Extreme metamorphism and metamorphic facies series at convergent plate boundaries: Implications for supercontinent dynamics

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ABSTRACT

Crustal metamorphism under extreme pressure-temperature conditions produces characteristic ultrahigh-pressure (UHP) and ultrahigh-temperature (UHT) mineral assemblages at convergent plate boundaries. The formation and evolution of these assemblages have important implications, not only for the generation and differentiation of continental crust through the operation of plate tectonics, but also for mountain building along both converging and convergent plate boundaries. In principle, extreme metamorphic products can be linked to their lower-grade counterparts in the same metamorphic facies series. They range from UHP through high-pressure (HP) eclogite facies to blueschist facies at low thermal gradients and from UHT through high-temperature (HT) granulite facies to amphibolite facies at high thermal gradients. The former is produced by low-temperature/pressure (T/P) Alpine-type metamorphism during compressional heating in active subduction zones, whereas the latter is generated by high-T/P Buchan-type metamorphism during extensional heating in rifting zones. The thermal gradient of crustal metamorphism at convergent plate boundaries changes in both time and space, with low-T/P ratios in the compressional regime during subduction but high-T/P ratios in the extensional regime during rifting. In particular, bimodal metamorphism, one colder and the other hotter, would develop one after the other at convergent plate boundaries. The first is caused by lithospheric subduction at lower thermal gradients and thus proceeds in the compressional stage of convergent plate boundaries; the second is caused by lithospheric rifting at higher thermal gradients and thus proceeds in the extensional stage of convergent plate boundaries. In this regard, bimodal metamorphism is primarily dictated by changes in both the thermal state and the dynamic regime along plate boundaries. As a consequence, supercontinent assembly is associated with compressional metamorphism during continental collision, whereas supercontinent breakup is associated with extensional metamorphism during active rifting. Nevertheless, aborted rifts are common at convergent plate boundaries, indicating thinning of the previously thickened lithosphere during the attempted breakup of supercontinents in the history of Earth. Therefore, extreme metamorphism has great bearing not only on reworking of accretionary and collisional orogens for mountain building in continental interiors, but also on supercontinent dynamics in the Wilson cycle.

INTRODUCTION

Temperature (T) and pressure (P) are two principal variables for regional metamorphism at convergent plate boundaries (e.g., Brown, 2001, 2007, 2014; Brown and Johnson, 2018), leading to different grades of metamorphism at different thermobaric ratios (Brown and Johnson, 2019a, 2019b). Extreme metamorphism is subdivided into ultrahigh-pressure (UHP) metamorphism occurring at mantle depths and ultrahigh-temperature (UHT) metamorphism occurring at crustal depths (Fig. 1). UHP metamorphism is defined by metamorphic pressures higher than the polymorphic transition of quartz to coesite (e.g., Coleman and Wang, 1995; Carswell and Compagnoni, 2003; Chopin, 2003; Liou et al., 2009), and UHT metamorphism is defined by metamorphic temperatures higher than 900 °C (e.g., Harley, 1998, 2008; Brown, 2007; Kelsey, 2008). The regional metamorphism of crustal rocks at these two extremes produces two types of mineral assemblages at lithospheric depths. One is at UHP eclogite facies, and the other is at UHT granulite facies. These two extremes are particularly important in metamorphic geology and continental dynamics because they have bearing on physicochemical mechanisms for mineralogical transformation of crustal rocks under different P-T conditions (e.g., Sajeev and Santosh, 2006; Santosh, 2011; Zheng and Chen, 2017).

Available studies indicate that identification of UHP metamorphism at previously convergent (converged) plate boundaries has important implications for the way in which the buoyant continental crust behaves during lithospheric subduction, continental collision, and crustal exhumation (e.g., Chopin, 2003; Liou et al., 2009, 2014). Although the geodynamic mechanism of UHT metamorphism is still controversial (e.g., Kelsey and Hand, 2015; Zheng and Chen, 2017; Harley, 2021), resolution of this issue may revolutionize ideas about the thermal and rheological behaviors of crustal basement under extreme P-T conditions.
that were previously considered impossible so far beyond the beginning of crustal anatexis. However, the physicochemical properties of metamorphic P-T conditions are similar, not only below or above the polymorphic transition from quartz to coesite for UHP metamorphism, but also below or above 900 °C for UHT metamorphism. In either case, there is no significant change in the geodynamics of plate boundaries at such conversions. For this reason, it is often questioned whether these two types of extreme metamorphism are diagnostic of any geodynamic change at plate boundaries (e.g., Brown, 2014; Brown and Johnson, 2018).

Within the framework of metamorphic facies series at convergent plate boundaries (Miyashiro, 1961, 1973), the two types of extreme metamorphism can be respectively linked to their lower-grade counterparts in the same metamorphic facies series (Zheng and Chen, 2017). One is the low-T/P Alpine facies series, which consists of blueschist- to eclogite-facies high-pressure (HP) to UHP metamorphic rocks, and the other is the high-T/P Buchan facies series, which consists of amphibolite- to granulite-facies high-temperature (HT) to UHT metamorphic rocks. As these two metamorphic facies series were formed at contrasting thermobaric ratios, their occurrence has bearing on the geodynamic style at convergent plate boundaries (Stern, 2005; Brown, 2006; Zheng and Chen, 2017; Zheng, 2021a). It is the difference in metamorphic thermobaric ratios that provides us with an excellent opportunity to decipher the tectonic evolution not only from oceanic subduction to continental collision, but also from convergent to divergent plate boundaries. This has important implications for the operation of plate tectonics (Brown et al., 2020a, 2020b).

In order to address both macroscale and microscale issues involving these two types of extreme metamorphism at convergent plate boundaries, this review highlights the thermal state of convergent plate boundaries and its relationships to dynamic regimes with respect to the origin of intracontinental orogens. In doing so, we have made the temporal distinction between actively convergent (converging) and previously convergent (converged) plate boundaries and thus between ongoing subduction (subducting) and fossil subduction (subducted) zones. The difference between converging and converged plate boundaries is particularly emphasized because it is related to the secular change in the metamorphic thermal gradients during the tectonic evolution of convergent plate boundaries. Furthermore, the metamorphic facies series are correlated with different types of mountain building in either compressional or extensional regimes. This leads to a more sophisticated understanding of the tectonic controls on the properties of regional metamorphism at convergent plate boundaries. The results provide insights into geodynamic mechanisms, not only for reworking of accretionary and collisional orogens through active rifting in continental interiors, but also for the assembly and breakup of supercontinents in the Wilson cycle.

### N. METAMORPHIC FACIES SERIES AT CONVERGENT PLATE BOUNDARIES

#### Metamorphic Facies

The concept of metamorphic facies was defined by Eskola (1915, 1920) for the same mineral assemblages in crustal rocks of equivalent chemical composition at similar metamorphic grades (e.g., Turner, 1948; Fyfe et al., 1958). Although its use in metamorphic petrology was controversial at the dawn of plate tectonics (e.g., Lambert and Sw, 1965; Fyfe and Turner, 1966), its applicability has been...
Metamorphic facies series were originally proposed by Miyashiro (1961, 1973) for mineral assemblages showing similar metamorphic field gradients at convergent plate boundaries. They were generalized into three major types (Miyashiro, 1994): (1) andalusite-sillimanite facies series (Abukuma or Buchan type); (2) kyanite-sillimanite facies series (Barrovian type); and (3) jadeite-glaucophane facies series (Alpine or Sanbagawa type). Thus, a metamorphic facies series is a sequence of metamorphic facies that were formed over a range of P-T conditions that plot in a P-T diagram along a thermobaric array (Fig. 2). While a single metamorphic facies is assigned to a given grade of metamorphic rocks that were formed over a limited range of P-T conditions, each metamorphic facies series is characterized by diagnostic thermal gradients for different grades of metamorphic rocks (Zheng and Chen, 2017; Zheng, 2021a). For this reason, the categorization of metamorphic facies by Miyashiro (1961, 1973) into the three thermobaric series represents important progress in recognizing the properties of metamorphic terranes at convergent plate boundaries (Miyashiro, 1994).

Metamorphic P-T conditions are recorded by metamorphic mineral assemblages at thermodynamic equilibrium. They can be transformed to metamorphic thermobaric ratios, \( T/P \text{ in } ^{\circ}\text{C}/\text{GPa} \) (Brown and Johnson, 2019a, 2019b), which are denoted as \( \frac{dT}{dP} \) in the literature (e.g., Brown, 2007, 2014; Brown and Johnson, 2018). Metamorphic thermal gradients are defined as ratios of temperature (\( T \)) and pressure (\( P \)).

Figure 2. Pressure-temperature (P-T) conditions for three types of regional metamorphism at different thermal gradients. The present diagram is revised from Zheng and Chen (2017), in which the boundary between Alpine and Barrovian facies series is defined by the metamorphic reaction of albite = jadeite and quartz, corresponding to thermal gradients of 10–11 \( ^{\circ}\text{C}/\text{km} \), and the boundary between Barrovian and Buchan facies series is defined by the polymorphic transition of \( Al_2SiO_5 \) between kyanite and andalusite at lower temperatures and between kyanite and sillimanite at higher temperatures, corresponding to thermal gradients around 30\( ^{\circ}\text{C}/\text{km} \) in consideration of overstepping during metamorphism. Thus, the Alpine facies series consists of blueschist and eclogite facies, the Barrovian facies series consists of kyanite-present amphibolite and granulite facies, and the Buchan facies series consists of kyanite-absent amphibolite and granulite facies. The P-T conditions for the polymorphic transition of \( SiO_2 \) are after Bose and Ganguly (1996), those for \( Al_2SiO_5 \) are after Pattison (1992), and those for the mineral reaction of \( Ab = Jd + Qz \) are after Holland (1980). Green lines denote the thermal gradients at 5 \( ^{\circ}\text{C}/\text{km} \), 10 \( ^{\circ}\text{C}/\text{km} \), and 30 \( ^{\circ}\text{C}/\text{km} \), respectively, corresponding to thermobaric ratios at 165 \( ^{\circ}\text{C}/\text{GPa} \), 335 \( ^{\circ}\text{C}/\text{GPa} \), and 1000 \( ^{\circ}\text{C}/\text{GPa} \). Mineral abbreviations: Coe—coesite; Qz—quartz; Jd—jadeite; Ab—albite; Ky—kyanite; Sil—sillimanite; And—andalusite.
to depth \( z \) for metamorphism at lithospheric depths, and they are normally denoted as \( T/z \) in °C/km (e.g., Stüwe, 2007; Zheng and Chen, 2017). In general, metamorphic thermal gradients can be translated from metamorphic thermobaric ratios through the relation \( z = P \rho g \), in which \( P \) is the lithostatic pressure, \( \rho \) is the average density of the rock column, and \( g \) is the gravitational acceleration. This translation is only valid under the assumption that metamorphic pressure is quantitatively determined by lithospheric thickness (i.e., lithostatic pressure) with \( P = 0 \) at \( z = 0 \). Such an assumption is related not only to the origin of heat flow but also to the operation of dynamic pressure (non-lithostatic pressure) during metamorphism. Crustal rocks rich in radioactive elements (such as Th, K, and U) can generate higher heat flow than average crust (Jaupart et al., 2016). On the other hand, high heat flow can be provided by upwelling of the asthenospheric mantle, resulting in high thermal gradients in areas where the lithospheric mantle is thinner, such as at mid-oceanic ridges, in backarc basins, and in continental rifts (e.g., Hyndman et al., 2005; Zheng and Chen, 2017).

By taking a range of metamorphic thermal gradients as an indication of the characteristic metamorphic \( P-T \) conditions, Zheng and Chen (2017) classified all metamorphic rocks at convergent plate boundaries into three facies series, i.e., Alpine, Barrovian, and Buchan (Fig. 2). This was achieved by defining two important facies series boundaries at higher and lower pressures, respectively (Brown, 2007; Kelsey and Hand, 2015). One is the metamorphic reaction from albite to jadeite and quartz between Barrovian and Alpine facies series at higher pressures. This boundary corresponds to lower thermal gradients of \(-10\text{–}11 \) °C/km for eclogite facies but as high as \(-13\text{–}14 \) °C/km for blueschist facies. The other is the polymorphic transition of Al\(_2\)SiO\(_5\) at lower pressures for boundaries between Barrovian and Buchan facies series at lower and higher temperatures, respectively. In detail, the andalusite to kyanite transformation marks the Buchan-Barrovian boundary at higher thermal gradients of \(30\text{–}35 \) °C/km, and the kyanite to sillimanite transformation marks the Barrovian-Buchan boundary at lower thermal gradients of \(25\text{–}30 \) °C/km. On average, a thermal gradient of \(30 \) °C/km approximates the boundary between the two facies series (Zheng and Chen, 2017). Consequently, Alpine facies series comprises blueschist facies, HP eclogite facies, and UHP eclogite facies. Everything in Alpine-type metamorphism is in reference to these three facies, or specific metamorphic \( P-T \) conditions. Both Barrovian and Buchan facies series comprise the same three facies, i.e., greenschist facies, amphibolite facies, and granulite facies. Nevertheless, Barrovian-type metamorphism proceeds at higher pressures than Buchan-type metamorphism at the same temperatures (Zheng and Chen, 2017; Pattison and Spear, 2018), resulting in the formation of kyanite in the former but andalusite and sillimanite in the latter. Therefore, the differences in pressure and petrology are translated to the differences in metamorphic thermal gradients (Zheng and Chen, 2017; Zheng, 2021a) and metamorphic thermobaric ratios (Brown and Johnson, 2019a, 2019b).

For either metamorphic thermobaric ratios or metamorphic thermal gradients, they are a critical variable in the origin of metamorphic rocks with respect to their petrogenetic link to the geodynamics of plate tectonics. Thus, they can be used to assess temporal changes in the dynamic regime of regional metamorphism at convergent plate boundaries (e.g., Brown, 2007, 2014; Zheng and Chen, 2017; Brown and Johnson, 2018, 2019a, 2019b; Zheng and Zhao, 2020). Using thermobaric ratios, Brown and Johnson (2019a, 2019b) categorized all metamorphic rocks into three types (Fig. 3): (1) a low-\( T/P \) type (<375 °C/GPa), including HP blueschist and low-\( T \) (LT) eclogite, and UHP metamorphic rocks; (2) an intermediate-\( T/P \) type (375–775 °C/GPa), including HP granulate and medium-\( T \) (MT) to HT eclogites; and (3) a high-\( T/P \) type (>775 °C/GPa), including HT and UHT granulites. This trichotomous classification is a modification of a previous classification by Brown (2007) based on natural groups. In doing so, thermobaric ratios for the HT eclogite stability field are expanded from \(-375 \) °C/GPa to \(<775 \) °C/GPa, but no attention is paid to the boundary of phase changes between Barrovian and Buchan facies series.

In essence, the trichotomous classification of Zheng and Chen (2017) by thermal gradients is almost identical to that of Brown and Johnson (2019a, 2019b) by thermobaric ratios, if pressure can be directly translated to depth at convergent plate boundaries. For this reason, Zheng (2021a) merged the two kinds of classification for metamorphic rocks into one (Figs. 2 and 3): (1) low-\( T/P \) Alpine facies series, comprising blueschist- to eclogite-facies HP to UHP metamorphic rocks; (2) moderate-\( T/P \) Barrovian facies series, comprising kyanite-bearing amphibolite-facies to HP granulate-facies metamorphic rocks; and (3) high-\( T/P \) Buchan facies series, comprising kyanite-absent amphibolite-facies to granulite-facies HT to UHT metamorphic rocks. Therefore, Alpine, Barrovian, and Buchan facies series are respectively used to register the three types of regional metamorphism at both converging and converged plate boundaries. In doing so, the thermobaric boundary between Barrovian and Buchan facies series is placed along the polymorphic progression boundaries of kyanite-andalusite at lower temperatures and kyanite-sillimanite at higher temperatures (Brown, 2007; Kelsey and Hand, 2015; Zheng, 2021a). In this regard, kyanite-bearing HP granulites formed at temperatures above 900 °C do not belong to the category of UHT metamorphic rocks.

Although thermobaric ratio and thermal gradient are the most important variables for the characterization of regional metamorphism in subduction zones, sufficient attention has not been paid to their changes with time during the lifetime of convergent plate boundaries. Available studies indicate that the trichotomous classification of metamorphic facies series is linked to the three-fold categorization of orogenic belts into accretionary, collisional, and rifting (Zheng and Chen, 2017; Zheng and Zhao, 2017, 2020), highlighting the causal relationships between metamorphism, magmatism, and orogenesis in both space and time during tectonic development from converging to converged plate boundaries. As a consequence, the two types of extreme metamorphism are tectonically coupled with the two types of dynamic regimes (compressional vs. extensional) and thermal states (cold vs. hot) at plate boundaries (Zheng, 2021a, 2021b). For this reason, the formation and evolution of metamorphic rocks are correlated with the generation and differentiation of continental crust through the operation of plate tectonics (e.g., Miyashiro, 1994; Brown, 2001, 2014; Zheng and Chen, 2017; Brown and Johnson, 2019a, 2019b; Holder et al., 2019; Zheng and Zhao, 2020).
Lithospheric Geotherms at Convergent Plate Boundaries

Lithospheric geotherms are quantified by thermal gradients, which are correlated with surface heat flow. Their temperatures at the same depths (e.g., at an average Moho depth of ~35 km) vary significantly from 400 °C to 1150 °C (Fig. 4), depending on lithospheric thickness. In general, thicker lithosphere correlates with lower surface heat flow. Thus, low heat flow (down to 30–40 mW/m²) is prominent in cratonic regions where lithosphere is thickest, whereas high heat flow (up to 80–100 mW/m²) prevails in rifting zones where lithosphere is thinnest (e.g., Stüwe, 2007; Artemieva, 2009; Tesauro et al., 2009; Jaupart and Mareschal, 2011; Balling et al., 2013; Furlong and Chapman, 2013). On average, continental lithosphere has surface heat flow of 60 mW/m², corresponding to the normal geotherm of 15 °C/km. Nevertheless, lithospheric geotherms at plate boundaries are highly variable, so they are classified into five types: (1) ultracold (<5 °C/km), (2) cold (5–10 °C/km), (3) warm (11–30 °C/km), (4) hot (30–60 °C/km), and (5) ultrahot (>60 °C/km).

Crustal rocks undergo prograde and retrograde metamorphism under different P-T conditions. While metamorphic pressures are approximated by lithostatic pressures at lithospheric depths, metamorphic temperatures are determined by lithospheric geotherms at plate boundaries. The lithospheric geotherms exert key control on major geological processes such as deformation, metamorphism, and magmatism, which are correlated with the geodynamic mechanism of tectonic processes at both convergent and divergent plate boundaries. While a steady-state heat flow prevails in cratonic and stable continental regions, transient heat flow is common in active tectonic settings, where asthenospheric heat flow may occur over a variety of time scales affecting different depths of the lithosphere (e.g., Stüwe, 2007; Artemieva, 2009; Tesauro et al., 2009; Jaupart and Mareschal, 2011; Balling et al., 2013; Furlong and Chapman, 2013). This is particularly so in lithospheric weak zones, which are generally fossil suture zones between previously convergent plates. Typically, continental rifts form along the weak zones, where the lithosphere is thinned during divergent lateral motions. This causes the lithosphere-asthenosphere boundary to move upward and heat flow to be advected from the hot mantle to the cold crust. Thus, rifts form as soon as the lithosphere attempts to rupture. The asthenospheric mantle and its derived material beneath the rift center start to move upward, increasing the lithospheric geotherms at shallower depths.

Lithospheric geotherms at convergent plate boundaries are usually characterized by steady-state heat flow, but often elevated by transient heat flow. The transient heat flow is derived from a number of heat sources, including not only internal ones, such as radioactive heat production elements in the orogenic crust (radioactive heating) and shear heating along the converging plate interface, but also external ones, such as heat flow from magmas of either mafic or felsic

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**Figure 3.** Pressure-temperature (P-T) diagrams of metamorphic facies series in terms of (A) petrology and (B) tectonics at convergent plate boundaries. Such combined petrologic-tectonic diagrams were originally derived from Stüwe (2007), the three types of metamorphic facies series are based on the classification of Zheng and Chen (2017), and the P-T data are derived from the compilation of Brown and Johnson (2019c) for metamorphic rocks of Archean through Proterozoic to Phanerozoic age. Bands delineating fields in A indicate that phase boundaries vary according to bulk composition of metamorphic assemblages, and red curve denotes the granite wet solidus. Lines in B denote the thermal gradients. Mineral abbreviations: And—anardslse; Coe—coesite; Ky—kyanite, Sill—sillimanite; Qz—quartz; Dia—diamond; Gr—graphite.
First-order role in Buchan-type metamorphism. As a consequence, Buchan facies series is produced at shallower depths than Barrovian facies series at the same thermal gradients (Zheng and Chen, 2017). In general, cold to warm geotherms prevail during lithospheric subduction for compressional heating, resulting in Alpines- to Barrovian-type metamorphism along converging plate boundaries. The compression is exerted by lithostatic pressure between two converging plates, which is proportional to the thickness of the overlying lithosphere and thus is a top-down property. Dynamic pressure becomes locally significant if solid material (either rock or mineral) can be carried into the rigid lithosphere rather than being subducted to lithospheric depths, where there is the weak interface between the two converging plates. As burial heating of crustal rocks is associated with the normal geotherm at the low thermal gradients. On the other hand, lithospheric extension during active rifting produces higher geotherms than the normal geotherm, giving rise to Buchan-type metamorphism at high thermal gradients. Because the increase of metamorphic thermal gradients in the continental crust is spatially associated with thinning of the lithospheric mantle, asthenospheric heating plays a first-order role in Buchan-type metamorphism. As a consequence, Buchan facies series is produced at shallower depths than Barrovian facies series at the same temperatures (Zheng and Chen, 2017; Pattison and Spear, 2018).

In general, cold to warm geotherms prevail during lithospheric subduction for compressional heating, resulting in Alpines- to Barrovian-type metamorphism along converging plate boundaries. The compression is exerted by lithostatic pressure between two converging plates, which is proportional to the thickness of the overlying lithosphere and thus is a top-down property. Dynamic pressure becomes locally significant if solid material (either rock or mineral) can be carried into the rigid lithosphere rather than being subducted to lithospheric depths, where there is the weak interface between the two converging plates. As burial heating of crustal rocks is associated with the normal geotherm in continental lithosphere, it is usually assumed to be the tectonic mechanism for Barrovian-type metamorphism. By contrast, hot to ultrahot geotherms prevail during lithospheric rifting for extensional heating, leading to Buchan-type metamorphism at converged plate boundaries. The extension is caused by active rifting, which also exerts bottom-up compression on the thinned lithospheric base due to underplating of the upwelling asthenosphere. The internal heating by radioactive elements and shearing becomes locally significant if it can be sufficiently accumulated in addition to the external heating.

The trichotomy of metamorphic thermal gradients for regional metamorphism at convergent plate boundaries is linked not only to the lithospheric thickness for lithostatic pressure and temperature, but also to the transient heat flow from asthenospheric upwelling (Zheng and Chen, 2017). In comparison, the trichotomy of metamorphic thermobaric ratios for regional metamorphism is not linked to given sources of pressure and temperature (Brown and Johnson, 2019a, 2019b), so it can be influenced not only by lithostatic and dynamic pressures but also by steady-state and transient heat flows. Generally, metamorphic thermobaric ratios increase from Alpine through Barrovian to Buchan facies series (Figs. 1–3), but this increase can be caused either by an increase of temperature or by a decrease of pressure. With respect to an increase of metamorphic thermal gradients from Alpine through Barrovian to Buchan facies series, this progression is commonly coupled with decompression in response to extension (Zheng and Chen, 2017; Zheng, 2021a). For this reason, prograde metamorphism is defined as a progressive increase of metamorphic grade along identical thermal gradients (or thermobaric ratios) inside the same facies series, resulting in the formation of metamorphic minerals with higher crystallinity and refractory. However, there are different definitions of retrograde metamorphism for different metamorphic facies series. Typically, retrogression of Alpines and Barrovian facies series is defined as decompressional metamorphism at elevated thermal gradients (or thermobaric ratios), and it often gives rise to Buchan facies series at high thermal gradients (or thermobaric ratios). Available observations indicate that ultracold subduction results in metamorphic mineral assemblages inside the forbidden zone in P-T space (Fig. 1). Ultracold to cold geotherms are common at converging plate boundaries such as modern oceanic subduction zones (Syracuse et al., 2010) and continental subduction zones (Faryad and Cuthbert, 2020), sometimes leading to the formation of lawsonite eclogites (Tsujimori et al., 2006; Tsujimori and Ernst, 2014). On the other hand, hot to ultrahot geotherms are remarkable not only in diverging plate boundaries such as mid-ocean ridges and backarc rifts, but also in converged plate boundaries such as continental rifts (Zheng and Chen, 2017; Perchuk et al., 2018). Ultrahot rifting may be responsible for pseuococontact metamorphism for part of the Buchan facies series (Harte and Hudson, 1979; Vite et al., 2013; Johnson et al., 2015). A warm realm is prominent in metamorphic soles in the base of ophiolites produced during subduction initiation (Agard et al., 2016). It may have been more common in pre-Neoproterozoic than Phanerozoic subduction zones (Zheng and Zhao, 2020).

The two types of extreme metamorphism at convergent plate boundaries leave their thermal records (P-T paths), which are relevant to overall metamorphic architecture (derived from mineralogy, P-T data, and age constraints). If either internal heat or dynamic pressure is insignificant during regional...
metamorphism at convergent plate boundaries, metamorphic thermal gradients can be approximated by metamorphic thermobaric ratios in the $P-T$ phase diagram with respect to their relationship to metamorphic facies series for crustal rocks (Figs. 1, 2, and 3). Therefore, peak metamorphic $P-T$ conditions are critical variables for individual facies series. They can be directly represented by peak thermobaric ratios for metamorphic rocks (Brown and Johnson, 2019a, 2019b). In this context, the dynamic regime and thermal state of tectonic processes can be retrieved from metamorphic rocks in both ongoing and fossil subduction zones (e.g., Brown, 2007; Zheng and Chen, 2017; Brown and Johnson, 2018; Holder et al., 2019; Zheng and Zhao, 2020). These rocks represent exposed slices of the crust that were subducted from forearc depths of <60–80 km to subarc depths of 80–200 km for Alpine-type metamorphism at low thermal gradients (Zheng and Chen, 2017; Zheng, 2021a). Subsequently, they were exhumed at first through slice thrust to Moho depths for Barrovian-type metamorphism at moderate thermal gradients and then through domical uplift to shallow crustal depths for Buchan-type metamorphism at high thermal gradients (Zheng and Chen, 2017; Zheng, 2021b).

## ALPINE-TYPE METAMORPHISM

Typically, crustal rocks undergo low- $T/P$ Alpine-type metamorphism during cold subduction (e.g., Ernst, 1988; Peacock, 1993; Maruyama et al., 1996; Chopin, 2003; Penniston-Dorland et al., 2015; McCarthy et al., 2018; Faryad and Cuthbert, 2020). Metamorphic $P-T$ conditions for Alpine facies series are well bracketed by the thermal gradients of modern subduction zones in the Phanerozoic (Fig. 5), indicating their production between two cold lithospheres at converging plate boundaries. Burial heating lead to metamorphic dehydration at temperatures below the solidus of crustal rocks, giving rise to mineral assemblages characteristic of blueschist to eclogite facies (Fig. 1). This is indicated by the formation of glaucophane in blueschist and omphacite in eclogite, with the disappearance of plagioclase during eclogitization. Typically, eclogitization is marked by the metamorphic reaction of albite to jadeite and quartz (Fig. 1), corresponding to thermal gradients of ~10–11 °C/km. For this reason, the Alpine facies series is generally produced at low thermal gradients of 5–11 °C/km (Figs. 2 and 3), equivalent to low thermobaric ratios between 150 °C/GPa and ~370 °C/GPa. Nevertheless, blueschist can be produced at elevated thermobaric ratios of 400–450 °C/GPa depending on the bulk composition of mafic protoliths (Ganne et al., 2012; Palin and White, 2016). This moves the upper limit of metamorphic thermal gradients to 13–14 °C/km (Figs. 1 and 3), close to the normal geotherm of 15 °C/km.

### Blueschist- to Eclogite-Facies Metamorphism

Originally, Alpine metamorphism was used to describe blueschist facies and eclogite facies in collisional orogens (e.g., Hietanen, 1967; Ernst, 1971; England, 1978). Historically, the blueschist facies was associated with the widest range of petrological interpretations: from outright rejection as a facies, through use in a restricted sense, without a definitive role for sodic amphibole, to use in a comprehensive form (Evans, 1990). Blueschist, also called glaucophane schist, is a kind of LT/HP metamafic rocks that is primarily produced by crustal subduction under $P-T$ conditions of ~1.0 to ~2.0 GPa and ~200 to ~500 °C (e.g., Maekawa et al., 1993; Maruyama et al., 1996), translating to lithospheric depths of ~30 to ~60 km (Fig. 1). The formation of lawsonite in blueschist assemblages indicates cold subduction, whereas epidote is stable in warm subduction zones. Blueschist facies is bounded by greenschist facies on the lower $P-T$ side and eclogite facies on the higher $P-T$ side (Bucher and Grapes, 2011). Blueschist-facies metamorphic rocks are the typical product of low- $T/P$ metamorphism in oceanic subduction zones (Ernst, 1971, 1973, 1975, 1988; Brothers and Blake, 1973; Maruyama et al., 1996; Monié and Agard, 2009). They occur at convergent plate boundaries as lithotectonic units in faulted contact with greenschist- or eclogite-facies metamorphic rocks.

Eclogite is a kind of HP to UHP metabasite with the characteristic mineral paragenesis of garnet and omphacite without plagioclase (Coleman et al., 1965; Smith, 1988; Carswell, 1990; Godard, 2001). It starts to form at pressures of 1.4 GPa and temperatures of >400 °C (Bucher and Grapes, 2011), corresponding to lithospheric depths of >50 km (Fig. 1). However, the eclogite $P-T$ range encompassed by the omphacite + garnet paragenesis is often expanded from...
Alpine facies to Barrovian facies (e.g., Palin and Dyck, 2018; Palin et al., 2020). For this reason, Coleman et al. (1965) classified eclogites into A, B, and C groups, and Carswell (1990) suggested separation into HT, MT, and LT eclogites. According to the stability of hydrous minerals based on experimental petrology, the eclogite facies was also subdivided into lawsonite-, epidote-, amphibole-, and dry-eclogite (Liu et al., 1998; Okamoto and Maruyama, 1999). For mafic rocks of crustal origin, there are paragonite-quartz eclogite, kyanite-quartz eclogite, coesite eclogite, and diamond eclogite. It appears that there are a variety of eclogites on Earth. Nevertheless, the Alpine-type eclogite forms through regional metamorphism at low thermal gradients during plate convergence. Thus, it corresponds to the C-type eclogite of Coleman et al. (1965) and the LT and MT eclogites of Carswell (1990). Its protolith can be oceanic crust or continental crust. It can be produced by either metamorphic dehydration or dehydration melting of the mafic crust. As an unusually dense rock, eclogite can gravitationally sink into the mantle and thus play an important role in driving plate tectonics. Although eclogite is widespread in cold subduction zones, it typically occurs as small boudins and lenses within amphibolite-facies gneisses, with rare preservation of eclogite-facies mineral assemblages in warm to hot orogens (Zheng, 2021b).

Metamorphic transformation from LT/HP blueschist facies to MT/HP eclogite facies takes place through the breakdown of glaucophane and the formation of omphacite and garnet (e.g., Makekawa et al., 1993; Poli, 1993; Maruyama et al., 1996), with release of water from the subducting crust in the garnet stability field (e.g., Schmidt and Poli, 2014; Zheng et al., 2016; Bang et al., 2021). In doing so, anorthite in plagioclase is transformed to garnet, kyanite, and quartz through metamorphic reactions; albite in the plagioclase is decomposed to jadeite and quartz, whereby omphacite is produced through incorporation of the jadeite component into diopside (Bucher and Grapes, 2011). Lawsonite may be produced at blueschist to eclogite facies (Tsujimori and Ernst, 2014). As a hydrous mineral, lawsonite may be metastable without decomposition during fast subduction, but it may be decomposed at elevated thermobaric ratios due to slow subduction. As the HP polymorph of the aluminosilicate minerals, kyanite is produced at HP to UHP conditions (Fig. 1). In typical LT eclogites, omphacite and garnet often coexist with glaucophane and clinozoisite over a fairly wide range of P-T conditions. At elevated temperatures, hydrous minerals are dominated by phengite and zoisite in UHP eclogites.

The transformation from blueschist to eclogite may play an important role in subduction zone magmatism owing to the dehydration and densification of descending oceanic slabs (Ringwood and Green, 1966; Ernst, 1970, 1971; Miyashiro, 1973; Råheim and Green, 1975). This is because: (1) large amounts of aqueous solutions are released by dehydration at the transition from HP to UHP conditions (e.g., Peacock, 1993; Schmidt and Poli, 1998), thought to be the biggest step in triggering partial melting of the overlying mantle wedge for mafic arc magmatism (e.g., Peacock, 1996; Manning, 1998); (2) the released aqueous solutions dissolve fluid-mobile incompatible trace elements in subducting slabs and thus contribute to arc geochemical signatures in mafic arc volcanics (e.g., Manning, 2004; Zheng, 2019); and (3) the formation of eclogite, which has the highest density of common crustal rocks, produces a negative buoyancy relative to the mantle peridotite, driving subduction of the oceanic slab (e.g., Hacker, 2003; Chen et al., 2013).

**UHP Metamorphism**

The transformation from HP to UHP eclogite-facies metamorphism is characterized by the polymorphic transition of quartz to coesite (Fig. 1), indicating an increase of depth from 50–80 km in the lower continental crust to >80 km in the lithospheric mantle. Thus, coesite serves as an index mineral in crustal rocks that were metamorphosed to UHP conditions (Chopin, 1984; Smith, 1984), corresponding to an increase of metamorphic pressure to >2.8 GPa. The occurrence of microdiamond inclusions in metamorphic minerals indicates pressures of >3.3 GPa (Sobolev and Shatsky, 1990; Xu et al., 1992), corresponding to mantle depths of >120 km (Fig. 1). The occurrence of majoritic garnet in eclogite indicates pressures of >6.0 GPa (Ye et al., 2000; Scambelluri et al., 2008), corresponding to depths of >200 km (Fig. 6). The occurrence of stishovite after coesite in metasedimentary rocks indicates pressures of >9.0 GPa (Liu et al., 2007, 2018), corresponding to depths of >300 km (Fig. 6).

The metamorphic P-T conditions of most UHP terranes are well established (e.g., Carswell and Compagnoni, 2003; Liu et al., 2009; Hermann and Rubatto, 2014; Faryad and Cuthbert, 2020). Estimating metamorphic pressures from exhumed HP to UHP metamorphic rocks has been one of the major tasks in reconstructing the tectonic architecture of continental subduction zones. In
general, pressures estimated from the thermodynamics of phase changes for a metamorphic rock are the total pressure composed of both lithostatic and dynamic pressures (e.g., Li et al., 2010; Gerya, 2015; Bauville and Yamato, 2021). As these two pressures cannot be discriminated at present, it is critical to examine whether metamorphic pressures approximate lithostatic pressures, which can be directly converted into lithospheric depths for the regional metamorphism at converging plate boundaries (Brown and Johnson, 2019a). Numerical geodynamic simulations suggest that the effects of crustal rheology and mechanical heterogeneities may yield significant dynamic pressures during crustal subduction to form HP to UHP metamorphic rocks (e.g., Li et al., 2010; Schmalholz and Podladchikov, 2013; Reuber et al., 2016), which could be significant at depths of >75 km (Bauville and Yamato, 2021). On the other hand, natural subduction zones are characterized by non-steady behavior over time and space (Marques et al., 2018). In particular, metamorphic rocks are mechanically weak under the $P$-$T$ conditions required for UHP metamorphism during plate convergence (e.g., Beltrando et al., 2007; Agard et al., 2009; Guillot et al., 2009; Rubatto et al., 2011). As a consequence, the dynamic pressure may be only transiently significant during continental subduction and collision (Warren, 2013). Therefore, the petrological methods do not lead to considerable overestimates of the maximum subduction/exhumation depth for regional metamorphism. If the dynamic pressures could be significant during the regional metamorphism in continental subduction zones, the metamorphic products would show prograde pressure-temperature-time ($P$-$T$-$t$) paths characterized by unusual near-isothermal excursions to very high pressures. This would be associated with a large decrease in metamorphic thermobaric ratios but a small change in temperature, corresponding to a decrease of thermal gradients from warm to cold, i.e., different from only cold ones for UHP metamorphic rocks (Zheng and Chen, 2016; Faryad and Cuthbert, 2020).

Typically, eclogite-facies UHP rocks are only found in continental subduction zones (Chopin, 2003; Liou et al., 2009; Zheng, 2012), and they are absent from oceanic subduction zones (Agard et al., 2009; Erdman and Lee, 2014). This difference does not mean that no oceanic crust was subducted to the subarc depths for UHP metamorphism. Instead, it indicates the preferential exhumation of eclogite-facies UHP slices with continental protoliths (Guillot et al., 2009; Zheng et al., 2013). This involves a switch in both dynamic regime and thermal state from compressional heating during subduction to extensional heating during exhumation at the slab-mantle interface (e.g., Yamato and Brun, 2017; van Keken et al., 2018). Nevertheless, blueschist- to eclogite-facies HP metamorphic rocks are common in both continental and oceanic subduction zones. So far, about 30 UHP terranes have been found on Earth, mostly in collisional orogens of the Eurasian continent (Carswell and Compagnoni, 2003; Liou et al., 2014). The age of UHP metamorphism is dominantly in the Phanerzoic (Carswell and Compagnoni, 2003; Liou et al., 2014), leaving only two in the late Neoproterozoic (Pompano in western Africa and Mali in Brazil). This scarcity of Precambrian UHP terranes may reflect a problem either in the preservation of mineral assemblages that are thermodynamically unstable at elevated thermal gradients or in the production of mineral assemblages that cannot be produced at moderate thermal gradients (e.g., Brown, 2014; Zheng and Chen, 2017; Brown and Johnson, 2018).

Since the discovery of UHP metamorphic rocks in the Italian Alps (Chopin, 1984), the Norwegian Caledonides (Smith, 1984), and the Chinese Dabie-Sulu (Okay et al., 1989; Wang et al., 1989) more than three decades ago, the exhumation of such UHP rocks has been considered to occur within the framework of continental collision (Zheng, 2021b). However, there are active debates on the tectonic mechanism for their exhumation (e.g., Little et al., 2011; Hacker and Gerya, 2013; Warren, 2013; Zheng et al., 2013, 2019; Faryad and Cuthbert, 2020), mostly focusing on two mechanisms of slice thrust versus domical uplift (Zheng, 2021b). Nevertheless, it is generally agreed that the exhumation of HP to UHP metamorphic rocks is primarily achieved by the positive buoyancy of continental crust (Ernst and Liou, 2008), and the exhumation rate depends on buoyant forces overcoming the boundary tractions (Warren et al., 2008). Inspection of UHP rocks produced at subarc depths in a time span from the early Paleozoic to the late Tertiary indicates that exhumation would mainly occur via slice thrust in an extensional regime during subduction channel processes (e.g., Agard et al., 2009, 2018; Guillot et al., 2009; Warren, 2013; Zheng et al., 2013; Li et al., 2016). Whereas HP to UHP metamorphic slices were exhumed in the late stage of collisional orogenesis (Zheng et al., 2013), low-grade metamorphic rocks were exhumed during the early stage of continental collision (Zheng et al., 2005). As a consequence, all of the exhumed crustal slices occur in the form of orogenic wedges between two collisional continents (Zheng, 2021c).

With respect to the tectonic evolution from subduction to exhumation of crustal rocks along colliding continental boundaries, there are two-stage processes for collisional orogenesis (Fig. 7). The early stage is the compressional continental cold subduction (Fig. 7A), resulting in Alpine type metamorphism at low thermal gradients. The late stage is the extensional heating during exhumation of subducted crustal slices (Fig. 7B), leading to metamorphic superimposition at elevated thermal gradients (generally at Barrovian facies series). Rollback of the subducting oceanic slab may have been responsible for the extensional exhumation (Brun and Faccenna, 2008; Husson et al., 2009). In this regard, the subduction dip would have changed from lower angles in the early stage to higher angles in the late stage (Fig. 7). Because of the buoyant effect of crustal slices (Ernst and Liou, 2008; Warren et al., 2008), extension leads to a significant drop in pressure at the onset of exhumation. Therefore, the extensional exhumation of HP to UHP metamorphic rocks is probably the rule.

The duration of UHP metamorphism at mantle depths is estimated based on a variety of petrochronometers (Zheng et al., 2009; Kylander-Clark et al., 2012). The results are categorized into two groups, one is short at 5–10 m.y. for small UHP terranes such as Dora Maira, New Guinea, Kahan, Tso Morari, Kokchetav, Erzgebirge, and North Qaidam, and the other is long at 15–20 m.y. for large UHP terranes such as the Dabie-Sulu orogenic belt of China and the Western Gneiss Region of Norway (Kylander-Clark et al., 2012; Zheng et al., 2013). In either case, the longevity of UHP metamorphism provides minimum estimates for the time scale of collisional orogeny because the initial exhumation of UHP metamorphic rocks marks the onset of the extensional
A. Low-angle subduction

B. High-angle subduction

Figure 7 Schematic cartoons showing the two-stage tectonic evolution during plate subduction for regional metamorphism. While the present diagram is adapted from Zheng and Zhao (2020), the original idea was abstracted from Zheng (2019). (A) A compressional regime prevails in the early stage of plate subduction when there is tectonic coupling between the subducting slab and the mantle wedge to cause crustal metamorphism at low thermal gradients. (B) An extensional regime prevails in the later stage of plate subduction due to rollback of the subducting slab in association with the tectonic decoupling between the two plates, to produce the space for buoyant exhumation of deeply subducted crustal slices with metamorphic overprinting at elevated thermal gradients.

regime after switch from the compressional regime (Yamato and Brun, 2017). Nevertheless, there are different interpretations for the difference between the two groups of metamorphic duration. Kylander-Clark et al. (2012) linked the small UHP terranes to rapid, high-angle subduction during the early stages of continental collision, when the negative buoyancy of subducting oceanic lithosphere still prevailed, pulling the continental lithosphere down; the large UHP terranes were ascribed to slow, low-angle subduction during the later stages of collisional orogenesis, when large resistance develops to prohibit the further subduction of massive, buoyant continental lithosphere. However, their linkage does not stand under close scrutiny with respect to the tectonic evolution of continental subduction zones (Zheng et al., 2013). In terms of the dynamic regime during continental collision, its early stage is featured by compressional heating during cold subduction at low angles (Fig. 7A), making exhumation of the deeply subducting crust impossible. Exhumation only becomes possible in the late stage when active extension has prevailed (Fig. 7B). Therefore, a feasible mechanism for the exhumation of UHP terranes are the subduction channel processes (Warren, 2013; Zheng et al., 2013), in which crustal slices are detached at different depths from subducting lithosphere and then sequentially exhumed through overthrusting along the same subduction channel to crustal depths (Zheng, 2021b). Therefore, it is possible that small and thin UHP terranes were exhumed faster and thus earlier than large and thick ones (Zheng et al., 2013). Furthermore, the small UHP terranes would be exhumed rapidly, whereas the large UHP terranes would be exhumed slowly. In either case, the exhumation of UHP terranes would proceed in two stages (Zheng, 2021b), with the first one through slice thrust from mantle to crustal depths after crustal detachment at subarc depths and the second one through domical uplift from the lower to upper crust levels. Nevertheless, it merits pointing out that the subduction channel model emphasizes the pathway of crustal slices during their subduction and exhumation (Zheng et al., 2013), whereas the orogenic wedge model mainly focuses on the occurrence of exhumed crustal slices between two collisional continents (Zheng, 2021c).

Clockwise P-T paths are common for HP to UHP eclogite-facies metamorphic rocks (e.g., Hermann and Rubatto, 2014; Zheng and Chen, 2017), reflecting a secular evolution crossing from lower to higher thermal gradients at convergent plate boundaries (Zheng and Chen, 2017; Brown and Johnson, 2018). Their subduction paths are characterized by synchronous increases in both P and T prior to peak pressures within Alpine facies series, indicating the top-down compressional heating of crustal rocks during subduction to the maximum depths (e.g., Hermann and Rubatto, 2014; Faryad and Cuthbert, 2020). Such interplate compressional heating is also diagnostic of prograde Alpine- and Barrovian-type metamorphism at low and moderate thermal gradients (Zheng and Chen, 2017). However, their exhumation paths show variable shapes (e.g., Hermann and Rubatto, 2014; Faryad and Cuthbert, 2020), with three types in general (Fig. 8): (1) decompressional cooling, (2) isothermal decompression, and (3) decompressional heating. In all cases, there are two stages of exhumation with significant increases in metamorphic thermal gradients from Alpine through Barrovian to Buchan facies series for most UHP terranes, which have different implications for exhumation mechanism (Zheng, 2021b). The decompressional cooling leads to synchronous decreases in both P and T subsequent to the peak pressures (Fig. 8A), which is common for exhumation of small UHP metamorphic terranes through slice thrust from mantle to crustal depths in the first stage (e.g., Zheng et al., 2013; Zheng, 2021b). The isothermal decompression is only associated with a decrease in P (Fig. 8B), which is common for exhumation of large UHP terranes through slice thrust from mantle to Moho depths in the first stage (e.g., Zheng et al., 2013; Zheng, 2021b). Decompressional heating results in a P decrease but a T increase until the peak temperatures occur (Fig. 8C), which is common for exhumation of retrograded UHP slices from deeper to shallower crustal depths through domical uplifts in the second stage (Zheng, 2021b). Therefore, these three types of exhumation paths have different bearings on the geodynamic evolution of convergent plate boundaries. As the maximum pressure (P_{max}) commonly occurs before the maximum temperature (T_{max}) in Alpine-type HP to UHP metamorphic rocks, P_{max} and T_{max} do not coincide along clockwise P-T paths (e.g., Zheng and Chen, 2017; Brown and Johnson, 2018). Although many published ages have been inferred to record the timing of peak UHP metamorphism, some may record retrograde P-T conditions at elevated thermobaric ratios.
**BUCHAN-TYPE METAMORPHISM**

Originally, Buchan metamorphism was interpreted to be responsible for andalusite-sillimanite amphibolites (Atherton, 1977; Harte and Hudson, 1979; Hudson, 1980, 1985). It was then extended to HT to UHT granulite facies at pressures below the kyanite-sillimanite boundary (Fig. 1). In view of the high thermobaric ratios for Buchan facies series (Figs. 2 and 3), Buchan-type metamorphism is now defined to occur in the andalusite stability field at lower temperatures and in the sillimanite stability field at higher temperatures (Zheng, 2021a). Thus, its lower $T/P$ limits are marked by the kyanite-andalusite phase boundary at lower temperatures and by the kyanite-sillimanite phase boundary at higher temperatures. Because of the large slope for the kyanite-sillimanite phase boundary, the lower $T/P$ boundary is associated with variable thermobaric ratios from 775 to 1100 °C/GPa, corresponding to variable thermal gradients from 23 to 33 °C/km. As a consequence, high-$T/P$ Buchan-type metamorphism would take place at high thermobaric ratios between ~1000 °C/GPa and 2000 °C/GPa, equivalent to high thermal gradients of ~30–60 °C/km (Zheng and Chen, 2017).

**Historical Evolution**

Buchan metamorphism was originally introduced by Read (1927) to describe HT/low-$P$ (LP) facies amphibolites in the Buchan area of northeastern Scotland, consisting of biotite, cordierite, andalusite, and sillimanite zones (Tilley, 1925; Kennedy, 1948; Read, 1952). These four zones show different mineral assemblages from Barrovian facies amphibolites first described by Barrow (1893, 1912) elsewhere in northeast Scotland. In the Buchan amphibolites, the polymorphic transition from andalusite to sillimanite is common at low pressures with elevating temperatures (Atherton, 1977; Hudson, 1980). Thus, the metamorphic rocks in the Buchan facies are associated with hotter geotherms than elsewhere in the Scottish Highlands, where Barrovian metamorphism dominates (Richardson and Powell, 1976; Harte and Hudson, 1979; Hudson and Kearns, 2000; Viets et al., 2010). Petrographically, the two facies are distinguished by the absence of kyanite in Buchan facies but the presence of kyanite in Barrovian facies. However, such distinction between Buchan facies and Barrovian facies has not been recognized sufficiently, leading to

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![Pressure-temperature (P-T) diagrams showing exhumation P-T paths of some high-pressure (HP) to ultrahigh-pressure (UHP) eclogite-facies metamorphic rocks in continental subduction zones. The P-T paths are revised from Faryad and Cuthbert (2020), and the trichotomous classification of metamorphic thermal gradients is after Zheng and Chen (2017).](http://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/17/6/1647/5475988/1647.pdf)
some confusion between the two types of regional metamorphism in the literature (e.g., Atherton, 1977; Hudson, 1980; Vorhies and Ague, 2011; Kelsey and Hand, 2015; Hyndman, 2019). It must be emphasized that moderate-\(T/P\) Barrovian-type metamorphism proceeds at higher pressures than high-\(T/P\) Buchan-type metamorphism (Fig. 2), with the formation of kyanite by the former but andalusite and sillimanite by the latter (Spear, 1993). Kyanite is produced by prograde Alpine- and Barrovian-type metamorphism during compressional heating. It is not stable during decompressional heating at thermal gradients of >25–30 °C/km, commonly converting to sillimanite.

In general, sillimanite develops following the first appearance of either andalusite at lower temperature or kyanite at higher pressure (e.g., Fyfe et al., 1958; Spear, 1993). The coexistence of andalusite and sillimanite is common in Buchan-type sequences, indicating that these sequences would develop along a \(P-T\) path of nearly isobaric heating for the polymorphic transition from andalusite to sillimanite. On the other hand, the coexistence of kyanite and sillimanite is common in transitional Barrovian-Buchan-type sequences, indicating that such sequences would develop along a decompressional \(P-T\) path to enter the sillimanite stability field. Nevertheless, kyanite is only metastable in the sillimanite field, being probably delayed to an elevated thermal gradient of 30–35 °C/km. Because of this kinetic effect on mineralogical transformation in nature (Carlson et al., 2015; Spear and Pattison, 2017; Pattison and Spear, 2018), the disappearance of kyanite would postdate the termination of Barrovian-type metamorphism and thus the onset of Buchan-type metamorphism in the same orogens. Nevertheless, the polymorphic transition from kyanite to sillimanite is characteristic of the metamorphic sequence transitional from Barrovian type to Buchan type. If kyanite-sillimanite parageneses still occur in metamorphic rocks, they represent the product of retrograde Barrovian-type metamorphism at elevated thermal gradients. Likewise, the polymorphic transition from andalusite to sillimanite is characteristic of prograde Buchan-type metamorphism, whereas the sillimanite to andalusite transition is indicative of retrograde Buchan-type metamorphism.

Amphibolite- to Granulite-Facies Metamorphism

In nature, both amphibolite- and granulite-facies metamorphic rocks occur in Buchan and Barrovian facies series (Figs. 1 and 2). They are common lithologies of metamorphic core complexes that mainly consist of gneiss domes, in which the lower-crustal basement rocks of higher metamorphic grade were tectonically juxtaposed against low-grade upper-crustal rocks (e.g., Teyssier and Whitney, 2002; Brown, 2008; Whitney et al., 2013; Platt et al., 2015; Searle and Lamont, 2020). Nevertheless, kyanite is absent in Buchan facies series but present in Barrovian facies series (Fig. 2). In northeastern Scotland, Buchan-type metamorphism was contemporaneous with ductile deformation, emplacement of mafic magmas, crustal anatexis for migmatization, and eventually generation of pelaluminous granites (e.g., Johnson et al., 2001a, 2001b; Carty et al., 2012; Johnson et al., 2015). Amphibole, diopside, epidote, plagioclase, alkamidine and grossular garnet, and wollastonite are minerals typically found in the metamorphic rocks of amphibolite facies. The boundary with the higher-grade, granulite-facies metamorphic rocks is determined by the appearance of orthopyroxene in metabasites. While metamorphic reactions at temperatures below their wet solidus produce the amphibolite-facies assemblages, peritectic reactions on and above the solidus bring about the granulite-facies assemblages (Fig. 1).

Granulite is a kind of HT to UHT metamorphic rock with granoblastic texture and gneissose to massive structure (Mehnert, 1972; Harley, 1989). The presence of feldspar and the absence of primary muscovite are critical, and cordierite may also be present; dominant Fe-Mg silicate minerals are nominally anhydrous (e.g., clinopyroxene and orthopyroxene), but amphibole is often present in minor amounts; quartz is present sometimes, especially in HP granulites. A common type of mafic granulite contains pyroxene, plagioclase, and accessory garnet, oxides, and possibly amphiboles, where both clinopyroxene and orthopyroxene may be present. In fact, the coexistence of clinopyroxene and orthopyroxene in metabasites (metamorphosed basalts) was used to define the granulite facies. Nevertheless, a common mineral assemblage of pelitic granulites is cordierite + garnet + K-feldspar + quartz + plagioclase + sillimanite + biotite, though not all these minerals necessarily occur together. In either case, granulite-facies metamorphism proceeds in a large range of temperature from as low as 650–700 °C for the wet solidus of felsic rocks to as high as 1050–1100 °C for the dry solidus of these rocks at low to high pressures (Harley, 1998, 2008).

UHT Metamorphism

The appearance of sapphire and quartz in metapelitic rocks is a diagnostic feature of UHT metamorphism (Dallwitz, 1968; Ellis et al., 1980; Harley, 1998), corresponding to an increase of metamorphic temperature to >900 °C (Fig. 1). The lower temperature limit is based on extant experimental and natural rock evidence on the stability of the mineral assemblage orthopyroxene + sillimanite + K-feldspar + quartz, formed through biotite-controlled fluid-absent dehydration melting reactions involving garnet, sillimanite, and cordierite (Harley, 1998). The mineral assemblage of Al-rich orthopyroxene-sillimanite-quartz in metaclastic rocks is characteristic; so are the assemblages containing the spinel-quartz pair (Brown, 2007). Geothermometries can also provide UHT metamorphic temperatures of >900 °C (Hensen, 1971; Hensen and Green, 1973; Ellis, 1980; Harley and Green, 1982). Metamorphic pressures are placed below the kyanite-sillimanite transition (Brown, 2007; Kelsey and Hand, 2015; Zheng, 2021a), precluding leucogranulites composed of garnet + kyanite + plagioclase/mesoperthite + quartz.

The metamorphic \(P-T\) conditions of many UHT terranes are well established (Harley, 2008, 2016; Kelsey and Hand, 2015). The breakdown of hydrous minerals is prominent during UHT metamorphism. This leads to partial melting of the crustal rocks, with significant extraction of hydrous felsic melts, leaving
These age clusters may be correlated with tectonic episodes of supercontinent Antarctica, the Anoysen of Madagascar, and the In Ouzzal terrane of Algeria Variscides in Europe (e.g., the Bohemian Massif, the Ivrea zone), the Triassic (16 Ma), and xenoliths from the Tibetan Plateau (6 Ma). Therefore, UHT metamorphism is generally defined under variable \( P-T \) conditions of 0.5–1.6 GPa and 900–1100 °C, with a synchronous change in the upper \( P-T \) limits defined by the kyanite-sillimanite boundary (Fig. 1) until the upper temperature limit occurs, defined by the dry solidus of felsic rocks (Harley, 1998; Kelsey and Hand, 2015).

So far over 50 UHT terranes have been recognized globally (Kelsey and Hand, 2015; Harley, 2016). The majority of them are relatively isolated outcrops (<10 km\(^2\)) or areas of restricted proven extent (<100 km\(^2\)), such as the Labwor Hills in Uganda, the Anápolis-Itaucu Complex in Brazil, the Voronesh Massif in Russia, the Gruf Complex in the Alps, Saint Maurice in Canada, the Mather supracrustals and Rundvagshetta in Antarctica, the Lace kimberlite pipe xenoliths of South Africa, Seram in Indonesia, and the Uweinat-Kamil inlier of Egypt-Sudan. Nevertheless, there are 15 large UHT terranes on regional scales varying from ~500 to >10,000 km\(^2\). The largest ones are the Napier Complex of Antarctica, the Anoysen of Madagascar, and the In Ouzzal terrane of Algeria (all >10,000 km\(^2\)); the Arequipa of Peru, the Kontum Massif of Vietnam, and the west Musgrave Province of Australia (all >5000 km\(^2\); the Bakhuis block of Surinam, Wilson Lake in Canada, the Eastern Ghats Belt of India, and the Khondalite Belt of northern China (all >2000 km\(^2\)); and the Madurai block of southern India (~1000 km\(^2\)). Other extensive areas include the Kramenataur area of the Snowbird tectonic zone in Canada, the Sutam block of the Aldan Shield in Siberia, Epupa in Namibia, the Highland Series of Sri Lanka, the Kerala Khondalite Belt of southern India, and the Scourian of NW Scotland (all ~500 km\(^2\)).

Geochronological studies indicate that the occurrence of UHT metamorphic rocks was abundant in the Archean and Palaeoproterozoic but rare in the post-Cambrian (Harley, 2008, 2016; Kelsey and Hand, 2015). Although the data indicate no temporal change of UHT metamorphism in response to the secular change in the thermal evolution of Earth (Brown and Johnson, 2019b), there are clusters of Precambrian UHT metamorphism at 2.8–2.45 Ga in the late Archean, at 2.1–1.8 Ga and 1.7–1.6 Ga in the late Palaeoproterozoic, at 1.1–0.95 Ga in the latest Mesoproterozoic to the earliest Neoproterozoic, and at 0.65–0.5 Ga in the late Neoproterozoic to the earliest Phanerozoic (Brown and Johnson, 2018). These age clusters may be correlated with tectonic episodes of supercontinent amalgamation and breakup, specifically those of Columbia/Nuna, Rodinia, and Gondwana. Phanerozoic UHT metamorphic complexes include the 350–330 Ma Variscides in Europe (e.g., the Bohemian Massif, the Ivrea zone), the Triassic Kontum Massif in Vietnam, and the Cenozoic Gruf Complex (32 Ma), the Hidaka Complex of Hokkaido (19 Ma), Japan, the Khibin granulites of Seram Island (16 Ma), and xenoliths from the Tibetan Plateau (6 Ma).

The duration of UHT metamorphism is estimated to range from <10 m.y. to >100 m.y. based on a variety of petrochronometers (Kelsey and Hand, 2015; Harley, 2016). This range in the longevity of UHT metamorphism is quite large, indicating that anomalous heat supply for anatectic metamorphism varies from very short to very long time scales at crustal depths. The partial melting of crustal rocks is prominent under UHT conditions if any hydrous minerals such as biotite and hornblende are present during the metamorphism (Zheng and Chen, 2017). The occurrence of nominally anhydrous mineral assemblages indicates significant removal of hydrous melts from crustal sources. As soon as melts are sufficiently produced to reach a critical volume of ~7 vol% (Rosenberg and Handy, 2005), their connectivity transition is achieved for drainage and separation from the sites of partial melting and thus transport upward by buoyancy for felsic magmatism (Brown, 2013). The production and migration of partial melts may result in partial to complete modification of primary \( P-T \) records under UHT conditions. Therefore, the highly variable durations of UHT metamorphism have bearing on the nature of tectonic settings for the time scales of anomalous heat supply.

Both clockwise and counterclockwise \( P-T \) paths have been reported for UHT metamorphic rocks (e.g., Santosh et al., 2012; Kelsey and Hand, 2015; Harley, 2016), and different interpretations have been given for their genetic relationships to tectonic settings (Fig. 9). Although counterclockwise \( P-T \) paths are not so popular for UHT granulites, their occurrence indicates a secular evolution crossing from higher to lower thermal gradients (e.g., Zheng and Chen, 2017; Brown and Johnson, 2018). While bulk \( P-T \) paths show different shapes, they can be separated into two segments to see changes of both thermal state and dynamic regime with time at convergent plate boundaries. In general, their first half is characterized by nearly compressional heating within the same Alpine or Barrovian facies series, pointing to an origin of subduction/collision during plate convergence. However, there is a large variation in their

Figure 9. Possible tectonic settings for ultrahigh-pressure (UHP) eclogite-facies and ultrahigh-temperature (UHT) granulite-facies metamorphism at convergent and divergent plate boundaries, respectively. While the present diagram is adapted from Santosh (2011), it is revised following arguments in Zheng and Chen (2017).
second half, often similar to the paths of decompressional heating for UHP metamorphic terranes (Fig. 8C). This can be ascribed to the difference in heat sources: (1) internal heat (radioactive accumulation), resulting in isobaric heating; or (2) external heat (asthenospheric upwelling), giving rise to underplating compressional heating. In either case, $T_{\text{max}}$ commonly occurs before $P_{\text{max}}$ along counterclockwise $P-T$ paths. This suggests that at least two-stage processes are responsible for UHT metamorphism (Zheng and Chen, 2017): (stage 1) removal of the thickened orogenic root via a certain mechanism (e.g., slab breakoff, lithospheric delamination, or asthenospheric erosion); (stage 2) underplating of the asthenospheric mantle to transfer the high heat flow into the thinned lithosphere. Since thermal gradients are much higher in the second stage than in the first stage, the extent of dehydration melting is much larger in the late stage than in the early stage.

**Thermotectonic Mechanism**

The thermotectonic causes for high-$T/P$ Buchan-type metamorphism have been very much debated (e.g., Harte and Dempster, 1987; Bohlen, 1991; Will et al., 2004; Lyubetskaya and Ague, 2010; Craven et al., 2012; Santosh et al., 2012; Gorczyk et al., 2015; Kelsey and Hand, 2015; Zheng and Chen, 2017; Harley, 2021). Both internal and external heat sources have been taken into account in petrogenetic interpretation. Typically, internal heating is attributed to accumulation of radioactive heat production elements in collision-thickened crust (e.g., Fyfe, 1973; Loosveld and Etheridge, 1990; Sandiford et al., 1998; McLaren et al., 1999; Henk et al., 2000; McKenzie and Priestley, 2008; Vilà et al., 2010; Bea, 2012; Alessio et al., 2018), whereas external heating is linked to emplacement of felsic or mafic magmas into thickened crust (e.g., Thompson and England, 1984; Loosveld and Etheridge, 1990; Sandiford et al., 1998; McLaren et al., 1999; Henk et al., 2000; McKenzie and Priestley, 2008; Vilà et al., 2010; Bea, 2012; Alessio et al., 2018), whereas external heating is linked to emplacement of felsic or mafic magmas into thickened crust (e.g., Thompson and England, 1984; Loosveld and Etheridge, 1990; Sandiford et al., 1998; McLaren et al., 1999; Henk et al., 2000; McKenzie and Priestley, 2008; Vilà et al., 2010; Bea, 2012; Alessio et al., 2018), whereas external heating is linked to emplacement of felsic or mafic magmas into thickened crust (e.g., Thompson and England, 1984; Loosveld and Etheridge, 1990; Sandiford et al., 1998; McLaren et al., 1999; Henk et al., 2000; McKenzie and Priestley, 2008; Vilà et al., 2010; Bea, 2012; Alessio et al., 2018), whereas external heating is linked to emplacement of felsic or mafic magmas into thickened crust (e.g., Thompson and England, 1984; Loosveld and Etheridge, 1990; Sandiford et al., 1998; McLaren et al., 1999; Henk et al., 2000; McKenzie and Priestley, 2008; Vilà et al., 2010; Bea, 2012; Alessio et al., 2018), whereas external heating is linked to emplacement of felsic or mafic magmas into thickened crust (e.g., Thompson and England, 1984; Loosveld and Etheridge, 1990; Sandiford et al., 1998; McLaren et al., 1999; Henk et al., 2000; McKenzie and Priestley, 2008; Vilà et al., 2010; Bea, 2012; Alessio et al., 2018), whereas external heating is linked to emplacement of felsic or mafic magmas into thickened crust (e.g., Thompson and England, 1984; Loosveld and Etheridge, 1990; Sandiford et al., 1998; McLaren et al., 1999; Henk et al., 2000; McKenzie and Priestley, 2008; Vilà et al., 2010; Bea, 2012; Alessio et al., 2018), whereas external heating is linked to emplacement of felsic or mafic magmas into thickened crust (e.g., Thompson and England, 1984; Loosveld and Etheridge, 1990; Sandiford et al., 1998; McLaren et al., 1999; Henk et al., 2000; McKenzie and Priestley, 2008; Vilà et al., 2010; Bea, 2012; Alessio et al., 2018). In either case, their efficiency is still controversial (e.g., Platt et al., 2003; Clark et al., 2011; Craven et al., 2012; Zheng and Chen, 2017). If radioactive self-heating is the principal driver, the time scale required to cause Buchan-type metamorphism depends on the age, composition, and volume of orogenic crust.

Advectional heating through magma emplacement has been suggested to play a significant role in the development of Buchan-type metamorphism in orogenic settings (e.g., Wickham and Oxburgh, 1985; Sandiford and Powell, 1986; De Yoreo et al., 1989). However, it is common for Buchan-type metamorphism to be associated with bimodal magmatism (Barton and Hanson, 1989; De Yoreo et al., 1991; Collins and Vernon, 1991; Sandiford et al., 1991; Kemp et al., 2007; Pownall et al., 2014; Glasson et al., 2019). In either case, their efficiency is still controversial (e.g., Platt et al., 2003; Clark et al., 2011; Craven et al., 2012; Zheng and Chen, 2017). If radioactive self-heating is the principal driver, the time scale required to cause Buchan-type metamorphism depends on the age, composition, and volume of orogenic crust.

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Although cratonic regions have low heat flow, high heat flow is common at mid-ocean ridges, in backarc basins, and in continental rifts (Fig. 10). It is in such extensional settings that high heat flow is transferred from the asthenospheric mantle into the lithospheric crust (e.g., Wickham and Oxburgh, 1985; Sandiford and Powell, 1986; Bodorkos et al., 2002; Platt et al., 2003; Collins and Richards, 2008; Langone et al., 2010; Craven et al., 2012; Okay et al., 2014; Gorczyk et al., 2015; Zhang et al., 2018; Cipar et al., 2020). In doing so, shallowing of the lithosphere-asthenosphere boundary layer is associated with asthenospheric upwelling for active rifting (e.g., Sengör and Burke, 1978; Wickham and Oxburgh, 1985; Olsen, 1995; Collins, 2003; Thybo and Nielsen, 2008), leading to the high thermal gradients for Buchan-type metamorphism at crustal depths (Zheng and Chen, 2017; He et al., 2018; Cipar et al., 2020; Zheng, 2021a). This
rifting tectonism may take place in ongoing subduction zones where the subducting slab rolls back for backarc extension, in fossil subduction zones where the thickened orogenic lithosphere was thinned for continental rifting, or in mid-ocean ridges where the asthenosphere upwells for seafloor spreading. In either site, the high heat flow from the asthenospheric mantle is conducted upward for extensional heating to cause the HT to UHT metamorphism at low pressures (e.g., Kemp et al., 2007; Clark et al., 2011; Pownall et al., 2014; Johnson et al., 2015; Zheng and Chen, 2017; Zheng, 2021a).

In backarc basins, the lithosphere is thinned by extension in response to rollback of the subducting oceanic slab, leading to upflowing of the asthenospheric mantle through the mantle wedge into the crust in active continental margins (e.g., Collins, 2002; Hyndman et al., 2005; Currie and Hyndman, 2006; Currie et al., 2008; Johnson et al., 2017). Thus, the high heat flow is linked to extension of the overriding plate in the orogenic hinterland (Sizova et al., 2014; Pownall, 2015; Chowdhury et al., 2017). In mid-ocean ridges, rifting of the oceanic lithosphere is spatially and temporally associated with asthenospheric upwelling, resulting in high temperatures of 800–1100 °C at very low pressures of <0.5 GPa (Manning et al., 1996, 2000; Nicolas et al., 2003; Bosch et al., 2004). Nevertheless, the rate of seafloor spreading is variable. If it is so low that supply of basaltic magmas derived from decompressional melting of the asthenospheric mantle is not efficient enough to accommodate lithospheric extension, then the result is the formation of oceanic core complexes (e.g., Tucholke et al., 1998; Whitney et al., 2013; Brun et al., 2018). In continental riffs that develop along previously subducted zones, asthenospheric upwelling leads to active rifting as soon as orogenic roots are foundered downward, and the previously thickened lithospheric mantle thins (Zheng and Chen, 2017; He et al., 2018; Cipar et al., 2020). All the above three settings point to extensional rather than compressional regimes for the Buchan-type metamorphism of crustal rocks in the thinned lithosphere. Furthermore, the extensional regime is prominent not only during lithospheric rifting, but also in the late stage of plate convergence, whereas the compressional regime is common in the early stage of plate convergence (Zheng, 2021a, 2021b). With respect to the global operation of plate tectonics on a spherical Earth, on the other hand, the extensional regime in mid-ocean ridges is temporarily linked to the compressional regime in subduction zones. Typically, subduction pulls the lithosphere downward, resulting in convergent motions on one side and divergent motions on the other side (passive rifting for seafloor spreading) and making the space for regional upwelling of the asthenosphere (active rifting for backarc extension).

**Buchan-Type Metamorphism in Intracontinental Orogens**

Buchan facies series are common in intracontinental orogens that were developed from previously accretionary and collisional orogens, where thinning of the thickened orogenic lithosphere results in continental rifting (Fig. 11), enhancing the advective heat transfer from the underlying asthenosphere into the thinned lithosphere (Zheng and Chen, 2017; He et al., 2018; Cipar et al., 2020). This results in regional thermal anomalies in the continental crust with high thermal gradients of >30 °C/km (e.g., Sandiford and Powell, 1986; Olsen, 1995). For this reason, Wickham and Oxburgh (1985, 1986, 1987) proposed crustal-scale rifting, possibly in a strike-slip setting, to explain the Pyrenean Buchan metamorphism in the Variscan orogen of Europe. However, this proposal was rejected by arguing against the existence of a rifting event during the Variscan orogeny in the Pyrenees, where not only was the compressional regime prominent during collisional orogeny, but also the envisaged continental rift failed to develop into rupture (Matte and Mattauer, 1987; Kriegsman, 1989). As a consequence, the causal linkage between Buchan-type metamorphism and continental rifting in continental interiors remains uncertain.

He et al. (2018) identified for the first time Buchan-type metamorphic rocks in China, which show the occurrence of both andalusite and sillimanite in a successful rift during the breakup of supercontinent Rodinia at ca. 750 Ma. Their thermobarometric study yielded metamorphic P-T conditions of 100–350 MPa.
and 560–660 °C, corresponding to high thermobaric ratios of >1600–2000 °C/GPa. Their heat flow estimates from crustal Th, U, and K contents are significantly lower than that from peak mineral assemblages, indicating that anomalously high heat flow from the asthenospheric mantle did occur in the successful continental rift where the South China Block was separated from Rodinia. Therefore, continental rifts are important sites for high thermal gradients and thus for Buchan-type metamorphism in continental interiors (Zheng and Chen, 2017; He et al., 2018; Cipar et al., 2020; Zheng, 2021a). In addition, bimodal magmatism and asthenospheric upwelling are the features typical of continental rifts in association with extensional thinning of the lithosphere (Wilson, 1989).

The rock record of continental rifting (either abortive or successful) is common in high- T/P Buchan facies series and its associated granites and migmatites (Zheng and Chen, 2017; He et al., 2018; Cipar et al., 2020; Zheng and Zhao, 2020). Before continental rifting, the lithosphere of convergent plate boundaries would be thickened by subduction/collision (Fig. 11A), generating orogenic roots that may be as deep as 200–300 km. The collision-thickened lithosphere would be thinned by removal of the orogenic roots (Fig. 11B) through various mechanisms such as delamination (Bird, 1979), sagduction (Chardon et al., 2009), peeling off (Chowdhury et al., 2017), and convective erosion (Houseman et al., 1981). While the former three mechanisms are powered by gravitational foundering of the eclogitized mafic crust, the fourth one is driven by thermal erosion of the asthenospheric mantle. In either case, they may result in the initiation of continental rifting along the thinned orogenic lithosphere with extensional heating (e.g., Spohn and Schubert, 1982; Platt and England, 1994; Morency et al., 2002; He et al., 2018; Cipar et al., 2020), giving rise to the Buchan facies series at convergent plate boundaries (Zheng and Chen, 2017; Zheng and Zhao, 2020; Zheng, 2021a).

Although the majority of Buchan facies series do not occur along successful continental rifts that have developed into seafloor spreading to form not only divergent plate margins, but also a globally linked continuous system, the following three observations conform with the active rifting model (Zheng and Chen, 2017): (1) the very high temperature at shallow crustal depths of <15–20 km for metamorphic dehydration and partial melting, requiring an enhanced heat flux from the asthenospheric mantle; (2) the long length/narrow width with the dome-and-keel structure for mountain chains, indicating a linear but heterogeneous thermal anomaly from upwelling of the asthenospheric mantle along the thinned orogenic lithosphere; and (3) compressional structures in the cores of gneiss domes but extensional structures in the overlying carapace, suggesting that the asthenospheric mantle was protruding into the thinned lithospheric mantle. As a result, compression prevails in deeper crust with higher P-T conditions, whereas extension prevails in shallower crust with lower P-T conditions due to the extrusion of metamorphic core complexes. This may explain common observations in collisional orogens where domal structures expose ductilely deformed high-grade metamorphic rocks in the core that is surrounded by a ductile-to-brittle high-strain detachment (Searle and Lamont, 2020).

Because active rifting is advocated as the geodynamic mechanism of Buchan-type metamorphism at convergent plate boundaries (Zheng and Chen, 2017; Zheng, 2021a), it is important to distinguish it from passive rifting within the framework of plate tectonics (Sengör and Burke, 1978; Turcotte and Emanuel, 1983; Wilson, 1989; Nicolas et al., 1994; Ziegler and Cloetingh, 2004). Active rifts were defined to develop in response to lithospheric extension due to impingement of upraising hot mantle on the base of the cold lithosphere. Along convergent plate boundaries where the lithosphere was thickened by accretionary or collisional orogenesis, active rifting commonly takes place subsequent to removal of orogenic roots to cause thinning of the thickened orogenic lithosphere (Dunbar and Sawyer, 1988; Olsén and Morgan, 1995; Rey, 2001; Thybo and Nielsen, 2009; Zheng and Chen, 2017). Nevertheless, the orogenic lithosphere was only thinned from its lower part (one side) and this process predate the lithospheric extension resulting in active rifting. As a consequence, the thinned lithosphere undergoes heterogeneous extension at different depths, with the maximum extension in its top but the minimum extension in its bottom where both heating and compression are imparted from the upwelling asthenosphere. This leads to the upward heat transfer into the orogenic crust for Buchan-type metamorphism at the high thermal gradients. At the same time, asthenospheric doming develops to place significant compressional stresses on the adjacent continental margins (Mondy et al., 2018). For this reason, the compression prevails in the deeper crust with higher P-T conditions for high-grade metamorphism, whereas the extension would prevail in the shallower crust with lower P-T conditions for low-grade metamorphism. In either case, the direction of both strain action and material transport is deep to shallow rather than shallow to deep, the latter being characteristic of subduction for active compression at the low to moderate thermal gradients.

In contrast, passive rifts were defined to develop in response to lithospheric extension by far-field stresses. In this case, the extension is homogeneous at different depths of the lithosphere that is thinned in the two sides (McKenzie, 1978; McKenzie and Bickle, 1988; Nicolas et al., 1994). As soon as the lithosphere is considerably thinned by extension, the space is produced for the asthenospheric upwelling, with possible protrusion onto the lithospheric base for both heating and compression. Thus, the lithospheric extension by passive rifting is synchronous with the lithospheric thinning to cause the asthenospheric upwelling. This differs from the active rifts where the lithospheric extension due to the asthenospheric upwelling postdates the lithospheric thinning. In either case, high heat flow is transferred by the asthenospheric doming into the thinned lithosphere, leading to high-grade metamorphism at higher thermal gradients. Furthermore, a transition from passive to active rifting can result from changes in asthenospheric buoyancy forces due to localized thinning of the lithosphere (Huismans et al., 2001).

**BARROVIAN-TYPE METAMORPHISM**

Moderate- T/P Barrovian-type metamorphism generally proceeds at moderate thermal gradients of 11–30 °C/km (Figs. 2 and 3), equivalent to intermediate thermobaric ratios between 370 °C/GPa and around 1000 °C/GPa. Its upper T/P...
limit is defined by the andalusite-sillimanite boundary at lower temperatures and by the kyanite-sillimanite boundary at higher temperatures, whereas its lower $T/P$ limit is defined by the metamorphic reaction of albite to jadeite and quartz. The Barrovian facies series is not the product of extreme metamorphism in the Phanerozoic in terms of the thermal gradients for modern subduction zones (Fig. 5). Nevertheless, Barrovian facies series comprising medium-$P$ (MP) amphibolite and HP granulite have been paired with high-$T/P$ Buchan facies series comprising LP amphibolite and granulite to indicate the operation of plate tectonics since the Archean (Brown, 2006, 2007, 2010, 2014). In doing so, the two facies series are regarded as the products of regional metamorphism at contrasting thermal gradients along subduction zones. For this reason, it is necessary to explain the petrogenetic relationship between the two metamorphic facies series.

**Historical Evolution**

Barrovian facies was originally defined to represent a sequence of metamorphic mineral reactions that were recorded by successive mineral assemblages in metapelites from the Barrovian terrain in northeastern Scotland (Barrow, 1893, 1912). This sequence has been intensively investigated for its petrology in one century (e.g., Tilley, 1925; Read, 1927, 1952; Kennedy, 1948; Chinner, 1960, 1966; Harte and Johnson, 1969; Richardson and Powell, 1976; Atherton, 1977; Harte and Hudson, 1979; Hudson, 1980, 1985; Baker and Droop, 1983; Hudson and Kearns, 2000; Viets et al., 2010; Vorhies and Ague, 2011; Carty et al., 2012; Aoki et al., 2014; Johnson et al., 2015; Palin et al., 2018). The results consistently confirm that the classical Barrovian sequence was packaged to start from chlorite, biotite, and garnet in the low-grade rocks, through staurolite and kyanite to sillimanite (with muscovite) in the medium-grade rocks, to sillimanite + K-feldspar (without muscovite) and garnet + corundite + K-feldspar in the high-grade rocks. Although these minerals and rocks do represent an increase in metamorphic grade from greenschist facies to amphibolite facies (Figs. 1 and 2), the appearance and disappearance of kyanite in this sequence essentially indicate its two-stage evolution from compression heating to decompression heating. In particular, the appearance and uniqueness of sillimanite indicate completeness of the kyanite-sillimanite transition during decompression heating. In this regard, the kyanite-sillimanite association is characteristic of the metamorphic transformation from Barrovian facies to Buchan facies during decompression in association with extension. The polymorphic transition from kyanite to sillimanite in the type sequence is consistent with a steady-state geotherm of 28 °C/km, as calculated by Richardson and Powell (1976) in their modeling of possible heat flow in Daldrian metamorphism. This has challenged the traditional interpretation of Barrovian facies as the metamorphic product of crustal thickening thus causing compressional heating (e.g., Hietanen, 1967; Miyashiro, 1973; Brown, 1993, 2001; Stiwe, 2007).

Later on, eclogites with metamorphic thermobaric ratios of 368–518 °C/GPa were found in Barrovian facie zones (Table 1), looking as if these eclogites could have been produced by Barrovian-type metamorphism at elevated thermal gradients of 11–15.5 °C/km. The term “supra-Barrovian” metamorphism was thus used to denote the retrograde overprinting at crustal depths after UHP metamorphism at mantle depths (Walsh and Hacker, 2004; Walsh et al., 2007). The highest thermobaric ratio of 518 °C/GPa was obtained for eclogite inclusions hosted by migmatites in the cores of gneiss domes at Montagne Noire in the French Massif Central (Whitney et al., 2015). However, phase equilibrium calculations yielded metamorphic conditions of ~680 °C and 1.6–1.7 GPa for the dome-margin eclogite (Whitney et al., 2020). This results in metamorphic thermobaric ratios of 400–425 °C/GPa, corresponding to metamorphic thermal gradients of 12.0–12.8 °C/km. Inspection of these moderate-$T/P$ eclogites indicates that they are mostly associated with granulite-facies rocks in the field and thus underwent variable degrees of retrogression or overprinting at granulite facies (e.g., O’Brien and Rotzler, 2003; Kwon et al., 2009; Corrie et al., 2010; Cruciani et al., 2012; Wang et al., 2017; Loose and Schenk, 2018; Liu et al., 2021). Therefore, they belong to the B-type eclogite (Coleman et al., 1965), equivalent to HT eclogites (Carswell, 1990; Agard et al., 2009, 2018) or HP granulite-associated eclogites (Brown and Johnson, 2018, 2019a, 2019b). As such, their metamorphic $T/P$ ratios were elevated by granulate facies overprinting. This does not mean their original formation by warm subduction at the moderate thermal gradients.

It is known from metamorphic petrology that eclogitization of mafic rocks is realized by the disappearance of plagioclase through two metamorphic reactions at high pressure (e.g., Spear, 1993; Bucher and Frey, 1994; O’Brien and Rotzler, 2003). The first is the decomposition of albite in plagioclase to form jadeite and quartz, with incorporation of the jadeite component into omphacite; it starts under $P-T$ conditions along thermal gradients of 10–11 °C/km (Fig. 1). The second is the decomposition of anorthite in plagioclase to form grossular, kyanite, and quartz, with incorporation of the grossular component into garnet. This reaction takes place under somewhat variable $P-T$ conditions below the first reaction curve, depending on the composition of metamorphic lithology (e.g., De Paoli et al., 2009, 2012). Because the formation of omphacite is the key to eclogitization, the first metamorphic reaction is used to mark the boundary between Alpine and Barrovian facies (Figs. 2 and 3). For rocks of mafic composition, the thermobaric transition from eclogite- to granulate-facies conditions results in the production of secondary plagioclase domains through the decomposition of both garnet and clinopyroxene. Nevertheless, such plagioclase domains have distinctly different compositions in each of the reaction domains (O’Brien and Rotzler, 2003). In addition to the common occurrence of coronitic, symplectitic, and kelyphitic reaction textures in such overprinted rocks, a common textural feature in the granulitized eclogites is the replacement of omphacite by symplectitic intergrowths of sodic plagioclase and clinopyroxene with lower Na and Al contents than the initial clinopyroxene (e.g., O’Brien and Rotzler, 2003; Corrie et al., 2010; Wang et al., 2017; Liu et al., 2021).

The occurrence of omphacite in moderate-$T/P$ eclogites is prominent at convergent plate boundaries (Table 1), indicating the metastable survival of
TABLE 1. ECLOGITES SHOWING MODERATE THERMOBARIC RATIOS FOR BARROVIAN-TYPE METAMORPHISM AT CONVERGENT PLATE BOUNDARIES

| Location                                                      | Age (Ma) | $T$ (°C) | $P$ (GPa) | $T/P$ (°C/GPa) | Thermal gradient (°C/km) | Reference |
|---------------------------------------------------------------|----------|----------|-----------|----------------|------------------------|-----------|
| Wandamen Peninsula, western Papua New Guinea                  | 8.1      | 770      | 2.0       | 385            | 11.6                   | François et al. (2016) |
| Greater Himalaya, Dinge, China                                | 14       | 760      | 2.0       | 380            | 11.4                   | Wang et al. (2017)     |
| Greater Himalaya, Arun Valley, eastern Nepal                  | 21       | 670      | 1.5       | 447            | 13.4                   | Corrie et al. (2010)   |
| Gyeonggi Massif, South Korea                                  | 231      | 840      | 2.0       | 420            | 12.6                   | Kwon et al. (2009)     |
| Cabardes in Montagne Noire, French Massif Central             | 315      | 680      | 1.7       | 400            | 12.0                   | Whitney et al. (2020)  |
| Termé de Fourcaric in Montagne Noire, French Massif Central   | 315      | 725      | 1.4       | 518            | 15.5                   | Whitney et al. (2015)  |
| Punta de Il Tulchi, Sardinia                                  | 403      | 700      | 1.9       | 368            | 11.0                   | Cruciani et al. (2012) |
| Baeudong, Hongseong, southern Gyeonggi, South Korea           | 405      | 1000     | 2.3       | 435            | 13.0                   | Oh et al. (2014)        |
| Bakersville, Eastern Blue Ridge, Southern Appalachians, United States | 460   | 700      | 1.6       | 438            | 13.1                   | Page et al. (2003)      |
| Guabaowan, Chinese Beishan, southern Altaiids                 | 465      | 755      | 1.6       | 472            | 14.2                   | Qu et al. (2011)        |
| Ubendian Belt, Tanzania                                       | 593      | 740      | 1.8       | 411            | 12.3                   | Boniface and Schenk (2012) |
| Eastern segment, Sveconorwegian orogen, southwest Sweden      | 915      | 870      | 1.8       | 483            | 13.0                   | Tual et al. (2017), Möller et al. (2015) |
| Glenelg-Attadale inlier, northwest Scotland                   | 1000     | 1000     | 2.5       | 400            | 12.0                   | Sajeev et al. (2010)    |
| Glenelg-Attadale inlier, northwest Scotland                   | 1010     | 780      | 2.0       | 390            | 11.7                   | Storey et al. (2005)    |
| Ubendian Belt, Tanzania                                       | 1870     | 700      | 1.5       | 467            | 14.0                   | Boniface et al. (2012)  |
| Athabasca granulite terrane, Snowbird tectonic zone, Canada   | 1904     | 800      | 1.6       | 500            | 15.0                   | Baldwin et al. (2004)   |
| Usagararan Belt, Tanzania                                    | 1999     | 750      | 1.8       | 417            | 12.5                   | Collins et al. (2004)   |
| Nyong complex, Ebunian orogen                                 | 2000     | 800      | 1.9       | 421            | 12.6                   | Loose and Schenk (2018)  |

Note: Barrovian-type metamorphism proceeds at moderate thermal gradients of 11–30 °C/km, equivalent to moderate thermobaric ratios of 370 to 1000 °C/GPa. $T$—temperature; $P$—pressure.
of metamorphic zircon is much more prominent at granulite facies than at eclogite facies. As a consequence, zircon U-Pb dating yields the younger ages of 25–13 Ma at granulite facies compared to 30–17 Ma at eclogite facies (Li et al., 2019a; Wang et al., 2021). Petrological studies indicate that these eclogites were originally produced by eclogitization at ~720–770 °C and >20–21 kbar and then underwent HT to UHT granulite-facies overprinting at 750–970 °C and 6–11 kbar (Wang et al., 2017, 2021; Li et al., 2019a). This corresponds to an increase of metamorphic thermal gradients from ~10.8–11.0 °C/km for Alpine facies series to >26.5–37.5 °C/km for Buchan facies series (Fig. 12). For those exhumed HP to UHP eclogites in the bulk Himalayan orogen (e.g., Najman et al., 2010; Grujic et al., 2011; Jain, 2014; Kohn, 2014; Zheng and Wu, 2018; Wang et al., 2021), it is possible that eclogite-facies P-T conditions would have been maintained for variable time scales from ~5 to 30 m.y. They may have started from continental subduction resulting in compressional heating to the eclogite facies at 55 ± 10 Ma and terminated upon the onset of extensional heating to cause granulite-facies overprinting at variable ages from 30 to 17 Ma. As a consequence, the metamorphic superimposition is spatially and temporally associated with lithospheric extension in the collisional orogen. Therefore, the timing of peak granulite facies overprinting at the highest thermal gradients may provide a temporal proxy for the climax of lithospheric extension after its thinning along the convergent plate boundary.

The conventional viewpoint holds that Barrovian-type amphibolite-facies metamorphic rocks were produced by burial heating to crustal depths where the normal geotherm of 15 °C/km prevails. This corresponds to warm subduction at the moderate thermal gradients (Figs. 1 and 2). As a consequence, the disappearance of epidote and the increase of calcium in plagioclase are interpreted as the result of an increase in the metamorphic intensity through the metamorphic facies during prograde subduction. This leads to the illusion as if water was lost from their protolith during the prograde amphibolite-facies metamorphism. As amphibole is a hydrous mineral, its formation requires significant incorporation of water into the metamorphosing rocks under amphibolite-facies conditions. In this regard, amphibolite-facies rocks can be produced either by prograde metamorphism at moderate thermal gradients from greenschist facies with considerable dehydration or by retrograde metamorphism from eclogite facies or granulite facies with significant hydration (metasomatism). In the latter case, the water can be derived either from decompressional dehydration of UHP eclogite-facies rocks during their exhumation to moderate thermal gradients (Zheng et al., 2009) or from thermal dehydration of crystalline basement rocks while their underlying rocks are heated during granulite-facies metamorphism at high thermal gradients (Zheng and Chen, 2017).

As both amphibolite and granulate of the Barrovian facies series can be produced by MP to HP metamorphism in the kyanite stability field, they are readily distinguished from their counterparts in the Buchan facies series produced by LP metamorphism in the sillimanite stability field. Nevertheless, it is important to distinguish the prograde amphibolite from the retrograde amphibolite in Barrovian facies series. On the other hand, the eclogite of Alpine facies series may keep metastable with partial Fe-Mg exchange between garnet and omphacite during Barrovian-type metamorphism, resulting in the illusion that eclogite could be produced under P-T conditions below the metamorphic reaction boundary of albite to jadeite and quartz. In reality, the eclogite would be retrograded to garnet amphibolite at amphibolite facies and to garnet pyroxenite at granulate facies (Fig. 1). Therefore, it is critical to distinguish the primary eclogite from the metastable eclogite in Barrovian facies series.

**Prograde and Retrograde Metamorphism**

The traditional model for the origin of Barrovian metamorphism assumes thermal equilibration of the overthickened crust through burial heating during continental collision (e.g., Richardson and Powell, 1976; England and Richardson, 1977; England and Thompson, 1984; Thompson and Ridley, 1987; Brown, 1993; Jamieson et al., 1998). This is responsible for prograde Barrovian metamorphism along the normal geotherm, corresponding to top-down compressional heating in the early stage of collisional orogenesis (Figs. 7A and 11A). For this reason, the traditional definition of Barrovian facies accords with a synchronous increase in metamorphic P-T conditions within the kyanite stability field (Zheng and Chen, 2017). This leads to the definition of moderate-P Barrovian facies series with a given range of thermal gradients that are bounded between Alpine and Buchan facies series (Fig. 2). Such P-T conditions...
correspond to the intermediate- $T/P$ type of Brown and Johnson (2019a, 2019b) in their studies of bimodal metamorphism. As a consequence, the Barrovian facies series is usually used to represent a prograde sequence from greenschist facies at low pressures through amphibolite facies at medium pressures to granulite facies at high pressures (Figs. 1 and 2). However, more and more studies indicate the occurrence of Barrovian facies series in retrograded UHP metamorphic rocks on the one hand (e.g., Walsh and Hacker, 2004; Zheng, 2021b) and in metamorphic core complexes on the other hand (e.g., Whitney et al., 2013; Platt et al., 2015; Searle and Lamont, 2020).

With respect to tectonothermal mechanisms, Barrovian metamorphism is usually ascribed to the burial heating of crustal rocks during continental collision (e.g., Richardson and Powell, 1976; England and Richardson, 1977; England and Thompson, 1984; Thompson and Ridley, 1987; Brown, 1993; Jamieson et al., 1998). Both steady-state and transient heat flows were taken into account in this model, which is simple and classic but only valid for prograde Barrovian-type metamorphism in view of crustal subduction at moderate thermal gradients (Fig. 2). Except for subduction initiation through ridge-trench conversion along divergent plate boundaries (Agard et al., 2016), such warm subduction is generally lacking in the Phanerozoic, but it prevailed in the Archean and Paleoproterozoic (e.g., Anderson et al., 2012; Brown, 2014; Chowdhury and Chakraborty, 2019; Holder et al., 2019; Zheng and Zhao, 2020). This difference can be ascribed to high mantle temperatures in the Archean as compared to those in the Phanerozoic (Holder et al., 2019; Zheng and Zhao, 2020). High mantle temperature was prominent in the Archean (Herzberg et al., 2010; Korenaga, 2013; Ganne and Feng, 2017; Aulbach and Arndt, 2019), making the operation of plate tectonics unlike the modern style (e.g., Cawood et al., 2018; Brown and Johnson, 2018; Stern, 2018; Holder et al., 2019; Zheng and Zhao, 2020). In this regard, prograde Barrovian-type metamorphism may be common in Archean high-grade terranes. Following the thermal models of Syracuse et al. (2010) for modern subduction zones (Fig. 5), in contrast, the prevalence of low thermal gradients makes prograde Barrovian-type metamorphism uncommon in the Phanerozoic.

On the other hand, there is also superimposition of HP granulite-facies metamorphism at temperatures >900 °C on eclogite-facies UHP metamorphic rocks in continental subduction zones, such as the Bohemian Massif in central Europe (Kotková et al., 2011; Haßler and Kotková, 2016; Faryad and Cuthbert, 2020), the Kokchetav Massif in Kazakhstan (Schertl and Sobolev, 2013; Stepanov et al., 2016), and South Altyn in western China (Dong et al., 2018, 2019). Such polymetamorphism records a dramatic increase in metamorphic thermal gradients from low to moderate (Fig. 1), corresponding to decompressional exhumation of the UHP slices from mantle to crustal depths. If this decompressional tectonics should take place at converging plate boundaries, where ongoing subduction zones would change their geotherms from cold to warm over a short time scale (e.g., van Hunen and Allen, 2011; Magni et al., 2017; Garzanti et al., 2018), a possible mechanism for the retrograde Barrovian-type metamorphism at temperatures >900 °C may be slab breakoff (Zheng and Chen, 2017) or slab rollback (Sizova et al., 2019). In either case, the HP granulite-facies superimposition would be subsequent to the eclogite-facies UHP metamorphic event. On the other hand, there are collisional orogens where HP granulite-facies superimposition significantly postdates HP to UHP eclogite-facies metamorphism, such as in the Dabie orogen (Gao et al., 2017) and the Tongbai orogen (Zhang et al., 2020). This points to continental rifting as the mechanism responsible for metamorphic superimposition in continental interiors (Zheng and Chen, 2017). As a consequence, eclogites were completely retrograded to garnet pyroxenites during the granulite-facies superimposition, with only rare survival of omphacite as crystal inclusions inside refractory minerals such as garnet and zircon.

In nature, Alpine-type UHP metamorphic rocks were generally exhuming at moderate thermal gradients, resulting in their decompressional dehydration causing Barrovian-type superimposition at amphibolite to granulite facies. This gives rise to a kind of retrograde Barrovian facies series in the late stage of collisional orogenesis (Fig. 7B). As illustrated by the thermobarometric study of Walsh and Hacker (2004) for HP to UHP metamorphic rocks in the Western Gneiss Region of Norway, the decompressional exhumation transformed the UHP eclogite-facies rocks to HP amphibolite- to granulite-facies rocks. There is widespread evidence for decompressional melting of the UHP rocks during their exhumation (Kylander-Clark and Hacker, 2014). Such retrograde HP amphibolite- to granulite-facies overprinting is common in UHP metamorphic rocks in the Dabie-Sulu orogenic belt, where HP to UHP eclogite-facies metamorphism occurred during compressional heating in the Middle Triassic, whereas HP amphibolite- to granulite-facies retrogression took place subsequent to decompressional exhumation of the UHP slices to Moho depths in the Late Triassic (Zheng et al., 2009, 2019). In either case, moderate thermal gradients prevail at Moho depths in collisional orogens, causing the exhumed UHP slices to undergo retrograde Barrovian-type retrogression by internally derived liquid phases (such as aqueous solutions and hydrous melts) at granulite facies (outflux) and amphibolite facies (influx). Such metamorphic superimposition is common in collisional orogenesis (e.g., Cawood et al., 2010; Korenaga, 2013; Ganne and Feng, 2017; Aulbach and Arndt, 2019), orogenic crust by asthenospheric heating, providing widespread availability of liquid phases during the Buchan-type HT to UHT metamorphism. This kind of retrograde Barrovian facies series was contemporaneously formed with Buchan-type metamorphism at different thermal gradients in an extensional regime (Viete et al., 2013; Zheng and Chen, 2017). This is consistent with the common observations that retrograde Barrovian facies often coexist with Buchan facies in the same orogen, such as the Scottish Caledonides (Oliver et al., 2000; Baxter et al., 2002; Viète et al., 2013), New England (Florence et
The contemporaneity of Barrovian and Buchan metamorphism in the Scottish Caledonides is demonstrated by zircon U-Pb dating (Oliver et al., 2000) and garnet Sm-Nd isochron dating (Baxter et al., 2002). On the other hand, it is common for Barrovian metamorphism to predate Buchan metamorphism (e.g., Langone et al., 2010; Peixoto et al., 2018; de Hoym de Marien et al., 2019). This indicates superimposition of Barrovian facies by Buchan metamorphism. Whereas the contemporaneity between Barrovian and Buchan metamorphism indicates a retrograde Barrovian facies series, the sequential Barrovian to Buchan metamorphism points to prograde Barrovian metamorphism. In view of the current debates on the origin of Barrovian metamorphism (e.g., Jamie- son et al., 1998; Viete et al., 2013; Ryan and Dewey, 2019), it is critical to make the distinction between prograde and retrograde Barrovian facies series in collisional orogens. Therefore, caution must be exercised when constraining the timing of collisional orogeny by dating Barrovian facies rocks in poly-metamorphic terranes with superimposition by Buchan-type metamorphism.

It is also common for Alpine to Barrovian facies series to be superimposed by Buchan-type metamorphism in the postcollisional stage. This results in such occurrences where HP to UHP eclogites and HP granulite-facies rocks are hosted by felsic gneisses with domical structures in collisional orogens, such as the Western Gneiss Region in Norway (Cuthbert et al., 2000; Walsh and Hacker, 2004), the European Variscides (e.g., O’Brien and Carswell, 1993; Stipska et al., 2008; Gaggero et al., 2009; Whitney et al., 2015), the North Qaidam in western China (Mattinson et al., 2006), the Dabie-Sulu in eastern China (Wang et al., 2011; Groppo et al., 2015; Ji et al., 2017), the Asian Himalayas (Groppo et al., 2007; Cottle et al., 2009; Corrie et al., 2010), and the Woodlark Rift in Papua New Guinea (Baldwin et al., 2008; Little et al., 2011). Host gneisses typically do not record eclogite-facies HP to UHP metamorphic conditions, leading to long-standing controversy about the relationship between eclogite inclusions and host gneiss (Eskola, 1921; Lappin and Smith, 1978; Brueckner, 2018; Whitney et al., 2020). Geochronological studies indicate that high-grade metamorphic rocks in collisional orogens may experience two to three episodes of polymetamorphism at contrasting thermal gradients. Therefore, the superimposition of polymetamorphism involves the tectonic transition from compressional heating during collisional orogeny to extensional heating during rift events such as delamination or extensional heating, or thermal conditions (Fig. 1). This process may take place during delamination of the lithosphere, cooling, or heating conditions. In either mechanism, the lithospheric thinning at convergent plate boundaries can be indicated by the poly-metamorphic transition from kyanite to sillimanite in amphibolite- to granulite-facies metamorphic rocks.

### BIMODAL METAMORPHISM AND PLATE TECTONICS

#### Bimodal Metamorphism

Bimodal metamorphism is defined for two types of regional metamorphism at contrasting thermal gradients and dynamic regimes along convergent plate boundaries. One would proceed at lower thermal gradients in association with compressional heating and the other would proceed at higher thermal gradients in association with extensional heating (Zheng, 2021a). According
TABLE 2. DIFFERENCES BETWEEN TWO TYPES OF EXTREME METAMORPHISM AT CONVERGENT PLATE BOUNDARIES

| Property                  | Alpine type                           | Buchan type                           |
|---------------------------|---------------------------------------|----------------------------------------|
| Metamorphic facies        | Blueschist to eclogite                | Kyanite-absent amphibolite to granulite |
| Metamorphic pressure      | High to ultrahigh                     | Low to moderate                        |
| Metamorphic temperature   | Low to moderate                       | High to ultrahigh                      |
| Thermobaric ratios        | 165–370 °C/GPa                        | 1000–2000 °C/GPa                      |
| Thermal gradients         | 5–11 °C/km                            | 30–60 °C/km                            |
| Strain regime             | Compression                           | Extension                              |
| Rheology of plate margins| Rigid                                 | Ductile                                |
| Width of metamorphic zone| Small                                 | Large                                  |
| Strength of metamorphic zone | Low                                 | High                                   |

Paired metamorphic belts were originally found in circum-Pacific subduction zones (Fig. 13), such as the Sanbagawa and Ryoke belts in Japan, where an outboard low-T/P Alpine facies series is juxtaposed against an inboard high-T/P Buchan facies series (Miyashiro, 1961, 1973). The duality of metamorphic belts was extended to the duality of thermal state at convergent plate boundaries (Oxburgh and Turcotte, 1970, 1971). Recognizing such a duality is very important in dealing with the tectonic evolution of subduction zones, where the lower-plate material registers the low thermal gradients, whereas the upper-plate material registers the high thermal gradients. Therefore, the occurrence of paired metamorphic belts was used to mark the operation of plate tectonics (Dewey and Bird, 1970; Katz, 1972; Miyashiro, 1972).

Originally, bimodal metamorphism at convergent plate boundaries, pairing low-T/P Alpine facies series with high-T/P Buchan facies series (Miyashiro, 1961, 1973), appeared to result from the difference in compressional heating between two converging plates, generating cold geotherms, and extensional heating, still between two converging plates, giving rise to hot geotherms (Table 2). This leads to Alpine-type metamorphism outboard and Buchan-type metamorphism inboard. Later on, moderate-T/P Barrovian facies belts were paired with high-T/P Buchan facies belts in both accretionary and collisional orogens, highlighting the duality of metamorphic belts as the hallmark of plate tectonics since the Archean (Brown, 2006, 2007, 2010; Brown and Johnson, 2018, 2019a; Holder et al., 2019). In doing so, the bimodal metamorphism was associated with the duality of thermobaric ratios along convergent plate boundaries.
a feature that is regarded as the hallmark of one-sided subduction in plate tectonics (Brown, 2006; Gerya et al., 2008). In either case, the Buchan facies series is one of the indispensable components in bimodal metamorphic belts.

In paired metamorphic belts, the cold belts were generated on the trench side with low thermal gradients for low-T/P, Alpine-type, blueschist- to eclogite-facies metamorphism, whereas the hot belts were created in the backarc side with high thermal gradients for high-T/P, Buchan-type, amphibolite- to granulite-facies metamorphism (Fig. 13). As a consequence, the two belts were produced by the spatial juxtaposition of metamorphic terranes with different facies series along ongoing subduction zones. Later on, the difference in metamorphic facies series was also found in fossil subduction zones that have been converted into collisional orogens (Dewey and Bird, 1970), typically between the Alpine and Hercynian orogens, leading to the dichotomous classification of orogens as either Alpinotype or Hercynotype (Zwart, 1969). Further studies indicated that such polymetamorphic belts were produced by superimposition of metamorphic facies series at convergent plate boundaries, one through cold subduction for lower-T/P metamorphism along converging plate boundaries (Fig. 14A) and the other through hot rifting for higher-T/P metamorphism along converging plate boundaries (Fig. 14B). Therefore, the contrasts in metamorphic thermal gradients for the products of bimodal metamorphism are also the reflection of both dynamic regime and thermal state at convergent plate boundaries.

Polymetamorphism is prominent in the Woodlark Rift in eastern Papua New Guinea, which is the only place on Earth today where an active mid-ocean ridge terminates into a zone of continental rifting, low-angle normal faults are seismically active, and HP to UHP eclogite-facies metamorphic rocks of Pliocene age are exposed on the surface (Little et al., 2007; Ellis et al., 2011; Wallace et al., 2014). The youngest known UHP terrane on Earth is being actively exhumed at the active boundary between the Australian and Pacific plates (Baldwin et al., 2008, 2012). The Papuan UHP terrane was produced at low thermal gradients by cold subduction of the Australian continental margin to mantle depths. The UHP metamorphic rocks were overprinted at granulite facies to form metamorphic core complexes in the D’Entrecasteaux Islands of the Woodlark Rift (Little et al., 2007, 2011). It appears that active rifting has resulted in the granulite-facies superimposition at high thermal gradients. Therefore, the Woodlark Rift is a polymetamorphic belt that was produced by superimposition of higher-T/P metamorphic facies series on lower-T/P facies series at the active plate boundary. The metamorphic superimposition provides evidence for the change in the dynamic regime from compression during subduction to extension during rifting at the convergent plate boundary.

**Implications for the Operation of Plate Tectonics**

The traditional paired metamorphic belts were established in the circum-Pacific accretionary orogens of Phanerozoic age, where cold subduction is registered by Alpine facies series in the early stage of oceanic subduction zones, whereas hot rifting is recorded by Buchan facies series in the late stage of oceanic subduction zones. Therefore, cold subduction for blueschist facies and UHP eclogite facies is viewed as a characteristic feature of modern plate tectonics (Stern, 2005, 2007, 2018). On the other hand, Barrovian facies series have widely appeared since the Archean (Brown, 2006, 2007, 2010; Brown and Johnson, 2018; Holder et al., 2019), indicating the operation of warm subduction and thus ancient plate tectonics (Zheng and Zhao, 2020). In either case, hot rifting is a necessary process for the operation of plate tectonics since the Archean. Substantially, bimodal metamorphism at convergent plate boundaries manifests the two fundamentally different types of mountain building at the contrasting thermal gradients (Fossen et al., 2017; Zheng and Chen, 2017; Will et al., 2018). In this context, the diachronous property of bimodal metamorphism is evident in both ongoing and fossil subduction zones. However, this diachronous property was not obvious in the last century because of the low resolution in geochronology between bimodal metamorphic events.

Although the paired metamorphic belts in accretionary orogens are not produced at the same time (Miyashiro, 1994; Brown, 1998, 2010), bimodal metamorphism would have proceeded sequentially in the lifetime of oceanic subduction beneath continental margins. In other words, Alpine-type

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**Figure 14. Schematic cartoons showing the formation of polymetamorphic belts along convergent plate boundaries:**

**(A)** Cold subduction for metamorphism at lower thermal gradients

**(B)** Hot rifting for metamorphism at higher thermal gradients

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**Additional Information**

- **A. Cold subduction for metamorphism at lower thermal gradients**
  - Ocean
  - Continent
  - HP
  - UHP
  - P (GPa)
  - T (°C)
  - 300 600 900

- **B. Hot rifting for metamorphism at higher thermal gradients**
  - Continent
  - Continent
  - HT
  - UHT
  - P (GPa)
  - T (°C)
  - 300 600 1000

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**Research Paper**

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metamorphism would take place during cold subduction to cause compressional heating in the early stage (Fig. 7A), whereas Buchan-type metamorphism would occur during hot rifting to cause extensional heating in the late stage (Fig. 7B). On the other hand, the polymetamorphic belts in collisional orogens are produced at different ages (Zheng and Chen, 2017), where Alpine- to Barrovian-type metamorphism proceeds during cold to warm subduction to cause compressional heating in the collisional stage (Fig. 11A), whereas Buchan-type metamorphism proceeds during hot rifting to cause extensional heating in the postcollisional stage (Fig. 11B). In effect, bimodal P–T trajectories represent metamorphic field gradients in bimodal metamorphic belts (Brown, 1998, 2010, 2014). In particular, polymetamorphism is common along fossil subduction zones where the lower-T/P Alpine to Barrovian facies rocks were superimposed by the higher-T/P Buchan-type metamorphism. Therefore, bimodal metamorphism is primarily dictated by secular changes in both thermal state and dynamic regime at convergent plate boundaries.

Paired metamorphic belts were originally assumed to form contemporaneously during accretionary orogeny (Miyashiro, 1961, 1973). However, more and more studies indicate that low-T/P Alpine-type metamorphism considerably predates high-T/P Buchan-type metamorphism along converging plate boundaries (Brown, 2006, 2007, 2010). This type of bimodal metamorphism in separate belts can be regarded as the hallmark of one-sided subduction (Brown, 2006; Gerya et al., 2008). It is temporally and spatially associated with accretionary orogeny above oceanic subduction zones. In general, accretionary orogeny starts with the formation of accretionary wedges and subsequent low-T/P metamorphism, and it terminates with continental arc magmatism and subsequent high-T/P metamorphism. In this regard, the duration of paired metamorphism provides a proxy for the time scale of accretionary orogeny. Nevertheless, it is more common for bimodal metamorphism to be diachronous along converged plate boundaries, where the high-T/P metamorphism during rifting orogeny was spatially superimposed on low- to moderate-T/P metamorphic facies series produced by collisional orogeny (Zheng and Chen, 2017). This leads to polymetamorphism along fossil subduction zones. In either case, the products of bimodal metamorphism in the geological record are the characteristic imprint of plate tectonics through the duality of gravity-driven subduction (lithospheric downwelling) and buoyancy-driven rifting (asthenospheric upwelling) along plate boundaries (Zheng and Zhao, 2020).

Interrogation of geological records for bimodal metamorphism at convergent plate boundaries indicates that Alpine-type metamorphism predates Buchan-type metamorphism in both accretionary and collisional orogens. Nevertheless, there is no spatial superimposition of the paired metamorphic belts in accretionary orogens. In contrast, the cold and hot orogens were generated one after the other, with the spatial superimposition of granulite facies on eclogite facies in collisional orogens, leading to polymetamorphic assemblages. This is exemplified in the Himalayan orogen (Zheng and Wu, 2018) and the Dabie orogen (e.g., Zheng et al., 2019). Although the contrasting tectonic imprints in the metamorphic rocks represent discrete events due to changes in the thermal gradient, bimodal metamorphism is indeed responsible for the duality of dynamic regime at convergent plate boundaries (Zheng and Chen, 2017; Zheng and Zhao, 2017, 2020). In this regard, polymetamorphism can be viewed as one of the essential characteristics of plate tectonics with respect to gravity-driven subduction and buoyancy-driven rifting at convergent plate boundaries (Zheng and Zhao, 2020).

The operation of plate tectonics is characterized by the duality of both dynamic regime and thermal state along plate boundaries (Zheng and Zhao, 2020). One aspect is compressional heating at lower thermal gradients during plate convergence due to either oceanic subduction for accretionary orogeny or continental subduction for collisional orogeny (Cawood et al., 2008; Zheng and Chen, 2017), resulting in the asymmetric distribution of metamorphic products along converging plate boundaries. The other aspect is extensional heating at higher thermal gradients during the attempted plate divergence for rifting orogeny, either at backarc sites in the hinterland or at continental rifts along fossil suture zones (Zheng and Chen, 2017; Zheng and Zhao, 2017), leading to the symmetric distribution of metamorphic products along convergent plate boundaries. This duality is recorded in paired metamorphic belts, in which the high-T/P Buchan facies series was originally paired with the low-T/P Alpine facies series in accretionary orogens (Miyashiro, 1961, 1973), and later on with moderate-T/P Barrovian facies series in collisional orogens (Brown, 2006, 2007, 2010). In either case, their formation along convergent plate boundaries is diachronous in the time, but their spatial distribution may be separate or superimposed at convergent plate boundaries.

By using bimodal metamorphism as a proxy for secular change in the style of plate tectonics, Holder et al. (2019) made a statistical study of peak T/P ratios for bimodal metamorphic belts. Their results indicate that the style of plate tectonics has gradually changed with secular cooling of the mantle from the Archean through the Proterozoic to the Phanerozoic. In order to validate their methodology in identifying the operation of plate tectonics from the ancient rock record, Holder et al. (2019) firstly quantified the bimodal distribution of metamorphic T/P ratios in the past 200 Ma. Such a distribution is manifested in bimodal metamorphic belts on modern Earth and is considered a diagnostic feature of plate tectonics. Afterward, they evaluated the emergence of bimodal metamorphism through a statistical evaluation of the distributions of metamorphic T/P ratios through time. Their results indicate neither an abrupt transition in the style of plate tectonics in the Neoproterozoic nor the operation of modern plate tectonics since the Paleoproterozoic. Nevertheless, Holder et al. (2019) did find that the statistical distribution of metamorphic T/P ratios has gradually become wider and more distinctly bimodal from the Neoarchean to the Cenozoic, and the average metamorphic T/P ratios have decreased since the Paleoproterozoic. In detail, the distribution of metamorphic T/P ratios is wide and less distinctly bimodal in the Archean, whereas that is narrow and more distinctly bimodal in the Phanerozoic. This difference points to a change in the style of plate tectonics from the Archean through the Proterozoic to the Phanerozoic. As generalized by Zheng and Zhao (2020), modern plate tectonics prevails since the Neoproterozoic (1.0 Ga–present) when converging plate boundaries are characterized by cold subduction at low thermal gradients for
Alpine-type metamorphism, whereas ancient plate tectonics would operate from the Eoarchean to Mesoproterozoic (4.0–1.0 Ga) when converging plate boundaries are characterized by warm subduction at moderate thermal gradients for Barrovian-type metamorphism. In either case, active rifting would develop along convergent plate boundaries since the Eoarchean (Zheng and Zhao, 2020), resulting in the high thermal gradients for Buchan-type metamorphism (Zheng and Chen, 2017; Zheng, 2021a).

The nature of bimodal metamorphism at convergent plate boundaries can be deciphered by an integrated study of mineralogy, petrology, geochronology, and geochemistry. Premetamorphic protoliths may be oceanic materials in oceanic subduction zones (e.g., the circum-Pacific orogens) or continental materials with transitional oceanic remnants in continental subduction zones (e.g., the Alpine-Zagros-Himalayan orogens). They may have undergone or are undergoing bimodal metamorphism in these two types of orogenic systems, which are the two major regions of lithospheric subduction into the asthenosphere (e.g., Collins, 2003; Dickinson, 2004; Ernst, 2005). If the difference in thermobaric ratio between metamorphic processes is linked to the difference in orogenic size between orogenic processes at convergent plate boundaries, this results in a conceptual model in which cold to warm orogens are relatively small, whereas hot to ultrahot orogens are relatively large (Fig. 15). Inspection of available observations indicates that the small orogens were produced at lower thermobaric ratios in the early stage of accretionary and collisional orogenies, whereas the large orogens were generated at higher thermobaric ratios in the late stage of rifting orogeny. In accretionary orogens, typically, both orogenic size and thermobaric ratio increase from the lateral growth of cold, small accretionary wedges in the early stage to the vertical growth of hot, voluminous arc magmas in the late stage of oceanic subduction (Zheng, 2021c). In collisional orogens, on the other hand, both orogenic size and thermobaric ratio increase from the thrusting of cold, small crustal slices in the synollisional stage to the uplifting of large, hot gneiss domes in the postcollisional stage (Zheng, 2021b). As a consequence, this conceptual model provides an alternative to the causal relationship between orogenic temperature and magnitude as previously modeled by Jamieson and Beaumont (2013).

By presenting a time-series analysis of thermobaric ratios of metamorphic rocks ranging in age from ca. 3.7 Ga (Eoarchean) to 15 Ma (Cenozoic), Brown et al. (2020b) found shifts in mean T/P ratios that show significant drops at 1.91 Ga, 902 Ma, 540 Ma, and 515 Ma, but rises at 1.83 Ga, 604 Ma, and 525 Ma. These mean T/P shifts can be related to changes in the thermal gradients at convergent plate boundaries, where the mean T/P drops are associated with continental convergence resulting in compressional heating, whereas the mean T/P rises are associated with continental rifting due to extensional heating. In this regard, the mean T/P drop at 1.91 Ga marks the widespread occurrence of dehydration metamorphism at lower thermal gradients during the assembly of supercontinent Columbia, whereas the mean T/P rise at 1.83 Ga indicates the widespread occurrence of anatectic metamorphism at higher thermal gradients due to abortive rifting before the breakup of Columbia. Likewise, the mean T/P drop at 902 Ma is synchronous with the assembly of supercontinent Rodinia. Nevertheless, the two mean T/P rises at 604 Ma and 525 Ma predate the two drops at 540 Ma and 515 Ma, respectively, marking thermal maxima and minima during the tectonic transition from the latest Neoproterozoic to the earliest Phanerozoic. Such warm-cold alternation may provide temporal links between the thermal state and dynamic regime of convergent plate boundaries, with important implications for the lift evolution on Earth during the Ediacaran and Cambrian.

**INSIGHTS INTO SUPERCONTINENT DYNAMICS**

The assembly and breakup of supercontinents such as Pangea, Rodinia, and Columbia (Nuna) have been studied extensively from geological records in orogens surrounding cratons (e.g., Gurnis, 1988; Murphy and Nance, 2003, 2013; Santosh and Zhao, 2009; Cawood et al., 2016; Zhao et al., 2018; Pastor-Galán et al., 2019). In general, the supercontinent cycle can be categorized by assembly through subduction for compressional tectonism and breakup through rifting for extensional tectonism. However, interpretations of these records have been obscure, leading to much confusion in the reconstruction of supercontinents. Although this involves various alternates between eustroversion and introversion (Murphy and Nance, 2008; Li et al., 2019b), the key is to discriminate the geological records between converging and converged plate boundaries. As supercontinents are composed of cratons and orogens, correctly deciphering the geological records of orogens is a key factor in deciphering the supercontinent dynamics in both time and space, unravelling the
period of supercontinent cycles, and determining links to tectonic processes along convergent plate boundaries.

In principle, supercontinent assembly and breakup are two separate processes that are coupled in time but decoupled in space with respect to the movement of tectonic plates on spherical Earth (Zheng and Zhao, 2020). While assembly is associated with lithospheric subduction in a compressional regime, breakup is associated with seafloor spreading in an extensional regime. In either case, the extensional regime is necessary, not only for partial melting of the mantle wedge to cause arc magmatism, but also for backarc and continental rifting subsequent to thinning of the overlying mantle lithosphere. Despite various debates on the geodynamic mechanism for supercontinent breakup, the following two types of hypotheses are popular (e.g., Gurnis, 1988; Storey, 1995; Murphy and Nance, 2013; Buiter and Torsvik, 2014; Cawood et al., 2016; Zilio et al., 2018): (1) the bottom-up process, driven by mantle plumes beneath the supercontinent (Fig. 16A); (2) the top-down process, driven by the development of encircling subduction zones (Fig. 16B). In either case, the loci of supercontinent breakup are dictated by the positions of preexisting suture zones showing nonrigid features in the continental lithosphere (Wilson, 1966; Yoshida, 2013; Huang et al., 2019). This gave rise to the global network of mobile belts for the operation of plate tectonics (e.g., Hakesworth and Brown, 2018; Stern and Gerya, 2018). Whereas the development of subduction zones around a supercontinent is a prerequisite for supercontinent assembly (Murphy and Nance, 2013; Merdith et al., 2019), its relationship to supercontinent breakup remains enigmatic.

It is known that the mantle wedge beneath the continental crust is significantly cooled by plate subduction with active compression in its early stage (Figs. 7A and 11A), and it is only heated in its late stage due to active extension in response to rollback of the subducting slab (Figs. 7B and 11B). Therefore, there are significant lags of hundreds of million years from assembly to breakup. For example, Rodinia assembly was mainly completed at ca. 1.1–0.9 Ga (McLelland et al., 1996; Rivers, 1997; Condie, 2003; Li et al., 2008; Bradley, 2011; Spencer et al., 2013). Although Rodinia experienced episodes of abortive rifting at 830–800 Ma, its final breakup was realized through successful continental rifting at ca. 750 Ma (Li et al., 2008; Merdith et al., 2017). There are lags of at least 150–350 m.y. from the initial assembly to the final breakup of this supercontinent.

Figure 16. Schematic cartoons illustrating the two types of mechanisms for supercontinent breakup. (A) Mantle plume–driven bottom-up model. (B) Subduction-driven top-down model.
by finding the metamorphic rocks of Buchan facies series in continental rifts (He et al., 2018; Cipar et al., 2020), but also by documenting the formation of low-$\delta^{18}$O granites in continental rifts (He et al., 2016, 2019; Zhou et al., 2018, 2020). In this context, cold to warm subduction leads to amalgamation and assembly rather than breakup and dispersal of supercontinents. As such, supercontinent assembly would variably predate the granitic magmatism associated with Buchan-type HT to UHT granulite-facies metamorphism at convergent plate boundaries. Thus, continental rifting is a key stage of the Wilson cycle between the assembly and breakup of supercontinents, because continental rifts must develop first along fossil suture zones in continental interiors and eventually result in supercontinent breakup (Zheng and Chen, 2017; Zheng et al., 2019). When a supercontinent is episodically heated by mantle upwelling along its weak zones, continental rifting can eventually develop into successful breakup; otherwise, failed (attempted) breakup occurs. Therefore, continental rifting has played a fundamental role in the destruction of supercontinents. This is also the geodynamic mechanism for rifting and breakup of continental interiors.

Although collisional orogeny is always associated with supercontinent assembly, accretionary orogeny would have developed around significant parts of a supercontinent before continental collision. Thus, supercontinent assembly would experience two-stage processes from accretionary orogeny along the hanging wall of oceanic subduction zones to collisional orogeny along the footwall of continental subduction zones. The two types of orogeny would mainly develop in the compressional regime and thus considerably predate the episodes of continental rifting in the extensional regime resulting in supercontinent breakup. As a consequence, regional metamorphism at low to moderate thermal gradients would take place during plate subduction for lithospheric thickening to form the accretionary and collisional orogens, eventually leading to supercontinent assembly. In contrast, regional metamorphism at high thermal gradients would occur in lithospherically thinned zones, such as backarc and continental rifts, to form rifting orogens, which in turn would tend to cause supercontinent breakup. This linked history between the interior and exterior of a supercontinent, together with the widespread development of rift magmatism during its breakup, points to the dominance of bottom-up processes for continental rifting (Zheng and Chen, 2017). Although backarc rifting is associated with subduction of an oceanic slab, it develops in active continental margins rather than continental interiors. In this regard, the subduction is not a direct cause for supercontinent breakup. Underplating of mantle plumes may provide a strong power for supercontinent breakup, but such a power can also be acquired from asthenospheric upwelling through active rifting along lithospheric weak zones. As such, asthenospheric upwelling along thinned orogenic lithosphere can provide sufficient heat for continental rifting at convergent plate boundaries (Thybo and Nielsen, 2009; Zheng and Zhao, 2017; Zheng, 2021a). In this regard, hot to ultrahot orogens would result from continental rifting that is responsible for supercontinent breakup (or attempted breakup) rather than assembly.

Following the advent of plate-tectonic theory in the mid-1960s, Wilson (1966) proposed that a series of orogenic processes can be regarded as a cycle that starts with lithospheric rupture and seafloor spreading, develops into passive Atlantic-type continental margins, followed by subduction of oceanic lithosphere, and terminates with continental collision. This series of processes is known as the Wilson cycle (Dewey and Burke, 1974; Nance et al., 1988; Burke, 2011; Wilson et al., 2019). No matter whether continental rifting runs into rupture, it results in Buchan-type metamorphism and the production of bimodal magmatic rocks with dominance of granites in fossil suture zones (Zheng and Zhao, 2017; Cipar et al., 2020). If continental rifting develops into rupture, it evolves to opening of a new oceanic basin (He et al., 2018). This corresponds to the first stage of a Wilson cycle in its tectonic evolution (Zheng et al., 2019), which may develop either to successful rifts for supercontinent breakup or aborted rifts for attempted breakup. In this regard, rifting orogeny in continental interiors is an early phase of the first stage in a Wilson cycle.
whereas the opening of a new ocean is a late phase of the first stage for creation of new passive continental margins. As such, a Wilson cycle is composed of the three stages of orogeny from rifting through accretionary orogeny to collisional orogeny along convergent plate boundaries (Fig. 18). This results in a refined pattern of cyclicity in the creation and destruction of supercontinents along orogenic belts in the theory of plate tectonics.

Generally speaking, the progression of a supercontinent cycle on a spherical planet like Earth is recorded through the operation of continental rifting and supercontinent breakup on the one side and subduction and assembly of a new supercontinent on the other side. These are responsible for the different types of orogeny at plate boundaries (Zheng and Chen, 2017; Zheng and Zhao, 2017, 2020). Whereas continental rifting and supercontinent breakup lead to crustal reworking, not only for high-$T/P$, Buchan-type metamorphism, but also for granitic magmatism along converged plate boundaries, the processes of subduction and supercontinent assembly lead to not only crustal reworking for low-$T/P$, Alpine-type metamorphism, but also crustal growth through mafic arc magmatism along converging plate boundaries. As a consequence, series of composite orogens are generated along converged plate boundaries, with superimposition of rifting orogeny on accretionary and collisional orogens. Because collisional orogeny during the continental subduction would postdate accretionary orogeny during the oceanic subduction, accretionary orogens would often be superimposed by collisional orogeny at first and then by rifting orogeny at last. As such, composite orogens contain the records of Wilson cycles in their geological history.

It appears that a supercontinent cycle is spatially and temporally associated with the three types of orogenies within a Wilson cycle. It also keeps pace with changes in the thermal state and also dynamic regime at both converging and converged plate boundaries. In this context, a pattern of alternating extensional

Figure 18. Schematic cartoons showing the tectonic evolution of a Wilson cycle. While the present diagram is revised from Zheng et al. (2019), the original one was integrated from Cawood et al. (2009) and Guillot et al. (2009), as well as from Zheng and Chen (2017). (A) Successful rifting for the opening of a new oceanic basin along a former suture zone, resulting in seafloor spreading, mid-ocean-ridge basalt magmatism, and high-temperature/pressure ($T/P$) Buchan-type metamorphism. (B) Subduction of oceanic lithosphere causing accretionary orogeny, resulting in the emplacement of an accretionary wedge, low-$T/P$ Alpine-type to moderate-$T/P$ Barrovian-type metamorphism, and mafic arc magmatism. (C) Subduction of continental lithosphere inducing collisional orogeny, resulting in low-$T/P$ Alpine-type to moderate-$T/P$ Barrovian-type metamorphism without mafic arc magmatism. (D) Abortive rifting for reactivation of fossil subduction zones, resulting in high thermal gradients with high heat flow for rifting orogeny. Upwelling of the asthenospheric mantle along the thinned orogenic lithosphere is responsible for the generation of high thermal gradients, which occur not only along continental rifts for high-$T/P$ Buchan-type metamorphism and granitic magmatism at convergent plate margins but also along mid-ocean ridges and backarc basins for mafic magmatism at divergent plate margins.
and compressional processes can be proposed for a supercontinent cycle with coupled metamorphism and magmatism as follows: (1) prevalence of active extension for seafloor spreading and mid-ocean-ridge basalt (MORB)–type magmatism during supercontinent breakup (Fig. 18A), accompanied by higher-T/P metamorphism of the MORB; (2) dominance of active compression resulting in lower-T/P metamorphism in the early stage of oceanic subduction but a prevalence of active extension for mafic arc magmatism in the late stage (Fig. 18B), accompanied by higher-T/P metamorphism in backarc sites; (3) dominance of active compression resulting in lower-T/P metamorphism during continental collision after the closure of oceanic basins (Fig. 18C); and (4) prevalence of active extension resulting in higher-T/P metamorphism and granitic magmatism in aborted rifts (Fig. 18D), where supercontinents attempted to break up but were not successful along converged plate boundaries. Therefore, over the course of successive cycles, the orogenic style is coupled with the type of regional metamorphism at plate boundaries, where no mafic arc magmatism is accompanied by lower-T/P metamorphism but abundant felsic magmatism is associated with higher-T/P metamorphism.

In principle, extensional tectonism for supercontinent breakup would have taken place in a geodynamic realm that was independent of the pericontinental subduction resulting in compressional tectonism occurring during supercontinent assembly. The former results in regional metamorphism at higher thermal gradients, whereas the latter leads to regional metamorphism at lower thermal gradients. The periods of lithospheric extension and compression are respectively correlated around and within the margins of a supercontinent. This correlation has important implications for the geodynamic processes of a supercontinent cycle. As a consequence, subduction tectonism is responsible for continental assembly, whereas rifting tectonism is responsible for continental breakup. In detail, the amalgamation of continental blocks is meshed with cold to warm metamorphism during the accretionary and collisional orogens. In contrast, the breakup of supercontinents is completed with hot to ultrahot metamorphism and its synchronous granitic magmatism during the rifting orogeny. In this context, both the top-down and bottom-up processes are relevant in the geodynamic mechanism of supercontinent cycles. Nevertheless, the considerable temporal lag between the two processes in a supercontinental cycle makes their tectonic relationship more independent than dependent.

The top-down process is powered by gravity-driven lithospheric downwelling for amalgamation and assembly of supercontinents. In contrast, the bottom-up process is powered by buoyancy-driven asthenospheric upwelling for continental rifting along fossil suture zones. Thus, active compression during plate convergence always considerably predates active extension during plate divergence along the same suture zones. Although the majority of continental rifts fail to develop into lithospheric rupture for seafloor spreading, a few of them did develop successfully into supercontinent breakup. In either case, it is common for granulite-migmatite-granite associations to form in the continental rifts, where the tectonic reactivation by lithospheric extension for asthenospheric upwelling is prominent at converged plate boundaries (Zheng and Chen, 2017; Zheng and Zhao, 2017, 2020). Therefore, the assembly and breakup of supercontinents are coupled with the two types of extreme metamorphism, which are intrinsically related to secular changes in both the thermal state and the dynamic regime along plate boundaries.

### CONCLUDING REMARKS

By linking the UHP and UHT metamorphic products to their lower-grade counterparts in the same metamorphic facies series, the two types of extreme metamorphism are linked to the two types of dynamic regimes and thermal states at convergent plate boundaries. One is gravity-driven subduction resulting in interplate compression and heating, leading to blueschist- to eclogite-facies HP to UHP metamorphism at low thermal gradients. The other is buoyancy-driven rifting in association with extension (underplating compression) and heating, resulting in amphibolite- to granulite-facies HT to UHT metamorphism at high thermal gradients. As a consequence, the metamorphic P-T conditions of crustal rocks change with the dynamic regime and thermal state of plate boundaries. Therefore, metamorphic facies series offer the means to trace subduction and rifting processes at convergent plate boundaries. Nevertheless, it is critical to assess how metamorphic rocks were produced at different depths in given dynamic regimes and thermal states during orogeny.

The extreme metamorphism of crustal rocks at convergent plate boundaries is generally characterized by two fundamentally different types of metamorphic transformation that follow each other in time. An early Alpine/Baroian-type metamorphic event is caused by lithospheric subduction resulting in compressional heating at lower thermal gradients. Its products are not only blueschist- to eclogite-facies HP to UHP rocks in the Phanerozoic, but also MP amphibolite- to HP granulite-facies rocks in the Precambrian. A later Buchan-type metamorphic event is produced by lithospheric rifting due to extensional heating at higher thermal gradients. Its products are LP amphibolite- to granulite-facies HT to UHT rocks, together with an abundance of coeval granites and migmatites since the Archean. The bimodal metamorphism, one colder and the other hotter, would operate one after the other by such tectonic processes at convergent plate boundaries. This dualism of regional metamorphism is responsible for lithospheric subduction and rifting, respectively, at converging and convergent plate boundaries. It also explains the superimposition of rifting orogeny on both accretionary and collisional orogens.

One-sided subduction creates asymmetry in the thermal structure of convergent plate boundaries. During oceanic subduction resulting in accretionary orogeny, lower-T/P metamorphism takes place early in the foreland compressional zone, whereas higher-T/P metamorphism happens late in the hinterland extensional zone. These two distinct thermal states and dynamic regimes are sequentially imprinted in the rock record as contrasting types of regional metamorphism with no spatial overlap, resulting in the classic definition of paired metamorphic belts. Bimodal metamorphism is also common in collisional orogens, where lower-T/P metamorphism occurs early during continental subduction resulting in collisional orogeny, whereas higher-T/P
metamorphism happens late during rifting orogeny to cause the postcollisional reworking. These two distinct thermal states and dynamic regimes are spatially superimposed in the rock record as contrasting types of metamorphism. Because subduction zones may change their P-T conditions in both time and space, their dynamic regimes and thermal states are the two key variables to the observation of orogenic processes during the evolution of convergent plate boundaries. In particular, collisional orogeny is always associated with continental assembly, which is indicated by the operation of cold to warm metamorphism but the absence of synchronous arc magmatism. In contrast, rifting orogeny is generally associated with supercontinent breakup, either successful or abortive, which is indicated by the operation of hot to ultra-hot metamorphism with contemporaneous granitic magmatism. Therefore, gravity-driven subduction resulting in active compression cannot serve as the geodynamic mechanism for supercontinent breakup. On the other hand, buoyancy-driven rifting resulting in active extension does lead to reactivation of convergent plate boundaries for coupled metamorphism and magmatism, though continental rifting does not always result in supercontinent breakup. Given the differences in geological records at convergent plate boundaries in different ages, it is essential to develop the means to decipher how plate tectonics has operated via either top-down or bottom-up processes at both converging and convergent plate boundaries. Therefore, characterization of the dynamic regimes and thermal states at such places represents a major step forward in this effort.

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