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Alpine and Pacific styles of Phanerozoic mountain building: subduction-zone petrogenesis of continental crust

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ABSTRACT

A broad continuum exists between two distinct end-member types of mountain building. Alpine-type orogenic belts develop during subduction of an ocean basin between two continental blocks, resulting in collision. They are characterized by an imbricate sequence of oceanward verging nappes; some Alpine belts exhibit superimposed late-stage backthrusting. Sediments are chiefly platform carbonates and siliciclastics, in some cases associated with minor amounts of bimodal volcanics; pre-existing granitic gneisses and related continental rocks constitute an autochthonous–parautochthonous basement. Metamorphism of deeply subducted portions of the orogen ranges from relatively high-pressure (HP) to ultrahigh-pressure (UHP). Calcalkaline volcanic–plutonic rocks are rare, and have peraluminous, S-type bulk compositions. In contrast, Pacific-type orogens develop within and landward from long-sustained oceanic subduction zones. They consist of an outboard oceanic trench–accretionary prism, and an inboard continental margin–island arc. The oceanic assemblage consists of first-cycle, in-part mélanged volcaniclastics, and minor but widespread cherts ± deep-water carbonates, intimately mixed with disaggregated ophiolites. The section recrystallized under HP conditions. Recumbent fold vergence is oceanward. A massive, slightly older to coeval calcalkaline arc is sited landward from the trench complex on the stable, non-subducted plate. It consists of abundant, dominantly intermediate, metaluminous, I-type volcanics resting on old crust; both assemblages are thrown into open folds, intruded by comagmatic I-type granitoids, and metamorphosed locally to regionally under high-*T*, low-*P* conditions. In the subduction channel of collisional and outboard Circumpacific terranes, combined extension above and subduction below allows buoyancy-driven ascent of ductile, thin-aspect ratio slices of HP–UHP complexes to midcrustal levels, where most closely approached neutral buoyancy; exposure of rising sheets caused by erosion and gravitational collapse results in moderate amounts of sedimentary debris because exhumed sialic slivers are of modest volume. At massive sialic buildups associated with convergent plate cusps (syntaxes), tectonic aneurysms may help transport HP–UHP complexes from mid- to upper-crustal levels. The closure of relatively small ocean basins that typify many intracratonic suture zones provides only limited production of intermediate and silicic melts, so volcanic–plutonic belts are poorly developed in Alpine orogens compared with Circumpacific convergent plate junctions. Generation of a calcalkaline arc mainly depends on volatile evolution at the depth of magma generation. Phase equilibrium studies show that, under typical subduction-zone *P–T* trajectories, clinoamphibole ± Ca–Al hydrous silicates constitute the major hydroxyl-bearing phases in deep-seated metamorphic rocks of MORB composition; other hydrous minerals are of minor abundance. Ca and Na clinoamphiboles dehydrate at pressures of above approximately 2 GPa, but low-temperature devolatilization may be delayed by pressure overstepping; thus metabasaltic blueschists and amphibolites expel H_2O at melt-generation depths, and commonly achieve stable eclogitic assemblages. Partly serpentinized mantle beneath the oceanic crust dehydrates at roughly comparable conditions. For reasonable subduction-zone geothermal gradients however, white micas ± biotites remain stable to pressures >3 GPa. Accordingly, attending descent to depths of >100 km, mica-rich quartzofeldspathic lithologies that constitute much of the continental crust fail to evolve substantial amounts of H_2O, and transform incompletely to stable eclogite-facies assemblages. Underflow of amphibolitized oceanic lithosphere thus generates most of the deep-seated volatile flux, and the consequent partial melting to produce the calcalkaline suite, along and above a subduction zone; where large volumes of micaceous intermediate and felsic crustal materials are carried down to great depths, volatile flux severely diminishes. Thus, continental collision in general does not produce a volcanic–plutonic arc whereas in contrast, the long-continued contemporaneous underflow of oceanic lithosphere does.

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Introduction

The geological complexities of contractional orogenic belts have been studied for two centuries, yet our understanding of them is still rapidly changing. Most mountain belts form at or near the active edges of continents and/or fringing island arcs. Virtually all result from consumption of lithospheric plates, involving the underflow of oceanic crust, and the consequent transport and descent of spreading centres, oceanic plateaus, exotic island arcs, far-travelled terranes, microcontinents and salients of continental crust. Long-continued subduction allows the construction of a calcalkaline volcanic–plutonic arc on the stable, hangingwall plate. No two orogenic belts are identical. Indeed, most mountain chains are unique, and exhibit important structural and petrological contrasts along their lengths. A few contain mineralogical relics reflecting ultrahigh pressure (UHP) stages of recrystallization, but reflecting thorough and complete retrograde reaction, many other compressional mountain belts which may have been subjected to similar *P–T* conditions of metamorphism fail to

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retain any evidence of such conditions. Interpretation of such subduction-zone complexes for which critical evidence of deep subduction is exceedingly fragmentary, or totally obliterated, is a serious challenge.

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 mountain chains. Although Precambrian analogues may have resulted from the operation of comparable kinds of lithospheric plate motions, with systematic lithotectonic contrasts being related to the higher geothermal gradients of the early Earth, the ancient rock record is less clear. Accordingly, I concentrate on Phanerozoic mountain building in this synthesis. Two main end members are distinguished (e.g. Bally, 1981), but it is clear that all gradations exist between continent collisional and Circumpacific contractional orogens.

Overview of Alpine and Pacific-type orogenic belts

Convergent lithospheric plate junctions

With the development of plate-tectonic theory, Earth scientists realized that oceanic trenches represent the bathymetric expression of the zone of impingence between stable, continental crust-capped and downgoing, oceanic crust-capped lithospheric plates (Isacks *et al.*, 1968). The process of lithospheric underflow has been called Circumpacific or Pacific-type subduction (Bally's type-B subduction). Intracratonal orogenic belts of the Alpine type (Bally's type-A subduction) were recognized as representing ocean-margin suture zones formed by continental collision (Dewey and Bird, 1970; Molnar and Tapponnier, 1975). The amalgamation of continental, microcontinental, and/or island-arc crustal entities occurs in response to the consumption of intervening oceanic lithosphere initially lying between sialic crust-capped plates. Moreover, as emphasized in the terrane concept (Irwin, 1972: Coney *et al.*, 1980; Jones, 1983; Howell, 1985), head-on convergence in general is only possible at specific sites along curvilinear consumptive plate boundaries, reflecting the spherical nature of the globe-encircling lithospheric plates; many sutured terrane amalgams are produced by oblique convergence or by virtually pure strike-slip motion. Thus, Alpine and Pacific-type subduction-induced orogenic belts represent two end-members of a more complex continuum of actual plate-tectonic configurations.

In cases where vast expanses of oceanic lithosphere are consumed without volumetrically significant sialic terranes being transported into the subduction zone, as is characteristic of the Mesozoic + Cenozoic histories of the eastern, NW and SW borders of the Circumpacific, nearly end-member Pacific-type convergent plate boundaries have evolved over time (Engebretson *et al.*, 1985; Stock and Molnar, 1988). This plate-tectonic situation gives rise to an outboard, active-margin accretionary wedge as a largely sedimentary trench complex, a medial, longitudinal forearc basin, and an inboard, roughly contemporaneous calc-alkaline volcanic–plutonic arc (e.g. Kusky *et al.*, 1997). The trench section is typified by a narrow, low-temperature, low-heat-flow belt, and the magmatic arc by a broad region characterized by high temperatures and elevated thermal flux (Miyashiro, 1961, 1967). The former is deposited on oceanic crust, whereas the latter is constructed on pre-existing basement of the continental margin or island arc ± older oceanic crust.

Sea floor spreading may cause the transoceanic transport, arrival at a convergent plate junction, and underflow of a continent, microcontinent, or island arc beneath continental crust capping the stable non-subducted plate (the mantle wedge); the sialic terrane generally descends to considerable depth, drawn down by the leading, high-density ocean lithosphere. The leading edge of the subducting sialic crust may be covered by a passive-margin sedimentary platform that is sutured against an active-margin accretionary wedge forming at the leading edge of the stable, non-subducted plate. Accompanying collision, the aggregate continental thickness is increased by such amalgamation and contraction. Major mountain belts formed along collisional sutures are characterized by numerous allochthonous thrust sheets and a paucity of calcalkaline igneous activity. Typical examples include the Urals, Alps and Himalayas (Hamilton, 1970; Dal Piaz *et al.*, 1972; Ernst, 1973; Molnar *et al.*, 1987; Burchfiel *et al.*, 1989; Searle, 1996; Searle *et al.*, 2001).

Modern transitional convergent plate junctions

Convergent plate-tectonic regimes transitional between these orogenic end-member types are common. An example is the present-day junction between the Arabian Peninsula and Iran (Boudier *et al.*, 1985, 1988; Chemenda *et al.*, 1996; Gnos *et al.*, 1997; Miller *et al.*, 1998; Searle *et al.*, 2004). Another in-progress example of transitional Alpine–Pacific-type subduction is represented by the Indonesian island arc (Hamilton, 1979; Charlton, 1991; Maruyama *et al.*, 1996; Snyder *et al.*, 1996). The strongly curved eastern part of the Australian–Eurasian collisional suture zone between Timor and Seram is the site of underflow of old, cool continental crust of the Australian plate, whereas farther west along the Sumba–Java segment of the convergent plate boundary, oceanic crust-capped lithosphere is descending beneath Indonesia. Relationships are presented schematically in Fig. 1. The exposure of exhumed high pressure (HP) blueschists in the eastern, strongly curved suture zone reflects the decoupling and return towards the surface of subducted, largely quartzofeldspathic material. The driving force for ascent reflects rupturing and accelerated sinking of the oceanic-crust-capped lithospheric slab, as well as the decrease in shear stress along the base of the buoyant, ductile sialic material as it warms within the upper mantle. Lithospheric slab breakoff has been documented seismically by Osada and Abe (1981), Milsom and Audley-Charles (1986) and Widiyantoro and van der Hilst (1996). Loss of the dense, leading, oceanic portion of the Australian plate may be partly responsible for a shallowing of the angle of northward subduction, aiding in the exhumation of continental crust. Most importantly, ductile, relatively low-density quartzofeldspathic material is sandwiched between dense, relatively rigid mantle peridotite making up both hangingwall and footwall, hence ascent of an approximately 5 km thick slice of buoyant sialic material has

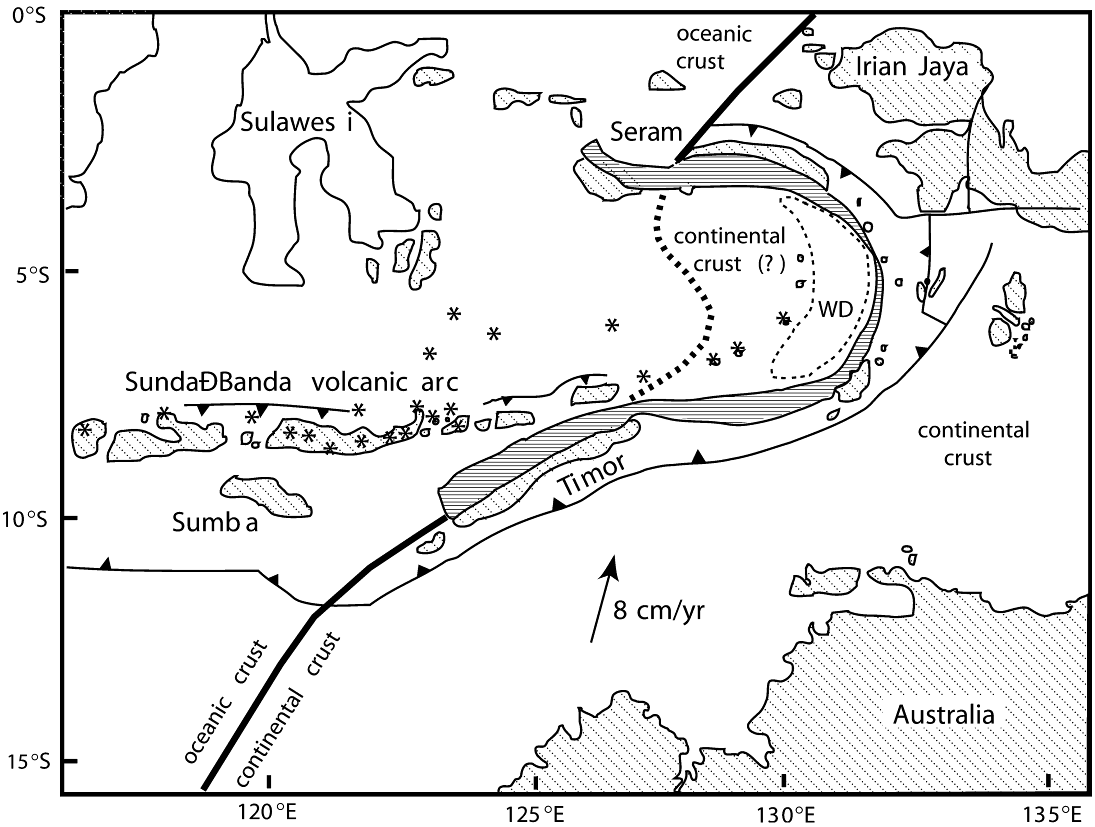

Fig. 1 Modern example of continental collision and exhumation of a transitional Alpine–Pacific type blueschist belt, after Osada and Abe (1981), Charlton (1991), and Maruyama *et al.* (1996). The convergent plate-tectonic junction (barbs on stable, nonsubducted plate) and exhumed HP blueschist belt (shaded pattern) are indicated. Australian continental crust is currently being subducted beneath the Timor–Seram segment of the Indonesian suture zone. The most active part of the Sunda–Banda arc lies above descending oceanic, not the continental, lithosphere of the Australian lithospheric plate.

taken place, confined between the tectonic boundaries of the active subduction channel (Cloos, 1993).

Regarding the generation of calcalkaline melts, the role played by subducting, amphibole-rich oceanic crust in contrast to micaceous continental crust along this convergent plate junction is apparent. As evident in Fig. 1, the Banda inner volcanic arc is extremely active west of Timor where oceanic lithosphere is descending beneath Indonesia; volcanism dies out eastward, and the strongly curved part of the archipelago is typified by volcanic quiescence in the vicinity of the Weber Deep, where Australian sialic crust-capped lithosphere is underthrusting the Eurasian plate.

Modelling the subduction–exhumation process

The dynamothermal evolution of convergent plate junctions has been investigated by many workers, employing chiefly numerical simulations (Shreve and Cloos, 1986; Cloos and Shreve, 1988a,b; Stüwe and Barr, 1998) and also including experimental studies (Tapponnier *et al.*, 1982; Chemenda *et al.*, 1995, 1996). The various models are a complex function of several controlling parameters, some not well constrained. Among them are the physical architectures of the converging lithospheric plates (temperature zonation, dimensions, viscosities, densities), rates of convergence, states and amount of sedimentary loading, extent of frictional dissipation, fluid expulsion and flow (channelled or pervasive), advective heat transport, kinetics of subsolidus reactions and of partial melting, and extents of contemporaneous erosion (subcrustal and surficial). Nevertheless, a broad quantitative understanding has emerged from these studies. Computed temperature–depth structures of convergent plate regimes have been studied extensively (e.g. Oxburgh and Turcotte, 1971: Anderson *et al.*, 1978; Thompson and Ridley, 1987; van den Beukel and Wortel, 1988; Peacock, 1990, 1992; Hacker *et al.*, 2003a). Although differing in details, reflecting contrasting assumptions and data constraints employed, all workers have demonstrated that a coherent descending lithospheric slab remains cool at upper mantle depths compared with its surroundings, whereas the landward volcanic–plutonic arc occupies a broad, shallow thermal high in the non-subducted crust and subjacent mantle wedge.

Scale modelling and numerical simulations of subduction-zone environments have generated a plethora of possible mass-transport scenarios, depending on different values for the various parameters noted above. Contemporaneous with ongoing subduction, return of subducted packets from great depth to crustal levels also is sensitive to the input variables, and reflects the nature of the lithotectonic units involved. The behaviour of a slab of low-density sialic material resting on a sinking lithospheric plate has been experimentally modelled in the laboratory by Chemenda *et al.* (1995, 1996) employing proportional scaling factors combined with a three-density-layer convergent system underflow of the structurally coherent, continent-like slab progresses until buoyancy forces exceed the frictional resistance to decoupling from the downgoing plate; at this stage, the modelled low-density material, either a graywacke trench complex, or an old microcontinental fragment, begins to return surfaceward along the subduction channel serving as a stress guide. Although somewhat simplified, the model successfully duplicates the general imbricate structural architecture of the Himalayas (Fig. 2) and many other subduction zones. For Circumpacific accretionary wedges and collisional mountain belts, both calculations and dynamic laboratory experiments realistically simulate the observed thickening of the sialic crust. Computations also have reproduced the backfolding (antithetic nappes) well known in the Western Alps and the Himalayas. Such finite-element modelling produces familiar convergent plate junction architectures, as illustrated in Fig. 3 (Beaumont *et al.*, 1996, 1999). Differing final developments reflect variable extents of erosion as well as numerical input parameters.

The episodic, erratic return of subducted quartz ± graphite-bearing HP blueschist–eclogite complexes and/or coesite ± diamond-bearing eclogitic UHP slabs, therefore, must be a sensitive function of the inherent physical properties of the subducted material, the configuration of the plate junction, the rate of closure, the degree of frictional dissipation along the upper and lower fault boundaries of the structural slice, and the magnitude of erosion-aided exhumation as a result of tectonism and/or buoyancy. Although most workers agree on the subduction path, exhumation mechanisms are still debated.

Nature of exhumed subduction complexes

Most resurrected subducted terranes appear to possess relatively tabular to sheetlike aspect ratios (Ernst *et al.*, 1997). The episodic return to relatively shallow crustal levels occurs in response to several processes: tectonic contraction or wedge extrusion (Maruyama *et al.*, 1994, 1996); buoyancy propulsion (Ernst, 1970, 1988; England and Holland, 1979; Hacker, 1996); and extensional-erosional collapse (Platt, 1986, 1987, 1993). As some relatively old, descending oceanic plates and their marginal deep-sea trenches seem to be retreating oceanward (rollback) more rapidly than the encroaching stable, non-subducted lithosphere (Molnar and Atwater, 1978; Seno, 1985; Busby-Spera *et al.*, 1990; Hamilton, 1995), compression of subducted continental flakes in the jaws of a convergent plate junction cannot account for ascent of the associated exhumed subduction complexes. Moreover, although extensional collapse and erosion help uncover deeply buried terranes, this phenomenon is incapable of explaining marked pressure discontinuities observed across boundaries between subducted and non-subducted lithological units (Ernst, 1970; Ernst *et al.*, 1970; Suppe, 1972). A body force such as buoyancy, however, provides sufficient motive force for the exhumation of low-density sialic sections carried to profound depths. Both numerical simulations and laboratory scale models demonstrate the efficacy of this mechanism. Slab breakoff (Sacks and Secor, 1990; von Blanckenburg and Davies, 1995) may slow the descent of the continental plate, thereby aiding the initiation of decoupling of buoyant sialic material, and its subsequent rise.

Virtually all exhumed HP–UHP metamorphic complexes appear to consist of relatively thin slabs or sheets sandwiched between rock units of much lower grade. An example from the western syntaxis of the Himalayas (O'Brien *et al.*, 2001) published by Kaneko *et al.* (2003) is illustrated in Fig. 4. Terry *et al.* (2000a,b) provided another illuminating cross-section through a UHP metamorphosed composite nappe in the Western Gneiss Region of coastal Norway. Similar cross-sections of imbricate low-P/high-T, HP and UHP slices in the Kokchetav massif of northern Kazakhstan were presented by Katayama *et al.* (2000) and Maruyama and Parkinson (2000). Clearly, the juxtaposition of higher and lower pressure lithotectonic units must have taken place subsequent to the UHP stage, and involved ductile attenuation and differential tectonic transport of the deeply buried rocks.

Based on combined thermobarometric and radiometric data, Kaneko *et al.* (2003) demonstrated that, in the western Himalayas, rapid exhumation to midcrustal levels was characterized by decompression under relatively low-temperature conditions (Fig. 5). This speedy ascent seems to be characteristic of preserved HP and UHP terranes globally (Ernst *et al.*, 1995; Gebauer, 1996; Gebauer *et al.*, 1997;

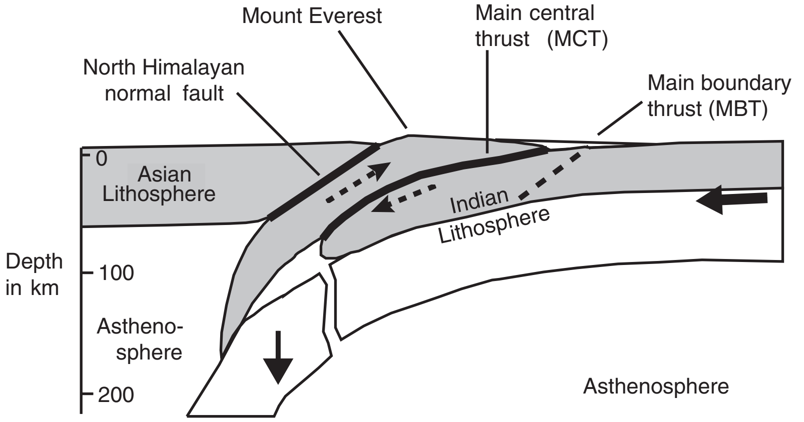

Fig. 2 Structural evolution of the Himalayan orogen, modified after scale-model experiments by Chemenda *et al.* (1995). The mantle lithosphere beneath the Indian subcontinent is unshaded.

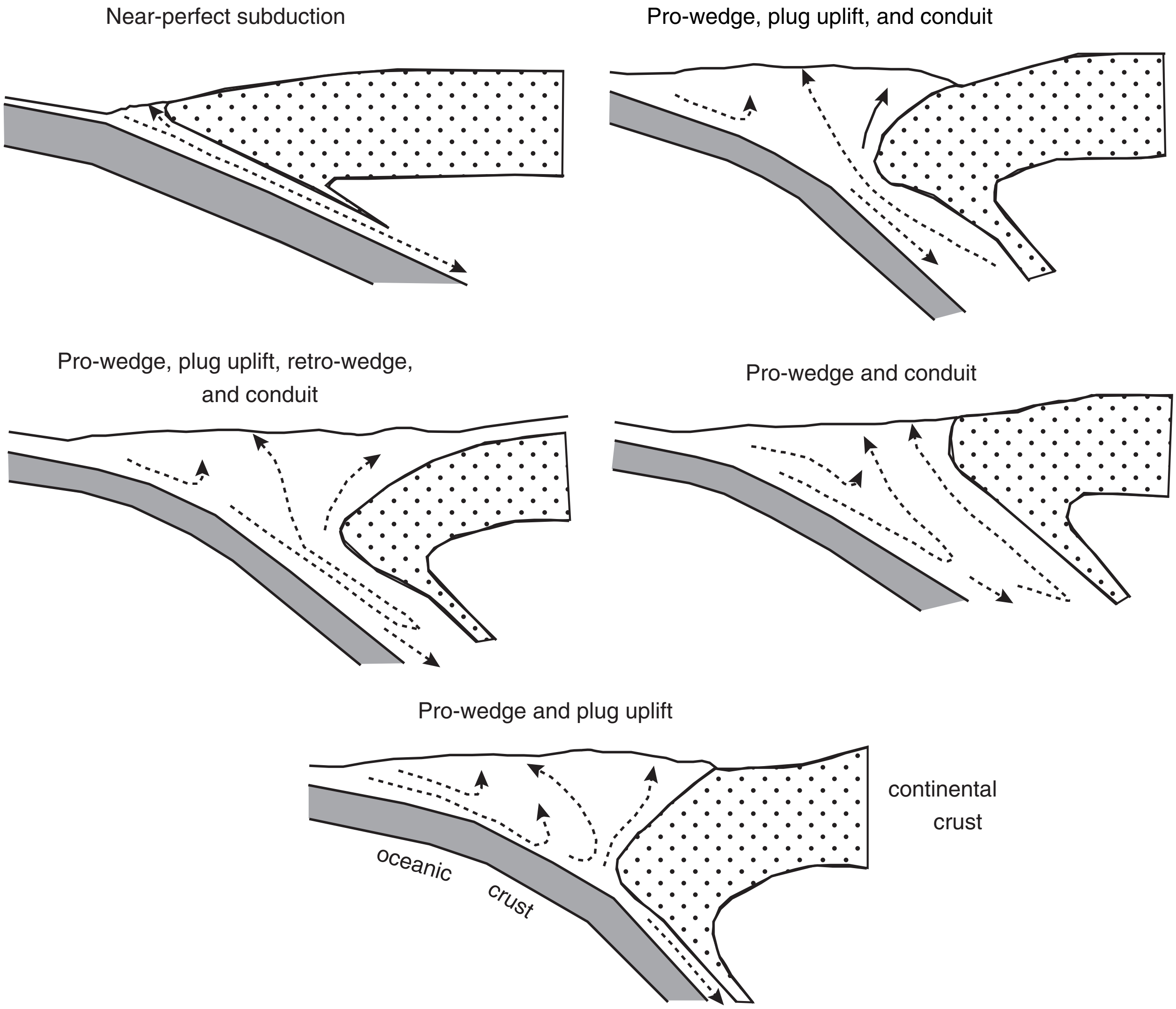

Fig. 3 Generalized, smoothed diagram of computer simulations of convergent lithospheric plate junctions involving a wide range of different input parameters, modified after Beaumont *et al*. (1996, 1999. Crustal units and simplified flow paths in the subduction complex approximated by present author.

Hacker *et al*., 2000, 2003b; Rubatto and Hermann, 2001).

Exhumation of over-thickened sialic crust as domical uplifts is occurring in some young collisional belts where curvilinear convergent plate-tectonic boundaries intersect at large angles, producing excess crustal mass. Such uplifts have been described as tectonic aneurysms (Zeitler *et al*., 2001). Certain post-Paleozoic arc cusps are characterized by the domical exposure of HP–UHP terranes (Maruyama *et al*., 1996), indicating that buoyant rise of unusually deeply buried, thick sections of low-density, ductile crust has taken place. UHP minerals and/or phase assemblages have been reported from some of these locales, including: the western Himalayan syntaxis (O'Brien *et al*., 2001; Kaneko *et al*., 2003); the Western Alps (Chopin, 1984; Compagnoni *et al*., 1995); and the Dabie–Sulu belt (Liou *et al*., 1996; Zhang *et al*., 2003). All these areas also contain imbricate nappe structures, so it is conceivable that exhumed HP–HUP terranes may reflect the operation of both earlier low-angle thrusting and later domical uplift.

Associated igneous activity

Massive calcalkaline arcs built on the landward non-subducted plate characterize the post-Paleozoic Circumpacific realm, attesting to thecon-sumption of thousands of kilometres of oceanic crust-capped lithosphere. Accumulating geochronological data from the Ryoke plus Sanbagawa belts in SW Japan, and the Sierra Nevada batholith plus Franciscan terrane of California show that the magmatic arcs are approximately coeval with outboard trench deposits, but that calcalkaline volcanism–plutonism began before and continued throughout growth of the outboard subduction complex (Stern *et al*., 1981; Brown, 1998, 2002; Barth *et al*., 2003; Joesten *et al*., 2004). Igneous bulk-rock isotopic signatures show that, although partially fused contin-

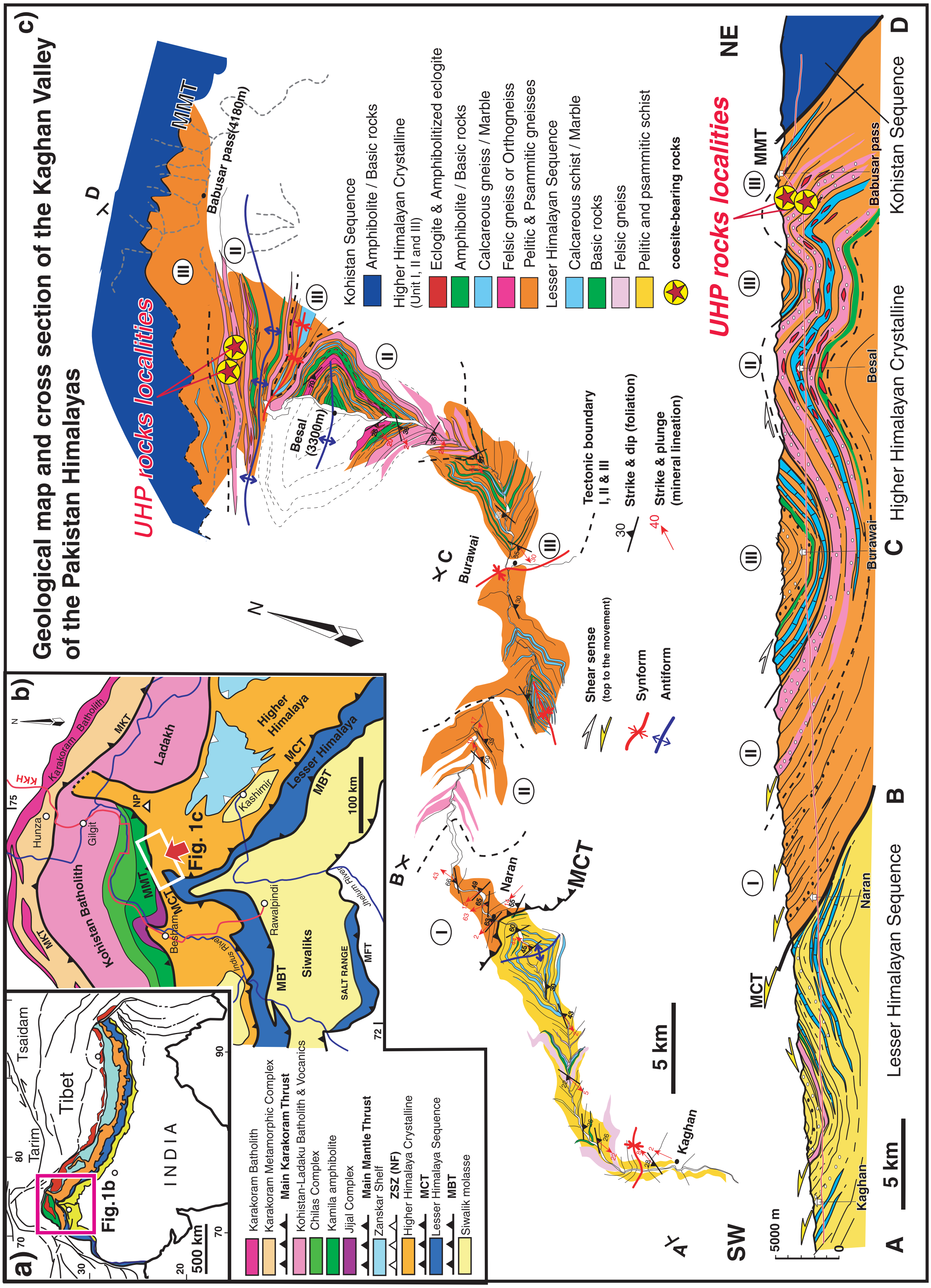

Fig. 4 General geological map and cross-section through the Kaghan Valley, western Himalayan syntaxis, Pakistan, from Kaneko *et al.* (2003). Index maps are shown in (a) and (b). A geological map and cross-section are presented as (c). Note that ultrahigh-pressure sheets are approximately 1 km thick.

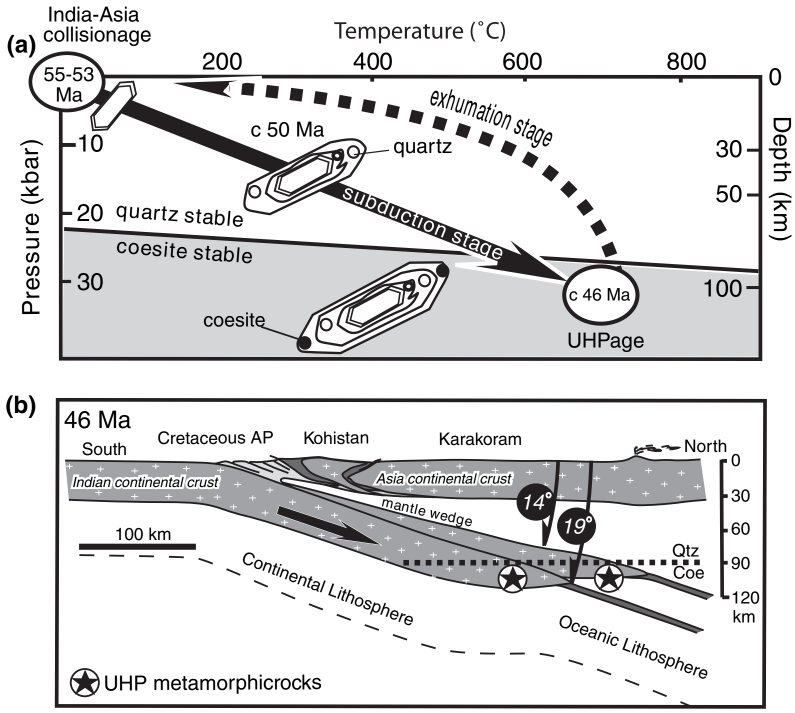

Fig. 5 Pressure–temperature time path for subduction and exhumation of Kaghan Valley ultrahigh-pressure rocks to midcrustal levels, from Kaneko *et al.* (2003). The *P–T* decompression path is shown in (a). The timing of underflow and unloading are illustrated in (b). Map and cross-section are shown as Fig. 4.

ental crust contributes to mature island-arc and continental-margin secondary magmas, the calcalkaline suite also contains an important primary contribution from the mantle wedge and/or subducted oceanic crust (DePaolo, 1981; Drummond and Defant, 1990; Hawkesworth *et al.*, 1993; Clift *et al.*, 2001; Hickey-Vargas *et al.*, 2002). These intermediate volcanics and plutonics are metaluminous (I-type) and some represent new additions to the continental crust; other igneous rock types are chiefly reworked crustal materials (S-type). Outboard, great deposits of first-cycle, volcaniclastic sediments make up fringing aprons of accretionary material (Dickinson, 1971, 1972, 1976; Kusky *et al.*, 1997); these active-margin sediments fill the forearc basin, and spill over into the trench and seaward, converging oceanic plate. Associated igneous rocks in the subduction complex and accretionary wedge are mostly ophiolites and other representatives of the oceanic lithosphere (Coleman, 1977). In particular, the underflow of oceanic fracture zones and spreading centres allows the tectonic insertion of mafic–ultramafic complexes into the accretionary prism (Cloos, 1993; Bradley *et al.*, 2003; Kusky *et al.*, 2003).

Calcalkaline plutons and related volcanics are scarce in Alpine-type collisional complexes. Unlike Circumpacific convergent plate margins, the igneous rocks are not set back on the stable, non-subducted plate; instead, peraluminous (S-type) granitoids occur within the root zone of the imbricate nappe edifice itself. Variable amounts of ophiolitic rocks occur in collisional orogens, but in some such mountain belts, fragments of oceanic lithosphere are conspicuously missing (e.g. Liou *et al.*, 1996). Lack of ophiolites and calcalkaline arc rocks in a collisional complex may reflect the consumption of only a small ocean basin. In addition, passive-margin sedimentary strata previously deposited on the leading edge of the downgoing continental crust are juxtaposed against the active-margin clastic apron derived from the stable, non-subducted plate.

Associated metamorphic activity

Essentially coeval paired metamorphic belts characterize the sustained underflow of Circumpacific subduction zones. However, ages of the landward high-*T*, low-*P* metamorphism range from older than to as young as recrystallization ages in the subduction-zone HP belt (Ernst, 1992; Maruyama *et al.*, 1996; Brown, 1998; Barth *et al.*, 2004; Joesten *et al.*, 2004). On the non-subducted lithosphere, calcalkaline plutons are emplaced into a consanguineous volcanic cover series and underlying, pre-existing basement; both lithic sequences have recrystallized at relatively high temperatures and low pressures. Recrystallization may be regarded as a series of superimposed contact metamorphic aureoles surrounding individual plutons (Barton *et al.*, 1988). Hornfelsic recrystallization and upright, cylindrical folding typify such inboard metamorphic terranes (Ernst, 1992). The seaward subduction complex consists of units recrystallized at relatively low *T* and high *P*. Reflecting its dynamic plate-tectonic setting within the subduction channel, penetrative deformation and foliation–lineation are typical of the higher grades of metamorphism. Décollements dip beneath the mantle wedge, and folds verge oceanward, reflecting the shear couple defined by a convergent plate junction.

Alpine subduction complexes include old continental massifs, as well as penetratively deformed, superjacent autochthonous and allochthonous sialic sections. Metamorphic belts of the Alpine type are not paired (Frey *et al.*, 1974). Relatively high-*P* subsolidus recrystallization is the rule, with parageneses comparable with those of outboard Circumpacific subduction complexes; however, some collisional orogens contain microcontinental slices and sheets that retain mineralogical relics attesting to much higher pressures and somewhat higher temperatures than characterize Pacific-type subduction complexes (Chopin, 1984; Smith, 1984; Sobolev and Shatsky, 1990; Coleman and Wang, 1995b; Liou *et al.*, 1998; Liu *et al.*, 2004).

Role of aqueous fluids in subduction-zone metamorphism and partial melting

General Statement

Consumption of an oceanic crust-capped plate allows the subduction

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of entrained sedimentary material and underlying basaltic crust, either offloaded into the accretionary prism, as is most of the low-density sedimentary debris, or returned to the deep upper mantle, as is most of the relatively high-density oceanic crust. Where microcontinental entities are part of the downgoing lithosphere, conduction to the site of subduction leads ultimately to collisional suturing; in such cases, continental crust may be carried beneath the stable, non-subducted plate to depths exceeding 100 km. Regardless of whether the descending materials are unconsolidated sediments or old sialic crust, pressure within the slab increases proportionately with depth, whereas temperature only rises slowly. Volatiles, principally $H_2O \pm CO_2$, are added to relatively dry portions of the downgoing plate, and are driven off from the more volatile-rich parts (Schreyer *et al.*, 1987; Peacock, 1995; Bebout, 1996). Shallow-level, pre-eclogitic stages of these reactions have been treated quantitatively by Frey (1987) and Ernst (1990). Attending the deep subduction of crustal lithologies, retained H_2O is sequestered exclusively in relatively refractory silicate phases (Poli and Schmidt, 1997; Okamoto and Maruyama, 1999). Eventually some of the low-density, subducted quartzofeldspathic material, recrystallized under HP or UHP conditions, may disengage from the descending slab and rise to intermediate levels of the continental crust. Generally, the weakly consolidated trench mélange characteristic of many Pacific-type junctions decouples at shallow depths (20–50 km), whereas old continental crust-capped slabs may continue sinking to considerably greater depths (up to at least 90–140 km). During exhumation, volatiles gain access to the rising quartzofeldspathic mass, or diffuse away from it, depending on the composition, average grain size, permeability, and extent of shearing of the various constituent units. Some evolved fluids rise into and through the overlying mantle wedge, whereas the rest migrates back up the conduit provided by the subduction channel (Cloos, 1984; Vrolijk *et al.*, 1988).

Below, I briefly summarize experimental phase equilibrium constraints on H_2O-bearing minerals stable in subducting crustal sections (e.g. Massone, 1995). Laboratory studies demonstrate that reaction rates are strongly influenced by the presence or absence of an aqueous fluid (Rubie, 1986, 1990). Beyond relatively shallow depths where compaction and much of the devolatilization have been completed (approximately 20 km), aqueous fluid is probably not present as a separate phase, and free H_2O becomes available only to the extent that OH-bearing minerals undergo dehydration. Three principal lithological assemblages constitute the spatially associated volatile-bearing subducted, recrystallized non-carbonate crust: (i) metabasalts and metagabbros; (ii) serpentinized peridotite; and (iii), metagraywackes ± pelitic schists, high-grade gneisses and metagranitoids.

Metabasalts and metagabbros

Glaucophane-crossite, barroisite and hornblende are major constituents, and are associated with epidote and/or lawsonite of lesser abundance in lower grade eclogite precursor assemblages. Although common as low-*T*, low-*P* prograde and retrograde minerals, most other hydrous phases such as actinolite, micas, chlorite and chloritoid are only minor UHP phases in metabasalts of normal composition. Natural parageneses have been documented worldwide through a generation of petrological studies (e.g. Banno, 1964; Ernst *et al.*, 1970; Cortesogno *et al.*, 1977; Dal Piaz and Ernst, 1978; Sobolev and Sobolev, 1980; Hirajima *et al.*, 1993; Beane *et al.*, 1995; Chopin and Sobolev, 1995; Dobretsov *et al.*, 1995; Krogh and Carswell, 1995; Liou *et al.*, 1996; Zhang *et al.*, 1997).

Phase-equilibrium studies have been performed on metabasaltic bulk compositions over the *P–T* range 300–1050 °C and 0–3.5 GPa (Essene *et al.*, 1970; Helz, 1973, 1979; Liou *et al.*, 1974; Spear, 1981; Gilbert *et al.*, 1982; Apted and Liou, 1983; Moody *et al.*, 1983; Poli, 1993; Liu *et al.*, 1996; Poli and Schmidt, 1997; Okamoto and Maruyama, 1999). As shown by long-term experiments and reaction reversals for fresh, chemically pristine MORB, barroisitic to pargasitic amphibole is the major hydrous mineral at and above approximately 650 °C. The Ca-amphibole stability field contracts dramatically at HP, and disappears at approximately 1.6–2.4 GPa, depending on the temperature. Phase relations for MORB + H_2O to 3.5 GPa are illustrated in Fig. 6 (Liu *et al.*, 1996); this diagram is applicable for relatively unmetasomatized oceanic basalts and gabbros. Substantial amounts of epidote ± lawsonite and sodic amphibole characterize such rocks at moderate pressures and at temperatures below 500–600 °C (Maresch, 1977; Maresch and Gerya, in press), but the glaucophane stability field is limited by elevated pressure, analogous to the *P–T* behaviour of Ca amphiboles (Koons, 1982; Carmen and Gilbert, 1983; Welch and Graham, 1992). Above moderate *T* and *P*, partial melting of eclogite to produce andesitic–dacitic liquids proceeds to the extent that free H_2O is available.

Lawsonite evidently remains stable in rocks of basaltic composition under UHP conditions (Okamoto and Maruyama, 1999), but only at extremely low temperatures defined by a geothermal gradient of about 2–5 °C km^{-1}. Phase relations for MORB + H_2O to 12.0 GPa are shown in Fig. 7. In synthesis experiments involving more aluminous mafic igneous rocks, epidote and chlorite possess broad *P–T* ranges at temperatures below approximately 600–650 °C (Pawley and Holloway, 1993; Schmidt and Poli, 1994; Poli and Schmidt, 1997) where they apparently coexist with actinolite at low pressures, and blue amphibole at intermediate pressures; these phases markedly decrease in amount with elevated pressures. Where metabasalts are more potassic than MORB, white mica persists to relatively elevated pressures, and may be a minor associate of garnet, omphacite and rutile in the HP–UHP paragenesis of altered mafic rocks. Chloritoid has been reported in some synthesis experiments, but this mineral probably does not represent an important phase in subducted mafic rocks because of its unusual Al_2O_3 + FeO-rich composition and restricted thermal stability.

Ca and Na amphiboles break down at pressures approaching UHP values, so metabasaltic rocks generate modest amounts of H_2O by dehydration, conducive to the production of eclogitic assemblages; accordingly, the prograde

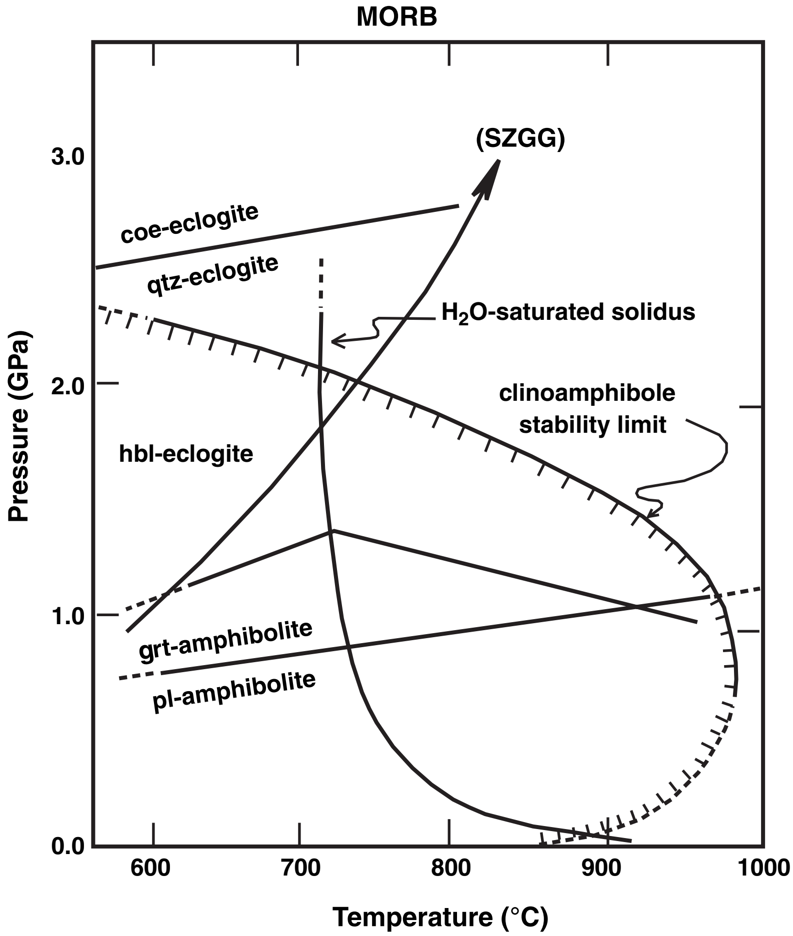

Fig. 6 Petrogenetic grid for the amphibolite-to-eclogite transformation in the basalt–H_2O system with oxygen fugacity defined by the $FeSiO_4$–Fe_3O_4–SiO_2 buffer, experimentally determined by Liu *et al.* (1996). Run lengths lasted up to 1630 hours at low temperatures; reaction reversals demonstrate a close approximation to chemical equilibrium. coe, coesite; grt, garnet; hbl, hornblende; pl, plagioclase; qtz, quartz. The illustrated overall P–T trajectory for a prograde subduction-zone geothermal gradient (SZGG) is approximately 8 °C km^{-1}. Note that progressive subduction-zone metamorphism involves devolatilization of garnet amphibolite and hornblende eclogite at physical conditions approaching those of the quartz–coesite *P–T* transition boundary.

reaction is thought to be favoured except in dry, impermeable protoliths such as granulite-facies metagabbros (Austrheim, 1990; Hirajima *et al.*, 1993; Zhang and Liou, 1997).

Metaserpentinites

At temperatures above the lowest metamorphic grades, antigoritic serpentinites dehydrate to orthopyroxene or talc + olivine, liberating large amounts of H_2O at approximately 550–600 °C and 2.0–3.0 GPa, depending on the geothermal gradient (Johannes, 1975; Evans *et al.*, 1976; Ulmer and Trommsdorff, 1995; Wunder and Schreyer, 1997; Pawley, 1998; Evans, 2004). Similar to the behaviour of clinoamphiboles, antigorite is pressure limited in its stability range. Equilibrium phase relations for the antigorite bulk composition to 7.0 GPa are presented in Fig. 8, after Wunder and Schreyer (1997).

But to what extent is the dunite–harzburgite–lherzolite depleted mantle lithosphere that underlies the oceanic crust hydrated? Most recovered specimens of suboceanic mantle are partly serpentinized. However, sampling has chiefly involved oceanic fracture zones, transform faults and on-land ophiolites, sections characterized by unusually intense deformation and fluid circulation (e.g. Bonatti, 1976). Fully hydrated mantle would have an aggregate density < 2.5, hence should be virtually unsubductable. Thus, only 10–20% serpentinized peridotite seemingly could descend along a convergent plate junction. This conclusion is supported by seismic transmission velocities of 8.1 km s^{-1} measured for the oceanic mantle, in contrast to a Vp of 5.5 km s^{-1} for pure serpentinite (Coleman, 1971; Christensen and Salisbury, 1975). Nevertheless, modest amounts of hydrated ultramafics may be present in the uppermost part of the oceanic lithosphere, so subduction and devolatilization of hydrated peridotite probably contributes to the aqueous fluid flux of a descending plate (Ulmer and Trommsdorff, 1995).

Metapelites, metagraywackes, gneisses and metagranitoids

Laboratory studies of quartzose feldspathic lithologies over a wide range of experimental conditions demonstrate that chlorite dehydrates above approximately 550–600 °C, whereas potassic white micas ± biotites remain stable to considerably higher temperatures (Le Breton and Thompson, 1988; Vielzeuf and Holloway, 1988; Vielzeuf and Montel, 1994; Gardien *et al.*, 1995; Nichols *et al.*, 1996; Patiño Douce and Beard, 1996; Skjerlie and Johnston, 1996; Luth, 1997; Massone and Szpurka, 1997; Patiño Douce and McCarthy, 1998). A simplified petrogenetic grid showing phase relations for metagraywackes and pelitic schists to 2.0 GPa, generalized after Vielzeuf and Holloway (1988), is illustrated in Fig. 9. In aluminous metapelitic rocks, the assemblage biotite + kyanite is replaced by muscovite + garnet at pressures above approximately 1.7 GPa, but in more quartzofeldspathic compositions, biotite does not break down at elevated pressures. For typical subduction-zone geothermal gradients (~8 °C km^{-1}), muscovite–phengite solid solutions remain stable in peraluminous lithologies to at least 3.5–4.0 GPa (e.g. Massone and Szpurka, 1997); for metaluminous rocks, biotite evidently persists to similar pressures. As white mica and biotite constitute

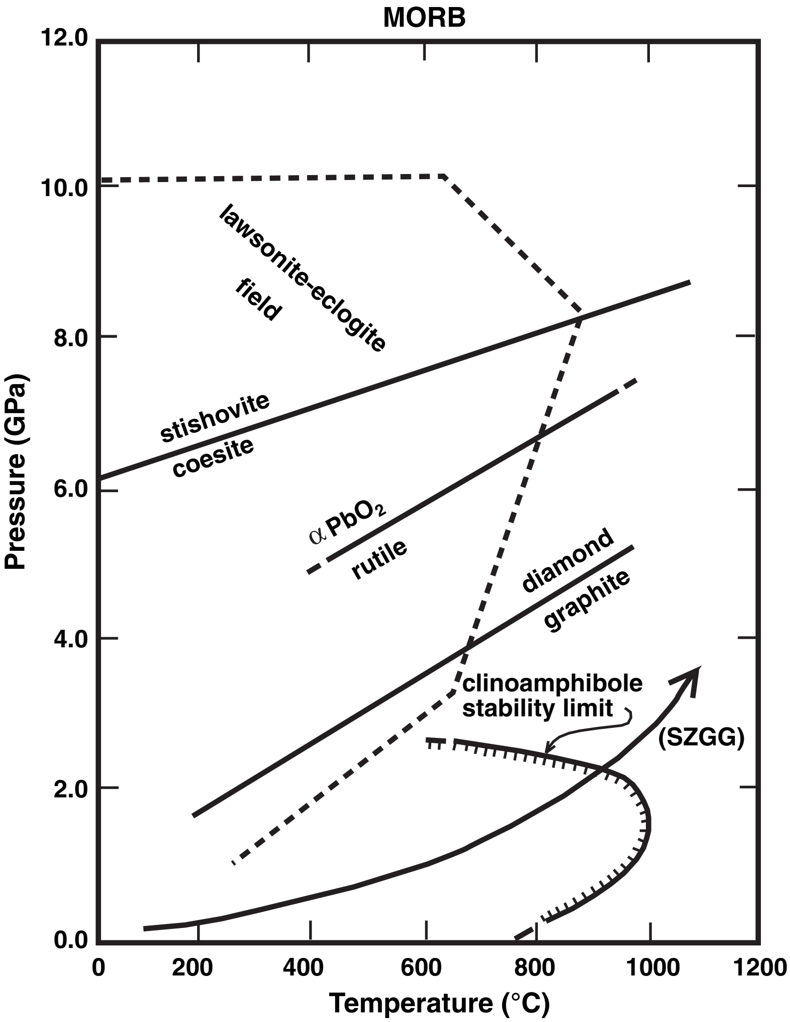

Fig. 7 Synthesis fields and postulated stability relations of lawsonite in the basalt–H_2O system, after Okamoto and Maruyama (1999). The rutile–srilankite (αPbO_2) transition is indicated after Akaogi *et al.* (1992). White mica would join this assemblage for K_2O + Al_2O_3–rich bulk–rock compositions. The illustrated *P–T* trajectory for a prograde subduction-zone geothermal gradient (SZGG) is approximately 8 °C km^{-1}.

the volumetrically most important hydroxyl-bearing phases in recrystallized shaley and quartzofeldspathic phase assemblages, units that dominate the continental crust may fail to evolve significant amounts of H_2O during UHP metamorphism at depths of 100 km or more (e.g. Tilton *et al.*, 1997); thus, in the absence of a catalytic aqueous fluid, such lithologies might retain metastable feldspar and quartz in micaceous rocks, and fail to develop the stable prograde UHP mineralogical assemblage. Supporting this hypothesis of limited prograde recrystallization, mineral inclusions in zircons separated from UHP gneisses exhibit a regular relationship involving relict quartz and other lower *P* phases in the cores, coesite inclusions in the zircon mantles, and retrograde minerals in the rims (e.g. Katayama *et al.*, 2000; Liu *et al.*, 2004).

Petrological inferences regarding the evolution of H_2O

Subducting low-grade, hydrous metamorphic rocks lose most of their volatiles at shallow depths, and by the stage in which they reach HP–UHP conditions, mica-retaining quartzofeldspathic and pelitic lithologies are relatively deficient in H_2O-producing phases. Thus, clinoamphibole-bearing oceanic crust ± serpentinized peridotite (not micaceous sialic crust) probably represents the most significant dehydrating rock type at depths approaching and exceeding 100 km. Devolatilization of clinoamphiboles ± serpentine should begin at approximately 70–80 km, but if reactions are sluggish on the timescale of lithospheric plate descent in the cool subduction-zone environment, aqueous fluid evolution probably reaches a maximum at somewhat greater depths. Such pressure overstepping appears to be characteristic of many UHP terranes (Austrheim, 1998). The long-continued descent of thousands of kilometres of Pacific-type oceanic lithosphere probably provides the greatest proportion of deep-seated aqueous fluid rising through the subduction channel as well as diffusing into the mantle wedge. Closure of an oceanic basin prior to continental impaction provides lesser amounts of volatiles, depending on the amount of consumed oceanic lithosphere. In the presence of an H_2O-rich fluid, partial fusion of the eclogitic oceanic crust and/or the refertilized mantle wedge evidently results in generation of the I-type magmas of the calcalkaline suite (Drummond and Defant, 1990; Kushiro, 1990; Hawkesworth *et al.*, 1993; Morris, 1995).

In fact, the fluid flux evolved under HP–UHP conditions probably is curtailed or terminated by the arrival, profound underflow and suturing of a sizeable mass of sialic crust. Continental collision, therefore, would not be expected to generate a substantial calcalkaline arc, but might involve the partial melting of overthickened, accretionary quartzofeldspathic and pelitic crust, producing peraluminous, S-type granitoids. True continental growth thus appears to be the result of clinoamphibole ± serpentine dehydration within the descending oceanic lithosphere, followed by volatile-induced partial melting of oceanic crust and/or the overlying, metasomatized mantle wedge (Ernst, 1999).

Kinetics of hydration–dehydration reactions

Measurement of the reaction rates that consume or evolve H_2O seems

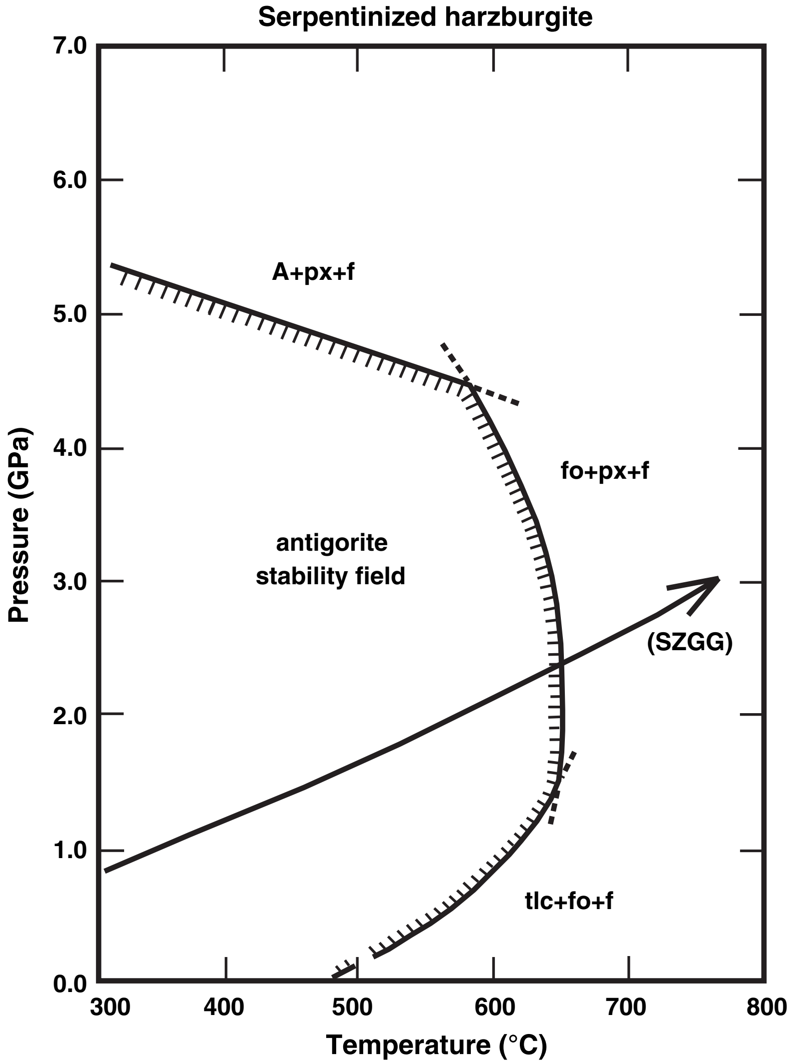

Fig. 8 Phase diagram for the stability of antigorite, after Wunder and Schreyer (1997). Reactions have been reversed, demonstrating equilibrium. A, phase A (hydrous magnesium silicate); f, aqueous fluid; fo, forsterite; px, clinoenstatite; tc, talc. See also: Johannes (1975); Evans *et al.* (1976); Ulmer and Trommsdorff (1995); Pawley (1998); and Evans (2004). The illustrated *P–T* trajectory for a prograde subduction-zone geothermal gradient (SZGG) is approximately 8 °C km^{-1}.

to be the key to understanding deep-seated phase changes in Alpine- and Pacific-type convergent plate junctions. Evidence of widespread disequilibrium during eclogite-facies metamorphism (e.g. Austrheim, 1990, 1998; Rubie, 1990; Hacker, 1996; Leech, 2001; Liu *et al.*, 2004) highlights the need for quantitative experimental data on reaction kinetics. From the study of HP and UHP complexes, the availability of fluid rather than *P–T* stability appears to control the rate at which reactions occur and the extent of mineralogical transformations. Although the catalytic effect of H_2O is well known and specific mechanisms have been proposed (e.g. Rubie, 1986), little is known about the dependence of reaction rates and mechanisms on fluid concentrations and chemistries in systems characterized by very low fluid–rock ratios attending the profound subduction of accretionary prisms or coherent continental crust.

It is generally assumed that dehydration either occurs rapidly or is controlled by the rate of enthalpy production (Ridley, 1985). Most laboratory experiments, employing powdered starting materials, yield rates that appear to be too rapid when extrapolated to geological conditions (e.g. Wegner and Ernst, 1983; Schramke *et al.*, 1987; Jové and Hacker, 1997; Mosenfelder and Bohlen, 1997). In contrast, experimental dehydration reactions employing rocks as starting materials yields slower rates and different mechanisms (Brearley, 1987; Rubie and Brearley, 1987; Brearley and Rubie, 1990; Hacker, 1990, 1991). Extrapolation of solid-rock results to longer time scales and to temperatures close to actual values is problematic at present. However, because dehydration reactions generate a rate-enhancing fluid, prograde transformations proceed far more rapidly than the equivalent retrograde reaction. This coupled with the fact that mafic–ultramafic lithologies carry substantial amounts of pressure-limited clinoamphibole and serpentine, whereas micaceous felsic rocks are not so pressure constrained, yields an important conclusion: Pacific-type plate junctions are characterized by the deep-seated evolution of aqueous fluids, whereas the deeper portions of Alpine-type collisional sutures in general remain dry. Moreover, it is quite probable that the H_2O given off from a rapidly sinking, dehydrating oceanic slab is released over a range of depths greater than that predicted by phase equilibria, perhaps 100, rather than 70–80 km. Of course, in cases where oceanic ridge subduction takes place, the elevated thermal regime in subducting, Pacific-type lithological assemblages would promote shallower levels of devolatilization (e.g. Sisson *et al.*, 2003).

The ease with which aqueous fluids gain access to, or egress from subducting and ascending low-density sialic masses is a complex function of rock compositions, relative permeabilities, extent of shearing, aggregate grain sizes, and availability of volatiles. These quantities reflect the past geological histories and dynamics of lithotectonic units moving along a convergent plate junction. Based on thermal modelling (Peacock, 1992, 1995; Ernst and Peacock, 1996) and laboratory phase-equilibrium studies (Vielzeuf and Holloway, 1988; Massone, 1995; Schreyer, 1995; Liu *et al.*,

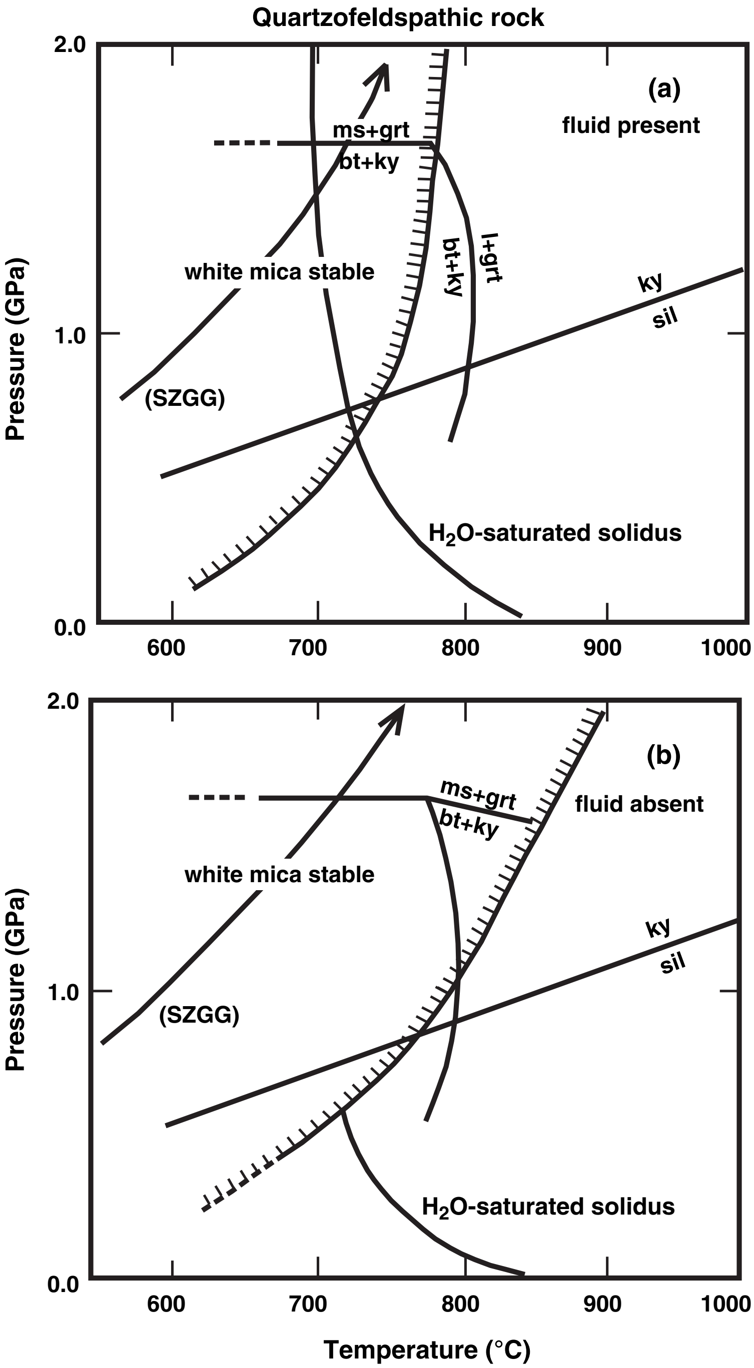

Fig. 9 Experimentally determined and computed petrogenetic grid under fluid-excess and fluid-absent conditions for quartzose metagraywackes + pelitic schists possessing intermediate bulk-rock Fe/Mg ratios, simplified after Vielzeuf and Holloway (1988). bio, biotite; grt, garnet; ky, kyanite; l, liquid; ms, white mica; sill, sillimanite. Phengitic micas are stable to pressures > 3.5–4.0 GPa in quartzofeldspathic rock types, as documented by: Le Breton and Thompson (1988), Vielzeuf and Montel (1994), Gardien *et al.* (1995), Patiño Douce and Beard (1996), Skjerlie and Johnston (1996) and Massone and Szpurka (1997). Biotite is stable as a solidus phase at least up to 3.5 GPa in ferruginous metapelites according to Nichols *et al.* (1996), and for chemically simplified metaluminous bulk compositions to approximately 9.0 GPa (Luth, 1997). The illustrated *P–T* trajectories for a prograde subduction-zone geothermal gradient (SZGG) are approximately 8 °C km^{-1}.

1996; Patiño Douce and McCarthy, 1998), what we understand semiquantitatively is the evolving *P–T* structure of a convergent plate junction, including subducting oceanic crust ± pelitic and quartzofeldspathic sedimentary load or colliding continental material, as well as stability fields of the mineral assemblages for the major rock types. Not well known are the rates at which mineralogical transformations take place, and therefore, the conditions and tectonic realms under which volatile constituents are evolved and consumed. Given locally anhydrous circumstances, low-*P* assemblages can be retained under prograde HP–UHP conditions (Austrheim, 1990), and HP–UHP relics can be preserved under retrograde low-*P* conditions (Hirajima *et al.*, 1993; Liou and Zhang, 1996; Ye *et al.*, 2000; Liu *et al.*, 2004). Quantification of the kinetics for the controlling reactions is the key, and these data are only available for a few, chemically simple mineral systems.

Tectonics of Alpine- and Pacific-type subduction zones: the thin-slab model

Previous works (e.g. Chopin, 1987; Ernst *et al.*, 1995; Rumble *et al.*, 2003) have demonstrated that the deep subduction of low-density quartzofeldspathic material accounts for the generation of Circumpacific and collisional HP and UHP metamorphic belts. A major petrotectonic problem consists of clarifying the manner in which these sialic terranes have been returned to shallow levels while preserving relics of the HP–UHP phase assemblages. The two-way movement of terranes along subduction zones was recognized long ago (Ernst, 1970; Suppe, 1972; Willett *et al.*, 1993). Two intergradational end-member convergent plate junction configurations (Bally, 1981), both of which may involve large components of arc-parallel slip, are recognized: (i) subduction of a Pacific-type greywacke–pelitic complex carried to depth on oceanic lithosphere; and (2), underflow and Alpine-type collisional suturing of a continent, microcontinent, or island arc beneath the stable, nonsubducted lithospheric plate. The former gives rise to the great accretionary prisms that mark the Pacific Rim (Dickinson, 1971, 1972, 1976). The latter apparently involves the underflow of salients or peninsulas of old continental crust, thoroughly imbedded in chiefly cool, oceanic-crust-capped plates (Yin and Nie, 1993; Ernst and Liou, 1995); they descend relatively rapidly, generating HP–UHP mineral assemblages (Peacock, 1995; Ernst *et al*., 1997). Sialic materials are carried down to great depths where buoyancy forces tending to drive them surfaceward balance dynamic forces tending to subduct them still further. In a contrasting type of attempted subduction of muddy mélange, where the low-density materials are incompetent and very poorly coupled to the descending lithosphere, upper plate indentation occurs instead (Ellis, 1996).

For Alpine and Pacific convergent plate junctions involving a coherent, competent greywacke-rich section or well-bonded continental crust + mantle, entrance of growing amounts of low-density material into the subduction zone enhances the braking effect because of increasing buoyancy; this in turn results in rupture and loss of the high-density lithosphere leading the downgoing plate at intermediate upper mantle depths, where the plate is in extension (Isacks *et al*., 1968). Slab breakoff (Sacks and Secor, 1990; von Blanckenburg and Davies, 1995) enhances the effective buoyancy and the sialic subduction complex, or a slice thereof, decouples from the descending plate and moves back up the subduction channel (van den Beukel, 1992; Davies and von Blanckenburg, 1998). Exhumation may be aided in part by a progressive shallowing of the ruptured and now buoyant, rebounding continental-crust-capped lithosphere, and perhaps more importantly, because of reduction of the shear force acting along its base as a result of its increasingly ductile behaviour as the slab gradually warms in the upper mantle (Stöckhert and Renner, 1998).

Because of continued subduction-induced refrigeration tectonically beneath the rising HP–UHP complex, and extensional faulting against the overlying, cooler non-subducted plate during ascent, relatively thin slices of such terranes effectively loose heat along both upper and lower surfaces by thermal conduction; thus, these complexes may nearly retrace the subduction-zone *P–T* trajectory during decompression (Rubie, 1984; Ernst, 1988; Ernst and Peacock, 1996). Schematic relationships are shown in Fig. 10, and apply equally well to the exhumation of HP and UHP terranes. A normal fault above and a reverse fault below bound the thin-aspect-ratio slab. Such shear senses seem required by structural relations, for example, in the Franciscan Complex (Bailey *et al*., 1964; Ernst, 1970; Jayko *et al*., 1987; Ring and Brandon, 1994), the Dora Maira Massif of the western Alps (Henry, 1990; Compagnoni *et al*., 1995; Michard *et al*., 1995), the Dabie Shan (Ernst and Liou, 1995; Hacker *et al*., 1995), the Kokchetav massif of northern Kazakhstan (Maruyama and Parkinson, 2000), and the western Himalayas (Searle *et al*., 2001; Kaneko *et al*., 2003). The decrease in overall density of the continental lithosphere subsequent to slab breakoff results in a shallowing of the angle of the conver-

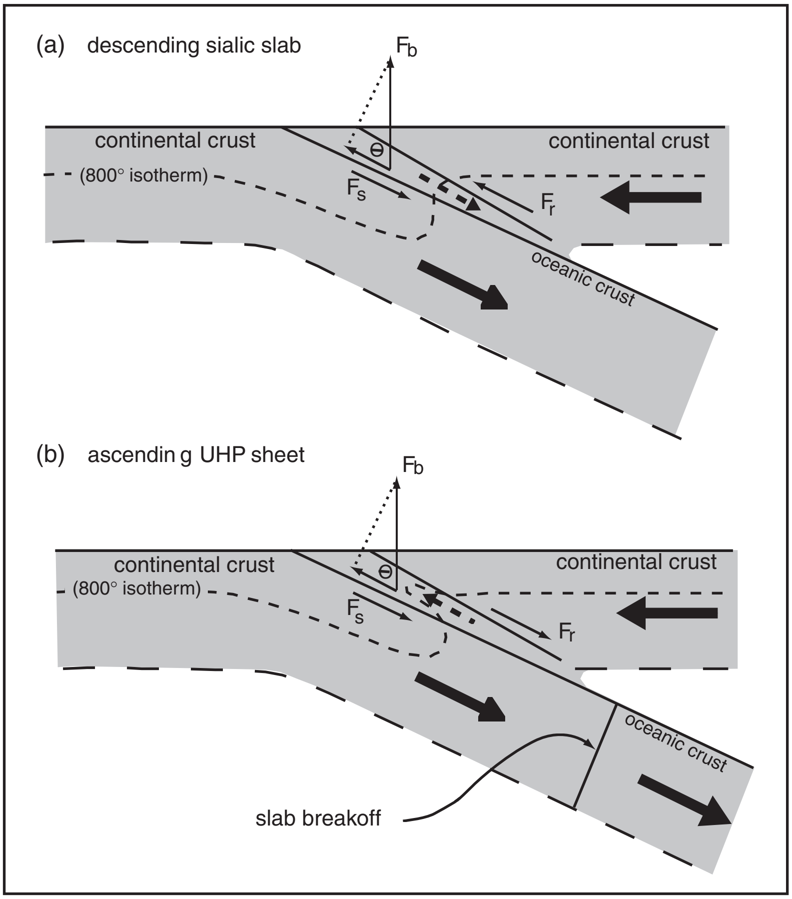

Fig. 10 Schematic convergent lithospheric plate-boundary diagram, modified after Ernst and Peacock (1996) for active subduction: (a) deep burial and thermal structure of a subducted prism of quartzofeldspathic strata, island arc, microcontinent, or continental salient; (b), later decompression cooling of a rising slice of low-density sialic material. Differential motion of plates and slices indicated by arrows (the subducting plate actually is sinking and rolling backwards; Hamilton, 1995). Accompanying ascent of the thin HP–UHP terrane (thickness exaggerated for clarity), cooling of the upper margin of the sheet takes place where it is juxtaposed against the shallower, lower temperature hangingwall plate (mantle wedge); cooling along the lower margin of the sheet takes place where it is juxtaposed against the lower temperature, subduction-refrigerated lithospheric plate. Tectonic exhumation of low-density slices requires erosive denudation and/or gravitational collapse and a sialic root at depth. The resolution of forces acting on the sialic slab in stages (a) and (b) are discussed in the text.

gent plate junction, and may be partly responsible for some of the late doming noted in many resurrected subduction complexes (e.g. Fig. 4). Another exhumation scenario involves the antithetic faulting typical of some compressional orogens, in which double vergence is produced during end stages of the collision and ascent of sialic crust (e.g. Dal Piaz *et al*., 1972; Beaumont *et al*., 1996).

Where thin HP–UHP slices are exhumed during continued subduction/refrigeration, the ascending quartzofeldspathic complex approximately follows the prograde metamorphic *P–T* path in reverse; this trajectory has been documented, for instance, in HP–UHP terranes of the western Alps and the California coast ranges (e.g. Ernst, 1988; Chopin *et al*., 1991; Coleman and Wang, 1995a). For thick, more nearly equidimensional ascending bodies (>10 km thick?), the ratio of cooling surface to mass is low, and central portions are likely to remain sufficiently hot during decompression for the complete obliteration of all evidence of deep-seated metamorphism, and perhaps even for partial melting to ensue; accordingly, such exhumed masses retain none of the precursor HP–UHP mineral assemblages, but may generate anatectic, peraluminous (S-type) granitoids. Reflecting this total recrystallization at lower pressures, any earlier UHP histories for such masses would never be suspected.

The one-atmosphere densities of unaltered oceanic crust, approximately 3.0, quartzofeldspathic continental material, approximately 2.7, and anhydrous mantle, approximately 3.2, increase with elevated pressure, reflecting the progressive transformation of framework silicates to layer, chain and orthosilicates. Typical HP–UHP mineralogical assemblages and computed (ambient) rock densities appropriate for burial depths of approximately 100 km are (Ernst *et al*., 1997): metabasaltic eclogite = 3.65; eclogitic quartzofeldspathic gneiss = 3.05 and garnet peridotite = 3.25. Completely transformed, K-feldspar + jadeite + coesite-bearing granitic gneiss remains less dense than garnet or spinel lherzolite, whereas metabasaltic eclogite is considerably more dense than the mantle. Accordingly, both shallowly and deeply subducted sialic crustal sections are buoyant relative to the surrounding mantle and tend to rise, whereas eclogitized oceanic crust becomes negatively buoyant and continues to sink (e.g. see Massone, in press). Of course, if the prograde conversion of quartzofeldspathic ± pelitic slices to dense mineral assemblages is incomplete because of lack of a catalytic aqueous fluid, sialic material should be even more buoyant than indicated above, whereas oceanic crust would be less negatively buoyant, depending on the extent of transformation to dense eclogitic phases.

Forces acting upon a sheet of subducted low-density material (Fig. 10) are as follows. Underflow of a basaltic crust-capped plate carries a passive, superjacent load of chiefly graywacke and pelitic sediment, or closure of an oceanic basin results in the ultimate entrance of a sialic terrane into the subduction zone. Descent of the low-density material occurs provided shear forces caused by underflow exceed the combined effects of buoyancy and frictional resistance along the hangingwall of the subduction channel. Decoupling of a slice (not necessarily the entire low-density section), followed by its ascent, occurs provided buoyancy is markedly positive and exceeds the combined effects of shearing along its base against the outboard, sinking lithosphere and resistance to movement along its upper surface against the inboard hangingwall. For the HP–UHP mineral assemblages to be partly preserved on exhumation, even in the absence of an aqueous fluid (which would promote retrograde reaction during decompression), the rising slice must be sufficiently thick to promote rapid, buoyancy-driven ascent, yet thin enough that heat is efficiently removed by thermal conduction across the sheet boundaries, i.e. the upper normal and lower reverse faults. A documented example of this type of structure is illustrated in Fig. 3.

The described process is similar to the slab-extrusion mechanism that Maruyama *et al*. (1994) advanced to account for the Dabie Shan UHP rocks of east-central China and the Kokchetav massif of northern Kazakhstan. It is also comparable with tectonic models advanced for the Himalayas and the Alps (Burg *et al*., 1984; Burchfiel and Royden, 1985; Merle and Guillier, 1989; Wheeler, 1991). For the present tectonic model, however, body-force propulsion of the UHP sheet of continental crust or accretionary complex back up the subduction zone, followed by extensional collapse and erosion (e.g. see Platt, 1986, 1987, 1993; England and Molnar, 1993; Chemenda *et al*., 1995, 1996; Beaumont *et al*., 1996, 1999), is postulated as caused by slab buoyancy, in contrast to a mechanism involving compressional extrusion.

Characteristics of Alpine- and Pacific-type orogenic belts

Petrotectonic models proposed to explain the origin, plate-tectonic setting, and *P–T* evolution of Circumpacific and collisional metamorphosed complexes must account for the following general observations regarding such terranes (Bailey *et al*., 1964; Banno, 1964; Frey *et al*., 1974; Liou *et al*., 1994; Coleman and Wang, 1995a; Dobretsov *et al*., 1995; Liou and Zhang, 1995; Maruyama, 1997; Rumble *et al*., 2003).

1 UHP complexes are developed within old, relatively cool sialic crust and are confined to present-day intracontinental orogenic belts, whereas HP terranes frame the Pacific Rim, and to a lesser extent, mark some intracratonic collisional sutures as well.

2 Low-density quartzofeldspathic and pelitic rocks constitute the most abundant lithologies for both HP and UHP metamorphic belts. Outboard Pacific-type orogens are characterized by metagraywackes and dark, metashaley mélanges, whereas these rocks and higher metamorphic grade sialic equivalents representing older protoliths, such as orthogneisses, paragneisses–paraschists, migmatites and metagranitoids, characterize some Alpine-type orogens, followed by lesser amounts of metabasaltic rocks and metacalcareous strata.

3 Mafic and ultramafic rocks are of minor importance in many, but not all collisional complexes, whereas tectonized, disaggregated ophiolites are ubiquitous in Circumpacific margin belts.

4 For UHP terranes, massive metamafic rocks and/or siliceous schists typically contain most of the surviving relict UHP silicate phases. Metacarbonates carry very rare UHP minerals; schistose metapelites and gneissose quartzofeldspathic lithologies almost completely lack such relics, although rare UHP micro-inclusions in garnet and/or zircon have been described. For HP and UHP belts, metabasaltic and metagabbroic lithologies tend to preserve minerals and phase assemblages indicating HPs more completely than most quartzofeldspathic and pelitic lithologies. Metacherts typically lack diagnostic HP phases, and if initially present, aragonite is poorly preserved in marbles.
5 Post-metamorphic products of erosion are present but do not necessarily occur in enormous quantities in either Alpine or outboard Pacific-type orogens.
6 Well-developed slightly older to coeval calcalkaline volcanic–plutonic terranes are ubiquitous, in all cases are immense, and are paired with a seaward trench complex in Pacific-type paired belts. Such coeval voluminous arcs generally are not associated with collisional terranes.
7 Although rocks of the volcanic–plutonic arc may preserve a semi-continuous record of igneous activity marking landward belts of the Pacific Rim, the exhumation of seaward HP and UHP terranes commonly is erratic and episodic in both Alpine and Pacific-type subduction complexes.
8 All deeply buried, then exhumed subduction-zone belts are characterized by nappes and low-angle décollements juxtaposing lithotectonic units of strongly contrasting metamorphic P–T conditions. Many of these imbricate stacks have been exposed by post-thrusting domical uplift combined with erosion.

Circumpacific margin convergence zones are characterized by widespread blueschist belts, tectonic mélanges and dismembered ophiolite complexes, reflecting the underflow of thousands of kilometres of ocean crust-capped lithosphere, an environment termed type-B subduction by Bally (1981); in contrast, Bally's type-A zones of continental collision involve the consumption of intervening ocean basins, in many cases of only modest size. Table 1 lists some aspects of the contrasting geological natures of Alpine and Pacific-type subduction zone assemblages. Metamorphic prograde and retrograde P–T paths followed by such terranes during subduction and later exhumation are topologically similar, but some intracratonic collisional complexes retain relics of UHP metamorphism (P_{max} = 2.6–4.0 GPa) whereas recovered outboard Circumpacific orogens typically consist of only HP belts (P_{max} = 0.6–1.5 GPa).

Discussion

The various similar and contrasting petrotectonic features of collisional and Pacific Rim orogenic complexes just described reflect their dynamic generation and exhumation histories. Genetic relationships are discussed in the order listed above.

1 Mechanical analyses demonstrate that sialic crust several km or more thick may be readily subducted if it enters a convergent plate junction (England and Holland, 1979; Molnar and Gray, 1979; Beaumont *et al*., 1996). A favourable geological environment for UHP metamorphism would involve insertion into the subduction zone of a narrow salient or promontory of continental crust as an integral part of an

Table 1 Generalized comparison of Phanerozoic subduction–zone tectonic end-member regimes, extensively modified after Maruyama *et al*. (1996)

	Alpine type (collisional)	Pacific type (Circumpacific)
Protoliths		
Shallow-marine sediments	Widespread platform carbonates	Reef (atoll) limestones
Clastic wedges	Multicycle siliciclastics, per-aluminous shales ± active-margin clastics	Voluminous first-cycle graywackes, muddy olistostromes
Deep-sea sediments	Uncommon	Widespread bedded cherts, Mn-nodules, deep-water carbonates
Igneous rocks	Uncommon bimodal basalts + rhyodacites	Abundant MORBs, seamounts (OIBs)
Basement	Granitic gneiss complexes	Ophiolite complexes
Age of rock formation	Ancient to young*	Exclusively young*
Tectonics	Oceanward vergence of nappes, late-stage backfolding, landward-dipping thrust faults	Oceanward vergence of nappes, landward-dipping thrust faults
Rapid exhumation	During final collisional suturing	Episodic, erratic
Petrology		
Typical maximum pressure	2.6–4.0 GPa (UHP)	0.6–1.5 GPa (HP)
Associated calcalkaline volcanic–plutonic belt	Rare or absent (S-type)	Outboard: absent Inboard: huge (mostly I-type)
Degree of retrogression	Nearly complete	Incomplete
Anhydrous mantle fragments	Garnet lherzolite	Spinel, plagioclase lherzolite, harzburgite
Serpentinite bodies	Generally uncommon	Outboard: ubiquitous Inboard: generally uncommon
Roughly coeval paired belts	Absent	Invariably present

*Young, generation of the lithologic section *c.* 0–200 Myr prior to time of underflow.

old, thermally relaxed, largely oceanic slab (Ernst and Liou, 1995). Although rare, the existence of UHP mineralogical relics in metagranitoids and quartzofeldspathic gneisses (e.g. Sobolev and Shatsky, 1990; Biino and Compagnoni, 1992; Wallis *et al*., 1997; Ye *et al*., 2000) demonstrates that such subduction does occur, and that sialic material may be carried down to depths of at least 90–140 km. UHP metamorphism is a predictable consequence of this process, and is independent of lithology. In the model presented here, the presence of low-density rocks is a prerequisite for the subsequent buoyant uplift–exhumation of the deeply buried terrane. Large masses of eclogitic oceanic crust would remain negatively buoyant, hence would continue descending into the deep mantle. For this reason, resurrected UHP complexes are restricted to zones of continental and microcontinental collision, the tectonic environment of Alpine-type subduction.

2 Pacific-type subduction of a quartzofeldspathic ± pelitic accretionary prism involves the geologically simultaneous underflow of a longitudinal belt of low-density material. Entrance of a voluminous section of greywacke-dominated material into the convergent plate junction would almost certainly terminate subduction unless offloading and/or decoupling of the low-density material occurred. Because of the weak bonding between the relatively incompetent sedimentary section and the sinking oceanic crust-capped plate, decoupling would be expected to take place episodically at relatively shallow approximately 20–50 km depths.

3 Lithospheric slabs descend to depths of c. 650–700 km along inclined Benioff–Wadati zones, so where underflow exceeds a few centimetres per year, HP–UHP recrystallization is nearly inevitable (Peacock, 1995). Oceanic crust transformed to eclogite-facies mineral assemblages is denser than mantle peridotite, and garnet lherzolite is slightly denser than plagioclase- and spinel-bearing analogues, so the oceanic lithosphere will continue sinking after phase transformations occur. Buoyancy will overcome frictional resistance and promote the ascent of entrained sialic masses after disengagement from the downgoing plate only where a sufficiently large volume of low-density material is subducted (the body force must exceed the frictional resistance to upward motion). Accordingly, recovered collisional complexes, and Circumpacific HP terranes as well, consist chiefly of quartzofeldspathic bulk compositions and phase assemblages, possessing aggregate densities considerably less than that of the mantle they dynamically displaced during subduction. Metabasaltic–metagabbroic ± metaperidotitic complexes constitute only a minor portion of resurrected HP–UHP terranes, or the latter would not be sufficiently buoyant to return to midcrustal levels.

Thus all known UHP complexes, and most HP terranes worldwide consist dominantly of low-density sialic materials, with mafic ± ultramafic rock types constituting no more than 5–15% by volume (Bailey *et al*., 1964; Banno, 1964; Frey *et al*., 1974; Coleman and Wang, 1995b; Lennykh *et al*., 1995; Hacker *et al*., 1996; Maruyama *et al*., 1996; Terry *et al*., 2000a,b; Maruyama and Parkinson, 2000; Kaneko *et al*., 2003). This is a reflection of the buoyancy-driven ascent process that allows fragments of such profoundly subducted terranes to be returned toward the Earth's surface.

4 The relative incompetence of shaley mélange and greywacke-rich, imbricated subduction complexes compared with old continental crust promotes the insertion of fragments of the underlying oceanic crust ± serpentinized harzburgite basement and tectonically overlying, variably hydrated peridotite of the mantle wedge into outboard Circumpacific terranes; in contrast, the structural integrity of descending Alpine sialic entities generally disfavours faulting of unrelated mafic and ultramafic rocks into the collisional complex. Tectonic incorporation of oceanic crust into an accretionary mélange typically results from the shearing off of oceanic plateaus, escarpments and seamounts during subduction (Cloos, 1993) and many more such oceanic crustal asperities are sampled attending sustained Pacific-type underflow compared with Alpine-type closure of small ocean basins.

5 Micaceous sialic crust probably remains dry under UHP P–T conditions, hence the prograde conversion to UHP phase assemblages would be kinetically inhibited. In contrast, during prograde subduction-zone metamorphism, clinoamphibole-bearing mafic lithologies must devolatilize, favouring the crystallization of stable UHP minerals. Consequently, oceanic crustal lithologies are more apt to develop coesite and other UHP indicator minerals than are sluggishly transforming continental rock types.

Most quartzofeldspathic and pelitic rocks are strongly foliated, whereas carbonates, cherts, and coarse-grained mafic igneous rocks are massive and nearly anhydrous. Thus during decompression and retrogression, the former are relatively more permeable to aqueous fluids than are the latter, so once formed, the retention of UHP relics as inclusions in zircons and other strong, unreactive container minerals (e.g. garnet, clinopyroxene) is kinetically favoured in eclogites, impermeable siliceous schists, and some marbles, but disfavoured in ductiley sheared sialic units. The absence of H_2O and closed-system recrystallization in situations where UHP relics are preserved is corroborated by the anomalously low $\delta^{18}O$ values measured in garnet, rutile and omphacite from Dabie–Sulu belt UHP coesite eclogites (Yui *et al*., 1995; Baker *et al*., 1997; Rumble, 1998; Zheng *et al*., 2003). Lack of a separate aqueous phase after UHP recrystallization is also suggested by experimental rate studies of the conversion of coesite to quartz which indicate that, at moderate temperatures, even H_2O contents of 400–500 ppm in SiO_2 are sufficient to cause conversion to the low-P polymorph on decompression (Mosenfelder and Bohlen, 1997). Even so, very rare relict coesite has been discovered as micro-inclusions in zircons from Dabie Shan quartzofeldspathic gneisses (Tabata *et al*., 1998; Ye *et al*., 2000; Liu *et al*., 2004).

6 To elucidate the thermal regime and possible retrograde *P–T* trajectories allowing preservation of HP–UHP relics, the exhumation of both Alpine- and Pacific-type terranes has been modelled (Ernst and Peacock, 1996) as extensional along the upper bounding surface of a thin sheet attending subduction-refrigeration along its lower bounding surface (Fig. 10). The process does not require the wholesale uplift of a lithospheric plate, or even the full thickness of continental crust or accretionary prism. Accordingly, post-metamorphic erosional debris, while considerable in many cases, need not be voluminous in either Pacific margin or continental collisional orogens.

7 During Circumpacific subduction, thousands of kilometres of oceanic lithosphere return to the mantle, so sufficient time and magma production are available for the landward development and maturation of a calcalkaline igneous plumbing system. Pressure-overstepped dehydration of abundant clinoamphiboles in the downgoing oceanic crust-capped slab at depths exceeding 70–80 km (probably reaching a maximum near approximately 100 km because of sluggish devolatilization attending rapid subduction at low temperatures) would provide sufficient H_2O necessary to account for the generation of largely I-type calcalkaline magmas through the partial fusion of oceanic crust and/or undepleted mantle wedge (Drummond and Defant, 1990; Kushiro, 1990; Hawkesworth *et al*., 1993; Morris, 1995). Vertical rise of such melts results in construction of andesitic–granitic arcs inboard from Pacific-type convergent plate junctions. The roughly contemporaneous subduction complexes consist chiefly of quartzofeldspathic debris sourced from the landward calcalkaline arcs, so their accretion, underflow and exhumation must follow early stages of volcanism–plutonism. Thus, so-called paired belts exhibit a longer time span for magmatic arc construction, and formation of the derivative HP–UHP subduction complex cannot precede it during Pacific-type underflow.

In contrast, continental collision in many cases involves relatively short-lived underflow of an old, thermally relaxed oceanic crust-capped lithospheric section, hence an igneous arc may not have had time to develop on the stable, non-subducted plate prior to crustal suturing. Moreover, the micas that characterize the sialic crust remain stable to subduction depths of at least approximately 140 km; thus, the arrival of continental crust at a convergent plate junction and its deep underflow would be expected to severely reduce the evolution of aqueous fluids in the *P–T* subduction realm where hydrous calcalkaline magmas are generated.

8 Long-sustained oceanic underflow along Pacific-type subduction zones produces a massive, temporally representative, calcalkaline volcanic–plutonic arc because magmas move up into the crust semicontinuously. In contrast, outboard quartzofeldspathic subduction complexes of both collisional and seaward Circumpacific orogenic belts decouple from the descending lithosphere episodically; the return of tectonic slices of such low-density material to midcrustal levels is a sensitive function of the changing geometry and physical properties of various units making up the architecture of the convergent junction.

9 Open folding is typical of the wall rocks and remobilized, deformed basement of calcalkaline island arcs and Andean margins, reflecting the dominantly vertical tectonics accompanying invasion of magma bodies, and solidification as both superjacent (mostly early-stage) and subjacent (chiefly late-stage) units. In contrast, outboard Pacific- and Alpine-type subduction complexes exhibit oceanward fold vergence, and landward inclination of thrust faults, reflecting the sense of shear in the tectonic environments in which they formed and were penetratively deformed. Their piecemeal exhumation apparently is driven by buoyancy; as the subducted units warm in the upper mantle, the strengths and viscosities of subducted lithotectonic units decrease, hence tectonic squeezing cannot be a major factor in their uplift. Because of rapid ascent from great depth and moderate temperatures, the critical requirement for preservation of HP–UHP relict phases in even fragmentary manner is that heat must be conducted away efficiently, which in turn requires decompressing rock masses to be characterized by large surface/volume ratios. Thus for HP and UHP complexes to survive decompression, transport to midcrustal levels in décollement-type structures evidently must occur, allowing substantial cooling of the UHP assemblages. In some cases, this may be followed by domical ascent (as tectonic aneurysms?) toward the Earth's surface.

Final Statement

The described petrotectonic features appear to be compatible with the proposed mountain building processes for the formation, thermal evolution and ultimate tectonic exhumation – exposure through erosion and/or gravitational collapse of collisional and marginal Circumpacific terranes (e.g. Ernst, 1970; Platt, 1986, 1987, 1993; Maruyama, 1997), as well as the formation of voluminous calcalkaline arcs inboard from Pacific-type plate junctions (Ernst, 1999). Clearly, the natures of materials carried down subduction channels, extents of deep-seated devolatilization, and rates of transformation strongly influence the resultant configuration of HP–UHP metamorphic belts. Exhumation to midcrustal levels appears to have been driven principally by buoyancy, and the ascent in most cases was rapid. Exhumation rates greater than 5 mm per year averaged over more than 20 Myr and seemingly required by HP and UHP geochronologic data (Ernst *et al*., 1995, 1997; Gebauer, 1996; Gebauer *et al*., 1997; Hacker *et al*., 2000, 2003b; Rubatto and Hermann, 2001) substantially exceed currently measured uplift and erosion rates in the Himalayas (Le Fort, 1996; Searle, 1996), but are compatible with rates of exhumation of approximately 4 mm per year calculated by Genser *et al*. (1996) for the eastern Alps. Yet higher ascent rates have been proposed for the Dabie–Sulu belt (Liou and Zhang, 1995; Grasemann *et al*., 1998), the Kokchetav massif (Dobretsov *et al*., 1995), eastern Taiwan (Lin and Roecker, 1998) and the southern Alps, South Island, New Zealand (Blythe, 1998).

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