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Ultrahigh-Pressure Metamorphism: an Overview

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ABSTRACT

The ultrahigh-pressure (UHP) metamorphism was defined as the metamorphism occurring under pressure conditions higher than the quartz-coesite equilibrium curve. Based on the discovery of microdiamond and coesite inclusions, UHP metamorphic rocks of crustal origin have been reported from five continent-collision orogenic belts around the world. This shows that crustal rocks can be subducted to mantle depths (>100 km) during continental collision. Several minerals/mineral assemblages characteristically relevant to UHP conditions have also been established through experimental studies. A fast exhumation rate, the absence of a fluid phase and the continuous refrigeration effect during exhumation are required to preserve UHP minerals at the Earth's surface. However, the restricted occurrence of UHP minerals makes it difficult to confidently establish a regional scale for some UHP terranes. It was demonstrated that water activity in fluid phase during UHP metamorphism was low. "Excess Ar" was also shown to be commonly incorporated within UHP minerals, causing difficulties in Ar geochronology. Age dating of UHP rocks combined with petrologic pressure-temperature studies revealed that the exhumation rate of UHP rocks from mantle depths to the lower crustal level was higher than 7-14 mm/yr. The geodynamic model for the exhumation mechanism(s) of these UHP terranes, however, remains controversial due to the lack of key constraints.

Key Words: mineral inclusion, tectonic exhumation, ultrahigh-pressure metamorphism

I. Introduction and Historical Background

According to the Glossary of Geology, metamorphism means "The mineralogical, chemical and structural adjustment of solid rocks to physical and chemical conditions which have generally been imposed at depth below the surface zones of weathering and cementation, and which differ from the conditions under which the rocks in question originated" (Bates and Jackson, 1987). This definition clearly states two important points concerning metamorphism: that metamorphism occurs in solid state within the Earth, and that metamorphism is due to changing physical and chemical conditions. The so-called physical conditions mainly refer to variables, such as pressure and temperature. Both of these intensive variables, in general, increase with depth within the Earth. Because the protoliths of metamorphic rocks may be sedimentary, igneous or even metamorphic rocks, metamorphism is described as prograde or retrograde, depending on whether the *P-T* conditions change in an increasing or decreasing direction, respectively.

Based on the observation that different mineral assemblages in metamorphic rocks may characterize different *P-T* regimes, a series of names now defined as metamorphic facies have been proposed (e.g., Eskola, 1920). Although the temperature conditions in the postulated *P-T* regimes could be as high as 800-1000 °C, the pressure conditions were generally constrained to less than 10 kbar, i.e., approximately equivalent to the average thickness of the continental crust, 35 km (e.g., Miyashiro, 1973). With the advance of plate tectonic theory in the 1960s, some previously recognized metamorphic facies, such as the blueschist facies and the eclogite facies, were interpreted to have probably resulted from high-pressure(/low-temperature) conditions prevailing within convergent subduction zones (e.g., Ernst, 1973). Geophysical data also showed that some Benioff zones at convergent plate boundaries may reach a depth of 900-1200 km (Creager and Jordan, 1984), and that the crustal thickness may reach 70 km in continental collision zones (Chun and Yoshii, 1977). Awkwardly, despite all these observations/interpretations, metamorphism of crustal rocks with pressure conditions higher than 10 kbar was never seriously considered before 1984. There are several probable explanations: (1) this might be due to the belief that sediments would be scraped off the subducting oceanic plates, and that all continental materials would be subjected to "buoyant uprising" due to their low densities relative to the heavier mantle rocks; (2) it might have been inconceivable that such deeply buried rocks could ever be sampled at the Earth's surface; and (3) "quantitative" geothermobarometers and relevant experimental mineralogical/petrological studies were still lacking (Schreyer, 1995).

The above situation lasted until 1984, when Chopin first presented indisputable evidence of coesite inclusions in magnesian pyrope within quartzite from the Dora Maira Massif of the Western Alps. This finding clearly demonstrated that crustal rocks in continental collision zones could have been metamorphosed under pressure conditions not less than 28 kbar, which is significantly higher than the 10 kbar existing in conventional *P-T* regimes for metamorphism. At the same time, Smith (1984) also reported coesite inclusion in the clinopyroxene of eclogites in western Norway and suggested similar metamorphic pressure conditions. The extreme of metamorphic pressure conditions was later further pushed higher than 40 kbar when microdiamond inclusions were found in zircon within biotite gneisses and schists from the Kokchetav Massif of northern Kazakhstan by Sobolev and Shatsky (1990). Further studies conclusively showed that coesite inclusions existed in eclogitic crustal rocks in the Dabie-Sulu area, east central China (Wang *et al.*, 1989; Yang and Smith, 1989), the Zermatt-Saas zone of the Western Alps (Reinecke, 1991) and the Pan-African belt in northern Mali (Caby, 1994) whereas diamond inclusions were also found in the Dabie area, east-central China (Xu *et al.*, 1992). Although these ultrahigh-pressure indicators have mostly been found as relic minute inclusions in minerals, such as garnet and zircon, their presence implies that crustal rocks subjected to such ultrahigh-pressure (UHP) metamorphic conditions may be a significant and widespread geologic feature. Liou *et al.* (1994), therefore, concluded that it would be convenient to separate this UHP metamorphism from conventional high-pressure (HP) metamorphism by means of quartz-coesite equilibrium (Fig. 1). Carswell (1990), however, extended the eclogite facies of HP metamorphism to cover all the possible *P-T* conditions of UHP metamorphism (Fig. 1).

If the metamorphic pressure acting on the rocks in question is solely due to the lithostatic component resulting from the mass of the overburden, pressure conditions higher than 30 kbar would be equivalent to a depth of over 100 km. This implies that those UHP crustal rocks mentioned above must have been meta-

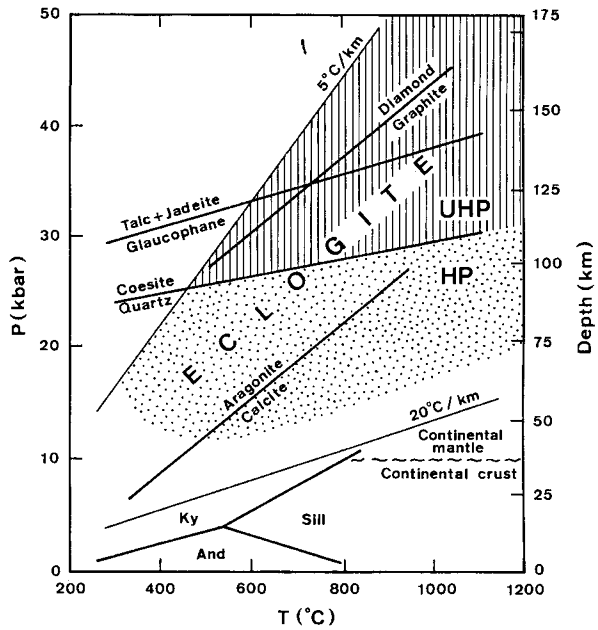

Fig. 1. *P-T* regimes for ultrahigh-pressure (UHP) and high-pressure (HP) metamorphism. The depth-pressure relation is based on an average density of 2.8 g/cm^3. Geotherms of 5 °C/km and 20 °C/km are indicated. Stabilities of diamond (Bundy, 1980), coesite (Bohlen and Boettcher, 1982), aragonite (Hacker *et al.*, 1992), talc+jadeite (Holland, 1988) and andalusite(And)-sillimanite(Sill)-kyanite(Ky) (Bohlen *et al.*, 1991) are shown for reference.

morphosed at mantle depths through subduction. The conventional wisdom that crustal rocks may survive by means of intracrustal recycling (melting) processes but can hardly be subducted beyond the crust-mantle boundary then needs revision. In this regard, the idea of "tectonic overpressure" has been revived in an attempt to reconcile the above discrepancy. "Tectonic overpressure" is an old idea related to HP metamorphism. It was postulated that during rock deformation, a differential stress could lead to an increase of the effective total pressure; consequently, some competent rocks might record pressure conditions higher than the actual lithostatic pressure. If the overpressure reached a value of 10 kbar, then the above mentioned UHP rocks could have actually been metamorphosed at the base of mountain roots, instead of at mantle depths. The most recent arguments supporting the idea of tectonic overpressure were mainly based on two observations from Kokchetav Massif: (1) microdiamond inclusions were found not only in UHP minerals, such as garnet and zircon, but also in non-UHP minerals, such as quartz, biotite, plagioclase, amphibole and chlorite; (2) most diamond-bearing rocks are confined to felsic rocks in narrow shear zones (Dobrzhinetskaya *et al.*, 1994). However, through detailed petrographic studies, Zhang *et al.* (1997) found that microdiamond inclusions in non-UHP minerals in the Kokchetav Massif are mainly due to metastable persistence of microdiamond in altered host minerals, such as garnet, during decompression/exhumation. The shear zones most probably developed during exhumation of the Massif, instead of during peak metamorphism. Schreyer (1995) also pointed out that if the protoliths of the UHP rocks contained a significant amount of talc or dolomite, the rocks could not have acted as competent. It seems that the tectonic overpressure may not have played an important role, and that these UHP metamorphic rocks may indeed have been metamorphosed at mantle depths.

It should be mentioned here that microdiamond and coesite as UHP indicators have actually long been recognized in kimberlite pipes (e.g., Smyth and Hatton, 1977) as well as in meteoric craters (e.g., Chao *et al.*, 1960). In addition, some diamond-bearing eclogite nodules in kimberlites and some diamondiferous orogenic ultramafic rocks might have formed at mantle depths directly through magmatic processes (e.g., Carswell, 1990; Pearson, 1993). These occurrences have somewhat different origins. To avoid ambiguity, they will not be included in the following discussion. The present paper will focus on the "metamorphism of crustal rocks at mantle depths through continental collision", which will lead to some important geodynamic implications/questions under the plate tectonic framework, such as: the mechanisms of subduction and exhumation of low-density crustal rocks to and from depths greater than 100 km, the preservation of UHP minerals at the Earth's surface, the possible genetic relation between UHP rocks and acidic igneous rocks during/after collision, the possible causes of intermediate-depth earthquakes, the role in geochemical cycling and the formation/destruction of continents during geologic time, etc. Much progress has been made on these questions since 1984. This overview, however, can not and will not include every detail. Those who have special interest in subjects not included here can consult the recent book "Ultrahigh Pressure Metamorphism" (Coleman and Wang, 1995) as well as the references therein.

II. Uncontroversial UHP Terranes

Coesite and diamond inclusions in minerals are unambiguous proof of UHP metamorphism (Fig. 1). They were found in the Western Gneiss Region (Smith, 1984), the Dora Maira Massif (Chopin, 1984), the Dabie-Sulu terrane (Wang *et al.*, 1989; Yang and Smith, 1989; Xu *et al.*, 1992), the Kokchetav Massif (Sobolev and Shatsky, 1990), the Zermatt-Saaş zone (Reinecke, 1991) and the Pan-African belt in northern Mali (Caby,

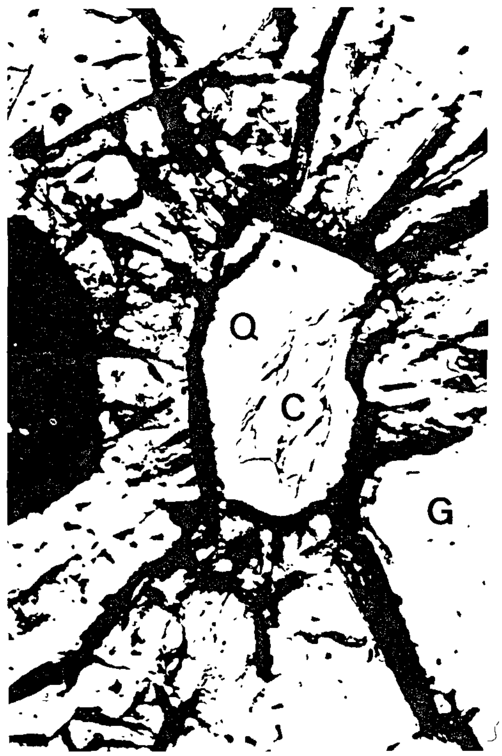

Fig. 2. Photomicrograph showing a coesite inclusion in a garnet host from an eclogite enclosed within a marble layer from the Dabie terrane. Note that coesite was partially inverted to quartz, and that radical cracks developed around the coesite inclusion. C: coesite, Q: quartz and G: garnet. Field of view: 0.23×0.36 mm. (Photo was kindly provided by R.Y. Zhang.)

1994). In all these uncontroversial UHP terranes, coesite inclusions are all partly inverted to low-pressure equivalent quartz. In some cases, the inversion is complete. It has been noted that the inversion always leads to fine-grained, partly feathery, polycrystalline aggregates of quartz surrounded by conspicuous radical cracks in the host minerals (Fig. 2). The radical cracks were interpreted to result from ~10% volume increase during coesite to quartz inversion. This kind of quartz pseudomorph inclusion accompanied by radical cracks in the host minerals was, therefore, employed as an indicator for UHP metamorphism for eclogites from the Saxonian Erzgebirge, Germany and Czechoslovakia (Schmadicke, 1991) and from the northeastern Bohemian Massif in Poland (Bakun-Czubarow, 1992; Brocker and Klemd, 1996), as well as for a garnet-chloritoid-talc schist from Tienshan, Kirghiz (Tagiri and Bakirov, 1990). However, some polycrystalline inclusions from the Erzgebirge were later found to be mixtures of quartz and feldspar and were interpreted to be the result of crystallizing from former melt inclusions, thus causing expansion cracks (Massonne, 1993). In addition, radical cracks can also form if inclusions and host minerals exhibit different elastic behaviors during decompression (Wendt *et al.*, 1993). Positive identification of coesite inclusions, instead of "polycrystalline quartz pseudomorphs", is therefore necessary to unambiguously establish a UHP terrane. In the following, an overview of general geology will be given only for those terranes with reported microdiamond- and coesite-inclusions.

1. Western Gneiss Region, Norway

The Western Gneiss Region, about 300 km long and 150 km wide, is located in the southwestern part of Norway (Fig. 3). It lies within the Caledonian Baltica-Laurentia collision zone, where the Western Gneiss Region is overridden by the Greenland plate. The Gneissic Complex consists of interlayered pelite, migmatite, marble, quartzite and amphibolite, with tectonic inclusions of gabbro, anorthosite and peridotite. Although most of the protoliths are of Precambrian age, the presence of minor Paleozoic rocks can not be excluded (Krill, 1980). Eclogites are widely spread as nodules, lenses, layers and larger bodies concordantly enclosed within gneisses. Based on the geochemical, isotopic and mineralogical data, Agrinier *et al.* (1985) concluded that these eclogites are most probably of crustal origin. The *P-T* estimates for these eclogites define a regional gradient from <600 °C and ~12 kbar in the southeast to >800 °C and ~18 kbar in the northwest (Griffin *et al.*, 1985). The hosting gneissic rocks, however, usually exhibit amphibolite facies assemblages, but relics of high-*P* assemblages, such as garnet, clinopyroxene, K-feldspar and quartz, also occur (Griffin *et al.*, 1985). Coesite inclusions, 100 to 200 μm in size, in eclogites were only found in Grytting and nearby Straumen in the westernmost part of the Western Gneiss Region (Smith, 1984, 1988). Recently, Dobrzhinetskaya *et al.* (1995) recovered three grains of microdiamond, less than 50 μm in size, from a high-grade gneiss sample from an island 100 km northeast of the coesite localities. Although the occurrence of UHP indicators is restricted, their presence is indisputable. Carswell (1992) proposed that these UHP rocks may only be a tectonic slice within the Western Gneiss Region, but that the whole region suffered extensive retrograde reactions during exhumation is an alternative possibility (Dobrzhinetskaya *et al.*, 1995).

Except for those eclogites associated with gneisses, some eclogites were enclosed within ultramafic rocks. They might have formed at 700-850 °C and at least 30 kbar (Smith, 1988). These eclogites and peridotites, however, were also subjected to the same Caledonian collision event.

Sm-Nd dating of clinopyroxene-garnet pairs and

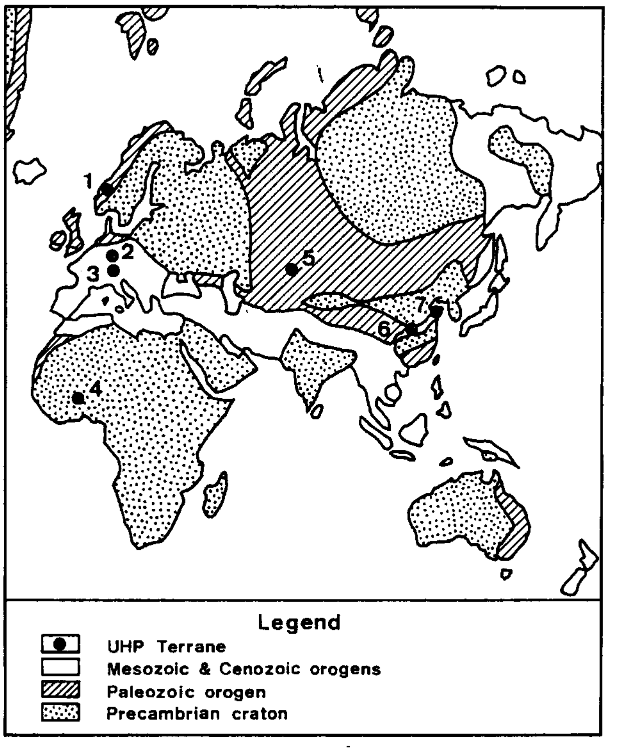

Fig. 3. Map showing the distribution of uncontroversial UHP terranes around the world: 1. the Western Gneiss Region, 2. the Zermatt-Saas Zone, 3. the Dora Maira Massif, 4. the Pan-African belt, northern Mali, 5. the Kokchetav Massif, 6. the Dabie Terrane, and 7. the Sulu Terrane.

U-Pb dating of zircon separates from eclogites of the Western Gneiss Region yielded ages from 400 to 450 Ma, reflecting the time of peak metamorphism (Gebauer *et al.*, 1985; Griffin and Brueckner, 1985). The $^{40}Ar/^{39}Ar$ mineral ages range from 390 to 410 Ma, which may indicate the time of retrograde amphibolite facies metamorphism (Chauvet and Dallmeyer, 1992; Boundy *et al.*, 1996). It should be noted that all these dating analyses were carried out on samples which do not contain coesite.

2. Dora Maira Massif, Western Alps

The Dora Maira Massif, together with the Monte Rosa and Gran Paradiso, form the internal crystalline massifs of the Penninic zone of the Western Alps (Fig. 3). The massif consists of a Paleozoic basement and a Mesozoic cover series, both of which underwent Alpine high-pressure/low-temperature metamorphism as a result of the Mesozoic-Cenozoic convergence of Europe and Adria. The high-pressure belt of the southern Dora Maira Massif has been studied in detail by Chopin *et al.* (1991) and divided into three crustal slices, differing in metamorphic grade, by low-angle south-dipping faults. The major foliation in these slices roughly parallels the lithologic and tectonic boundary and developed mainly under greenschist facies conditions; i.e., it postdates the HP assemblages. It is noted that blueschist facies metamorphism dominated in the upper (P = 10-12 kbar) and lower (P < 10 kbar?) crustal slices whereas eclogite was found only in the middle one. The eclogite crustal slice was further divided into the lower coesite-bearing hot eclogite (UHP) unit, covering an area of about 5×10 km, and the upper cold eclogite unit. The former mainly consists of gneiss, metapelite, eclogite, marble and some Hercynian intrusives. The UHP mineral inclusions, such as coesite (with grain size from 300 to 1000 μm)/quartz pseudomorphs, ellenbergerite and Mg-staurolite, were found within pyrope quartzite boudins in gneisses and within schist and eclogite (Chopin, 1984; Chopin *et al.*, 1991; Schertl *et al.*, 1991). The peak P-T conditions were estimated to be ~37 kbar and ~800 °C, and the retrograde P-T conditions were 7-9 kbar and 500-600 °C (Schertl *et al.*, 1991) (Fig. 4). Coesite inclusions, however, were not found in gneisses, which consist mainly of greenschist facies mineral assemblages, but UHP conditions were inferred from the coexistence of relic minerals of grossular-rich garnet and rutile in one such gneiss sample by Chopin *et al.* (1991) and Schertl *et al.* (1991). The peak P-T conditions for the cold eclogite unit were significantly lower, i.e., about 15 kbar and 500 °C (Chopin *et al.*, 1991).

Based on $^{40}Ar/^{39}Ar$ dating of phengitic mica, the time of UHP metamorphism was considered to be 80-110 Ma, and the time of major retrograde metamorphism to be 35-41 Ma (Monie and Chopin, 1991). However, Tilton *et al.* (1991) and Gebauer *et al.* (1997) reported ages of 240-300 Ma for protoliths, 35-38 Ma for UHP metamorphism and 28-30 Ma for retrograde metamorphism based on the Pb-Sr-Nd isotopic systematics and ion probe zircon dating. The latter results were supported by recent laser-Ar studies showing excess Ar in Dora Maira phengite (Scaillet *et al.*, 1992; Arnaud and Kelley, 1995).

3. Dabie-Sulu Terrane, East-Central China

The Dabie-Sulu metamorphic complex represents the Triassic collision zone between the Sino Korean and the Yangtze cratons (Fig. 3). The Dabie terrane was separated from the Sulu terrane by the Tanlu fault for a distance of at least 500 km, probably during the Cretaceous and/or Cenozoic era. Based on differences in the metamorphic grade, the Dabie terrane can be further divided into three units: a northern migmatitic belt (P < 20 kbar), a central UHP coesite and diamond-bearing eclogite belt (P = 20-40 kbar) and a southern HP blueschist-eclogite belt (P = 5-12 kbar). The UHP metamorphic rocks in both the Dabie (covering an area of > 400 km^2) and the Sulu (covering an area of > 10000 km^2) terranes include mainly eclogite, marble, (para)gneiss, amphibolite and ultramafics.

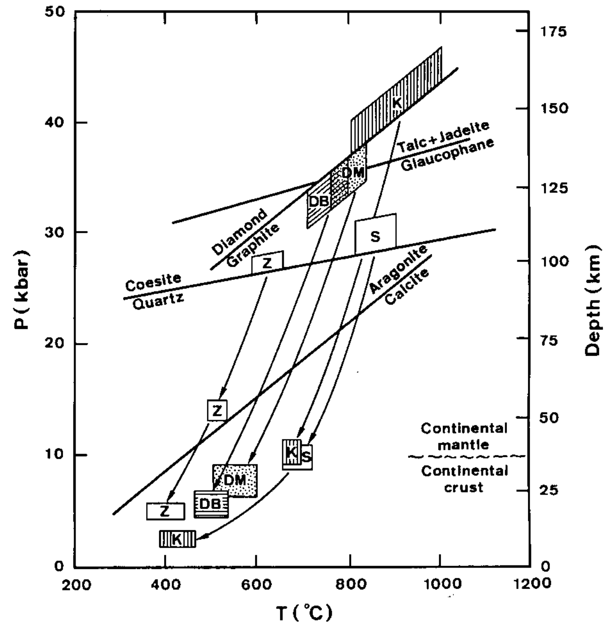

Fig. 4. Representative P-T regimes of UHP metamorphic conditions and retrograde overprints referred to in the text. The stability curves are the same as in Fig. 1. Data sources are: K, Kokchetav Massif (Zhang *et al.*, 1997); DM, Dora Maira Massif (Schertl *et al.*, 1991); DB, northern Dabie terrane (Wang *et al.*, 1992); S, eastern Sulu terrane (Enami *et al.*, 1993); and Z, Zermatt-Saas Zone (Reinecke, 1991).

Coesite, 50 to 300 μm in size, and quartz pseudomorphs inclusions in zircon, garnet, omphacite, kyanite, epidote, zoisite and dolomite were found not only in eclogite but also in marble, schist and gneiss in the Dabie and the Sulu terranes (Wang *et al.*, 1989; Yang and Smith, 1989; Wang and Liou, 1991; Schertl and Okay, 1994). Relic intergranular coesite grains were also recently reported in eclogites from the Sulu area (Liou and Zhang, 1996). This UHP mineral, although rare, is widely distributed. On the other hand, diamond inclusions, ranging in size from 150 to 700 μm, in garnet were only found in marble-hosted eclogites in the Dabie terrane (Xu *et al.*, 1992). It should be emphasized that UHP minerals may not be restricted only to the Dabie and Sulu areas because coesite inclusions were also found in the Hongan block (Zhang and Liou, 1994), which is located to the west of the Dabie terrane.

On the basis of garnet-omphacite Fe-Mg thermometry, a regional gradient in maximum temperature conditions for the eclogites in the Sulu terrane was defined, with a southwestward decrease from 810-880 °C to 500-750 °C (Enami *et al.*, 1993). A concomitant decrease of minimum pressure conditions, i.e., from > 28 kbar to > 26 kbar, could also be inferred based on the quartz-coesite equilibrium curve. In the Dabie terrane, the estimated *P-T* conditions for eclogites were higher in the north (i.e., > 35 kbar, 700-800 °C) but lower in the south (i.e., 20-25 kbar, 550-600 °C). Wang *et al.* (1992) suggested that this *P-T* difference is gradational in space. However, Okay (1993) and Carswell *et al.* (1997) preferred a scenario of abrupt change and postulated that the coesite/diamond-free cold eclogite zone in the south was tectonically overlain on top of the coesite/diamond-bearing hot eclogite zone in the north. It should be noted that although UHP minerals are not uncommon in the Dabie and the Sulu terranes, all UHP rocks show extensive amphibolite/greenschist retrograde overprint. The retrograde *P-T* conditions vary: $P > 9$ kbar and $T > 680$ °C for the eastern Sulu terrane, $P < 10$ kbar and $T < 580$ °C for the western Sulu terrane, and $P = 6$ kbar and $T = 470$-530 °C for the Dabie terrane (Wang *et al.*, 1992; Enami *et al.*, 1993) (Fig. 4).

Sm-Nd and U-Pb dating of eclogites and gneisses for both the Dabie and the Sulu terranes yielded consistent UHP metamorphic ages of 200-244 Ma (Ames *et al.*, 1993, 1996; Li *et al.*, 1993). Excess Ar was also found in phengite in the UHP eclogites (Li *et al.*, 1994). $^{40}Ar/^{39}Ar$ dating further suggested that HP blueschists/eclogites and UHP gneisses recorded cooling to about 300 °C at 180-236 Ma (Eide *et al.*, 1994; Hacker and Wang, 1995; Hacker *et al.*, 1995). The protoliths of the UHP eclogites were also shown to be of crustal origin and of the Proterozoic age (700-800 Ma) (e.g., Li *et al.*, 1993; Yui *et al.*, 1995; Ames *et al.*, 1996; Jahn *et al.*, 1996).

4. Kokchetav Massif, Kazakhstan

The Kokchetav Massif in northern Kazakhstan is a large (300×150 km) fault-bounded metamorphic complex surrounded by Caledonian rocks of the Ural-Mongolian fold belt (Fig. 3). The central part of this massif consists of Proterozoic rocks metamorphosed to amphibolite/greenschist facies conditions. The massif core is composed of a variety of coarsely crystalline schist, gneiss, eclogite, amphibolite, granulite, quartzite and marble. Microdiamond inclusions, mostly smaller than 20 μm, were found in garnet and zircon in garnet-pyroxene and garnet-pyroxene-carbonate rocks, garnet-biotite gneisses and schist and marble (Sobolev and Shatsky, 1990; Zhang *et al.*, 1997). Coesite/quartz pseudomorph was reported as inclusions in zircon and garnet in both eclogite and gneiss (Claoue-Long *et al.*, 1991; Zhang *et al.*, 1997). Microdiamond inclusions, however, have not been found in eclogites yet. Available carbon isotope data (from -10 to -19‰ $\delta^{13}C$) suggest a crustal organic origin for the microdiamond inclusions (Sobolev *et al.*, 1989). All these diamond/coesite-bearing rocks mainly occur as lenticular bands or lens-like bodies within garnet-biotite gneiss. Field occurrences suggest that the hosting gneiss may actually be a tectonic slice with thickness less than 5 km (Ernst *et al.*, 1995). Zhang *et al.* (1997) speculated that UHP metamorphism might have taken place during the middle Cambrian, probably involving the collision and profound underflow of a narrow salient of the Proterozoic Kazakhstan-North Tienshan microcontinent+superjacent uppermost Proterozoic and Lower Cambrian passive margin section beneath a now-largely vanished immature oceanic island arc. The intense deformation/mylonitization observed in the massif mainly developed during subsequent exhumation.

The metamorphic peak *P-T* conditions for these UHP rocks might have been $P > 40$ kbar and $T = 800$-1000 °C while those for the retrograde overprint were $P < 10$ kbar and $T = 650$-680 °C (Zhang *et al.*, 1997) (Fig. 4). U-Pb and Sm-Nd dating of eclogite, biotite gneiss and diamond-bearing rocks yielded a UHP metamorphic age of 530-533 Ma (Claoue-Long *et al.*, 1991; Shatsky *et al.*, 1998). Retrograde mica from diamond-bearing garnet-biotite gneisses yielded an $^{40}Ar/^{39}Ar$ age of 517 Ma (Shatsky *et al.*, 1998).

5. Zermatt-Saas Zone, Western Alps

The Zermatt-Saas Zone is a part of the former Piemonte-Ligurian ocean that was closed by subduction underneath the Austroalpine Sesia Zone during the late Mesozoic-Cenozoic era (Fig. 3). It is an eclogite-facies meta-ophiolitic unit overlying the Monte Rosa, Gran Paradiso and Dora Maira crystalline massifs. Coesite (up to 60 μm) and quartz-pseudomorphs were found as inclusions in pyrope-rich garnet and tourmaline of a manganiferous quartzite (Reinecke, 1991). The quartzite occurs as a 2×5 m sized lens, together with several eclogites and clinopyroxene-rich rocks, within carbonate-garnet-mica schists, which overlie a thick sequence of eclogites and outcrop at Lago di Cignana, Valtournanche, Western Alps, Italy (i.e., west of the Monte Rosa Massif). The peak metamorphic conditions were estimated to be 590-630 °C and 26-28 kbar. Retrograde metamorphic overprints were noticeable. Two stages of partial re-equilibration were suggested at about 500 °C/13-15 kbar and 360-430 °C/4-6 kbar (Reinecke, 1991) (Fig. 4). This is the first finding of UHP metamorphic rocks within sediments of former oceanic crust and provides direct proof of very deep subduction of relatively soft oceanic sediments to a considerable depth of about 90 km. Reinecke (1991) also mentioned rare coesite inclusions in omphacite from eclogites near the quartzite, suggesting that the metaophiolites in the Zermatt-Saas Zone perhaps also experienced this UHP metamorphism.

Recent Sm/Nd dating on an eclogite sample from the Zermatt-Saas Zone yielded an age of 52±18 Ma. The dated eclogite sample, however, does not contain coesite and was collected 22 km northeast of Lago di Cignana (Bowtell *et al.*, 1994).

6. Pan-African Belt in Northern Mali

The southern Saharan Segment of the Pan-African belt in northern Mali is the suture between the west Africa plate and an island arc or an Adean-type continental margin (Fig. 3). The age of this suture zone may not be older than 630 Ma (Caby *et al.*, 1989). In the east part of the Gourma area, a complex edifice of flat-lying nappes, which form an elongated belt up to 100 km wide, run along the suture zone. These nappes comprise the lower internal nappes and the upper external nappes. High pressure eclogitic rocks, 25×100 km in size, occur in the internal nappes. Coesite, up to ~150 μm in size, occurs as inclusions in the omphacite of a basic nodule enclosed within an impure marble. Possible quartz pseudomorphs after coesite were also found in garnets from the surrounding eclogitic mica schist and eclogite boudins (Caby, 1994). The estimated metamorphic peak conditions were 700-750 °C and >27 kbar. $^{40}Ar/^{39}Ar$ dating of phengite from the quartz pseudomorph-bearing mica schist yielded an age of 1048 Ma (Caby, 1994). However, considering the possible presence of excess Ar, the geologic meaning of this date is dubious at the moment. Nevertheless, the discovery of coesite in the Pan-African belt demonstrates that UHP metamorphism, or subduction of crustal rocks to mantle depths, might not be confined to the Phanerozoic period of the Earth's history.

III. Characteristic Minerals Relevant to UHP Metamorphism

The confirmation that crustal rocks may have subducted to mantle depths is primarily based on experimental mineralogy and petrology at high pressures and temperatures, through which (semi)quantitative pressure and temperature constraints on UHP conditions can be obtained. Such experiments conducted under mantle *P-T* conditions have conventionally focused on the mantle materials, but have recently been extended to crustal compositions (e.g., Schreyer, 1988a). The available data show that some minerals or mineral assemblages are characteristically relevant to UHP metamorphism. For simplicity, these minerals or mineral assemblages and their possible *P-T* stability fields are listed in Table 1. Note that if quartz-coesite equilibrium is taken as the *P-T* boundary between UHP and HP metamorphism (Liou *et al.*, 1994), then minerals/mineral assemblages, such as diamond, coesite, phase A, phase D, clinohumite-OH, topaz-OH, MgMgAl-pumpellyite, wollastonite II m, wollastonite II tc, ellenbergerite-Ti, ellenbergerite-Zr, high K clinopyroxene, talc+jadeite and grossular+rutile, would be diagnostic to UHP metamorphic rocks (Table 1). It is also interesting to note that some of these diagnostic minerals have not yet been reported in nature (Table 1). More detailed petrographic studies on UHP rocks to determine their possible occurrences are certainly warranted.

Besides those listed in Table 1, there are many minerals which may also be present in UHP rocks but are less informative. Almandine, grossular, spessartine, corundum, chlorite, epidote, zoisite, talc, phlogopite, chloritoid, Fe-staurolite, Ca-pumpellyite, K-feldspar, pyroxenes, magnesite, dolomite etc. may be examples. This list certainly indicates that it is quite possible that with appropriate bulk chemical compositions, UHP rocks may actually contain absolutely "normal" mineral assemblages. Detailed petrographic studies on rocks of different bulk chemical compositions are, therefore, of prime importance for identifying UHP terranes.

It should be pointed out that the above mentioned minerals/mineral assemblages are only representative examples. More complete and detailed discussion on this issue has been given by Schreyer (1988a) and Massonne (1995). One thing which should be emphasized here is that most experimental studies that have been carried out involving crustal compositions were largely limited to simple chemical systems, such as C, SiO_2, CaO-SiO_2, and (K_2O)-MgO-Al_2O_3-SiO_2-H_2O (Table 1). More complex chemical systems, such as those which include Fe, Na and CO_2, or even rock-specific mineral reactions, that help in deciphering the solid solution effect of minerals on the *P-T* stability field, need to be studied.

Besides the above qualitative/semi-quantitative UHP minerals, there are also a few quantitative geothermobarometers applicable to UHP metamorphic conditions. Examples are the garnet-clinopyroxene Al barometer (e.g., Wood and Holloway, 1984), garnet-orthopyroxene Al barometer (e.g., Boyd and England, 1964), phengite (+talc+kyanite+quartz/coesite, or +phlogopite+kyanite+quartz/coesite) barometer (e.g.,

Table 1. Minerals/Mineral Assemblages Relevant to UHP Metamorphism

Minerals/Mineral assemblages	Mineral composition	Approximate stability range, kbar, (a_{H_2O}=1)	Reference
Diamond	C	>27 (>500 °C)	Bundy (1980)
Coesite	SiO_2	25-85 (>500 °C)	Bohlen and Boettcher (1982)
Phase A*	$Mg_7Si_{12}O_{14}H_6$	>55 (<935 °C)	Horiuchi *et al.* (1979)
Phase D*	$Mg_5Si_2O_{10}H_2$	>29 (>730 °C)	Yamamoto and Akimoto (1977)
"Clinohumite-OH"*	$Mg_9Si_4O_{18}H_2$	>29 (>730 °C)	Yamamoto and Akimoto (1977)
Phase Pi*	$Al_3Si_2O_7(OH)_3$	22-60 (450-700 °C)	Wunder *et al.* (1993a)
"Topaz-OH"*	$Al_2SiO_4(OH)_2$	>55 (400-900 °C)	Wunder *et al.* (1993b)
Mg-carpholite	$MgAlSi_2O_6(OH)_4$	7-50 (<650 °C)	Schreyer (1988a)
Mg-chloritoid	$MgAl_2SiO_5(OH)_2$	18-58 (400-800 °C)	Schreyer (1988a)
Mg-staurolite	$Mg_4Al_{18}Si_8O_{46}(OH)_2$	>13 (700-950 °C)	Schreyer (1988a)
Pyrope	$Mg_3Al_2Si_3O_{12}$	>14 (>700 °C)	Schreyer (1988a)
"MgMgAl-pumpellyite"*	$Mg_5Al_5Si_6O_{21}(OH)_7$	>37 (<800 °C)	Schreyer (1988a)
"Wollastonite II m"*	$CaSiO_3$	>30 (600-900 °C)	Chatterjee *et al.* (1984)
"Wollastonite II tc"*	$CaSiO_3$	>30 (>900 °C)	Chatterjee *et al.* (1984)
Lawsonite	$CaAl_2Si_2O_7(OH)_2{\cdot}H_2O$	>3 (<700 °C)	Chatterjee *et al.* (1984)
"K-cymrite"*	$KAlSi_3O_8{\cdot}H_2O$	>25	Massonne (1992)
Jadeite	$NaAlSi_2O_6$	>10-25	Netwon and Kennedy (1968)
Glaucophane	$Na_2Mg_3Al_2Si_8O_{22}(OH)_{22}$	10-30 (<700 °C)	Holland (1988)
"Ellenbergerite-Ti"	$Mg_{6.66}Ti_{0.66}Al_6Si_8O_{28}(OH)_{10}$	27-42 (650-725 °C)	Chopin *et al.* (1992)
"Ellenbergerite-Zr"	$Mg_{6.66}Zr_{0.66}Al_6Si_8O_{28}(OH)_{10}$	>28 (<770 °C)	Burchard (1993)
"Ellenbergerite-P"	$Mg_{13}P_8O_{28}(OH)_{10}$	>9 (<750 °C)	Beller (1987)
Aragonite	$CaCO_3$	>8-30 (>400 °C)	Hacker *et al.* (1992)
Calderite	$Mn_3Fe_2Si_3O_{12}$	>15 (>500 °C)	Schreyer and Baller (1981)
high K clinopyroxene	0.17 wt% K_2O	65 (1200 °C)	Doroshev *et al.* (1992)
Talc + Kyanite		>4 (<800 °C)	Schreyer (1988a)
Talc + Phengite (+ SiO_2)		>10 (<600-800 °C)	Massonne and Schreyer (1989)
Talc + Jadeite		>30 (>350 °C)	Holland (1988)
Grossular + Rutile		25-85 (>500 °C)	Chopin *et al.* (1991)

*Not found in nature yet.

Massonne and Schreyer, 1989), two pyroxene solvus thermometer (e.g., Boyd, 1973), garnet-clinopyroxene Fe-Mg thermometer (e.g., Krogh, 1988), garnet-orthopyroxene Fe-Mg thermometer (e.g., Harley, 1984), garnet-olivine Fe-Mg thermometer (e.g., O'Neil and Wood, 1979) and garnet-phengite Fe-Mg thermometer (e.g., Krogh and Raheim, 1978) (see Carswell (1990) for a detailed review). Most of these thermobarometers, however, are only suitable for basic/ultrabasic rocks. It should also be remembered that quantitative *P-T* estimates might be difficult to obtain even with minerals/mineral pairs suitable for geothermobarometers present as relics in inclusions because it is usually hard to prove the state of mineral equilibrium or the presence of buffering assemblages during UHP metamorphism.

IV. Preservation of UHP Minerals and Fluid Composition during UHP Metamorphism

The preservation of coesite or diamond (or any other UHP mineral) as inclusions in minerals, such as garnet and zircon, in UHP rocks through tectonic exhumation from mantle depths to the Earth's surface has been a puzzling problem. It is generally agreed that in addition to the requirement of "fast exhumation" with "simultaneous cooling", the rigid host minerals must have acted as a "pressure vessel". The "pressure vessel" can not only restrict the phase inversion, such as coesite to quartz or diamond to graphite, to only moderate amounts, but also prevent the UHP mineral inclusions from accessing the fluid phase, which would be an effective catalyst for the phase inversion processes (Chopin, 1984; Gillet *et al.*, 1984; Wang *et al.*, 1989). The different *P-T*-time paths, which may reflect exhumation rate and retrograde *P-T* conditions, and the different extent of retrograde deformation, which may enhance the fluid-influx ability, may therefore both control the degree of preservation of UHP mineral inclusions in different UHP rocks. It is important to note that, recently, Liou and Zhang (1996) found intergranular coesite grains in an eclogite body from the Sulu region. This rare occurrence of intergranular coesite in a rock matrix clearly demonstrates that the lack of fluid during rapid exhumation of UHP rocks may have played a more important role (Liou *et al.*, 1997). The absence of UHP minerals in some country rock gneisses could therefore be interpreted as resulting from more extensive retrograde overprinting, most probably facilitated by the presence of minor amounts of a fluid phase.

The presence or absence of a fluid phase is crucial not only during the retrograde but also during the prograde stage. Based on fluid inclusion studies, Philippot *et al.* (1995) documented a change from low salinity fluids to complex saline brines during prograde UHP metamorphism. Stable isotope data also showed heterogeneous $\delta^{18}O$ values on meter scales despite homogeneous δD values for the UHP rocks (Philippot, 1993; Sharp *et al.*, 1993; Rumble and Yui, unpublished data). The fluid flow within the UHP rocks was, therefore, suggested to be short-ranged and limited, which was thought to result from the formation of dense hydrous silicates, such as epidote and phengite, during deep subduction (Philippot, 1993). The water activity of the fluid phase under UHP conditions was also suggested to be lower than 1.0. The important role of water as a reaction catalyst during eclogitization has been illustrated by Austrheim (1987). Zhang and Liou (1997) further presented a gradational change from incipiently metamorphosed gabbro in the core to completely recrystallized coesite-bearing eclogite along the margins of a single 30 m block. This clearly indicates that some low-*P* mineral assemblages may persist metastably under UHP conditions due to sluggish reactions resulting from a lack of water during prograde metamorphism. This means that disequilibrium assemblages on both thin section and outcrop scale may have prevailed under UHP conditions. This would cause additional difficulties in differentiating the in situ or allochthonous model in the following discussion. Most importantly, the water activity in fluid phase has a profound effect on the stability fields of UHP minerals/mineral assemblages, which have mostly been experimentally determined under fluid-excess conditions with a water activity equal to 1.0 (Table 1).

V. Structural Coherence of UHP Terranes

The question of whether the UHP terranes were subducted and exhumed as "coherent structural units" (i.e., the in situ model) or as "tectonic melanges" (i.e., the allochthonous/exotic model) has been controversial during the past decade. The main reason for this debate has been that while diagnostic UHP minerals were found in eclogitic nodules, mineral assemblages of the enclosing country rocks mainly recorded greenschist or amphibolite facies metamorphic conditions (Smith, 1988). Contrasting *P-T* estimates for the peak metamorphism but comparable *P-T* conditions for the retrograde metamorphism between eclogitic nodules and the enclosing country rocks, therefore, were interpreted as resulting from tectonic juxtaposition during exhumation (e.g., Li *et al.*, 1993). Sheared borders around the eclogitic nodules as well as possible different metamorphic ages for nodules vs. country rocks reinforced the arguments for tectonic emplacement (Smith, 1988). However, with more detailed work in some well-exposed UHP terranes, such as the Dabie-Sulu terrane, it was shown that the diagnostic UHP minerals were not only present in the eclogitic nodules, but were also found, although rarely, within the country rocks, such as gneiss, marble and schist (e.g., Wang and Liou, 1991). The inferior ability of the enclosing gneisses to retain UHP minerals could be ascribed to the presence of minor amounts of fluid derived from some spontaneous retrograde dehydration reactions (Yui *et al.*, 1997). In addition, dating of zircons from both the eclogitic nodules and the enclosing gneisses showed identical ages of UHP metamorphism (Ames *et al.*, 1996). Structural concordance post-dating the UHP metamorphism between lithological units containing UHP minerals and the enclosing gneisses with amphibolite facies mineral assemblages was also confirmed (Xue *et al.*, 1996). All this evidence would favor the "in situ" model. The occurrences of coesite/diamond inclusions in orogenic belts, therefore, have been inferred as reflecting a regionally extensive UHP metamorphism (Wang and Liou, 1991).

It should be noted that the UHP indicators are distributed in a rather limited way, especially in western Norway, the Zermatt-Saas Zone and northern Mali (see the section "Uncontroversial UHP Terranes"). The generalization of local indicators of UHP conditions to regional scales must be viewed with caution. As a matter of fact, the evidence for or against the in situ or allochthonous model may appear convincing for different UHP terranes. This may partly be due to the differing extent of the studies and may also result from authentic differences in the tectonic history of different UHP terranes. The interpretation of regionally structural coherent UHP terranes must rely heavily on extensive field-mapping, petrologic, geochronologic and, especially, structural studies. In this context, it is a challenging task to distinguish between those gneisses which experienced UHP metamorphism but were later overwhelmingly overprinted by retrograde assemblages and those gneisses which only witnessed prograde low *P-T* conditions but were later tectonically emplaced within the UHP rocks. Besides these uncertainties, Harley and Carswell (1995) also mentioned that the presence of some mantle-derived peridotite lenses in both the Western Gneiss Region and the Dabie-Sulu terrane may not be in accord with the suggestion that these two terranes are tectonometamorphic regions structurally coherent on a large scale. The main issue of "in situ or allochthonous" may, therefore, be re-cast as "the (largest) scale of the structurally coherent UHP unit during subduction and exhumation". The answer to this question would be an important aid in physical modeling of the kinematics of UHP tectonics. Based on the spatial distribution of the eclogites and their country rocks with abnormally low $\delta^{18}O$ values, the scale of the structurally coherent UHP unit was estimated as large as 25×100 km in east-central China, although the thickness of the unit was not known (Rumble and Yui, unpublished data).

VI. Geochronology of UHP Terranes

In both the Dora Maira Massif and the Dabie-Sulu terranes, the $^{40}Ar/^{39}Ar$ plateau ages of the UHP phengites are mostly distinctly older than the ages derived from U-Pb, Sm-Nd and Rb-Sr systems (e.g., Mattauer *et al.*, 1991; Monie and Chopin, 1991; Tilton *et al.*, 1991; Ames *et al.*, 1993, 1996). Although in conflict with the concept of blocking temperature in geochronology, these results were further confirmed by Li *et al.* (1994), who employed different isotope systems on the same UHP eclogite samples and also got older $^{40}Ar/^{39}Ar$ ages. Li *et al.* (1994) therefore suggested that despite the "plateau" age spectrum, the UHP phengites must contain a significant amount of excess Ar. Laser Ar datings on mineral grains from the Dora Maira Massif also revealed that excess Ar might have been present in all the UHP minerals, and that it was most probably incorporated into these UHP phases during their formation (Arnaud and Kelley, 1995). It seems that the radiogenic Ar produced within or some external Ar entrapped by the protoliths of the UHP rocks was largely retained after the rocks in question subducted over the depths of blueschist facies *P-T* conditions. This also implies that the Ar partial pressure must be high under most UHP environments.

In studying the fluid behavior within subducted materials, Philippot (1993) postulated that the top of a subducted crust will be characterized by an inverted thermal gradient, and that at sub-arc depths (i.e., 50 to 150 km), the contact between a cold/competent/fluid-rich subducted crust and the above hot/relatively weak/anhydrous mantle material will form an interface acting as an impermeable barrier for fluid. The fluid phase within the subducted crust will, therefore, be largely kept either through hydration/carbonation reactions or by downward forced flow. This postulated fluid behavior was supported by the unusually low $\delta^{18}O$ values of some UHP rocks in the Sulu region (Yui *et al.*, 1995) and can also be employed to explain the observation of excess Ar within crustal rocks at mantle depths. It should be emphasized, however, that UHP minerals containing excess Ar might not be the rule, and that exceptions exist. For example, some phengites from UHP terranes seem to yield reasonable $^{40}Ar/^{39}Ar$ ages, which were interpreted as the time of retrograde metamorphism/cooling (e.g., Hacker and Wang, 1995; Boundy *et al.*, 1996). Whether these "well-behaved" phengites were fortuitous, were related to retrograde recrystallization or were due to different kinds of UHP environments is not yet clear. More detailed geochronological studies are obviously needed. In this context, future studies on the physical properties of crustal materials under UHP conditions may also be important in deciphering the possible effects of rheological change on fluid behavior throughout UHP metamorphism.

With the problem of the possible presence of excess Ar in UHP rocks in mind, extensive geochronological studies have only recently been carried out for the Dora Maira Massif, the Kokchetav Massif and the Dabie terrane. The ages of the peak/retrograde metamorphism for these three UHP terranes were found to be 35-38/28-30 Ma, 530-533/517 Ma and 200-244/180-236 Ma, respectively (see the section "Uncontroversial UHP Terranes"). The pressure conditions for the peak/retrograde metamorphism were estimated to be 37/7-9 kbar, >40/10 kbar and >35/6 kbar, respectively. Despite the poor precision of age dating in some cases, the exhumation rates from mantle depths to crustal levels for these three UHP terranes would then be greater than 7-14 mm/yr. There is no doubt that it is too simple to use only one averaged exhumation rate to describe a decompression distance of 50-100 km for a metamorphic complex. Since most UHP rocks record multiple stages of metamorphic overprint, more precise age dating, more advanced techniques, such as using ion or laser probes, applied to minerals with chemical zoning, and, possibly cleavage/foliation dating will be challenging tasks and will help to unravel the exhumation processes of UHP terranes.

VII. Geodynamics of UHP Terranes

The discovery of UHP terranes, composed mainly of continental materials, leads to three basic problems in the context of geodynamics: (1) How are continental crustal rocks subducted to mantle depths? (2) How (why) does subduction stop? (3) How do crustal rocks exhume back to the Earth's surface?

Continental materials could be subducted to great depths if the buoyancy effect due to the density contrast between continental and mantle materials could be overcome. England and Holland (1979) showed that if the crust and the slab/mantle remain coupled, peninsulas or microcontinents could be subducted completely by the negative buoyancy of the lithosphere. Based on buoyancy analysis with the assumption of local isostasy and slab-pull driven subduction, Cloos (1993) further demonstrated that continental margins or oceanic island arcs with crustal thickness greater than 15-20 km would jam the subduction zones, but that a crust with thickness less than 15-20 km could be deeply subducted. In addition, continental crust of normal thickness could also be tectonically sliced into the buoyant upper crust, which could be accreted to the overriding plate, and the lower crust, which could adhere to the slab/mantle materials and subduct to be transformed into UHP rocks. In this respect, it would be interesting to see if a UHP terrane contains rocks of upper or lower crust affinity.

Since the present ocean floor is younger than ~200 Ma, and since less than 0.001% subducted oceanic crust was incorporated into mountain belts, it is generally accepted that most of the oceanic crust that formed during the Earth's history was destroyed during subduction (Coleman, 1977). The subducted continental crust, therefore, would theoretically continue to undergo subduction until destruction and would not exhume back to the Earth's surface unless the continental crust was decoupled from the down-going slab/mantle material through various processes. Such decoupling processes are assumed to occur mainly due to different rheological behaviors between crustal acidic rocks and adjacent mafic/ultramafic rocks as temperature increases during subduction. The crustal rocks would become more ductile and weak zones will eventually form leading to decoupling and exhumation. Nevertheless, this does not mean that all subducted continental crust would detach from the down-going slab and return back to the Earth's surface. Schreyer (1988b) postulated that the incorporation of UHP rocks in orogenic belts may only be a special case and that some of UHP rocks may have coupled with the subducted slab to a deeper level within subduction zones, where they finally melted and interacted with the mantle material. The latter case would have special implications on the issue of geochemical recycling. In this respect, it may not be a coincidence that some geochemical characteristics of oceanic island basalts could be well explained if some ancient subducted continental crusts were incorporated in the source region within the mantle (e.g., Zindler and Hart, 1986).

The exhumation of UHP rocks from mantle depths to the Earth's surface has been a stimulating as well as a controversial issue during the past decade. Both *P-T* path studies (Fig. 4) and thermal modeling required that the exhumation rate be high, and that the exhumation be accompanied by continuous cooling (refrigeration) to preserve UHP minerals (Chopin *et al.*, 1991; Hacker and Peacock, 1995). Any exhumation model has to meet these two requirements as well as be consistent with the structural, petrological and geochronological constraints.

Although the opinions on the exhumation of UHP terranes are diverse, they can be roughly grouped into two schools of thought: (1) "contractional" exhumation through thrusting and erosion (e.g., Hsu, 1991; Okay and Sengor, 1992) and (2) "extensional" exhumation through low-angle (listric) normal faulting (e.g., Platt, 1986; Andersen and Jamtveit, 1990). Since it is generally agreed that continuous shortening has been taking place across the Alpine transect from 100 Ma up to the present (e.g., Dewey *et al.*, 1989), the thrusting/erosion model was, therefore, an involuntary choice to explain the exhumation of UHP rocks in the Dora Maira Massif as well as in other UHP terranes. This model, however, requires a large degree of crustal erosional stripping. The lack of appropriate amounts of coeval sediments to account for the necessary stripping was, therefore, thought as a major flaw for the "contractional" exhumation model. To overcome this difficulty, the importance of extensional orogenic collapse as a mechanism for the exhumation of UHP rocks was proffered (e.g., Platt, 1986). "Extensional" exhumation can be triggered by two processes: lithospheric delamination and continuous underplating. The former process is considered implausible because it is likely to impose a high-temperature thermal structure at the base of the thickened crust and would tend to cause heating during decompression. The latter process, however, could be attractive because it may cause internal extension with concomitant refrigeration within an overall compressional tectonic setting. The extensional tectonics can also well explain the common occurrence of UHP rocks which are overlain by substantially lower-pressure rocks in UHP terranes (Platt, 1986). However, it should be emphasized that during continental collision, variations in the subduction rate, the thickness of inflowing crust and sediments, the mechanism of accretion (frontal or underplating) and the rheology of subducted crust may all affect the orogenic processes in different ways, so that periods of internal contraction and extension may alternate, thus producing a complex deformation/exhumation history (Platt, 1986). Recent detailed structural studies have shown that most structural elements observed in UHP terranes actually postdate UHP metamorphism and developed even after substantial exhumation of UHP rocks (Chopin *et al.*, 1991; Michard *et al.*, 1993; Hacker *et al.*, 1995; Xue *et al.*, 1996). Opinions against all-contractional or all-extensional exhumation of UHP terranes, therefore, have appeared (e.g., Michard *et al.*, 1993), leading to a two-stage exhumation model which includes: (1) an initial stage characterized by rapid exhumation from mantle depths to the lower crustal level through forced return, or extrusion tectonics, in an on-going subduction setting, and (2) a second stage characterized by a slower exhumation rate from the lower crust to the Earth's surface through extension or doming (e.g., Michard *et al.*, 1993; Maruyama *et al.*, 1994). It is obvious that the diverse opinions on the exhumation mechanisms for UHP rocks are, in fact, a reflection of the lack of conclusive constraints. It is also uncertain if one exhumation model can be universally applied to all UHP terranes. At the moment, it seems that any exhumation model proposed for any specific UHP terrane could only be considered as a working hypothesis. Systematic structural, petrological and geochronological investigations on every UHP terrane are needed to solve this problem.

VIII. Concluding Remarks

The discovery of microdiamond and coesite inclusions in metamorphic rocks of crustal origin testifies that crustal rocks can be subducted to mantle depths (>100 km) through continental collision. This discovery has also induced many challenges within the realms of experimental petrology, crust-mantle-fluid interactions, UHP geochronology and geodynamic modeling of UHP terranes. The results of these studies have brought much new insight into global geodynamics as well as global geochemical cycling, and more insight will certainly be gained from additional UHP case studies in the future. Among the issues studied, the mechanism(s) that causes exhumation of UHP terranes from mantle depths to the Earth's surface is most intriguing and fascinating. However, due to a lack of sufficient constraints, consensus has not yet been reached.

Up to the present, UHP metamorphic rocks have only been identified from five continent-collision orogenic belts around the world, with ages ranging from the late Proterozoic through the Paleozoic/the Mesozoic to the Cenozoic. Based on the available knowledge and with more detailed studies, more occurrences of UHP metamorphism can surely be expected in the near future. In this respect, some early Proterozoic HP rocks were recently reported from the Snowbird tectonic zone of Canada (2600 Ma) (Williams *et al.*, 1994) and the Usagaran belt of Tanzania (2000 Ma) (Moller *et al.*, 1995). This may suggest that the geothermal gradient of the Earth during the early Proterozoic may not be as high as that during the Archean. The presence of some UHP rocks of the middle to early Proterozoic age is not impossible.

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超高壓變質作用簡介

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摘　要

超高壓變質作用是指變質時壓力條件超過石英—柯石英平衡界線之變質作用。目前在世界上含有微鑽石及柯石英包裹體之地殼源超高壓岩石出露於五個大陸碰撞造山帶中，證實了地殼岩石可因大陸碰撞而隱沒至超過一百公里的地函深度。由實驗礦物／岩石學亦確立了與超高壓變質條件特別相關的一些礦物（群）。這些礦物（群）能保存於地表的溫壓條件，主要肇因於超高壓岩體的快速上升，以及上升時環境缺乏液相和持續的冷凍效應。由於在某些造山帶中，超高壓礦物（群）發現的地點相當侷限，因此實不易據以界定超高壓岩體的區域規模。此外，目前的資料亦顯示超高壓變質作用發生時，環境中水的活性應相當低，且變質液相中亦含有相當量的氬氣，造成超高壓礦物在形成時均含“超氬”而不利於氬氣定年。根據定年工作及變質溫壓的估計，超高壓變質岩體由地函深度上升至下地殼深度的速度應大於7-14釐米／年；唯因缺乏一些關鍵性的控制條件，對超高壓岩體上升的地體構造機制模式，目前尚無定論。