemblages are commonly preserved as inclusions in rock-forming or accessory minerals. This has proved to be particularly important for retrieving reliable information from ultrahigh-pressure metamorphic rocks.

In the discussion below, 214 $P-e$ data, which is about a 50% increase over the size of the dataset used by Brown (2007a), have been grouped into three different types of metamorphism, as follows. (1) High-pressure—ultrahigh pressure (HP-UHP) metamorphism, characterized by lawsonite blueschists/lawsonite eclogites (27 examples) or low-temperature eclogites/UHP metamorphic rocks (59 examples), where $T$ is that registered at maximum $P$, and the age of metamorphism is generally interpreted to be that around peak $P$. (2) Eclogite—high-pressure granulite (E-HPG) metamorphism (52 examples), characterized by medium-temperature eclogite or high-pressure granulite mineral assemblages, where peak $T$ generally was achieved after maximum $P$, and the age of metamorphism is generally interpreted to be that around or slightly post peak $T$. (3) Granulite—ultrahigh-pressure (G-UHT) metamorphism (76 examples), characterized by granulite or UHT mineral assemblages, where $P$ is that registered at maximum $T$, and age of metamorphism is generally interpreted to be around or post—peak $T$, probably recording post-peak cooling across an elevated solidus in many cases, particularly in melt-depleted UHT metamorphic rocks.

As a general statement for the Phanerozoic, lawsonite blueschists/lawsonite eclogites have been interpreted to record oceanic subduction, whereas low-temperature eclogites/UHP metamorphic rocks have been interpreted to record continental subduction and to define the suture formed during the transition to terminal continent—continent collision. However, with more studies during the past decade, this distinction is becoming less clear (Rebay et al., 2010; Wei and Clarke, 2011). Thus, the development of lawsonite appears to be controlled more by H2O-saturation versus under-saturation for a fixed bulk rock chemical composition than by tectonic setting (Rebay et al., 2010), and the development of blueschist versus eclogite seems to relate more to the bulk rock chemical composition—Ca-rich versus Ca-poor, respectively (Wei and Clarke, 2011). During the Proterozoic and the Phanerozoic, metamorphism associated with terminal continent—continent collision is generally characterized by medium-temperature eclogites and high-pressure granulites. In contrast, normal granulites and UHT metamorphic rocks were most likely generated in oceanic/continental arc—backarc settings or in orogenic hinterlands, such as the Altiplano or Tibetan plateau.

I follow Miyashiro (1961) in believing that it is useful to classify metamorphic belts by type based on apparent thermal gradient in order to identify tectonic setting and understand secular change in global geodynamics. Apparent thermal gradient is the slope of the line from 0 °C/0 GPa through the peak $P-e$ conditions as defined above; this is neither the metamorphic field gradient nor the piezothermic array (England and Richardson, 1977; Richardson and

Figure 5. Peak metamorphic $P-T$ conditions (as defined in the text) of metamorphic belts by type plotted on the $P-T$ diagram of Fig. 1, extended to 6.0 GPa; in (b) four apparent thermal gradients are shown, as discussed in the text. Types of metamorphism are: granulite—ultrahigh temperature (G-UHT — orange circles) metamorphism; eclogite—high-pressure granulite metamorphism (E-HPG — light purple diamonds); and, high-pressure—ultrahigh-pressure metamorphism (HP-UHP), comprising lawsonite-bearing blueschists and eclogites (pale blue squares) and eclogites and ultrahigh-pressure metamorphic rocks (dark blue squares).
3.2.2. Results

Figs. 3–6 show plots of metamorphic belts classified by type, based on a judgment of the current best estimate of representative peak $P–T$ conditions and age associated with these conditions for each metamorphic belt (Michael Brown, unpublished; based on the dataset from Brown (2007a, Tables 1–5) as updated in December 2013). From Fig. 3, it is clear that temperatures of metamorphism were uniformly high from the Neoarchean to the Paleozoic, but there is a dramatic drop in the temperature of metamorphism with the appearance of HP-UHP metamorphism in the Cryogenian–Ediacaran. This is unlikely to be related to the exclusion of data from low-temperature metamorphism at low-presures, a conclusion supported by the data in Figs. 4 and 5. From Fig. 4, it is clear that pressures of metamorphism were uniformly low (G-UHT metamorphism) or moderate (E-HPG metamorphism) from the Neoarchean to the Paleozoic, but there is a dramatic increase in the pressure of metamorphism with the appearance of HP-UHP metamorphism.

In Fig. 5 it can be seen that although there is some variation in apparent thermal gradient within each type of metamorphism, nonetheless each falls within its own range of apparent thermal gradients, as follows. HP-UHP metamorphism mostly lies between apparent thermal gradients of 150 °C/GPa and 350 °C/GPa, with mean values of 242 °C/GPa for lawsonite blueschists/lawsonite eclogites, and 251 °C/GPa for low-temperature eclogites/UHP metamorphic rocks, respectively. E-HPG metamorphism mostly lies between apparent thermal gradients of 350 °C/GPa and 750 °C/GPa, with a mean value of 555 °C/GPa. G-UHT metamorphism mostly lies between apparent thermal gradients of 750 °C/GPa and 1500 °C/GPa, with a mean value of 1092 °C/GPa. Notwithstanding that this reflects the robustness of the metamorphic facies concept and is to be expected, it implies that each type of metamorphism is likely to have a characteristic tectonic setting with which it was associated, as discussed above. This becomes important when the temporal distribution of the different types of metamorphism is considered.

The temporal record of apparent thermal gradients retrieved from crustal rocks also provides information about secular change in thermal regimes, as shown in Fig. 6, which plots apparent thermal gradient against age. Simple linear regression suggests there is no significant change of apparent thermal gradient with age for G-UHT metamorphism. However, for E-HPG metamorphism, the apparent thermal gradient appears to decline slightly from the Neoarchean to the Paleozoic, although the slope of the linear regression is only 27 °C/GPa/Ga. Although simple linear regression suggests there is no significant change of apparent thermal gradient with age for low-temperature eclogites/UHP metamorphic rocks, the apparent thermal gradient appears to decline slightly through the Phanerozoic for lawsonite blueschists/lawsonite eclogites, although the slope of the linear regression is only 60 °C/GPa/Ga. However, based on average apparent thermal gradient, there is no significant difference between the thermal gradients recorded by low-temperature eclogites/UHP metamorphic rocks and lawsonite blueschists/lawsonite eclogites. Rather than recording “abnormally low geothermal gradients” (Tsujimori et al., 2006), both low-temperature eclogites/UHP metamorphic rocks and lawsonite blueschists/lawsonite eclogites record the characteristic thermal gradients of Phanerozoic subduction and differences in mineralogy appear to be largely a function of H₂O-saturation and bulk chemical composition rather than thermal gradient, as discussed above (Rebay et al., 2010; Wei and Clarke, 2011).

Using these data we may investigate aspects of the thermal history of Earth, as recorded by the geological record, and the evolution of the lithosphere since the Mesoarchean. Furthermore, these data provide information that will help to clarify Precambrian geodynamics.

3.2.3. Mesoarchean to Neoarchean geodynamics

In the Mesoarchean–Neoarchean, the sporadic appearance of two types of metamorphism—E-HPG and G-UHT—with distinct apparent thermal gradients marks a transition to one-sided subduction and plate tectonics as the dominant tectonic regime (Brown, 2006, 2007a,b, 2008). This transition and subsequent changes in tectonic style may be investigated using 2-d numerical geodynamic modeling (van Hunen and van den Berg, 2008; Sizova et al., 2010). In a series of experiments, Sizova et al. (2010) systematically investigated the dependence of geodynamic regime at a convergent ocean–continent plate margin on ambient upper-mantle temperature and a variety of additional parameters that have been suggested to be different in the Archean.

With increasing ambient upper-mantle temperature, the numerical experiments show unequivocally that one-sided subduction is destabilized principally by weakening of the oceanic lithosphere consequent upon increased melt flux from the underlying asthenospheric mantle. Inverting the sequence, there is a first-order transition from a no-subduction regime through a pre-subduction regime, involving limited underthrusting of oceanic
Figure 7. Snapshots from a geodynamic model used to investigate the transition from mobile-lid to stagnant-lid plate tectonics with increasing temperature of the ambient upper mantle. In the experiments, the upper-mantle temperature ($\Delta T$) and the crustal radiogenic heat production ($H$) are varied independently from the present day values. The three experiments shown identify three different tectonic regimes: (a) the modern subduction regime ($\Delta T = 0$–175 K); (b) a pre-subduction regime characterized by underthrusting but not subduction ($\Delta T = 175$–250 K); and, (c) a no-subduction regime ($\Delta T > 250$ K). The lower boundary of the melt-bearing asthenospheric mantle (red areas) in (a) and (b) is located at depths of 100–150 km, whereas in (c) this boundary extends to depths of 200 km (i.e. below the bottom of the model). Figure reproduced from Sizova et al. (2010) with permission (RightsLink License Number 3292490716639).
Figure 8. Snapshots from a geodynamic model used to investigate the transition from the modern collision regime to a hot collision regime with increasing temperature of the ambient upper mantle. Shown are two examples of different exhumation styles for UHPM rocks in the modern collision regime (large-scale crustal stacking and trans-lithospheric diapirism) and the consequences of a hotter mantle for the depth of slab breakoff. (a) Large-scale evolution of the large-scale crustal stacking model with a 600 km long oceanic plate and an abrupt (50 km wide) passive margin. The detailed evolution of the collision zone for this model is shown in Sizova et al. (2014, Fig. 2). (b) Large-scale evolution of the crustal stacking model at $\Delta T = 100$ K. A detailed snapshot of the collision zone for this model is shown in Sizova et al. (2014, Fig. 3e). (c) Large-scale evolution of the trans-lithospheric diapirism model with a 600 km long oceanic plate and a gradual (150 km wide) passive margin. The detailed evolution of the collision zone for this model is shown in Sizova et al. (2014, Fig. 6). (d) Large-scale evolution of the trans-lithospheric diapirism model at $\Delta T = 150$ K. A detailed snapshot of the collision zone for this model is shown in Sizova et al. (2014, Fig. 7c). Figures reproduced from Sizova et al. (2014) with permission (RightsLink License Number 3292490560289).
lithosphere along an imposed zone of weakening at the edge of the continental lithosphere, to subduction as ambient upper-mantle temperatures decline from >200 to <175 °C warmer than present-day ambient upper-mantle. The experimental results are summarized in the series of snapshots from three numerical experiments with ambient upper mantle temperature shown in Fig. 7.

Neither higher radiogenic heat production nor an increased thickness of ocean crust have any significant effect on the ambient upper mantle temperature at which subduction becomes impossible. Furthermore, although increasing the density of the ocean crust via the transformation of basalt to eclogite, increasing the velocity of the ocean plate and decreasing the density of the subcontinental lithospheric mantle allow subduction to continue to ~50 °C higher ambient upper mantle temperature, subduction eventually becomes impossible at ambient upper-mantle temperatures >200 °C warmer than the present-day ambient upper-mantle regardless of changes to these parameters (Sizova et al., 2010).

Taking a present-day mid-point value of 1340 °C for the mantle potential temperature of ambient mantle below oceanic ridges (Herzberg et al., 2007), the transition from a stagnant lid regime dominated by Rayleigh–Taylor instabilities to subduction and plate tectonics likely occurred at a mantle potential temperature in the range 1550–1500 °C, which corresponds to ambient mantle potential temperatures during the Neoarchean–Paleoproterozoic (Herzberg et al., 2010; Johnson et al., 2014). Thus, the transition to dominantly steep-slab subduction occurred during the late Archean to early Proterozoic as Rayleigh–Taylor instabilities became increasingly inefficient with declining ambient mantle potential temperature (Johnson et al., 2014). This transition is consistent with the scarcity of eclogites in the orogenic rock record before 1.9 Ga, whereas eclogites recording P up to 2.0 GPa become more common through the remainder of the Proterozoic. This is the beginning of the Proterozoic plate tectonics regime.

One feature argued to be characteristic of the Archean is the reported similarity of metamorphic facies and erosion level from craton to craton (e.g., Binns et al., 1976; Galer and Mezger, 1998). This postulate may be examined by reference to Figs. 3, 4 and 6. From Fig. 3 it is clear that peak temperature varies widely within and overlaps considerably between G-UHT and E-HPG types of metamorphism. In contrast, from Fig. 4 we see that pressure does separate these types of metamorphism with minimal overlap, even in the Archean. Thus, in Fig. 6 these two types of metamorphism are seen to have distinct apparent thermal gradients. Thus, the postulate of similarity of metamorphic facies and erosion level from craton to craton is not supported by the data, except that the average apparent thermal gradients for the paired E-HPG and G-UHT metamorphism are similar from throughout the Neoarchean.

### 3.2.4. Proterozoic geodynamics

Although subduction is generally considered to be an ongoing close-to-static-state process, it is clearly interrupted from time to time by plate collisions, such as that occurring at the present day along the northern edge of the Australian plate, and cyclically by the formation of supercontinents, although to what extent these processes change the relative lengths of the different types of plate boundaries and/or the rates of spreading and subduction is uncertain (Korenaga, 2008; Silver and Behn, 2008). Therefore, it is no surprise to discover that the age distribution of exposed metamorphic rocks during the Proterozoic is clustered. In Fig. 6 it is apparent that there are gaps in the data in the intervals 2.45–2.05 Ga, 1.53–1.15 Ga and 0.92–0.66 Ga, which correspond in turn to supercraton breakup (Bleeker, 2003) and the supercontinent cycle from Nuna (Columbia; Zhao et al., 2004) to Rodinia (Li et al., 2008) and Gondwana–Pangea (Ernst et al., 2013). This suggests that during the Proterozoic collision-related metamorphic rocks with apparent thermal gradients of 350–750 °C/GPa were preserved as a result of orogenesis associated with supercontinent assembly (Brown, 2008, 2009). Thus, the metamorphic record does not allow us to address the issue of whether, once established, Proterozoic plate tectonics continued without interruption or was shut down by supercontinent assembly, as argued by Silver and Behn (2008), or was episodic, driven by mantle avalanche events, as argued by O’Neill et al. (2007). Furthermore, this issue cannot be resolved by arguments based on apparent gaps in the record of magmatic activity, as argued by Condie et al. (2009), or its perceived continuity, as argued by Partin et al. (2014), since the record may be incomplete due to differential preservation of igneous rocks from tectonic setting to tectonic setting (Hawkesworth et al., 2009; Cawood et al., 2013).

### 3.2.5. Proterozoic versus Phanerozoic orogenesis

There are differences in the style of collision tectonics between Proterozoic and Phanerzoic orogens, such as the widespread occurrence of G-UHT in the former and of blueschists and UHPM in the latter. What controls these differences in style may be investigated using 2-D numerical geodynamic modeling. To provide a starting point it is necessary to understand how contemporary continental collisions work on Earth (Sizova et al., 2012) and then extrapolate these models back in time. To do this, Sizova et al. (2014) varied four first-order parameters in a series of numerical experiments. Parameters investigated were ambient upper-mantle temperature (increased in a series of steps up to 150 °C higher than the present-day value), radiogenic heat production in the crust (increased to one and a half times the present value), thickness of the continental lithosphere (increased from 140 km to 160 km), and the chemical density contrast between the subcontinental lithospheric mantle and the asthenospheric mantle (increased from a difference of 20 kg/m3, corresponding to the present-day, to 50 kg/m3, corresponding to the Proterozoic (Djomani et al., 2001)). Results of their 2-D numerical experiments show that an increase in the mantle potential temperature to >80–100 °C above present values, corresponding to conditions of the late Proterozoic, leads to shallow slab-breakoff (Fig. 8) and modes of collision tectonics that are distinctly different from contemporary continental collision. Thus, it is the transition from shallow slab breakoff to deep slab breakoff during the Cryogenian–Cambrian and stronger crust–mantle coupling that enabled continental subduction to mantle depths as upper mantle temperature declined to <100 °C warmer than at present. This is the beginning of the modern plate tectonics regime.

The transition to the modern plate tectonics regime is shown dramatically if the data from metamorphic belts are separated into three age groups—older than the Cryogenian (>0.850 Ga), Cryogenian to Cambrian (0.850–0.485 Ga) and younger than the Cambrian (<0.485 Ga)—and replotted in P–T space, as shown in Fig. 5. The Cryogenian to Cambrian period represents the global transition from shallow to deep slab breakoff as the overall decline in ambient upper mantle temperatures allowed the lithosphere to strengthen sufficiently to maintain coherency during subduction to greater mantle depths before breakoff occurred. Deeper subduction prior to slab breakoff generates a colder environment in the subduction zone, as recorded by the appearance of HP-UHP metamorphism in the geological record (Maruyama and Liou, 1998). This regime is not only characterized by cold subduction and metamorphism of continental crust at mantle depths, but also by the decline in occurrence of ultrahigh-temperature metamorphism and its virtual absence from the geological record after the Paleozoic (Figs. 6 and 9).
Figure 9. Metamorphic belts plotted by age in relation to thermal gradients where white symbols are belts where peak $P$–$T$ conditions were recorded at $>0.850$ GPa, gray symbols are belts where peak $P$–$T$ conditions were recorded at $<0.485$ GPa, and filled symbols are belts where peak $P$–$T$ conditions were recorded at $<0.485$ GPa. Types of metamorphism are: G-UHT metamorphism — circles; E-HPG metamorphism — diamonds; and, HP-UHP metamorphism — squares.

Sizova et al. (2014) identified two different tectonic regimes at upper mantle temperature $>80$–100 °C warmer than at present, first a truncated hot collision regime, controlled by the presence of strong lower continental crust, and second a two-sided hot collision regime, controlled by the presence of weak lower continental crust. Shallow slab breakoff in both regimes precludes formation of UHPM rocks (Sizova et al., 2014). This transition to the modern collision regime would have occurred during the late Proterozoic, suggesting that collisional tectonic regimes in the late Proterozoic were different compared to those in the Phanerozoic.

There is a range of different styles of collisional tectonic regimes recorded by Proterozoic orogens, which may reflect the variation in mantle potential temperature globally. Some Proterozoic orogens record heating prior to thickening, expressed as CCW metamorphic $P$–$T$–$t$ paths that reach G-UHT metamorphic conditions, followed by close-to-isobaric cooling; these orogens have similarities with ongoing convergence following truncated hot collision in the geo-dynamic models. Examples of this first style of Proterozoic orogenesis include the Paleoproterozoic Khondalite belt in the North China craton and the late Mesoproterozoic—early Neo-proterozoic Eastern Ghat Province—Rayner Province of India and East Antarctica. Other Proterozoic orogens are characterized by CW looping metamorphic $P$–$T$–$t$ paths and extensive granite magmatism sourced from metasedimentary and plutonic crust (syn-orogenic) and LM (post-orogenic); these orogens have similarities with two-sided hot collision in the geodynamic models. Examples of this second style of Proterozoic orogenesis include the Paleoproterozoic orogens in the southwest of the American continent (Mojave, Yavapai and Mazatzal provinces), the Paleoproterozoic Svecofennides of southern Finland, and the late Mesoproterozoic Namaqua orogen in Namibia. Finally, there are some Proterozoic orogens where the collision style was similar to large Phanerozoic orogens, but without generating HP-UHP metamorphic conditions along the future suture, perhaps identifying locations where contemporary ambient upper mantle temperature was cooler. Examples of this style of orogenesis in the Proterozoic include the Paleoproterozoic Trans—Hudson orogen of North America and the Mesoproterozoic Grenville orogen of North America.

4. Concluding remarks

The data used in this study of secular change in the $P$–$T$ conditions retrieved from metamorphic belts is taken from the literature, which was culled to 500 articles that were judged to provide robust estimates of $P$–$T$ conditions and reliable age information. These articles are strongly skewed towards the twenty-first century, with 23% dating from 2010, 45% dating from 2010 to 2001, 22% dating from 2000 to 1991 and only 10% dating from before 1991. Furthermore, the dataset used in the present study is approximately 50% larger than the one used in the first iteration 8 years ago (214 $P$–$T$–$t$ data in this study versus 144 $P$–$T$–$t$ data used in the study of Brown, 2006, 2007a). This is largely a reflection of the wider application of the advances discussed in the early part of this review, particularly those relating to thermodynamic modeling of phase equilibria and the combination of in situ geochronology with trace element geochemistry of accessory minerals, which have led to many more precise estimates of $P$–$T$ and a significant increase in the number of age determinations that are linked to these estimates. Consequently, this type of data compilation and analysis could not have been done reliably 30 years ago.

The trends of secular change in metamorphism identified by Brown (2006, 2007a,b) are confirmed by this study. Superimposed on the secular decrease in ambient upper mantle temperature from the late Archean to the present are two changes in the style of metamorphism that are interpreted to record differences in geo-dynamic regime. The first change in the style of metamorphism is interpreted to register the transition from a stagnant lid to subduction and a plate tectonics geodynamic regime that occurred during the Neoarchean (Sizova et al., 2010). This transition is reflected in changes in many different elements of the Earth system during this time (Reddy and Evans, 2009; Condie and O’Neill, 2010; Bradley, 2011; Keller and Schoene, 2012; Condie and Kröner, 2013; Young, 2013), in addition to the metamorphic record (Brown, 2006, 2007a,b). The second change in the style of metamorphism is interpreted to register the transition from shallow to deep slab breakoff that occurred during the late Neoproterozoic (Sizova et al., 2014). This transition also is reflected in changes in many different elements of the Earth system during same period (Maruyama and Liu, 1998; Campbell and Squire, 2010; Bradley, 2011; Young, 2013), in addition to the metamorphic record (Brown, 2006, 2007a,b).

However, there remain some fundamental issues that must be resolved if we are to fully understand the evolution of the lithosphere in the Precambrian.

First, we need to determine how the heat for regional-scale UHT metamorphism was generated. Recent 1-D numerical thermal modeling implicates high internal concentrations of heat producing elements and low rates of erosion as essential contributors to power UHT metamorphism, as might occur, for example, in long-lived orogenic hinterlands (Clark et al., 2011). In contrast, ridge
subduction and slab breakoff are currently popular in the literature as mechanisms to drive UHT metamorphism, probably in association with advected heat carried by magmatic magmatism, as recently reviewed by Santosh et al. (2012). Thus, we need to determine the relative importance of enhanced radiogenic heat production and slow erosion versus ridge subduction and slab breakoff as driving forces for UHT metamorphism. In addition, alternative templates for the generation of hot orogens, such as those suggested by the 2-d numerical geodynamic modeling of Sizova et al. (2014), should be investigated in more detail.

Second, we need to determine how to interpret pressures calculated from UHP metamorphic terranes in relation to lithospheric depth. Do the recorded pressures simply reflect the lithostatic load or is there a component of the total pressure that is attributable to tectonic overpressure that was sustained for long enough to be recorded by the mineral assemblage (Moulas et al., 2015)?

Third, we need to obtain quantitative information on the P–T–t evolution of early Archean crust. For high-grade greenstone terrains, this will only be achieved once we have thermodynamic models for investigating suprasolidus conditions in a much wider range of protolith compositions than is possible at present, particularly for tonalitic and basaltic protoliths that are ubiquitous in these terrains. Work on these models is currently underway.

In spite of an incomplete geological record and limitations on our knowledge of geodynamic processes in the past, we are making progress in understanding lithospheric evolution, particularly in respect of our comprehension of the Precambrian. The way forward is clear.

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566 M. Brown / Geoscience Frontiers 5 (2014) 553–569


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