PetroTECTONICS OF ULTRAl-HIGH-PRESSURE CRUSTAL AND UPPER-MANTLE ROCKS—IMPLICATIONS FOR PHANEROZoIC COLLISIONAL OROGENS

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ABSTRACT

Ultrahigh-pressure (UHP) metamorphic terranes in contractional orogens reflect descent of continental crust bonded to a dense, dominantly oceanic plate to depths of 90–140 km. All recognized well-documented UHP complexes formed during Phanerozoic time. Rocks are intensely retrogressed to low-pressure assemblages, with rare relict UHP phases retained in tough, refractory host minerals. Resurrected UHP slabs consist chiefly of quartzofeldspathic rocks and serpentinites; dense mafic + ultramafic lithologies comprise <10% of exhumed masses. Associated garnet-bearing ultramafic lenses are of four general origins: type A peridotite + eclogite pods reflect premetamorphic residence in the mantle wedge; type B masses were mantle-derived ultramafic-mafic magmas that rose into the crust prior to subduction; type C tectonic lenses were present in the oceanic lithosphere prior to underflow; and type D garnet peridotites achieved their deep-seated mantle mineralogy long before—and independent of—the subduction event that produced the UHP-phase assemblages in garnet peridotite types A, B, and C. Geochronology constrains the timing of protolith, peak, and retrograde recrystallization of gneissic, ultramafic, and eclogitic rocks. Round-trip pressure-temperature (P-T) paths were completed in <5–10 m.y., where ascent rates approximated subduction velocities. Exhumation from profound depth involves near-adiabatic decompression through P-T fields of much lower-pressure metamorphic facies. Many complexes consist of thin, allochthonous sheets, but those in eastern China and western Norway are about 10 km thick. Ductilely deformed nappes generated in subduction zones allow heat to be conducted away as sheet-like UHP complexes rise, cooling across both upper and lower surfaces. Thicker UHP massifs also must be quenched. Ascent along the subduction channel is driven mainly by buoyancy of low-density crustal material relative to the surrounding mantle. Rapid exhumation prevents establishment of a more normal geothermal regime in the subduction zone. Lack of H₂O impedes back reaction, whereas its presence accelerates transformation...
to low-P phase assemblages. Late-stage domal uplifts characterize some collisional terranes; erosion, combined with underplating, contraction, tectonic aneurysms, and/or lithospheric plate shallowing, may further elevate mid-crustal UHP terranes toward the surface.

**Keywords:** ultrahigh-pressure metamorphism, subduction-zone metamorphism, continental collision, exhumation of UHP rocks.

**INTRODUCTION**

Most compressional mountain belts form at or near the active edges of continents and/or fringing island arcs. Virtually all result from the underflow of oceanic lithosphere and the consequent transport and descent of spreading centers, oceanic plateaus, island arcs, far-traveled microcontinental terranes, and/or continental crustal salients beneath the continental lithosphere. The downgoing slab is subjected to relatively high-pressure (HP), low-temperature subduction-zone metamorphism, which produces lawsonite and jadeitic pyroxene-bearing assemblages, and mafic blueschists. Long-continued subduction results in the construction of a massive calc-alkaline volcanic-plutonic arc on the crust of the stable, hanging-wall plate, but consumption of a small intervening ocean basin prior to collision does not generate a substantial arc. Most mountain chains are a reflection of their specific geography and unique plate-tectonic history; each orogen tends to exhibit major structural and petrologic contrasts along its length. Some sialic collisional belts contain mineralogic relics reflecting ultrahigh-pressure (UHP) stages of prograde recrystallization. Ultrahigh-pressure conditions are defined as those in which the high-pressure polymorphs of silica and carbon (i.e., coesite and diamond) are stable. Other dense phases and mineral assemblages, including Si- and K-bearing pyroxene, Mg-rich garnet, and eclogite-facies rocks are stable under such remarkable pressure-temperature ($P$-$T$) conditions.

Two main end members have been defined, but it is clear that all gradations exist between continent collisional (Alpine-type) and circum-Pacific (Pacific-type) compressional mountain belts. Similar to Pacific HP metamorphic belts, UHP Alpine orogens mark convergent plate junctions (e.g., Hacker et al., 2003a; Ernst, 2005). The former are characterized by the underflow of thousands of kilometers of oceanic lithosphere, whereas the latter involve the consumption of an intervening ocean basin followed by the submerging of an outboard island arc, microcontinent, or promontory of sialic crust against the nonsubducted continental margin. During collision, crustal sections may reach depths approaching 90–140 km, as indicated by the metamorphic crystallization of UHP indicator minerals, phases, and assemblages that are only stable at pressures exceeding ~2.5 GPa. On resur- rection, many collisional UHP terranes consist of an imbricate stack of tabular sheets (Ernst et al., 1997). The Dabie-Sulu belt of east-central China, the Western Alps, the Kokchetav Massif of northern Kazakhstan, the western Himalayan syntaxis of northern Pakistan, and the Western Gneiss Region of Norway constitute the best-documented examples of exhumed UHP rocks. In all these complexes, scattered UHP phases are partially preserved in strong, tough, refractory zircon, pyroxene, and garnet—minerals characterized by great tensile strength and low rates of intracrystalline diffusion. Armoring of the UHP inclusions subjects them to high confining pressure, provides spatial separation from the recrystallizing matrix minerals and rate-enhancing intergranular fluids, and thus protects them from back reaction during decompression.

This review tries to assess the nature of the orogenic process from a general petrotectonic viewpoint, concentrating on the architectures and rock assemblages of Phanerozoic UHP complexes. Although Precambrian analogues may have resulted from the operation of comparable lithospheric plate motions, where systematic lithotectonic contrasts were related to the higher geothermal gradients that attended a younger, hotter Earth, the ancient rock record is less clear; for this reason, unambiguously ancient UHP complexes have not yet been well documented; accordingly, we concentrate on Phanerozoic collisional mountain belts in this synthesis.

Exhumation of deeply subducted UHP complexes involves near-adiabatic decompression through the $P$-$T$ fields of much lower-pressure metamorphic facies. Thus, back reaction, especially where kinetically enhanced by the presence of an aqueous fluid, causes recrystallization and obliteration of the earlier UHP phases. Although volumetrically dominant in exhumed UHP complexes, quartzofeldspathic and pelitic rocks generally retain very few relics of the maximum physical conditions, whereas eclogites and some anhydrous peridotites, because they are relatively impervious to the diffusion of H$_2$O, have more fully preserved effects of the deep-seated processes (Ernst et al., 1998; Liu et al., 1998). The index minerals coesite and diamond are largely lacking in mafic and ultramafic rock types; hence we attempted to quantify the $P$-$T$ conditions of putative UHP rocks by employing thermobarometric computations as well as phase-equilibrium experiments on rocks and minerals.

**PRESSURE-TEMPERATURE CONDITIONS OF ULTRAHIGH-PRESSURE METAMORPHIC COMPLEXES**

HP and UHP terranes are typified by the presence of mafic (and/or ultramafic) eclogite-facies rocks. However, $P$-$T$ determinations on eclogites are inherently difficult because most contain only two silicate phases, garnet and clinopyroxene. Measuring the Fe-Mg exchange between these two minerals enables cal-
culation of temperature, but additional phases such as phengite or kyanite are required for barometry. Even for simple Fe-Mg exchange reactions, two problems render temperature calculation via this method tenuous: (1) diffusional reequilibration during retrogression ensures that recovery of the peak temperature is unlikely—especially at the highest temperatures; and (2) the P-T position of an Fe-Mg exchange reaction cannot be calculated accurately unless the Fe$^{3+}$/Fe$^{2+}$ ratios of the iron-bearing phases, particularly clinopyroxene, are known. The former problem is well known (Pattison et al., 2003), but the magnitude of the latter problem perhaps is not widely appreciated. Krogh Ravna and Paquin (2003) summarized the results of half a dozen studies that compared ferrous/ferric ratios calculated by charge balance with ratios measured by Mössbauer, micro-XANES (X-ray absorption near edge structure), or titration. They found that Fe-Mg garnet clinopyroxene temperatures calculated without knowledge of mineral Fe$^{3+}$/Fe$^{2+}$ typically had uncertainties of ±100 °C. Proyer et al. (2004) used the Mössbauer milliprobe to demonstrate that the problem can be even worse, with apparent temperatures as much as 300 °C too high. Unfortunately, only a handful of Fe$^{3+}$/Fe$^{2+}$ measurements on UHP eclogites have been made, so this method has not found general application.

A better solution to both of these difficulties with eclogite thermobarometry is to use net-transfer reactions rather than exchange equilibria, although garnet activities for Ca-rich solid solutions also can be problematic. The retrograde diffusional reequilibration problem is solved or at least reduced because the increase in diffusive length scale from grain scale in net-transfer reactions to grain-boundary scale in exchange reactions vastly increases the ability to capture peak temperature, and the problem with ferrous/ferric ratios is solved by using equilibria that involve Mg rather than Fe. In eclogites, the two principal equilibria of choice are (Nakamura and Banno, 1997; Ravna and Terry, 2004):

\[ \text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} + \text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12} + \text{K}\text{MgAl}_2\text{Si}_4\text{O}_{10}(\text{OH})_2 = \text{CaMgSi}_2\text{O}_6 + \text{KAl}_2\text{Al}_2\text{Si}_6\text{O}_{18}(\text{OH})_2 \]

(pyrope + grossular + celadonite = diopside + muscovite);

and

\[ \text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} + \text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12} + \text{Si}_2 = \text{Al}_2\text{Si}_2\text{O}_5 + \text{CaMgSi}_2\text{O}_6 \]

(pyrope + grossular + coe/qtz = kyanite + diopside).

Unfortunately, kyanite-phengite eclogites make up only a small portion of the total eclogite population, greatly restricting the applicability of this method. This limitation is offset, however, by the great advantage of the robust pressures and temperatures determined by this method.

We applied this method to calculate accurate eclogite P-T conditions from microprobe mineral analyses presented in the literature. The positions of net-transfer equilibria were calculated using two approaches: (1) using THERMOCALC v. 3.1 (Powell et al., 1998) with the May 2001 updated database of Powell and Holland (1988); and (2) employing the spreadsheet of Ravna and Terry (2004), which depends on the same data set, but involves the Ganguy et al. (1996) garnet activity model rather than the Newton and Haselton (1981) model used by THERMOCALC. Only data from the latter model (Hacker, 2006) are shown in Figure 1; P-T data for the former are similar but are slightly more dispersed. In samples for which a range of mineral compositions was reported, we calculated P-T conditions using the most jadeite-rich omphacite, the most Si-rich white mica, and garnet with the highest a$_{py}$$a_{gr}$ (prp = pyrope; gr= grossular), following the logic outlined by Carswell et al. (2000). Where possible, we supplemented these data with other robust temperature determinations (e.g., oxygen isotope temperature measurements from Dora Maira by Sharp et al., 1993).

Several important conclusions can be obtained from this diagram. As determined by this technique, the temperature range of UHP kyanite-phengite eclogites is 550–1000 °C, although most values are 600–750 °C; the maximum pressure is slightly in excess of 4 GPa for collisional terranes. This P-T field is smaller than that determined for kyanite-phenelite-free eclogites for which Fe$^{3+}$/Fe$^{2+}$ had to be assumed. None falls on the high-pressure side of the “forbidden zone,” defined as the array of geotherms less than 5 °C/km (Liou et al., 2000; but see Schmid et al., 2003). Within uncertainty, most of the determinations fall along the granite, tonalite, and metasediment solidi. This may indicate that: (1) UHP rocks that experienced hypersolidus temperatures recrystallized continuously in the presence of melt, and then froze in mineral compositions during cooling; (2) UHP rocks that have been subjected to hypersolidus temperatures are rarely exposed at Earth’s surface; or (3) few UHP rocks are produced at hypersolidus temperatures. In contrast, as described farther on, some garnet peridotites have computed conditions of crystallization that fall within the high-pressure realm of the “forbidden zone.”

**GENERATION AND EXHUMATION OF UHP METAMORPHIC COMPLEXES**

Ductilely deformed nappes and thrust sheets formed in subduction channels (e.g., Koons et al., 2003; Hacker et al., 2004; Terry and Robinson, 2004) make up the architecture of most recovered HP-UHP complexes; others may represent coherent, non-nappe sections of continental lithosphere (Young et al., 2007). Ascent to shallow crustal levels reflects one or more of several processes: tectonic extrusion (Maruyama et al., 1994, 1996; Searle et al., 2003; Mihalynuk et al., 2004); corner flow blocked by a hanging-wall backstop (Cowan and Silling, 1978; Cloos and Shreve, 1988a, 1988b; Cloos, 1993); underplating combined with extensional or erosional collapse (Platt, 1986, 1987, 1993; Ring and Brandon, 1994, 1999); and/or buoyant ascent (Ernst, 1970, 1988; England and Holland, 1979; Hacker, 1996; Hacker et al., 2000, 2004). Old, thermally relaxed, sinking oceanic lithosphere appears to roll back oceanward more
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rapidly than the nonsubducted plate moves forward (Molnar and Atwater, 1978; Seno, 1985; Busby-Spera et al., 1990; Hamilton, 1995), so compression and extrusion of subducted sialic slabs in such convergent plate junctions cannot be responsible for the exhumation unless the oceanic lithosphere tears away. Constriction by a backstop requires buoyancy or tectonic contraction to produce the return flow of subducted sections. Extension and erosion help to unroof HP–UHP terranes once they reach crustal levels, but these processes do not produce the major pressure discontinuities (up to >2 GPa) that mark the major fault boundaries between deeply subducted and nonsubducted crust (Ernst, 1970; Ernst et al., 1970; Suppe, 1972).

Buoyancy coupled with erosional decapitation provides a plausible mechanism for the exhumation of low-density crustal slices propelled upward from great depth by body forces. Geologic relationships, laboratory scale models (Chemenda et al., 1995, 1996, 2000), and numerical simulations (Beaumont et al., 1996, 1999; Pysklywec et al., 2002), illustrated schematically in Figure 2, document this process (see also volumes edited by: Parkinson et al., 2002; Carswell and Compagnoni, 2003; and Malpas et al., 2004). The strengths and integrity of the subducted lithospheric materials, extents of deep-seated devolatilization, and rates of recrystallization strongly influence the characteristics of the resultant UHP metamorphic belts (Ernst et al., 1998). The petrotectonic features of Phanerozoic UHP complexes thus reflect their plate-tectonic settings and P-T histories (Table 1).

Attending circum-Pacific subduction of a largely sedimentary mélange, devolatilization and increased ductility cause

decoupling of subducted HP materials from the downgoing oceanic plate at ~20–50 km, followed by piecemeal ascent. In contrast, for a continental salient well bonded to the lithosphere, disengagement of a coherent crustal slice from the descending oceanic plate may be delayed to a depth of 90–140 km. The insertion of increasing amounts of low-density material into the subduction zone gradually reduces the overall negative buoyancy of the lithosphere. Attainment of neutral buoyancy at moderate upper mantle depths, where the plate is in extension (Isacks et al., 1968), may result in rupture and accelerated sinking of the dense oceanic lithosphere. Slab breakoff (Sacks and Secor, 1990; von Blanckenburg and Davies, 1995) increases the net buoyancy of the updip, relatively low-density sialic UHP complex and allows sheets to disengage from the oceanic plate and move back up the subduction channel (van den Beukel, 1992; Davies and von Blanckenburg, 1998). During collision, decoupling and exhumation also may be enhanced as the continental crust warms in the upper mantle and passes through the brittle-ductile transition (Stöckhert and Renner, 1998).

The two-way migration of terranes along subduction channels is well known (Ernst, 1970; Suppe, 1972; Willett et al., 1993). Similar to the subduction of circum-Pacific metaclastic mélanges, low-density sialic crustal sections descend at plate-tectonic rates, and at great depth, generate the distinctive HP-UHP prograde mineralogy of Alpine continental collisional complexes (Peacock, 1995; Ernst and Peacock, 1996). Return of these decoupled sections up the subduction channel during exhumation obviates the need to remove 50–100 km of the overlying hanging wall (the mantle wedge acts as a stress guide) by erosion, extensional collapse, or tectonism.

Densities (g/cm³) of unaltered oceanic crust, 3.0, continental material, 2.7, and anhydrous mantle, 3.2, increase with elevated pressure, reflecting the transformation of open framework silicates to more compact layer-, chain-, and orthosilicates. Stable UHP mineralogic assemblages and computed rock densities appropriate for burial depths of ~100 km and 700 °C are roughly as follows: metabasaltic eclogite, 3.55; eclogitic granitic gneiss, 3.05; and garnet peridotite, 3.35 (Ernst et al., 1997; Hacker et al., 2003a). Even when transformed completely to a UHP assemblage, K-feldspar + jadeite + coesite-bearing granitic gneiss remains ~0.30 g/cm³ less dense than garnet lherzolite, whereas metabasaltic eclogite is ~0.20 g/cm³ denser than upper mantle lithologies. Evidently, subducted packets of UHP metamorphosed sialic crust are buoyant enough to overcome the traction of the oceanic plate carrying them downward because quartzofeldspathic nappes are now exposed at the Earth’s surface.

Continental crustal rocks contain muscovite and biotite, minerals stable to 800–1100 °C at subduction depths >140 km (Stern et al., 1975; Nichols et al., 1994; Patiño Douce and McCarthy, 1998), as the main hydrous phases; therefore, such rocks do not devolatilize completely during normal subduction (Ernst et al., 1998). In the absence of a rate-enhancing aqueous fluid, such lithologies are unlikely to transform rapidly, or totally, to UHP mineral assemblages (Hacker, 1996; Austrheim, 1998). In contrast, the main H₂O-bearing phase in mafic rocks is hornblende, which is a pressure-limited mineral that devolatilizes at moderate temperatures where depths exceed ~70–80 km. In the presence of this evolving aqueous fluid, metabasaltic eclogites are far more likely to recrystallize to the stable prograde HP-UHP assemblage than are sialic units. Consequently, at upper-mantle depths, continental crust converted completely or incipiently to UHP-phase assemblages remains buoyant relative to the surrounding mantle and should rise to mid-crustal...
levels; in contrast, eclogitized oceanic crust becomes negatively buoyant compared to both near-surface oceanic basalt and garnet herzolite and continues to sink. This relationship explains why exhumed HP-UHP terranes worldwide consist of ~90% low-density felsic material and contain only small proportions of dense mafic and anhydrous ultramafic rock types.


TABLE 1. SUMMARY DATA FOR ULTRAHIGH-PRESSURE (UHP) METAMORPHIC COMPLEXES

<table>
<thead>
<tr>
<th>Terrane characteristic</th>
<th>Dabie-Sulu belt, coesite-eclogite unit</th>
<th>Kokchetav Massif, UHP unit</th>
<th>Dora Maira Massif, L. Venasca nappe</th>
<th>Western Gneiss region</th>
<th>Western Himalayan syntaxis, Kaghan V.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Protolith formation age</td>
<td>Chiefly 800–650 Ma</td>
<td>2.3–2.2 Ga</td>
<td>Ca. 300 Ma</td>
<td>1.8–0.4 Ga</td>
<td>&gt;170 Ma</td>
</tr>
<tr>
<td>Temperature of metamorphism</td>
<td>650–750 °C</td>
<td>900 ± 75 °C</td>
<td>725 ± 50 °C</td>
<td>600–800 °C</td>
<td>750–780 °C</td>
</tr>
<tr>
<td>Depth of metamorphism</td>
<td>90–125 km</td>
<td>~140 km</td>
<td>90–110 km</td>
<td>90–130 km</td>
<td>~100 km</td>
</tr>
<tr>
<td>Time of metamorphism</td>
<td>236–226 Ma</td>
<td>535 ± 3 Ma</td>
<td>35 Ma</td>
<td>410–405 Ma</td>
<td>44 Ma</td>
</tr>
<tr>
<td>Crustal annealing</td>
<td>230–195 Ma</td>
<td>529 Ma</td>
<td>32 Ma</td>
<td>Ca. 402 Ma</td>
<td>40–42 Ma</td>
</tr>
<tr>
<td>Rise time to mid-crust</td>
<td>6 m.y.</td>
<td>6 m.y.</td>
<td>3 m.y.</td>
<td>3–8 m.y.</td>
<td>2–4 m.y.</td>
</tr>
<tr>
<td>Exhumation rate†</td>
<td>≥10 mm/yr</td>
<td>15–30 mm/yr</td>
<td>~20 mm/yr</td>
<td>8–20 mm/yr</td>
<td>&gt;15 mm/yr</td>
</tr>
<tr>
<td>Coesite inclusions</td>
<td>Rare, locally abundant</td>
<td>Relatively abundant</td>
<td>Absent</td>
<td>2 localities</td>
<td>Absent</td>
</tr>
<tr>
<td>Diamond inclusions</td>
<td>Very rare</td>
<td>Relatively abundant</td>
<td>Absent</td>
<td>2 localities</td>
<td>Absent</td>
</tr>
<tr>
<td>Areal extent</td>
<td>&gt;400 × 50 km</td>
<td>120 × 10 km</td>
<td>35 km²</td>
<td>165 × 50 km</td>
<td>30 × 70? km</td>
</tr>
<tr>
<td>Thickness of individual UHP units</td>
<td>5–15 km</td>
<td>1–3 km</td>
<td>1–2 km</td>
<td>&gt;10 km?</td>
<td>1 km</td>
</tr>
</tbody>
</table>


Average exhumation rates were estimated by dividing depth of UHP metamorphism by time of ascent to 10–15 km crustal depth.

CONTINENTAL COMPLEXES

Considerable effort has been expended to measure the exhumation rates of UHP terranes by radiometric investigations and, to a lesser extent, by diffusion modeling. In general, the most comprehensive studies infer relatively rapid exhumation, approaching plate-tectonic rates. This poses a challenge for geochronologists for several reasons: (1) uncertainties in the decay constants for some radiometric clocks (i.e., 40K and 176Lu) increase the difficulty of obtaining sufficiently accurate ages for pre-Cenozoic rocks; (2) accurate Lu/Hf and Sm/Nd mineral-isochron ages require unzoned, unaltered phases that formed at a single, known P-T stage; and (3) U/Pb ages must have high temporal precision and come from discrete crystal volumes formed at a specific pressure. Advances are being made along all of these fronts, but none of these problems has yet been solved; accurate decay constants (e.g., Begemann et al., 2001) and the ability to analyze subcrustal volumes that can be tied to specific pressures are required. However, the best-documented cases show that exhumation to mid-crustal levels is rapid, with minimum average exhumation rates of tens of millimeters per year (Table 1).

UHP complexes with relatively few geochronological data paint a fairly simple picture. The exhumation rate of the Dora Maira Massif is constrained by U/Pb (chiefly sensitive high-resolution ion microprobe [SHRIMP]), Lu/Hf, and fission-track ages to ~20 mm/yr (see review by Rubatto et al., 2003). The Kokchetav Massif, investigated by Sm/Nd, U/Pb SHRIMP, and 40Ar/39Ar techniques, rose at 15–30 mm/yr (Hermann et al., 2001; Katayama et al., 2001; Hacker et al., 2003b). Sm/Nd and Rb/Sr ages indicate that the Lago di Cignana eclogites of the Lepontine Alps were exhumed at 26 mm/yr (Amato et al., 1999). Pliocene U/Pb ages for UHP rocks in Papua New Guinea indicate exhumation rates of 10–20 mm/yr (Baldwin et al., 2004). The Tso Morari complex of the NW Indian Himalaya was exhumed within 2–4 m.y. (Treloar et al., 2003), evidently at an average rate approaching 20 mm/yr. Geochronological data from the giant UHP terranes in China and Norway are vastly more abundant, and, as a result, more complex, but exhumation rates in the Dabie-Sulu UHP terrane of China certainly exceeded 10 mm/yr (Hacker et al., 2000, 2006), as did those in the Western Gneiss Region of Norway (Carswell et al., 2003a, 2003b; Root et al., 2004, 2005). Such speedy unloading exceeds present-day regional exhumation and erosion rates (Blythe, 1998), implying that modern erosion rates are mischaracterized or that erosion alone did not expose the known UHP complexes.
CONDUCTIVE COOLING OF UHP CONTINENTAL COMPLEXES

Diffusion modeling studies demonstrate that Himalayan UHP rocks were subjected to temperatures >600 °C for only short times during decompression (O’Brien and Sachan, 2001; Massonne and O’Brien, 2003). These complexes evidently were quenched during exhumation. Poor thermal conductivities of rocks account for high-pressure prograde conditions attending underflow, but this property of Earth materials also dictates that deeply buried units retain heat on rapid exhumation. During decompression, UHP complexes exhibit pervasive mineralogic overprinting and assemblages characteristic of heating (typically granulite-facies), maintenance of constant temperature (amphibolite facies), or only modest cooling (greenschist facies). As an example, Figure 3 illustrates prograde and nearly isothermal retrograde P-T-time (t) trajectories calculated for the Paleogene subduction complex of the western Himalayan syntaxis. On decompression, the presence of a rate-enhancing aqueous fluid would have resulted in virtually complete obliteration of all pre-existing UHP-phase assemblages. Lack of catalytic, grain-boundary H₂O in a complex subjected to rapid ascent substantially decreases the rate of retrogression (Rubie, 1986, 1990; Ernst et al., 1998; Mosenfelder et al., 2005), but even so, heat must be effectively withdrawn from the rocks at some early stage during exhumation while the complex is relatively hot, or mineralogic evidence of former HP-UHP conditions would be lost. The preservation of UHP relics in a rising subduction complex is favored by juxtaposition against cooler rocks, such as by extensional faulting against a colder hanging wall and/or by thrusting against a colder footwall (Hacker and Peacock, 1995). Nevertheless, only in optimally favorable kinetic circumstances are any relict UHP phases and/or mineral assemblages preserved.

To first order, the thermal history of a UHP body during decompression is determined by its minimum dimension (i.e., thickness), its rate of ascent, and the temperature of the medium through which it ascends (e.g., Root et al., 2005). Relatively thin ascending slices will exchange heat more effectively than will thicker units. If a thin UHP body ascends slowly through a typical (cool) subduction thermal gradient, the P-T path during ascent can simply be the reverse of that during compression. However, if a thin UHP body ascends slowly through a zone of much hotter rocks—say, through interior portions of the mantle wedge—it may become hot enough that the evidence of UHP metamorphism is obliterated. Most well-characterized UHP terranes show neither of these types of behavior but, instead, near-isothermal decompression down to ~1 GPa. If a UHP body is thick, the heating or cooling of the body interior will be reduced proportional to the square of its thickness. If the rate of ascent is more rapid, the heating or cooling of the body interior will be reduced, following the square root of the ascent rate. In general terms, for a UHP complex to ascend without significant heating or cooling, its minimum dimension (radius or half-thickness) must exceed the characteristic diffusion distance

\[
u = \sqrt[6]{\frac{k \Delta z}{(dz/dt)}}
\]

where \( k \) is thermal diffusivity, \( \Delta z \) is the vertical ascent distance, and \( dz/dt \) is the vertical ascent rate. For example, a UHP body with a minimum dimension of 15 km must ascend at >10 km/m.y., and a UHP body with a minimum dimension of 2 km must ascend at >500 km/m.y. The rapid ascent rates required mean that thin UHP sheets cannot have ascended near-isothermally from 100 km depth in their present shape, but must have cooled, approximating in reverse the subduction-zone prograde P-T trajectory (Chopin, 1984; Rubie, 1984; Ernst, 1988; Ernst and Peacock, 1996). The manner in which thick, decompressing slabs are quenched remains problematic. Of course, for UHP phases to be preserved in even fragmentary form, the ascending complex—thick or thin—must be quenched prior to complete back reaction. Examples of thin and thick UHP sheets are presented in Figures 4–6.

Well-studied exposures in the western syntaxis of the Himalayas (O’Brien et al., 2001; Parrish et al., 2003; Kaneko et al.,
2003) and the Central and Western Alps (Henry, 1990; Michard et al., 1995) include nappes and imbricate slices of UHP continental crust less than 1–2 km thick. Similar aspect-ratio coesite-bearing thrust sheets have also been documented from the northern Western Gneiss Region of coastal Norway (Terry et al., 2000a, 2000b; Terry and Robinson, 2004), and the Kokchetav Massif of northern Kazakhstan (Kaneko et al., 2000; Maruyama and Parkinson, 2000; see also Dobretsov et al., 2006). In contrast, UHP sections at least 5 km thick have been proven by drilling in the Sulu belt of east-central China (Liu et al., 2004, 2007; Z. Zhang et al., 2006). Moreover, other intensively mapped portions of UHP terranes of comparable thickness include the Hong’an-Dabie terrane of eastern China (Hacker et al., 2000, 2004), and major tracts of the southern Western Gneiss Region (Root et al., 2005). Geologic maps and cross sections of Figures 4–6 illustrate the imbricate nature common to all these UHP complexes; the most striking contrast involves the differing thicknesses of the various UHP nappes.

Schematic relations shown in Figure 7 apply to the underflow and later exhumation of HP-UHP sheets. Descent of the low-density crust occurs only if shear forces caused by underflow (F_s) exceed the combined effects of buoyancy (F_b) and frictional resistance along the hanging wall of the subduction channel (F_r). Here, F_s > F_b cos θ + F_r. Decoupling and ascent of a slice of the low-density crust take place where buoyancy is greater than the combined effects of shearing along its footwall and resistance to movement along its upper, hanging-wall surface. In this case, F_s cos θ > F_b + F_r. The mantle wedge guides exhumation, and the rising nappe is emplaced oceanward (outboard) from the site of metamorphism. Where the angle of subduction decreases, the effect of buoyancy lessens during both underflow and exhumation. For HP-UHP complexes to be returned to shallow depths and partly preserved, the rising slab must overcome frictional resistance to sliding, so it must be thick enough for buoyancy-driven ascent, yet thin enough that heat is efficiently removed by conduction across the bounding faults—upper normal and lower reverse. Such kinematic structural relationships have been mapped in many resurrected, relatively thin-aspect-ratio subduction terranes, i.e., the Himalayas (Burchfiel et al., 1989; Searle, 1996; Searle et al., 2001; Kaneko et al., 2003); the Franciscan Complex (Ernst, 1970; Suppe, 1972; Platt, 1986; Jayko et al., 1987); the Western Alps (Henry, 1990; Compagnoni et al., 1995; Michard et al., 1995); the Sanbagawa belt (Kawachi, 1968; Ernst et al., 1970; Banno and Sakai, 1989); and the Kokchetav Massif (Kaneko et al., 2000; Ishikawa et al., 2000; Ota et al., 2000; Maruyama and Parkinson, 2000). Nappes have also been described from the Western Gneiss Region of Norway (Harley and Carswell, 1995; Krogh and Carswell, 1995; Terry et al., 2000a, 2000b); and the Dabie-Sulu belt (Liou et al., 1996; Hacker et al., 1995, 1996, 2000; Webb et al., 1999).
Recrystallized, retrogressed UHP complexes, although less dense than anhydrous mantle, become neutrally buoyant at approximately middle levels of the sialic crust (Walsh and Hacker, 2004). In some cases, further exhumation of such slabs may be the product of contractional tectonism (Maruyama et al., 1994, 1996) or low-density crustal underplating—in either case combined with isostatically compensated regional exhumation and erosional decapitation (Platt, 1986, 1987, 1993). In addition, a drop in overall density of the subducting lithosphere after plate breakoff results in a shallowing of the downgoing, increasingly buoyant slab, and may be partly responsible for the late doming recognized in many exhumed convergent plate junction regimes (Ernst et al., 1997; O’Brien, 2001; O’Brien et al., 2001). Yet another unloading mechanism involves the antithetic faulting typical of some con-

Figure 5. Geologic sketch map (A) of the western and central Alps, and diagrammatic cross section (B) through the southern Dora Maira Massif (DM; after Henry, 1990; Michard et al., 1995). In B, the transect across the Dora Maira Massif, numbers indicate the upward change in recorded pressure in GPa relative to the adjacent underlying unit. The lower Venasca ultrahigh-pressure (UHP) nappe is shown in the gridiron pattern.
Figure 6. High-pressure–ultrahigh-pressure (HP-UHP) domains in the (A) Hong’an area of China and (B) Western Gneiss Region of Norway, after Hacker et al. (2000) and Root et al. (2005), respectively. Gray shades are used to distinguish different units. Cross sections provide a measure of the relatively great thickness of these HP-UHP complexes.
tractional orogens, in which double vergence is produced during terminal stages of the ascent of low-density crust (e.g., Dal Piaz et al., 1972; Ring and Brandon, 1994, 1999).

Exhumation of domal or diapiric bodies of granitic crust appears to be occurring along convergent plate junctions where curvilinear arcs intersect at large angles. At such lithospheric boundary cusps, overthickened continental crust gradually warms and loses strength. Basal portions may partially melt, but in any case, the crust softens, becomes even more buoyant, and rises more-or-less like a salt dome. Such uplifts, shown diagrammatically in Figure 8, have been termed tectonic aneurysms (Zeitler et al., 2001; Koons et al., 2002; Chamberlain et al., 2002). Some appear to be the sites of exhumed UHP terranes (Ernst, 2006).

**Figure 7.** Schematic convergent lithospheric plate-boundary diagram for active subduction, after Ernst and Peacock (1996). (A) Deep burial and thermal structure of a subducted sheet of continental crust. (B) Later decompression cooling of a rising slice of the sialic material. Relative motions of plates and slices are indicated by arrows (the subducting plate actually is sinking and rolling backward; Hamilton, 1995). During ascent of the HP-UHP terrane (thickness exaggerated for clarity), cooling of the upper margin of the sheet takes place where it is juxtaposed against the lower-temperature hanging wall (the mantle wedge); cooling along the lower margin of the sheet takes place where it is juxtaposed against the lower-temperature, subduction-refrigerated lithosphere. Exhumation of low-density slices requires erosive denudation and/or gravitational collapse and a sialic root at depth. The resolutions of forces acting on the sialic slab in stages A and B are discussed in the text. Lithosphere is shaded (crust-mantle boundary not indicated); asthenosphere is unshaded. Degrees in Celsius.

**TECNOLOGIC SIGNIFICANCE OF GARNET PERIDOTITES IN UHP CONTINENTAL COMPLEXES**

Studies of volumetrically minor mafic eclogite boudins and layers in subducted continental crust have provided important quantitative constraints regarding the UHP conditions that attended metamorphism of the enclosing, largely quartzofeldspathic complex. The occurrences of spatially associated garnet-bearing peridotite bodies are less well understood. Such ultramafic rocks occur as tectonic massifs, pods, and lenses in many ancient collisional mountain belts. HP-UHP examples include the Caledonian, Variscan, and Alpine orogens of Europe, the Kokchetav Massif of Kazakhstan, and the Triassic Dabie-Sulu terrane in east-central China (for reviews, see Medaris, 1999; Brueckner and Medaris, 2000; O’Brien, 2000). These garnet peridotites are of contrasting origins. Some have been interpreted as mantle-derived bodies tectonically emplaced into sialic crustal sequences (Ernst, 1978; Carswell and Gibb, 1980), whereas, others are regarded as products of prograde HP metamorphism of spinel peridotite, or their serpentinized equivalents, previously emplaced in the crust (e.g., Evans and Trommsdorff, 1978; England and Holland, 1979). Medaris (1999) subdivided garnet peridotites from Eurasian HP-UHP terranes into four general types: (1) serpentinites or ultramafic igneous complexes emplaced in the crust prior to subduction, followed by underflow and UHP metamorphism; (2) mantle wedge spinel and/or garnet peridotites inserted into a downgoing lithospheric plate; (3) low-pressure, high-temperature spinel peridotites that may reflect the upwelling of asthenospheric material; and (4) HP garnet peridotites tectonically extracted from the deepest portions of the continental crust–capped lithosphere.

Quantitative compositional and structural data for the subcontinental lithospheric mantle provide crucial information for the erection of realistic large-scale models describing Earth’s geochemical and tectonic evolution (Griffin et al., 1999). Our knowledge of mantle compositions and heterogeneities has been obtained mainly through the study of xenoliths and xenocrysts from kimberlites and volcanic rocks of deep origin. However, detailed, integrated petrochemical, mineralogic, and geochronologic studies of orogenic garnet peridotites provide important constraints on mantle processes, and the chemical-mineralogic compositions and evolution of the mantle wedge overlying a subduction zone. The discovery of phases of very deep origin, such as majoritic garnet, HP clinopyroxenite, and olivine containing elevated concentrations of FeTiO₃ rods in garnet peridotites from several UHP terranes (e.g., Dobrzhinetskaya et al., 1996; Bozhilov et al., 1999; van Roermund et al., 2000, 2001; Massonne and Bautsch, 2002) has provided important information about mantle dynamics. How these deep-seated (>200 km) mantle rocks were transported to shallow depths, and by what means they were incorporated in subducted continental slabs of contractional mountain belts remain unclear. Some of these garnet peridotites and the enclosing continental crust have been postulated to have undergone subduction-zone UHP metamor-
phism characterized by extremely low thermal gradients, on the order of \(\leq 5\) °C/km (e.g., Liou et al., 2000; Zhang et al., 2004). High-pressure experiments reveal that numerous hydrous phases may be stable in such HP environments, the so-called forbidden zone (Liou et al., 1998). Thus, unusually cold subduction zones might well represent the sites of major recycling of H\(_2\)O back into the mantle. These findings have advanced our quantitative understanding of the thermal structure of subduction zones and of the return of volatiles to the mantle.

Most Eurasian HP-UHP garnet peridotites are rich in Mg and Cr and represent depleted-upper-mantle materials, but several are more Fe-rich and originated as igneous mafic-ultramafic complexes (Medaris, 1999). These peridotites are polymetamorphic, with UHP garnet-bearing assemblages extensively replaced by a succession of retrograde mineral assemblages generated during exhumation and cooling. Some peridotites also contain evidence for a pre-UHP stage, evidenced by spinel and/or Ti-clinohumite inclusions in garnet. Equilibration conditions of peak-UHP stages have been calculated from garnet-bearing peridotites by employing the olivine-garnet Fe-Mg exchange thermometer and the Al-in-orthopyroxene barometer (but see section dealing with \(P-T\) conditions of UHP metamorphism). Garnet peridotites occur as meter- to kilometer-sized blocks and lenses in gneisses; the quartzofeldspathic host rocks also have been subjected to UHP metamorphism and exhibit massive, granoblastic or porphyroblastic textures. Most garnet peridotites are deformed and are partially to almost fully serpentinized. Relict garnet-bearing assemblages on the surface are more completely preserved in the central parts of such ultramafic boudins; some occupy up to 30 vol% of the entire body. Garnet peridotite samples from drill holes, however, tend to be relatively less intensely serpentinized. Exsolution microstructures in olivine, garnet, and diopside, and clinohumite polymorphs of orthopyroxene are common (Dobrzhinetskaya et al., 1996; Zhang and Liou, 1999, 2003; Zhang et al., 1999, 2003; Spengler et al., 2006). Geochronologic data for garnet peridotites are poorly constrained, but associated mafic eclogites have been dated by various methods, such as Sm-Nd mineral isochrons and SHRIMP zircon U-Pb dating (Katayama et al., 2003; Zhang et al., 2005a, 2005b; Z. Zhang et al., 2006; Zhao et al., 2007).

Figure 8. Diagrammatic cross section of the Neogene tectonic aneurysm at the western Himalayan syntaxis, simplified from Zeitler et al. (2001). Erosion-induced rapid unloading of high mountains overlying deep-seated, thickened crust causes upward flow of thermally softened, buoyant crust. Numbered features are as follows: (1) hot, ductile, devolatilizing metamorphosed crust enters flow regime, and (2) passes through high-strain zone, incipiently melting and degassing further. (3) Crust enters region of rapid exhumation as unloading and further melting take place, with granitoids (4) possibly inserted into massif along NW and SE shear zones. (5) Strain focusing leads to accelerated upward advective transport of the lower, ductile crust, carrying along its thermal structure. (6) High topography surmounting the weak diapiric zone is partly removed by vigorous erosion, exposing back-reacted, decompressed migmatites. Also involved laterally is a strong meteoric circulation system (not illustrated). MMT—main mantle thrust.
Numerous tectonic origins for garnet peridotites in UHP terranes have been proposed (e.g., Brueckner, 1998; Medaris, 1999; Zhang and Liou, 1999). Similar to, but slightly different from the classification of Medaris, we infer contrasting origins for HP-UHP garnet peridotites based on their modes of occurrence, petrochemical characteristics, and tectonic histories; the range of properties for some of these bodies is summarized in Table 2. The four general types of ultramafic rock, now recrystallized to garnet peridotite, are as follows: type-A, hanging-wall (mantle wedge) fragments; type-B, crustal mafic-ultramafic igneous complexes; type-C, tectonic blocks from the footwall mantle lithosphere; and type-D, ancient mantle complexes tectonically emplaced in the crust prior to subduction. Inasmuch as many of the Eurasian garnet peridotites listed in Table 2 are incompletely characterized tectonically, geochemically, and/or are undated (particularly by Re-Os isotopic systematics), our assignment of tectonic type must be considered tentative. Moreover, several types of garnet peridotite may occur in certain HP-UHP terranes. For example, in the Western Gneiss of Norway, type-B garnet peridotites with Caledonian HP assemblages occur in addition to type-D Proterozoic UHP assemblages (Jamtveit, 1987). Global locations of some of these garnet peridotites in UHP metamorphic belts are indicated in Figure 9.

**Type-A Garnet Peridotites**

These ultramafic rocks originated in the mantle wedge above a subduction zone. Type-A uppermost mantle peridotites are either residual mantle fragments, or they are peridotite and pyroxenite bodies differentiated from mantle-sourced magma; they possess isotopic and geochemical signatures of the hanging-wall mantle. For example, garnet peridotites from eastern China are in fault contact with enclosing country-rock granitic gneisses, they are massive and relatively homogeneous without layering, they exhibit either near-equitigranular or porphyroplastic textures, and they contain lenses of biminerical coesite-bearing eclogite. Most such garnet peridotites belong to the Mg-Cr type of Medaris and Carswell (1990) and contain more MgO and Cr₂O₃ and less fertile elements such as TiO₂, Al₂O₃, CaO, and FeO than primitive mantle as defined by Ringwood (1975). Type-A garnet lherzolites and pyroxenites preserve mantle δ¹⁸O value ranges for garnet, olivine, and clinopyroxene of +4.8‰–+5.6‰, +4.7‰, and +4.5‰–5.6‰, respectively (Zhang et al., 1998, 2000, 2004). They tend to have low ⁸⁷Sr/⁸⁶Sr (0.7038–0.7044) and ¹⁴⁴Nd/¹⁴⁴Nd (0.5123–0.5124) values. However, some exhibit unusually high isotopic ratios and plot outside the range of mantle values; these anomalous isotopic compositions may be due to later metasomatism and/or contamination by crustal materials. It should be noted that low-pressure, high-temperature garnet peridotites reported by Medaris (1999), such as those from the Bohemia Massif, are included here (see also Carswell and O’Brien, 1993; O’Brien and Röttzler, 2003). Some of these high-temperature bodies were evolved from spinel peridotites, contain abundant inclusions of spinel in garnet, and equilibrated at ~1000–1300 °C.

**Type-B Garnet Peridotites**

Such bodies were derived from ultramafic portions of pre-subduction crustal mafic-ultramafic complexes. The protoliths were produced by differentiation from mafic magma prior to UHP metamorphism (e.g., Z. Zhang et al., 2006); the continental crustal section was then subjected to underflow and HP-UHP metamorphism. Typically, garnet peridotites are interlayered with eclogites of various compositions. Garnet peridotites from the Dabieshan are characterized by: (1) well-developed compositional banding and/or layering; (2) the occurrence of low-pressure mineral inclusions in garnets (e.g., Okay, 1994); (3) the preservation of relatively light isotopic bulk-rock oxygen compositions (δ¹⁸O < 5‰); and (4) an old, presubduction age of intrusion (~500–300 Ma) into the sialic crust, as well as a Triassic (~230–220 Ma) UHP metamorphic age (Chavagnac and Jahn, 1996; Jahn et al., 2003). Type-B mafic-ultramafic igneous complexes exhibit a large range in major-element concentrations, and most contain lower MgO and higher SiO₂, CaO, TiO₂, Al₂O₃, and FeO values than type-A peridotites. Some type-B garnet peridotites contain well-preserved prograde low-pressure mineral assemblages as inclusions in UHP phases. For example, inclusions of sapphireine, corundum, clinohlore, and amphibole occur in garnet porphyroblasts from the Maowu area of the Dabieshan (Okay, 1993). In the Lotru garnet peridotite from the South Carpathians, garnet and orthopyroxene formed as reaction products at the boundary of partly serpentinitized olivine and pseudomorphs after plagioclase, now consisting of amphibole, zoisite, and chlorite (Medaris et al., 2003).

**Type-C Garnet Peridotites**

These tectonic entities were derived from the underlying mantle of subducted oceanic or continental lithosphere. Protoliths of the footwall mantle of a sinking slab in some cases were serpentinitized prior to HP-UHP metamorphism. The ultramafic rocks may represent part of an ophiolitic sequence that was emplaced in the downgoing plate prior to deep underflow. Thus, some Alpine garnet peridotites of the Western Alps are associated with eclogites that recrystallized from rodingitized gabbros, and retain geochemical evidence of earlier seawater alteration. Petrochemically, such garnet peridotites are difficult to distinguish from type-A hanging-wall mantle-derived analogues. Accordingly, only a few garnet peridotites from the Lepontine Alps (e.g., Cima de Gagnone and Monte Duria bodies) have been assigned to this group (Evans and Trommsdorff, 1978).

**Type-D Garnet Peridotites**

These deep-seated mantle fragments were emplaced tectonically at crustal levels prior to subduction. Relict high-pressure, high-temperature peridotitic lenses are present in the Western Gneiss region of coastal Norway; the ultramafic lithologies represent fragments of ancient depleted mantle, the
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<th>Terrane</th>
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<th>Modes of occurrence</th>
<th>Rock types</th>
<th>Mineral assemblage</th>
<th>Peak-stage T-P (°C, GPa)</th>
<th>Metamorphic age (Ma)</th>
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<td>A</td>
<td>Blocks in gneiss</td>
<td>LZ, DN</td>
<td>Grt+Olt+Opx+Cpx</td>
<td>820–920; 4–6</td>
<td>Triassic</td>
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<td>Triassic</td>
<td>Zhang et al. (2005a, 2005b)</td>
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<td>Rizhao</td>
<td>A</td>
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<td>B</td>
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<td>Grt+Ol+Op+Cpx+Chu+Mgs</td>
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<td>DN, HZ</td>
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<td>DN, CP, WR, WB</td>
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<td>PD</td>
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<td>1570–1540; &gt;3.4</td>
<td>Late Cretaceous</td>
<td>Abbott et al. (2005)</td>
</tr>
</tbody>
</table>

Note: Mineral abbreviations are after Kretz (1983).

origins of which are unrelated to the later Caledonian UHP metamorphism (Carswell and van Roermund, 2005). Several garnet peridotite bodies are especially noteworthy because they exhibit evidence of the former stability of megacrystic mineral assemblages that include now-exsolved high-pressure enstatite and majoritic garnet (van Roermund et al., 2000, 2001, 2002; Spengler et al., 2006). Early assemblages in these peridotites possess Sm-Nd Proterozoic ages (Brueckner and Medaris, 1998), close to the igneous crystallization ages of the host granitic gneisses. However, recent in situ Re-Os analysis of sulfides in the garnet peridotites yield a range of Proterozoic and Archean model ages. A Late Archean (3.1–2.7 Ga) protolith age also is supported by whole-rock Re-Os data for dunites from several such bodies (Beyer et al., 2004). The Archean ages bear testament to a process of partial fusion in the mantle that predated formation of the Proterozoic upper crust in the Western Gneiss Region. Apparently, some mantle blocks previously identified as Proterozoic subcontinental lithospheric mantle may represent metasomatized and refertilized Archean mantle.

**Contrasting Physical Conditions of Crystallization of Mafic and Ultramafic UHP Rocks?**

A comparison of the eclogite thermobarometry described in the text and presented in Figure 1 with computed garnet peridotite conditions of equilibration suggests that although some of the latter rock types (type-D and perhaps some type-A ultramafic bodies) evidently recrystallized under subduction-zone geothermal gradients less than 5°C/km, none of the eclogites and associated sialic crustal entities can be proven to have formed at so-called forbidden-zone $P$-$T$ conditions. The reasons for this disparity remain unclear. Possible explanations include: (1) systematic errors were made in the thermobarometric evaluations of peridotites or eclogites, or both lithologies; (2) eclogites and enclosing crustal units re-equilibrated during exhumation-decompression, yielding post-maximum pressures, whereas anhydrous peridotites did not; (3) peridotites characterized by “forbidden-zone” $P$-$T$ conditions of formation are exotic and formed in a mantle environment unrelated to that of the eclogites. Additional geochemical, geochronologic, and phase-equilibrium investigations are needed to address this problem; what is clear is that at least some zones of continental collision produced subducted HP-UHP assemblages that have been recovered from upper-mantle depths characterized by low prograde geothermal gradients.

**CONCLUSIONS FOR PHANEROZOIC CONTRACTIONAL OROGENS**

Continental collision involves the essential and substantial consumption of oceanic lithosphere and the transport of a salient of sialic crust, island arc, or microcontinental fragment to the convergent plate junction. Insertion of continental crust and
underflow to great depths result in the incipient-to-complete transformation of pre-existing low-pressure quartzofeldspathic and mafic-ultramafic lithologies to UHP-phase assemblages. During prograde metamorphism, evolution of H₂O due to the breakdown of hornblende and serpentine kinetically favors the conversion of mafic and ultramafic rock types to stable eclogitic-garnet peridotitic assemblages, whereas, reflecting the higher-pressure stabilities of biotite and muscovite, micaceous granitic gneisses may persist due to the lack of a free aqueous fluid. The spatial association of volumetrically minor amounts of garnet peridotite and mafic eclogite reflects active participation of both mantle and oceanic crust in the UHP subduction-zone deformation and recrystallization. However, worldwide, chiefly continental materials are regurgitated in exhumed UHP complexes, mirroring the low densities of sialic crust relative to mafic and ultramafic lithologies. Such quartzofeldspathic crustal assemblages are propelled upward by body forces, i.e., buoyancy. Characteristic decompression rates exceeding 10 mm/yr in general are comparable to rates of subduction. Many exposed ultrahigh-pressure complexes consist of ~1–2-km-thick allochthonous sheets, but the largest, in east-central China and western Norway, are ~10 km thick. For thin, ductilely deformed nappes, heat is efficiently conducted away as the UHP complexes rise, cooling the sheets across upper and lower fault-bounded surfaces. For such geometries, the rate of ascent need not be especially rapid. In contrast, the manner in which enormous, much thicker, rapidly decompressing UHP complexes like the Western Gneiss Region and the Dabie-Sulu belt are quenched, preserving relict UHP phases, remains enigmatic. In either case, however, surviving UHP bodies must be relatively dry during the ascent; the absence of a separate aqueous fluid accompanying exhumation would retard back reaction in the complex, allowing the scattered retention of early stage, UHP phases.

The significance of tectonic aneurysms is speculative, but it deserves consideration with regard to the mechanism of final exhumation of UHP terranes. Most recognized UHP collisional complexes bear extensive evidence of prior nappe emplacement, so exposure of the deep-seated terranes may reflect the operation in varying degrees of subduction-zone slab imbrication and/or buoyant massif ascent followed by late domal uplift aided by locally vigorous erosion. Due to relatively rapid decompression at moderately high temperatures, the critical requirement for preservation of UHP relict assemblages in at least fragmentary form is effective heat removal; this, in turn, requires that less rapidly decompressing complexes be characterized by large surface/volume ratios. Massif-type buoyant bodies must rise from great depths at near-adiabatic P-T conditions, i.e., extremely rapidly. For recognizable UHP terranes, transport to mid-crustal levels either in décollement-type structures or as giant slabs must occur first, allowing substantial cooling (quenching) of the UHP mineral assemblages. This event is followed by further exhumation combined with erosional collapse; possible late-stage processes include structural contraction, crustal underplating, shallowing of the dip of the subducting lithosphere, crustal back-folding or faulting, or domal ascent as tectonic aneurysms.

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