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Dunk, Dunkless and Re-dunk Tectonics: A Model for Metamorphism, Lack of Metamorphism, and Repeated Metamorphism of HP/UHP Terranes

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Abstract

High-pressure metamorphic terranes form when slivers of continental crust are pulled into the mantle by previously subducted oceanic lithosphere. Old, cold lithosphere subducts into the mantle at steep angles, dragging sialic crust to deep levels, quickly followed by steep, rapid exhumation. Young thin lithosphere subducts at shallow angles, pulling crustal slabs slowly into the shallow mantle, followed by slow exhumation, retrogression, and melting. Terranes that follow inboard-dipping subduction zones will subduct once, and will be overprinted by subsequent subduction and collisions. Terranes above outboard-facing subduction zones will be underthrust by the continental margin. Subsequent collisions will result in the re-subduction of an increasingly complex continental margin characterized by previously accreted terranes.

Statement of the Problem

A growing number of mountain systems have joined the high-pressure (HP) metamorphic club and the even more exclusive ultrahigh pressure (UHP) metamorphic club (see, for example, Liou et al., 2002 and Parkinson et al., 2002 for recent summaries). Joining the club requires that the mountain system contain at least one metamorphic terrane that contains medium-temperature (MT) or high-temperature (HT) eclogite-facies assemblages or relicts thereof. This restriction excludes terranes with low-temperature (LT) "blueschist" eclogites of the type that occur in accretionary prisms, because they probably form through a different tectonic mechanism. There are a few mountain systems that lack HP/UHP terranes, such as much of the Cordilleras of the contiguous United States. However, these systems should be searched more thoroughly before designating them as "have-not" orogens. HP/UHP terranes may remain undiscovered for two primary reasons: (1) anhydrous terranes subducted to eclogite-facies depths will not necessarily generate regional eclogite-facies assemblages for kinetic reasons; and (2) eclogite-facies terranes can be almost completely overprinted by LP and/or HT assemblages either during exhumation or during subsequent metamorphic events. The result of both processes is that evidence for metamorphism under eclogite-facies conditions can be subtle and elusive (Parkinson, et al., 2002). For example, anhydrous terranes may contain eclogite-facies assemblages only along fractures that allowed the access of H_2O, as is the case, for example, in the Bergen Arcs and the Lofoten Islands of the Norwegian Caledonides (Austrheim, 1987, Boundy et al., 2003). Overprinted terranes can be so thoroughly recrystallized that evidence for previous HP/UHP events may be preserved only as inclusions in refractory minerals such as zircons (Katayama et al., 2002).

The model adopted in this paper for the development of HP/UHP metamorphic terranes involves the subduction of sialic crust into the mantle to pressures and temperatures appropriate to eclogite-facies metamorphism followed by buoyancy-aided exhumation ("eduction") of the sialic crust toward the surface (Ernst, 1970; Andersen et al., 1991; Chemenda et al., 1996; Maruyama et al., 1996; Matte, 1998; Brueckner, 1998). These slabs or "pips" (Platt, 1987) are characteristically thin relative to their lateral extent (Ernst, 2005), are bordered by units that recrystallized under significantly lower pressures and temperatures, and are bounded by low-angle normal faults above and thrust faults below (Wheeler, 1991). These subduction/eduction events or "dunks" (Brueckner and Van Roermund, 2004) can occur several times during the closure of an ocean basin—i.e., during

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collisions between sialic terranes that occur within ocean basins (active and relict volcanic arcs, microcontinents, and composite terranes formed by the suturing of these units), during collisions between these terranes and the continents that bound the ocean basins and, finally, during the culminating continent-continent collision that closes the ocean basin. The question arises whether or not all collisions will result in deep-crustal subduction and the development of HP or UHP terranes and if not, why not? This question is relevant for mountain systems such as the Cordilleras of the contiguous United States, which lacks HP/UHP terranes despite the fact that several collisions were involved in the evolution of the system (e.g., the Antler, Sonoma, Sevier, and Laramide orogenies). In contrast, mountain systems such as the Caledonides of Scandinavia, the Variscan Belt of central Europe, and the Alpine Chain of southern Europe appear to have two, three, or possibly more HP/UHP terranes that formed at different times and/or in different places. This paper explores the possibility that the nature of metamorphism depends on the thermal structure of the subducting oceanic lithosphere that preceded collision and on the vergence of the colliding systems through time as the ocean closes. A simple consideration of the architecture of colliding terranes suggests that certain configurations can result in the possible, even likely, re-subduction of crustal terranes ("re-dunks") and hence their re-metamorphism under either HP conditions or UHP conditions.

The Subduction/Eduction Model

The subduction/eduction model (Figs. 1 and 2) is assumed to be the mechanism for forming HP/UHP sialic terranes in the ensuing discussion. HP and UHP conditions can be achieved easily if crust is pushed, pulled, or sheared to depths of ~90–160 km or more in the mantle. Other mechanisms have been proposed for generating eclogite-facies pressures, including two way channel flow, crustal thickening, and tectonic overpressure. Two-way channel-flow models generally assume that material is being subducted and educted at the same time (i.e. the "two-way street of Ernst et al., 1997) and appears to be an attractive mechanism for the relatively unconsolidated material that enters an accretionary prism and undergoes the essentially chaotic deformation that results in mélanges with blueschist-type eclogites (Ernst and Dal Piaz, 1978; Shreve and Cloos, 1986; Beaumont et al., 1999). However, the HP/UHP sialic terranes discussed in this paper retain coherent tectonostratigraphies over large lateral distances, suggesting behavior as a single plate that undergoes "one-way" movement into the mantle (subduction) followed, after a brief stasis, by a "return trip" back toward the surface (eduction) along more or less the same path the plate went down. The Cretaceous Sanbagawa belt of Shikoku, Japan is generally interpreted as an ocean-floor assemblage that should have evolved through the two-way channel flow mechanism, but recent studies suggest that in many ways it appears to have evolved through the rigid-body subduction/eduction mechanism described above in that it is a subhorizontal tectonic slice sandwiched between low-grade units and bounded above by a low-angle normal fault and below by a thrust fault (Ota et al., 2004).

Crustal thickening seems conceivable where the pressures necessary to form HP assemblages (> 1.0 GPa) can be achieved at the base of the continental crust by tectonically thickening it to thicknesses believed to prevail beneath the Tibetan Plateau (~60 km). However, UHP eclogite-facies terranes contain minerals, including coesite and microdiamonds, that require pressures in excess of 2.5 and 3.3–4.0 GPa, respectively, at high temperatures. The crustal thicknesses required to generate these pressures (> 100 km) could be achieved by tripling or quadrupling the thickness of the continental crust, but there is no region on the present-day planet that has such a thick crust, presumably because it is gravitationally unstable and would collapse under its own weight. The concept of tectonic overpressure rests on the assumption that the lower crust and perhaps the upper mantle are strong enough to sustain considerable deviatoric stress for significant periods of time (ca. millions of years). If the deviatoric stress is very high, the mean stress can be significantly higher than the pressure generated by depth alone, perhaps generating eclogite-facies pressures in the lower crust or uppermost mantle. However, it is difficult to imagine that stresses at depth can be maintained over millions of years without causing deformation to relieve the stresses. The problem seems particularly acute for sialic terranes because of the well-known weakness of quartz subjected to deviatoric stress at 350–400°C under hydrous conditions (Scholz, 1988).

The crustal thickening and tectonic overpressure models also do not provide simple mechanisms for transferring fragments of the upper mantle into the crust. Yet many, though admittedly not all, HP/UHP terranes are associated with "orogenic" peridotites that clearly were introduced into the crust from the mantle as solid bodies (Brueckner and Medaris, 2000). The crustal subduction/eduction model, on the other hand, results in a geometrical configuration that makes this transfer plausible and even likely. A mantle wedge must lie above subducted continental crust (Figs. 1 and 2) and there is shear along the boundary between this crust and mantle. It therefore seems probable that fragments can transfer from the dense mantle hanging-wall into the less dense crustal footwall through brittle (i.e. duplex transfer, subduction erosion) or ductile mechanisms (i.e., diapiric downwelling; Brueckner, 1998). It is more difficult to envision how peridotite fragments can be inserted into a crust that is undergoing crustal thickening because the transfer involves moving dense mantle material upward into less dense crust. Inserting peridotite into a crust that sustains enormous deviatoric stress is even harder to imagine since that crust cannot fault or flow without relieving that stress differential. No doubt mechanisms can be postulated that circumvent these obstacles, but they lack the simple elegance of transfer from an overlying dense mantle into an underlying slab of less dense crust.

In addition to the concrete evidence cited above, the subduction of sialic crust into the mantle is almost required by plate tectonic geometry. It is unlikely that two converging crustal slabs will simply smash into each other in a "head-on" collision, because this process either involves lateral shortening and vertical thickening of both crust and the underlying mantle or a delamination mechanism that allows the mantle to sink so that only the crust will thicken. Most mountain systems display a pronounced asymmetry, which suggests that one collisional terrane overrides the other. Some models, for example for the Himalayas, show one crustal slab thrust over the other crustal slab, requiring the thrust surface of the overriding slab to be the boundary between the crust and mantle (i.e., the Moho), which in turn involves decoupling of the crust and the underlying mantle and the roll-back of that mantle. The geometrically simplest way to view collision is that one crustal terrane follows the oceanic lithosphere that preceded it down the subduction zone The pertinent questions that arise from these considerations is how steep, how deep, and how long will crustal slabs be subducted, how rapidly will they return toward the surface, and what will happen to them during subsequent collisions?

Slab Pull

The crustal subduction model appears to contradict the commonly held assumption in plate tectonic theory that sialic continental crust, initially at least, should be too buoyant to subduct because the density of continental crust (~2.7 to 2.8 gm/cm^3,) is significantly less than the density of the upper mantle (~3.3 gm/cm^3). However, as discussed in considerable detail by Cloos (1993), the continental crust is attached to the underlying continental lithosphere and represents a tiny fraction of the total thickness of the subducted plate. For example, a thin crust of the type that occurs at the stretched edges of continents might be 5–10 km thick, whereas the underlying lithosphere could be 100 to 150 km thick, so that the percentage of crust to the total mass of the lithosphere could be as little as 3 to 10%. If the continental lithosphere is cold, as would be the case along the passive margin of a mature ocean basin, it would be denser than the underlying asthenosphere by as much as 0.08 gram/cm^3 (Cloos, 1993), which would give it sufficient negative buoyancy to subduct deeper into the mantle, especially if that lithosphere had been enriched in rift-related basaltic rocks that converted to very dense eclogite as the lithosphere cooled. Buoyancy analyses by Cloos (1993) suggest that sialic crust would have to be greater than 15 km thick in order to reduce the negative buoyancy of a 100 km thick lithosphere to a neutral or positive value, which would in effect stop subduction. The attached crust is essentially carried into the mantle passively and will remain at depth, returning toward the surface only if it decouples from the underlying mantle so that it can slide toward the surface more or less along the same route it came down. Delamination can occur at the crust-mantle boundary (i.e., the Moho) or within the crust—for example, along the boundary between mafic lower crust and the quartz-rich, and hence weaker, upper crust. Even delamination will not necessarily allow the sialic crust to rise buoyantly toward the surface if eclogitization and the extraction of melt resulted in the densification of the upper crust to values close to that of the surrounding mantle. Slabs that do not return toward the surface will, eventually, heat up through a combination of radioactive decay and the conduction of heat from the adjacent mantle, perhaps eventually melting to produce the late- to post-orogenic granites that occur in many mountain systems. Kinetic factors thus play an important role in whether or not a terrane becomes sufficiently eclogitized so that it remains in the mantle and eventually melts or is only partially eclogitized so that a substantial portion of a subducted terrane retains sufficient buoyancy to return toward the surface once it has detached from the underlying lithosphere.

A second, and perhaps more important, consideration for explaining the subduction of continental crust is that the subduction of oceanic lithosphere precedes the subduction of continental lithosphere. Oceanic lithosphere becomes thicker as the mantle moves away from a mid-ocean ridge and cools. The "thermal lithosphere" can achieve a thickness of ~50 km in 30 to 40 m.y. and will reach an equilibrium thickness of 80–100 km after ~70–80 m.y. (Sacks, 1983; Cloos, 1993). Cloos (1993) estimated that oceanic lithosphere as young as 10 million years is sufficiently dense to subduct. Presumably, older and thicker oceanic lithosphere, particularly when capped by eclogitized oceanic crust, will have enough negative buoyancy to drag the adjacent and linked continental lithosphere into the mantle. This mechanism, called "slab pull," is assumed herein to be the primary mechanism for the initiation and maintenance of subduction of sialic crust. If so, the angle that oceanic lithosphere penetrates into the mantle should influence the subduction angle of the trailing continental lithosphere. The subduction angle of oceanic lithosphere can be affected by several variables (i.e. slab-pull, ridge-push, flow in the underlying asthenosphere, flow in the overlying mantle wedge, roll back, etc.). However, if the simplifying assumption is made that slab-pull is the most important (Carlson, 1995; Lithgow-Bertelloni and Richards, 1995), then the angle, depth, and speed that a slab of continental crust subducts should be a function of the thickness of the oceanic lithosphere that preceded it into the mantle.

Subduction Angle, Rate, and Depth

The subduction angle of oceanic lithosphere in modern tectonic settings varies from 30 to nearly 90° (Isaacs and Barazangi, 1977; Uyeda, 1983). The classic study of subduction zones by Uyeda (1983) established that thick, old oceanic lithosphere sinks into the mantle at a steeper angle than young oceanic lithosphere. Hence older (> 50 Ma) oceanic lithosphere has the potential to pull the trailing continental lithosphere into the mantle at a steep angle and to deep levels before breaking off and sinking away (Fig. 1A). Subducted continental crust that follows this steep and deep trajectory would be expected to move rapidly to deep levels, where it would be exposed to UHP metamorphic conditions. The rapid rate of subduction is the result not only of the greater pull exerted by thick oceanic lithosphere, but also by the steep angle of subduction. Upon delamination, the crust would be expected to rise rapidly (Fig. 1B) as well, because the buoyancy force is directed upward more or less parallel to the return path the educting slab takes as it rises through the mantle.

Old, cold lithosphere occurs along the passive margins of mature ocean basins—for example, along both margins of the Atlantic Ocean. The older the ocean basin, the more likely this lithosphere has achieved thermal equilibrium at a thickness of ~100 km. The adjacent passive margin is characterized by continental crust that was thinned when continents rifted apart to form the original ocean. If this reasoning is correct, UHP terranes should mark the final closure of relatively mature ocean basin and the UHP lithologies should be thinned continental margins composed of old crystalline basement rocks and their passive margin cover. The protoliths of the mafic eclogites that form during collision should be old mafic bodies that had formed during the pre-rift evolution of this crust as well as rift-related mafic rocks that intruded during and just before the opening of the ocean basin. Several UHP terranes contain all or most of these characteristics, including Dora Maira (Variscan basement that formed the southern edge of Europe), the Western Gneiss Region (Proterozoic basement rocks that formed the west edge of Baltica), Dabie Shan/Sulu (Paleozoic basement rocks that formed the southern edge of the Yangtze craton), the Kokchetav Massif (Paleozoic basement rocks that formed the southern edge of the Siberian plate), and the Bohemian Massif (Early Paleozoic basement of the northern margin of the Gondwana craton).

Younger, and therefore relatively thinner, oceanic lithosphere will have a shallower angle of subduction, which will pull the adjacent crustal terrane into the mantle at a shallow angle (Fig. 2A), which in turn delays the rate a crustal terrane will be pulled to eclogite-facies depths and which also raises the possibility that "pull" will not be sufficient to bring a crustal slab to UHP depths. The terrrane may achieve depths appropriate only to HP

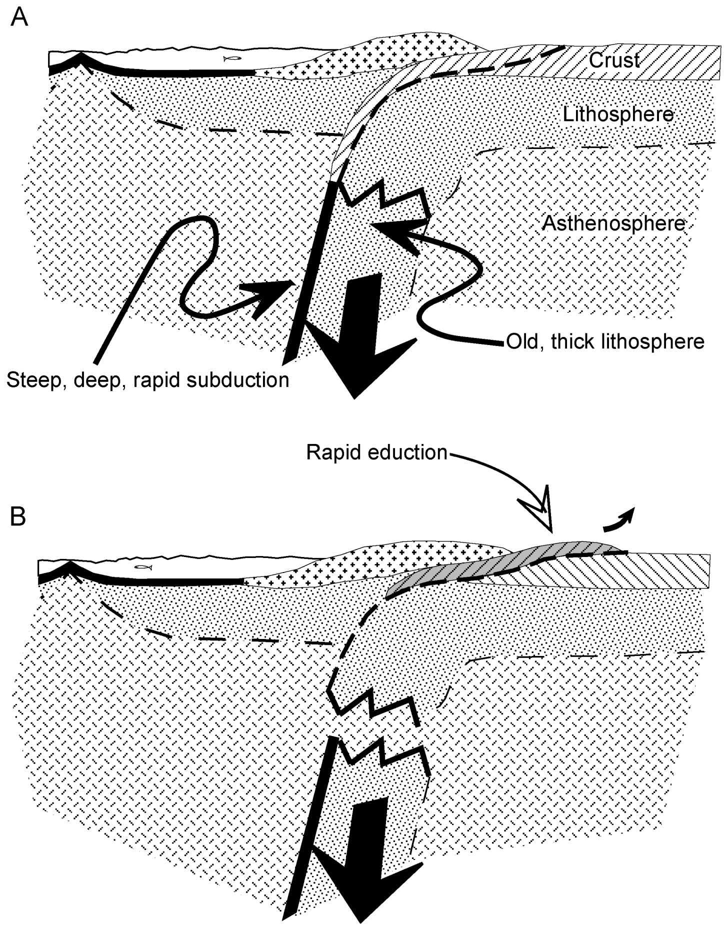

FIG. 1. Steep subduction/eduction model for UHP metamorphism.

metamorphism. The subduction history of this terrane may last longer than a steeply subducted terrane, which could mean a greater spread in "ages" that date the formation of eclogites. Finally, a shallow-dipping delaminated slab of sialic crust may return toward the surface more slowly (Fig. 2B) than it went down because the vertical buoyancy force of the slab will cause it to press up against the overlying mantle wedge, causing the upper plate and lower plate to be more strongly coupled. The slow subduction and eduction of such a terrane would allow heat to penetrate the terrane so that the late stages of the subduction/eduction history may involve HT metamorphism and melting, particularly during decompression. Thin oceanic lithosphere would be characteristic of relatively young and narrow ocean basins (such as today's Red Sea), or near spreading ridges in mature oceans (such as the Juan de Fuca plate west of the Cascades and the Nazca and Pacific plates off the southern and northern Andes of South America). Much of the evolution of the Pan African system has been modeled as the closure of intracontinental basins or relatively narrow and very young, ocean basins. However, the discovery of two Pan African eclogite terranes within this system (Parkinson, et al., 2001; John and Shenk, 2003; John, et al., 2003) suggests that at least some of the basins contained older oceanic lithosphere (i.e., John et al., 2003).

Dunk versus Dunkless Tectonics

It seems likely that some collisions will not result in the deep subduction of a quartz-rich terrane into

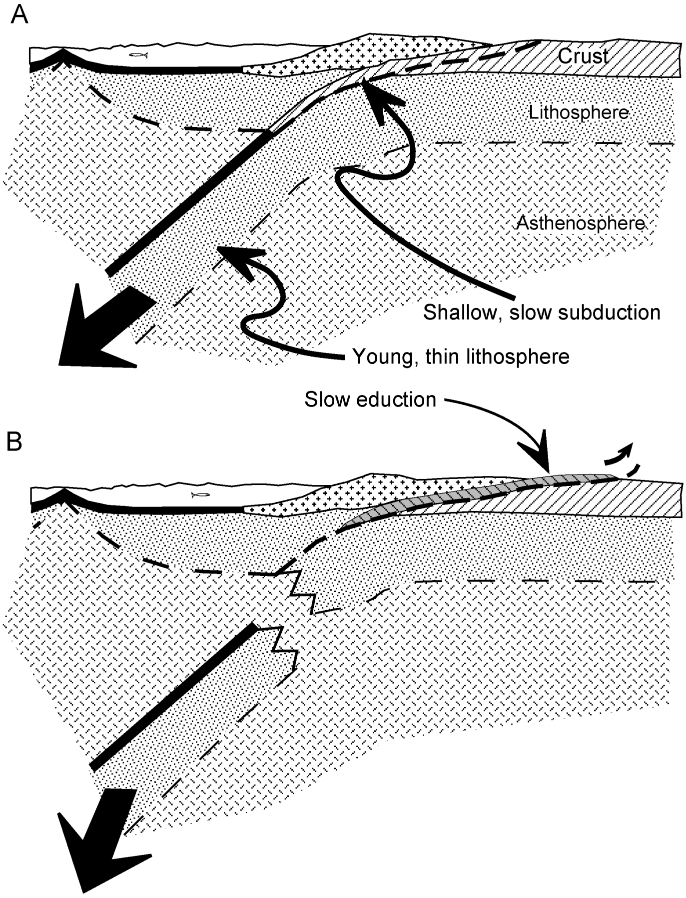

FIG. 2. Shallow subduction model for HP metamorphism.

the mantle. This situation could apply to terranes that begin to follow recently generated oceanic crust down a subduction zone. The high temperature of the mantle near the surface would result in thin oceanic lithosphere that cannot exert enough pull to subduct a trailing sialic crustal terrane. Cloos (1993) has estimated that oceanic lithosphere as young as 10 Ma is dense enough and thick enough to subduct. However, it will not necessarily have enough pull to cause the trailing sialic terrane to subduct or, if it does subduct, to do so slowly at low angles to relatively shallow levels. These terranes are likely to be subjected to Barrovian-style metamorphism because sufficient time could pass to allow the terranes to warm up to thermal gradients of 25–30°C/km, much hotter than those that characterize eclogite-bearing terranes (as low as 5–10°C/km; Liou et al., 2000). If sufficient penetration is generated to juxtapose sialic crust with a thin overlying mantle wedge, it is likely that the wedge will be subjected to a large flux of hydrous fluids released by the dehydrating crust, resulting in the generation of hydrous ultramafic assemblages such as the talc-tremolite and chlorite schists and serpentinites that occur in many Barrovian metamorphic terranes. This model, which emphasizes metamorphism during subduction, compliments the models of Maruyama et al. (1996) and Parkinson et al. (2002), who point out that Barrovian metamorphism can overprint HP/UHP assemblages during exhumation through the introduction of water from bounding low-grade hydrated units.

The extreme case of "non-dunk" tectonics may involve the closure of backarc basins, where the hot mantle above subducting zones has the potential to be further weakened by water rising from the underlying subducting plate leading to a very thin lithosphere (Hyndman et al., 2005). The lack of sufficient pull to subduct a sialic terrane into the mantle would cause the crustal terranes to overthrust each other, resulting in, at most, relatively low grade metamorphism. Most, if not all, of the suspect and exotic terranes in the Cordilleras of North America lack HT eclogite-facies rocks (there are numerous LT/HP terranes). The paucity of high-grade metamorphism can be attributed in large part to oblique convergence with a high component of strike-slip motion, as appears to have happened during the docking events that assembled many of these terranes (Coney et al., 1980). However not all collisions involved oblique convergence or strike-slip juxtaposition. The Antler, Sonoma, and Nevadan orogenies, for example, are generally interpreted as collisions between east-facing arcs (present coordinates) and the western margin of the North American craton. Each collision thrust ocean-floor and arc assemblages eastward over North America, but there is no evidence that the edge of North American was pulled to mantle depths during these events. The width of the ocean basins separating each of these arcs from North America is not known with certainty (cf. Brueckner and Snyder, 1985 and Miller, et al., 1984, for contrasting views on the size of the Havallah basin that closed during the Sonoma Orogeny). The collisions occurred at ~360, 245, and 160 Ma, and the long intervals between these collisions could easily have generated old dense oceanic crust between the arcs and western North America. However, if the eastern Pacific in the late Paleozoic and early Mesozoic was composed of a series of successive marginal basins that opened shortly before collision, similar to those that occur today in parts of the western Pacific, closure of these basins of limited width and floored by thin lithosphere would have resulted in collisions without significant subduction into the mantle.

There is, however, evidence against this "non-dunk" hypothesis. The Moldanubian Gföhl Nappe in the Bohemian Massif is thought to have originated in a marginal basin (Franke, 2000) and yet was metamorphosed to eclogite-facies assemblages and invaded by orogenic lherzolites (Medaris et al., 2006). Although the crustal rocks only reached pressures of 15–20 kbar, prograde metamorphism and exhumation were both very rapid.

Inboard versus Outboard Subduction

The closure of ocean basins as part of the Wilson Cycle will result in multiple collisions because most ocean basins become littered with active and inactive volcanic arcs, composite arcs, microcontinents, oceanic plateaus, etc. as the ocean matures. The nature, number, and vergence of these collisions will vary, but as closure continues the terranes and composite terranes that occur within an oceanic system will tend to accrete to the edges of the bounding continents and, if the ocean basin achieves complete closure, all collisional terranes will end up along a single more or less narrow more or less linear belt (i.e., Urals, Himalayas, Caledonides, etc.).

Inboard subduction

The vergence of the accretionary events at the margins of the bounding continents plays a critical role in the final architecture of the mountain and the metamorphic histories of its terranes. At one extreme is "inboard subduction," where subduction zones dip beneath the continent (Fig. 3A), in which case a volcanic arc will be on the continent, as is the case in the Andes, or subduction zones will dip beneath an arc that has rifted away from a continent by the opening of a marginal basin, as is the case with the Japanese archipelago. Incoming terranes (the ensuing discussion assumes for simplicity that the continent inboard of the subduction zone is fixed) will be forced to subduct beneath the continent (Fig. 3B) or the rifted arc outboard of the continent. As discussed previously, the rate, depth, and angle of subduction will vary as a function of thermal state of the oceanic lithosphere that precedes the incoming crustal terrane into the subduction zone as well as the thermal state of the mantle beneath the crustal terrane that enters the subduction zone. The incoming terranes can be relict arcs, microcontinents, oceanic plateaus, hot spot tracks, mid-ocean ridges, even relatively coherent slabs of accretionary wedges, etc., but only the low-density terranes composed of sialic or quartz-rich crust will delaminate and return buoyantly toward the surface (a notable exception is the Zermatt-Saas ophiolite of the central Alps). Buoyancy-driven exhumation would drive the HP terrane back along the route it went down, backthrusting it over the unsubducted part of the incoming terrane (Fig. 3C). In extreme cases. where the incoming terrane was narrow, backthrusting would place the HP/UHP terrane over the oceanic crust that occurred on the other side (Fig. 6B). The resultant geometry in the latter case would

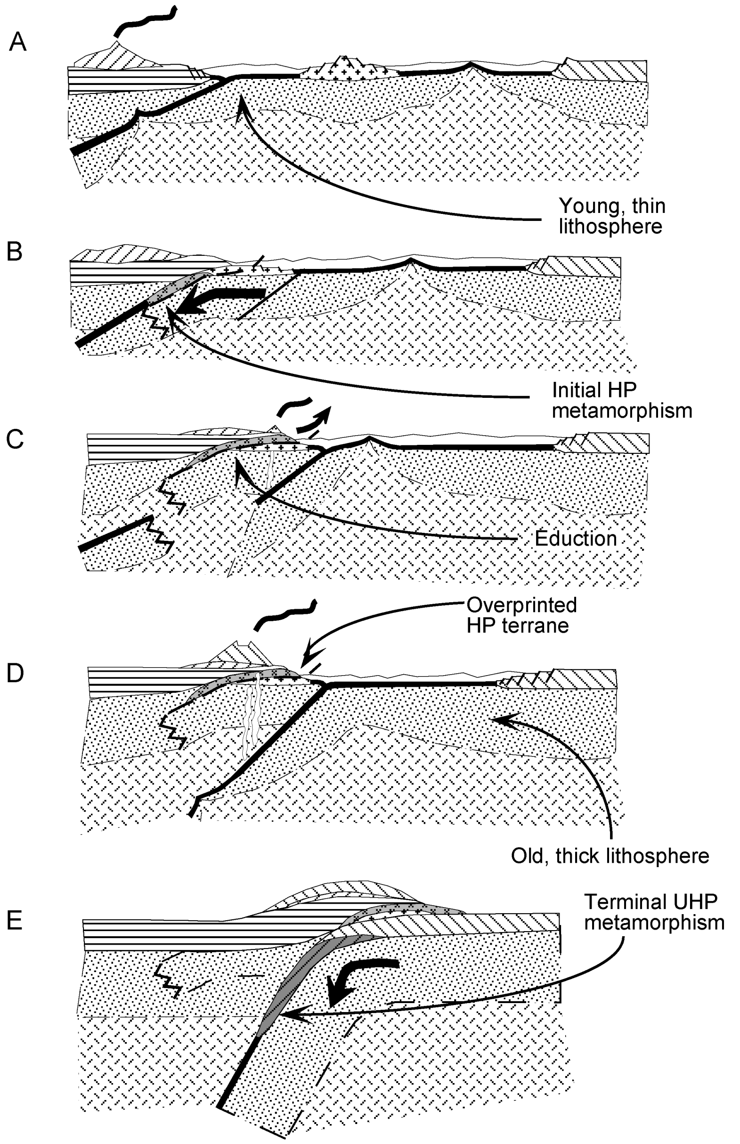

FIG. 3. Sequential collision model for inboard-facing subduction.

be a HP terrane on top of an ophiolite complex. The Austroalpine Sesia-Lanzo HP terrane of the western Alps lies structurally above the Zermatt-Saas ophiolite complex, for example, and could have formed through the subduction of a microcontinent beneath Apulia, followed by exhumation and thrusting the Sesia-Lanzo terrane above the Piemontese oceanic crust (Fig. 6B).

It seems likely that oceanward backthrusting during the exhumation of a sialic terrane will create a geometry that leads to the initiation of a new subduction zone of continental vergence on the other side of the subducted and educted terrane (Fig. 3C). If subduction is sufficiently prolonged, a new arc complex will develop on top of the accreted terrane (Fig. 3D). The advection of heat from the subduction zone upward into the accreted terrane will overprint the overlying HP assemblages through LP/HT metamorphism (Abakuma type). It may be difficult to discern the earlier HP event in such an overprinted terrane.

In addition, subsequent terranes will approach the subduction zone and will subduct beneath the previously accreted terrane (Fig. 3E). If they retain buoyancy despite being metamorphosed under UH/UHP conditions, they too will be backthrust oceanward. If the vergence of the subduction zone continues to be towards the nearby continent throughout the closure of the ocean basin, the result will be terranes that were subducted to HP/UHP conditions only once (single "dunks") and that were likely to be overprinted by LP/HT metamorphism and the intrusion of igneous rocks as a result of their position above subsequently developed subduction zones and underthrust terranes. An example of a HP/UHP terrane that may have undergone this type of evolution is the Moldanubicum of the Bohemian Massif in the Variscan Belt, which was characterized by HT metamorphism and granitic intrusions after HP metamorphism.

Outboard subduction

The other configuration, called "outboard subduction," can result in the re-subduction of terranes to HP/UHP conditions. It occurs when subduction zones dip away from a passive continental margin, which will result in a volcanic arc on the other side of the subduction zone (Fig. 4). Incoming terranes can include active arcs, either island arcs or arcs on microcontinents or continents. In each case, the passive continental margin will subduct beneath the arc (Fig. 4B). The act of subduction will shut off subduction-related melting events, effectively ending the igneous evolution of the arc system (Ernst, 2005). The edge of a continent or microcontinent is likely to be sialic and hence buoyant even if eclogitized, and hence is more likely to return toward the surface or educt than other more mafic terranes. Furthermore, continental margins are adjacent to old oceanic lithosphere: so subduction is likely to be fast, steep, and deep resulting in rapidly subducted and rapidly educted UHP terranes. This configuration is a common one and every mountain system that formed through the Wilson Cycle should have at least one such terrane, particularly at the end because final closure is accomplished by a continent-continent collision where the subducting continent is likely to follow old, thick oceanic lithosphere.

Each cycle of subduction and eduction will result in the exhumed terrane being backthrust over the continent (i.e., inboard thrusting rather than outboard backthrusting; cf. Fig. 4C and 3C). Outboard of the resultant composite terrane is the remnant of the ocean basin (Figs. 4C and 4D). If this oceanic lithosphere continues to develop subduction zones that dip away from the continent, the accreted terrane will develop, with time, a new passive margin. This margin will not be overprinted by a subsequent arc-related thermal event unless a subduction zone develops beneath the continent, as discussed above. If successive subduction events dip away from the continent, the resultant collisions will involve the re-subduction of the continental crust (Figs. 4D and 4E). In principle, re-subduction, or re-dunks, can occur several times (Fig. 4E). However, each event will result in thickened crust and it is unlikely that thick crust can be re-subducted. So sufficient time must pass for erosion, extensional collapse, or some other process to thin the edge of the continent so that it can undergo another cycle of subduction. This thinning may require significant intervals between collisions. In the case of the Scandinavian Caledonides, HP events are separated by ~50 m.y., which may be sufficient to create a thinned margin by erosion. Alternatively, the accreted terranes during subduction and eduction may be wide enough so that the outboard edge remains thin.

The Scandinavian Caledonides

Several of the concepts developed in the preceding sections appear to fit well with the Scandinavian Caledonides, which developed several HP/UHP terranes during their evolution (Brueckner and Van Roermund, 2004). The system evolved during the closure of Iapetus (Fig. 5A) and ended with the culminating collision of a western continent, Laurentia, and an eastern continent, Baltica (present coordinates) at roughly 400 Ma during the Scandian Orogeny (Fig. 5C). This closure involved at least four episodes of eclogite-facies metamorphism, one of which involved western Iapetus and the Laurentian margin (Krogh et al., 1990; Roberts, 2003) and three of which involved eastern (or central) Iapetus and the Baltic Margin (Roberts, 2003; Brueckner and Van Roermund, 2004, 2006). The western collision is generally believed to have occurred by outboard subduction when Laurentia was carried to eclogite-facies (HP and perhaps UHP) depths beneath an incoming arc (Roberts, 2003 and references therein)

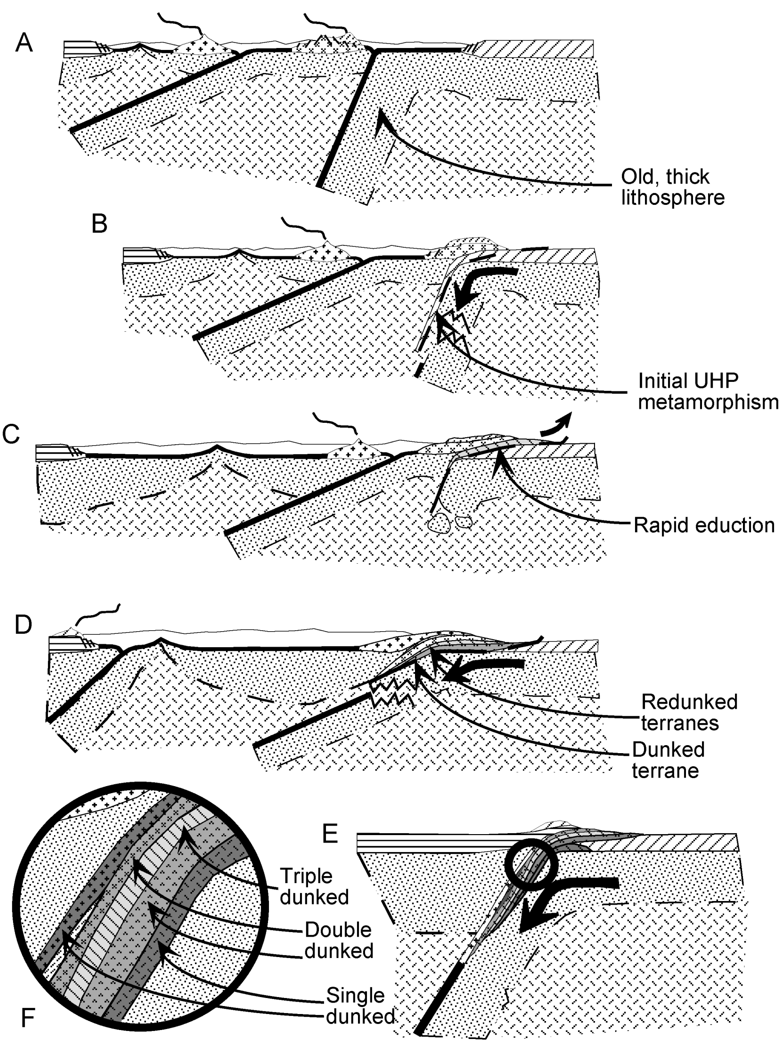

FIG. 4. Sequential collision model for outboard-facing subduction

at roughly 450 Ma (Corfu et al., 2003; Ravna et al., 2006). An alternative inboard subduction model (Fig. 5B), where a microcontinent that had rifted away from Laurentia was subsequently subducted eastward (using present coordinates) into the mantle beneath Laurentia, should also be considered.

The eastern orogenies are called the Finnmarkian (ca. 500 Ma), the Jämtlandian (ca. 450 Ma), and the Scandian (ca. 400 Ma). The Jämtlandian Orogeny at 450 Ma is not yet generally recognized (Brueckner and Van Roermund, 2006) and is noteworthy because it occurred at the same time as the collision between the arc and Laurentia in western Iapetus (Fig. 5B). The vergence for all three eastern orogenies involved outboard subduction where the Baltic margin (or, in the case of the Finnmarkian event, a microcontinent that had rifted away from Baltica; Brueckner and Van Roermund, 2004) was subducted westward beneath an incoming terrane. The first two eastern collisions resulted in HP metamorphism, and hence did not require the subduction of very old oceanic lithosphere, whereas the final Scandian collision was a UHP event that resulted in the formation of coesite and microdiamond-bearing assemblages (Root et al., 2005) and therefore probably involved the "pull" of old, thick oceanic

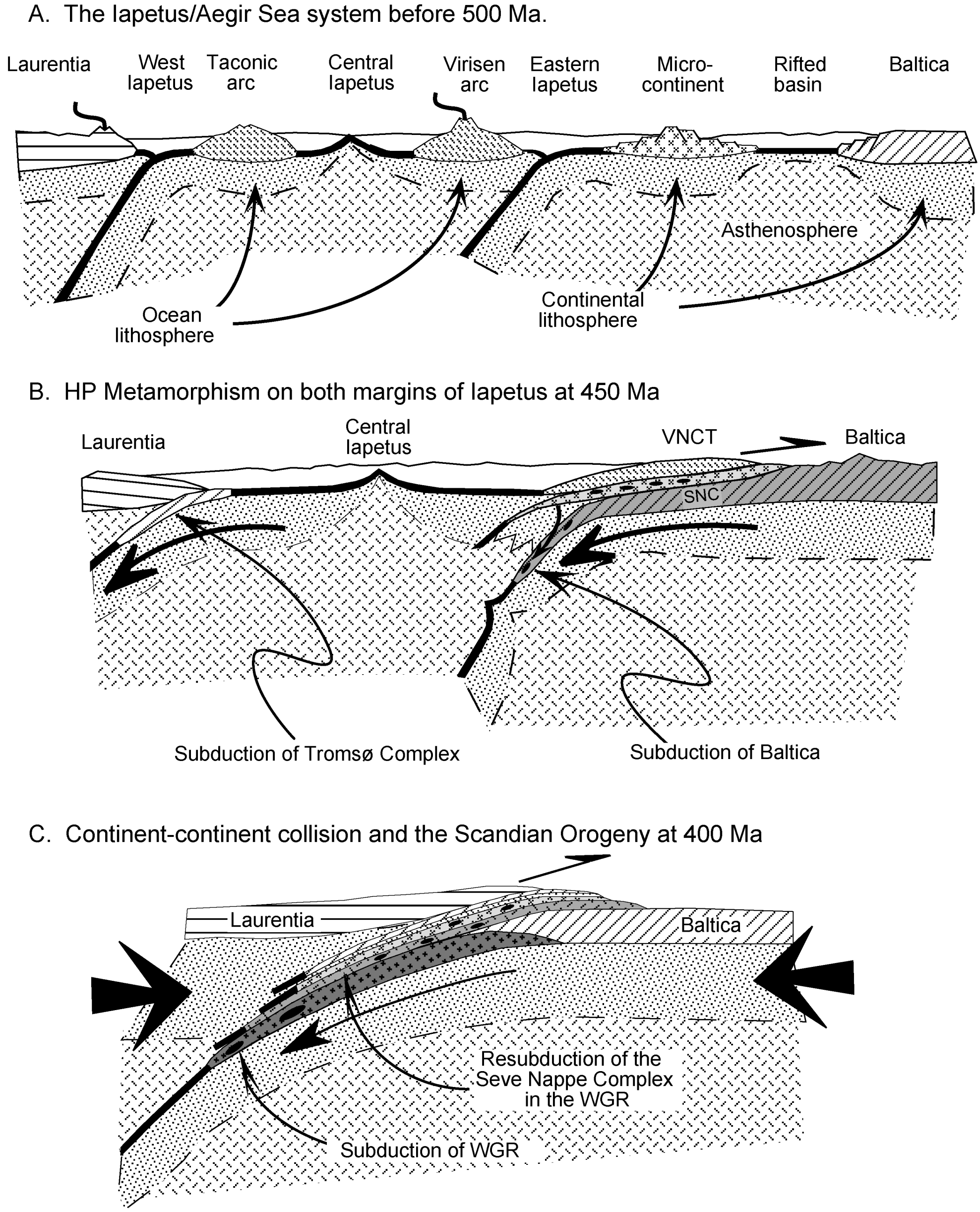

FIG. 5. A multiple collision model for the evolution of the Caledonides of Scandinavia modified from Brueckner and Van Roermund, 2004 and Roberts, 2003 and references therein. VNCT = Virisen/Norrbotten Composite Terrane; SNC = Seve Nappe Complex; WGR = Western Gneiss Region.

lithosphere (alternatively, UHP metamorphism during the Scandian Orogeny involved an intracontinental thrust; see Brueckner, 1999). The interval between the Jämtlandian Orogeny and Scandian Orogeny was approximately 50 million years, which may have been sufficient time to generate thick oceanic lithosphere with enough "pull" to subduct the eastern edge of Baltica to mantle depths where diamond is stable. In addition, the 50 million years between collisions could have allowed erosion to reduce the thickness of the Baltic margin to the relatively thin configuration required for subduction during subsequent collisions.

The final collision, the 400 Ma Scandian Orogeny, involved a Baltic Margin that had already accreted at least two eclogite-facies terranes when it was subducted beneath Laurentia (Fig. 5C). The simple two-dimensional models shown in Figure 5 implies that these terranes would have been re-subducted (re-dunked) during the Scandian Orogeny. However, the Scandinavian Caledonides is a very long mountain belt. The earlier terranes are known to be exposed within the Seve Nappe Complex of central and northern Sweden, whereas the Scandian collision developed HP and UHP assemblages in the Western Gneiss Region (WGR) of Norway, which forms only a small part of the Scandinavian Caledonides and occurs well to the south of the known HP terranes within the Seve Nappe Complex (see Fig. 1 in Brueckner and Van Roermund, 2004). Therefore, the distribution of the two previously formed HP terranes relative to the WGR is somewhat ambiguous. Recent mapping within the WGR has demonstrated that units of the Seve Nappe Complex (as well as the overlying Köli Nappe Complex) were re-subducted during the Scandian Orogeny (Robinson, 1995; Lutro et al., 1997, Terry et al., 2000). These studies have also established that eclogites occur within these re-subducted units as well as within the basement gneisses. The eclogites in the basement gneisses almost certainly formed during the Scandian Orogeny (Root et al., 2004), as would be predicted by the dunk tectonics model if the basement gneisses had been subducted only once (Fig. 5C). However, some of the eclogites that occur within the cover units could also have formed during the Finnmarkian and/or Jämtlandian orogenies. If so, they would comprise, as predicted, "re-dunked" HP/UHP terranes (Brueckner and Van Roermund, 2003). This possibility would explain why dates by a variety of radiometric techniques (see Root et al., 2005) from the eclogites of the WGR extend from ~400 Ma, taken to be the time of UHP metamorphism during the Scandian Orogeny, toward much older ages that approach 450 Ma (the time of the Jämtlandian Orogeny). Some of the older ages had been dismissed as inaccurate ("two-point" Sm-Nd isochrons), but recent high-precision U-Pb work confirms some of the older ages (Krogh et al., 2003). Additional support for both the older ages and the re-dunk model comes from field and textural observations that indicate many of the UHP eclogites of the WGR underwent an earlier HP and a later UHP metamorphism (Carswell et al., 2006).

The Western and Central Alps

The dunk tectonic model appears, at least initially, applicable to the evolution of the western and central Alps, which appears to have involved at least three eclogite-facies metamorphic events. However, a more detailed appraisal suggests that either aspects of the evolution of the Alpine system are still not well understood or, more likely, that the model is too simplistic and requires revision. The three collisions that resulted in the HP/UHP metamorphism are generally considered to be associated with the closure of, from north to south (present coordinates), the Valais Basin, the Piemontese-Ligurian Ocean, and Canavese Basin (Fig. 6A), and the ultimate collision at 35 to 40 Ma of a northern continent (called either Europe or Eurasia) and a southern continent (variously called Apulia, Adria, or Africa). Most recent models for the closure of these basins (Platt, 1987; Escher et al., 1997; Rubatto and Gebauer, 1998, Gebauer, 1999; Froitzheim, 2001; Liati et al. 2003; Meffan-Main, et al., 2004) show the southern edge of Europe as a passive margin throughout the closure of the ocean basins and was involved in orogeny only at the end when it was subducted beneath Adria. The northern edge of Adria, on the other hand, was actively involved in inboard (south-directed) subduction during and between the first two collisions, so that it overrode, successively, the Sesia-Lanzo microcontinent at 65 Ma (Fig. 6B) and the Briançonnais continent at 40 to 44 Ma, forming metamorphic terranes during each event (Fig. 6C). The subduction of the Sesia-Lanzo terrane developed HP assemblages presently exposed in the Dent Blanche and Sesia-Lanzo units of the Austroalpine Zone. Backthrusting of the Sesia-Lanzo terrane may have emplaced it above the Piemontese-Ligurian oceanic crust on the other side of the subducted microcontinent (Fig. 6B). This backthrusting can also be viewed as the beginning of the underthrusting the Piemontese-Ligurian beneath the composite Sesia-Lanzo/Adria terrane, resulting in the metamorphism at 44 Ma of the Piemontese-Ligurian crust to form the Zermatt-Saas UHP terrane (Fig. 6C).

.The subduction of the Piemontese-Ligurian lithosphere to UHP depths should have dragged the Briançonnais microcontinent (presently comprising the Bernhard nappe system) to HP/UHP depths, but

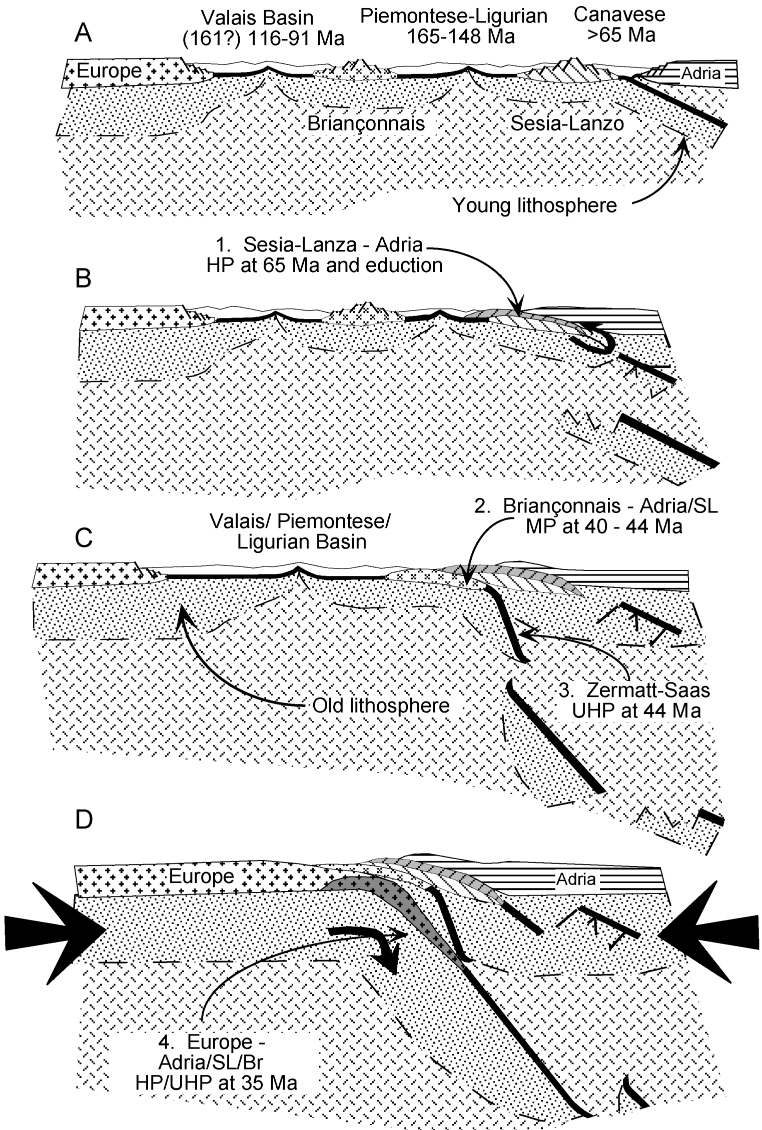

FIG. 6. A multiple collision model for the evolution of the western and central Alps based on models by Platt (1986) Escher et al. (1997), Rubatto and Gebauer (1998), Gebauer, (1999), Froitzheim (2001), Liati et al. (2003), Meffan-Main et al. (2004, and references therein). Abbreviations: SL = Sesia-Lanzo; Br = Briançonnais.

the Bernhard nappe system does not appear to have developed eclogite-facies assemblages unless the Monte Rosa Nappe is considered part of the system (i.e., part of the Briançonnais terrane) as proposed by Escher et al. (1997). One explanation is that the Piemontese-Ligurian broke off from the Briançonnais terrane before it could pull it deeply into the mantle, but that breakoff should also have resulted in the sinking of the Zermatt-Saas terrane to depths where it could not have been subsequently exhumed. It can only be speculated that the Briançonnais terrane was too buoyant to subduct, which in turn prevented the Zermatt-Saas terrane from breaking off and sinking away. The high density of the eclogitized Zermatt-Saas ophiolite creates problems regarding how this ophiolitic terrane was exhumed. One possibility is that it was scooped up by the subsequent subduction and eduction of the southern European margin, a possibility made likely by the subduction to UHP depths of part of the European margin, which could have underthrust the eclogitized ocean floor (Fig. 6D) and brought it back up "piggy-back" style (Meffan-Maine et al., 2004; Wheeler et al., 2005).

According to the models developed earlier in this paper, these earlier collisions should have been followed by subduction of oceanic lithosphere and the development of arc systems on the composite terranes that resulted from each collision. However, there appears to be very little evidence of arc-related igneous activity within any of these terranes (Ernst, 2005). A possible explanation is that the HP/UHP events occurred in relatively rapid succession at 60–70 Ma, 60–40, and 35–45 Ma, which are very short intervals compared to the long intervals between collisions in the Caledonides. The short periods of ~10–20 million years between collisions may not have allowed subduction to proceed long enough to develop conditions for continental arc magmatism. This explanation may also explain why there is little evidence of a later HT overprint for either Sesia-Lanzo or Zermatt-Sass, as would be predicted by the inboard subduction model presented in Figure 3.

The youngest ocean crust dated so far in the Alps is ~93 Ma, or late Cretaceous (Liati et al., 2003), roughly 30 m.y. before the earliest collision. Therefore, most of the oceanic lithosphere that separated the two continents must have been generated prior to the earliest collision. Much older dates of 155 to 163 Ma have been obtained from ophiolites within the western Alps (Antronian ophiolites; Liati et al., 2005) and if these dates are anywhere near the time the Piemontese-Ligurian Ocean was initiated, the ~110–120 Ma between initial opening and the first HP collision allows ample time to develop thick oceanic lithosphere capable of pulling sialic terranes to deep mantle levels. A picture emerges, then, of oceanic systems separating the various continents and microcontinents that were not wide enough to generate arc magmas during subduction, yet were old enough to develop the thick lithospheres necessary to pull the Sesia-Lanzo terrane as well as the edge of Europe deep enough into the mantle to generate eclogite-facies assemblages.

The southern margin of the Eurasian continent is generally considered to have been unaffected by the earlier collisions and acted as a passive margin until the final closure of the Valais Basin at 35 Ma (Fig. 6D), which subducted Europe beneath Adria and metamorphosed the internal Pennine Massifs to HP (Gran Paradiso and, if it is part of Europe as proposed by Froitzheim, 2001, Monta Rosa,) and UHP (Dora Maira) assemblages. The peridotite-bearing HP units of the Adula Nappe (Trescolmen, Alpe Arami, and Cima di Gagnone) are also believed to have been part of the southern European margin, but there is some controversy about the timing of this event. Ages of ~40 Ma (Becker, 1993; Gebauer, 1999) agree well with the ages from the internal massifs. However, older ages (~90 Ma) have also been determined (Becker and Mezger, 1995), which confuse the general picture. An attractive explanation that is consistent with the outboard subuction model presented in Figure 4 is that this terrane was subducted and metamorphosed at ~90 Ma, and then the resultant composite terrane was subducted again during final collision (redunked), but this explanation does not appear to fit with most modern models for the evolution of the Western and Central Alps.

It would seem likely that given the long time between the initiation of the Piemontese-Ligurian oceanic lithosphere and the first subduction event, the lithosphere immediately adjacent to the European passive margin should have been very old and thick, and hence would have pulled Europe to deep levels to form UHP assemblages. However, thus far UHP minerals have been found only in the Dora Maria Massif (Chopin, 1984). It is always possible that UHP relicts will be found in the other Internal Massifs, but another possibility is that the Valais Basin, which separated the Briançonnaise microcontinent (or peninsula) from Europe, also opened up the much older Piemontese-Liguria ocean (Liati et al., 2005) so that the southern European margin was adjacent to both old Piemontese-Lugurian lithosphere and relatively younger Valaisian lithosphere. The resultant slab-pull may therefore have varied along strike so that Europe was pulled at different angles to different depths beneath the oncoming composite Adriatic terrane.

Conclusions

This contribution is an attempt to provide a conceptual framework for understanding the development of multiple HP/UHP terranes during the evolution of mountain systems and to apply this framework to two mountain systems in particular. The framework is based on three premises: (1) there can be several collisional events that result in the subduction of continental (i.e., silica-rich) crust into the mantle during the closure of an ocean basin; (2) the thickness of the ocean lithosphere that precedes continental crust down a subduction zone will determine whether the crust will be metamorphosed under MP, HP, or UHP conditions as well as whether or not the peak assemblages developed during subduction will be preserved during exhumation toward the earth's surface; and (3) a terrane can be metamorphosed more than once, and the nature of that metamorphism (HP versus HT) will be determined by the architecture of the closing oceanic system (inboard versus outboard subduction). Here it is shown that while these models appear to work reasonably well for the Scandinavian Caledonides, they require considerable tweaking and *ad hoc* reasoning to work for the Western and Central Alps. Attempting to apply the models to other mountain systems that probably have undergone similarly complex evolutions (the rest of the Alpine-Himalayan system, the Variscan belt of central Europe, the Dabie/Sulu–Qinling-Qaidam–Kokchetav system of China and Central Asia, the Pan African Belts of Africa and South America, and the relatively recent HP/UHP terranes of New Guinea and the western Islands of Indonesia) will encounter similar difficulties, but my hope is that the attempts to apply these concepts will result in raising questions and opening new avenues of thought for further study. Two important complications should not deter us from realizing that many if not most mountain systems underwent multiple subduction/eduction events resulting in the formation of HP/UHP terranes. The first complication is that HP/UHP terranes occur in different parts of a mountain system with no obvious relationship to each other. The closure of an ocean basin as part of the Wilson Cycle almost requires that collisions will occur in different places at different times. If the cycle runs to completion, these scattered terranes will be collected into a single, more or less linear belt that may appear at first glance to be of deceptive simplicity. The second complication is that similar ages in two HP/UHP terranes do not automatically imply formation as a single unit. HP terranes of identical ages occur in the Caledonides of Scandinavia, yet one formed along the Laurentian margin whereas the other formed along the Baltic margin while Iapetus was still a major ocean basin (Brueckner and Van Roermund, 2006). They were brought into proximity by the collision that closed Iapetus, yet they initially may have formed thousands of kilometers apart.

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