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Preservation/exhumation of ultrahigh-pressure subduction complexes

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Abstract

Ultrahigh-pressure (UHP) metamorphic terranes reflect subduction of continental crust to depths of 90–140 km in Phanerozoic contractional orogens. Rocks are intensely overprinted by lower pressure mineral assemblages; traces of relict UHP phases are preserved only under kinetically inhibiting circumstances. Most UHP complexes present in the upper crust are thin, imbricate sheets consisting chiefly of felsic units ± serpentinites; dense mafic and peridotitic rocks make up less than ~10% of each exhumed subduction complex. Roundtrip prograde–retrograde *P*–*T* paths are completed in 10–20 Myr, and rates of ascent to mid-crustal levels approximate descent velocities. Late-stage domical uplifts typify many UHP complexes.

Sialic crust may be deeply subducted, reflecting profound underflow of an oceanic plate prior to collisional suturing. Exhumation involves decompression through the *P*–*T* stability fields of lower pressure metamorphic facies. Scattered UHP relics are retained in strong, refractory, watertight host minerals (e.g., zircon, pyroxene, garnet) typified by low rates of intracrystalline diffusion. Isolation of such inclusions from the recrystallizing rock matrix impedes back reaction. Thin-aspect ratio, ductile-deformed nappes are formed in the subduction zone; heat is conducted away from UHP complexes as they rise along the subduction channel. The low aggregate density of continental crust is much less than that of the mantle it displaces during underflow; its rapid ascent to mid-crustal levels is driven by buoyancy. Return to shallow levels does not require removal of the overlying mantle wedge. Late-stage underplating, structural contraction, tectonic aneurysms and/or plate shallowing convey mid-crustal UHP décollements surfaceward in domical uplifts where they are exposed by erosion. Unless these situations are mutually satisfied, UHP complexes are completely transformed to low-pressure assemblages, obliterating all evidence of profound subduction.

Keywords: UHP metamorphism; Exhumation of UHP complexes; Collisional orogens; Alpine–Himalayan belts; Buoyancy drive; Tectonic aneurysms

1. Generation of UHP metamorphic complexes

Similar to circum-Pacific-type high-pressure (HP) metamorphic belts, UHP Alpine-type terranes mark convergent plate junctions (e.g., Hacker et al., 2003a). The former are characterized by subduction of thousands of kilometers of oceanic lithosphere, whereas the latter involve the consumption of an ocean basin followed by insertion of an island arc, microcontinent, or promontory of sialic crust into the suture zone. During Alpine-type continental collision, subducted quartzo-feldspathic sections may reach depths of 90–140 km, as indicated by the solid-state crystallization of

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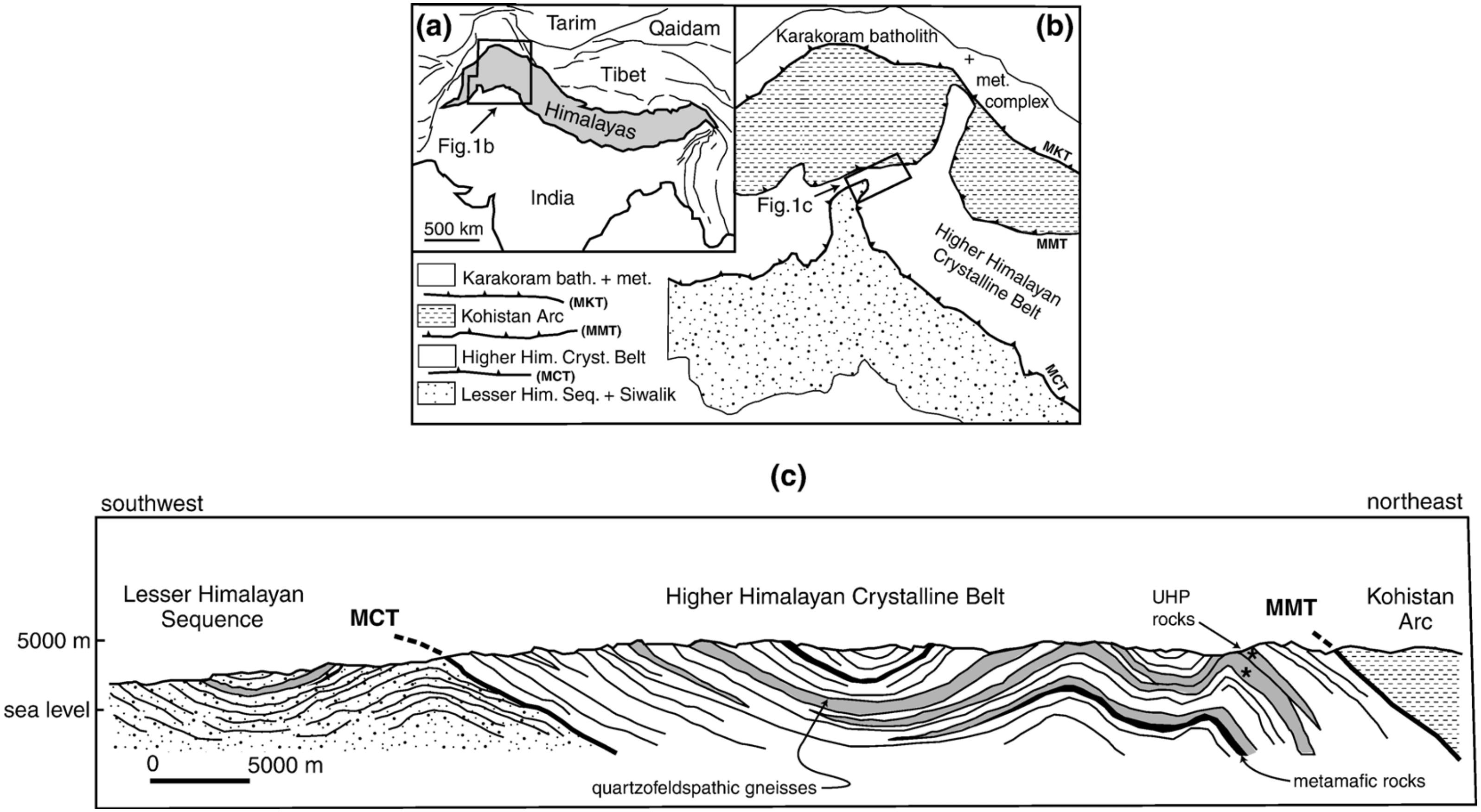

Fig. 1. Simplified (a) index map, (b) regional tectonic scheme, and (c) geologic cross-section looking NW through the Kaghan Valley, western Himalayan syntaxis (Nanga Parbat), Pakistan, after Kaneko et al. (2003). MKT = Main Karakorum thrust; MCT = main central thrust; MMT = main mantle thrust. The Lesser Himalayas and the Higher Himalayas imbricate nappes consist dominantly of pelitic, sandy, and calcareous schists, with lesser amounts of quartzo-feldspathic gneisses (gray pattern), and metamafic units (black pattern). On the NE, eclogitic pods are present in the gneisses. Sparse relict coesite inclusions occur in pelitic and felsic gneisses and in mafic eclogite lenses within the gneisses. Note that coesite-bearing ultrahigh-pressure thrust sheets (asterisks) are thinner than about 1 km.

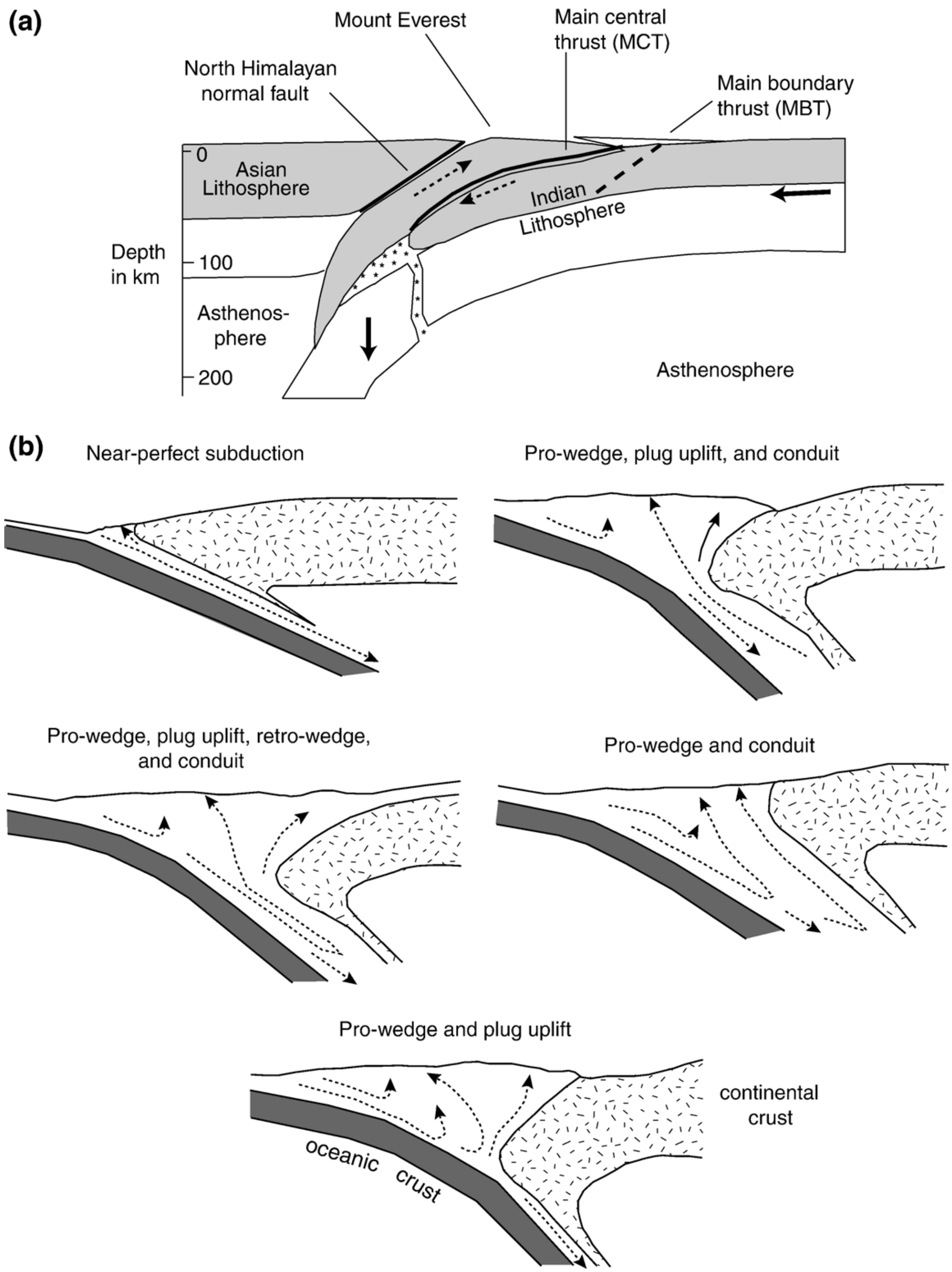

Fig. 2. (a) Simplified structural evolution of the Himalayan orogen, from scale-model experiments by Chemenda et al. (1995). Model shows ascent of a low-density subducted slab after lithospheric plate breakoff. (b) Computational modelling of convergent plate junctions for a wide range of different input parameters, generalized from Beaumont et al. (1996, 1999); all numerical simulations show subduction, followed by regurgitation of some of the low-density material (see also Stöckhert and Gerya, 2005).

neoblastic coesite and/or diamond ± unusual Si-rich garnets and/or K-bearing pyroxenes; such phases are only stable at elevated pressures. On resurrection to the near-surface, most recognizable collisional UHP terranes consist of an imbricate stack of tabular sheets (Ernst et al., 1997). Examples of exhumed ultrahigh-pressure nappes from the Dabie–Sulu belt of east-central China, the Western Alps, the Kokchetav Massif of northern Kazakhstan, the Bohemian Massif of central Europe, and the Western Gneiss Region of coastal Norway are exceptionally well documented. A geologic index map and cross-section through a similar UHP imbricate complex exposed near the western syntaxis of the Himalayas is illustrated schematically in Fig. 1 (Kaneko et al., 2003). Although heavily retrograded to crustal-level metamorphic phase assemblages during exhumation–decompression–recrystallization, two tectonic slices retain scattered relics of their UHP mineralogy. As in similar complexes elsewhere, ultra-high-pressure phases are partially preserved in strong, refractory zircon, pyroxene, and garnet-phases that are characterized by great tensile strength and low rates of intracrystalline diffusion. Armoring of such UHP inclusions provides spatial isolation from the recrystallizing matrix minerals ± intergranular fluids, and protects the inclusions by preventing back reaction during decompression (Rubie, 1986, 1990; Ernst et al., 1998; Mosenfelder et al., 2005).

Thin-aspect ratio, ductile-deformed nappes and thrust slices formed in subduction channels (e.g., Koons et al., 2003: Hacker et al., 2004) make up the upper crustal architecture of most recovered HP and UHP complexes. Ascent to shallow crustal levels seems to have been due to one or more of several processes: tectonic extrusion (Maruyama et al., 1994, 1996; see also Searle et al., 2003); corner flow blocked by a hanging wall backstop (Cowan and Silling, 1978; Cloos and Shreve, 1988a,b; 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). Relatively old, thermally relaxed, sinking oceanic plates and submarine trenches appear to be retreating oceanward (i.e., they rollback) more rapidly than encroachment of the nonsubducted (mantle wedge) lithosphere (Molnar and Atwater, 1978; Seno, 1985; Busby-Spera et al., 1990; Hamilton, 1995), so compression and extrusion of subducted sialic slabs in a convergent plate junction cannot be responsible for the exhumation of some UHP complexes. Constriction by a backstop requires buoyancy or tectonic squeezing to produce the return flow of subducted sections. Faulting associated with extensional collapse and erosional unroofing helps to uncover deeply buried terranes, but does not produce the major pressure

Table 1
Summary data for ultrahigh-pressure metamorphic complexes [a]

Terrane characteristic	Dabie–Sulu coesite–eclogite unit	Kokchetav Massif, UHP unit	Dora Maira Massif, l. Venasca nappe	W. Gneiss Region, Fjordane complex	Kaghan Valley, N. Pakistan	Bohemian Massif Erzgebirge unit
Protolith formation age	Chiefly 650–800 Ma	2.2–2.3 Ga	~300 Ma	1.6–1.8 Ga	>170 Ma	>400 Ma
Temperature of metamorphism	750 ± 75 °C	900 ± 75 °C	725 ± 50 °C	775 ± 75 °C	750–780 °C	650–800 °C
Depth of metamorphism	90–125 km	~140 km	90–110 km	90–130 km	~100 km	95–130+ km
Time of metamorphism	235 ± 5 Ma	531 ± 3 Ma	35–40 Ma	400–410 Ma	48–46 Ma	335–340 Ma
Crustal annealing	220 ± 5 Ma	525 ± 3 Ma	30 Ma	~395 Ma	44 Ma	325 ± 5 Ma
Rise time to mid-crust	15 ± 5 Myr	~6 Myr	3–4 Myr	10 ± 5 Myr	1–3 Myr	~10 Myr
Exhumation rate [b]	6–8 mm/year	>18 mm/year	>20 mm/year	8–12 mm/year	23–45 mm/year	>10 mm/year
Coesite inclusions	Relatively abundant	Rare, locally abundant	Relatively abundant	Rare	Rare	Rare
Diamond inclusions	Very rare	Relatively abundant	Absent	Very rare	Absent	Very rare
Areal extent	>400 × 75 km	~120 × 10 km	225 × 60 km	350 × 70 km	30 × 70? km	40 × 85 km
Max. thickness of individual UHP units	5–10 km [c]	1–3 km	1–2 km	>1 km?	~1 km	~5? km

[a] After Coleman and Wang (1995), Harley and Carswell (1995), Ernst and Peacock (1996), Hacker et al. (2000, 2003b), Maruyama and Parkinson (2000), Terry et al. (2000a,b), Rubatto and Hermann (2001), Massone and O'Brien (2003), and Parrish et al. (2003).

[b] Average exhumation (decompression) rates estimated by dividing depth of UHP metamorphism by time of ascent to 10–15 km crustal depth.

[c] Thickness possibly resulting from contraction and nappe folding during mid-crustal emplacement (Hacker et al., 2004, Fig. 8d).

discontinuities that mark sheared boundaries between deeply subducted nappes and nonsubducted sections (Ernst, 1970; Ernst et al., 1970; Suppe, 1972). However, buoyancy coupled with erosion and mass wastage provides a plausible mechanism for the exhumation of low-density crustal slices, transported upward from great depth by body forces. Geologic field relationships (Fig. 1), laboratory-scale models (Chemenda et al., 1995, 1996, 2000), and numerical simulations (Beaumont et al., 1996, 1999)–the latter two illustrated in Fig. 2–are all compatible with this process (see also volumes edited by: Parkinson et al., 2002; Carswell and Compagnoni, 2003; and Malpas et al., 2004). In addition, recent subduction zone computational modelling has faithfully reproduced the HP–UHP metamorphism and nappe architecture of the Western Alps, without necessitating massive continental collision (Stöckhert and Gerya, 2005).

Table 1 summarizes natures of protoliths, thicknesses of allochthonous ultrahigh-pressure slabs, and times of UHP metamorphism, followed by exhumation for six thoroughly studied Eurasian terranes. The depths listed in this tabulation are minimal values as defined by P–T fields for the stable occurrence of neoblastic coesite and/or diamond. Provided that thermobarometric computations based on Si contents of titanite from a diamondiferous Kokchetav marble (Ogasawara et al., 2002), and discovery of the α-PbO_2 structure in TiO_2 from an Erzgebirge quartzo-feldspathic gneiss (Hwang et al., 2000) are substantiated, considerably higher pressures (6–9 GPa) and depths of metamorphism are preserved. Such great depths (some in fact, even deeper) are supported by thermobarometric investigations of some associated garnet peridotite lenses (e.g., Liou et al., 1998; Zhang et al., 2003), but at least several of these might represent tectonic insertions of old rocks from the deep mantle unrelated to the metamorphism of the crustal section.

2. Body-force exhumation of subduction complexes

Studies (e.g., Ernst, 1971; Chopin, 1987) have shown that deep underflow of low-density material is responsible for the formation of both outboard Pacific- and Alpine-type metamorphic belts. During the circum-Pacific subduction of a chaotic, largely sedimentary mélange, devolatilization and increased ductility promote decoupling of subducted packets from the downgoing oceanic plate at the relatively shallow depths of 20–50 km, followed by ascent. In contrast, for a continental salient well bonded to the sinking lithosphere, disengagement of a crustal slice from the descending plate may be delayed until a depth of 90–140 km is reached. The insertion of increasing amounts of low-density material into the subduction zone gradually reduces the overall negative buoyancy of the lithosphere; achievement of neutral buoyancy at intermediate upper mantle depths, a realm where the plate is in extension (Isacks et al., 1968), may result in rupture and continued descent of the dense, oceanic crust-capped lithosphere. Slab breakoff (Sacks and Secor, 1990; von Blanckenburg and Davies, 1995) increases the effective buoyancy of the up-dip sialic subduction complex, and allows low-density sheets to be sliced off from the oceanic plate and move back up the subduction channel (van den Beukel, 1992; Davies and von Blanckenburg, 1998). Decoupling and exhumation during continental collision also may be enhanced as the quartzo-feldspathic 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 has been documented (Ernst, 1970; Suppe, 1972; Willett et al., 1993). Similar to Pacific-type metaclastic mélanges, low-density continental crust descends relatively rapidly, and disengages at great depth, generating the characteristic HP–UHP prograde mineralogy of Alpine-type collisional complexes (Peacock, 1995; Ernst et al., 1997). The depth at which buoyant sections disengage from the downgoing plate is probably a function of the loss of strength of the sialic crust in passing through the brittle–ductile transition, aided perhaps by breakoff of the oceanic lithosphere. Return of these deep-seated allochthons back up the subduction zone during exhumation obviates the need to remove 50–100 km of the overlying hanging wall (mantle wedge) plate by erosion, extensional collapse, or tectonism.

Ambient densities of unaltered oceanic crust, ~3.0, continental material, ~2.7, and anhydrous mantle, ~3.2, increase with elevated pressure, reflecting the progressive transformation of framework silicates to layer-, chain-, and orthosilicates. Stable ultrahigh-pressure mineralogic assemblages and computed rock densities appropriate for burial depths of about 100 km (Ernst et al., 1997) are: metabasaltic eclogite, ~3.6; eclogitic felsic gneiss, ~3.0; and garnet peridotite, ~3.3. Even fully transformed to an HP–UHP assemblage, K-feldspar+jadeite+coesite-bearing granitic gneiss remains less dense than garnet or spinel lherzolite, whereas metabasaltic eclogite is much denser than this upper mantle lithology. Subducted packets of continental material seemingly are sufficiently buoyant to overcome traction of the oceanic plate carrying them downward, because some such UHP quartzo-feldspathic nappes are exposed at the Earth's surface.

Perhaps more importantly, sialic crustal rocks contain potassic white micas ± biotite as the principal carriers of H_2O. Phengites in these metamorphic assemblages are stable to temperatures exceeding 800 °C at subduction depths of at least 140 km (Nichols et al., 1994; Massone, 1995; Patiño Douce and McCarthy, 1998), hence are unlikely to release aqueous fluids and transform rapidly, or totally, to UHP mineralogy. In striking contrast, the main H_2O-bearing phase in mafic rocks is hornblende, a pressure-limited mineral that devolatilizes at any temperature where depths exceed about 70–80 km; fluxed by rate-enhancing aqueous fluids, metabasaltic eclogites are far more likely to recrystallize to the stable prograde HP–UHP assemblage than are quartzo-feldspathic units. Thus, at moderate upper mantle depths, continental crust transformed completely (or perhaps only incipiently) to UHP phase assemblages would remain buoyant relative to the surrounding mantle, and would tend to rise to mid-crustal levels; in contrast, eclogitized oceanic crust would become even more negatively buoyant than near-surface oceanic basalt, and would continue to sink. This relationship explains why exhumed HP–UHP terranes worldwide consist of at least 90% low-density sialic material, and contain only small proportions of mafic and anhydrous ultramafic lithologies.

Times of UHP recrystallization in well-studied complexes ranges from about 530 Ma in northern Kazakhstan (Sobolev and Shatsky, 1990; Hermann et al., 2001; Katayama et al., 2001; Hacker et al., 2003b) to ~46 Ma in the western Himalayas (Kaneko et al., 2003; Schlup et al., 2003; Treloar et al., 2003), and approximately 35 Ma in the Western Alps (Tilton et al., 1991; Gebauer et al., 1997; Rubatto and Hermann, 2001). Older ultrahigh-pressure slabs eventually may be discovered, but the Earth's ancient geothermal gradient may have been too high to crystallize UHP mineral parageneses during much of Precambrian time.

3. Rate of ascent

Petrotectonic features of Phanerozoic UHP metamorphic belts reflect their plate-tectonic settings (Table 1). The kinds of materials carried down subduction channels, extents of deep-seated devolatilization, and rates of transformation strongly influence the resultant natures of UHP metamorphic belts (Ernst et al., 1998). Exhumation to mid-crustal levels appears to have been driven principally by buoyancy, and the ascent in most cases was surprisingly rapid. Average exhumation rates (i.e., the vertical component of the return path), approaching or exceeding 10 mm/year averaged over approximately > 5 Myr, are required by geochronologic data for HP–UHP terranes (Ernst et al., 1995, 1997; Gebauer, 1996; Gebauer et al., 1997; Hacker et al., 2000, 2003b; Rubatto and Hermann, 2001; Hermann et al., 2001; Rubatto et al., 2003; Schlup et al., 2003; Treloar et al., 2003). Such speedy unloading appears to be greater than currently measured uplift and erosion rates in the Himalayas (Le Fort, 1996; Searle, 1996), but is roughly compatible with exhumation rates of ~4 mm/year calculated by Genser et al. (1996) for the Eastern Alps. Even faster, more nearly comparable ascent rates have been proposed for the eastern Dabie–Sulu belt (Liou and Zhang, 1995; Grasemann et al., 1998), the Kokchetav Massif (Dobretsov et al., 1995), the Tso Morari complex of the NW Indian Himalayas (Massonne and O'Brien, 2003), eastern Taiwan (Lin and Roecker, 1998), the Southern Alps of South Island, New Zealand (Blythe, 1998), the Kaghan Valley of northern Pakistan (Parrish et al., 2003), and eastern Papua, New Guinea (Baldwin et al., 2004).

4. Conductive cooling by subduction zone imbricate faulting

Poor thermal conductivities of rocks are responsible for maintaining the high-*P*/low-*T* prograde metamorphic conditions accompanying underflow, but this property also dictates that deeply buried lithologic units tend to remain warm during rapid exhumation. Rocks cool slowly, and rising, buoyant subduction complexes decompress by passing through lower pressure (in some cases, higher temperature) crustal realms. For this reason, surviving UHP complexes exhibit the pervasive mineralogic overprinting typical of granulite-, amphibolite-, and greenschist-facies *P–T* conditions. For example, Fig. 3 schematically illustrates prograde and retrograde trajectories for the Himalayan Paleogene subduction complex of the western syntaxis. In the presence of a catalytic aqueous fluid, ascent should result in relatively high-temperature obliteration of preexisting UHP phase assemblages. Of course, lack of grain-boundary H_2O during the period of rapid uplift would substantially decrease the rate of retrogression. Nevertheless, heat must be withdrawn efficiently from subduction complexes as they are exhumed, or mineralogic evidence of preexisting HP–UHP conditions would be lost. The uncommon preservation of UHP relics in a rising subduction complex is favored by coeval extensional faulting against the overlying, cooler hanging wall combined with subduction–refrigeration

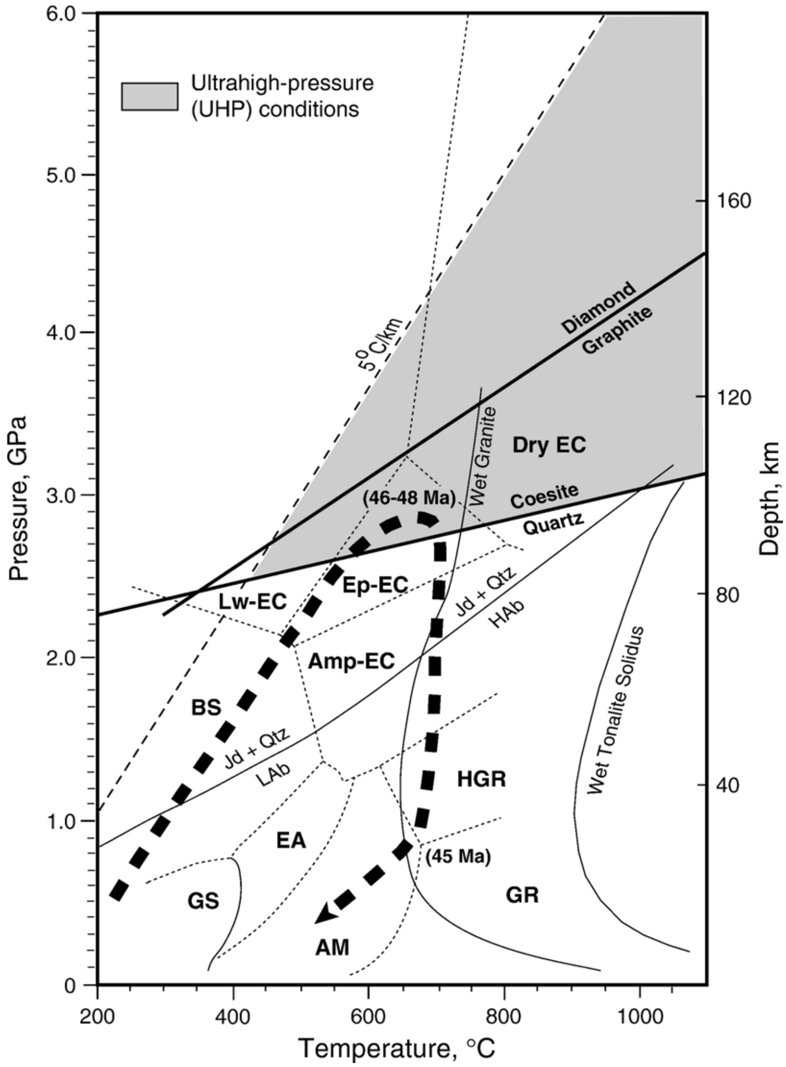

Fig. 3. Pressure–temperature–time path for subduction and exhumation of Kaghan Valley UHP rocks to mid-crustal levels, after Kaneko et al. (2003) and Parrish et al. (2003). See Fig. 1 for the geologic cross-section. The background petrogenetic grid for rocks of basaltic bulk composition is modified from Liou et al. (1998) and Okamoto and Maruyama (1999). An extremely low subduction zone geothermal gradient of 5 °C/km, and UHP conditions are also shown for reference. Citations to the experimental phase equilibria were provided by Liou et al. (1998). Mineral abbreviations are: Jd = jadeite; Qtz = quartz; LAb = low albite; and HAb = high albite. Metamorphic-facies abbreviations are: AM = amphibolite; Amp–EC = amphibolite–eclogite; BS = blueschist; EA = epidote amphibolite; EC = eclogite; Ep–EC = epidote–eclogite; GR = sillimanite–granulite; GS = greenschist; HGR = kyanite–granulite; Lw–EC = lawsonite–eclogite; and Px–Hf = pyroxene hornfels.

tectonically beneath it (Hacker and Peacock, 1995). In these cases, relatively thin, ascending slices such as those shown in Fig. 1 would loose heat along both upper and lower surfaces by thermal conduction. Some complexes rise virtually adiabatically, whereas others appear to have nearly retraced the prograde subduction *P–T* path during decompression (Rubie, 1984; Ernst, 1988; Ernst and Peacock, 1996).

Simple relations shown in Fig. 4 apply to the underflow and exhumation of many tabular HP–UHP terranes. Descent of the low-density sheet occurs only if shear forces caused by underflow (F_s) are greater than the combined effects of buoyancy (F_b) and frictional resistance along the hanging wall of the subduction channel (F_r). Here, $F_s > F_b \sin\theta + F_r$. Ascent of a slice of the low-density material takes place only where buoyancy exceeds the combined effects of shearing along its footwall base and resistance to movement along its upper surface. In this case, $F_b \sin\theta > F_s + F_r$. The mantle wedge guides exhumation, so the rising crustal slice is accreted oceanward from the inboard volcanic–plutonic arc. If subduction inclination decreases, so does the

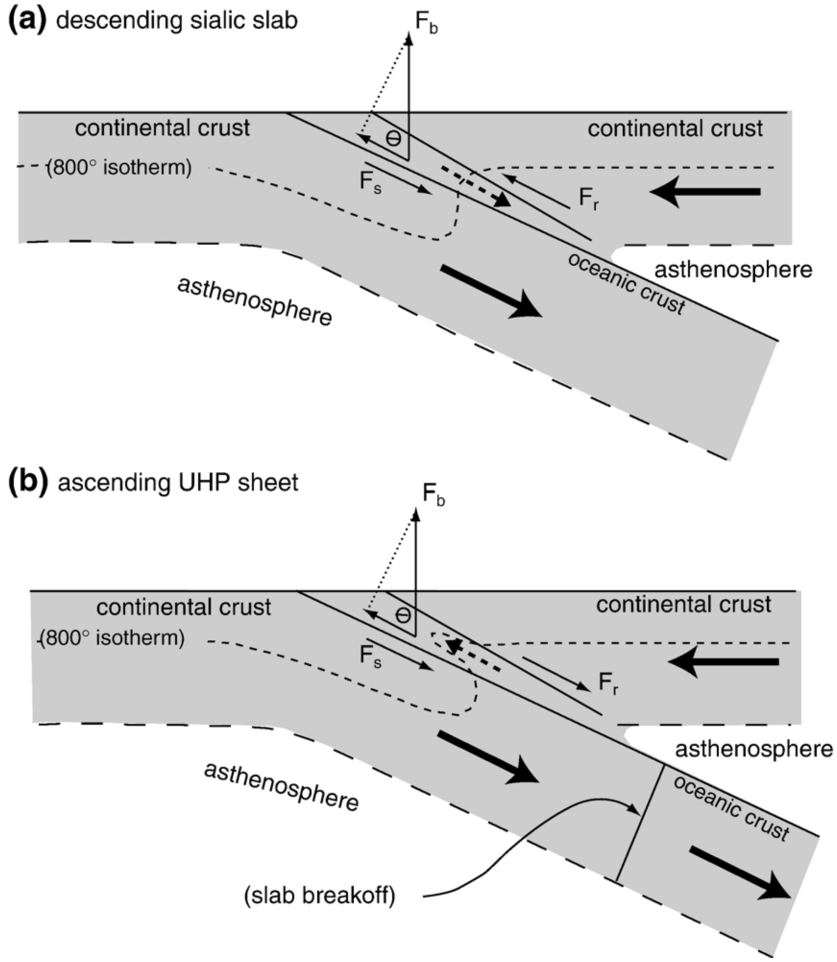

Fig. 4. Schematic convergent plate–boundary diagram after Ernst et al. (1997) during active subduction: (a) deep burial and thermal structure of a prism of quartzo-feldspathic strata, island arc, microcontinent, or continental salient; (b) decompression cooling of a rising slice of low-density continental material. Accompanying ascent of a thin HP–UHP terrane (thickness exaggerated for clarity), conduction cooling across the upper surface of the sheet takes place where it is juxtaposed against the lower temperature hanging wall plate; conduction cooling across the lower surface of the sheet occurs where it is juxtaposed against the lower temperature, subduction-refrigerated lithospheric plate. Buoyancy-driven exhumation of low-density felsic slices requires erosive denudation and/or gravitational collapse, and a quartzo-feldspathic root at depth. The resolution of forces acting on the sialic slab in stages (a) and (b) is discussed in the text. The lithosphere is shaded.

influence of buoyancy during both underflow and exhumation. For HP–UHP mineral assemblages to be returned to shallow depths and partly preserved, the rising slab must overcome the frictional resistance to sliding, so must be thick enough for buoyancy-driven ascent, yet thin enough that heat is effectively removed by conduction across the upper normal and lower reverse faults. Such shear senses and structural relations are well documented in many resurrected subduction terranes, e.g., 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), the Kokchetav massif (Maruyama and Parkinson, 2000; Ishikawa et al., 2000; Kaneko et al., 2000; Ota et al., 2000), the Western Gneiss Region of SW Norway (Harley and Carswell, 1995; Krogh and Carswell, 1995; Terry et al., 2000a,b), the Himalayas (Burchfiel et al., 1989; Searle, 1996; Searle et al., 2001; Kaneko et al., 2003), and the Dabie–Sulu belt (Liou et al., 1996; Hacker et al., 1995, 1996, 2000; Webb et al., 1999).

Recrystallized, retrograded UHP complexes, although less dense than anhydrous mantle, become neutrally buoyant at approximately middle levels of the sialic crust and stall there. In some cases, further uplift of these slabs toward the surface 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 or mass wastage (Platt, 1986, 1987, 1993). In addition, the decrease in overall density of the continental lithosphere after plate breakoff should result in a shallowing of the subduction angle, and may be partly responsible for the late doming recognized in some exhumed convergent plate junctions (Ernst et al., 1997). Yet another unloading mechanism involves the antithetic faulting typical of some compressional orogens, in which double vergence is produced during terminal stages of the buoyant ascent of low-density crust, its collapse and erosional removal (e.g., Dal Piaz et al., 1972; Ring and Brandon, 1994, 1999).

5. Tectonic aneurysms

Rapid uplift of domical bodies of continental crust seems to be occurring along convergent plate boundaries where curvilinear arcs intersect at large angles. These cusps evidently are loci of excess accumulations of sialic material. At such lithospheric plate junction discontinuities, the basal portions of overthickened continental crust gradually warm and soften. Deepest sections may partially melt, but in any case, the accumulated quartzo-feldspathic crust looses strength, becomes buoyant, and rises more-or-less like a salt dome. Uplifts of this sort have been termed tectonic aneurysms, and are explained as a ductile, lower crustal response to local unloading due to erosion by major river systems (Zeitler et al., 2001; Chamberlain et al., 2002; Koons et al., 2002; Fig. 5). However, intense fluvial down cutting may be equally well a function of regional crustal thickness, topographic elevation, increased precipitation and thus erosion–in other words, a consequence rather than a cause of the uplift. Whatever the cause-and-effect relationship, these convergent plate-tectonic cusps appear to have promoted diapiric uplift along the amalgamated margin of the nonsubducted continental plate. Incidentally, the so-called “mantle drip” postulated for the southern Sierra Nevada lithosphere (Saleeby and Foster, 2004; Zandt et al., 2004) may represent a negatively buoyant analogue of the same process that generates a sialic crustal tectonic aneurysm.

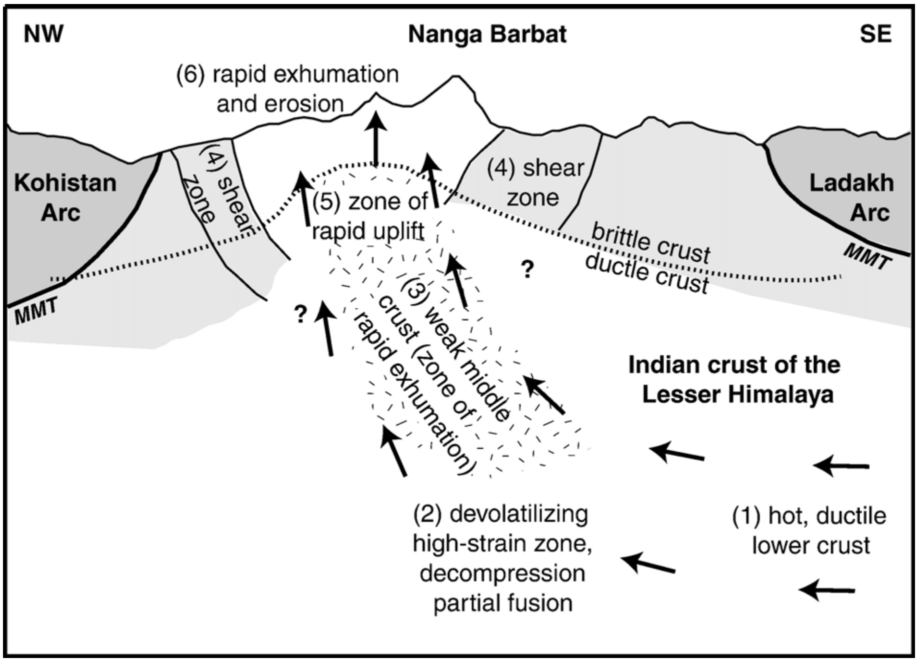

Fig. 5. Diagrammatic cross-section (looking NE) of the Neogene tectonic aneurysm at Nanga Parbat, western Himalayan syntaxis, simplified from Zeitler et al. (2001). Erosion-induced rapid unloading of high mountains overlying deep-seated, thickened crust causes upward flow of thermally softened, buoyant crust. Numbered features are: (1) hot, ductile, devolatilizing metamorphosed crust enters flow regime, and (2) passes through high-strain zone, degassing further. (3) Crust enters region of rapid exhumation as unloading and partial fusion takes place, with granitoids (4) possibly inserted into the massif along shear zones. (5) Strain focusing leads to accelerated upward advective transport of the hot, ductile crust, carrying along its thermal structure. (6) Units constituting the elevated topography above the weak diapiric zone are partly removed by vigorous erosion. This scenario results in the axial exposure of back-reacted low-P/high-T decompressed migmatites, and laterally, a strong meteoric circulation system.

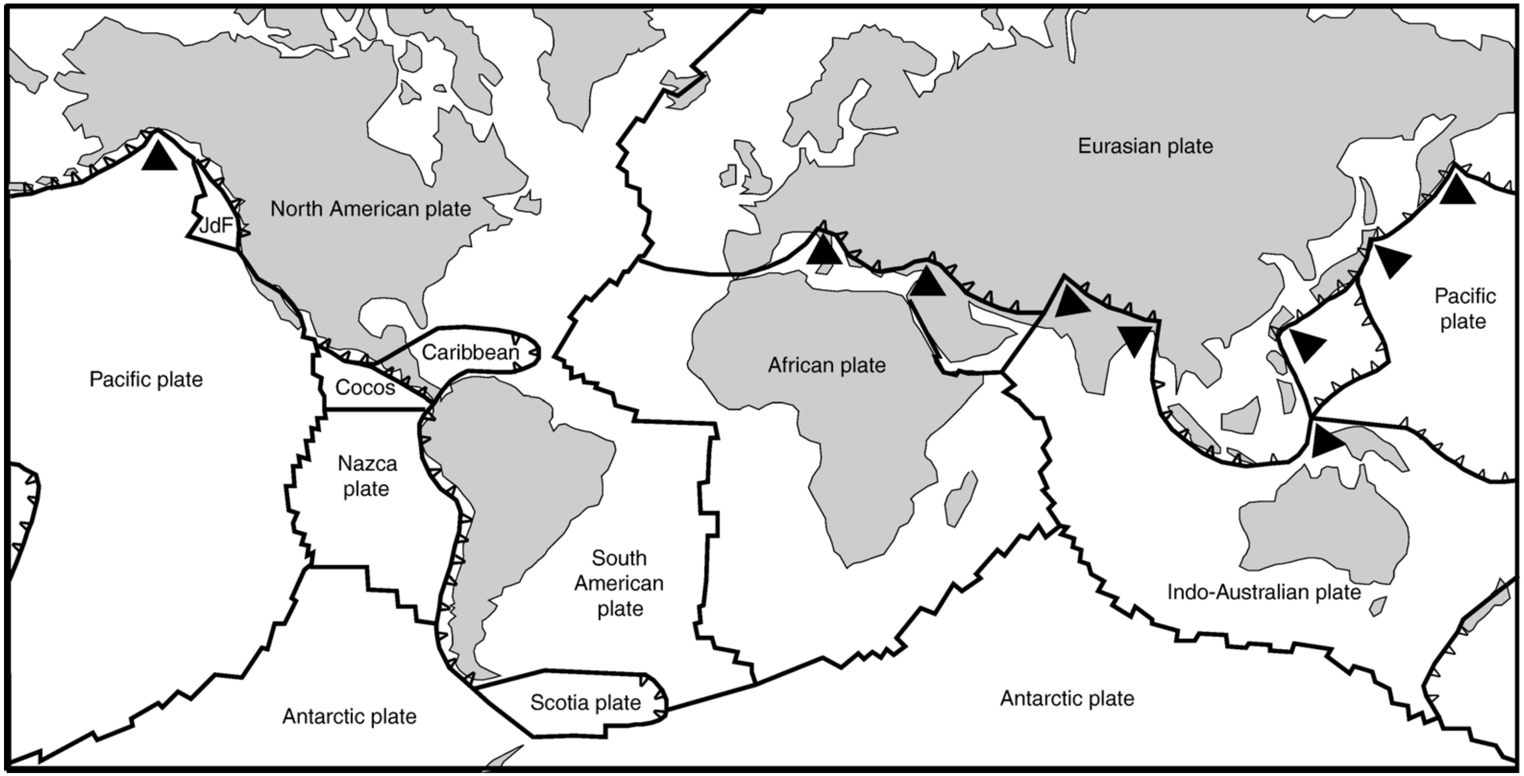

Fig. 6. Generalized plate boundary map of the world (barbs on upper plates at subduction zones), showing convergent arc cusps (black triangles) as possible sites of especially thick continental crust and consequent localization of tectonic aneurysms. JdF = Juan de Fuca plate.

The eastern (Namche Barwa) and western (Nanga Parbat) syntaxes of the Himalayan–Tibetan suture zone display cusp-like compressional plate-tectonic discontinuities, and constitute illuminating examples of the associated rapid uplift and erosion (Burg et al., 1997; Zeitler et al., 2001; Koons et al., 2002). Similar sites include convergent plate junction intersections of the Ryukyu arc–northeastern Taiwan, the Marianas arc-central Honshu, the western Aleutians–Kamchatka Peninsula, the Gulf of Alaska, the Greece–Turkey–Iran contractional suture zone, and possibly the collision zone between the Indonesian archipelago and Australia (see Fig. 6). Extremely rapid uplift and erosion in northeastern Taiwan (Hartshorn et al., 2003; Dadson et al., 2003, 2004) may reflect the presence of a crustal diapir at depth. The areal bulge of the southern Kamchatka Peninsula suggests the possibility that decompression partial melting of a tectonic aneurysm may be responsible for the local breadth of the volcanic–plutonic arc. In addition, the oroclinal bend of the Western Alps and the offset of the Dabie–Sulu orogen (modified by post-collision sinistral slip on the Tan Lu fault) also might represent pre-Cenozoic, now-deformed syntaxes.

The significance of convergent plate cusps is highly speculative, but deserves investigation with regard to the mechanism of final uplift of ultrahigh-pressure terranes. Many of these arc intersections are typified by domical exposures of UHP mineral assemblages (Maruyama et al., 1996), indicating diapiric rise of formerly deeply buried sections of low-density, ductile crust. Ultrahigh-pressure minerals and/or phase assemblages have been reported from: the western Himalayan syntaxis (O'Brien et al., 2001; Kaneko et al., 2003); the eastern Himalayan syntaxis (Zhong and Ding, 1996; Ding and Zhong, 1999; Ding et al., 2001); Greece (Mposkos and Kostopoulos, 2001); Indonesia (Kardarusman and Parkinson, 2000; Parkinson, in press); the Western Alps (Chopin, 1984; Schertl et al., 1991; Compagnoni et al., 1995); and the Dabie–Sulu belt (Wang et al., 1989; Liou et al., 1996; Zhang et al., 2003). These areas bear evidence of prior nappe emplacement, so exposure of the HP–UHP terranes may reflect the operation in varying degrees of (1) early subduction zone slab imbrication/regurgitation and (2) late domical uplift + erosion.

Due to rapid ascent from great depth at moderately high temperatures, the critical requirement for preservation of UHP relict assemblages in even fragmentary form is that heat must be conducted away effectively; this in turn requires rapidly decompressing rock masses to possess large surface/volume ratios. Hence, for surviving ultrahigh-pressure complexes, transport to mid-crustal levels in décollement-type structures must occur first, allowing substantial cooling (quenching) of the UHP assemblages. This event is followed by further exhumation combined with erosional collapse; possible late-stage processes include tectonic compression, crustal underplating, shallowing of the dip of the subducting lithosphere, crustal backfolding or faulting, or domical ascent as tectonic aneurysms.

6. Final words

The geologic complexities of contractional orogenic belts have been studied for nearly two centuries, yet our understanding of them is still evolving. No two are identical; indeed, most mountain chains are unique, and individually exhibit marked petrotectonic contrasts and age relationships along their lengths. A few contain mineralogic relics reflecting UHP stages of recrystallization but because of thorough and complete back reaction, many other compressional lithotectonic belts may have been subjected to similar UHP metamorphism without retaining clear evidence of such conditions. Exhumation of such profoundly subducted complexes involves contractional tectonism coupled with intense surface erosion. How can we interpret subduction zone complexes for which critical evidence of deep burial is at best exceedingly fragmentary, or now perhaps totally obliterated? Evidently we must intensify our search of compressional orogenic realms for new clues reflecting past P–T conditions and early-stage tectonic features.

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