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Exhumation Settings, Part I: Relatively Simple Cases

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Abstract

Exhumation has been recognized as an important problem in tectonics and can be achieved by normal faulting, ductile thinning, and erosion. The exhumation of ultrahigh-pressure (UHP) rocks is the most challenging and least understood aspect of the exhumation problem. In this review, we focus on a variety of geologic settings with the goal of characterizing exhumation rates and maximum depths of exhumation in different settings. In Part I of this series of articles, we summarize exhumation in some relatively simple divergent (rifts and mid-ocean ridges) and convergent regions (subduction and collision zones). Our examples from rifts are from the East African rift, the Galicia margin, and the Cyclades in the Aegean. The Mid-Atlantic Ridge serves as an example of exhumation at a slow-spreading ridge. The examples from subduction settings are the Mariana ocean-ocean subduction zone, and the ocean-continent subduction zones of the Olympic, Franciscan, and Hikurangi complexes. For relatively simple collision belts, we describe the arc-continent collision of Taiwan and the continent-continent collision of the Southern Alps of New Zealand. For these relatively simple cases, we discuss near–end-member scenarios of exhumation settings. These examples may be divided into four classes: (1) rifting; (2) buoyancy-driven diapirism; (3) erosional exhumation associated with the development and maintenance of orogenic topography; and (4) underplating in convergent wedges. In general, it seems to be important that multiple processes cause exhumation. Asthenospheric upwelling associated with rifting appears to be significant for the initial ~100 km of exhumation of UHP peridotite.

Introduction

FIELD GEOLOGY is the geology of exhumed rocks, and is still by far the most important discipline in Earth Sciences. Therefore, exhumation processes play a vital role in our subject and have been recognized as an important problem in tectonics. Exhumation occurs in a variety of tectonic settings. A first broad subdivision can be made between exhumation settings controlled by divergent plate motions (rift zones and mid-ocean ridges) and convergent settings (subduction and collision zones). Ring et al. (1999) showed that each tectonic setting has a maximum exhumation depth. In collision zones, the most spectacular exposure of very dense rocks from depth >185 km are reported (van Roermund and Drury, 1998; van Roermund et al., 2001). The exhumation of these ultrahigh-pressure (UHP) rocks is the most challenging aspect of exhumation (Ernst, 2001, 2006). The fact that UHP rocks are exposed in collision belts implies that exhumation processes in collision zones are more efficient than in other settings. Figure 1 shows that in continental collision zones a large variety of tectonic processes can operate, some of which occur simultaneously. In rifts, only a restricted number of processes appear to operate. Accordingly, the three exhumation processes—normal faulting, near-horizontal ductile flow, and erosion (Ring et al., 1999) (Fig. 2)—interact in different ways in these settings.

This article complements a review of ours, which focused on processes that achieve exhumation (Ring et al., 1999). Other reviews on this topic include Platt (1993) and Coleman and Wang (1995), Ernst and Liou (2004), and Ernst (2001, 2006). Platt (1993) concentrated on concepts and processes of the exhumation of high-pressure rocks, whereas the other reviews mainly dealt with the metamorphism and tectonics of high-pressure and UHP rocks. Herein, the emphasis is on characteristic geologic aspects of exhumation in divergent and convergent settings with the goal of characterizing exhumation rates and maximum depths of exhumation in differ-

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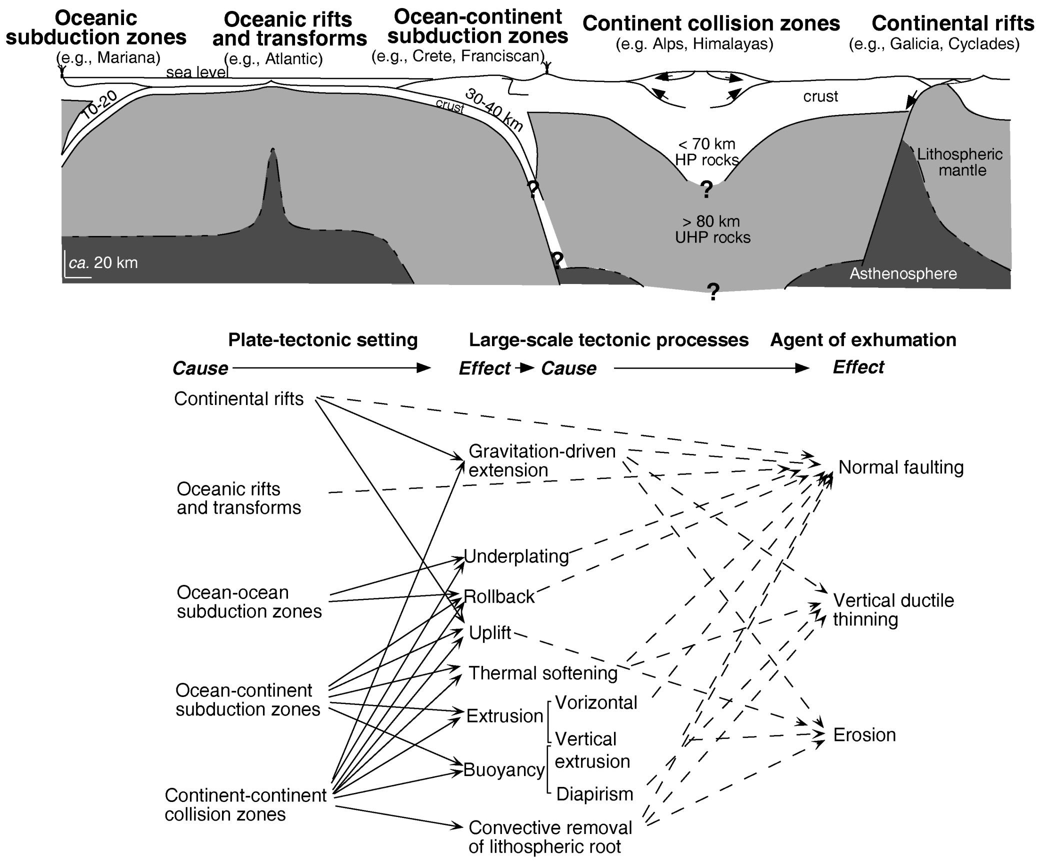

FIG. 1. Interaction of exhumation mechanisms in various plate-tectonic settings.

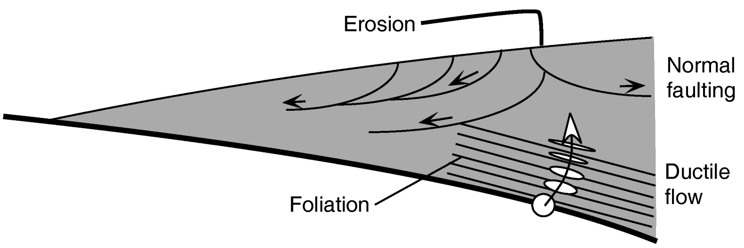

FIG. 2. Schematic illustration of the three exhumation processes: normal faulting, ductile flow, and erosion. Normal faulting covers both brittle normal faulting in the upper crust and normal-sense ductile shear zones in the deeper crust. Ductile thinning refers to wholesale vertical shortening in the wedge. The circle shows an undeformed particle accreted at the base of the wedge, which becomes deformed (indicated by ellipses), along the exhumation path.

ent geologic settings. Our paper distinguishes itself from earlier review articles in the following features: (1) discussion of divergent and convergent settings; (2) inclusion of a broad range of examples drawn from around the world; and (3) emphasis on erosion as an important exhumation process. We deliberately do not pay too much attention to normal faulting because the importance of normal faulting for exhuming deep-seated rocks has been stressed in numerous papers and reviews (e.g. Platt, 1986,

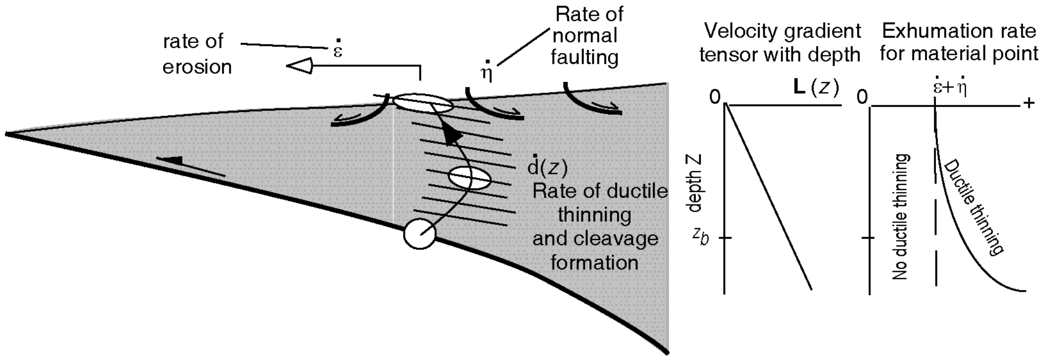

FIG. 3. Exhumation model of Feehan and Brandon (1999) illustrating the relationship between basal accretion, exhumation, ductile flow, and erosion. A particle enters the base of the wedge at z_b unstrained and aquires its finite ductile strain as it rises through the wedge. The vertical or exhumation velocity, Zdot, is divided into two components: (1) erosion rate (εdot) and the rate of brittle normal faulting (ηdot); and (2) the rate of thinning due to ductile flow (ddot). An upward moving particle follows an exhumation path while it aquires ductile strain. The model assumes a steady-state wedge, where accretion, ductile flow, erosion, and normal faulting remain constant.

1993, 2007) and thus is probably the best understood exhumation process.

In Part I of this series of papers, we first examine relatively simple cases of exhumation in divergent and convergent orogens (see Ring et al., 1999, for our usage of orogen) in which the principal setting did not change with time. We look at: (1) divergent zones, i.e., rift zones, mid-ocean ridges, and transform faults; and (2) convergent settings, i.e., ocean-ocean subduction zones, ocean-continent subduction zones, and relatively simple collision belts. For each of the settings, we present relatively simple near–end-member examples, in which the interaction between the exhumation processes appears, at least to a certain degree, resolvable. In the summarizing discussion, we outline models (rifting, buoyancy-driven diapirism, erosion of mountainous topography, underplating in convergent wedges) for the relatively simple settings and discuss the relevance of these settings for the exhumation of UHP rocks. In a forthcoming article—i.e., Part II of this paper series—we outline more problematic non–end-member examples like the Hellenic subduction zone in the Aegean, the Apennines, the Alps, the Himalayas, and the Basin and Range province. The rocks in these orogens underwent a complicated exhumation history and the settings in which exhumation occurred changed in time and space.

Terminology

Collapse

As noted by Willett (1999), the term "collapse" is poorly defined and is probably overused in the literature. Variations of this term include "extensional collapse," "gravitational collapse," "orogenic collapse," and "post-orogenic collapse"—all of which may, or may not, have the same meaning. We avoid the term collapse in this article and use "gravitation-driven extensional deformation" to refer to tectonic lowering of an orogen due to excess topography or a reduction of the yield strength within the orogen. Note that extensional deformation implies horizontal extension of a material line, which may be achieved by normal faulting or by near-horizontal ductile flow.

Vertical ductile thinning

With exhumation by ductile flow, we mean vertical thinning of the ductile crust associated with a subhorizontal foliation. Feehan and Brandon (1999) outlined potential problems of using vertical strains to quantify exhumation. In the simplest case, where exhumation is entirely controlled by ductile thinning, exhumation is given by the average stretch (final length/initial length) in the vertical, because this tells how much the thickness has changed. However, in the general case of exhumation by the interaction of various processes (Fig. 3) it is more difficult to quantify the contribution of ductile thinning to exhumation. In this case, the vertical rate at which a rock moves through its overburden and the rate of thinning of the remaining overburden at each step along the exhumation path have to be considered (Feehan and Brandon, 1999). The general conclusion is that the contribution of ductile thinning to exhumation will always be considerably less than that estimated from the vertical stretch only. Simple one-dimensional calculations show that the contribution of vertical ductile thinning in subduction complexes from western North America and Chile is less than half of the estimated finite vertical shortening in the rock (Feehan and Brandon, 1999; Ring and Brandon, 1999; Ring and Richter, 2004; Richter et al., 2007). The same holds true for the Aegean extensional province (Ring and Kumercis, 2008). It follows that huge vertical strains are needed for vertical ductile thinning to make a significant contribution to exhumation. Note that gravitation-driven extensional deformation can be achieved, at least theoretically, by vertical ductile thinning alone and must not involve normal faulting.

Buoyancy

"Buoyancy" is a concept developed as a result of the Archimedes principle. The term buoyancy refers to body-force-controlled relative upward rise of material due to lower bulk density (ρ) than the surrounding medium, and is restricted to material that can flow in a viscous manner (however, the rising/sinking body can of course be rigid). The basic exhumation process by which buoyancy exhumes rocks is ductile thinning.

The terms buoyancy and diapirism were first introduced to modern geology by Ramberg (1967, 1972, 1980, 1981). The formation of load casts, salt domes, as well as mud and serpentinite diapirs are classic examples of buoyancy-driven movements. Later, buoyancy was discussed as an exhumation agent for eclogite in the Tauern Window of the Eastern Alps by England and Holland (1979). Because eclogite is generally the densest rock in an orogen (ρ ~3600 kg m^{-3}; Christensen and Mooney, 1995) it will sink within a ductile body of lighter material. Exhumation of eclogite requires that this high-density rock is entrained in a huge mass of lighter, constantly upwelling material, which overcomes the negative buoyancy of the eclogite. The most critical factors for eclogite exhumation are the viscosity of the matrix rocks surrounding the eclogite and the exhumation rate of the eclogite-matrix sequence. The lower the viscosity of the matrix, the higher the exhumation rate has to be to hinder dense eclogite from sinking. For the classic case of the Eclogite zone in the Tauern Window, Glodny et al. (2005, 2008) showed that minimum exhumation rates of 36 km Myr^{-1} hindered the sinking of eclogite blocks and facilitated their exhumation.

Settings

One of the major problems of subdividing this review into different settings is that clear boundaries between the settings are hard to define. It is usually difficult to tell whether or not tectonic exhumation was caused by late-collisional gravitation-assisted extension or by extension due to plate divergence (rifting), which usually is also assisted by differences in potential energy. Gravitation-assisted extension must drive contraction on the periphery of the area of high potential energy, either during plate convergence or after plate convergence has ceased. In orogens, especially the older ones, where relative plate motions are imperfectly known, it is difficult to tell whether contractional deformation in foreland regions was still due to plate convergence or occurred after the cessation of the latter and caused by gravity spreading in the internides. It follows that late continental collision and rifting are hard to distinguish (see Henk, 1999, for an example from the mid-European Variscan orogen).

The end of subduction of oceanic crust is not always easy to separate exactly from the onset of subsequent underthrusting of continental crust, i.e., incipient continental collision. In some cases it appears that the leading thinned edge of a colliding continent has been subducted to great depth while the thermal gradient remained $\leq 10°C$ km^{-1} (see Schmid et al., 2004 and O'Brien et al., 2001 for UHP metamorphism in the Adula nappe of the Alps and the Indian continental margin, respectively; Ring et al., 2002, for eclogite-facies metamorphism of the passive margin of East Gondwana in eastern Africa).

We use the following definitions for rifting, subduction, and collision in this paper. Rifting means horizontal extensional deformation caused by divergent plate motions. By subduction we mean the underthrusting of oceanic crust either below oceanic or continental crust. The underthrusting of thinned continental crust underneath an overriding continent during still low thermal gradients is referred to as the subduction/collision transition. This stage seems very important for the formation and exhumation of high-pressure and UHP rocks. Continental collision describes the underthrusting of continental crust underneath continental crust of normal thickness (~40 km according to Christensen and Mooney, 1995). The terms pro-wedge and retro-wedge will be used according to Willett et al. (1993).

Continental Rifts

Continental rifts, like oceanic rifts (see below), are characterized by rather shallow exhumation depths. In a number of cases, exhumation seems to be dominated by slow normal faulting, with slip rates generally <1 km Myr^{-1}. However, in hot intra/back-arc rifts, normal faulting is typically more rapid, with rates exceeding 3 km Myr^{-1}. Erosion of rift flanks is relevant for subaerial continental rifts and appears also to be slow. Ductile thinning only plays a role in relatively young and hot rifts, mainly back-arc rifts and within the mantle in the form of diapirism. The examples we discuss here are the East African rift, the rift/drift transition of the Galicia margin in the Bay of Biscay, and the back-arc rifts of the Cyclades in the Aegean.

East African rift

The Cenozoic East African rift developed in continental crust, which had normal thickness at the beginning of the rifting process. The current crustal thickness in the central Kenya rift thins from ~36 km outside the rift to 30 km beneath the rift valley. To the north, the crustal thickness underneath the rift is 20 km (Maguire et al., 1994).

The rift shoulders of the Cenozoic East African rift expose mainly middle to lower crustal Precambrian gneiss, but those rocks are known to have been dominantly exhumed prior to the onset of Miocene to Recent rifting. This is indicated by mean apatite-fission-track ages of ~65 Ma, ~110 Ma, and ~180 Ma from the rift shoulders (Foster and Gleadow, 1996). Noble et al. (1997) showed that during several episodes of Mesozoic rifting in the Anza rift in Kenya, on which the Cenozoic East African rift is superimposed, regionally extensive erosion of the basement occurred. Miocene to Recent rifting appears to have caused only minor exhumation as indicated by the unreset apatite-fission-track ages in the Kenya rift (Foster and Gleadow, 1996). The unreset ages verify that late Cenozoic exhumation was less than 4–5 km (assuming a geothermal gradient of ~25°C km^{-1} and a closure temperature of 110°C for apatite).

In the northern part of the Malawi sector of the East African rift, Wheeler and Karson (1989) estimated a cumulative vertical throw at the Livingstone border fault of 5–6 km. It is not known how much of this throw occurred during late Cenozoic rifting, which commenced in northern Malawi at ~8.6 Ma (Ebinger et al., 1993). Apatite fission-track ages from the Livingstone escarpment are between 30 Ma and 216 Ma. Modeling of these ages suggests a total erosion of 2.2 ± 0.4 km since the Miocene (van der Beek et al., 1998). Combining the 2.2 ± 0.4 km of erosion with the 8.6 Ma for the onset of rifting yields an average mean erosion rate of ~0.25 km Myr^{-1}. Because the apatite-fission-track ages are older than the beginning of rifting at 8.6 Ma, exhumation since then was between 2.2 km (estimated amount of erosion) and 4–5 km (closure depth for apatite fission tracks). If erosion caused 2.2 ± 0.4 km of this exhumation, the maximum amount of exhumation caused by normal faulting was ~2–3 km. The surprising conclusion is that Cenozoic exhumation at the 60°-dipping Livingstone border fault was not dominated by normal faulting.

Another somewhat extreme example that shows the importance of erosion in exhuming rift mountains are the Rwenzori Mountains of western Uganda. The Rwenzori Mountains form a promontory on the eastern rift shoulder of the Albertine rift just north of the equator, and its highest peak, Margherita, rises to 5119 m. Topography is pronounced, with high peaks and deeply incised glacially formed valleys. The Rwenzori Mountains represent the only uplifted equatorial basement block in Africa that is glaciated and also provides evidence for glaciations in the past (Osmaston, 1989; Osmaston and Harrison, 2005). Ring (2008) discussed that the onset of glaciation in the Rwenzori Mountains in the Middle Pleistocene was important for extreme rift mountain uplift. In the Rwenzoris, equilibrium line altitudes during Pleistocene glaciations were high and glacial processes were limited to the peak regions in the central Rwenzoris. Glacial erosion rates in the peak regions were 1.5–4 km Myr^{-1} and the center of the Rwenzoris expierenced a high rate of rock uplift. Isostatic compensation of material removed by erosion elevated the remaining terrain, transforming it into one with high peaks and deeply incised valleys. We propose that strong glacial erosion and the retreat of the glaciers during interglacial periods caused a removal of loads leading to isostatic rebound, which reduced horizontal stresses promoting normal faulting and enhanced rift-mountain uplift.

Rift/drift transition at the Galicia margin

The Galicia margin off the Spanish/Portuguese coast documents a stage of extreme rifting and the transition from continental to oceanic spreading (rift/drift transition). The Galicia margin consists of a number of tilted blocks resulting from stretching of continental crust just before sea-floor spreading started between Iberia and Newfoundland at ~114 Ma (Boillot and Winterer, 1988). Between the inferred oceanic crust (Grau et al., 1973) of the Iberian abyssal plain and the continental Galicia margin, serpentinized peridotite is largely buried beneath sediments but also crops out locally (Boillot et al., 1980; Perez-Gussinyé et al., 2003; Peron-Pinvidic et al., 2007). Because of the thin crust of oceanic lithosphere, shallow exhumation from depths of ~10 km is sufficient to expose the lithospheric mantle in this setting. According to Boillot et al. (1980), the exhumation of serpentinized peridotite is thought to be the result of serpentinite diapirism and tectonic unroofing of mantle rocks along the rift axis of the margin.

The interpretation of seismic reflection profiles reveals that the Galicia margin is characterized by small blocks bound by oceanward-dipping normal faults (Thommeret et al., 1988; Manatschal, 2004). The fault blocks reach a vertical thickness of 3–4 km and are underlain by a series of apparently undulating bright reflections, which were interpreted to be low-angle detachment faults (de Charpal et al., 1978). The detachment faults are thought to have controlled the final break-up of the continent west of Galicia (Reston et al., 1996; Reston, 2007).

This summary suggests that the continental crust was primarily thinned by normal faults. However, the exhumation of mantle rocks was also aided by diapiric rise of serpentinitic peridotite. Therefore, the major exhumation processes at the Galicia margin appear to have been normal faulting and diapirism. For diapiric rise, the buoyancy caused by serpentinization of mantle peridotite is thought to drive exhumation of lithospheric mantle rocks. Serpentinization can cause a decrease in density from 3300 kg m^{-3} to as low as 2600 kg m^{-3} (–22%; Christensen and Mooney, 1995).

Continental intra/back-arc rifts

Deeper exhumation in continental rift zones occurred for instance in the Cyclades of Greece and the D'Entrecasteaux islands of eastern Papua New Guinea. High-pressure rocks in these two orogens had a complex history, and their exhumation took place in different settings. Based on dating and metamorphism we know that most of their exhumation happened early on compared to the present rift setting of the exhumed rocks (Avigad et al., 1997). In this section, we aim to unravel that part of the exhumation history that occurred when the high-pressure rocks of the Cyclades were in a back- and intra-arc rift setting.

In the Cyclades, blueschist and eclogite from ~50–60 km depth are exposed (Schliestedt et al., 1987; Okrusch and Bröcker, 1990; Will et al., 1998; Schmädicke and Will, 2003). Global Position system (GPS) data (LePichon et al., 1995) demonstrate that the present Cyclades are a true continental rift in the back arc of the Hellenic subduction zone. We know that 30–40 km of the exhumation of the Cycladic high-pressure rocks occurred during cold thermal conditions in the Eocene and Early Oligocene, shortly after the rocks were subducted and accreted (Avigad et al., 1997; Gessner et al., 2001). The oldest normal faults are dated at 42–32 Ma (Raouzaios et al., 1996; Ring et al., 2007), are related to wedge extrusion in a convergent setting and this stage of wedge extrusion accommodated the 30–40 km of early exhumation of the Cycladic blueschist unit. It is important to note here that normal-sense shearing is a geometric effect of the extruding wedge only and not an effect of net extension of the region, which instead was undergoing shortening in the Late Eocene. After this stage of Eocene to Oligocene wedge extrusion, the Cyclades were subjected to a phase of extension-related normal faulting that commenced in the Late Oligocene and Early Miocene and occurred intermittently until the Recent (Lister et al., 1984). We discuss exhumation processes in the intra/back-arc rift setting since the Miocene.

In the Oligocene, the Hellenic subduction zone retreated from a position in the present Cyclades to its Late Oligocene/Early Miocene position underneath Crete (Thomson et al., 1999). The Middle Miocene Cyclades were responding to flow in the asthenosphere as evidenced by the establishment of a magmatic arc at ~14 Ma. According to the flow-line modeling of retreating subduction zones by Garfunkel et al. (1986), the asthenospheric flow in the arc and back-arc region behind retreating slabs is forced to accelerate in the horizontal to fill the free space causing plate divergence (rifting) (Fig. 4). We speculate that this process probably was and still is the dominant cause for extending the upper plate in the Aegean. More importantly, the flow-line model-

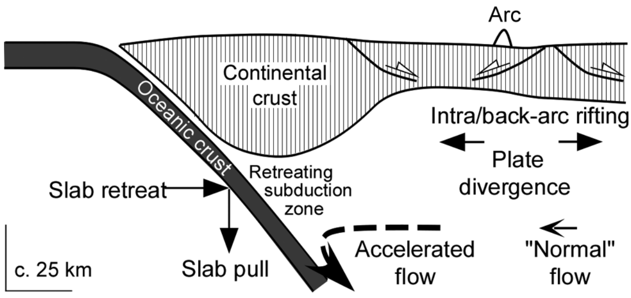

FIG. 4. Simplified sketch showing accelerated corner flow in the asthenosphere behind retreating subduction zones, causing relative plate divergence in the back-arc area.

ing implies a Middle Miocene age for onset of plate divergence in the Cyclades. Arc-related magmatism was coeval with, but also followed extensional deformation (Lister et al., 1984; Avigad et al., 1997). Oligocene and Early Miocene subduction-related volcanic activity has been reported from southernmost Bulgaria and the Greek mainland north of the Aegean Sea, but not from the Cyclades (Fytikas et al., 1984), strongly suggesting that the Cyclades were still in a fore-arc position in the Oligocene and Early Miocene (Ring et al., 2001), and consequently exhumation at this time did not occur in an intra/back-arc setting. Age data for a greenschist/amphibolite–facies metamorphic overprint at 4–8 kbar in the Cycladic blueschist unit cluster around 22–16 Ma (Wijbrans and McDougall, 1988; Bröcker and Franz, 2000; Ring and Layer 2003). This metamorphic overprint occurred at greater depths in the central Cyclades (Naxos and Paros at ~25–35 km depths; Buick and Holland, 1989) than on the islands to the east and west (at ~12–14 depth; Kumerics et al., 2005). This provides an upper limit for maximum exhumation of the Cycladic blueschist unit in an intra/back-arc setting. Fission-track dating on a number of islands in the Cyclades indicates that the Cycladic blueschist unit passed the brittle-ductile transition (as approximated by the zircon fission-track age) during and shortly after the onset of arc-related granite intrusions at 14–12 Ma (Kumerics et al., 2005; Brichau et al., 2006, 2007, 2008). The relationship between arc magmatism and greenschist- to amphibolite-facies metamorphism and a few quantitative estimates indicate more than 20 km of exhumation since the onset of arc magmatism on Naxos/Paros and less than 15 km of exhumation on the neighboring islands (Avigad et al., 1997; Kumerics et al., 2005; Brichau et al., 2007). It follows that most of the exhumation of the Cycladic blueschists occurred prior to back-arc rifting.

Mid-ocean Ridges and Transform Faults

Mid-ocean ridges are generally classified into slow-spreading (<20 km Myr^{-1}) and fast-spreading (>20 km Myr^{-1}) ridges (e.g., Nicolas, 1995). Slow-spreading ridges are magmatically drier than fast-spreading ones, and are characterized by the occurrence of lherzolite instead of harzburgite and by high degrees of lithospheric stretching. As noted by Ranero and Reston (1999), the ends of slow-spreading ridges are characterized by inside corners, which are elevated 0.5–1.5 km above conjugate outside corners. Some inside corners expose deep crustal and mantle rocks and have faults with larger throws and spacings than outside corners. This across-axis asymmetry is usually explained by large-scale detachment faulting at inside corners (e.g., Tucholke and Lin, 1994). Tucholke et al. (1998), Blackman et al. (1998), Nooner et al. (2003), Hopper et al. (2004) and Ildefonse et al. (2007) described large-offset, low-angle (0–30° dip angle) normal faults along the slow-spreading, currently active Mid-Atlantic Ridge. Fault offset is on the order of 30 km (Blackman et al., 1998; Tucholke et al., 1998), indicating that the amount of exhumation caused by detachment faulting is probably ~10 km.

Deep exhumation also occurs along transform boundaries, which is thought to be due to two factors: (1) Oceanic lithosphere of different age is commonly juxtaposed along transform faults, giving rise to high scarps. Along the Mendocino fracture zone in the eastern Pacific, those scarps can be a few kilometers high. At the bottom of these scarps, oceanic mantle might be exposed. (2) Transform faults are a favorable site for serpentinite diapirism and a classic example of this diapirism is the occurrence of ophicalcite (Bernoulli and Weissert, 1985). An ophicalcite is a breccia made up of mafic and ultramafic clasts set in a pelagic carbonate matrix (e.g., Lemoine, 1980). The clasts indicate that the underlying oceanic lithosphere was exposed on the seafloor and thus records ancient examples of exhumed oceanic lithosphere at transform faults.

Ocean-Ocean Subduction Zones

We use the Mariana subduction zone as a type example. Serpentinite diapirism appears to be an important exhumation process in this setting and has been studied quantitatively by Phipps and Ballotti (1992). The Marina subduction zone is the only setting known to us where the role of buoyancy has been studied quantitatively. Hence, it has important implications for all buoyancy-driven models. In general, ocean-ocean subduction zones and associated accretionary wedges are below sea level and, therefore, erosion is not an important exhumation process.

The Mariana convergent margin of the western Pacific is an active island-arc and back-arc basin system beneath which the Pacific plate is being subducted at the eastern edge of the Philippine Sea. The system is virtually non-accretionary. Fryer and Fryer (1987) concluded that some parts of the Mariana fore-arc were at blueschist-facies grade during the early stages of subduction. Large volumes of fluids were released from sediments of the downgoing plate. These fluids exceed the capacity of the fore arc to absorb and react with this volume (Fryer et al., 1995). The fluids caused and are still causing active protrusions of serpentine mud, which form large, in part gigantic serpentinite seamounts (some hundreds to 1950 km^2 in size; Fryer, 1996). Entrained in the seamounts are small clasts of metamorphic rocks and abundant mantle peridotite in mélange-like zones. The clasts are surrounded by serpentinite. Phipps and Ballotti (1992) showed that rising serpentinite mud could transport up to 20 m large blocks of serpentinized peridotite of a density of up to 2600–2700 kg m^{-3} upward against the force of gravity. If the material stops to well upward from within the conduit of the seamount, the blocks will fall back downward into the conduit. Some of the serpentine mud volcanoes have low densities of <2000 kg m^{-3} and are composed of sequences of unconsolidated serpentine mud flows (Phipps and Ballotti, 1992). These low-density seamounts contain clasts with blueschists-facies mineral assemblages from depths of 12–18 km (Fryer, 1996). These blueschist-facies rocks are the first samples from a still active convergent margin and therefore tell us something about ongoing processes of blueschist exhumation in wedges related to oceanic subduction. Nonetheless, it is important to note that the high-pressure blocks are very small in size (up to a few meters) and do not form coherent tracks of high-pressure rocks. For the Mariana case, the exhumation of mantle and high-pressure rocks appears to be entirely driven by the buoyancy caused by serpentinization of mantle peridotite and resulting serpentinite-mud volcanism. Serpentinite protrusion is facilitated by a dense array of faults in the Mariana fore-arc.

Ocean-Continent Subduction Zones

The Miocene to modern Olympic Mountains along the northwestern coast of the coterminous United States are a useful example of a deeply exhumed accretionary wedge that is exposing high-pressure rocks exhumed from ~14 km depths (Brandon et al., 1998). There is no evidence for normal faulting in the Olympics. Ductile flow hindered exhumation because it is associated with a steeply dipping foliation that causes thickening in the vertical (Brandon and Fletcher, 1998). Hence, the Olympic Mountains are a type example for erosional exhumation above an ocean-continent subduction zone.

On the basis of age-elevation trends and paired apatite and zircon fission-track cooling ages, Brandon and Vance (1992) and Brandon et al. (1998) showed that the exhumation rate in the central part of the Olympics remained fairly constant at 0.7–0.8 km Myr^{-1} since the last 14 Myr. Erosion has caused pronounced vertical extension in the most deeply exhumed part of the accretionary wedge, which caused the subvertical foliation (Brandon et al., 1998; Brandon and Fletcher, 1998). Mountainous topography and erosion have been sustained by continued accretion and thickening in the underlying Cascadia accretionary wedge. The rivers that drain the modern Olympic Mountains indicate that most of the eroded sediment is transported into the Pacific Ocean, where it is recycled back into the accretionary wedge. Because of that drainage pattern, the Puget Sound fore-arc basin does not receive sediment eroded from the fore-arc high. The influx of accreted sediments equals the outflux of eroded sediment, indicating that the Olympic segment of the Cascadia margin is close to a topographic steady state (Brandon et al., 1998).

Ring and Brandon (1994, 1999) argued that the Late Cretaceous Franciscan subduction complex in coastal California developed in a fairly similar manner as the Olympic Mountains. However, the blueschists of the Franciscan, which were exhumed from depths of 25–30 km (Ernst, 1993), have a subhorizontal foliation indicating that ductile flow aided exhumation and contributed ~10% or ~3 km (Ring and Brandon, 1999; Bolhar and Ring, 2001). Recently, Unruh et al. (2007) showed that the normal faults in the adjacent fore-arc basin caused ~6 km of exhumation in the early Tertiary. About 60–70% of blueschist exhumation was accomplished by erosion, showing that the Franciscan is another example where high-pressure rocks were dominantly exhumed by erosion of a fore-arc high. Predicted erosion rates of 0.4–0.8 km Myr^{-1} (Ring and Brandon, 1999) are in the same range as those determined for the Olympic Mountains. Even slower exhumation rates of ~0.1 km Ma^{-1} for Franciscan-type blueschist and eclogite were reported by Baldwin and Harrison (1989) from Baja California, Mexico.

The role of normal faulting above ocean-continent subduction zones remains controversial. The Hikurangi subduction zone in the North Island of New Zealand is an example where normal faulting occurs in the fore arc (Pettinga, 1982; Ballance et al., 1982; Walcott, 1979, 1987). The Hikurangi subduction zone accommodates oblique convergence between the Pacific and Australian plates. The subducting Pacific plate dips very shallowly and much of the fore arc is above sea level. Geodetic measurements (Walcott, 1987, 1998), historic earthquakes (Henderson, 1993), and Holocene coseismic uplift along the coast (Wallace et al., 2007) show that the contractional fore arc is presently tectonically active. Local extension is accommodated by listric normal faults (Cashman and Kelsey, 1990). The vertical throw on single normal faults in Pliocene rocks does not exceed 150 m and the magnitude of extension is on the order of 1–5% (Cashman and Kelsey, 1990). Walcott (1987) attributed this small-scale extension in the fore arc to underplating in the subduction zone, whereas Henrys et al. (2007) favor kinking of the subducting slab.

The fore-arc high of the Hellenic subduction zone in Crete is another, but more complicated example where normal faulting is important. We will return to this example in Part II of this paper series.

We have shown that erosion can be a viable syn-orogenic exhumation mechanism in emergent subducting-related accretionary wedges with a well-drained landscape. In accretionary wedges along the west coast of North America, erosion at slow to moderate rates, ranging from ~0.1 to <1 km Myr^{-1}, largely exhumed blueschist-facies rocks. Despite this small rate, the rocks almost perfectly preserved their high-pressure mineralogy. Normal faulting in the rear part of the Hikurangi wedge occurs at even smaller rates of <0.1 km Myr^{-1}.

It should be mentioned that there is an example of UHP rocks from an ocean-continent subduction zone in Sulawesi, Indonesia (Parkinson et al., 1998). Unfortunately aspects of the exhumation of these UHP rocks are not well known. The age of UHP metamorphism of quartzofeldspathic gneiss associated with garnet-bearing peridotite is thought to be Cretaceous (140–115 Ma; Kadarusman and Parkinson, 2000). The peridotite has been interpreted as representing the mantle wedge that formerly overlay the subducting continental quartzofeldspathic rocks. It has been assumed that the UHP rocks did not form in the present setting but are related to a Cretaceous continent collision zone (Kadarusman and Parkinson, 2000).

Relatively Simple Collision Belts

Pronounced mountainous topography typically develops as a result of large-scale underplating of continental crust in collision zones. Late- to post-collisional thermal and mantle processes commonly enhance mean elevation. To a first approximation, collision settings may be divided into mountain ranges characterized by high elevation and also regionally large extent (e.g., Himalayan/Tibetan orogen) and those that are relatively small in extent and characterized by relatively low mean elevation (e.g., Central Range of Taiwan, Southern Alps of New Zealand). In general, exhumation in high-elevation mountain belts appears to be more complicated. We discuss examples from the Apennines, the Alps, and the Himalayan/Tibetan orogen in Part II of this paper series. Here we focus on two relatively simple cases of the arc-continent collision in Taiwan and the continent-continent collision of the Southern Alps of New Zealand. Both examples highlight the importance of erosion for exhuming the metamorphic interior of these orogens.

For the Central Range of Taiwan, Willett et al. (2003), Schaller et al. (2005) and Hebenstreit (2006) showed that average short-term erosion rates of 5 km Myr^{-1} equal the rate of mass accretion. The bulk of the sediment leaving the Central Range was supplied by mass wasting (Hovius et al., 2000). Mass-balance calculations suggest that the Taiwan wedge is in topographic steady state and that erosion accounts for the bulk exhumation of the metamorphic interior. Nevertheless, late-stage normal faulting is currently operating, as suggested by both GPS surveys and field studies (Yu et al., 1995; Crespi et al., 1996). No quantitative data on the throw at the normal faults are available, but the mass balance calculations of Willett et al. (2001) suggests that erosion basically accounts for the exhumation. The Taiwan example is important because it highlights the significance for constraining rates of the various exhumation mechanisms; the existence of normal faults alone does not mean that they contribute to exhumation in any substantial way.

The continent-continent collision zone of the Southern Alps in New Zealand is an impressive example illustrating differential exhumation controlled by rain shadow effects in mountain belts. Exhumation is controlled by erosion of orogenic topography (Kamp et al., 1989). Due to westerly winds bringing in moist air from the Tasman Sea, the western part of the Southern Alps has precipitation rates of about 3,000–10,000 mm yr^{-1}, whereas the eastern side is much dryer with annual rates of 500–1,500 mm. Likewise, erosion rates in the west reach values of ≥10 km Myr^{-1}, whereas in the east they are on the order of 1–2 km Myr^{-1} (Kamp et al., 1989). Most of the erosion in the west is by landslides, which are mainly triggered by earthquakes (Hovius et al., 1997; Larsen et al., 2005; Korup et al., 2006). In accord with the erosion rates, uplift rates in the west may reach ≥10 km Myr^{-1}, whereas in the east they are considerably smaller (Suggate et al., 1979; Bull, 1991). In general, the erosion rates balance uplift rates (C. J. Adams, 1980; J. Adams, 1980; Batt, 2001; Koons et al., 2003) suggesting topographic steady state. The pronounced differences between the western and eastern Southern Alps are also manifested in the very different topography: steep slopes and strongly incised valleys in the west compare with rounded hills and wide valleys in the east. These data indicate that the exhumation pattern must be asymmetric as well. Amphibolite-facies Alpine Schist from depths of about 25 km is exposed in the west (Holm et al., 1989; Grapes and Watanabe, 1995; Vry et al., 2004), whereas at the crest of the Southern Alps and to the east of it, lower greenschist-facies and non-metamorphic rocks are exposed.

The examples from Taiwan and New Zealand demonstrate that erosion can exhume rocks at rates exceeding 10 km Myr^{-1} in some collisional settings. Mass wasting is the primary erosion mechanism. There is no reason that great rates of erosion cannot be sustained over long periods of time as long as uplift continues and the landscape remains well drained. This profoundly contradicts a commonly held opinion that erosion is a slow mechanism that cannot exhume rocks at rates exceeding 1–2 km Myr^{-1} (e.g., Avigad and Garfunkel, 1991).

Discussion and Models for Relatively Simple Cases

In the above-outlined examples, we focused on exhumation in tectonic settings that did not change through time. In some of these relatively simple "single-setting" examples, the rocks were exhumed by one process (i.e., erosion in the case of the Olympic Mountains), whereas in some of these simple examples exhumation was more complex and occurred by multiple processes.

Rifting

Collectively it appears that individual episodes of continental rifting (including back-arc rifts) do not exhume rocks from depths >30–35 km. Relatively shallow exhumation in rift settings appears to be counter-intuitive given that at least some rifts are characterized by very large scale horizontal extension (up to 250 km, or >100% of extension in the Aegean according to McKenzie, 1978) and the formation of metamorphic core complexes. Nonetheless, the examples demonstrate that large-scale horizontal extension achieved by listric or shallowly dipping normal faults is commonly not an efficient exhumation process. In the Cyclades, rifting was accompanied by pronounced magmatism, and exhumation led to the exposure of associated plutons. In the East African rift, rifting is also associated with magmatism, but plutons have not yet been exhumed. A surprising finding is that slow exhumation in the East African rift was, at least in parts of the rift valley, strongly aided by erosion.

Models for continental rifts commonly distinguish passive from active rifts. During passive rifting, continental lithosphere is stretched and thinned and the asthenosphere wells up to fill the free space; i.e., plate divergence leads to asthenospheric rise and the Cyclades serve as an example (Lister et al., 1984). Active rifting requires primary asthenospheric upwelling, which in turn causes regional uplift and magmatism leading to a broad regional culmination or dome. The Kenyan sector of the East African rift is an example (Williams, 1970). In the so-called "Kenya dome," elevation rises to 3–4 km. The topography develops above the upwelling and hot asthenosphere and triggers gravitation-induced extension. In the Kenya rift, this relationship is well documented by a lag time of ~10–15 m.y. between initial volcanism and the onset of faulting (Ring and Betzler, 1995).

The difference in the amount of stretching and exhumation between continental rifts like the Cyclades and the East African rift system might be the thickness of the strong brittle part of the lithosphere. According to the modeling of Lavier et al. (1999), a small thickness should lead to the formation of a core complex and a large thickness to half-graben structures. As rifting proceeds, a half graben may develop with time into a full graben. In the course of continued stretching (rift/drift transition), detachment faults may also develop to accommodate increasing extension (Lister et al., 1986; Lavier and Manatschal, 2006). Flexural rotation due to isostatic readjustment causes footwall uplift, which in subaerial exposed continental rifts, enhances erosion of the rift flanks. Due to the development of pronounced topography, erosion may be important in active rifts. Even in semi-arid regions like East Africa, erosion appears to be important. In well-drained landscapes, erosion may even become the dominant exhumation process.

A possible reason why rifting in the cited examples did not cause deep exhumation might be that the crust was not syn-exhumationally thickened. Rifting processes start to operate after the cessation of convergence and crustal thickening, and it is a common observation that high-pressure rocks are significantly exhumed soon after their metamorphic overprint—i.e., the crust may be largely "unthickened" again before rifting commences. An additional reason might be the thermal structure in rifts. Magmatic activity heats up and weakens the middle and lower crust, which is then likely to flow. In such a setting, crustal detachments root in the highly ductile middle crust and decouple the latter from the lower crust, preventing exhumation of lower crustal rocks.

In the oceanic realm, crustal extension at fast-spreading mid-oceanic ridges commonly exposes the basaltic section of the oceanic crust—i.e., the upper crust. This implies that continuous magma emplacement keeps pace with extension. In slow-spreading ridges, the Moho is locally exhumed, indicating deep exhumation. The supply of magma is thought to be rhythmic, with a periodicity of 1–2 Ma (Nicolas, 1995). In times of little or no magma supply, tectonic extension starts to replace the supply of magma in order to maintain spreading. The magma chamber solidifies and the still hot gabbro becomes stretched and thinned. Continued stretching causes the exhumation of mantle lherzolite along the ridge crest. Normal faults appear to be the dominant agent of exhumation, but serpentinite protrusions may also play, at least locally, an important role.

The relevance of rifting for the exhumation of UHP rocks is that rocks from the asthenosphere are initially exhumed by convection or diaprism within the mantle (Nicolas, 1995). van Roermund and Drury (1998) suggested that ~100 km of exhumation of garnet peridotite from >185 km depth in the Western Gneiss region of Norway was accomplished by asthenospheric upwelling. Likewise, Vissers et al. (1995) ascribed the first ~90 km of exhumation of the Ronda peridotite in the Betic Cordillera of southern Spain from ~150 km depth to Jurassic break-up in the western Mediterranean. In both cases, dense ultramafic rocks were brought to the base of the crust by buoyancy forces associated with upwelling, hot, and thus light material in the asthenosphere. The density contrast between cold and hot peridotite is on the order of ~100 kg m^{-3} (Molnar et al., 1993).

Buoyancy-driven diapirism

Rifting-related asthenospheric upwelling and very significant exhumation of mantle rocks to the base of the crust is the most extreme case of buoyancy-driven exhumation. The formation of oceanic crust during the advanced stages of rifting and hydrothermal processes causing serpentinization of this crust promote serpentinite diapirism in oceanic settings. Serpentinite diapirism in some cases may become a locally important exhumation process. In oceanic rifts and transforms, serpentinite diapirs may emplace mantle peridotite at the bottom of the ocean floor. In oceanic subduction zones, they may locally exhume blueschist-facies rocks. However, it is important to note that the diapirically exhumed high-pressure rocks do not form regionally coherent belts (Platt, 1993).

Ramberg (1972) showed that small density contrasts of ~100 kg m^{-3} are sufficient to make an underlying lower density unit rise with a velocity, which is small, but yet large enough to account for the development of domes and diapirs. Ramberg's experiments showed that the denser overburden of a rising body has to be sufficiently ductile. The rise of a buoyant layer commonly results in elongated anticlines which, after they reach a certain height, split up into rows of individual domes. Such geometries are well illustrated by salt diapirs (Ramberg, 1967). A characteristic and thus distinguishing feature of diapirs are the curtain-like folds restricted to a rather narrow zone along the periphery of the diapir; mature diapirs should additionally have developed a mushroom-like geometry.

Due to the work of Ramberg (1967, 1972), buoyancy-driven diapirism appears to be a well-understood mechanism and might explain the local exhumation of some metamorphic rocks. Gneiss domes and serpentinite protrusions are probably the best examples of this class. However, whereas a gneiss dome perfectly fits the original view of diapirs by Ramberg, protruding serpentinite does not. This is because shallow oceanic crust overlying the serpentinite is brittle. The ascent of the diapirs must have been controlled by fractures in the overburden. Fryer (1992) suggested that the protrusions are related to periods of seismic activity on faults underlying the seamounts, and that materials making up the resultant flows may be derived from the portion of the fault along which movement took place.

Does the diapiric exhumation of blueschist in the Mariana subduction zone have any general relevance for exhumation in subduction zones? In the Franciscan subduction complex, eclogite- and blueschist-facies mafic rocks indeed occur, at least locally, as isolated blocks (the so-called "knockers"), in large part associated with serpentinized peridotite. Pressure estimates indicate exhumation from 50 km and deeper. The Franciscan "knockers" are usually 10–30 m wide and have a typical protolith of oceanic basalt. An actinolite rind indicates that the tectonic blocks resided for some time in a surrounding serpentinite matrix (Coleman and Lanphere, 1971; Moore, 1984). A key issue is whether or not the high-pressure knockers come from the upper or lower plate of the subduction zone. The hanging wall of the subduction zone is, at least in some cases, invoked as a source for these deep rocks (Coleman and Lanphere, 1971; Platt, 1975; Moore, 1984). However, the fact that the knockers were derived from oceanic basalt suggests that eclogite blocks were formed by accretion of mafic crust from the subducting slab into the mantle of the overriding plate. The exhumation of these rocks remains poorly understood, although the observation of exhumation of high-pressure metamorphic rocks in the still active Mariana fore arc suggests that serpentinite diapirs might be involved in exhumation of peridotite, eclogite, and blueschist blocks as well. A limiting factor for diapiric rise of high-density blocks is the rheologic data for serpentinite mud of Phipps and Ballotti (1992). The data suggest that for a density contrast on the order of ~500–800 kg m^{-3}, the upwelling serpentine mud has to be ~>750,000 times larger in volume than that of the entrained blueschist clast. If the density contrast is smaller, this ratio will also be smaller. For the Franciscan example, this implies that a typical knocker must have been entrained in a ~1–20 km^3 large mass or serpentinite. Such huge masses of serpentinite do not surround the knockers in the present Franciscan.

Because the overburden of a buoyantly rising rock unit has generally to be sufficiently ductile, diapirism is probably restricted to deep crustal levels. The consequence is that this process cannot fully exhume rocks. As argued by Platt (1993), buoyancy appears unlikely to explain the exhumation of regionally coherent high-pressure belts, but Platt also noted that buoyant rise may be the most important exhumation mechanism of crustal rocks subducted into the mantle. UHP granitoid rocks typically have densities of ~3000 kg m^{-3} (Ernst et al., 1997) and, therefore, would in general be buoyant with respect to lithospheric mantle and eclogitized mafic lower crust. However, the relatively low temperatures of about 700–900°C recorded by most UHP rocks indicate that the surrounding mantle would be very strong. Ramberg's work makes it hard to envision how UHP granitoids rise diapirically through dense, cold, and brittle lithospheric mantle, and only some sort of fracture- or fault-controlled diapirism appears plausible. To circumvent this problem, Ernst et al. (1997) proposed that UHP terranes ascend as slabs back up the subduction zone, with the mantle wedge as a stress guide.

Eclogite formed in the lower crust and upper mantle is sufficiently weak at temperatures of 700–900°C. Experimental work by Stöckhert and Renner (1998) showed that the strength of jadeite

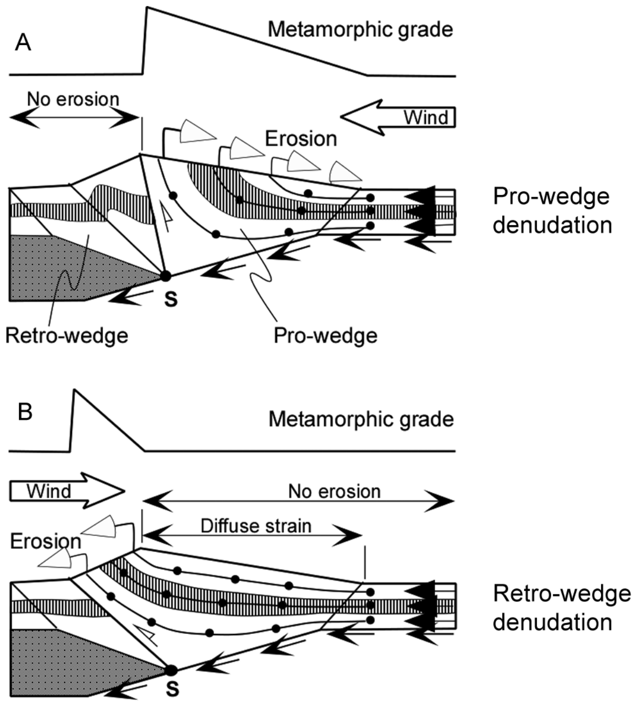

FIG. 5. Schematic illustration of convergent wedges showing the S point and the pro- and retro-wedge (Willett et al., 1993). A. Pro-wedge denudation. B. Retro-wedge denudation. The lines in A and B are material trajectories. Note that depending on the actual array of the flow lines, the base of the wedge in both examples might not be sampled and barometric data from the exposed high-grade rocks only supply a minimum value for wedge thickness.

approaches that of quartz at temperatures of ~600°C, and jadeite might even become weaker than quartz at higher temperatures. The experiments support the inferred low strength of omphacite under natural conditions, as indicated by lattice-preferred orientation of omphacite in deformed eclogite (Phillipot, 1993; Godard and van Roermund, 1995; van der Klauw et al., 1997). Given that the modal abundance of jadeite/omphacite is usually about 50% in eclogite (Ernst et al., 1997), and taking additional quartz and phengite into account, it becomes quite likely that eclogite behaves sufficiently ductile at temperatures >600°C and might aid diapiric rise of underlying high- and UHP granitoid.

Erosion of mountainous topography in convergent mountain belts

The development of mountainous topography above subduction zones and in collision belts makes it possible that erosion becomes an important exhumation process. If the development of mountainous topography is maintained over geologically meaningful periods of time and is associated with a well-drained landscape, erosion may become an efficient and in some cases dominant denudation process. The models of Chapple (1978), Davis et al. (1983), Dahlen (1984), Platt (1986), and Willett et al. (1993) invoke a strong interdependence between accretion/underplating, active deformation, and topographic profile in orogens—i.e., tapering of orogenic wedges. Because erosion denudes mountains, it tends to reduce wedge taper and thus suppresses horizontal extension, especially the development of normal faults. Furthermore, the velocity field in an orogen has to adjust to replace the eroded material, which will in turn result in enhanced rates of contractional deformation and exhumation, creating a self-perpetuating system.

The Central Range of Taiwan has been used by Willett et al. (1993) and Willett (1999) as the type example for steady-state pro-wedge denudation (Fig. 5A). The characteristics of such a wedge are the occurrence of the highest-grade rocks in the central, highly elevated part of the orogen close to the boundary zone between the rapidly denuding pro-wedge and the non-eroding retro-wedge. Because material trajectories in this case are short-circuited (Fig. 5A), residence times in the wedge are relatively short. The Southern Alps is an example for steady-state retro-wedge denudation (Fig. 5B) (Willett et al., 1993). In these wedges, the highest-grade rocks occur at the retro-wedge deformation front, which is dynamically pinned by windward denudation. Because the replacement mass is derived mainly from the pro-wedge, the material must move across the entire width of the wedge, which results in relatively long residence times. Highest elevation does not correlate with the highest metamorphic grade of exposed rock.

Although the Central Range of Taiwan and the Southern Alps are extremely fast orogens, they are not characterized by high elevation (mean elevation is generally less than 2 km in both cases). The absence of high topography in these steady-state wedges is probably due to pronounced erosion. The large erosional and accretionary fluxes cause the fast rates but at the same time control the limited size of the orogen.

One of the major arguments against erosion being a viable exhumation process is that there is not much evidence for eroded material in mountain belts. There are a number of problems with such a statement. England (1981) already showed that sedimentary basins within and adjacent to orogens can usually hold less than about half of the sediment produced by erosional lowering of the orogen. This fact demands that the eroded sediment is being dispersed over distances of thousands of kilometers and into distant ocean basins. In the collisional Himalayan/Tibetan system, for instance, eroded sediment reaches the Java-Sumatra and Makran trenches and is being subducted beneath other orogens.

In the specific case of the subduction-related Franciscan high-pressure rocks, the sediments of the Great Valley fore-arc basin show hardly any evidence for erosional denudation of Franciscan blueschist (cf. Platt, 1986). There are two major reasons, which suggest that the sedimentary record in the Great Valley Group is not very useful for constraining aspects of blueschist exhumation in the subjacent Franciscan: (1) The sedimentary record of erosional denudation of the Franciscan would probably be fairly cryptic because the eroded section would have been dominated by the higher-level rocks of the Franciscan wedge, including the Great Valley Group, the Coast Range ophiolite, and low-grade metasedimentary rocks of the subduction complex (Ring and Brandon, 1994, 2008). (2) Evidence from other subduction complexes, especially the Olympic Mountains, indicates that most of the eroded material is transported back into the trench system and recycled into the accretionary wedge (sediment cannibalism). We believe that sediment recycling and wide sediment dispersal are important constraints for studies that attempt to quantify sediment budgets in convergent settings and to quantify the role of erosional exhumation from the sedimentary record.

Taiwan, the Southern Alps, and the Olympics appear to be relatively straightforward examples in which the metamorphic interior was largely exhumed by erosion of mountainous topography. The maximum depth of exhumation is <30 km in these examples. Nie et al. (1994) argued that the Triassic/Jurassic Songpan-Ganze flysch, which has a volume of $2.2 \cdot 10^6$ km^3, accumulated as a result of erosional denudation of the Dabie Shan UHP rocks. General models for the Triassic/Jurassic climate (Dickens, 1993; Parrish, 1993) show that the Dabie-Quinling orogen in China was part of a region characterized by highest precipitation associated with summer monsoons on the margins of Tethys and the eastern Pacific tropics. The orogenic regions near the coasts are characterized by increased runoff and are in marked contrast to arid continental interiors (Parrish, 1993). The climatic conditions and a likely pronounced topography resulting from extreme thickening make it indeed likely that the Dabie-Quinling orogen was extensively eroding in the Triassic and Jurassic.

It appears that in well-drained landscapes, erosion remains the primary agent of exhumation unless some process weakens the orogen and drives it into an extensional mode. Horizontal extension in orogens is typically initiated by some major change in the geodynamic system, for instance a change in the constitutive behavior of the crust that will lead to thermal (i.e., heating of the middle crust) or rheologic (i.e., development of high pore-fluid pressure or a change in deformation mechanisms at the basal decollement) softening. Such processes would reduce the strength of the crust and could make it weak enough to fail into a new configuration by horizontal extension. The basic process by which thermal softening affects an orogenic wedge is a

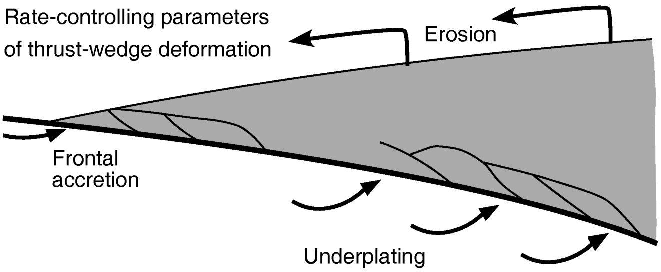

FIG. 6. Distribution of fluxes around an orogen wedge. Accretionary fluxes generally enter the wedge on its pro-side, either by frontal accretion, which tends to reduce wedge taper, or underplating, which tends to increase wedge taper. Erosion preferentially occurs at the upper rear part of a subareal wedge and tends to reduce wedge taper.

reduction in shear stress at the base or within an orogenic wedge. As shown by Davis et al. (1983), a reduced shear stress at the base of a wedge would be able to support a gentler slope than existed before the reduction in strength. Such a reconfiguration of the wedge would result in horizontal extension, either by normal faulting, ductile flow, or a combination of both and is discussed in the next section.

Underplating in convergent wedges (supercritical tapering)

The deformation within a convergent orogen is largely controlled by the distribution of fluxes around the orogen. Accretion and surface erosion control the fluxes into and out of an orogen (Fig. 6). Accretionary fluxes generally enter into an orogen on its pro-side, either by frontal accretion (offscraping) or basal accretion (underplating). A significant erosional flux only occurs when there is subaerial topography, and when that topography has sufficient runoff to develop an integrated river system to carry out eroded material. As a general rule, frontal accretion and erosion both tend to promote horizontal contraction across an orogen. This case is well illustrated by the strain calculations in Dahlen and Suppe (1988).

Tectonic exhumation by horizontal extension may occur in the rear part of an orogenic wedge thickened by underplating (Platt, 1986). We believe it is important to note that horizontal extension must not necessarily be in the form of normal faulting in the upper rear part of the wedge and may also be expressed as vertical ductile thinning or by a combination of the two. Platt (1987) envisioned a general flow field that is either mixed mode or thinning (Fig. 7). Such flow fields preferentially develop when the rate of frontal accretion and erosion at the rear of the wedge are small and the rate of underplating is dominantly controlling the flow field in the wedge (Platt, 1986, 1993; Brandon and Fletcher, 1998). Hence, a diagnostic feature for underplating-related horizontal extension would be a wedge that shows almost no frontal accretion and/or almost no erosion at the top. Normal faulting should be controlled by gradients in topography and thus preferentially directed top-to-the toe of the wedge.

The Altiplano might be a useful example. In the present central Andes, this high plateau in the rear part of the Andean orogenic wedge is almost non-erosive (Lamb and Hoke, 1996). The fore arc is non-accretionary (von Huene and Scholl, 1991). Structural and geophysical evidence is interpreted to show that crustal thickening reaches values of up to 70–80 km below the high Andes (Zandt et al., 1994; Baumont et al., 2002; Gilbert et al., 2006), which, if a steady-state Andean wedge was assumed, would indicate considerable underplating. This would make the Andes a prominent candidate for underplating-controlled normal faulting in the upper rear of the Andean wedge. Normal and strike-slip faulting in the Andes is currently occurring in the Altiplano above mean elevations of ~4 km and apparently balanced by bivergent thrusting at lower elevations (Allmendinger et al., 1990; Marrett et al., 1994; Ege et al., 2007).

Despite the Altiplano, there is other ample field evidence that normal faulting occurs in convergent wedges (see Hikurangi and Taiwan examples above and summary in Platt, 1993, 2007). In order to evaluate the role that normal faults play in the exhumation of the deep interior of a wedge, the throw at the normal faults and the timing of faulting is critical. The aforementioned examples of normal faulting in Taiwan and at the Hikurangi margin show that the

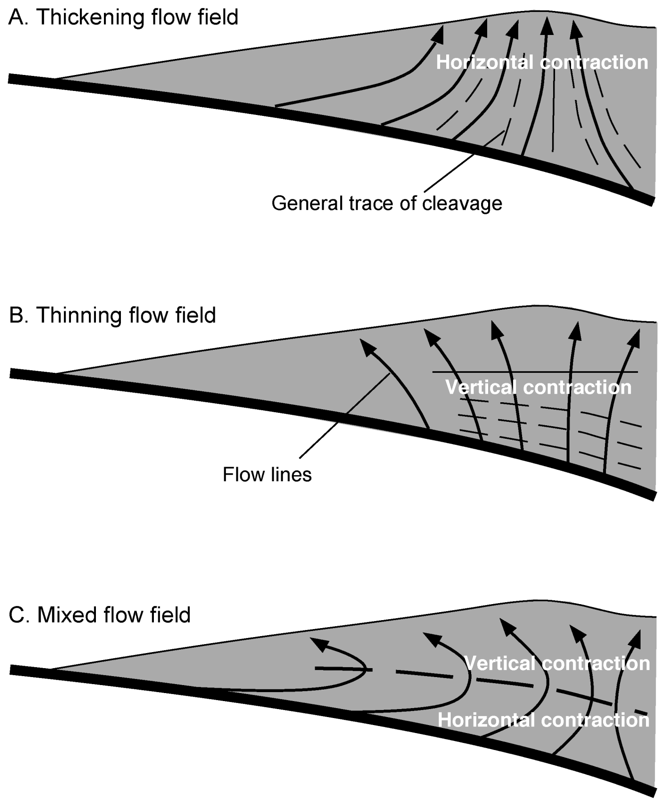

FIG. 7. Schematic illustrations of flow fields for three end-member convergent wedges (modified from Feehan and Brandon, 1999). A. A mixed-mode flow field characterized by horizontal contraction at the base of the wedge and vertical contraction near the top. The dashed line marks the zone where the reversal in strain occurs. B. A thickening flow field is characterized by converging flow lines and widespread horizontal contraction. C. A thinning flow field is characterized by diverging flow lines, which indicates widespread vertical contraction and horizontal extension. Note that horizontal extension can be achieved by either ductile flow or normal faulting.

normal faults there do not cause much exhumation. This shows that the casually held opinion that normal faulting is *a priori* faster and more efficient than erosion is not justifiable.

The Franciscan subduction complex had a thinning flow field controlled by a generally higher rate of underplating than erosion (Brandon and Fletcher, 1998; Ring and Brandon, 1999). Therefore, the Franciscan would be another good candidate for underplating-related horizontal extension. However, vertical contraction in the rear part of the Franciscan wedge is counteracted by pervasive mass loss, and limited horizontal extension occurs (Unruh et al., 2007). This example of pervasive mass loss in accretionary wedges shows that horizontal and vertical strains are not necessarily linked quantities.

Another critical assumption in wedge models appears to be the rheology of the material that makes up the wedge. For shallow wedges thinner than about 15–20 km, a plastic or Coulomb rheology is commonly envisioned (Davis et al., 1983; Dahlen, 1984; Dahlen and Suppe, 1988)—i.e., the wedge essentially deforms by brittle fracture. Dahlen (1984) and Willett (1999) showed that underplating alone is hardly capable of driving a convergent Coulomb wedge from a thrust-dominated taper to a taper where normal faults become active. Pavlis and Bruhn (1984) and Platt (1986) argued that wedges thicker than about 15 km should be dominated by thermally activated deformation mechanisms at the base of the wedge. Platt (1986), therefore, envisioned a linear-viscous wedge rheology—i.e., that solution-mass-transfer deformation or pressure solution is the dominant deformation mechanism in the wedge. Numerical modeling by Willett (1999) showed that horizontal extension in convergent orogens is possible in linear-viscous wedges, whereas in wedges controlled by power-law viscous materials, extension is suppressed.

An important conclusion of the work by Willett et al. (1993) and Willett (1999) appears to be that horizontal extension only occurs, or is strongly aided, when the material behaves ductile in very shallow levels of the wedge. The case study on the thermal evolution of parts of the Alboran Sea basin by Platt et al. (1998) is an impressive example, which implies a very weak middle crust and a shallow (i.e., <10 km) brittle/ductile transition. The heat input for thermally weakening the middle crust may come from: (1) pronounced magmatic activity; (2) conductive heating from the mantle, especially when parts of the mantle were delaminated or convectively removed; (3) radiogenic heat production of crustal rocks; or (4) rapid exhumation of lower crustal rocks, which advects heat from below.

A variant of thermal softening would be rheologic softening. Davis et al. (1983), Pavlis and Bruhn (1984), Platt (1986), and Willett et al. (1993) have argued that an orogenic wedge may be thrown into a supercritical configuration when its basal decollement becomes very weak. A weak base of an orogen might result from rheologic changes, such as a change from brittle deformation mechanisms to thermally activated deformation mechanisms at the base of accretionary wedges due to temperature increase caused by an increased depth of burial. As a consequence of a viscous base of an orogen, a low-angle plateau will form over the interior of the orogen. With increasing viscosity, the angle of taper will decrease and, if the rate of heating is too high for deformation (vertical ductile thinning in this case) to adjust the taper, normal faulting will occur in order to regain critical taper (Platt, 1986; Willett et al., 1993).

Final Remarks

We believe that the most important conclusion of our review is that erosion is, at least in some settings, a very important agent of exhumation. The discussed examples from the ocean-continent subduction zones of the Olympics and Hikurangi complexes provide ample evidence that erosion of an emergent fore-arc high is capable of exhuming pristine blueschist-facies rocks. The Paleozoic coastal accretionary wedge of Chile would be another example that demonstrates the importance of erosion as an exhumation process (Richter et al., 2007; Willner et al., 2007). The collision belts of Taiwan and the Southern Alps of New Zealand are superb examples demonstrating the importance of erosion in exhuming the high-grade interior of mountain belts. Interestingly, it seems that erosion was also important for rift flank uplift and exhumation in the East African rift.

Erosion appears to be a viable and efficient exhumation agent that is associated with the development and maintenance of orogenic topography. An essential feature is that the orogenic landscape is well drained. It is widely accepted that there is a strongly coupled feedback between erosional processes and tectonic forcing resulting in a distinct connection between slope morphology, relief, and erosion rates, which Ahnert (1970) used to formulate a functional relationship between topography and rates of erosion for temperate regions. Montgomery and Brandon (2002) elaborated on Ahnert's work and showed that there is a strongly non-linear power-law relation between mean local relief and erosion rate. This emphasizes the importance of erosion as an exhumation mechanism in well-drained, high-relief mountain chains.

Low-angle normal faulting is only a viable exhumation mechanism if associated with large-magnitude displacement. This is a simple and straightforward conclusion that results from the low-angle nature of faulting. The discussed quantitative data from the Cyclades in the Aegean show that relationship and the well-studied low-angle normal faults in the Basin and Range underpin this statement (Wernicke, 1981, Foster and John, 1999). However, in oceanic settings low-angle normal faulting exhumes the mantle, which, due to the thin crust in this setting, is not associated with large-magnitude displacement. What is commonly lacking are quantitative data that constrain the displacement and the amount of exhumation cuased by normal faulting. Buoyancy-driven asthenospheric upwelling in rift setting appears to be an extremely important exhumation agent for the initial stages of ultrahigh-pressure peridotite and may account for as much as 100 km of exhumation.

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