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Metamorphic conditions in orogenic belts: A record of secular change

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ABSTRACT

In general, Archean rocks exhibit rather ordinary moderate-P–high-T facies series metamorphism; neither blueschists nor any record of deep continental subduction and return are documented. However, the abundance and scale of ultrahigh temperature (UHT) metamorphic belts from the Neoarchean to the Cambrian imply a significant change in geodynamics during the Neoarchean Era after which transient sites of high heat flow were available at intervals throughout this period of Earth evolution. Many Neoproterozoic–Cambrian UHT metamorphic belts appear to have developed in settings analogous to modern back arcs that were closed and inverted during crustal aggregation and formation of the Gondwana supercontinent. If back arcs were the general setting for UHT metamorphism, then on a hotter Earth the cyclic formation of supercratons (in the Neoarchean Era) and supercontinents (in the Proterozoic Eon) required the destruction of oceans floored by thinner lithosphere that may have generated hotter back arcs than those associated with the current destruction of the Pacific Ocean on modern Earth. The inherent weakness of the lithosphere in a hotter thermal regime inevitably localized magmatism and deformation at these sites contemporaneously with UHT metamorphism. Medium-temperature eclogites of crustal derivation and associated high-pressure granulites are also first recognized in the Neoarchean, for example within the Belomorian Mobile Belt, and they occur at intervals throughout the Proterozoic, for example in the Orosirian Usagaran Orogen and in the Grenvillian belts of the Proto-Atlantic region, and Paleozoic, for example in the circum-North Atlantic Caledonides and European Variscides. Eclogite–high pressure granulite (E-HPG) metamorphism is predominantly a Proterozoic–Paleozoic phenomenon—complementary to but sparser than UHT metamorphism to begin with, but extending further into the Paleozoic than does UHT metamorphism—that is inferred to record subduction-to-collision orogenesis?

Blueschists appear in the Neoproterozoic Era, becoming common through the Phanerozoic Eon; they record the low thermal gradients associated with modern subduction. Lawsonite-bearing blueschists and eclogites, and ultrahigh pressure (UHP) metamorphism characterized by coesite or diamond are predominantly Phanerozoic phenomena related to deep subduction within subduction-to-collision orogens. In general, UHP metamorphic belts in subduction-to-collision orogens are not associated with a contemporary magmatic arc in the hanging wall; this suggests that deep subduction of continental crust may inhibit the generation of calc-alkaline magmas, a feature that may have enabled preservation during exhumation of the mineralogical evidence for extreme pressures. However, an enigma concerning UHP metamorphism is the first evidence of deep subduction of continental crust in the rock record. At issue is the recycling implied by the geochemistry of Archean and Proterozoic diamonds (since entrained from mantle to upper crust in younger magmatic events), which requires a supracrustal component. Are these diamonds evidence of deep subduction of continental crust and an early record of UHP metamorphism, or was some

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other mechanism (e.g., delamination of underthrust lithosphere, some form of slab break-off) responsible for taking a supracrustal component deep into the mantle source?

The Archean and Proterozoic Eons were characterized by higher but decreasing mean mantle temperatures and radioactive heat production (RHP), and a thinner TBL with a shorter residence time than modern Earth. Modeling the effect of increased RHP on the thermal evolution of crust instantaneously doubled in thickness predicts that metamorphic rocks in Archean collisional orogens should have experienced maximum temperatures several hundreds of degrees Celsius higher than those recorded by metamorphic rocks in modern collisional orogens. However, there is no evidence of this – the Archean record is dominated by rather ordinary P–T conditions and crustal melting at relatively low temperatures probably fluxed by water. I argue that a duality of metamorphic belts—reflecting a duality of thermal environments—is the characteristic metamorphic imprint of plate tectonics in the rock record, and it appears only since the Neoarchean Era. Based on the occurrence of both UHT and E-HPG metamorphic belts since the Archean-to-Proterozoic transition, I suggest this transition records the onset of a *'Proterozoic plate tectonics regime'*, although the total ridge length and the number of plates undoubtedly were larger in the beginning than on modern Earth and the style of tectonics likely involved some differences. The Neoproterozoic transition to the *'modern plate tectonics regime'* registers a change to subduction of continental crust deeper into the mantle and its (partial) return from depths of up to 300 km, a change perhaps related to whole mantle convection as oceanic lithosphere became thicker with decreased thermal gradients. Although there was overlap in time and space between E-HPG and UHP metamorphism during the Paleozoic, since the Carboniferous UHP metamorphism has dominated and extreme thermal metamorphism is rare.

DEDICATION

This review is dedicated to honoring the career contributions of J.G. (Louie) Liou. Of course, this is not the end of Louie's contributions, and the timing of the Liou Symposium in his honor at Stanford University on December 10-11, 2005 was determined more by Louie's change in status to Emeritus Professor than any expectation that he will slow down in research. However, I do anticipate that the greater flexibility associated with an Emeritus faculty rank will give Louie more time to travel, and to contribute research and write papers – his research career is but partway through! An occasion like the Liou Symposium provides an opportunity for retrospection, for example, to review our current understanding of the thermal conditions and depth of burial implied by the mineralogic record of exhumed deep crust now exposed in eroded remnants of metamorphic belts, and for prospection, for example, by asking how much our view of metamorphic conditions might change within the next decade or two!

INTRODUCTION

Metamorphism is concerned with the interpretation of information recorded by the mineralogy, microstructure and chemistry of rocks and the application of this information to understand the burial and thermal history of the crust and underlying mantle. Metamorphic processes occur at scales that range from atomic (in minerals) to global (interactions between tectonic plates). On modern Earth, these processes occur at: (1) convergent plate boundaries, where subduction (e.g., the Andes) or continental collision after a phase of subduction (e.g., the Himalayas) takes crustal rocks deep into Earth's mantle during orogeny; (2) at plate boundaries that involve predominantly lateral displacement or that also may involve transpression or transtension (e.g., the San Andreas fault system and associated structures in the Transverse Ranges of the western USA, and the Kermadec Trench and the Alpine Fault of New Zealand); and, (3) at divergent plate boundaries, where newly generated oceanic crust is metamorphosed following hydrothermal circulation of sea water (ocean floor metamorphism alongside the mid-ocean ridge system). These diverse tectonic environments generate metamorphic signatures in rocks that span a range in temperature from several hundred to more than a thousand degrees Celsius and in depth from near-surface to several hundred kilometers (Brown, 2001).

Field relations, rock fabrics, mineral assemblages, reaction microstructures, and mineral and whole-rock compositional variation provide a record of orogenic evolution. Our ability to interpret the evidence recorded by metamorphic rocks is critical to understanding their history and to constraining models of orogenesis. The metamorphic cycle involving burial (increasing pressure and temperature) and exhumation (decreasing pressure and temperature) is referred to as a pressure-temperature-time-deformation (P–T–t–d) path. When viewed in a metamorphic P–T diagram, these paths may be clockwise or counter-clockwise in sense according to the relative rates of burial and heating versus cooling and exhumation.

We are concerned with two scales of interaction. At the global scale, similarities in type of metamorphism are likely to reflect geodynamic processes and changes in type with time are likely to reflect changes in geodynamic processes. At the regional scale, differences in type of metamorphism from terrane to terrane enable us to characterize the tectonic settings involved and the sequence of events that have produced a particular orogenic collage.

WHY IS THIS TOPICAL NOW?

My aim in this paper is to review the record of metamorphism back through Earth history to evaluate secular change in thermal regimes and assess the link between the metamorphic record and global tectonic regimes. One may question whether there is anything new to say in reviewing metamorphic conditions, particularly since this year is the 50th Anniversary of de Roever's landmark paper "Some differences between post-Paleozoic and older regional metamorphism" (de Roever, 1956) and the 10th Anniversary of the massive "Blueschists and eclogites of the world and their exhumation" by Maruyama, Liou and Terabayashi (Maruyama et al., 1996). Furthermore, the relationship between secular change in Earth's thermal structure as it loses heat and the metamorphic conditions recorded by exhumed and exposed metamorphic belts, and the relationship between these and global tectonic regimes have generated a succession of contributions during the past five decades. These contributions include those by de Roever, Miyashiro and Ernst in the early days of the paradigm shift in Earth Sciences with the plate tectonics revolution (de Roever, 1956, 1965; Miyashiro, 1961, 1973; Ernst, 1972), to the more recent contributions by Maruyama and Liou (Maruyama et al., 1996; Maruyama and Liou, 1998, 2005), and by Stern (2005), with only the single paper by Grambling (1981) in the intervening period. However, although the topic of metamorphic conditions and secular change has a fifty-year history, several advances have led to: (1) a dramatic increases in the maximum calculated pressures (and implied depths of burial; e.g., Chopin, 2003) and highest calculated temperatures (Harley, 2004) recorded by terrane-scale metamorphic rocks in orogens; and, (2) a higher level of confidence in the absolute values of retrieved maximum pressures and temperatures (e.g., Green et al., 2000; Powell et al., 2005), and a better understanding of the relationship between this P–T point in the tectonic evolution and retrieved age data that may relate to P–T points along the prograde, around the peak or along the retrograde segments of the P–T path (e.g., Hodges, 2003; Müller, 2003; Rubatto, 2002).

These advances include: (1) the ability to identify, analyze and date indirectly (if contained in zircon) micro-inclusions and assemblages of micro-inclusions in rock-forming minerals, such as garnet and pyroxene (e.g., Chopin, 2003), and accessory minerals, such as zircon (Liu et al., 2002, 2003; McClelland et al., 2005); (2) significant improvements in thermobarometry (e.g., Pattison et al., 2003; Ravna and Terry, 2004) and in the range of P–T space accessible using comprehensive internally-consistent thermodynamic datasets (e.g., Holland and Powell, 1998), particularly with reference to activity-composition models for minerals characteristic of extreme P–T conditions (e.g., Holland and Powell, 1996a, b, 2003; Powell and Holland, 1999;Wei and Powell, 2003, 2004, 2006; Kelsey et al., 2004; Dale et al., 2005); (3) the development of new trace-element-based thermometers (e.g., Watson et al., 2006); and, (4) improvements in the precision of U-Pb age data both by TIMS and SIMS (e.g., Baldwin et al., 2004; Collins et al., 2004), and the refinement of monazite dating by EPMA (e.g., Williams and Jercinovic, 2002; Dahl et al., 2005; Baldwin et al., 2006).

For these reasons, it is a worthwhile exercise to review types of metamorphic belt in relation to tectonic setting, compile the most recent estimates of P–T conditions and age of peak metamorphism for selected metamorphic belts, consider the implications of any change in thermal environment of metamorphism inferred from these data, and to relate any changes to possible non-linear switches in the pattern of mantle convection and the tectonics of the thermal boundary layer. It is through this process of periodic review and integration of data from different specialties in the Earth sciences that we identify unanswered questions that may direct future research.

METAMORPHISM AT EXTREME P–T CONDITIONS—DEFINITIONS

It is only a little more than twenty-five years since our community in metamorphic geology was 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 (UHT) metamorphism. Therefore, UHT 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 UHT 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 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 (UHP) metamorphism as:

> *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. 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).

THE METAMORPHIC REALM

The metamorphic realm traditionally has been divided into facies, each represented by a set of mineral assemblages repeatedly associated in space and time that reflect equilibration within a range of P–T conditions (e.g., Turner, 1968). The standard metamorphic facies are well established; ignoring very low grade and contact metamorphism, these are the blueschist, greenschist, eclogite, amphibolite and granulite facies (Fig. 1). Furthermore, the idea that orogenic belts might be characterized by different metamorphic facies series is well understood (Miyashiro, 1961; Zwart, 1967), although the concept of 'paired' metamorphic belts is a result of terrane accretion (Brown, 1998a, b, 2002b; and implicit in Miyashiro, 1961) rather than in situ generation as has sometimes been implied.

New discoveries in the last twenty-five years (discussed above) have led us to realize that conditions of crustal metamorphism have been more extreme than we previously imagined. Thus, the granulite facies now extends to temperatures above 1,000°C, and at the extreme is only limited by the liquidus for crustal rocks (UHT metamorphism; e.g., Harley, 1998), and the eclogite facies extends to pressures of at least 6 GPa, and possibly as much as 10 GPa, for crustal terranes returned to the surface, and extends to at least these pressures for xenolithic materials (UHP metamorphism; e.g., Chopin, 2003; Liu et al., 2005), but this facies does not have a defined upper pressure limit anyway. There is a transition between the two referred to as medium-temperature eclogite – high-pressure granulite (E-HPG) metamorphism (e.g., O'Brien and Rötzler, 2003). These extremes in the metamorphic realm reflect a range of thermal regimes characterized by high dT/dP to low dT/dP gradients – essentially the paleo-thermal gradients (Fig. 1).

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 secular change 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 how the metamorphic facies are distributed through Earth history and evaluate changes in thermal regime over time.

METAMORPHIC CONDITIONS DETERMINED BY THERMOBAROMETRY

One approach to the record of crustal metamorphism is to calculate P and T using an internally-consistent thermodynamic dataset applied to individual samples that have been well characterized in terms of mineralogy, microstructure and mineral chemistry (e.g., Powell and Holland, 1994). Although uncertainty associated with thermobarometry is large when calculating absolute P and T—10% or more (e.g., Worley and Powell, 2000)—when used for comparative purposes among a suite of samples, thermobarometry yields a good estimate of the range of P–T conditions for different types of metamorphism. I illustrate this point with two examples.

Figure 2 shows a plot of P–T conditions for individual samples of granulite and ultrahigh-temperature (UHT) metamorphic belts calculated by Pattison et al. (2003, their Appendix 2), using Al-solubility in Opx, corrected for retrograde Fe-Mg exchange. These data for individual samples define a limited range of P—about 1 GPa—across more than 400°C, and reflect thermal gradients with high dT/dP.

For metamorphic belts that exhibit mineralogical characteristics indicative of high-pressure (HP) and ultrahigh-pressure metamorphism (UHP), Figure 3 shows the range of P–T conditions calculated using the methodology proposed by Ravna and Terry (2004), based on Grt-Cpx-Ky-Phe-Qtz/Coe thermobarometry, and reported by these authors. Additional data are included from Hacker (2006) and Tsujimori et al. (2006b). The spread of P–T conditions recorded from HP and UHP metamorphic belts suggests a range of depths of burial are recorded by individual samples, together with some variation in thermal gradient, although in all cases the metamorphism is characterized by low values of dT/dP.

TYPES OF METAMORPHISM CLASSIFIED BY TERRANE

I follow Miyashiro (1961) in believing that it is useful in terms of understanding crustal-scale geodynamics to classify metamorphic belts by type. However, I am cognizant of the advice offered by Francis J. Turner in his Presidential Address to the Mineralogical Society of America on

November 11, 1969 *". . . much of petrological effort has been concentrated on attempts to discern, define and interpret regularly recurring patterns of . . .metamorphic petrogenesis . . . I do not wish to deny the existence of discernable order in petrological phenomena . . . but . . . every geological event . . . is unique. If we concentrate too heavily on discernment of order and pattern, may we perhaps overlook or underrate the unique quality of each . . . metamorphic episode? Possibly uniqueness may have as great a significance as conformity to pattern when we attempt to fit the phenomena of petrology into the broad framework of geology."* Accepting this caveat, for the purpose of this paper, I classify metamorphic belts into three types.

Figures 4-6 show plots of metamorphic belts classified by type [granulite–ultrahigh temperature (G-UHT) metamorphism, medium-temperature eclogite – high-pressure granulite (E-HPG) metamorphism and high-pressure to ultrahigh-pressure (HP-UHP) metamorphism], 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 (listed in Tables 1-5). Different thermal gradients are implied by the metamorphic grade evaluated by terrane for each type of metamorphism (Fig. 7); one implication of these different thermal gradients is that each type of metamorphism is likely to have a characteristic tectonic setting with which it was associated.

Granulite and ultrahigh temperature (G-UHT) metamorphism

Common granulite facies metamorphism (Fig. 4) is ubiquitous throughout Earth history, and granulites are represented widely in the Archean rock record (e.g., Harley, 1989; Pattison et al., 2003). On modern Earth, granulites of Cretaceous age occur in the deeper parts of oceanic plateaux (e.g., Gregoire et al., 1994; Shafer et al., 2005) and granulites apparently are being generated today under orogenic plateaux, such as Tibet, based on the inference of mica breakdown melting derived from interpretations of multiple geophysical datasets (e.g., Nelson et al., 1996; Unsworth et al., 2005).

In contrast, UHT metamorphism (Fig. 4; metamorphic conditions of >900°C and from ~0.5 to ~1.5 GPa) is predominantly a Proterozoic phenomenon, probably associated with crustal aggregation into supercontinents, although lower crustal UHT granulites of Cenozoic age are sampled as metapelitic xenoliths retrieved from Quaternary volcanoes in central Mexico (e.g., Hayob et al., 1989), which may relate to the Mesozoic–Cenozoic eastward migration of the Antilles arc from the Pacific to the Atlantic and the formation of the Caribbean plate. A common diagnostic mineral assemblage in metaclastic rocks is Al-rich orthopyroxene–sillimanite–quartz, but more exotic assemblages including sapphirine–quartz and spinel–quartz also are characteristic.

Common granulite and UHT metamorphic mineral assemblages imply thermal regimes with high dT/dP of >>750°C/GPa; this is approximately equivalent to the more familiar form of the thermal gradient expressed with reference to depth of >>20°C/km, assuming a density for the lithostatic load equivalent to average continental crust and no significant tectonic overpressure, both of which are reasonable assumptions in this case. P–T paths may be CW or CCW, but the limited range of pressures for common granulite and UHT metamorphism leads to an element of near-isobaric cooling from peak P–T in many cases, followed by decompression during exhumation.

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

E-HPG eclogites and granulites (Fig. 5) occur in the Neoarchean Belomorian Mobile Belt of the Fennoscandian Shield (Earth's oldest crustal omphacite-garnet rocks at ca. 2.72 Ga; Volodichev et al., 2004), in the Proterozoic Orosirian Usagaran Orogen of Tanzania (ca. 2.0 Ga; Collins et al., 2004) and in the Grenvillian belts of the Proto-Atlantic region and in the Paleozoic Caledonian and Variscan belts. E-HPG metamorphism (metamorphic conditions of ~750 to >1000°C and from ~1.5 to >2.5 GPa) is predominantly a Proterozoic–Phanerozoic phenomenon related to subduction and collision. The E-HPG field of metamorphism for rocks of quartz–tholeiitic and high-Al basaltic compositions is characterized by garnet and Ca–Tschermaks clinopyroxene-bearing assemblages, whereas metapelitic and quartzo-feldspathic rocks are characterized by garnet–kyanite–K-feldspar assemblages.

These observations translate to an environment for E-HPG metamorphism with thermal gradients of 350-750°C/GPa, which is approximately equivalent to the more familiar form of the thermal gradient expressed with reference to depth of 10-20°C/km, assuming a density for the lithostatic load equivalent to average continental crust and no significant tectonic overpressure – assumptions that may result in an underestimate of the thermal gradient in this case. The higher pressures of metamorphism and colder thermal regime of E-HPG metamorphism in comparison with common granulite and UHT metamorphism are consistent with subduction of crustal rocks into the mantle. P–T paths are more commonly CW than CCW, with peak T after peak P, which suggests that burial to maximum pressures was followed by thermal relaxation as subduction was blocked by collision and erosion or orogenic collapse facilitated exhumation.

Blueschist (BS) and high-pressure to ultrahigh-pressure (HP-UHP) metamorphism

BS metamorphic belts appear for the first time in the Neoproterozoic Era and become increasingly common through the Phanerozoic Eon; they record low- to very-low thermal gradients due to subduction at convergent plate margins during the pre-collisional accretionary phase of the tectonic cycle. BS rocks may be associated with low-temperature eclogites.

Lawsonite and inferred pseudomorphs after lawsonite are reported in low-temperature HP blueschist facies rocks world-wide (e.g., Ballèvre et al., 2003), where glaucophane–lawsonite is a lower temperature association compared to glaucophane–epidote; more rarely lawsonite occurs in low-temperature HP eclogite facies rocks found in subduction-to-collision orogens and young active convergent margins (Table 4; see also Tsujimori et al., 2006b). Lawsonite is stable over a broad range of very low-temperature – high-pressure conditions developed during subduction (Fig. 6), although high water content likely is required to stabilize lawsonite-bearing assemblages and cold, possibly rapid exhumation likely is required to preserve them. The oldest lawsonite eclogites are Ordovician in age (Table 4), which suggests that subduction zone thermal regimes evolved over a few hundred million years from cold enough to generate ordinary blueschists during the Neoproterozoic to colder still during the early Paleozoic Era. This is confirmed by the recent discovery of Fe-Mg carpholite in supracrustal rocks of the lower unit on Motalafjella, in the Svalbard Caledonides, which is by far the oldest known occurrence of this mineral that is indicative of very low thermal gradients (Agard et al., 2005).

UHP metamorphism (Fig. 6; metamorphic conditions of ~600 to >900°C and from ~2.6 to >10.0 GPa) is predominantly a Phanerozoic phenomenon related to deep subduction of continental crust and its (partial) return to normal crustal depths. Approximately twenty HP-UHP metamorphic belts have been documented within major continental plate collision belts. The record of HP-UHP metamorphism is preserved in pods and slabs of eclogite and garnet peridotite within gneisses (e.g., Cuthbert et al., 2000); where present, ophiolitic eclogites are older than continental crustal eclogites. A few UHP rocks contain minute inclusions of coesite in zircon, garnet and omphacite, and microdiamond in garnet and zircon (e.g., Chopin, 2003), although the microdiamond in UHP metamorphic rocks most likely was precipitated from a fluid phase rather than being simply recrystallized graphite (Dobrzhinetskaya et al., 2004, 2005). Recent discoveries of pyroxene exsolution in garnet and coesite exsolution in titanite from some HP-UHP metamorphic belts indicate pressures on the order of 6 GPa (e.g., Chopin, 2003), but the inference of stishovite from at least one locality in the Altyn Tagh indicates pressures >10 GPa (Liu, 2005) and suggests that under the right circumstances supracrustal rocks may be returned from depths of approximately 300 km, from around the ‘point of no return’ along the descent into the mantle.

The environment of BS and HP-UHP metamorphism suggests thermal gradients of 150-350°C/GPa; this would be approximately equivalent to 4-10°C/km, assuming a density for the lithostatic load equivalent to average continental crust and no significant tectonic overpressure, but the increases in density due to phase changes with depth means natural gradients will be lower. These very cold thermal gradients require subduction of crustal rocks into the mantle, but the younger granulite facies overprint that is common in some UHP metamorphic terranes suggests thermal relaxation as subduction was blocked by collision, and evolution of thermal gradients to values closer to Barrovian thermal gradients. However, the extreme pressures attributed to some UHP metamorphic belts may be misleading if tectonic overpressure was significant and if it was dynamically maintained (e.g., Mancktelow, 1995; Petrini and Podladchikov, 2000; Hartz et al., 2005), although tectonic overpressure remains a contentious issue (e.g., Green, 2005).

OROGENIC SYSTEMS AND METAMORPHISM

There are two principal styles of orogenic system on modern Earth (e.g., Wilson and Burke, 1972; Miyashiro, 1973; Murphy and Nance, 1991; Ernst, 2005): accretionary orogenic systems [sometimes called *'Pacific-type'* (e.g., Matsuda and Uyeda, 1971; Maruyama, 1997) or *'Cordilleran-type'* (e.g., Coney et al., 1980)], which form during ongoing plate convergence, and dominate in the period after supercontinent break-up and continent dispersal; and, collisional orogenic systems (sometimes called *'Himalayan-type'*), in which an ocean is closed, commonly during continent assembly and formation of supercontinents. Collisional orogenic systems may be superimposed on accretionary orogenic systems (e.g., Liou et al., 2004; herein termed *'subduction-to-collision orogenesis'*), although 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 minor collision due to subduction of ocean floor debris, including oceanic plateaus (e.g., Nur and Ben-Avraham, 1989; Cloos, 1993, Mann and Taira, 2004), or more moderate collision if an oceanic arc is involved. These alternatives generate the terminal and non-terminal sutures of Dewey and Kidd (1977; see also Burke et al., 1977), respectively.

Accretionary orogenic systems may be subdivided into two types. In extensional-contractional accretionary orogenic systems, continental growth is related to mantle-derived magmatism in dominantly extensional arc systems (e.g., Collins, 2002). In terrane accretion orogenic systems, allochthonous, commonly oceanic elements are accreted to continental margins and modified into continental lithosphere in arc systems that involve oblique to highly-oblique relative plate motion vectors (e.g., Monger, 1997; Cowan et al., 1997; Johnston, 2001).

In extensional-contractional accretionary orogenic systems, the metamorphic pattern is dominated by LP-HT metamorphism, with looping or ambiguous (CW or CCW) P–T–t paths. Peak metamorphic mineral growth occurred late in relation to tectonic fabrics, and there is an absence at surface outcrop of granulites (especially of UHT granulites), HP granulites and UHP metamorphic rocks, although blueschists may occur sporadically early in the orogenic cycle (e.g., Spaggiari et al., 2002a, b). Short-lived compressional-contractional phases of orogenesis probably relate to interruptions in the continuity of subduction by topographic features on the ocean plate, particularly oceanic plateaux (e.g., Nur and Ben-Avraham, 1989; Cloos, 1993; Collins, 2002). Granite magmatism is an extensive accompaniment to metamorphism. The rapid loading and unloading P–T–t paths of some Mesozoic metamorphic belts may reflect intra-arc overthickening of extended arc lithosphere (e.g., Hiroi et al., 1998; Whitney et al., 1999).

In terrane accretion orogenic systems, coeval 'paired' metamorphic belts have been argued to characterize the metamorphic pattern (Miyashiro, 1961, 1973). However, this is a false construct that results from failure to recognize orogen-parallel terrane migration and the limitations of the chronological datasets (Brown, 2002b), even though tectonic contacts between the belts were widely recognized (e.g., Miyashiro, 1961). Commonly a HP-LT (blueschist-eclogite) metamorphic belt occurs outboard and is separated from a LP-HT (andalusite-sillimanite type) metamorphic belt by a terrane boundary. In some of these systems an important additional feature of the orogenic process is ridge subduction (Brown, 1998a), which is reflected in both the pattern of LP-HT metamorphism and the associated magmatism normal to the orogen, which may have adakitic characteristics (Brown, 1998b). Extensive granite magmatism is common, but thin lithosphere suggests mantle delamination. Granulites may occur at the highest grade of metamorphism in the LP-HT belt, and are implied to occur at depth, but so far UHP metamorphic rocks appear to be rare in the outboard HP-LT belt (but see Ota et al., 2004b).

Collisional orogenic systems, as represented in the Phanerozoic geologic record, generally produce characteristic CW metamorphic P–T–t paths. Due to progress made during the past twenty years, we now know that on modern Earth initial collision between a subducting continental plate and an overriding plate may generate extreme UHP metamorphism of subducted continental crust, which may melt and if subducted past the 'point of no return' may be transported into the deep mantle (e.g., Irifune et al., 1994; Domanik and Holloway, 2000; Searle et al., 2001; Dobrzhinetskaya and Green, 2005; Liu et al., 2005; Hacker, 2006). Subduction of continental crust 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 (e.g., Ernst, 2005). It is not yet clear whether it is this factor that enables preservation and exhumation of UHP metamorphic rocks and their return back to normal crustal depths. The complementary thermal environment to the cold subduction zone environment is the backarc, and Hyndman et al. (2005a) have argued that since subduction zone backarcs have hot, thin and consequently weak lithosphere over a considerable distance, these are likely sites of crustal thickening and orogenesis during continental collision. Collision could generate CW or CCW metamorphic P–T–t paths, according to the relative rates of thickening and heat transfer.

An important question relates to the styles of orogenesis earlier in Earth history. Next I review metamorphic conditions of ancient orogenic systems. The pattern of crustal metamorphism in the geological record indicates secular change and records two major discontinuous modifications in the thermal structure of orogenic systems that separate three tectonic regimes in Earth history. However, I avoid the sterile semantics of what constitutes plate tectonics—it is possible that some form of thermal boundary layer tectonics has operated throughout most of Earth history (Davies, 2006).

METAMORPHIC REGIMES THROUGH TIME

Change in the metamorphic record broadly coincides with the change from the Archean to Proterozoic and Proterozoic to Phanerozoic Eons (Fig. 8).

Ultrahigh-temperature metamorphism (UHT) metamorphism

UHT metamorphism has been recognized during at least four distinct periods in Earth history (see Table 2 and references therein): 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 Braziliano and Pan-African belts of the southern continents at ca. 630-510 Ma). There is only one confirmed example of UHT metamorphism >>2.5Ga in age, which is the Voronezh Crystalline Massif of Sarmetia (ca. 3.2 Ga in age). UHT metamorphism is rare in the post-Cambrian rock record. In the Phanerozoic Eon, parts of the European Variscides evolve from E-HPG metamorphism to UHT metamorphism, but there is only one confirmed example of UHT metamorphism younger than Paleozoic, which is the Gruf Unit in the central Alps (ca. 33 Ma in age). The four main periods of UHT metamorphism show a remarkable coincidence with the first formation of supercratons (Superior/Sclavia) and then the supercontinent cycle (Nuna, Rodinia and Gondwana).

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

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 (see Table 3 and references therein). 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 UHT metamorphic conditions, both apparently without distinction between Proterozoic and Phanerozoic examples. Compositional zoning of minerals and reaction coronas, symplectites and kelyphites are common as multivariant reactions are crossed during this evolution. Three periods of E-HPG metamorphism coincide with the first formation of supercratons (Superior/Sclavia) and then the Nuna and Rodinia supercontinent cycles, but unlike UHT metamorphism *sensu stricto*, E-HPG metamorphism is characteristic not just of Gondwana assembly but also of the Lower Paleozoic Caledonides and Variscides. E-HPG metamorphism is rare in the post-Carboniferous rock record.

Blueschist (BS) and high-pressure to ultrahigh-pressure (HP-UHP) metamorphism

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).

HP-UHP metamorphism has been recognized during multiple periods in Earth history (see Table 5 and references therein): 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 UHP rocks of Miocene-Pliocene age (Fergusson Island, eastern Papua New Guinea). P–T–t paths from UHP metamorphic 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 CW paths with isentropic decompression segments 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.

What is the depth limit after which continental crust is irretrievably 'lost' to the mantle? The mineral assemblages of common continental crustal rocks at pressures around 10 GPa—based on experimental studies and assemblages calculated using internally-consistent thermodynamic datasets (Liu et al., 2005; Massonne and Brandelik, 2005)—suggest that continental crust will become denser than ambient mantle at a depth of ~300 km, inhibiting its buoyant return to a more normal depth. Part of the Altyn Tagh in western China may have been returned from pressures that were this extreme – around the 'point of no return' (Liu et al., 2005).

Is there any evidence of crustal recycling into the deep mantle in the Proterozoic or Archean Eons? Two studies show enigmatic chemistry of diamond that may record early UHP metamorphism. First, microdiamonds from the 1,800 Ma Akluilâk minette dykes (Nunavut, Canada) are similar to diamonds formed in subducted metamorphic rocks (Cartigny et al., 2004)—they have high concentrations of un-aggregated nitrogen and positive $\delta^{15}N$—suggesting that they formed within supracrustal rocks taken down to ultrahigh pressures before being sampled by the minette magma. The postulated UHP metamorphism may have been related to either the Thelon or Trans-Hudson orogenic system and, since a cold environment is required, it may indicate that something akin to modern subduction was active during the Paleoproterozoic Era on some parts of Earth. Second, populations of sulfide inclusions in Archean diamonds from the Orapa kimberlite pipe in the Kaapvaal-Zimbabwe craton, Botswana, preserve mass-independent sulfur isotope fractionations (Farquhar et al., 2002). The data indicate that material was transferred from the atmosphere to the mantle by sedimentation, lithification, metamorphism and convective stirring, but imply that sulfur was not well mixed in the diamond source regions. Evidence for these connections is preserved because sulfide was encased in diamond and transported back to the surface, but it is unclear whether these connections require a process akin to modern subduction.

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 [e.g., the ca. 3,800 Ma metamorphism recorded by the Isua Supracrustal Belt of southern West Greenland, Komiya et al. (2002); and, the ca. 3,230 Ma metamorphism recorded in the Barberton and related greenstone belts of South Africa, Dziggel et al. (2002), Diener et al. (2005)]. Thermal gradients were around 750°C/GPa (~20°C/km) – consistent with generally thinner lithosphere during the Archean. Although ordinary granulites are common throughout the Archean Eon, extreme thermal conditions are not recorded before the Neoarchean. Furthermore, blueschist facies rocks are not documented and there is no metamorphic imprint of subduction of continental crust into the mantle. 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, 2004).

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 & Harrison, 2005), suggests a tectonic setting and/or process particular to this Eon.

Inferences and implications

Overall, the restricted time span of different types of metamorphism through Earth history and the periods of metamorphic quiescence during the Proterozoic Eon (Fig. 8) suggest: (1) a link with major events in the mantle, supporting the postulate of discontinuous geodynamic evolution of Earth; (2) a different style of tectonics in the Archean Eon in comparison with the Proterozoic Eon; and, (3) a different style of tectonics in the Proterozoic Eon in comparison with the Phanerozoic Eon.

The metamorphic 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.*' Such a premise is consistent 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. This is also consistent with assembly of the continental 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, which period of Earth history was characterized by the assembly, partial dispersal and reconfiguration of much of the continental lithosphere into several supercontinents. 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 UHT metamorphic belts in the interval Neoarchean to Cambrian, and their broad correlation in time with amalgamation of continental crust into supercratons and supercontinents, suggests that UHT metamorphism, which commonly records tight clockwise P–T paths, may record closure and thickening of back arcs with their characteristic high heat flow (cf., Currie and Hyndman, 2005; Hyndman et al., 2005b). 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 UHT 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, 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, if 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, Caledonide 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 (Fig. 7). 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 UHP metamorphism (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 features are characteristic of the modern plate tectonics regime.

SECULAR CHANGE IN TECTONICS

Although there are differences in opinion about how a hotter early Earth may have dissipated heat, there is general agreement that the mean mantle temperature must have been higher, crustal heat production must have been higher, the mean thermal boundary layer thickness must have been lower and the mean residence time of oceanic lithosphere in the thermal boundary layer must have been much shorter (e.g., Pollack, 1997). There are multiple consequences for the organization of the thermal boundary layer and for Archean-Proterozoic tectonics.

On modern Earth, plates and plate tectonics partly represent a surface conductive thermal boundary layer, which may return to the interior of Earth via subduction, but where continents and lithospheric mantle are buoyant, due to chemical composition and, therefore, mineralogy, and do not readily sink just because they are cold. The plate mosaic is likely a self-organized network of plates and forces readily reorganized by changes in stress (Anderson, 2002). Hargraves (1986) has argued that younger lithosphere on early Earth would subduct more slowly, and that if heat flow at about the Archean-Proterozoic transition was three times the present heat flow, then ridge length in the ocean basins would have to have been as much as twenty-seven times the present ridge length to dissipate the heat. In an alternative calculation, Abbott and Menke (1990) suggest that the length of plate boundaries at 2.4 Ga was about 2.2 times the present boundary lengths, with more than sixty individual plates moving with velocities and associated sea-floor spreading rates about 16% slower than today. The point of these examples is that whatever the configuration of ridges and plate boundaries, and whatever the number of plates, tectonics in the Archean was likely to have been different in many of the details in comparison with the modern plate tectonics regime.

The effect of increased radiogenic heat production (RHP) on temperatures generated in subduction-to-collisional orogens through time has been modeled in a preliminary manner by Chacko (2003). Because of higher abundances of heat-producing elements, older crust would have produced 1.2, 1.6 and 2.3 times more heat 1.0, 2.0, and 3.0 Ga ago, respectively. The effect of increased RHP on the thermal evolution of crust instantaneously thickened from 35 to 70 km suggests that collisional orogens as old as 1.0 Ga would have experienced maximum mid-crustal (20-40 km) temperatures that were approximately 70°C hotter than those of modern orogens. Collisional orogens at 2.0 and 3.0 Ga would have produced mid-crustal temperatures that were approximately 180 and 370°C, respectively, hotter than modern orogens. However, there is no evidence of these significantly higher and progressively declining temperatures of metamorphism in the rock record, and in those circumstances where high-grade or UHT metamorphic conditions are attributed to radioactive heating alone, the U and Th contents are high and modeling the heat production at the age of metamorphism achieves the temperatures required (Andreoli et al., 2006; Kramers et al., 2001). Indeed, the Archean rock record is dominated by rather ordinary conditions of crustal metamorphism and there is accumulating evidence that thermal conditions in Earth's very early continental crust also were rather ordinary (e.g., Valley, 2005; Watson and Harrison, 2005, 2006; Watson et al., 2006), although the evidence is by no means without dispute (e.g., Nemchin et al., 2006); one inference we may draw from this observation is that a model of instantaneous doubling of the continental crust is probably inappropriate as a model of Archean orogenesis (cf., Cagnard et al., 2006).

It is outside the scope of this paper to speculate on Archean tectonics. However, there are documented differences in structural styles between Archean, Proterozoic and Proterozoic orogenic belts (e.g., Hamilton, 1998; Marshak, 1999; Cagnard et al., 2006) and dramatic changes in the metamorphic record through the Archean to Proterozoic and Proterozoic to Phanerozoic transitions as documented in this paper. These differences and changes probably were linked to and reflect non-linear switches in mantle convection, although exactly what these might be is a matter of speculation and modeling. For example, whether the Archean to Proterozoic transition affects a change from convective overturn events to slab-induced overturn events in a layered mantle (Davies, 1995) or affects a change to episodic catastrophic subduction of ocean lithosphere and consequent plume events (O'Neal et al., 2005) demonstrates that whatever the mechanism might be, two-layer convecting mantle models generate episodic whole mantle convection overturn events on timescales appropriate to major transitions in Earth history. Furthermore, it is possible that some form of oceanic 'plate tectonics' operated in the Archean despite the high temperature of the mantle (e.g., Davies, 2006). The relationship between superplume events and heterogeneities in the deep mantle (Burke and Torsvik, 2004) and the strongly exothermic perovskite–post-perovskite phase transition within the lower mantle (e.g., Yuen and Monnereau, 2005) remains enigmatic, but demonstrate the importance of coupling in Earth system behavior.

END NOTES

1. I propose that one defining feature of a plate tectonics regime is the development of two contrasting thermal environments where plates converge—on modern Earth these are the subduction zone and the back arc, respectively—that are registered in the rock record as contemporaneous different types of metamorphism, viz. E-HPG and UHT metamorphism in Proterozoic-Cambrian orogens and BS-to-UHP and moderate-P, high-T metamorphic belts in the Phanerozoic orogens (Figs. 7 and 8).
2. 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 UHP metamorphism 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.
3. I argue the change from a Proterozoic plate tectonic regime to a modern plate tectonic regime during the Neoproterozoic-Cambrian reflects a change in global geodynamics and pattern of mantle convection. The change was responsible for the reconfiguration of Rodinia and it's transformation into Gondwana—the last of the supercontinents—and the progressive destruction of a long-standing hemispheric-scale oceanic realm by closure of the Rheic Ocean (during the Paleozoic) and subduction successively along the Paleo-Tethyan and Tethyan trenches (during the Mesozoic; e.g., Collins, 2003).
4. Finally, UHT metamorphism clearly must be limited by the liquidus for crustal rocks, although for all practical purposes an upper limit of ~1,100°C seems to operate in nature. However, the maximum depth in the mantle from which crustal protoliths may be exhumed back to Earth's surface—the point of no return—is not well established, although it appears to be around 10 GPa. Clearly, we should determine the natural limits of terrane-scale crustal metamorphism at the extremes of temperature and pressure.

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Table 1. Peak P-T conditions and age of peak metamorphism for granulite (G) metamorphic belts.

Age (Ma)	Location	*P*(GPa) (from Pattison et al., 2003)	*T*(°C)	Original References
ca. 2,700	Minto block, Superior Province, Canada (average 7 samples)	6.7	840	Bégin and Pattison (1994); Pattison and Bégin(1994)
ca. 2,700	English River Domain, Superior Province, Canada (average 7 samples)	6.5	810	Perkins and Chipera (1985)
2690-2670	Quetico Metasedimentary Belt, Superior Province, Canada (average 6 samples)	8.0	880	Percival and McGrath (1986)
2685-2650	Ashuanipi Complex, Superior Province, Canada (average 10 samples)	7.8	880	Percival (1991)
ca. 2530	Karnataka, Dharwar Craton, India (average 7 samples)	9.3	840	Janardhan et al. (1982); Hansen et al. (1987)
ca. 2530	Nilgiri Hills, India (average 51 samples)	9.9	800	Raith et al. (1990)
2590-2510	Napier Complex, Enderby Land, Antarctica (average 25 samples)	8.9	950	Harley (1985); Harley et al. (1990)
ca. 2,500	Kasai Craton, Zaire (average 6 samples)	7.2	810	Bingen et al. (1988)
1,950-1930	Taltson Magmatic Zone, Canada (average 15 samples)	6.7	930	Berman and Bostock (1997)
ca. 1,850	Ketilidian Belt, Greenland	5.8	940	Dempster et al. (1991)

	(average 9 samples)			
ca. 1,100	Adirondack Highlands, Grenville Province, North America (average 11 samples)	9.2	820	Bohlen and Essene (1979, 1980)
640-630	Furua Complex, Tanzania (average 16 samples)	12.6	900	Coolen (1980)
610-550	Sri Lanka Highlands, Sri Lanka (average 10 samples)	7.7	800	Schumacher et al. (1990)
530-520	Kerala Khondalite Belt, India (average 34 samples)	7.9	920	Chacko et al., (1987, 1996); Harris et al. (1982); Hansen et al. (1987); Nandakumar and Harley (2000); Santosh (1987); Santosh et al. (1990); Srikantappa et al. (1985)
ca. 510	Prydz Bay, Princes Elizabeth Land, Antarctica (average 16 samples)	7.2	890	Fitzsimons and Harley (1994)
ca. 510	Madras Granulites, India (average 8 samples)	8.5	860	Sen and Bhattacharya (1984)

Table 2. Peak P-T conditions and age of peak metamorphism for ultrahigh temperature (UHT) metamorphic belts.

Age (Ma)	Location	*P*(GPa)	*T*(°C)	References
3,178 +/-38	Voronezh Crystalline Massif, Sarmatia	1.0-1.1	1,000	Fonarev et al. (2006)
2,720-2,715	Kaapvaal Craton, Southern Africa	0.9-1.1	1,000-1,100	Dawson and Smith (1987); Dawson et al. (1997); Schmitz and Bowring (2003)
ca. 2,700	Sutam Block, Central Aldan Superterrane, Russia	1.05	1,070	Perchuk et al. (1985); Sal'nikova et al. (2004)
ca. 2,690	Southern Marginal Zone, Limpopo Belt, southern Africa	0.9	950	Kreissig, et al. (2000); Tsunogae et al. (2004)
ca. 2,680	Sipiwesk Lake, Pikwitonei Granulite Domain, NW Superior Craton, Canada	1.0-1.1	1,060-1,100	Böhm (2004); Arima and Barnett (1984)
ca. 2675	Layered gneiss unit, Teton-Wind River Domain, Wyoming, USA	>1.1	>950	Frost et al. (2005); Hughes et al. (2005)
2,590-2,510	Napier Complex, E Antarctica	0.9-1.1	1,050-1,100	Ellis et al. (1980); Harley (1985); Harley and Motoyoshi (2000); Hokada (2001); Hokada and Harley (2004); Hokada et al. (2004); Kelly and Harley (2005); Suzuki et al. (2006)
ca. 2,500	Andriamena Mafic Unit, north-central Madagascar	1.05	1,000	Goncalves et al. (2004); Paquette et al. (2004)
ca. 2,480	Badcall Bay, Assynt terrane, Lewisian Complex, Scotland	1.4	1,000	Love et al. (2004); Cartwright and Barnicoat (1989)
2090-2040	Bhandara-Balaghat Granulite Domain, Central Indian	0.9	950	Bhowmik et al. (2005)

	Tectonic Zone			
ca. 2,070	Jequié Block, Bahia, Brazil	0.5-0.7	1,000-1,100	Barbosa et al. (2004)
2027 +/-6	Central Zone, Limpopo Belt, southern Africa	1.0	1,000	Kröner et al. (1999); Jaeckel et al. (1997); Hisada and Miyano (1996); Droop (1989); Harris and Holland (1984)
ca. 2,000	InOuzzal Terrain, NW Hoggar, Algeria	1.0-1.1	1,000-1,050	Ouzegane and Boumaza (1996); Ouzegane et al. (2003); Bertrand et al. (1992); Kienast and Ouzegane (1987)
1,950-1,930	Taltson Magmatic Zone, north-western Canada	0.6-0.8	>962	Farquhar et al. (1996)
ca. 1,880	South Harris, Lewisian Complex, Scotland	1.0	955	Hollis et al. (2006); Mason et al. (2004); Friend and Kinny (2001); Baba (1998, 1999, 2003)
ca. 1,650	Wilson Lake, Grenville Province, Canada	1.0-1.1	1,060-1,100	Korhonen and Stout (2004); James et al. (2002); Arima et al. (1986); Morse and Talley (1971); Leong and Moore (1972)
1,525-1,510	Epembe Unit, NW Namibia	0.95	970	Seth et al. (2003); Brandt et al. (2003)
ca. 1100	Eastern Ghats Province, India	0.9-1.0	985-1,085	Bhattacharya (2004); Bhattacharya and Kar (2002); Sengupta et al. (2004); Jarick (2000); Dasgupta et al. (1992, 1994)
1,200-940	Mollendo-Camana Block, Arequipa-Antofalla Basement, central Andean Margin	1.2	1,000	Loewy et al. (2004); Martingole and Martelat (2003)
ca. 770	Niquelândia and Barro Alto complexs, Northern Brasília Belt, Brazil	0.8	980	Pimentel et al. (2004); Moraes and Fuck (2000); Ferreira Filho et al.

				(1994, 1998)
645-635	Anápolis-Itauçu complex, Southern Brasília Belt, Brazil	0.875	980	Baldwin et al. (2005, unpublished data); Piuzana et al. (2003); Moraes et al. (2002)
ca. 635	Labwor Hills, Uganda	0.95	1075	Leggo et al. (1971); Sandiford et al. (1987)
610-550	Central Highland Complex, Sri Lanka	1.1-1.2	1,000-1,100	Sajeev and Osanai (2004); Osanai et al. (2003), Kriegsman and Schumacher (1999)
550-530	Southern Granulite Terrain, India	1.1	980-1,030	Mohan et al. (1986); Brown and Raith (1996); Raith et al. (1997)
530-520	Kerala Khondalite Belt, India	1.0	950	Shabeer et al. (2005); Santosh et al. (2005); Morimoto et al. (2004); Pattison et al. (2003); Chacko et al. (1996)
517 +/-9	Lutzow-Holm Bay, East Antarctica	1.1	1,000	Kawakami and Motoyoshi (2004); Frazier et al. (2000); Motoyoshi and Ishikawa (1997)
511 +/-4	Prydz Bay localities, East Antarctica	1.1-1.2	1,000-1,050	Kelsey et al. (2003a, b); Harley (1998)
ca. 510	Madurai Block, Southern India	1.0	1000	Mohan et al. (1986); Brown and Raith (1996); Raith et al. (1997); Tamashiro et al. (2004)
ca. 254	Kontum Massif, Vietnam	1.1	1000	Osanai et al. (2004); Nakano et al. (2004); Nam et al. (2001)
ca. 33	Gruf Unit, central Alps	1.1	980-1,030	Liati and Gebauer (2003); Droop and Bucher-Nurminen (1984)

Table 3. Peak P-T conditions and age of peak metamorphism for medium-temperature eclogite – high-pressure granulite (E-HPG) metamorphic belts.

Age (Ma)	Location	*P*(GPa)	*T*(°C)	References
2,720 +/-8	Gridino eclogite-bearing melange zone, Belomorian Mobile Belt, Fennoscandian Shield	1.75	770-880	Volodichev et al. (2004)
ca. 2,487	Jianping Complex, North China Craton	1.1	800	Wei et al. (2001); Kröner et al. (1998a)
2,700-2,400	S Badakhshan Block, Afghanistan	1.2-1.3	700-750	Faryad (1999)
ca. 2,400	Sharyzhalgai Complex, SW Margin, Siberian Craton	2.3-3.0	920-1030	Ota et al. (2004a)
1,999.1 +/-1.1	Usagaran orogen, Tanzania	1.8	750	Möller et al. (1995); Collins et al. (2004)
1,848 +/-6	Ubendian, Tanzania	1.2-1.4	840-900	Sklyarov et al. (1998); Boven et al. (1999); Muhongo et al. (2002)
1904.0 +/-0.3	Snowbird Tectonic Zone, Canadian Shield	1.8-2.0	920-1,000	Baldwin et al. (2003, 2004)
1,900-1,850	Sanggan area, North China Craton	1.1-1.5	750-870	Guo et al. (2002)
1,900-1,800)	Hengshan, North China Craton	1.35-1.55	770-840	O'Brien et al. (2005); Zhao et al. (2001)
ca. 1790	Rudall complex, central Western Australia	1.2	800	Bagas (2004); Smithies and Bagas (1997)

1134-1129	Llano uplift, Texas	1.5-2.1	750	Carlson (personal communication)
1050-1020	Manicouagan Imbricate zone, and Molsen Lake, Gagnon and Lelekuau terranes, Grenville belt, Canada	1.5-2.0	800-900	Indares (1993, 1995, 1997, 2003); Indares and Dunning (2001)
1,100-960	Zambezi belt, Zimbabwe	1.9	760	Dirks and Sithole (1999)
ca. 975	Sveconorwegian belt, SW Sweden	1.0-1.2	705-795	Möller (1998, 1999)
ca. 630	Guaxupé	1.4-1.5	1040	Campos Neto and Caby (1999, 2000); Del Lama et al. (2000); Moraes et al. (2003); Valeriano et al. (2004)
ca 630	Varghina nappe complex, Brasilia orogen, SE Brazil	1.75	950	Campos Neto and Caby (1999, 2000); Del Lama et al. (2000); Moraes et al. (2003); Valeriano et al. (2004)
610 +/-2	Dahomeyides, Ghana-Togo	1.4	800-900	Attoh (1998a, b); Attoh et al. (1997)
ca. 605	NE Ox Inlier, NW Ireland	1.4-1.5	850-900	Sanders et al. (1987)
560-520	Musgrave Block, Australia	1.2-1.3	700-750	Scrimgeour and Close (1999)
ca. 530	Lufilian Arc-Zambesi Belt, Zambia	1.0-1.3	700-750	John et al. (2004)
ca. 500	South Altyn Tagh, NW China	1.9-2.7	870-1050	Zhang et al. (2005)
ca. 470	Carolina terrane, Southern Appalachians	1.2-1.5	700-800	Shervais et al. (2003)
470-460	Moine, Scotland	1.2	700	Friend et al. (2000)
432 +/-6	Roan, Western Gneiss region, Norway	1.45	870	Johansson and Möller (1986); Möller (1988); Dallmeyer et al. (1992a)
410-390	Payer Land, NE Greenland	1.4-1.7	800-850	Elvevold and Gilotti (2000); Gilotti and Elvevold (2002); Elvevold et al. (2003)

440-390	Bragança-Morais, NE Portugal	1.3-1.4	800-850	Marques et al. (1996)
402 +/-1	Góry Sowie, Sudetes, Poland	1.5-2.0	900-1,000	Kryza et al. (1996); O'Brien (1997)
395-380	Munchberg Massif, Germany	1.8	925	Gayk et al. (1995)
370-360	Armorican Massif, France	2.0	950	Ballèvre et al. (1994); Bosse et al. (2000); Le Hébel (2002)
369-360	Snieznik, Sudetes, Poland	2.1-2.8	800-1,000	Kryza et al. (1996); Bakun-Czubarow (1992); Klemd and Bröcker (1999); Pouba et al. (1985)
350-340	Massif Central, France	1.0-1.2	700-750	Pin and Vielzeuf (1983, 1988); Costa et al. (1993); Dufour et al. (1985); Gardien et al. (1990)
341.5 +/-0.8	Granulitgebirge, Saxony, Germany	2.2	1,010-1,060	Rötzler (1992); Rötzler and Romer (2001); von Quadt (1993); Kröner et al. (1998b)
ca. 341	Erzgebirge, Germany	2.1	830	Willner et al. (1997); Kröner and Willner (1998); Nasdala and Massonne (2000)
342 +/-5	N. Bohemia, Czech Republic	1.5-1.7	800-850	Kotková et al. (1996)
339.8 +/-2.6	S. Bohemia, Czech Republic	1.5	900	Fiala et al. (1987); Vrána (1989); O'Brien (1999); Kröner et al. (2000)
ca. 341	Moravia, Czech Republic	1.4-2.0	750-1,050	Medaris et al. (1995a, b)
ca. 340	Lower Austria	1.4-1.6	950-1050	Scharbert and Carswell (1983); Dallmeyer et al. (1992b); Carswell and O'Brien (1993); Becker (1997); Cooke et al. (2000); Cooke and O’Brien (2001)
340-335	Schwarzwald, Germany	1.5	950-1,000	Kalt et al. (2000)
ca. 330	Ulten zone, Eastern Alps	1.45	700	Godard et al. (1996); Tumiati et al. (2003)
ca. 330	Santa Lucia, Corsica	1.2-1.5	800	Pin and Vielzeuf (1983, 1988); Libourel (1988)

125-110	Fjordland, New Zealand	1.4	750-800	Ireland and Gibson (1998); Clarke et al. (2000)
ca. 100	Jijal Complex, Pakistan	1.5-1.6	800-900	Jan and Karim (1995); Yamamoto (1993); Yamamoto and Yoshino (1998); Yoshino et al. (1998)

Table 4. Peak P-T conditions and age of peak metamorphism for lawsonite blueschists and lawsonite eclogites.

Age (Ma)	Location	*P*(GPa)	*T*(°C)	References
ca. 475	Motalafjella, Spitsbergen	1.5-1.6	380-400	Agard et al. (2005)
		1.8-2.4	580-640	Dallmeyer et al. (1989); Hirajima et al. (1988)
ca. 470	Port Macquarie, New South Wales, Australia	>1.8-<2.6	560	Och et al. (2003); Fukui et al. (1995)
ca. 400	Western Gneiss Region, Norway	2.9	710	Root et al. (2004); Carswell et al. (2003); Ravna (1982)
ca. 370	Ile de Groix, Armorican Massif, France	1.8-2.0	450	Bosse et al. (2005); Ballèvre et al. (2003)
ca. 350	Western Tianshan, NW China	1.5-1.9	540-580	Klemd et al. (2005); Wei et al. (2003); Klemd et al. (2002)
ca. 340	Maksyutovo Complex, Southern Urals, Russia	1.8-2.1	520-540	Schulte and Sindern (2002)
ca. 240	Huangzhen-Zhujiachong, South Dabie Terrane, China	3.3	670	Li et al. (2004); Castelli et al. (1998)
ca. 230	Qinglongshan, Sulu Terrane, China	3.0-3.5	600-700	Mattinson et al. (2004)
ca. 218	Pinchi Lake, British Columbia, Canada	1.3	500	Ghent et al. (1993)
	(Grt-Cpx-Phe)	2.2-2.7	450	Ghent (personal communication)
ca. 145	Cazadero, Northern California, USA	1.0-1.3	360-400	Anczkiewicz et al. (2004); Shibakusa and Maekawa (1997)
	(Grt-Cpx-Phe)	1.8-2.2	430-460	Tsujimori et al. (2006b)
ca. 145	Jenner, Northern California, USA	0.9	320	Anczkiewicz et al. (2004); Ravna et al. (1994)
	(Grt-Cpx-Phe)	2.2	435	Ravna and Terry (2004)

ca. 132	Motagua Fault Zone, Guatemala	2.6	480	Tsujimori et al. (2006a); Brueckner et al. (2005)
ca. 120	Bantimala, Sulawesi, Indonesia	2.1	520	Tsujimori et al. (2006b); Parkinson et al. (1998); Miyazaki et al. (1996)
ca. 95	Sistan Suture Zone, Iran	2.25	600	Fotoohi Rad et al. (2005)
ca. 87	Orhaneli, NW Turkey	2.1-2.7	430	Okay (2002); Okay et al. (1998)
ca. 85	Corsica, France	2.0	390	Tsujimori et al. (2006b); Caron and Péquignot (1986)
ca. 80	Samaná Peninsula, Dominican Republic	1.6	360	Zack et al. (2004); Goncalves et al. (2000)
ca. 80	Garnet Ridge, Colorado Plateau, USA	3.5	540	Tsujimori et al. (2006b); Usui et al. (2003)
ca. 50	Sifnos, Greece	2.0	550-600	Forster and Lister (2005); Schmädicke and Will (2003)
ca. 44	Diahot Terrane, Pam Peninsula, New Caledonia	1.5	410	Potel et al. (2006); Spandler et al. (2005)
	(metamorphic peak)	2.1-2.4	475-580	Fitzherbert et al. (2003, 2005)
ca. 32	Eclogite zone, Tauern Window, Eastern Alps, Austria	2.2	600	Kühn et al. (2005); Hoschek (2001, 2004)

Table 5. Peak P-T conditions and age of peak metamorphism for granulite and ultrahigh presure (UHP) metamorphic belts.

Age (Ma)	Location	*P*(GPa)	*T*(°C)	References
ca. 630	Pan-African Nappes of Minas Gerais State, SE Brazil	>2.7	800	Parkinson et al. (2001)
ca. 620	Pan-African Nappes of Northern Mali, West Africa	>2.7	700-750	Caby (1994); Jahn et al. (2001)
530-528	Kumdy-Kol, Kokchetav Massif (diamond-grade), northern Kazakhstan	4.0-6.0	920-1,000	Rubatto et al. (2005); Sobolev and Shatsky (1990); Shatsky et al. (1995)
508-500	Kulet, Kokchetav Massif (coesite-grade), northern Kazakhstan	3.8-4.0	730	Rubatto et al. (2005); Sobolev and Shatsky (1990); Shatsky et al. (1995)
503.9 +/-5.3	Jianggalesayi, south-west Altyn Tagh, west China	2.8-3.0	820-850	Liu et al. (2001, 2003a); Zhang et al. (2001)
498-490	Lanterman Range, Northern Victoria Land, Antarctica	>2.7	750	Ghiribelli et al. (2002); Palmeri et al. (2003)
452 +/-1.7	Tromsø Nappe, Norway	3.4	735	Ravna and Roux (2006); Corfu et al. (2003)
440-420	Dulan Belt, North Qaidam	2.9-3.3	630-740	Mattinson et al. (2005a, b); Song et al. (2003)
420-400	Massif Central, France	>2.8	740-780	Lardeaux et al. (2001)
ca. 400	Western Gneiss Region, Norwegian Caledonides	>2.8	790	Ravna and Terry (2004); Root et al. (2004); Smith (1984, 1988); Wain (1997); Cuthbert et al.

				(2000)
389 +/-4	Maksyutov Complex, southern Urals	3.2	650	Chesnokov and Popov (1965); Leech et al. (2002); Bostick et al. (2003)
361-347	North-east Greenland, Greenland Caledonides	3.6	970	McClelland et al. (2005); Gilotti and Ravna (2002); Gilotti (personal communication)
360 +/-7	Erzgebirge, Saxony, Germany	>4.2	900-1,000	Massonne (1999, 2001); Nasdala and Massonne (2000)
360-340	Leaota Massif, South Carpathians	>2.7	700-750	Săbău (2000); Medaris et al. (2003)
355-340	Sudetes Mountains, SW Poland	3.0	700-800	Bakun-Czubarow (1991, 1992)
355-340	Bohemian Massif, Czech Republic	4.5-4.9	1,050-1,150	Nakamura et al. (2004)
ca. 345	Makbal, Kyrghyzstan	2.6	>610-680	Tagiri et al. (1995)
ca. 345	At-bashy, Kyrghyzstan (younger?)	>2.5	660	Tagiri et al. (1995)
ca. 345*	Western Tianshan, China	>4.95-5.07	560-610	Zhang et al. (2002, 2003, 2005a, b, c); Gao and Klemd (2003)
ca. 240	Dabie Shan, Central China	4.0	700	Ravna and Terry (2004); Wang et al. (1989, 1995); Wang and Liou (1991); Okay (1993)
ca. 230	Su-Lu Terrane, Eastern China	3.5	660	Zhang et al. (1995); Liu et al. (2004a, b); Liu et al. (2003b)

*233-225	Western Tianshan, China	>4.95-5.07	560-610	Zhang et al. (2002, 2003, 2005a, b)
120-115	Bantimala Complex, Central Sulawezi, Indonesia (eclogite)	>2.7	720-750	Parkinson et al. (1998)
ca. 90	Pohorje Mountains, Eastern Alps, Slovenia	3.0-3.1	760-825	Janák et al. (2004); Thöni (2005)
73.5 +/-3.4	Kimi, East Rhodope, Northern Greece	>2.5	700-780	Mposkos and Kostopoulos (2001); Liati et al. (2002)
53.5 +/-7	Tso-Morari Complex, Indian Himalaya	>2.8	>640	Mukherjee and Sachan (2001); Mukherjee et al. (2003); Sachan et al. (2004); Leech et al. (2005)
52 +/-18	Zermatt-Saas Zone, Western Alps	2.6-3.0	590-630	Reinecke (1991); van der Klauw et al. (1997)
46.3 +/-0.2	Kaghan Valley, Pakistan Himalaya	2.7-3.2	700-770	O'Brien et al. (2001); Kaneko et al. (2003); Parrish et al. (2003)
35.4 +/-1	Dora-Maira Massif, Western Alps	4.35	740	Hermann (2003); Chopin (1984); Chopin et al. (1991)
8-3	Fergusson Island, eastern Papua New Guinea	2.75	700	Baldwin et al. (2005)

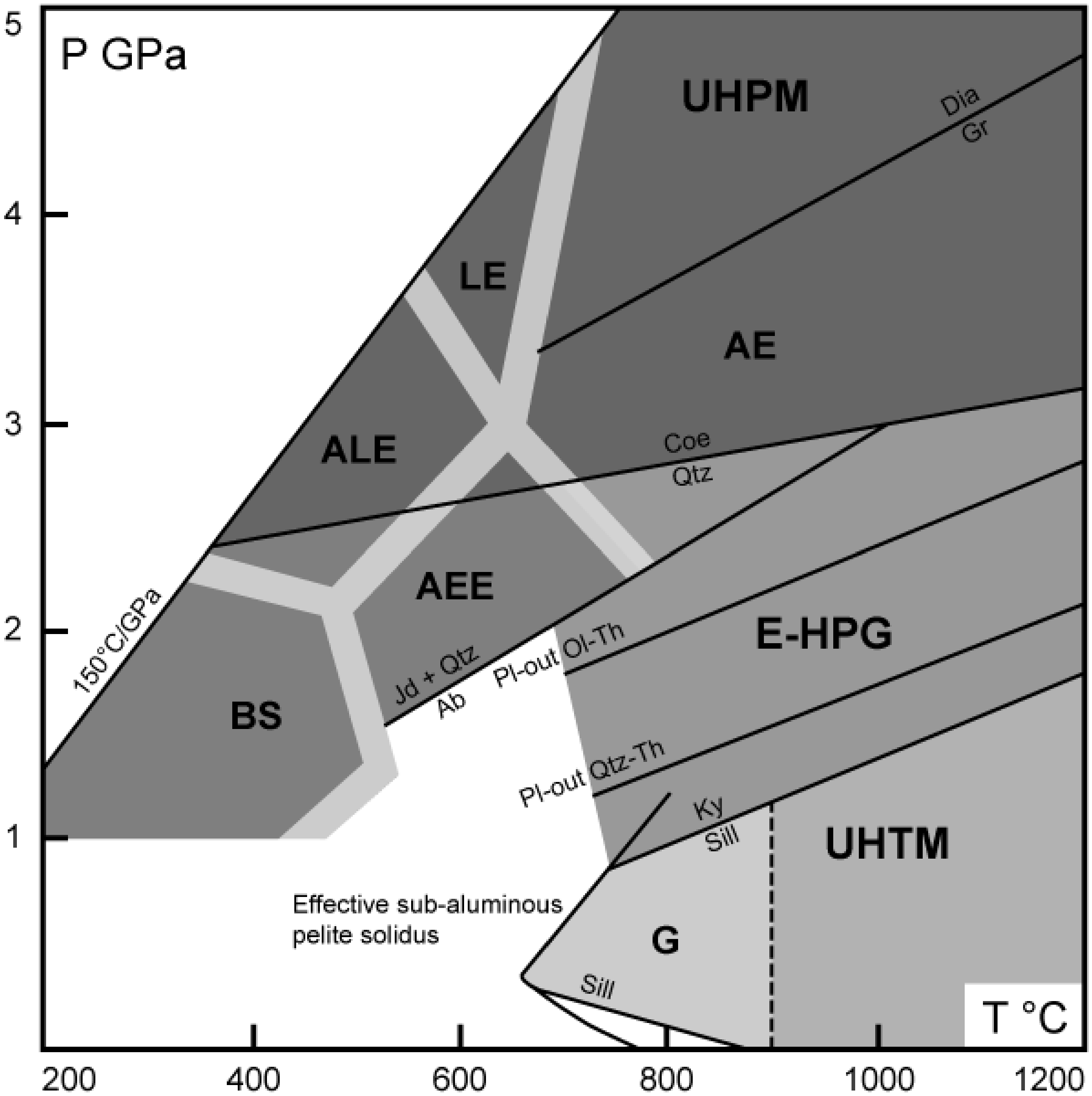

Figure 1. P–T diagram to show the P–T location of selected metamorphic facies and P–T ranges of different types of metamorphism. HP-UHP metamorphism includes the following – BS = blueschist, AEE = amphibole-epidote eclogite facies, ALE = amphibole lawsonite eclogite facies, LE - lawsonite eclogite facies,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

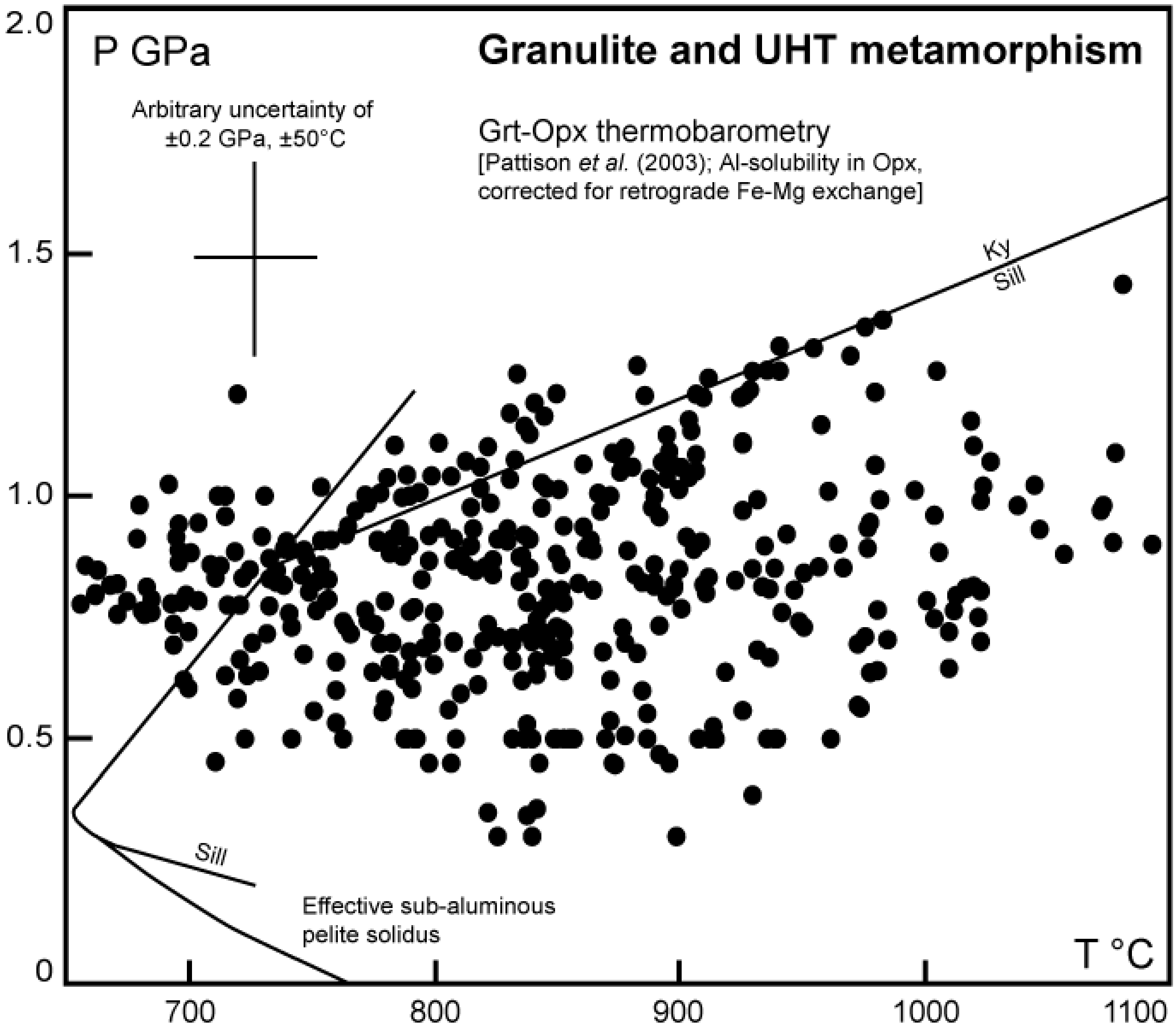

Figure 2. Internally-consistent thermobarometry of single rock samples of granulite and UHT metamorphism (Grt–Opx thermobarometry by Pattison et al. (2003) using Al-solubility in Opx corrected for retrograde Fe-Mg exchange).

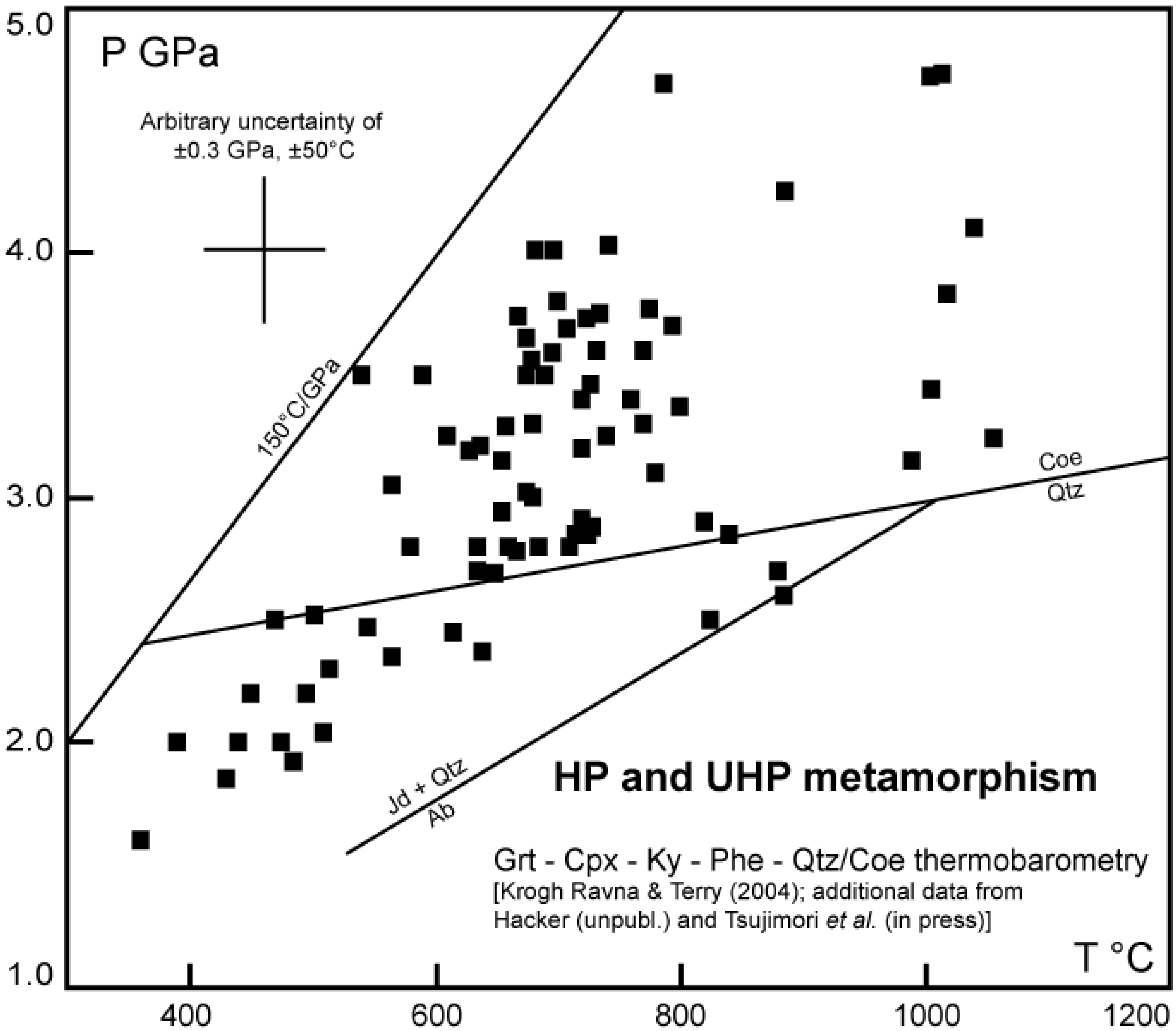

Figure 3. Internally-consistent thermobarometry of single rock samples for HP and UHP metamorphism [Grt–Cpx–Ky–Pag–Qtz/Coe thermobarometry using the calibration of Ravna and Terry (2004), with additional data from Hacker (2006) and Tsujimori et al. (2006b)].

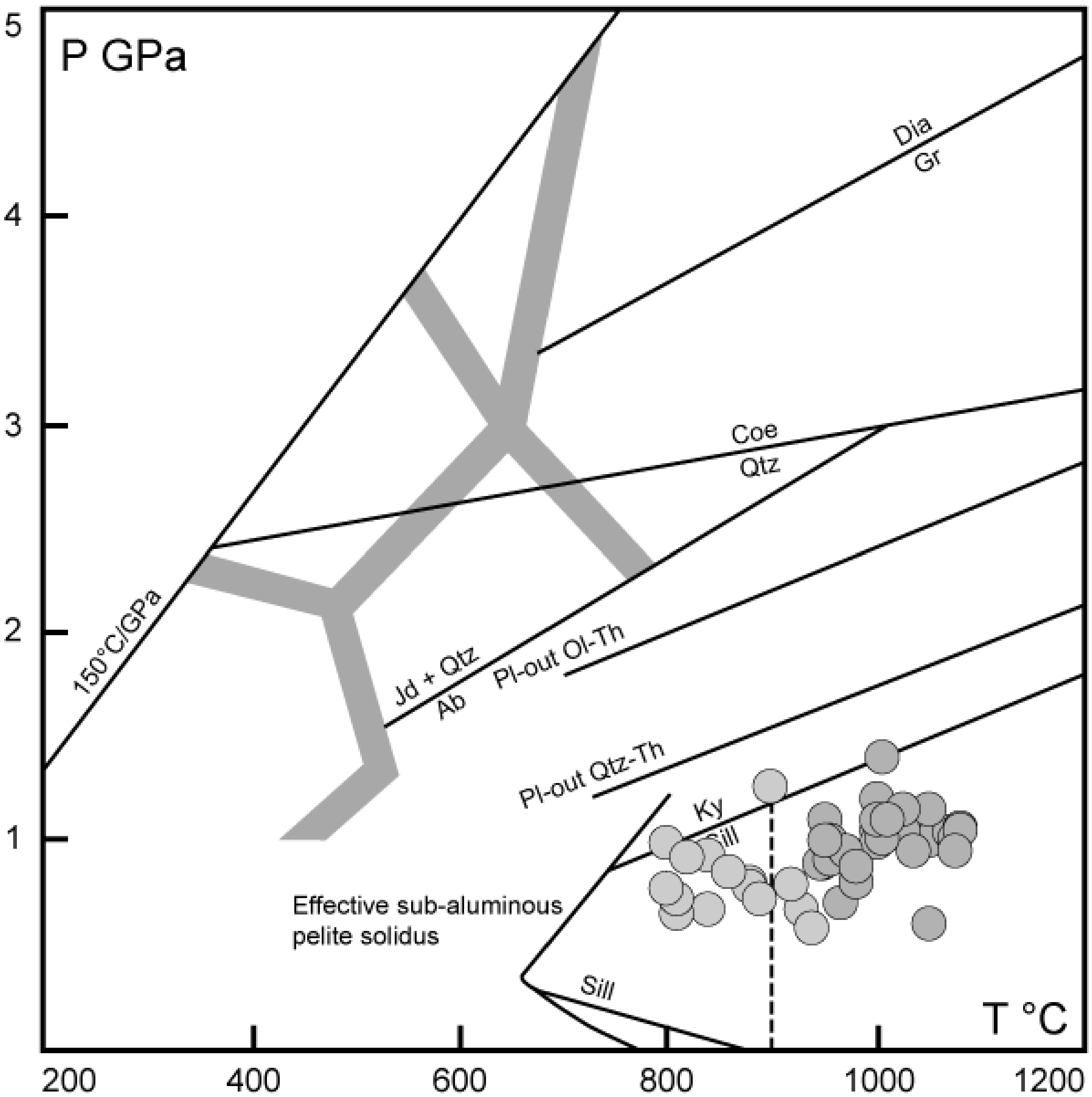

Figure 4. Peak metamorphic P-T conditions of metamorphic belts: granulite–ultrahigh temperature (G-UHT) metamorphism (lighter grey symbols are granulite belts from Table 1 and darker grey symbols are UHT metamorphic belts from Table 2).

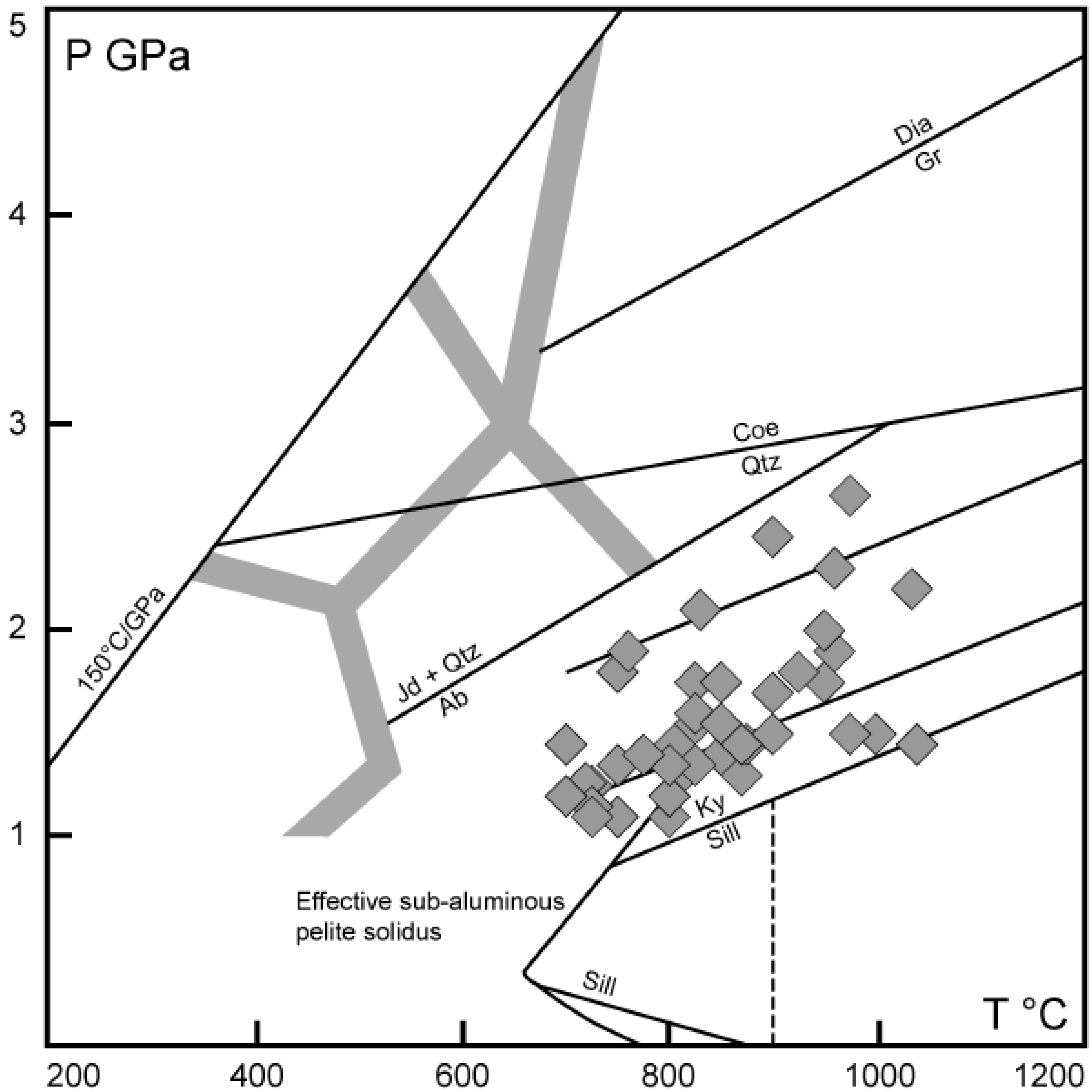

Figure 5. Peak metamorphic P-T conditions of metamorphic belts: eclogite–high-pressure granulite (E-HPG) metamorphic belts (data from Table 3).

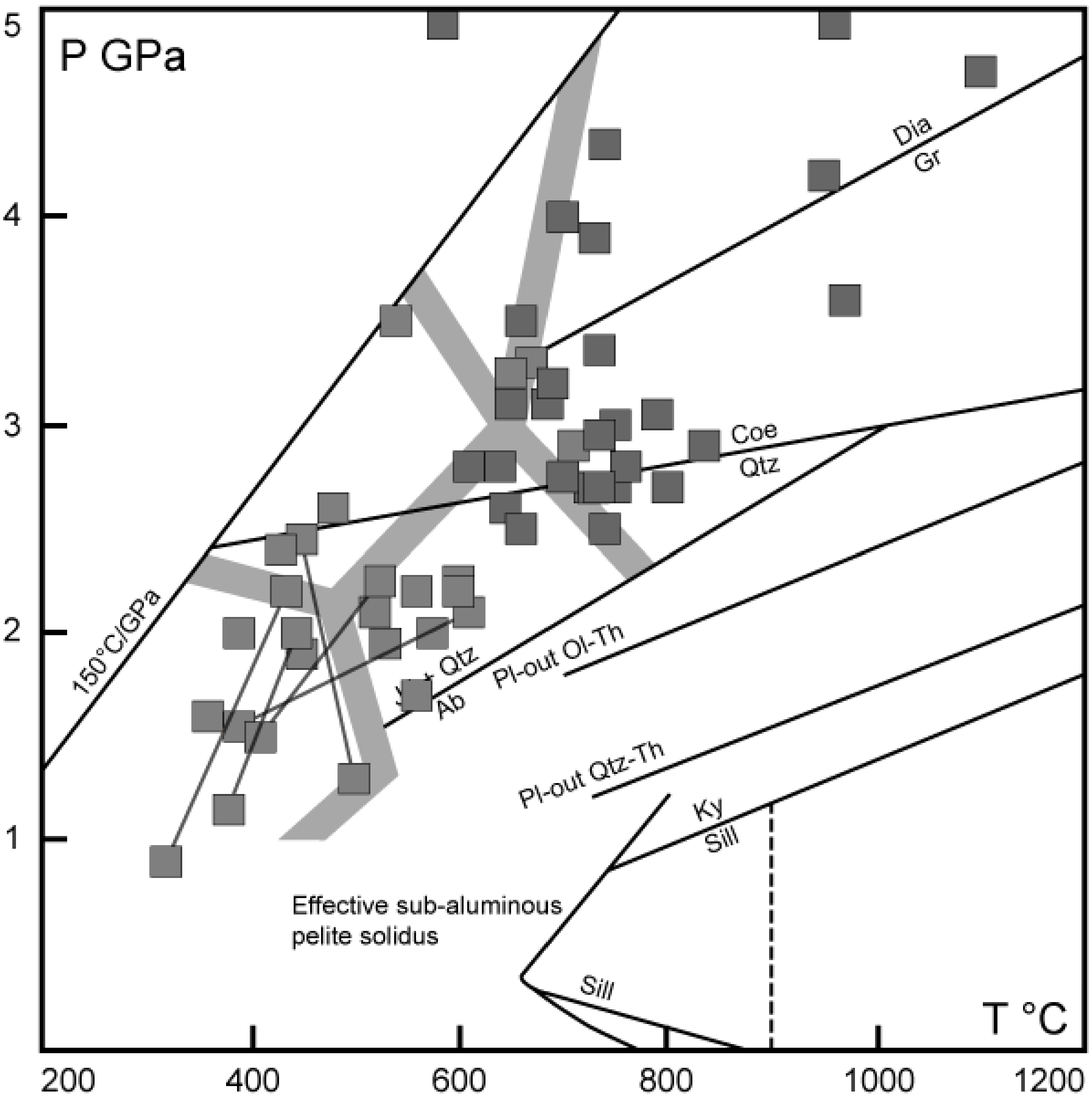

Figure 6. Peak metamorphic P-T conditions of metamorphic belts: high pressure–ultrahigh pressure (HP-UHP) metamorphism (lighter grey symbols are lawsonite blueschists and eclogites from Table 4, lines join prograde and peak P-T conditions for some localities, and darker grey symbols are UHP metamorphic belts from Table 5).

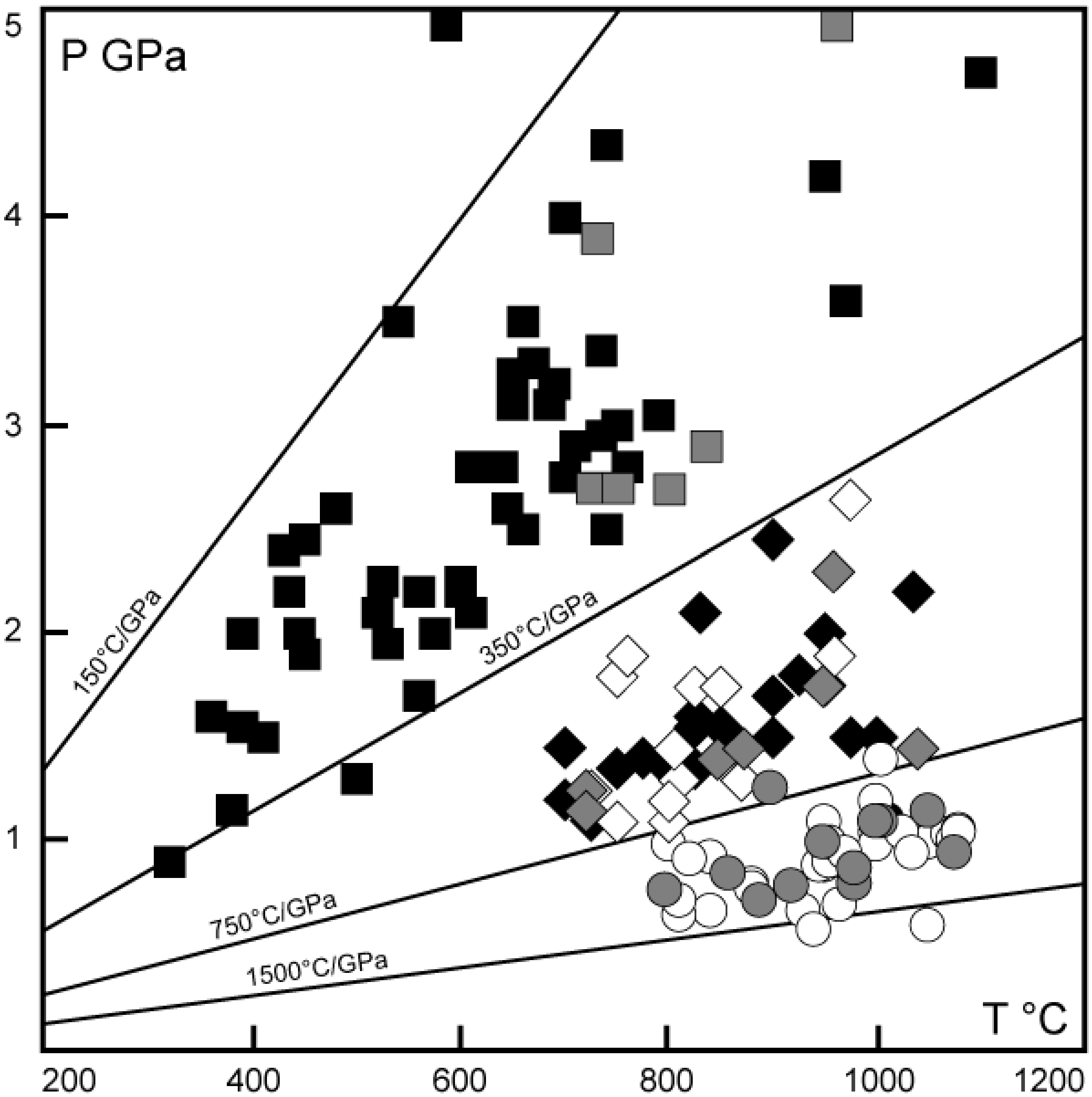

Figure 7. Metamorphic patterns by belt in relation to thermal gradients and age (filled symbols are belts where peak P-T conditions were recorded at < 488 Ma, grey symbols are belts where peak P-T conditions were recorded at < 850 but > 488 Ma, and open symbols are belts where peak P-T conditions were recorded at > 850 Ma).

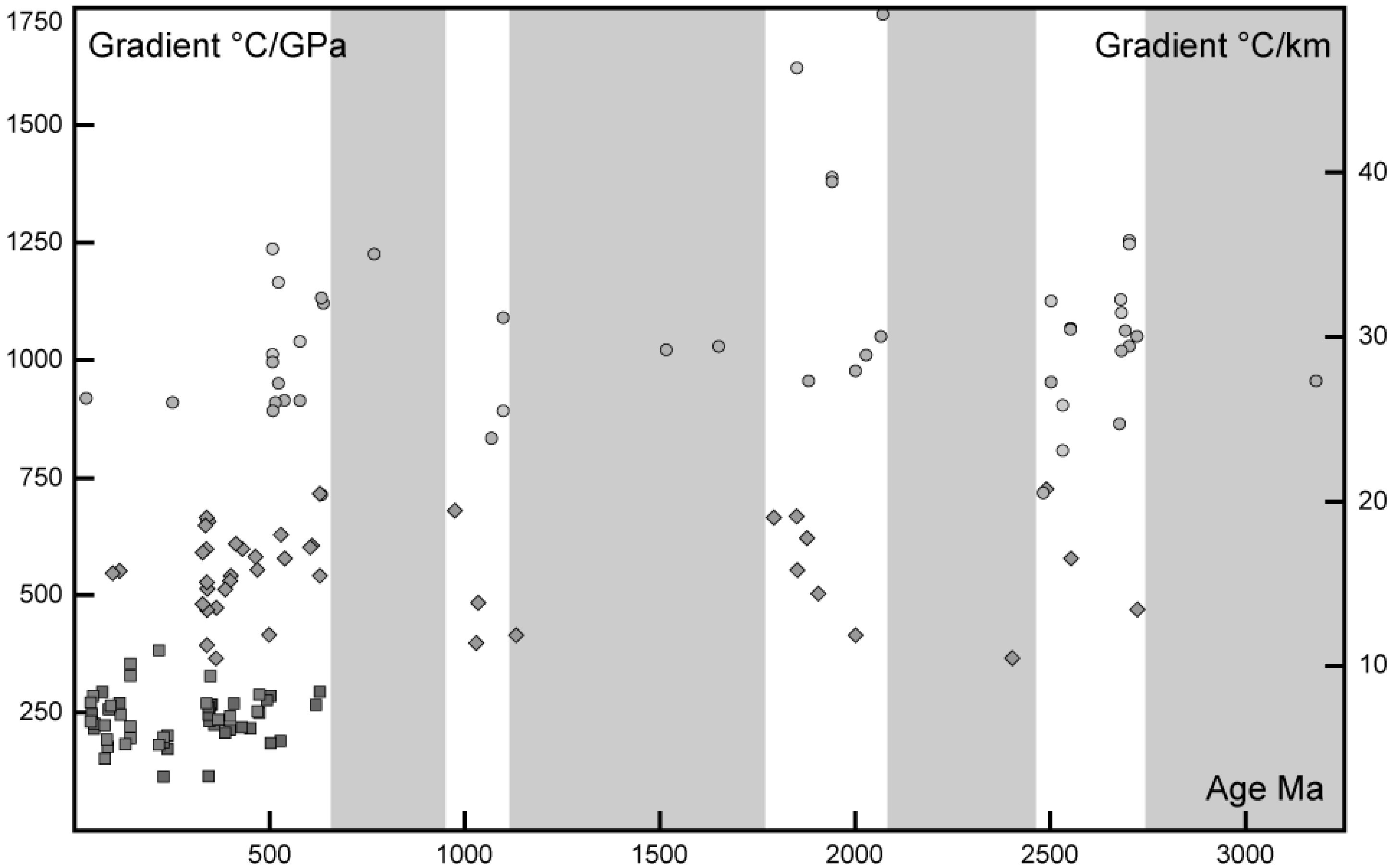

Figure 8. Plot of thermal gradients versus age for the three main types of extreme metamorphic belt – G-UHT metamorphism (circles), E-HPG metamorphism (diamonds) and HP-UHP metamorphism (squares); data from Tables 1-5.