Metamorphism, plate tectonics and the supercontinent cycle

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ABSTRACT

Granulite facies ultrahigh temperature metamorphism (G-UHTM) is documented in the rock record predominantly from Neoarchean to Cambrian; G-UHTM facies series rocks may be inferred at depth in younger, particularly Cenozoic orogenic systems. The first occurrence of G-UHTM in the rock record signifies a change in geodynamics that generated transient sites of very high heat flow. Many G-UHTM belts may have developed in settings analogous to modern continental backarcs. On a warmer Earth, the cyclic formation of supercontinents and their breakup, particularly by extroversion, which involved destruction of ocean basins floored by thinner lithosphere, may have generated hotter continental backarcs than those associated with the modern Pacific rim. Medium-temperature eclogite–high-pressure granulite metamorphism (E-HPGM) also is first recognized in the Neoarchean rock record, and occurs at intervals throughout the Proterozoic and Paleozoic rock record. E-HPGM belts are complementary to G-UHTM belts, and are generally inferred to record subduction-to-collision orogenesis. Blueschists become evident in the Neoproterozoic rock record; they record the low thermal gradients associated with modern subduction. Lawsonite blueschists and eclogites (high-pressure metamorphism, HPM), and ultrahigh pressure metamorphism (UHPM) characterized by coesite (+/- lawsonite) or diamond are predominantly Phanerozoic phenomena. HPM-UHPM registers the low thermal gradients and deep subduction of continental crust during the early stage of the collision process in Phanerozoic subduction-to-collision orogens. Although perhaps counterintuitive, many HPM-UHPM belts appear to have developed by closure of small ocean basins in the process of accretion of a continental terrane during a period of supercontinent introversion (Wilson cycle ocean basin opening and closing). A duality of metamorphic belts—reflecting a duality of thermal regimes—appears in the record only since the Neoarchean Era. A duality of thermal regimes is the hallmark of modern plate tectonics and the duality of metamorphic belts is the characteristic imprint of plate tectonics in the rock record. The occurrence of both G-UHTM and E-HPGM belts since the Neoarchean manifests the onset of a *'Proterozoic plate tectonics regime'*, although the style of tectonics likely involved differences. The *'Proterozoic plate tectonics regime'* evolved during a Neoproterozoic transition to the *'modern plate tectonics regime'* characterized by colder subduction and subduction of continental crust deep into the mantle and its (partial) return from depths of up to 300 km, as chronicled by the appearance of HPM-UHPM in the rock record. The age distribution of metamorphic belts that record extreme conditions of metamorphism is not uniform, and metamorphism occurs in periods that correspond to amalgamation of continental lithosphere into supercratons (e.g. Superia/Sclavia) or supercontinents (e.g. Nuna (Columbia), Rodinia, Gondwana and Pangea).

Keywords: continental backarcs, metamorphism, plate tectonics, subduction, supercontinents

INTRODUCTION

It is four decades since the plate tectonics revolution in the late 1960s introduced a robust paradigm within which to understand the Cenozoic-Mesozoic tectonics of the lithosphere. Since that time there has been much debate about when Earth might have adopted a plate tectonics mode of behavior. The first record of blueschist facies series metamorphism occurs in the Neoproterozoic, and this is commonly taken as one indicator of the beginning of the modern style of subduction (Stern, 2005). However, due to secular cooling, Brown (2006a, b) asked whether plate tectonics on a warmer Earth might have left a different imprint in the older rock record. To answer this question, it is necessary to establish the distinctive geological characteristics of plate tectonics that might be preserved or imprinted in the rock record. Brown (2006a, b) argues that the definitive hallmark of plate tectonics is a duality of thermal environments at convergent plate margins, and that the imprint of this hallmark will be found in the rock record as the contemporary occurrence of contrasting types of metamorphism. In effect, the metamorphic rock record will tell us when plate tectonics began on Earth.

Metamorphic rocks record evidence of the pressure and temperature evolution of the crust with time (the P–T–t path), reflecting the thermal environment(s) under which the mineral assemblage(s) crystallized and/or equilibrated (e.g. Brown, 1993, 2001). Different thermal environments characterized by different dT/dP register different metamorphic imprints in the geologic record. Thus, by characterizing the types of metamorphic facies series and the types of extreme metamorphism that are produced in the different thermal environments associated with convergent margins on modern Earth (<200 Ma) and by considering patterns of metamorphism in earlier Phanerozoic (>200 Ma) orogenic systems where plate tectonics reasonably may be inferred to have controlled heat loss from Earth, we may extend this analysis with some confidence to the Precambrian rock record to identify thermal environments earlier in Earth history. On this basis, Brown (2006a, b) suggests that the imprint of plate tectonics is evident in the rock record since at least the Neoarchean.

In this paper, I first discuss what we mean by plate tectonics after forty years of refinement of the original paradigm. Then, after definition of types of extreme metamorphism, I review sites of contemporary metamorphism (<200 Ma) and consider metamorphism in relation to different orogenic systems during the earlier part of the Phanerozoic rock record (>200 Ma) before extending this analysis back throughout Earth history. I discuss the implications of these data in relation to the question of what is the hallmark of plate tectonics and when did this imprint first appear in the rock record. Finally, I evaluate the age distribution of extreme metamorphism in the rock record and consider how this relates to the supercontinent cycle.

WHAT DO WE MEAN BY PLATE TECTONICS?

The concept of plates was introduced by Tuzo-Wilson (1965), and referred to rigid segments of Earth's lithosphere that are separated by three fundamentally different types of displacement reflecting spreading at ocean ridges, subduction at ocean trenches and transforms that connect these features to form a mosaic of about a dozen lithospheric plates. A number of implications of this hypothesis were expanded in a series of papers during 1967-1968 (McKenzie and Parker, 1967; Morgan, 1968; LePichon, 1968; Isaacs et al., 1968), and the paradigm of how the Earth's lithosphere behaves on modern Earth became known as plate tectonics.

Plate tectonics is based on the notion that crustal motions may be understood as rigid body rotations on a sphere. However, it had become clear by the early 1970s that the concept of rigid plates separated by narrow zones of high strain deformation could not be applied to the continents, where deformation is distributed over broad mobile belts. With the benefit of forty years reflection, we may consider plate tectonics as a mechanism for recycling of the mantle by creation and destruction of oceanic lithosphere, which controls heat loss from the Earth's interior; on modern Earth, plumes play a secondary role in heat removal, but plumes may have had a greater role in transferring heat from Earth's interior to space earlier in Earth history. From a modeling perspective (e.g. Lenardic et al., 2004; Hansen et al., 2006), plate tectonics is referred to as a *'mobile lid regime.'*

As identified above, an important question concerns when a plate tectonics mode began on Earth. A hotter Earth is likely to have produced thicker oceanic crust but thinner oceanic lithosphere that, as a result, may have been more buoyant but weaker than modern oceanic lithosphere (van Hunen et al., 2004). One consequence of these differences may have been an inability for more buoyant oceanic lithosphere to subduct. Indeed, early Earth may have been characterized by a different style of mantle convection in which a rigid but largely immobile layer develops at the surface of the mantle (Lenardic et al., 2004). In this tectonic mode, heat loss by vigorous convection beneath this layer may have been configured as plume-like (Hansen et al., 2006). From a modeling perspective, this type of tectonics is referred to as a *'stagnant lid regime.'*

A third behavior is an *'episodic regime'* in which the mode of convection may alternate between a stagnant lid and a mobile lid regime, or there may be a transition from one to the other as the dominant mode via an episodic regime (e.g. Hansen et al., 2006). A scenario of changing from a stagnant lid to a mobile lid mode via an episodic regime is one possible evolutionary path for Earth. Indeed, Davies (1995) has suggested a punctuated tectonic evolution for Earth in which the Archean might be characterized by periods of plate tectonics alternating with mantle overturn events; Davies hypothesized that the early rock record might register mainly the effects of the overturn events.

On modern Earth, there are two main zones of subduction into the mantle—the Alpine–Himalayan–Indonesian subduction system and the circum–Pacific subduction system—and two zones of upwelling—under Africa and under the central Pacific—which define a simple pattern of long-wavelength mantle convection. The Phanerozoic has mostly been distinguished by dispersion of the continental lithosphere with the breakup of Pangea. We know that the continental lithosphere influences mantle heat loss by sequestering heat-producing radioactive elements from the mantle and by acting as an insulator to the convecting mantle (e.g. Gurnis, 1988; Cooper et al., 2006). During periods of aggregation of continental lithosphere, we may anticipate longer wavelength convection and variation in mantle heat loss and mantle potential temperature (e.g. Grigné et al., 2004). Furthermore, there is some evidence that the presence of continental lithosphere may be a necessary condition for continuous stable plate-like behavior (e.g. Grigné and Tackley, 2005).

From a geological perspective, we may break down into several components the question of when plate tectonics began on Earth. For example, we may ask when did the lithosphere first behave as a mosaic of plates, that is, a mosaic of largely internally rigid lithosphere elements bounded by zones of generation, destruction or transform displacement, and how far back in

Earth history may we identify independent horizontal motions from different cratons? Also, we may ask when in the rock record do we first identify zones of convergence and subduction of oceanic lithosphere? However, care is required – what imprint in the rock record do we take to answer each of these questions? Furthermore, in a stagnant lid regime horizontal motions and convergence may occur due to non-rigidity, and this may generate symmetric subduction (in contrast to the asymmetric subduction of the plate tectonics paradigm) of the lower (mantle) part of oceanic lithosphere as postulated by Davies (1992). These issues take us into the realm of secular change on a cooling planet and the effect on geodynamics, which is the topic of a separate contribution, so I will now return to the imprint of metamorphism in the rock record and how this relates to when plate tectonics began.

OROGENIC SYSTEMS

There are two principal styles of orogenic system operating on Earth at the present time, and this has been the situation during at least the Cenozoic and Mesozoic Eras (Maruyama, 1997; Ernst, 2001; Liou et al., 2004). These are: (1) accretionary orogenic systems [variants have been called *'Pacific-type'* (e.g. Matsuda and Uyeda, 1971), *'Cordilleran-type'* (e.g. Coney et al., 1980) or *'Turkic-type'* (Şengör and Natal'in, 1996)], which form during ongoing plate convergence, as exemplified by the Phanerozoic evolution of the Pacific Ocean rim; and, collisional orogenic systems [sometimes called *'Himalayan-type'* (Liou et al., 2004)], in which an ocean is closed and arcs and/or continents collide, as exemplified by the Alpine–Himalayan system. Although collisional orogenic systems are superimposed on accretionary orogenic systems in *'subduction-to-collision orogenesis'*, the distinction is useful because some accretionary orogenic systems exist for hundreds of millions of years without disruption by collision whereas others have the continuity of subduction interrupted—but not necessarily terminated if a new trench is formed outboard of the collider—by subduction of ocean floor debris or an oceanic plateau or arc (e.g. Cloos, 1993; Collins, 2002, 2003). Orogens may be composed of metamorphic belts that exhibit different sequences of metamorphic facies with increasing metamorphic grade in relation to tectonic setting (e.g. Miyashiro, 1961; Zwart, 1967). However, orogens with 'paired' metamorphic belts generally result from juxtaposition of terranes with different metamorphic facies series rather than from formation in situ (Brown, 1998a, b, 2002b).

THE METAMORPHIC REALM

The metamorphic realm traditionally has been divided into facies, each represented by a group of mineral assemblages associated in space and time that are inferred to register equilibration within a limited range of P–T conditions. New discoveries in the last 25 years have led us to realize that conditions of crustal metamorphism have been more extreme than we previously imagined. Some rocks characteristic of the granulite facies record temperatures >1,000°C at pressures around 1 GPa [ultrahigh-temperature metamorphism (UHTM); e.g. Harley, 1998], whereas other eclogite facies rocks record pressures >5 GPa at temperatures of 600-1,000°C, and in one case mineralogical evidence suggests pressures of at least 10 GPa [ultrahigh-pressure metamorphism (UHPM); e.g. Chopin, 2003; Liu et al., 2005]. Also, we now recognize a transition between these two facies, referred to as medium-temperature eclogite–high-pressure granulite metamorphism

(E–HPGM; e.g. O'Brien and Rötzler, 2003). The P-T regimes for facies of extreme metamorphism are shown in Fig. 1. As I will discuss in more detail below, these different types of extreme metamorphism are generated in different plate tectonic settings on modern Earth with contrasting thermal regimes.

METAMORPHISM AT EXTREME P–T CONDITIONS—DEFINITIONS

It is only a little more than twenty-five years since we were introduced to the spectacular mineral assemblages reported from lower crustal rocks exposed on a regional scale in Antarctica that yielded estimates of peak temperature of >1000°C (e.g. Ellis, 1980; Ellis et al., 1980; Harley and Green, 1982), following earlier reports of assemblages with sapphirine + quartz (Dallwitz, 1968). Only a few years later, the exciting realm of ultrahigh-pressure metamorphism was revealed through the near simultaneous discovery of coesite in the Western Alps (Chopin, 1984) and Norway (Smith, 1984).

For the purposes of this paper, I follow the proposal by Harley (1998) to use 900°C as an arbitrary lower temperature limit for ultrahigh-temperature metamorphism (Fig. 1). Therefore, ultrahigh-temperature metamorphism is: *metamorphism of crustal rocks in which peak temperature exceeds 900°C, recognized either by robust thermobarometry or by the presence of a diagnostic mineral assemblage in an appropriate bulk composition and oxidation state, such as assemblages with orthopyroxene + sillimanite + quartz, sapphirine + quartz or spinel + quartz, generally at pressure conditions of sillimanite stability in metapelites.* Additionally, corundum + quartz has been argued to characterize ultrahigh-temperature metamorphic conditions (e.g. Mouri et al., 2003), but since this mineral pair is metastable relative to sillimanite (e.g. Harlov and Milke, 2002) this mineral pair is not included as part of the definition of UHT metamorphism herein. The current state-of-the-art procedure in robust thermobarometry for granulite metamorphism in general for a variety of bulk compositions is either the method proposed by Pattison et al. (2003) or the average P–T mode in THERMOCALC (Powell et al., 1998) with the internally-consistent thermodynamic dataset of Holland and Powell (1998, most recent update).

Also, I define ultrahigh-pressure metamorphism (UHPM) as (Fig. 1): *metamorphism of crustal rocks under conditions of pressure that exceed the stability field of quartz, recognized either by the presence of coesite or diamond, or by equivalent P–T conditions determined using robust thermobarometry.* Examples of robust thermobarometry currently in use, as recently evaluated by Hacker (2006), include the method of Ravna and Terry (2004) and the average P–T mode in THERMOCALC (Powell et al., 1998) with the internally-consistent thermodynamic dataset of Holland and Powell (1998, most recent update).

In between these two extreme types of metamorphism, the transition in pressure from granulite to high-pressure granulite facies conditions is not well defined, but it may be taken to correspond approximately to conditions of the change from sillimanite to kyanite stability in metapelitic rocks (Fig. 1). Furthermore, the field of high-pressure granulite facies metamorphism overlaps the transition from amphibolite to eclogite in basaltic rocks, which is a transition over more than 1 GPa according to chemical composition of the protolith (O'Brien and Rötzler, 2003).

CAVEATS

Metamorphic rocks record evidence of the pressure (from which we infer depth) and temperature evolution with time—the P–T path—which reflects the thermal environment under which the constituent mineral assemblage(s) equilibrated. The map distribution of peak P–T conditions provides information about the spatial variation in thermal regime(s). Our task is to piece together the changes in depth, temperature and kinematics with time from the mineralogical, microstructural, and chemical evidence preserved in the rock. This evidence may be incomplete, it may not be clear how one piece relates to another, evidence from two or more different segments of the path may overprint to create ambiguity in interpretation, and connections may be made between segments that do not correctly reflect the actual path. However, as a community during the past decade we have increased confidence in our ability correctly to interpret this fragmentary record of evidence (e.g. O'Brien, 1997, 1999; Brown, 2001, 2002a; White et al., 2002; O'Brien and Rötzler, 2003; Johnson and Brown, 2004; Baldwin et al., 2005; Powell et al., 2005); this confidence was lacking in the early days of studies of metamorphism under extreme P–T conditions (e.g. Green, 2005).

The 'age' of metamorphic events (including closure of minerals to diffusion) and pinning these dates to the depth of metamorphism is critical, since this yields rates of burial and exhumation (e.g. Hodges, 2003; Müller, 2003). When this information is combined with structural data to constrain the kinematics, the dataset may allow the particular tectonic setting in which the metamorphism occurred to be identified. Measured time scales for both thermal and burial–exhumation events have become shorter as precision in dating has increased, and as a result we now find acceptable rates for tectonic processes that would have been considered too fast only a few years ago (e.g. Hermann, 2003; Camacho et al., 2005).

The P–T–t paths recovered by this detective work on rocks are generally described as clockwise (in P–T space; CW) or counter-clockwise (CCW), although the P–T loops may vary from hairpin to open. CW paths may include a segment with significant isentropic decompression at high T, CCW paths may exhibit approximately isobaric cooling at moderate P, and portions of either type may be stepped; however, these features alone are not diagnostic of the tectonic setting. Without correct determination of the P–T–t evolution, including unraveling the effects of overprinting of multiple orogenic events (polymetamorphism) where present or suspected (e.g. Hensen and Zhou, 1995; Hensen et al., 1995), any evaluation of the imprint of metamorphism in the rock record will be potentially flawed.

Fortunately, as discussed above, the last decade has seen significant advances in our ability to interrogate rocks. Also, the wider availability of high-precision ages on accessory phases linked to specific microstructural sites and metamorphic P–T conditions, commonly through the identification of mineral assemblages preserved as micro-inclusions in the core, mantle or rim of zircon grains, has increased confidence that we know the true 'age' of peak metamorphism in metamorphic belts. Thus, we may now assess with some confidence the imprint of metamorphism through Earth history.

METAMORPHISM ON EARTH DURING THE PAST 600 Ma

Contemporary metamorphism (<200 Ma)

Metamorphism during the Cenozoic and Mesozoic Eras has taken place at: (1) divergent plate boundaries, where newly generated oceanic crust is metamorphosed following hydrothermal circulation of sea water; (2) convergent plate boundaries, where subduction or collision after a phase of subduction takes crustal rocks deep into the mantle during orogenesis; and, (3) at plate boundaries that involve mainly lateral displacement, although these also may involve transpression or transtension.

Rocks of the blueschist–eclogite facies series [lower temperature–higher pressure (LT–HP) metamorphism] are created in the subduction zone. We know this from samples of glaucophane schist entrained in serpentine mud volcanoes in the Mariana forearc (e.g. Shipboard Scientific Party, 2002). Incipient blueschists in the Mariana forearc with the assemblage lawsonite-pumpellyite-hematite yield estimated T of 150-250°C at P of 0.5-0.6 GPa (Maekawa et al., 1993) and dT/dP of 300-400°C/GPa (approximately equivalent to 9-11°C/km). These results are consistent with thermal models for subduction zones (e.g. Hacker et al., 2003) and thermal gradients derived from Eocene low-grade blueschists facies series rocks (e.g. Potel et al., 2006).

Ultrahigh-pressure metamorphic rocks—where the minimum P must exceed ~2.7 GPa at moderate T for coesite to occur—register deep subduction and exhumation of continental lithosphere during the early stage of collision. We know this from samples of coesite-bearing eclogites of Eocene age at Tso Morari in the frontal Himalayas (Sachan et al., 2004; Leech et al., 2005), and Pliocene age coesite eclogites in Eastern Papua New Guinea (Baldwin et al., 2004, 2005) that have been exposed exceptionally rapidly by rifting. Furthermore, Searle et al. (2001) have argued that deep earthquakes of the Hindu Kush seismic zone represent a tracer of contemporary coesite and diamond formation in ultrahigh-pressure eclogite facies metamorphism occurring at the present day.

Lawsonite blueschists and lawsonite eclogites (HP-UHPM) record colder subduction than typical blueschist–eclogite–ultrahigh-pressure facies series rocks. We know this from thermodynamic modeling of Eocene lawsonite blueschists and lawsonite eclogites exhumed to Earth's surface in the Pam Peninsular, New Caledonia (Clarke et al., 1997, 2006), and xenoliths of lawsonite eclogite of Eocene age inferred to record cold subduction to UHPM conditions in excess of 3.5 GPa brought up in Tertiary kimberlite pipes at Garnet ridge in the Colorado Plateau, USA (Usui et al., 2003, 2006).

Rocks of the granulite facies series [higher temperature–lower pressure (HT–LP) metamorphism] are scavenged from the deeper parts of Mesozoic oceanic plateaus (e.g. Gregoire et al., 1994; Shafer et al., 2005) and occur in exposed middle-to-lower crust of young continental arcs (e.g. Lucassen and Franz, 1996). Although the interpretation is contentious, metapelitic xenoliths retrieved from Neogene volcanoes in central Mexico (Hayob et al., 1989, 1990) and at El Joyazo in south-east Spain (Cesare and Gomez-Pugnaire, 2001) have been argued to record evidence of Cenozoic G-UHTM, and evidence of melt-related processes in lower crustal garnet granulite xenoliths from Kilbourne Hole, Rio Grande Rift, suggests contemporary G-UHTM in the lower crust of rifts (Scherer et al., 1997).

Modern continental backarcs are hot with uniformly thin and weak lithosphere over considerable widths (Hyndman et al., 2005). They represent an inevitable locus of deformation and an environment suitable for the generation of granulite metamorphism, particularly during terrane accretion and/or subduction-to-collision orogenesis. Thus, we may infer that granulites are being generated today under orogenic plateaus, such as Tibet and the Altiplano, based on the inference of mica breakdown melting derived from interpretations of multiple geophysical data sets (e.g. Nelson et al., 1996; Schilling and Partzsch, 2001; Unsworth et al., 2005).

Additionally, heating by viscous dissipation may be important in some subduction-to-collision orogenic systems, particularly in the early stages of subduction. Although viscous dissipation is a mechanism potentially to generate heat (e.g. Kincaid and Silver, 1996; Stüwe, 1998; Leloup et al., 1999; Burg and Gerya, 2005), it requires initially strong lithosphere (e.g. a differential stress of 100-300 MPa); England and Houseman (1988) suggest a differential stress of 100-200 MPa is necessary to generate a plateau, consistent with this requirement. The strong positive correlation between overall intensity of viscous heating in crustal rocks and the instantaneous convergence rate suggests that a significant contribution (>0.1 μW m-3) of viscous heating into the crustal heat balance may be expected when the convergence rate exceeds 1 cm yr-1, particularly if the lower crust is strong (Burg and Gerya, 2005). Therefore, heating by viscous dissipation may become a dominant heat source in the early stage of subduction-to-collisional orogenesis if convergence rates are rapid (e.g. the Himalayas).

We may conclude that contemporary metamorphism on Earth occurs in two very different thermal environments. The first environment is characterized by low dT/dP, mirrored by lower temperature–higher pressure metamorphism, corresponding to the subduction zone environment; this may be modified by and/or terminated by continental subduction and terminal suturing. The second environment is characterized by high dT/dP, registered by higher temperatures and lower pressures of metamorphism, corresponding to the continental arc–continental backarc environment, which will be inverted during collision; furthermore, a thick hot orogen may become modified by subsequent collapse and/or extension (e.g. Beaumont et al., 2004). This would be evidenced in rocks that experienced a similar history by a high-T decompression segment in a clockwise P-T-t path (cf. Harley, 2004).

The modern plate tectonics regime is characterized by a duality of thermal environments in which two principal types of regional-scale metamorphic belts are being formed. I consider this duality to be the hallmark of plate tectonics. Thus, the characteristic imprint of plate tectonics in the rock record will be the broadly contemporaneous occurrence of two contrasting types of metamorphism reflecting this duality of thermal environments (Brown, 2006b).

Metamorphism in the earlier Phanerozoic record (>200 Ma)

The oceanic lithosphere is <200 Ma old. A plethora of papers since the 1968 plate tectonics revolution suggests that we may reasonably extrapolate back through one or two ocean lithosphere cycles. Thus, we may expect similar tectonics since the last stages of Gondwana amalgamation at the dawn of the Phanerozoic Eon. Prior to this time the extrapolation becomes more contentious, particularly for rocks older than Neoproterozoic (Stern, 2005).

Accretionary orogenic systems may be subdivided into two types: extensional-contractional accretionary orogenic systems and terrane accretion orogenic systems. Metamorphism in extensional–contractional accretionary orogenic systems, such as the Lachlan Belt of Australia, is dominated by HT–LP metamorphism, with close-to-isobaric clockwise or counter-clockwise hairpin P–T–t paths. G-UHTM, E-HPGM and UHPM rocks are absent at outcrop, although blueschists may occur sporadically early in the orogenic cycle. Short-lived phases of orogenesis probably relate to interruptions in the continuity of subduction by topographic features on the ocean plate, particularly oceanic plateaus (Collins, 2002).

In terrane accretion orogenic systems, allochthonous elements were accreted to continental margins in convergent systems that involved oblique relative plate motion vectors (e.g. Johnston, 2001). Paired metamorphic belts are characteristic due to orogen-parallel terrane migration and juxtaposition by accretion of contemporary belts of contrasting type (Brown, 1998a, b, 2002b), with a LT–HP metamorphic belt outboard and a HT–LP metamorphic belt inboard, exemplified by the Mesozoic metamorphic belts of Japan (Miyashiro, 1961) and the Paleozoic–Mesozoic metamorphic belts of the western U.S. (Patrick and Day, 1995). In some systems an important additional feature was ridge subduction, which is reflected in both the pattern of HT–LP metamorphism and the associated magmatism (Brown, 1998b). Granulites may occur at the highest grade of metamorphism in the HT–LP belt, but UHPM rocks generally appear to be absent in the outboard LT–HP belt.

In subduction-to-collision orogenic systems, subduction and initial collision may generate HPM-UHPM of continental crust as it is being subducted and until the subduction zone is choked. Subduction of continental crust also inhibits generation of calc-alkaline magmas, because the high-pressure stability of micas in this thermal regime inhibits liberation of the volatile phase necessary to flux melting (Ernst, 2005). It is not yet clear whether it is this factor that enables preservation and exhumation of UHPM rocks and their return back to normal crustal depths. Deformation of the continental backarc may generate clockwise or counter-clockwise metamorphic P–T–t paths in granulite facies series rocks, including G-UHTM rocks, according to the mechanisms and relative rates of thickening and thinning versus heat transfer by conduction and advection. The European Variscides represent a classic example of a subduction-to-collision orogenic system (O'Brien and Rötzler, 2003).

Each of these older Phanerozoic orogenic systems preserves metamorphic belts with contrasting metamorphic facies series that may be inferred to record contrasting thermal regimes. These examples are consistent with the premise that plate tectonics may be extrapolated back through at least two ocean lithosphere cycles (to ca. 600 Ma); they also highlight the variations in imprint in the rock record generated by the complex interactions and diversity of processes along plate boundary zones.

METAMORPHIC REGIMES THROUGH EARTH HISTORY

Metamorphic belts are classified according to the characteristic metamorphic facies series registered by the belt into three types (Brown, 2006a, b), as follows: HPM-UHPM, characterized by lawsonite blueschist to lawsonite eclogite facies series rocks and blueschist to eclogite to ultrahigh-pressure facies series rocks, where T is that registered at maximum P; E-HPGM, characterized by facies series that reach peak P-T in the high-pressure granulite facies (O'Brien and Rötzler, 2003), where maximum P and T generally are achieved approximately contemporaneously; and, G-UHTM, characterized by granulite facies series rocks that may reach ultrahigh-temperature metamorphic conditions, where P is that registered at maximum T. Figure 2 is a plot of metamorphic belts classified by type (data from Brown, 2006a). Different apparent thermal gradients are implied by each type of metamorphism, as shown in Fig. 3. These are: HPM-UHPM metamorphism, which has apparent thermal gradients of 150–350°C/GPa (approximately equivalent to 4–10°C/km); E-HPGM, which has apparent thermal gradients of

350–750°C/GPa (10–20°C/km); and, G-UHTM, which has apparent thermal gradients >>750°C/GPa (>>20°C/km). The gradients are inferred to reflect different tectonic settings.

The distribution of these types of metamorphism throughout Earth history, based on the age of 'peak' metamorphism as defined above, is displayed in Fig. 4 (data from Brown, 2006a). Changes in the metamorphic record broadly coincide with the transitions from the Archean to Proterozoic and Proterozoic to Phanerozoic Eons (Fig. 4), and imply a different style of tectonics in the Archean Eon in comparison with the Proterozoic Eon and in the Proterozoic Eon in comparison with the Phanerozoic Eon. Overall, the restricted time span of different types of metamorphism through Earth history and the periods of metamorphic quiescence during the Proterozoic Eon suggest a link with the supercontinent cycle (Fig. 5) and major events in the mantle, supporting a postulate of discontinuous geodynamic evolution of Earth. This issue is discussed in more detail below.

Caveats

There are several caveats about possible bias in this record. It is commonly argued that going back through time increases loss of information by erosion of the older record. However, the data in Fig. 2 plot in particular periods, and there is a clear distinction between the Archean, where the data are minimal, and the Proterozoic and Phanerozoic, where the data are abundant, and between the pre-Neoproterozoic, where UHPM does not occur, and Neoproterozoic and younger belts, where UHPM is common. These observations are inconsistent with a progressively degraded record with increasing age. Extrapolation back in time also raises questions about partial-to-complete overprinting by younger events, which is a concern in any metamorphic study. However, our ability to recognize the effects of overprinting and to 'see through' them has improved significantly with better technical capabilities for rapid chemical mapping of mineral grains and high spatial resolution in situ geochronology. Overprinting has been avoided in compiling the dataset used for this analysis (Brown, 2006a). Finally, as discussed earlier, it is likely that some Phanerozoic G-UHTM rocks have not yet been exposed at Earth's surface, leading to bias in the younger part of the record.

Although patterns based on first occurrences may be challenged by new discoveries, the analysis below provides a set of compelling first-order observations. Based on the age distribution of metamorphic belts that record extreme P-T conditions, I argue that the modern era of cold subduction began in the Neoproterozoic, as registered by the occurrence of HPM-UHPM, but that cold subduction alone is not the hallmark of plate tectonics. In contrast, G-UHTM and E-HPGM are present in the exposed rock record back to the Neoarchaean, registering a duality of thermal regimes, which I argue represents the hallmark of plate tectonics. Thus, evidence of plate tectonics is present in the Neoarchean rock record, and this may manifest the beginning a global plate tectonics mode on Earth.

Granulite facies ultrahigh-temperature metamorphism (G-UHTM)

G-UHTM is dominantly a Neoarchean to Cambrian phenomenon documented during four distinct periods in Earth history (Fig. 4): the Neoarchean (e.g. Kaapvaal Craton, southern Africa, at ca. 2.72 Ga; and the Napier Complex, East Antarctica and the Andriamena Unit, Madagascar, at ca. 2.50-2.46 Ga); the Orosirian (e.g. the Hoggar, northern Africa, and the Lewisian Complex, Scotland, at ca. 2.0-1.88 Ga); the Ectasian to Stenian (e.g. the Eastern Ghats Belt, India and the

Mollendo-Camana Block, Peru, at ca. 1.4-1.0 Ga); and, the Ediacaran to Lower Cambrian (e.g. the Brasiliano and Pan-African belts of the southern continents at ca. 630-510 Ma). There is only one confirmed example of G-UHTM >>2.5Ga in age, which is the Voronezh Crystalline Massif of Sarmetia (ca. 3.2 Ga in age). G-UHTM is rare in the exposed post-Cambrian rock record, but may be inferred for the lower crust of large Cenozoic orogenic systems as discussed above. In the Phanerozoic Eon, parts of the European Variscides evolve from E-HPGM to G-UHTM, but there is only one confirmed example of G-UHTM younger than Paleozoic exposed at Earth's surface, which is the Gruf Unit in the central Alps (ca. 33 Ma in age). The four main periods of G-UHTM show a remarkable coincidence with the first formation of supercratons (Superior/Sclavia) and the supercontinent cycle (Nuna/Columbia, Rodinia and Gondwana).

Medium-temperature eclogite – high-pressure granulite metamorphism (E-HPGM)

There are about fifty documented terranes of this type, with approximately equal numbers of Proterozoic and Phanerozoic age, although the Phanerozoic examples are dominated by various segments of the European Variscides. About one-third of recognized high-pressure granulite belts have evolved from medium-temperature eclogite facies assemblages, and about one-quarter of the remainder evolve into granulite facies ultrahigh-temperature metamorphic conditions, both apparently without distinction between Proterozoic and Phanerozoic examples. Compositional zoning of minerals, and reaction coronas, symplectites and kelyphites commonly develop as multivariant reactions are crossed during this evolution. Periods of E-HPGM coincide with the first formation of supercratons (Superior/Sclavia) and the Nuna, Rodinia and Gondwana supercontinent cycles, and also occur in the Lower Paleozoic Caledonides and Variscides. E-HPGM is rare in the exposed post-Carboniferous rock record, but this may reflect lack of complete exhumation of young examples.

Blueschist (BS) and high-pressure to ultrahigh-pressure metamorphism (HPM-UHPM)

BS belts appear for the first time in the Neoproterozoic Era, where two definite examples, the Aksu Group, western China, and inliers of the Anti-Atlas belt in central Morocco, both are overlain by sediments of Ediacaran age (Maruyama et al., 1996). Other possible Neoproterozoic blueschists are those in the suture between the Yangtze and Cathaysia blocks of south China, within the Delhi Fold Belt of Rajasthan, India and in Anglesey, Wales. BS rocks may be associated with low-temperature eclogites (Maruyama et al., 1996).

HPM-UHPM has been recognized during multiple periods in Earth history: the Edicaran (e.g. Minas Gerais (?), Brazil; Mali, western Africa); the Cambrian (e.g. Kokchetav Massif, northern Kazakhstan; south-west Altyn Tagh, west China); the Ordovician-Silurian (e.g. Tromsø Nappe, Norway; Dulan Belt, North Qaidam, west China); the Devonian (e.g. western Gneiss Region, Norway, north-eastern Greenland; Maksyutov Complex, southern Urals; Saxonian Erzgebirge, Germany; Bohemian Massif, central Europe; and, the French Massif Central); the Permo-Triassic (e.g. Dabie-Sulu belt, east-central China; western Tianshan (?), west China); the Lower Cretaceous (e.g. central Indonesia, SW Sulawesi); through the Cenozoic Era (e.g. Dora Maira Massif, Italy; Zermatt-Saas Zone, western Alps; Tso-Morari Complex, India; Kaghan Valley, Pakistan); and, finally Earth's youngest UHPM rocks of Miocene-Pliocene age (Fergusson Island, eastern Papua New Guinea). P–T–t paths from UHPM terranes vary from hairpins, in which the exhumation path is close to the burial path in terms of dT/dP, and in the range 150-350°C/GPa (4-10°C/km), implying syn-subduction exhumation, to clockwise paths with isentropic decompression segments down to Barrovian-type thermal regimes of 350-750°C/GPa (10-20°C/km), consistent with thermal relaxation after initial fast exhumation to Moho depths.

Archean metamorphism

The Eoarchean through Mesoarchean rock record, and particularly the greenstone belts, generally records P–T conditions characteristic of low-to-moderate-P–moderate-to-high-T metamorphic facies. Examples include the ca. 3.75 Ga (?) metamorphism of southern West Greenland [the Isua Supracrustal Belt (Hayashi et al., 2000; P-T of 5-7 kbar, 500-550°C), and the Ameralik relict granulite facies assemblages (data of Griffin et al., 1980; av. P-T (THERMOCALC) of ~11 kbar, ~900°C)] and the ca. 3.23 Ga metamorphism of the Barberton and related greenstone belts of South Africa [Onverwacht Group greenstone remnants (Dziggel et al., 2002; P-T of 8-11 kbar, 650-700°C), southern Barberton greenstone belt (Diener et al., 2005, 2006; P-T of 9-12 kbar, 650-700°C), and the Inyoni Shear Zone (Moyen et al., 2006; P-T of 12-15 kbar, 600-650°C)]. Apparent thermal gradients were 500-800°C/GPa (~14-23°C/km), which is consistent with thinner lithosphere during most of the Archean Eon.

Blueschists are not documented and there is no metamorphic imprint of subduction of continental crust to mantle conditions and return to crustal depths (although this could be interpreted to mean that the subducted continental crust was not returned prior to the Neoproterozoic). However, the chemistry of Mesoarchean-to-Neoarchean eclogite xenoliths and the chemistry and paragenesis of Neoarchean diamonds in kimberlites within cratons suggest some process associated with supercraton formation at convergent margins was operating to take basalt and other supracrustal material into the mantle by the Neoarchean Era (e.g. Jacob, 2004; Shirey et al., 2004).

Although granulites are common throughout the Archean Eon, extreme thermal conditions typical of G-UHTM are not generally registered before the Neoarchean. This is a surprise, since we might expect the higher abundance of heat-producing elements would have led to higher crustal heat production and hotter orogens. We may infer from this that *'orogenic heat'* is related to *'mantle heat'*. Thus, it is the thermal regime and thermal environment of the tectonic setting that generates G–UHTM conditions that is the primary control in generating orogenic heat and not simple thickening of continental rocks in orogenic belts and radioactive decay.

An unusual feature of Archean granulite-amphibolite facies terranes is that the granulite-amphibolite facies isograd commonly is a retrograde feature (e.g. McGregor and Friend, 1997). This requires the widespread availability of an H_2O-rich fluid phase (volatile phase and/or melt phase) during the waning stages of the metamorphic evolution, and together with the suggestion that low-temperature water-saturated melting may be common early in Earth history (Watson and Harrison, 2005), may suggest a tectonic setting and/or process particular to this Eon although such a unique interpretation of the data recently has been questioned by Coogan and Hinton (2006).

INFERENCES AND IMPLICATIONS

The rock record indicates a dramatic change in the thermal environments of crustal metamorphism through the Neoarchean to the Archean-to-Proterozoic transition, which may reflect the switch to a '*Proterozoic plate tectonics regime*.' This premise is consistent with aggregation of the crust into progressively larger units to form supercratons (Bleeker, 2003), perhaps indicating a change in the pattern of mantle convection during the transition to the Proterozoic Eon. This transition resulted in stabilized lithosphere in which cratons form the cores of continents that subsequently grew dominantly by marginal accretion. Furthermore, the transition coincides with the first occurrence of ophiolite-like complexes in Proterozoic suture zones (Moores, 2002) and with the increase in $\delta^{18}O$ of magmas through the Paleoproterozoic, which Valley et al. (2005) argue may reflect maturation of the crust, the beginning of recycling of supracrustal rocks and their increasing involvement in magma genesis via subduction. Although the style of Proterozoic subduction remains cryptic, the change in tectonic regime whereby interactions between discrete lithospheric plates generated tectonic settings with contrasting thermal regimes was a landmark event in Earth history.

The concentration of granulite facies ultrahigh-temperature metamorphic belts in the interval Neoarchean to Cambrian (Fig. 4), and their broad correlation in time with amalgamation of continental crust into supercratons and supercontinents (Fig. 5), suggests that ultrahigh-temperature metamorphism, which commonly records tight clockwise P–T paths, may record closure and thickening of continental backarcs with their characteristic high heat flow (cf. Currie and Hyndman, 2005; Hyndman et al., 2005). Furthermore, continental backarcs seem to be more commonly developed around the margins of larger rather than smaller ocean basins (cf. the Pacific Ocean basin), and the development of G–UHTM may be related preferentially to supercontinent breakup by extroversion. This certainly appears to be true for the aggregation of the continental crust in Gondwana, where the geological evidence for tectonic setting is more complete. However, it is less clear whether this tectonic setting is appropriate for all ultrahigh-temperature metamorphic belts, although it is an attractive solution to the heat problem.

The thermal environment of E-HPG metamorphism [350-750°C/GPa (~10-20°C/km)] and maximum pressures of metamorphism suggest that crustal rocks have been taken to mantle depths during subduction-to-collision orogenesis, although the pressures could simply reflect collision and doubling in thickness by crustal stacking following subduction of oceanic plates. However, given the contrasting metamorphic record of the Proterozoic versus the Phanerozoic, it is difficult to disagree with Stern (2005) that subduction as we know it on modern Earth did not commence before the Neoproterozoic Era.

The transition to the '*modern plate tectonics regime*' through the Neoproterozoic is one in which ocean crust became thinner whereas the lithosphere became thicker (e.g. Moores, 2002), and the anastomosing pattern of Brasiliano–Pan-African orogens that created Gondwana was replaced by semi-circumferential large-scale plate-margin orogens (e.g. the Terra Australis, Caledonian and Tethyan orogens). These features may indicate a change to a smaller number of plates, perhaps reflecting a fundamental change in the pattern of mantle convection that led to a sharp decrease in the thermal gradient of subduction zone metamorphism and facilitated deeper subduction of continental crust during the Phanerozoic. Many authors have noted the confluence of other changes to the Earth system during this transition (e.g. Maruyama and Liou, 1998, 2005).

The repeated occurrence of HPM–UHPM (with the inference of subducted continental crust) since the Ediacaran Period suggests a common process in convergent tectonic settings, and the P–T conditions coupled with high-precision ages indicate fast rates of subduction to generate the low thermal gradient (150-350°C/GPa) and rates of exhumation similar to modern plate velocities (1-2 cm yr^{-1}). These rates are typical of the modern plate tectonics regime, but it is unclear whether they are a definitive characteristic of this regime. It is pertinent to note that the size of the ocean basins that have been subducted during production of HPM–UHPM belts generally appears to have been small (in comparison with the modern Pacific Ocean basin, for example), based on recent plate tectonics models for the Phanerozoic that integrate geologic, geodynamic and geophysical data (Stampfli and Borel, 2002, 2004; Hafkenscheid et al., 2006).

RELATIONSHIP OF EXTREME METAMORPHISM TO PLATE TECTONICS AND THE SUPERCONTINENT CYCLE

In the past, the continents have amalgamated into supercontinents on several occasions with dramatic effects on both surface and deep Earth processes. The relationship between plate tectonics, the supercontinent cycle and metamorphism at extreme P–T conditions is shown in Fig. 5. Extreme metamorphic conditions and contrasts in metamorphic facies series type are registered in the rock record from at least the Neoarchean, which records the start of the Proterozoic plate tectonics regime. The transition to colder thermal environments during the Neoproterozoic records the start of the modern plate tectonics regime characterized by cold subduction.

There are five principal periods of crustal amalgamation since the Neoarchean (Bleeker, 2003), as follows: formation of the supercratons Superia and Sclavia; and, formation of the supercontinents Nuna (Columbia), Rodinia, Gondwana and Pangea. There is a clear correspondence between periods during which extreme metamorphic conditions are registered in the rock record and the amalgamation of continental lithosphere by subduction-to-collision orogenesis.

One intriguing feature that requires further study is the manner in which supercontinents break up, disperse and reassemble, and whether this has an effect on the style of orogenesis or type of metamorphism. Two end-member models for the formation of supercontinents have been proposed by Murphy and Nance (2003). In one model the supercontinent simply breaks up and reassembles along the same (internal) contacts or introverts, whereas in the alternative model the supercontinent turns ‘inside-out’ or extroverts.

During introversion, the break up and dispersal of a supercontinent is accompanied by the generation of ocean lithosphere between the drifting continental blocks. It is the closure of this newly-generated ocean basin and subduction of the young lithosphere that leads to reassembly of the continents into a new supercontinent.

As an example, consider the Paleozoic Appalachian–Caledonian and Variscan orogenic systems, which represent the classical examples of the Wilson cycle of orogenesis, where the transformation of Rodinia to Pangea involved the breakout of Laurentia from the Rodinia supercontinent and its separation from Baltica and Siberia by creation of internally-generated ocean lithosphere – the Iapetus and Rheic ocean basins. The Iapetus ocean basin was closed along the Appalachian–Caledonian orogenic system and the intervening ocean lithosphere was consumed during reassembly of Laurentia, Avalonia and Baltica in the Early Devonian forming

Laurussia (Murphy and Nance, 2005; see also Collins, 2003). UHPM rocks have not yet been identified in the southern Appalachians—although HPM rocks are present (Dilts et al., 2006) and are a characteristic feature of the final (Alleghanian) phase of orogenesis—but UHPM rocks are a relatively common feature of the Norwegian Caledonides (e.g. Carswell and Compagnoni, 2003; Liou et al., 2004). The Rheic ocean basin was closed along the Variscan orogenic system and the intervening ocean lithosphere was consumed during assembly of Pangea by amalgamation of Gondwana with Laurussia by subduction-to-collision orogenesis during the Carboniferous. UHPM rocks are also a relatively common feature of the European Variscides (e.g. Carswell and Compagnoni, 2003; Liou et al., 2004).

Closure of relatively small internally-generated ocean basins also is a feature of the long-lived process of terrane accretion that formed the Altaides (Şengör and Natal'in, 1996; Badarch et al., 2002; Windley et al., 2002; Xiao et al., 2004; see also Collins, 2003), and of the polyphase collisional orogenesis that formed the Central Orogenic Belt of China (the Tienshan–Qinling–Dabie–Sulu orogenic system; e.g. Yang et al., 2005) and the Alpides (the Alpine–Zagros–Himalayan orogenic system; e.g. Şengör, 1987; Hafkenscheid et al., 2006). HPM and/or UHPM is associated with each of these orogens suggesting a geodynamic cause related to subduction of young ocean lithosphere.

In general, continental lithosphere aggregates over cold downwellings (geoid lows) that inhibit subduction and mantle cooling (Gurnis, 1988). As a consequence, the mantle overheats and fragments the continental lithosphere under tension. This is the basic principle that leads to extroversion of a supercontinent. Thus, the supercontinent breaks up and disperses due to dissipation of the mantle thermal anomaly caused by continental lithosphere insulation (Gurnis, 1988; Grigné and Labrosse, 2001; Lenardic et al., 2005), and the next supercontinent aggregates above areas of mantle downwelling by subduction-to-collision orogenesis (e.g. Gurnis, 1988; Collins, 2003). Increased mantle temperatures cause expansion (a geoid high) in the interior of the supercontinent and the ultimate breakup is preceded by and associated with widespread magmatic activity (e.g. the breakup of Pangea in the Jurassic). In contrast to the Wilson cycle, the supercontinent extroverts and the exterior continental margins of the supercontinent at the dispersal stage become the interior orogenic belts of the next supercontinent (for example, if the Pacific Ocean basin closes sometime into the future by continued subduction of the Pacific Ocean lithosphere and collision of the Americas with Australasia).

The transformation from Rodinia to Gondwana involved the breakup, dispersal and reassembly of the continental lithosphere by collisional suturing to form the network of Braziliano and Pan-African orogenic belts (e.g. Cordani et al., 2003; Collins and Pisarevsky, 2005). The internal geometry of Rodinia changed considerably during its few hundred million years of existence. Geologic and paleomagnetic data suggest that the supercontinent consolidated at 1100 to 1000 Ma and most likely disintegrated began between 850 and 800 Ma (e.g. Torsvik, 2003; Cordani et al., 2003). Reassembly to form Gondwana occurred by destruction of parts of the complementary superocean as Rodinia progressively disintegrated. In this case, subduction transported cratonic fragments to sites of collisional suturing against continental arcs–continental backarcs that now exhibit Brasiliano–Pan-African granulite facies ultrahigh temperature metamorphism (e.g. Brown, 2006a).

The transformation of Nuna (Columbia) to Rodinia is poorly constrained, but extroversion of Nuna to form Rodinia and the G-UHTM belts that are a common feature of the Grenville-age orogens (Brown, 2006a) is a reasonable hypothesis to consider. Further back in Earth history the situation is even more speculative, but I suggest that the generation of G-

UHTM conditions by inversion of continental backarcs during formation of Nuna—the first true supercontinent—and also during amalgamation of continental cratonic nucleii into the first supercratons in the Neoarchean is a reasonable hypothesis to test.

CONCLUSIONS

I propose that the hallmark of plate tectonics is the development of two contrasting thermal regimes where plates converge—on modern Earth these are the subduction zone and the continental arc–continental backarc system, respectively. The first imprint of this hallmark in the rock record is the appearance of contemporaneous contrasting types of metamorphism, viz. E-HPGM and G-UHTM, in Neoarchean belts. I propose a common genesis for E-HPGM and G-UHTM in former subduction zones choked by collision (subduction-to-collision orogenesis) and continental arcs–continental backarcs as far back as the Neoarchean. Plate tectonics has operated on Earth since the Neoarchean. This model may be changed with new findings and acquisition of new data, and there are inevitably local details of relative age and spatial location that are obscured by generalizing at a global scale.

There is a close relationship between the age of metamorphic belts characterized by extreme metamorphism and the supercontinent cycle, which reflects the fact that two contrasting thermal regimes are generated where plates converge and are registered in the rock record by subduction-to-collision orogenesis during amalgamation of the continental lithosphere into supercontinents. However, the breakup of a supercontinent by extroversion or introversion implies fundamentally different geodynamic processes, which may reflect, for example, the pattern of mantle convection. It is possible that introversion is a uniquely Phanerozoic phenomenon. The pattern of mantle convection may have been different in the Proterozoic prior to and during the formation of Gondwana than during and after the breakup of Gondwana in the Phanerozoic. A pre-Neoarchean tectonics regime may have involved plates, but if so these were probably thin with an ultramafic crust (Davies, 2006); if contrasting thermal regimes were generated, they do not appear to be preserved in the record of metamorphism.

I suggest we investigate further the implications of the crustal recycling implied by the geochemistry of Archean and Proterozoic diamonds to determine whether they represent (scant) evidence of UHPM before the Ediacaran. This will enable us to establish whether the Proterozoic plate tectonics regime involved deep subduction of continental crust—as might be implied by the diamonds—but somehow prevented its exhumation, or whether the warmer thermal environment—implied by the widespread occurrence of E-HPG metamorphism and the apparent absence of BS and UHP metamorphism—somehow prevented deep subduction of continental crust before the Neoproterozoic Era.

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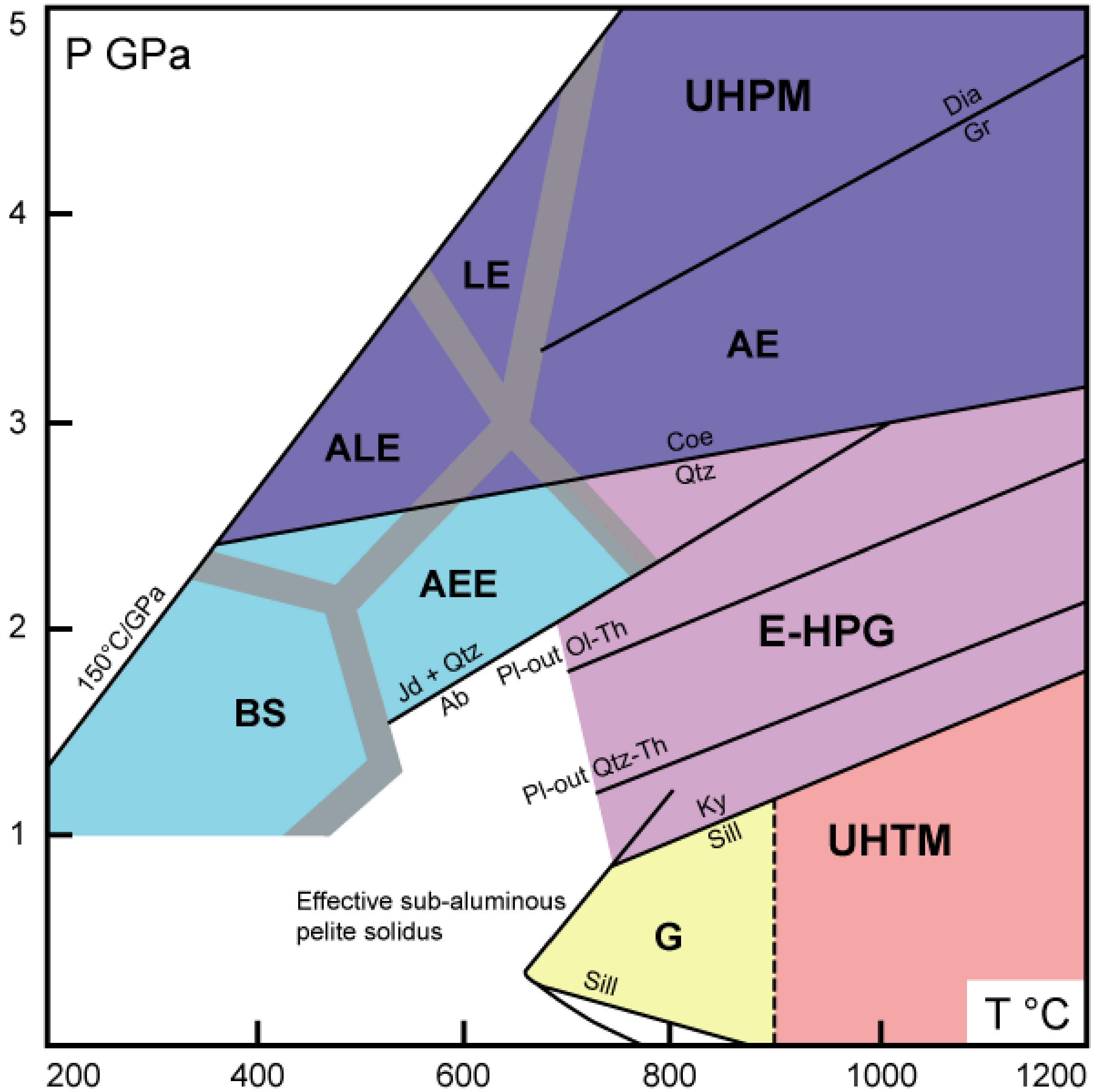

Figure 1. P–T diagram to show the P–T location of selected metamorphic facies and P–T ranges of different types of metamorphism. HP-UHP metamorphism includes the following – BS = blueschist, AEE = amphibole-epidote eclogite facies, ALE = amphibole lawsonite eclogite facies, LE - lawsonite eclogite facies, AE = amphibole eclogite facies, UHPM = ultrahigh-pressure metamorphism; E-HPG = medium temperature eclogite–high pressure granulite metamorphism; and, G = granulite facies, whereas UHTM = the ultrahigh-temperature metamorphic part of the granulite facies.

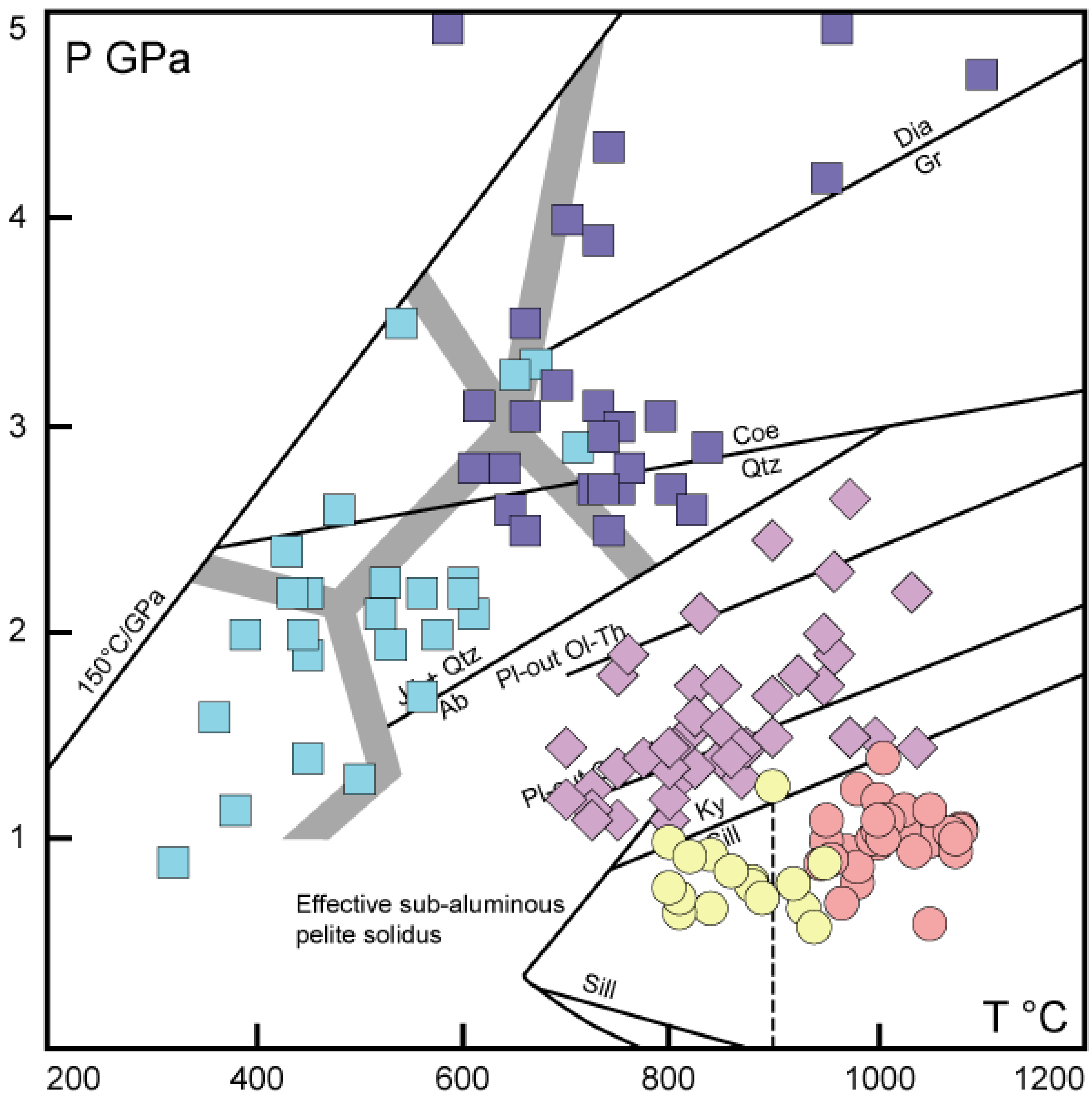

Figure 2. Metamorphic patterns based on representative 'peak' metamorphic P-T conditions of metamorphic belts in relation to the metamorphic facies in Fig. 1. Common granulite and granulite facies ultrahigh temperature metamorphism – G-UHTM belts (P at maximum T; yellow circles are common granulite belts and red circles are granulite facies ultrahigh temperature metamorphic belts – data from Tables 1 and 2 of Brown, 2006a); medium temperature eclogites–high-pressure granulites – E-HPGM belts (peak P-T; violet diamonds – data from Table 3 of Brown, 2006a); lawsonite blueschists–lawsonite eclogites and ultrahigh pressure metamorphic rocks – HPM-UHPM belts (T at maximum P; blue squares are lawsonite-bearing rocks and purple squares are ultrahigh pressure belts – data from Tables 4 and 5 of Brown, 2006a).

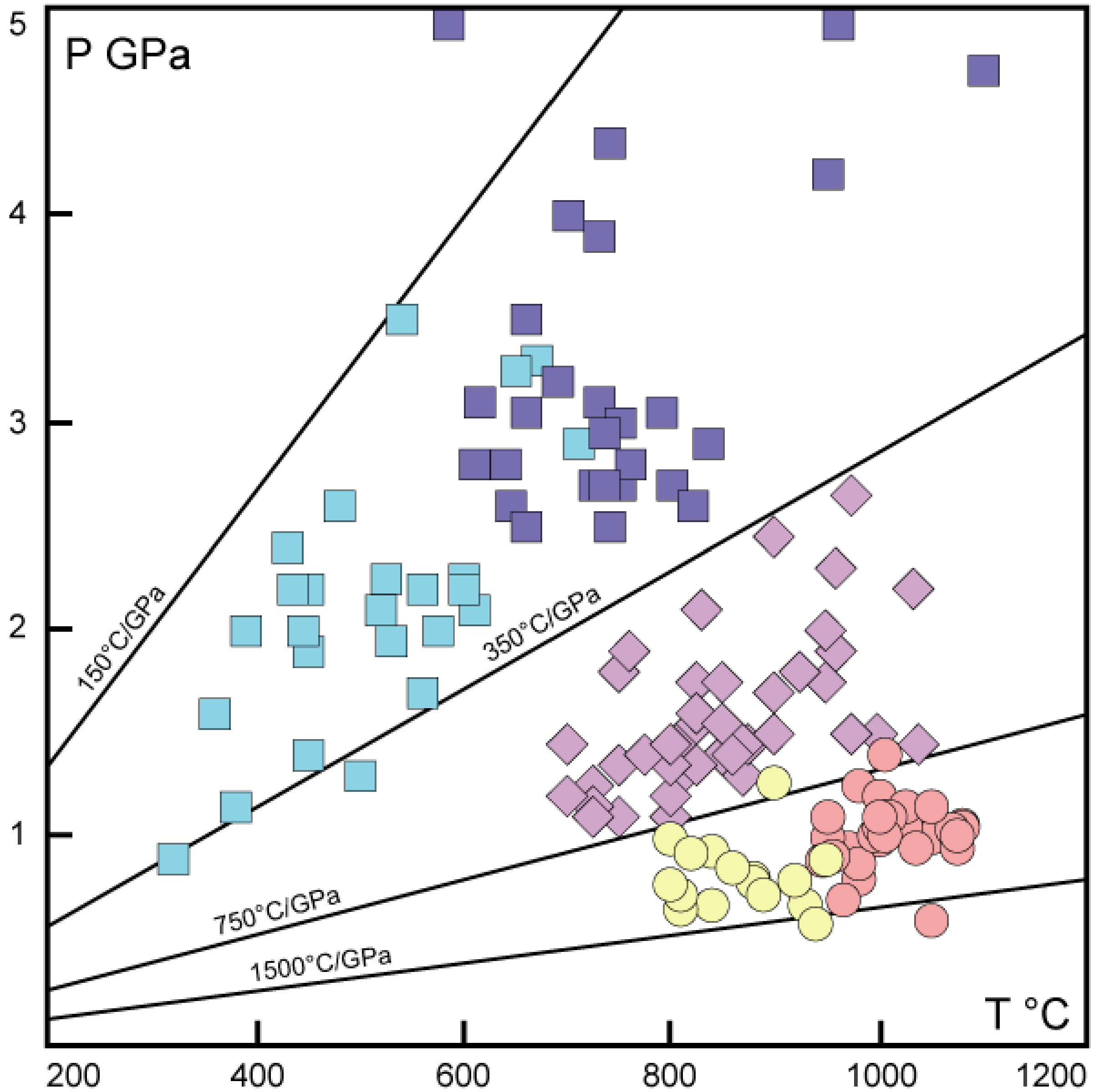

Figure 3. Metamorphic patterns by belt in relation to thermal gradients (symbols as in Fig. 2).

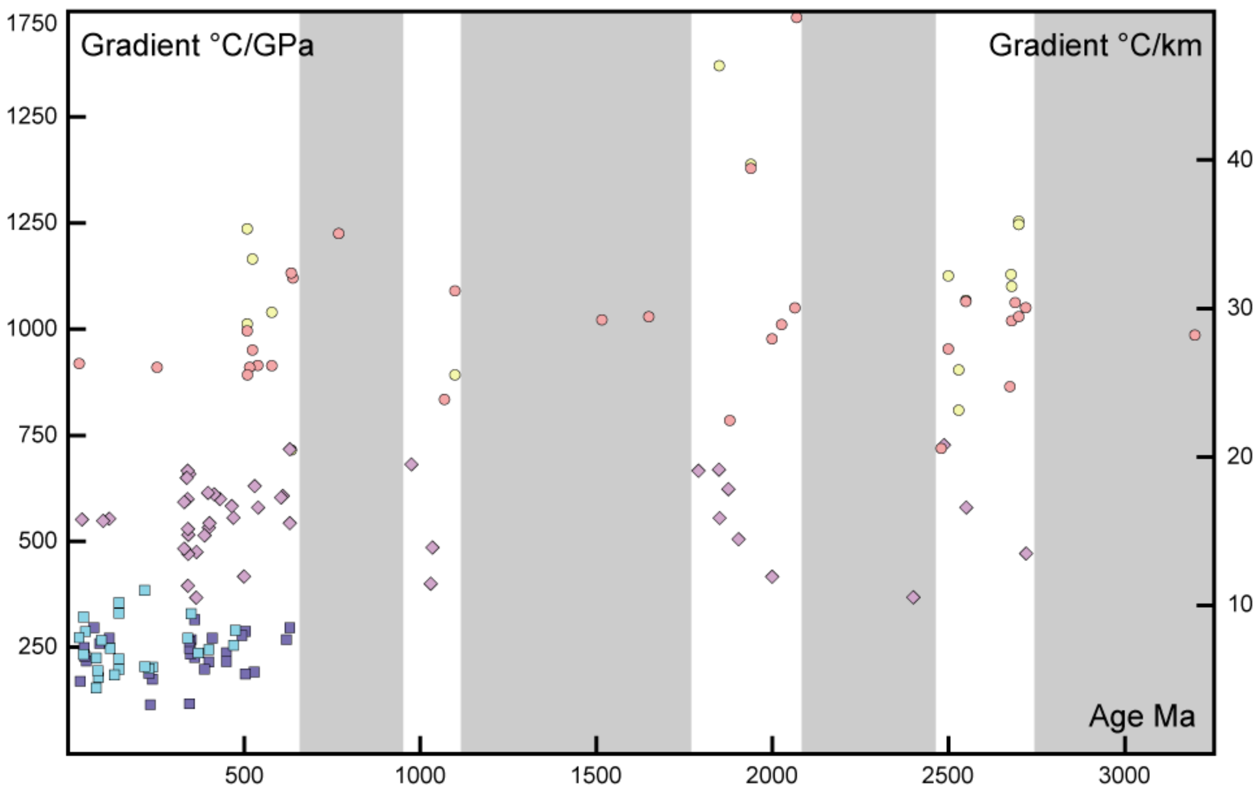

Figure 4. Plot of thermal gradient in ºC/GPa versus age in Ma for the three main types of extreme metamorphic belt – G-UHTM (circles), E-HPGM (diamonds) and HPM-UHPM (squares); an approximate conversion to ºC/km is also shown (data from Tables 1-5 of Brown, 2006a).

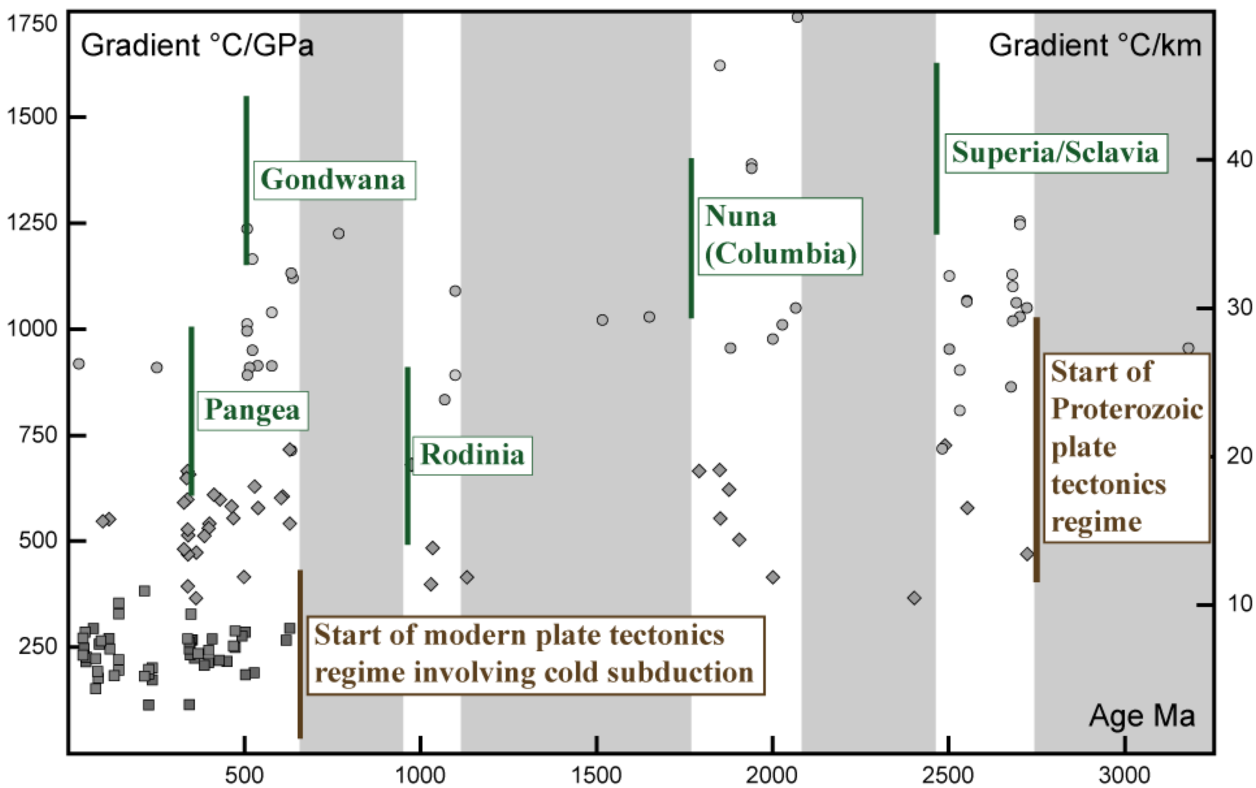

Figure 5. Timing of supercraton amalgamation (Superia/Sclavia) and supercontinent amalgamation (Nuna (Columbia), Rodinia, Gondwana and Pangea), and the start of the Proterozoic plate tectonics and modern plate tectonics regimes in relation to the thermal gradient in versus age plot from Fig. 4 for the three main types of extreme metamorphic belt.