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Subduction-zone metamorphism, calc-alkaline magmatism, and convergent-margin crustal evolution

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ABSTRACT

At the dawn of the age of plate tectonics, Akiho Miyashiro published a seminal paper in which he drew attention to significant differences in regional metamorphic phase assemblages, described contrasting geologic occurrences, and inferred their characteristic ranges of physical conditions. He advanced the paired metamorphic belt concept, involving an oceanward, narrow, high-P/low-T blueschist zone intimately intermixed with ophiolites, and a landward, broad, low-P/high-T realm associated with arc volcanic–plutonic rocks. More recent studies have illuminated the following: (1) Two main types of convergent plate junction can be distinguished, Pacific underflow of thousands of km of oceanic lithosphere, and Alpine closure of an intervening oceanic basin leading to continental collision. (2) Plate descent carries mafic + felsic lithologies to depths of ~35–120 km or more at relatively low temperatures, producing high-pressure (HP) Pacific-type and ultrahigh-pressure (UHP) Alpine-type metamorphic terranes, respectively. (3) Exhumed HP–UHP complexes display low-aggregate bulk densities, reflecting chiefly buoyancy-driven ascent of quartzofeldspathic allochthons. (4) At a depth of ~35 km in a warm-lithosphere subduction zone, sialic crust should melt if an aqueous fluid were abundant; it does not, hence the activity of H_2O must be low at such depths. (5) Global observations indicate that volcanic–plutonic arcs are sited 100 ± 20 km above subducting oceanic plates; devolatilization of sinking, hydrated oceanic crust ± mantle promotes the generation of calc-alkaline arc melts in this magmagenic zone. (6) Pacific-type underflow of basaltic crust-capped plates produces new and recycled continental crust, whereas Alpine-type convergence reshuffles collided terranes but does not generate juvenile sialic crust.

1. Introductory statement

Miyashiro (1961) published a landmark synthesis describing contrasting mineral parageneses in three principal metamorphic terrane end members, and showed that their geologic settings and inferred physical conditions were distinctly different. He called attention to the coeval, post-Paleozoic paired belts of contrasting *P–T* ranges that rim the Pacific Basin, and are especially well exemplified in the Japanese archipelago: an outboard, high-P/low-T ophiolitic blueschist belt parallels an inboard, broad, low-P/high-T belt spatially associated with intermediate and felsic igneous rocks. Since 1961, a wealth of new phase equilibrium data and thermochemical studies, coupled with wide ranging structural geologic + geophysical investigations, have allowed Miyashiro's original petrologic model to be placed in a plate-tectonic framework. His concept of paired metamorphic belts has stood the test of time remarkably well, although naturally, some aspects have been slightly modified by the acquisition of abundant new data.

2. Convergent plate junction end members

Pacific convergent boundaries mark curvilinear sutures where thousands of km of oceanic lithosphere return to the upper mantle without the arrival of important masses of sialic crust at the subduction zone (Fig. 1a). This plate-tectonic realm produces three subparallel belts, an outboard, largely metasedimentary accretionary trench complex, a medial, longitudinal forearc basin, and an inboard, massive, contemporaneous calc-alkaline arc. The long, narrow trench prism consists of a low-temperature, low-heat-flow belt in which folds verge oceanward, and thrust faults are controlled by the subduction zone and dip beneath the hanging-wall plate, whereas the broad magmatic arc is characterized by open folding and a high-temperature, high-thermal-flux regime (Miyashiro, 1961, 1967; Ernst et al., 1970; Dickinson, 1972). The trench complex is deposited on the sinking oceanic crust (*e.g.*, Franciscan, Aleutian, and Chile–Peru subduction systems); the forearc section and inboard volcanic–plutonic terrane are constructed on the nonsubducted continental margin or island arc (*e.g.*, Sierran, Indonesian, and Andean arcs).

Alpine convergent boundaries form where consumption of ocean lithosphere causes insertion of a continental promontory, microcontinent, or arc beneath the stable continental plate (Fig. 1b). A salient of the

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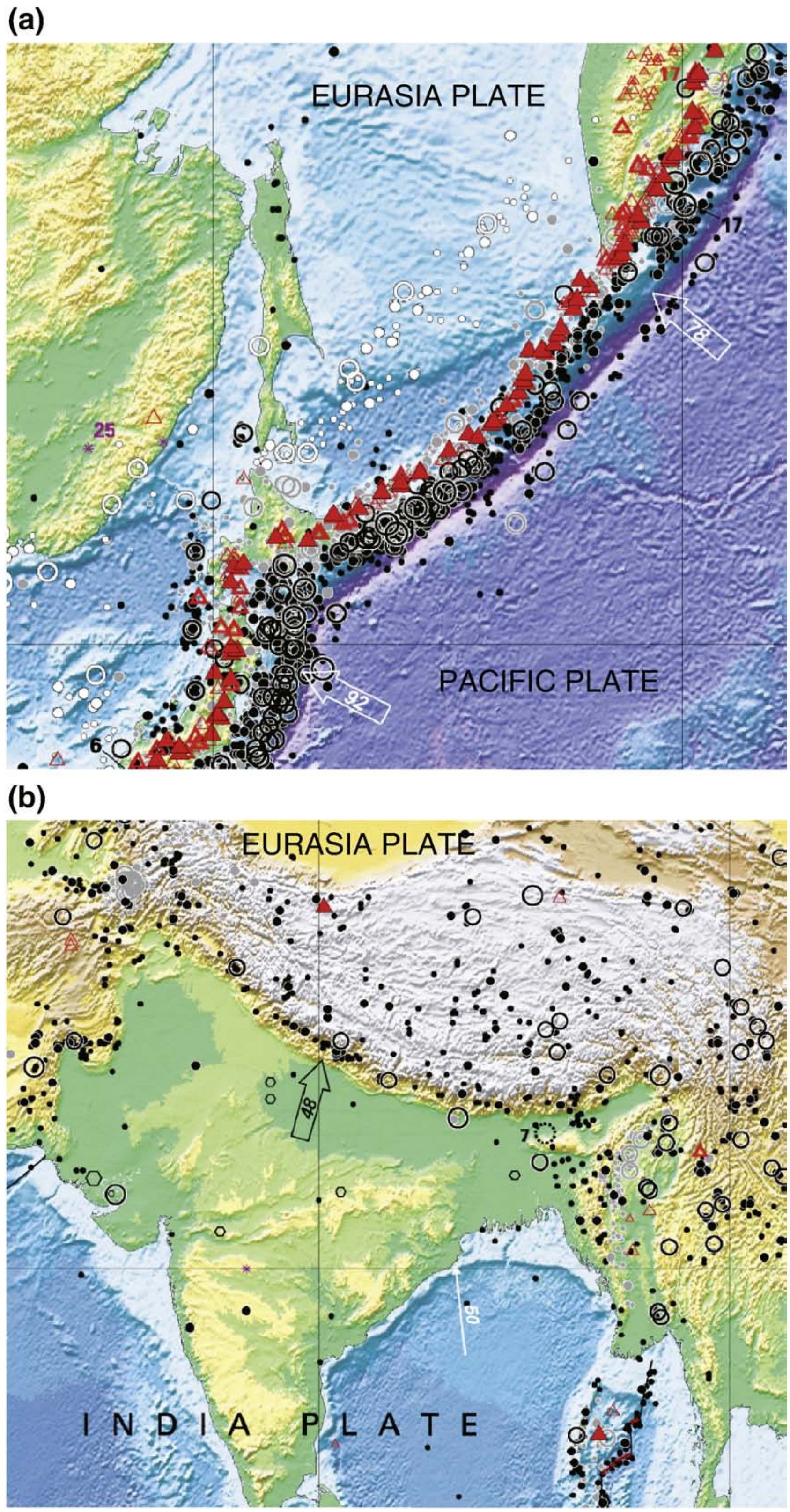

Fig. 1. Modern oceanic trenches, seismicity, and arc volcanos in (a) the Kamchatka–Kuriles–northern. Japan Pacific-type convergent plate junction and (b) the Indian–Tibetan Alpine-type collisional suture, from Simkin et al. (2006). Convergence rates = mm/yr; circles = earthquake epicenters; triangles = volcanos.

downgoing sialic terrane may descend to great depth along with the surrounding oceanic lithosphere, because the overall density of the dominantly oceanic plate exceeds that of the displaced asthenosphere. During underflow, the sialic crust is thickened by accretion, underplating, shortening, and amalgamation–continental collision. Such contractional orogens exhibit a distinct lack of coeval calc-alkaline igneous activity. Typical examples include the Urals, Alps, and Himalayas (Hamilton, 1970; Dal Piaz et al., 1972; Molnar and Tapponnier, 1975;

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Molnar et al., 1987; Burchfiel et al., 1989). Rarely, continental lithosphere may be carried beneath a young, hot, less dense oceanic crust-capped plate (*e.g.*, Oman and Sulawesi; Searle et al., 2004), but this is uncommon because most oceanic plates are denser than the asthenosphere, whereas continental plates are relatively buoyant.

Along the length of a convergent junction, underflow may range by degrees between purely Pacific- and Alpine-type. Moreover, although we tend to concentrate on the down-dip component of underflow, a large component of arc-parallel slip is present in nearly all orogenic belts because of the curvilinear nature of convergent plate junctions (*e.g.*, Fitch, 1972; Howell, 1985; Koons et al., 2003).

3. HP and UHP metamorphic belts

As noted above, paired metamorphic belts mark Pacific-type convergent plate junctions. The outboard subduction complex consists of poorly consolidated, mainly quartzofeldspathic units ± metashales ± serpentinites that recrystallized at low temperatures and high pressures. Reflecting the subduction channel setting directly below the hanging-wall lithosphere that acts as a stress guide, folds verge oceanward; shear zones, foliation, lineation, and thrust faults are inclined landward beneath the hanging-wall plate. Recovered depths of HP metamorphism lie in the range 35 ± 15 km (Bailey et al., 1964; Ernst, 1971; Aoya, 2001), indicating relatively shallow decoupling from the descending oceanic plate. Well inboard, voluminous granitoids invade the preexisting basement and an overlying, comagmatic volcanic cover series. Wall rocks show the effects of high-T, low-P (15 ± 10 km) recrystallization. Metamorphic mineral zonations reflect advective heat supplied by rising magmas, and consist of a series of superimposed contact aureoles encircling individual plutons (Barton et al., 1988). Hornfels and open, cylindrical folding typifies these upper-crust metamorphic terranes (Ernst, 1992).

Alpine-type collisional complexes consist of preexisting sialic massifs, as well as deformed, superjacent autochthonous and allochthonous units. Metamorphic belts are not paired (Frey et al., 1974). Subsolidus HP recrystallization is common, with mineral parageneses comparable to those of Pacific-type subduction complexes. Some contractional orogens include tectonic slices that retain tiny scattered mineralogic relics of coesite and/or microdiamond, chiefly as inclusions in strong, unreactive container phases like zircon and garnet. These neoblastic phases must have crystallized at ultrahigh pressures and somewhat higher temperatures, but prograde geothermal gradients appear to be similar to progressive *P–T* trajectories of Pacific-type subduction complexes (Chopin, 1984; Smith, 1984; Sobolev and Shatsky, 1990; Coleman and Wang, 1995). UHP conditions attend the subduction of old, cool continental crust well bonded to the downgoing oceanic plate to depths of 90–125 km or more; in the process, some quartzofeldspathic sections may decouple from the downgoing lithosphere and migrate back up the convergent plate junction.

HP and UHP metamorphic conditions are depicted in Fig. 2, a petrogenetic grid based on experimentally determined phase equilibria for rocks of more-or-less basaltic bulk composition. Contrasts in metamorphic-facies assemblages, and attendant prograde *P–T* gradients generated in subduction zones are clear. For comparison, a high-heat-flow geotherm, and the inferred calc-alkaline magmagenic zone are also illustrated. Subduction-zone *P–T* trajectories are slightly to substantially cooler than those that attend the production of volcanic–plutonic arc melts.

4. Buoyancy-driven exhumation of HP–UHP terranes

Ductilely deformed, relatively thin nappes and thrust sheets that formed in subduction channels (*e.g.*, Koons et al., 2003; Hacker et al., 2004; Terry and Robinson, 2004) make up the architecture of most recovered HP–UHP complexes; others may be coherent, massive sections of the continental crust (Young et al., 2007). Ascent to shallow crustal levels reflects one or more of several processes: tectonic extrusion (Maruyama et al., 1994, 1996; Searle et al., 2003; Mihalynuk et al., 2004; Masago et al., 2009); corner flow due to 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). Cool, sinking oceanic lithosphere evidently retreats seaward faster than advance of the nonsubducted plate (Molnar and Atwater, 1978; Seno, 1985; Busby-Spera et al., 1990; Hamilton, 1995), so compression and thrusting along such lithospheric junctions cannot account for the exhumation of subducted sialic slabs. Constriction by a backstop requires buoyancy or tectonic shortening to cause the return flow of subducted units. Extension and erosion help to expose HP–UHP terranes after they have risen to crustal levels, but do not produce the major pressure discontinuities (up to >2 GPa) that mark contacts juxtaposing deeply subducted and nonsubducted sections (Ernst et al., 1970).

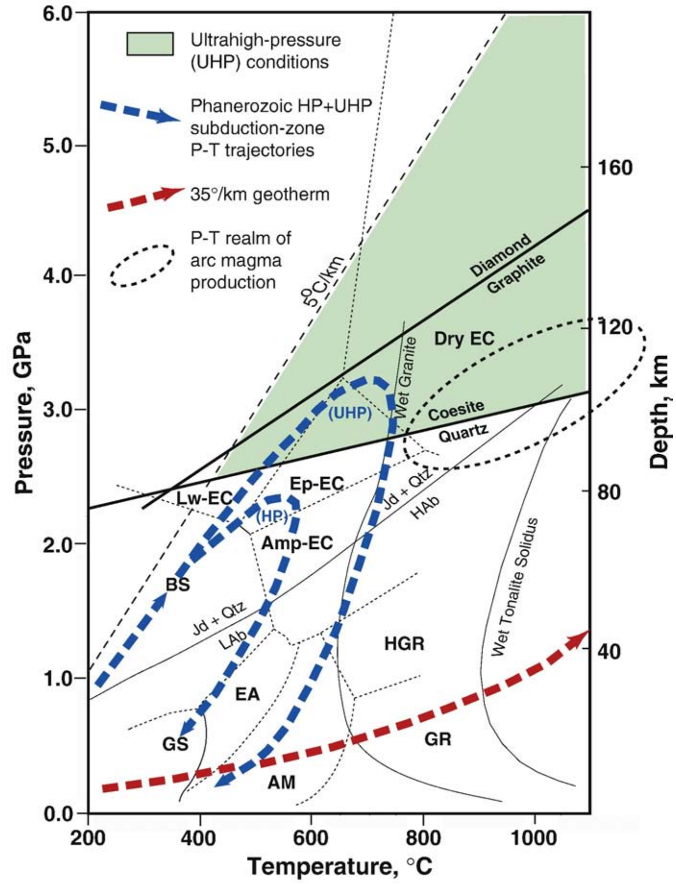

Fig. 2. Prograde subduction-zone *P–T*–time paths followed by exhumation to mid-crustal levels for UHP imbricate thrust slices such as exposed in the Kaghan Valley, western Himalayan syntaxis (O'Brien, 2001; Kaneko et al., 2003), and for oceanic plate underflow such as the Franciscan HP belt, central California (Ernst, 1993; Dalla Torre et al., 1996). A very low subduction-zone geotherm of 5 °C/km, and a 35 °C/km high-heat-flow geotherm are shown for reference. The petrogenetic grid for rocks of roughly basaltic composition is modified from Liou et al. (1998), Okamoto and Maruyama (1999), and Hacker et al. (2003b). Hack et al. (2007, fig. 28) reviewed experimental data with the solidus curves for granite + H_2O (and tonalite + H_2O) terminating at critical endpoints near 2.6 GPa, and for basalt H_2O at ~4.2 GPa (the critical endpoint for peridotite + H_2O exceeds 9 GPa; personal communication, P. J. Wyllie, 2009). Mineral abbreviations: Ar = aragonite; Cc = calcite; Jd = jadeite; Qtz = quartz; LAb = low albite; and HAb = high albite. Metamorphic-facies abbreviations: 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; Px–Hf = pyroxene hornfels.

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

Buoyant drive coupled with erosional unroofing provides a plausible mechanism to account for the exhumation of low-density crustal slices. Geologic relationships, laboratory scale models (Chemenda et al., 1995, 1996, 2000), and numerical simulations (Beaumont et al., 1996, 1999; Pysklywec et al., 2002) document this process (see also: Parkinson et al., 2002; Carswell and Compagnoni, 2003; Malpas et al., 2004). The strength, integrity, and composition of subducted lithospheric materials, extent of deep-seated devolatilization, and rates of recrystallization strongly influence the characteristics of the resultant ultrahigh-pressure metamorphic belts (Ernst et al., 1998). Petrotectonic features of UHP Phanerozoic complexes thus reflect their plate-tectonic settings and *P–T* histories (Maruyama et al., 1996; Ernst, 2005).

Densities (g/cm^3) of unaltered oceanic crust, 3.0, continental material, 2.7, and anhydrous mantle, 3.2, increase with elevated pressure, reflecting the transformation of open framework silicates to more compact layer-, chain-, and orthosilicates. Stable UHP mineralogic assemblages and computed rock densities appropriate for burial depths of about 100 km and 700 °C are roughly: metabasaltic eclogite, 3.55; eclogitic granitic gneiss, 3.05; and garnet peridotite, 3.35 (Ernst et al., 1997; Hacker et al., 2003a). Even when transformed completely to an HP–UHP assemblage, K-feldspar + jadeite + coesite-bearing granitic gneiss remains ~0.30 g/cm^3 less dense than garnet lherzolite, whereas metabasaltic eclogite is about 0.20 g/cm^3 denser than upper mantle lithologies. Consequently, some subducted packets of UHP metamorphosed sialic crust are sufficiently buoyant to overcome the traction of the oceanic plate carrying them downward, as evident from the fact that quartzofeldspathic nappes are now exposed at the Earth's surface. HP–UHP belts worldwide consist dominantly of such low-aggregate-density lithologies.

Thus, attending Pacific-type subduction of a largely sedimentary mélange, devolatilization and increased ductility cause decoupling of subducted HP materials from the sinking oceanic plate at depths approaching ~35 km, followed by piecemeal ascent. In contrast, for a continental salient that is an integral part of descending, largely oceanic lithosphere, disengagement of a crustal slice may be delayed to depths of 90–125 km or more. Insertion of increasing amounts of low-density material into the subduction zone gradually reduces the overall negative buoyancy of the sinking lithosphere. Attainment of neutral buoyancy at modest upper mantle depths, where the plate is in extension (Isacks et al., 1968), can result in rupture and accelerated sinking of the dense oceanic lithosphere. Slab breakoff (Sacks and Secor, 1990; von Blanckenburg and Davies, 1995) increases the net buoyancy of the low-density sialic UHP complex, and allows Alpine sheets to decouple from the oceanic plate and move back up the subduction channel (van den Beukel, 1992; Davies and von Blanckenburg, 1998). During collision, reduced frictional resistance also may enhance exhumation as the continental crust warms in the upper mantle, and passes through the brittle–ductile transition (Stöckhert and Renner, 1998).

5. Anatexis of the continental crust

Zeolites, clay minerals, and carbonates, the chief volatile-bearing weathering and low-grade metamorphic alteration products of igneous, sedimentary and metamorphic rocks in the near-surface environment, devolatilize at relatively shallow subduction depths (Ernst, 1990). Thus deeper than ~25 km, the principal hydrous minerals of downgoing continental crustal rocks include muscovite and biotite, phases stable to 800–1100 °C at subduction depths >125 km (Stern et al., 1975). Phase relations in felsic units are illustrated schematically in Fig. 3. This diagram shows that white mica and biotite remain stable to pressures exceeding 3 GPa for typical subduction-zone geothermal gradients of 5–10 °C/km and for higher geothermal gradients as well, especially in the absence of a separate H_2O phase (Vielzeuf and Holloway, 1988; Vielzeuf and Montel, 1994; Skjerlie and Johnston, 1996; Nichols et al., 1996; Patiño Douce and Beard, 1996; Patiño Douce and McCarthy, 1998; Massone and Szpurka, 1997; Luth, 1997). Reflecting the broad *P–T* stability ranges of these hydrous layer silicates, quartzofeldspathic rocks do not devolatilize extensively at uppermost mantle depths under normal subduction-zone conditions (Ernst et al., 1998). Accordingly, the partial fusion of sialic lithologies at typical prograde HP–UHP subduction depths would not be expected to occur, given the low activities of H_2O (Fig. 2).

In addition, lacking a rate-enhancing aqueous fluid, sialic units are unlikely to transform rapidly, or totally, to UHP mineral assemblages

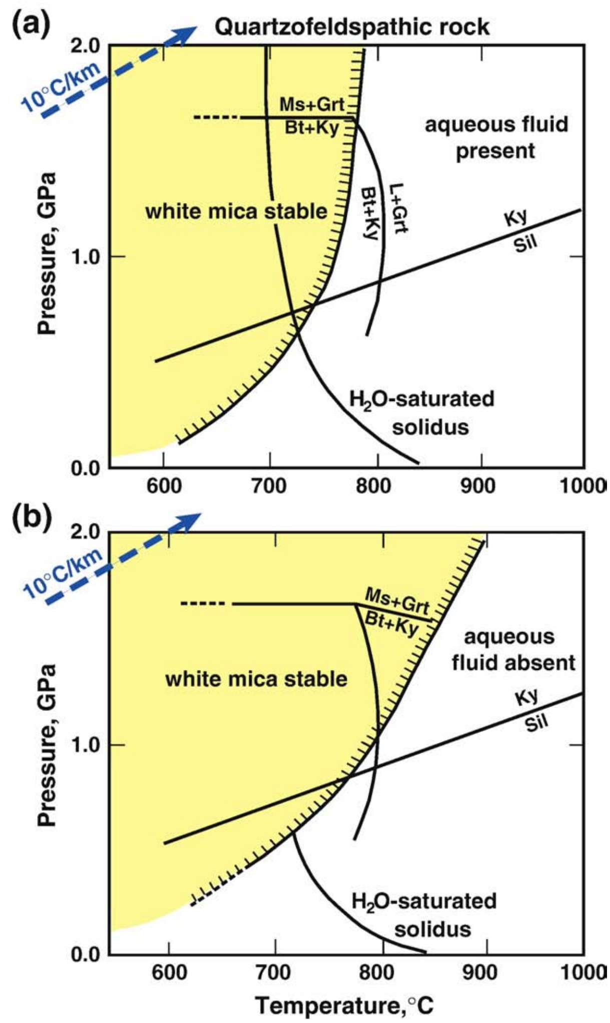

Fig. 3. Experimentally determined and computed petrogenetic grid under (a) H_2O fluid-excess and (b) H_2O fluid-absent conditions for metagraywackes + pelitic schists with intermediate bulk-rock Fe/Mg ratios, after Vielzeuf and Holloway (1988). *P–T* stability fields of white mica are shown in yellow. Abbreviations: Bt = biotite; Grt = garnet; Ky = kyanite; L = liquid; Ms = white mica; Sil = sillimanite. Phengitic micas are stable to pressures >3.5–4.0 GP in intermediate and felsic rocks, as shown by 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 at least up to 3.5 GPa in Fe-rich metapelites (Nichols et al., 1996), and for simple metaluminous bulk compositions to ~9.0 GPa (Luth, 1997). A prograde subduction-zone gradient of 10 °C/km is illustrated.

(Hacker, 1996; Austrheim, 1998). In any case, continental crust completely or incipiently converted to UHP phase assemblages at upper mantle depths, would still remain buoyant relative to the surrounding mantle, and should rise to mid-crustal levels. In contrast, the main hydrous phases in mafic and ultramafic rocks are hornblende, chlorite, and serpentine—pressure-limited minerals that devolatilizes at moderate temperatures where depths exceed ~70–90 km (see Figs. 4 and 5). In the local presence of an evolving aqueous fluid, metabasaltic eclogites are far more likely to recrystallize to the stable prograde HP–UHP assemblage than are the more refractory mica-bearing felsic rocks. Thus, eclogitized oceanic crust becomes negatively buoyant compared to both near-surface oceanic basalt and garnet lherzolite, and would continue to sink. This relationship explains why exhumed HP–UHP terranes worldwide consist of more than 90% low-density felsic material, and contain only small proportions of dense mafic ± anhydrous ultramafic rock types (Ernst, 2001; Hacker et al., 2003b).

6. Subduction of oceanic lithosphere and arc magma generation

Experimental phase equilibrium investigations summarized in Fig. 4 (Poli, 1993; Schmidt and Poli, 1994; Liu et al., 1996; Poli and Schmidt, 1997; Okamoto and Maruyama, 1999) show that, for typical prograde subduction-zone *P–T* trajectories in the system Ca–Mg–Fe–Al–Si–O–H, hydroxyl-bearing mafic phases are the main constituents of metabasaltic rocks. At geothermal gradients of 5–10 °C/km, Ca-, Mg- and Na-amphiboles and chlorite have high-P limits in the 2–3 GPa range. Serpentine, the major hydrous phase in altered mantle harzburgites of the uppermost mantle, also has a restricted high-P limit for geotherms of 10 °C/km (2.0 GPa) and 5 °C/km (3.9 GPa), as clear from Fig. 5 (Ulmer and Trommsdorff, 1995; Wunder and Schreyer, 1997; Pawley, 1998; Evans, 2004). Although these mafic chain and layer silicates should dehydrate at depths exceeding ~70–80 km, low-T fluid evolution may be retarded due to sluggish kinetics, (*i.e.*, pressure overstepping). Thus for relatively cold oceanic plates, most metabasaltic blueschists, eclogitic amphibolites, and serpentinites expel H_2O at the magmagenic depths of ~100 km (Hatherton and Dickinson, 1969), and would achieve stable eclogitic and anhydrous, depleted peridotitic phase assemblages (Ernst, 1999; Rondenay et al., 2008).

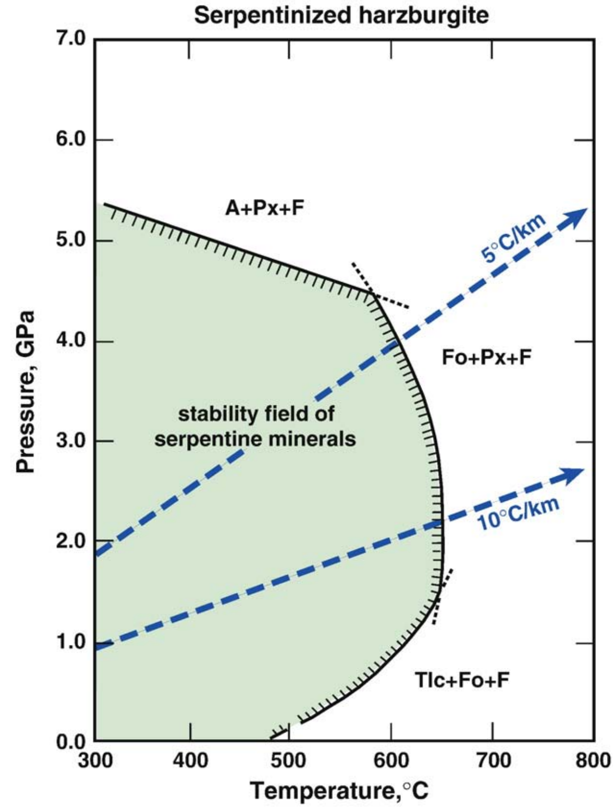

Fig. 5. Phase diagram for the stability of serpentine in the harzburgite + excess H_2O system, after Wunder and Schreyer (1997). Reactions were reversed, indicating chemical equilibrium. The serpentine *P–T* stability field is shown in blue. Abbreviations: A = phase A (hydrous magnesium silicate); F = aqueous fluid; Fo = forsterite; Px = clinoenstatite; Tlc = talc. See also Ulmer and Trommsdorff (1995), Pawley (1998), and Evans (2004). Prograde subduction-zone gradients of 5 and 10 °C/km are illustrated.

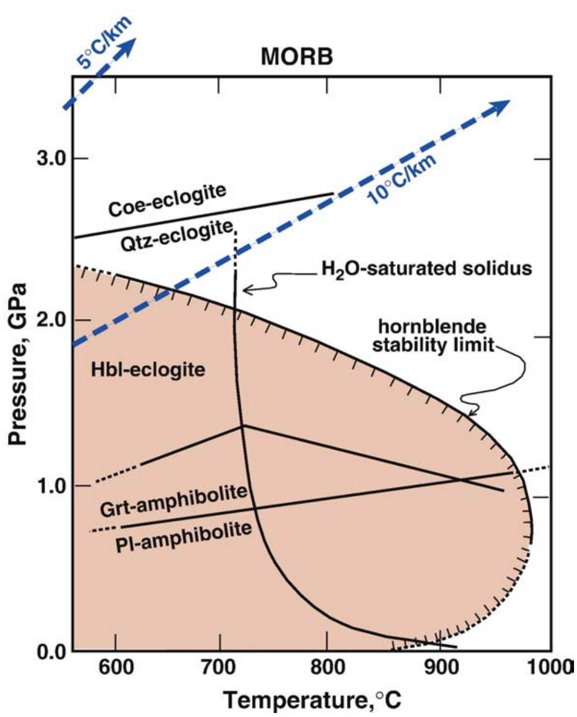

Fig. 4. Petrogenetic grid for the amphibolite–eclogite transformation in the system MORB basalt + excess H_2O with oxygen fugacity defined by the $FeSiO_4$–Fe_3O_4–SiO_2 buffer, after Liu et al. (1996). Experiments lasted up to 1630 h and reaction reversals suggest a close approach to chemical equilibrium. The Ca-amphibole *P–T* stability field is shown in orange. Prograde metamorphism involves the devolatilization of garnet amphibolite and hornblende eclogite under *P–T* conditions near those of the quartz–coesite transition. Abbreviations: Coe = coesite; Grt = garnet; Hbl = hornblende; Pl = plagioclase; Qtz = quartz. Prograde subduction-zone gradients of 5 and 10 °C/km are illustrated.

However, Fig. 3 shows that white mica and biotite remain stable to pressures exceeding >3.0 GPa for typical subduction-zone geothermal gradients. On descent to depths of as much as 125 km, mica-rich sialic crust fails to evolve important amounts of H_2O, so volatile flux severely diminishes where large volumes of micaceous intermediate and felsic crust are carried down to great depths. In stark contrast, subducted amphibolitized oceanic lithosphere and altered mantle become unstable, dehydrate, and generate most of the deep-seated volatile flux and the consequent partial melting that produce the calc-alkaline suite along and above a subduction zone. Thus, continental collision commonly does not produce a volcanic–plutonic arc whereas the long-continued underflow of mafic–ultramafic oceanic lithosphere does (Peacock et al., 1994; Ernst, 1999; Davidson et al., 2007).

The low-P, anhydrous melting temperature of granite, ~950 °C, lies 150–200 °C below the solidi of mantle peridotite + basaltic crust, ~1200 °C, and the presence of $H_2O \pm CO_2$ would further lower these temperatures (Fig. 2). Fig. 6 shows that with increasing fluid pressure,

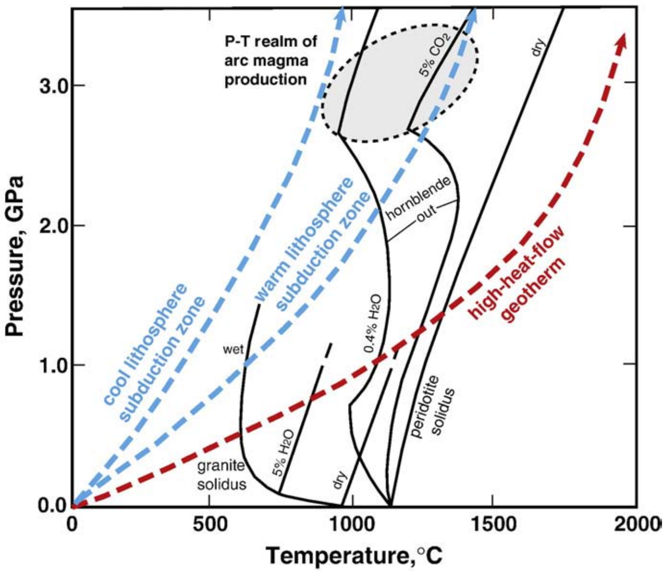

Fig. 6. Experimentally determined melting curves for fertile peridotite, dry and with H_2O or CO_2 present, and of granite, dry and with minor or excess H_2O (Lambert and Wyllie, 1972; Wyllie, 1979, 1981; Presnall and Gudfinnsson, 2005). The H_2O-saturated solidus of basalt lies ~200 °C above the H_2O-saturated granite solidus (Poli and Schmidt, 1995, 2002; Liu et al., 1996). Modern cool and warm subduction-zone *P–T* trajectories and a high-heat-flow geotherm (Clark and Ringwood, 1964; Ernst, 2005) are indicated; to a depth of 35 km, typical gradients are ~10, ~18, and ~35 °C/km. The gray ellipse indicates the inferred range of deep-seated physical conditions for calc-alkaline melt generation.

polar molecules such as H_2O become quite soluble in silicate melt, strongly lowering fusion temperatures. Generation of a calc-alkaline arc chiefly depends on the presence and incorporation of volatiles, which cause a lowering of the melting temperatures of transformed, partly hydrated oceanic crust, the upper mantle wedge, and/or lower crustal mafic protoliths at magmagenic depths along and above the sinking oceanic lithosphere. The breakdown of hornblende + chlorite ± serpentine apparently provides the *P–T–*H_2O conditions necessary for the partial fusion of basaltic crust and/or hanging-wall mantle (Fig. 6), and accounts for the origin of arc magmas in a warm subduction-zone environment.

The closure of a relatively small ocean basin such as typifies many continental collision zones generates only minor volumes of hydrous intermediate and silicic melt, so pre-impaction volcanic–plutonic belts are poorly developed in most Alpine-type convergent plate junctions compared with Circumpacific analogues. Where extensive tracts of oceanic lithosphere are consumed prior to continental collision, volcanic–plutonic arcs do develop, but the precursor igneous activity is extinguished by the eventual arrival and subduction of large volumes of sialic crust (*e.g.*, Kohistan arc; Khan et al., 2009).

7. Growth of the continental crust

Net growth of the Earth's sialic crust is a reflection of its rate of formation minus its rate of return to the mantle; both rates presumably are functions of the thermal evolution of the planet. Based largely on isotopic data, episodic or continuous additions of mass to the continental crust seem to have reached a zenith in the late Archean and early Proterozoic (~2.7 Ga), and to have decreased ever since (O'Nions et al., 1980; Lambert, 1980; Jacobson and Wasserburg, 1981; Nelson and DePaolo, 1984; Reymer and Schubert, 1986; McCulloch and Bennett, 1994; O'Neill et al., 2007; Condie, 2007; Rino et al., 2008; Condie et al., 2009). Due to the progressive depletion of crustal components in the convecting mantle source region over the past 3–4 Gyr, the rate of formation of juvenile crust has declined with time, and is now about equal to its rate of destruction (*i.e.*, its return to the mantle through viscous drag at subduction zones ± slower continental crust delamination). Fig. 7 schematically illustrates the progressive net accumulation of

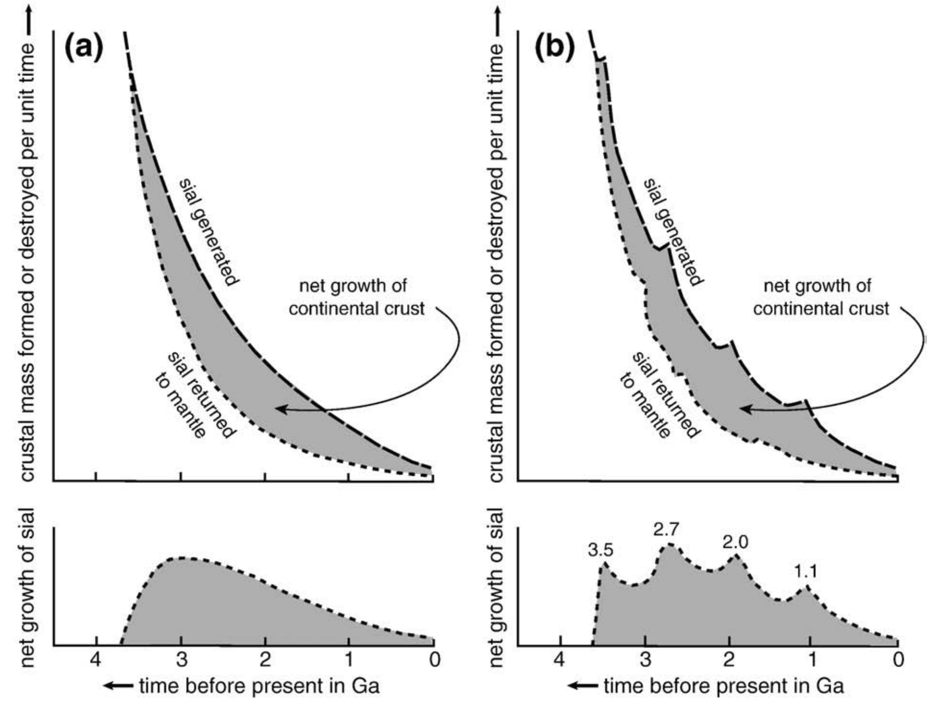

Fig. 7. Sketch of the competing processes of continental crust generation and return to the mantle (subduction + subcrustal erosion) of juvenile calc-alkaline arc rocks attending mantle–crust differentiation, illustrating net accumulation of sialic material over geologic time (*e.g.*, Jacobson and Wasserburg, 1981; Ernst, 1983, 2007; DePaolo et al., 1991; Condie, 1998, 2007; O'Neill et al., 2007, 2009). In (a), net preservation of felsic crust is assumed to be a smoothly varying function of planetary cooling. In (b), episodes of rapid overturn and greater loss of continental crust due to viscous coupling are conjectured to have followed by ~100 Myr the high rates of volcanic–plutonic activity and consequent high productivity of continental crust at 3.5, 2.7, 2.0, and 1.1 Ga. For a quantitative synthesis of global continental crust growth rates, see O'Neill et al. (2007).

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sialic crust, assuming growth as (a) a smoothly varying function reflecting thermal relaxation of the planet, and (b) a consequence of punctuated rapid mantle overturn. Insofar as the aggregation + suturing of supercontinental assemblies and their subsequent fragmentation-dispersal may strongly affect the rates of subduction and sea-floor spreading (Moresi and Solomatov, 1998; O'Neill et al., 2007; Silver and Behn, 2008), growth and preservation of the sialic crust undoubtedly are not simply a monotonic function of time. Some authorities (*e.g.*, Armstrong, 1991; Stern and Scholl, in press) even question whether or not, after an initial segregation from the differentiating planet, the mass of continental crust has grown over geologic time.

Deep-seated dehydration of the subducting amphibole + chlorite ± serpentine-rich oceanic plate (in contrast to limited dewatering of the micaceous continental crust-capped lithosphere) along the Indonesian, Aleutian, Kurile, Tonga, and most other modern convergent plate junctions results in the generation of arc magmas mainly associated with the sinking oceanic lithosphere, as noted above. Intermediate and felsic igneous activity reflects the fact that, under subduction-zone *P–T* conditions, hornblende breaks down at pressures exceeding ~2.3 GPa, chlorite and serpentine at similar depths, whereas white micas ± biotite are stable to pressures >3.0 GPa (Ernst, 1999; Li et al., 2008).

In summary, subducting amphibolitized oceanic crust and serpentinized mantle become unstable at the ~100 km depths of the magmagenic zone, and generate most of the deep-seated volatile flux and partial melting that produce the calc-alkaline suite within and above a subduction zone. Unless an H_2O-rich fluid is present as micaceous intermediate-felsic crust is carried down to great depths, volatile flux diminishes greatly. Thus, continental collision commonly does not produce abundant hydrothermal solutions or H_2O-bearing melts rising into and forming a coeval volcanic–plutonic arc, whereas the long-continued underflow of mafic–ultramafic oceanic lithosphere does (Peacock et al., 1994; Ernst, 1999; Davidson et al., 2007). Evidently Pacific-type underflow of basaltic crust-capped plates produces new and recycled sialic crust whereas Alpine-type convergence chiefly reshuffles collided terranes but does not generate new continental crust.

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