Exhumation processes

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Abstract: Deep-seated metamorphic rocks are commonly found in the interior of many divergent and convergent orogens. Plate tectonics can account for high-pressure metamorphism by subduction and crustal thickening, but the return of these metamorphosed crustal rocks back to the surface is a more complicated problem. In particular, we seek to know how various processes, such as normal faulting, ductile thinning, and erosion, contribute to the exhumation of metamorphic rocks, and what evidence can be used to distinguish between these different exhumation processes.

In this paper, we provide a selective overview of the issues associated with the exhumation problem. We start with a discussion of the terms *exhumation*, *denudation* and *erosion*, and follow with a summary of relevant tectonic parameters. Then, we review the characteristics of exhumation in different tectonic settings. For instance, continental rifts, such as the severely extended Basin-and-Range province, appear to exhume only middle and upper crustal rocks, whereas continental collision zones expose rocks from 125 km and greater. Mantle rocks are locally exhumed in oceanic rifts and transform zones, probably due to the relatively thin crust associated with oceanic lithosphere.

Another topic is the use of P–T–t data to distinguish between different exhumation processes. We conclude that this approach is generally not very diagnostic since erosion and normal faulting show the same range of exhumation rates, reaching maximum rates of >5–10 km Ma^{-1} for both processes. In contrast, ductile thinning appears to operate at significantly slower rates. The pattern of cooling ages can be used to distinguish between different exhumation processes. Normal faulting generally shows an asymmetric distribution of cooling ages, with an abrupt discontinuity at the causative fault, whereas erosional exhumation is typically characterized by a smoothly varying cooling-age pattern with few to no structural breaks.

Last, we consider the challenging problem of ultrahigh-pressure crustal rocks, which indicate metamorphism at depths greater than 100–125 km. Understanding the exhumation of these rocks requires that we first know where and how they were formed. One explanation is that metamorphism occurred within a thickened crustal root, but it does seem unlikely that the crust, including an eclogitized mafic lower crust, could get much thicker than *c.* 110 km while maintaining a reasonable Moho depth (<70 km, assuming that the seismically defined Moho would be observed to lie above the eclogitized lower crust). Diamond-bearing crustal rocks cannot be explained by this scheme. The alternative is to accrete the upper 10–40 km of lithospheric mantle into the orogenic root. This scenario will provide sufficient pressures for both coesite- and diamond-bearing eclogite-facies metamorphism, while maintaining a reasonable Moho depth (<70 km) and reasonable mean topography (≤3 km). We speculate that the detachment and foundering of the mantle root may contribute to the exhumation of any crustal rocks contained within the mantle root.

Exhumation is a generic term describing the return of once deep-seated metamorphic rocks to the Earth's surface. Field geology is, by definition, the geology of exhumed rocks. In fact, most of our understanding of crustal deformation and metamorphism is based on studies of exhumed rocks. Exhumation occurs by three processes: normal faulting, ductile thinning, and erosion. These processes are important, not only for the exhumation that they cause, but also for their influence on the formation of orogenic topography and the contribution to production of synorogenic sediments.

Exhumation can occur in virtually any geological setting, regardless of age or tectonic

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regime. Even in the earliest studies of alpine tectonics, erosion was recognized as an important process for unroofing the internal metamorphic zones of convergent mountain belts. Early geologists observed that mountainous regions eroded faster than adjacent lowlands, and that ancient mountain belts were commonly flanked by thick synorogenic deposits that could be traced by provenance to erosional sources within the orogen.

The term 'tectonic denudation' (Moores *et al.* 1968; Armstrong 1972) made its way into the literature with the discovery of metamorphic core complexes in the Basin-and-Range province of western United States. Early workers recognised that normal faulting (Fig. 1) was capable of unroofing mid-crustal rocks, and that the hallmark of this type of exhumation was the 'resetting' of footwall rocks to a common isotopic age. In fact, we now understand that the common isotopic age is caused by rapid cooling as the hanging wall is stripped away. This observation has lead to the widely held view that rapid cooling is a diagnostic feature of tectonic exhumation. Work over the last ten years has demonstrated that exhumation by normal faulting often occurs in convergent as well as divergent orogens.

A third exhumation process is ductile thinning (Fig. 1), which can contribute to unroofing of metamorphic rocks. This idea was at the centre of the debate about diapiric emplacement of migmatites and gneiss domes (Ramberg 1967, 1972, 1980, 1981). In this sense, diapiric emplacement of a pluton can also be viewed as a type of exhumation, given that the pluton is 'exhumed' by thinning of its cover. The role of ductile thinning has received less attention than other exhumation mechanisms, but it appears to be important in some cases. For example, there has been much debate recently about the possibility of buoyant rise of high-pressure and ultra-high-pressure quartzofeldspathic rocks, an idea that has close similarity to the diapiric model for gneiss dome emplacement (Calvert *et al.* this volume).

Our objective is to provide a selective review of the exhumation problem. We focus on five topics: (1) a review of the terminology used to discuss exhumation and its relationship to orogenesis, (2) identification of tectonic parameters relevant to the exhumation processes, (3) a summary of how exhumation varies as a function of tectonic setting, (4) the critical review of evidence that might be diagnostic of specific exhumation processes, and (5) a discussion of the origin and exhumation of ultra-high-pressure metamorphic rocks, which represent a particularly challenging example of deep exhumation.

Terminology

The exhumation problem is surrounded by a confusing and inconsistent terminology, which can leave even simple concepts, such as uplift (England & Molnar 1990) and extension (Wheeler & Butler 1994; Butler & Freeman 1996), difficult to follow. In this section, we examine the terminology and provide some simple definitions and suggestions for consistent usage.

The term *orogen* has broadened over the years, and now is commonly used to refer to any mountainous topography at the Earth's surface resulting from localized deformation. This usage includes convergent orogens like the European Alps and the Cascadia accretionary wedge of northwestern North America, and divergent orogens like the Basin-and-Range province and

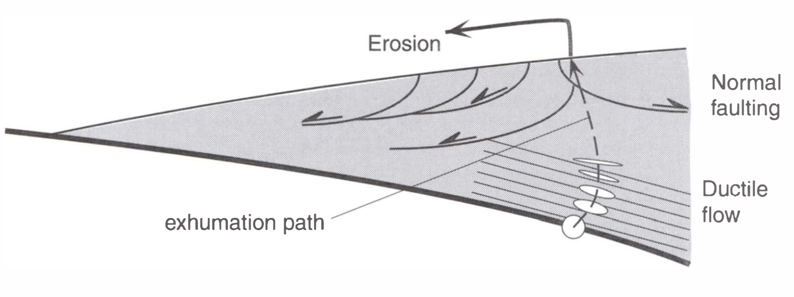

Fig. 1. Schematic illustration of the three exhumation processes: normal faulting, ductile flow, and erosion. Normal faulting covers both brittle normal faulting in the upper crust and normal-sense ductile shear zones in the deeper crust. Ductile thinning refers to wholesale vertical shortening in the orogenic wedge. The circle shows an undeformed particle accreted at the base of the wedge, which becomes deformed (indicated by ellipses) along the exhumation path.

the East African rift. Orogens are commonly described as *compressional* and *extensional*, which emphasises the horizontal stress regime within the orogen. Unfortunately, these terms are not complementary, given that compression refers to stress and extension to strain. More importantly, the simple reference to a dominant horizontal strain ignores the fact that some orogens are characterized by coeval horizontal extension and horizontal contraction of different levels within the same orogen (mixed-mode flow field in Fig. 2). Such mixed-mode deformation fields have been recognized in a number of convergent orogens, such as the Apennines of Italy (Royden 1993), the Betic Cordillera of Spain (Platt & Vissers 1989, Vissers *et al.* 1995), and the Himalayas (Burchfiel & Royden 1985).

Plate boundary terminology provides a better classification of an orogen. The terms *convergent*, *divergent* and *translational* refer to the kinematic relationship of the plates that bound the orogen, and carry no implications about the dominant state of stress or strain within the orogen. For example, one can describe the Apennines as a convergent orogen, even though the deformation field is mixed with some parts of

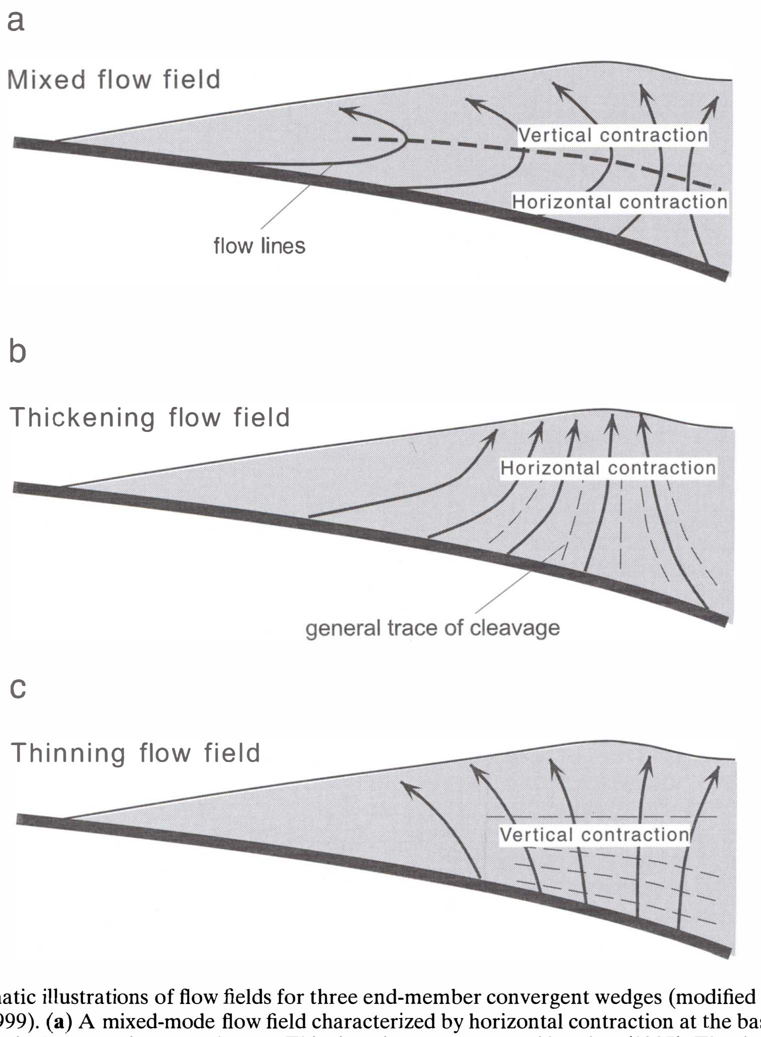

Fig. 2. Schematic illustrations of flow fields for three end-member convergent wedges (modified from Feehan & Brandon 1999). (**a**) A mixed-mode flow field characterized by horizontal contraction at the base of the wedge and vertical contraction near the top. This situation was proposed by Platt (1987). The dashed line marks the level where the principle strain-rate directions reverse. (**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.

the orogen showing horizontal contraction and others, horizontal extension.

Horizontal extension is commonly taken as evidence of tectonic exhumation. The implicit assumption is that horizontal extension indicates vertical shortening, which in turn indicates tectonic thinning of the overburden. There are two problems with this linkage of horizontal extension to tectonic exhumation. The first is that the strain geometry is not always as assumed. For instance, broad shear zones at translational plate boundaries are characterized by plane strain with the maximum extension and maximum shortening directions lying in the horizontal, and little to no strain in the vertical. The result is little to no vertical thinning or thickening of the overburden. In another example, Ring & Brandon (this volume) show that the Franciscan subduction complex was shortened by about 30% in the vertical, but that shortening was balanced by a nearly equal volume strain. The result was that the other principal strains were nearly zero. This example shows that vertical thinning can occur without any appreciable horizontal extension.

This ambiguity can be avoided by focusing on the vertical strain – either brittle or ductile – when assessing the role of tectonic exhumation. In this regard, the terms *thickening* and *thinning* are unambiguous descriptions of the change in overburden thickness caused by vertical strain. Vertical thickening causes tectonic burial and vertical thinning, tectonic exhumation.

There has been some discussion about the suitability of *exhumation* as a term for unroofing of rocks. In a strict sense, *exhumation* means 'to dig up or disinter' (Summerfield & Brown 1998). As such, it refers to the motion between a rock relative to the Earth's surface (England & Molnar 1990). An alternative term is *denudation*, which has a general meaning 'to make bare' and a geological meaning 'to expose rock strata by erosion'. Denudation and erosion are, in fact, widely used synonyms, although there is a tendency to restrict denudation to mean erosion at the regional scale (Summerfield & Brown 1998). The use of denudation in a more generic sense has clear precedent given the term *tectonic denudation,* which means to unroof by normal faulting (Moores *et al.* 1968; Armstrong 1972).

In our review of the literature, *exhumation* and *denudation* are used almost interchangeably, although a subtle distinction is sometimes made about the frame of reference. Exhumation is used to refer to the unroofing history of a rock, defined by the vertical distance transversed by the rock relative to the Earth's surface. On the other hand, denudation, and erosion as well, are frequently used to refer to the removal of material at a particular point on the Earth's surface. This distinction is not universally accepted, but we think it is useful and encourage its adoption. Walcott (1998) emphasized this usage in his synthesis of the Southern Alps orogen of New Zealand. The Southern Alps are characterized by a large gradient in surface erosion rates across the orogen, from *c.* 1 km Ma^{-1} on the eastern dry side of the range to *c.* 10 km Ma^{-1} on the western rainy side of the range. Most particle paths in the orogen have a significant westward component of motion, causing them to move horizontally through the orogen from the slowly eroding east side to the rapidly eroding west side. As stated by Walcott (1998) (our italics): 'In two dimensions the amount of rock removed by erosion at a *spatial* point on the Earth's surface may be vastly greater than the exhumation *experienced by the rock* and is better referred to as denudation *or erosion*'.

England & Molnar (1990) made the useful distinction between *surface uplift*, meaning the vertical motion of the Earth's surface relative to sea level, and *rock uplift,* meaning the vertical motion of rock relative to sea level. Erosion and denudation are then defined as the difference between rock uplift and surface uplift at a single spatial point. Since erosion is measured in the vertical, it is probably best viewed as a flux (i.e., velocity normal to the approximately horizontal surface of the Earth) rather than a true velocity. This usage is consistent with estimates of drainage-scale erosion rates, which are calculated by dividing sediment yield by drainage area.

As used here, exhumation is measured by integrating the difference between the rock velocity and surface velocity while following the rock along its material path (cf. England & Molnar 1990). Barometers and thermochronometers provide information about exhumation since they tell us about the unroofing history specific to a rock sample. For the simple case of a vertical particle path, exhumation and erosion are equivalent, but this situation cannot be expected in general. When exhumation is divided by time, the resulting exhumation rate can be viewed as a spatially and temporally averaged erosion rate, but this equivalence should be viewed with caution. The approximation becomes increasingly flawed as the closure depth for the relevant barometric and thermochronometric data increases. For instance, cosmogenic isotopes provide estimates of exhumation rates within meters of the surface (Cerling & Craig 1994), which means that the measured rates are essentially equal to erosion rates. Low-temperature thermochronometers have deeper closure depths, and thus provide a more averaged estimate of the erosion rate. For instance, the apatite (U–Th)/He thermochronometer (Lippolt *et al.* 1994; Wolff *et al.* 1997) has a very low closure temperature (*c.* 75°C), which implies closure depths of 2–3 km, assuming slow erosion rates. This depth will be even shallower in areas of fast erosion because of the upward advection of heat caused by erosion. Apatite fission-track ages have a higher closure temperature (*c.* 110°C) and deeper closure depths (<4–5 km), so exhumation rates determined from those data will have a more approximate relationship to surface erosion rates. Zircon fission-track ages (*c.* 240°C) and ^{40}Ar–^{39}Ar ages (>300°C) have greater closure temperatures, and an even more distant relationship to surface erosion rates.

To summarize, we recommend the following definitions.

Exhumation: the unroofing history of a rock, as caused by tectonic and/or surficial processes.

Erosion: the surficial removal of mass at a spatial point in the landscape by both mechanical and chemical processes.

Denudation: the removal of rock by tectonic and/or surficial processes at a specified point at or under the Earth's surface.

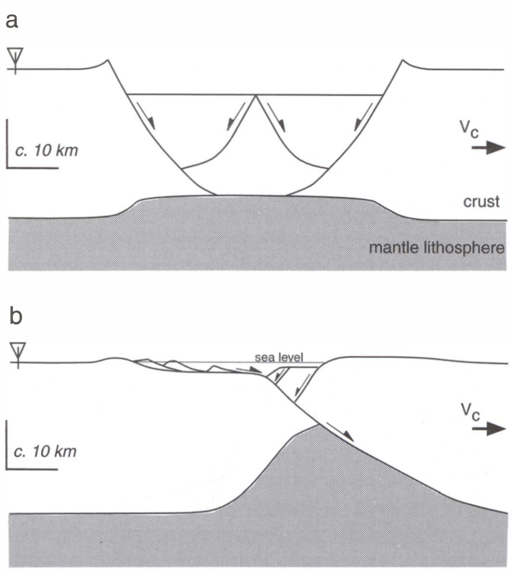

Fig. 3. Schematic illustration of divergent settings. (**a**) Symmetric rifting, resulting from coaxial extension of the lithosphere. (**b**) Asymmetric rifting, due to non-coaxial deformation of the lithosphere.

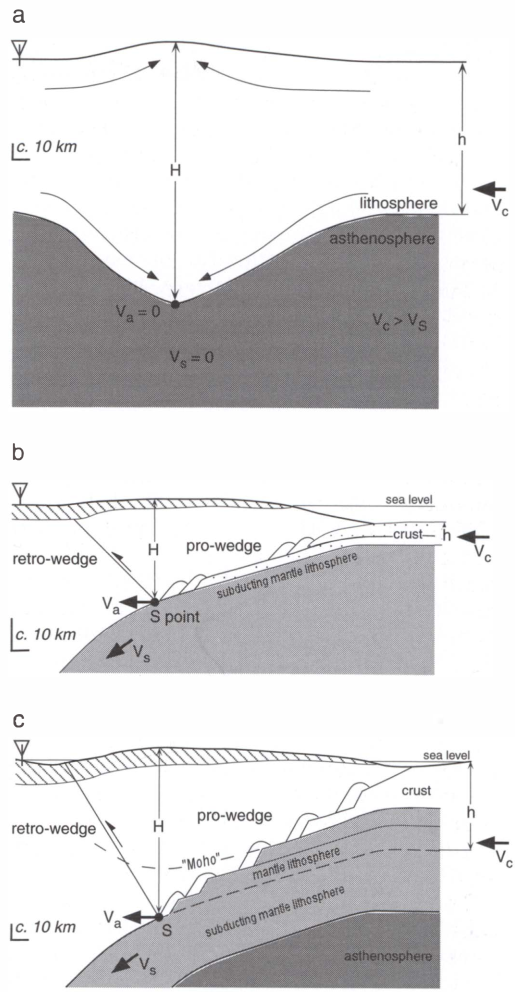

Fig. 4. Schematic illustration of convergent wedges showing the relationship between V_c, V_a, V_S and the S point. The vector for V_a shows the case for subduction-zone advance where $V_a > 0$. (**a**) Extreme case where both lithospheric plates are completely accreted into the orogen. V_S and the S point are undefined in this case. (**b**) Subduction-related accretionary wedge. The S point is at a depth of about 30 km. Incoming crustal section is about 6 km thick. (**c**) Convergent continental wedge. The S point is at a depth of about 120 km, which means that imbrication and accretion include both crust and lithospheric mantle from the underthrust plate.

Orogenic deformation and tectonic models

Plate-tectonic theory dictates that the velocity field at the Earth's surface is dominated by the motion of rigid plates. Deformation occurs mainly at the velocity gradients within the zones that define the plate boundaries. To a first order, orogens are either divergent (Fig. 3) or convergent (Fig. 4) depending on the sign of the relative velocity, V_c, defined here to be positive for convergence.

Divergent orogens display considerable complexity depending on the symmetry of internal structure (Fig. 3). In continental settings, divergence results in symmetric or asymmetric stretching and rifting. Symmetric stretching can result in significant amounts of ductile thinning. A more asymmetric style of stretching is influenced by the development of strongly localized normal faults, which impart a sense of vergence to the deformation. In both cases, the deformation, whether brittle or ductile, involves a strong component of horizontal extension and tectonic denudation of the footwall beneath major normal faults.

Similarly, convergent orogens show a broad diversity depending on both the boundary conditions and the internal response. The general expectation is that convergence causes pervasive horizontal contraction throughout the orogen. However, horizontal extension has been recognized as an important feature in a number of convergent orogens. For instance, in the special case of a retreating plate boundary (Royden 1993; Waschbusch & Beaumont 1996), oceanic subduction is commonly characterized by horizontal contraction in the accretionary wedge and simultaneous horizontal extension within the arc and back arc.

The complexity of deformation within convergent orogens, as well as the diversity of tectonic style and exhumation processes, has led to considerable effort in describing and modelling these orogenic systems. Much of the deformational response to convergence is dictated by how much of the lithosphere is involved in the deformation. At one limit, the entire lithosphere of one or both bounding plates is forced to contract and thicken (Fig. 4a). Whole-lithosphere thickening has been proposed as an important feature of large-scale continental collisions, such as the India–Asia collision (England & McKenzie 1982; England & Houseman 1986). England & Houseman (1986) have concluded that tectonic exhumation will occur when the gravitationally unstable mantle lithospheric root detaches, but they argue that the orogen otherwise shows no strong tendency for vertical thinning during convergence.

The other end member for convergent orogens is oceanic subduction zones, which are characterized by accretion of a thin layer, often limited to only the sedimentary cover of the downing plate (Fig. 4b). In this case, the orogen is the accretionary wedge, which grows slowly by accretion of sediments. Intermediate to these end-members are relatively small convergent orogens that form by accretion and contraction of sediment, crustal basement rocks, and sometimes lithospheric mantle as well (Fig. 4c). Well-studied examples include the European Alps, the Pyrenees, the Apennines, and the arc–continent collision of Taiwan.

For a subduction/accretion model, there is always a point or, more precisely, a locus of points at the base of the orogen, that mark the lower limit of mass transfer (i.e. accretion) across the plate boundary (Willett *et al.* 1993; Beaumont *et al.* 1994; Waschbusch & Beaumont 1996). Below this point, which we denote as S, subduction of the downgoing plate occurs beneath a discrete shear zone. The velocity of S, designated as V_a, describes retreat or advance of the subduction zone. As shown in Fig. 4, our convention is to define V_c and V_a relative to a fixed overriding plate, with positive towards the overriding plate. A positive V_a implies advance of the subduction zone towards the overriding plate; a negative V_a represents slab rollback and motion of the subduction boundary away from the overriding plate. Waschbusch & Beaumont (1996) used a slightly different convention. In their notation, $V_c = V_P - V_R$ and $V_a = V_S - V_R$, where V_P and V_R are the absolute velocities of the subducting (pro-) and overriding (retro-) plates, and V_S is the horizontal migration velocity of the subducted slab relative to the deep mantle.

Subduction-zone advance ($V_a > 0$) describes the motion of the S point towards the overriding plate. Advance must be accompanied by contraction of the overriding plate. A modern example is western South America where crustal thickening and growth of the Andean orogenic belt has occurred without any significant accretion from the subducting Nazca plate. Without accretion, shortening of the South American plate must be accompanied by eastward motion of the Nazca slab (Pardo-Casas & Molnar 1987; Isacks 1988; Pope & Willett 1998). Note that outward growth of an orogen is a natural consequence of the addition of mass by accretion and will occur independent of the velocity of S. As a result, the advance of the retro-side deformation front does not require the advance of the underlying subduction zone.

When $V_a < 0$, the S point migrates away from the overriding plate, defining the case of subduction retreat. Slab retreat or rollback has long been recognized as an important factor influencing the style of deformation in convergent orogens (Uyeda & Kanamori 1979; Dewey 1980; Jarrard 1986; Royden & Burchfiel 1989; Waschbusch & Beaumont 1996). Examples of retreating convergent orogens include the Mariana margin in the western Pacific and the Hellenic margin of Greece. In an oceanic setting, slab retreat leads to extension of the upper plate and formation of a back-arc basin. In a continental setting, the consequences are less clear, but as the slab retreats away from the overriding plate, the position of active contraction migrates away from the upper plate, creating additional space to accommodate accreted material. If retreat creates space faster than accretion can fill it, then the upper plate might respond by horizontal extension.

The deformation within a convergent orogen is also controlled by the distribution of fluxes around the orogen. Accretion and surface erosion control the fluxes into and out of the orogen. Accretionary fluxes generally enter into the orogen on its pro-side, either by frontal accretion (offscraping) or basal accretion (underplating). Accretion can also occur on the retro-side of the wedge (Willett *et al.* 1993), but the fluxes are usually much smaller than those on the pro-side, except for the case of subduction-zone advance. A significant erosional flux only occurs when there is extensive 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. A fast rate of frontal accretion will tend to cause pervasive contraction throughout the orogen. This case is nicely illustrated by the strain calculations in Dahlen & Suppe (1988). Waschbusch & Beaumont (1996) show that at retreating subduction zones, frontal accretion will also result in horizontal contraction in the accreted material. They also show that when rates of accretion are low or nil, subduction-zone retreat can cause mixed contraction and extension in pre-existing rocks that border the subduction zone. A fast rate of erosion in the Olympic Mountains of the Cascadia fore-arc high has caused pronounced vertical extension in the most deeply exhumed part of the accretionary wedge (Brandon *et al.* 1998; Brandon & Fletcher 1998). In contrast, a fast rate of underplating combined with a slow rate of erosion should cause horizontal extension in the upper rear part of the orogen (Platt 1986, 1993; Brandon & Fletcher 1998). A possible example might be the Apennine thrust belt, which shows active extension in the internal part of the orogen (Elter *et al.* 1975; Patacca *et al.* 1993). Fast underplating might be expected given that the Apennines saw a transition over the last 5 Ma from oceanic subduction to subduction of the passive margin of the Adriatic continental block (Dewey *et al.* 1989). An alternative explanation (Royden 1993) is that extension is caused by subduction-zone retreat.

What gets exhumed?

Exhumation occurs at a variety of tectonic settings (Fig. 5), but mainly at oceanic rifts and transform faults, continental rift zones, subduction zones, and at continent–continent collision zones. Here we summarize the types of rocks that are exhumed in these settings. In general, we find that each tectonic setting has a maximum exhumation depth, with oceanic rifts and transforms showing the shallowest exhumation depths (*c.* 10 km) and continental collision zones, the deepest (>125 km).

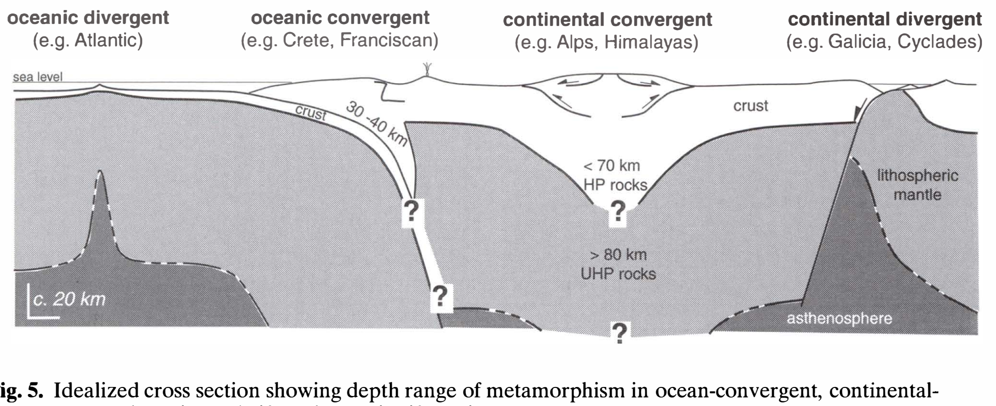

Fig. 5. Idealized cross section showing depth range of metamorphism in ocean-convergent, continental-convergent, and continental-rift and oceanic-rift settings.

Oceanic rifts and transforms

This setting is charaterized by shallow exhumation from depths of *c*. 10 km. Given the thin crust for oceanic lithosphere, this depth is sufficient to expose the mantle, as evidenced by spectacular local exposure on the seafloor of the serpentinized peridotites. In the modern oceans, exhumed mantle is most commonly found along long-offset transform faults and 'under-fed' ridge crests. The exposed peridotites are plagioclase- and spinel-bearing, indicating shallow mantle rocks.

Exhumed mantle has also been observed in the deep ocean in association with extreme rifting. Off the Spanish/Portuguese coast between the supposedly oceanic crust (Grau *et al.* 1973) of the Iberian abyssal plain and the Galicia margin, serpentinized peridotite derived from the upper mantle is largely buried beneath sediments but also crops out locally (Boillot *et al.* 1980). The Galicia margin is made up of a number of tilted blocks formed during continental rifting. According to Boillot *et al.* (1980), the serpentinized peridotite is thought to be the result of serpentinite diapirism and tectonic unroofing of mantle rocks along the rift axis of the margin just before sea-floor spreading started between Galicia and Newfoundland. The setting would thus be transitional from continental to oceanic rifting.

Ophicalcites record ancient examples of exhumed oceanic lithosphere. An ophicalcite is a sedimentary breccia made up of mafic and ultramafic clasts set in a pelagic carbonate matrix (e.g. Lemoine 1980; Bernoulli & Weissert 1985). They are thought to form in intra-oceanic settings. The clasts provide a clear record that all levels of the underlying oceanic lithosphere were exposed at the seafloor.

Continental rifts

The rift shoulders of the East African rift expose mainly middle to lower crustal Precambrian gneisses, but those rocks are known to have been mostly exhumed prior to the onset of Cenozoic rifting. The rift process itself appears to have caused only minor exhumation. Continental rifting in the Basin-and-Range province of western North America locally resulted in exhumation of the upper and middle crust (e.g. Applegate & Hodges 1995). However, despite large-magnitude extension (about 100%, Wernicke *et al.* 1988), no high-pressure metamorphic rocks (i.e. >10 kbar) are found at the surface.

Deeper exhumation in continental rift zones has occurred in the Cyclades of Greece, where blueschist and eclogite from *c*. 50 km depth are locally exposed (Schliestedt *et al.* 1987; Okrusch & Bröcker 1990). Divergence started there sometime after the middle Oligocene (Raouzaios *et al.* 1996; Thomson *et al.* 1998; Ring *et al.* 1999), but yet we know that much of the exhumation of the Cycladic high-pressure rocks occurred earlier, during the Eocene and Early Oligocene, shortly after the rocks were subducted and accreted (Avigad *et al.* 1997; Ring *et al.* 1999). Forster & Lister (this volume) argue that deep exhumation in the Cyclades is a result of multiple episodes of normal faulting, with any single event involving only a modest amount of exhumation.

Collectively, these examples suggest that individual episodes of continental rifting will exhume rocks from depths no greater than *c*. 25 km. Nevertheless, it may be difficult to tell whether or not tectonic exhumation was caused by divergent plate motions (rifting) or by synconvergent extension, as highlighted by the discussion between Andersen (1993) and Fossen (1993) concerning exhumation in the Norwegian Caledonides.

Subduction zones

Subduction zones expose a wider variety of deeply exhumed rocks. In a number of cases, the metamorphic grade is no greater than blueschist facies, indicating exhumation from depths of about 30–40 km, which is the maximum thickness of modern subduction-related accretionary wedges.

The Mariana convergent margin (Fryer 1996) displays an intriguing example of deep exhumation. Cold intrusions of serpentinite are found as conical volcano-like features littering the seafloor of the Mariana forearc. Rare blueschist minerals are locally found in association with these serpentinite diapirs. For this case, exhumation of mantle and high-pressure rocks is thought to be driven by the buoyancy caused by serpentinization of mantle peridotites. Serpentinization can cause a decrease in density from 3300 kg m^{-3} to as low as 2600 kg m^{-3} (–22%; Christensen & Mooney 1995).

Eclogite-facies mafic rocks (i.e. eclogites) have been recognized at many ancient subduction zones. These rocks may occur as isolated blocks, in association with lenses and blocks of serpentinized peridotite. The evidence usually indicates that the eclogites were severely disrupted and dismembered after metamorphism. Pressure estimates indicate exhumation from 50 km and deeper. The Central Belt of the Franciscan subduction complex of coastal California serves as a classic example of disrupted eclogites in a subduction zone setting. The 'knockers' are made up of typical oceanic basalts. Some are surrounded by an actinolitic rind, indicating that they resided for some time in a serpentinite matrix (Coleman & Lanphere 1971; Moore 1984).

The hanging wall of the subduction zone is, at least in some cases, invoked as a source for these deep rocks (e.g. Coleman & Lanphere 1971; Platt 1975; Moore 1984 for the Franciscan subduction complex). The reason is that modern subduction zones have well-defined Benioff zones indicating that the lower-plate mantle is subducted, not accreted. This observation 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 minerals in the Mariana forearc suggests that serpentinite diapirs might be involved in exhumation of eclogite blocks as well.

Continental collision zones

The internal zones of collisional belts expose the widest variety of exhumed rocks. Schist and gneiss from the upper and middle crust are commonly exhumed in this setting. Some convergent orogens, such as the Delamerian orogen of south Australia, the Mount Isa orogen of northeast Australia, and the Rocky Mountains of the western US, are almost entirely made up of upper and middle crustal rocks and generally lack exposed high-pressure granulite, blueschist, or eclogite (i.e. rocks characterized by metamorphic pressures <10 kbar). We suspect that high-pressure rocks are formed in these settings but the exhumation processes operating there were somehow unable to bring those rocks to the surface.

Other collisional belts show clear evidence of crustal thickening and deep exhumation during orogenesis, as indicated by the general occurrence of metamorphic rocks from > 40 km depth. A classic example is the Sesia zone in the Italian Alps (Compagnoni & Maffeo 1973), which contains high-pressure continental crustal rocks that were metamorphosed at the base of a thickened crustal root, and then subsequently exhumed. Pressure estimates indicate metamorphism at *c.* 70 km, which is consistent with the maximum Moho depths in modern orogens (e.g. Meissner 1986; Christensen & Mooney 1995).

Less common within collisional orogens are deeply exhumed oceanic assemblages. An interesting example is the Zermatt-Saas zone of the Swiss-Italian Alps (e.g. Bearth 1956, 1967, 1976), which consists of metabasalts, metacherts, and ultramafite. Metamorphic assemblages indicate both high pressure and ultrahigh-pressure conditions (locally >100 km depth). The lithological association suggests a subduction-zone environment, with initial accretion of crust and mantle from an oceanic lower plate, followed by low-temperature/high-pressure metamorphism. However, age constraints (Reddy *et al.* 1998) suggest that exhumation occurred during continental collision, which distinguishes these rocks from subduction-zone eclogites, like those of the Franciscan complex.

The most challenging examples of deeply exhumed rocks are ultrahigh-pressure (UHP) metamorphic rocks, which are usually found exhumed in continental collisional zones. These rocks were first discovered in the Cima-Lunga nappe (Ernst 1977; Evans & Trommsdorff 1978), the Dora-Maira massif (Chopin 1984), and the Zermatt–Saas zone (Reinecke 1991) of the Alps, and the Western Gneiss region of Norway (Griffin 1987; Smith & Lappin 1989). They are now known from a number of collisional orogens (Coleman & Wang 1995). Ultrahigh-pressure rocks are continental or oceanic crustal rocks that were metamorphosed within the stability field of coesite or diamond (Schreyer 1995; Coleman & Wang 1995). The exposure of these rocks at the Earth's surface demonstrates that continental and oceanic crust can be subducted to depths >100 km and then returned to the surface. A common debate is whether or not UHP rocks formed within highly overthickened crust or within the mantle (see section on UHP rocks below).

Garnet peridotites are also exposed in several collisional orogens such as the Alps (Alpe Arami, Ernst 1977; Evans & Trommsdorff 1978) and the Betic–Rif orogen (e.g. Ronda and Beni Bousera peridotites, Loomis 1975; Pearson *et al.* 1989). For the Ronda and Beni Bousera peridotites, graphite pseudomorphs after diamond indicate depths >125 km (Pearson *et al.* 1989; Tabit *et al.* 1990). The initial stages of exhumation are attributed to Mesozoic rifting (Vissers *et al.* 1995). However, the Alpe Arami peridotite was subjected to metamorphism at depths of >70–100 km during the Alpine orogeny (e.g. Ernst 1977; Evans & Trommsdorff 1978; Becker 1993), which would have postdated Mesozoic rifting. The presence of deep-seated ultramafic rocks at the Earth's surface indicates that orogenic wedges must have sufficient upward flow to overcome the negative buoyancy of these rocks.

General remarks

Our summary suggests a counter-intuitive result: continental rift zones seem to have only modest potential for deep exhumation, whereas continental collision zones seem to have the greatest potential. We see no simple explanation for this result.

Another highlight is the evidence from the Mariana subduction zone indicating that some high-pressure metamorphic rocks might be brought back to the surface by buoyancy-driven exhumation, triggered by sepentinization of upper-plate mantle. If such serpentinization does occur beneath the forearc region, then we will probably need to reconsider what the seismic Moho means in those settings. Serpentinization can cause a decrease in compressional wave velocities from about 7.7–8.2 km s^{-1} to 5.3–5.5 km s^{-1} (Christensen & Mooney 1995).

Diagnostic features of different exhumation processes

A difficult question in most orogenic belts, especially the older ones, is the relative contributions of different exhumation processes. Here, we assess features that might be diagnostic of erosion, normal faulting, and ductile thinning. We also evaluate some problems with quantifying the rates of these exhumation processes.

Erosion

The large volumes of detrital deposits found adjacent to almost all convergent continental orogens provides ample evidence that erosion is a significant exhumation process. The Himalayan foreland and the offshore Bengal and Indus fans preserve an important record of erosional exhumation of the Himalayas (Cerveny *et al.* 1988; Copeland & Harrison 1990; Burbank *et al.* 1993). Nie *et al.* (1994) proposed that the voluminous Songpan–Ganzi flysch was produced during exhumation of the Dabie Shan ultrahigh-pressure rocks. None of this evidence requires that erosion is the sole exhumation process or even the dominant process, but clearly erosion cannot be ignored as a contributing factor.

Erosion is often assumed to be a fairly slow exhumation process, but there is little support for this view. Figure 6 summarizes Milliman & Syvitski's (1992) compilation of modern sediment yield from a global distribution of 280 river drainages. The conclusion is that the average rate of mechanical erosion of the drained part of the continents is *c.* 0.052 km Ma^{-1}. However, more important is the fact that recorded erosion rates reach local values of 5–13 km Ma^{-1} (e.g. Southern Alps of New Zealand and the Taiwan Alps). In fact, 2% of the drained area of the continents has erosion rates > 0.5 km Ma^{-1}. It is important to stress that these rates represent drainage-scale averages. Thus, local rates within a drainage could be much greater. Erosion rates associated with warm-based alpine glaciers, such as those of southern Alaska, have rates of 1–100 km Ma^{-1} (Hallet *et al.* 1996). It is interesting to speculate that an increase in Alpine glaciation, caused by global cooling, increased precipitation, and/or growth of mountainous topography, might play a major role in exhuming metamorphic rocks and in limiting the maximum height of mountains. The important conclusion is that relative to tectonic exhumation, which may be no greater than *c.* 5–10 km Ma^{-1}, surficial erosion can locally be a very fast process.

Fast eroding regions tend to be mountainous, tectonically active, and wet. Conversely, arid climates tend to have slow erosion rates regardless of the amount of topography. At present, arid landscapes cover about one-third of the continents (p. 278 in Bloom 1998). Arid high plateaux, such as Tibet and the Altiplano in the

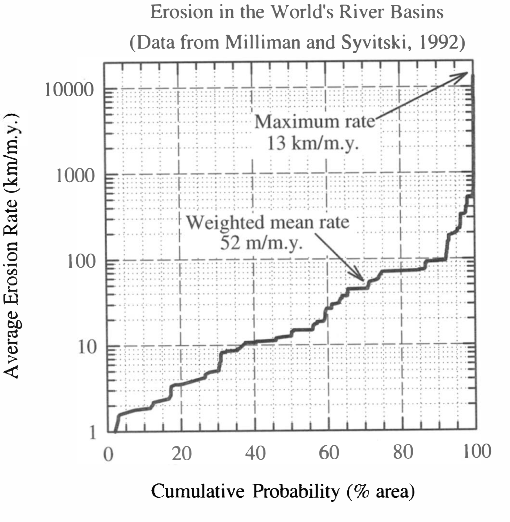

Fig. 6. Average erosion rate from the world's river basins plotted against cumulative probability (data compiled from Milliman & Syvitski 1992). The diagram shows that mechanical erosion is typically slow (weighted mean rate *c.* 0.05 km Ma^{-1}), but in some orogens, mechanical erosion can be extremely fast with rates locally exceeding 10 km Ma^{-1}.

Andes, serve as examples of mountainous landscapes with little to no erosion. The transition from an arid to a humid climate appears to be marked by a dramatic increase in erosion rates, primarily due to the enhanced transport capacity that comes with a well-drained landscape. Drainage-scale erosion rates are commonly assumed to scale with precipitation or runoff, but regional compilations do not provide clear support for this inference (e.g. Pinet & Souriau 1988; p. 134 in Allen 1997). In other words, erosion rates do not appear to be strongly influenced by climate as long as the climate is not arid. The rate of tectonically driven uplift appears to be a much more important factor in controlling local erosion rates. Recent work by Burbank *et al.* (1996) and Hovius *et al.* (1997) have called attention to the importance of bedrock landslides in producing transportable material in well-drained tectonically active landscapes. In this type of setting, the rate of erosion is probably directly driven by the rate of the tectonically-driven uplift, given the usual conclusion that in well-drained humid landscapes, river incision keeps pace with tectonic uplift.

An important question is whether modern erosion rates are representative of long-term erosion rates. We ignore human effects here; most compilations of modern erosion rates have included corrections to remove this factor. Perhaps the most important factor influencing erosion rates on the short-term is climate-induced changes in vegetation cover, which controls the amount of regolith and soil that can be stored on the hill slopes (Bull 1991). A loss of vegetation cover can result in very high short-term erosion rates, but those rates can only be sustained for a short period of time until the stored regolith is striped away. The major climate cycles have periods ranging from 21 to 100 ka, with the 100 ka glacial cycle being especially pronounced during the Quaternary. In this context, long-term erosion rates can be usefully defined as the average rate of erosion for a period of time >100 ka, to ensure that 'short-term' climate variations are averaged out.

Long-term erosion rates can be estimated by reconstructing eroded geological features, by isopach analysis of synorogenic sediments, and by thermochronologic dating of exhumed bedrock or detrital grains in synorogenic sediments. A number of studies have documented long-term syn-orogenic erosion rates ranging from 1 to 15 km Ma^{-1} (e.g. Kamp *et al.* 1989 for the Southern Alps of New Zealand; Copeland & Harrison 1990 for the Himalayas; Burbank & Beck 1991 for the Salt Range of Pakistan; Fitzgerald *et al.* 1995 for the central Alaska range; Johnson 1997 for the Betic Cordillera of southern Spain; Brandon *et al.* 1998 for the Olympic Mountains).

Our conclusion is that erosion cannot be excluded as an exhumation process on the basis of rate alone. For modern orogens, lack of subaerial relief (e.g. submarine accretionary wedge) or an arid climate or pronounced rain shadow (e.g. Tibet) would qualify as credible arguments for generally slow erosion rates, but these arguments are fairly difficult to apply to ancient settings. Isopach analysis provides a direct measure of erosional exhumation, although it is commonly found that synorogenic sediments end up being dispersed to great distances from the orogen. In the Himalayan system, the dispersal is over distances of thousands of kilometres, reaching distant ocean basins and subduction zones. The reason for the wide dispersal is that a collisional foreland basin can never hold more than about half of the sediment produced by erosional lowering of the orogen (England 1981). Wide dispersal makes it difficult, even in the best-preserved orogens, to make a useful comparison between the volume of eroded sediment and the depth of exhumation.

Another factor that has to be taken into account in subduction zone settings is the recycling of eroded material back into the subduction zone. In a similar manner, sedimentary basins adjacent to collision orogens might be overridden and hidden underneath large-slip faults.

Another method that has much promise for distinguishing between normal faulting and erosional exhumation is the use of isotopic cooling ages in synorogenic sediments (e.g. Garver *et al.* this volume). Detrital ages provide information about the lag-time distribution of the sediment, where lag time is defined as the difference between the cooling age of a detrital grain and its depositional age. For instance, the growth and decay of an erosion-dominated orogen should show a parallel decrease and increase in lag time for the sediment shed from the orogen. The changes observed in the lag-time distribution will be delayed relative to the actual growth and decay of the topography because some time is needed to erode away rocks that have pre-orogenic cooling ages after initial orogenic uplift, and to erode away rocks that have syn-orogenic cooling ages during the post-orogenic decay of the topography. In contrast, after orogenic uplift has stopped normal faulting should be marked by the rapid appearance of sediment with short lag time, and then long periods of time where the lag time increases at the same rate as stratigraphic age. These two features of the lag-time evolution mark the fact that normal faulting brings hot rocks to the surface without any need for erosion. Furthermore, if normal faulting is rapid, much of the upper crust will be reset to a common isotopic cooling age. Brandon & Vance (1992) reported an example like this for sediments shed from the Eocene metamorphic core complexes in the Cordilleran thrust belt of eastern Washington State (USA) and southeast British Columbia (Canada). Sandstone samples spanning from the Eocene through the Early Miocene show peaks in the fission-track grain-age distribution for detrital zircons that remain fixed at *c.* 43 and *c.* 57 Ma. These peaks were called static peaks because they did not move with depositional age. They can be related to the slow erosion of the Eocene core complexes after they formed.

Normal faulting

There is abundant evidence that normal faulting aids the exhumation of metamorphic rocks. The Basin-and-Range province (Armstrong 1972; Crittenden *et al.* 1980; Davis 1988; Foster & John this volume), the Aegean (Lister *et al.* 1984; Thomson *et al.* this volume; Forster & Lister this volume), the Betic Cordillera (Platt & Vissers 1989; Vissers *et al.* 1995) and the Alps (Mancktelow 1985; Selverstone 1985; Behrmann 1988; Ratschbacher *et al.* 1991; Ring & Merle 1992; Reddy *et al.* 1999) are well-documented settings where normal faulting contributed to exhumation.

The most commonly cited evidence for normal faulting is the presence of 'younger over older' or 'low-grade on high-grade' tectonic contacts, where large faults, with low to moderate dips, have placed younger above older rocks or low-grade rocks on high-grade rocks, and in the process have cut out a significant thickness of stratigraphic or metamorphic section. A more diagnostic feature for large-slip normal faults is the juxtaposition of a ductily deformed footwall with a brittlely deformed hanging wall (e.g. Lister & Davis 1989; Foster & John this volume). The evolution of superimposed sedimentary basins can be used to determine the relative displacement of footwall and hanging wall. These 'piggy-back' style basins also provide a record palaeohorizontal, which can be used to restore underlying faults back to their original orientation. This information, together with the shear sense of a fault, is essential for resolving the nature (reverse or normal in relative sense) of a fault.

The pattern of cooling ages in the field can also provide clues about the exhumation process. Exhumation due to normal faulting commonly results in abrupt breaks in the cooling-age pattern with younger ages in the footwall of the normal fault (Fig. 7a) (Wheeler & Butler 1994). Johnson (1997), Foster & John (this volume) and Thomson *et al.* (this volume) provide examples from the Betic Cordillera, the Basin-and-Range province and Crete, respectively. In contrast, erosional exhumation should be characterized by a relatively smooth variation in cooling ages across the eroded region (Fig. 7b). Brandon *et al.* (1998) report a useful example for the erosional exhumed fore-arc high above the Cascadia subduction zone.

Attenuation of stratigraphic or metamorphic units, by itself, is not diagnostic of tectonic thinning by normal faulting (Wheeler & Butler 1994; Ring & Brandon 1994; Ring 1995; Butler & Freeman 1996) (Fig. 8). Thrust faults can thin a stratigraphic or metamorphic section when that section is tilted rearward prior to thrusting. In this case, the thrusts can cut down-section in the direction of transport, while cutting up towards the Earth's surface as thrust faults should.

In some cases thrusts may be rotated by

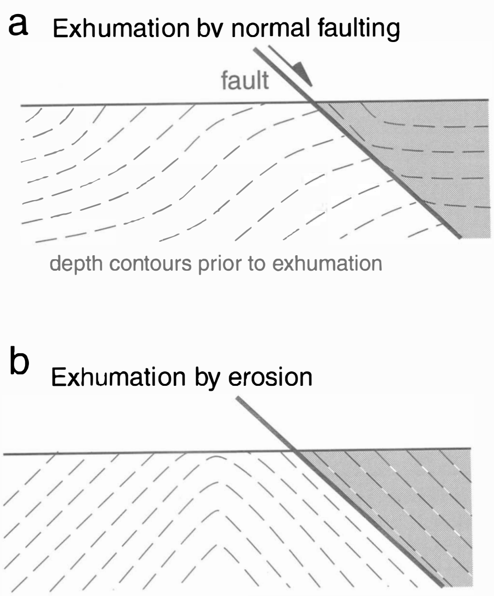

Fig. 7. Exhumation pattern as a function of process. Dashed lines show depth contours prior to exhumation. (**a**) Normal faulting will create an asymmetric exhumation pattern. Cooling ages will get younger as one approaches the normal fault (see Foster & John and Thomson *et al.* this volume). (**b**) Exhumation by erosion alone will commonly result in a broadly domed pattern, which should vary smoothly, independent of faults and local structures.

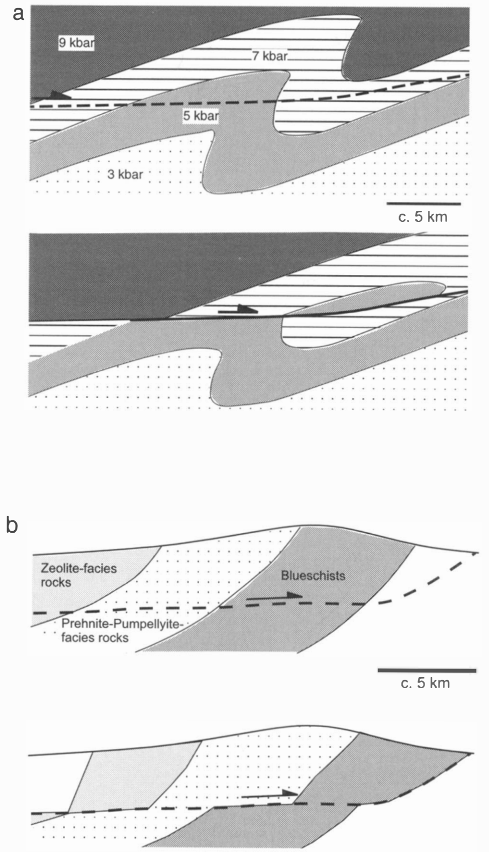

Fig. 8. Examples of attenuation of metamorphic section by contractional faults. (**a**) Normal nappe sequence characterized by the tectonic juxtaposition of higher-grade rocks over lower-grade rocks. Nappe sequence is further affected by post-nappe folding. Re-imbrication of the thrust and fold sequence; the re-imbricating contractional fault climbs upwards towards the foreland, but cuts downward through the tilted section resulting in faulted contacts where lower-grade rocks are thrust over higher-grade rocks. Such a multiphase contractional history is commonly found in the internal zones of orogens. (**b**) Contractional fault cutting down through an uplifted metamorphic section, while cutting upward relative to the foreland in the direction of tectonic transport, and thereby attenuating metamorphic or structural section.

folding, block tilting or differential exhumation until their geometry classifies them as normal faults at the front of a thrust culmination (e.g. Price 1981). Conversely, normal faults are sometimes domed by isostatic rebound due to unloading of their footwalls until the fault locally has a thrust sense of offset (Buck 1988; Lister & Davis 1989). Therefore, the kinematic development of a fault has to be related to the *P–T–t* evolution in its foot- and hanging wall at the time of tectonic transport, or displacement along a fault relative to the palaeosurface at the time of fault activity has to be demonstrated in order to distinguish between normal faults and thrusts (cf. Wheeler & Butler 1994).

A problem for quantifying the relative contribution of normal faulting to exhumation of deep-seated rocks is that the total offset and original dip of most crustal-scale normal faults, especially those involving deep crustal rocks, are typically poorly resolved. The offset is commonly approximated with barometric estimates from above and below the fault. In this case, it has to be demonstrated that faulting is actually contemporaneous with the pressure break across the fault. Reddy *et al.* (1999) report an example from the Zermatt–Saas zone where the break in metamorphic pressure across a shear zone allowed them to constrain a total of 30 km of exhumation by normal faulting within 9 Ma, indicating an average exhumation rate of 3.5 km Ma^{-1}.

Only a few studies have attempted to measure slip rates for normal faults. The greatest reported slip rates come from some extensional detachments in the Basin-and-Range province and are of the order of 7–9 km Ma^{-1} (Davis & Lister 1988; Spencer & Reynolds 1991; Foster *et al.* 1993; Scott *et al.* 1999; Foster & John this volume). There is growing evidence that many of these 'high-speed' extensional detachments had an initially gentle dip (<30°) (Scott & Lister 1992; Foster & John this volume). Other estimates for exhumation by extensional detachments in the Basin and Range indicate lower rates of about 1–2 km Ma^{-1} (see review in Foster & John this volume). Average rates from other orogens appear to be generally slow: (1) <2 km Ma^{-1} for a 10 Ma interval for the Tauern window in the Eastern Alps (Frisch *et al.* 1998) and, (2) 0.6–0.9 km Ma^{-1} and <0.6–0.7 km Ma^{-1}, respectively, for a 20 Ma interval for the Menderes Massif in western Turkey (Hetzel *et al.* unpublished data; Ring unpublished data).

Normal faults in long-lived continental rifts also appear to have relatively low slip rates and slip generally occurred along steeply dipping (*c.* 60°) normal faults. In the Miocene to Recent East African rift system, apatite fission-track ages are, in general, not reset by the young rifting events (Foster & Gleadow 1996), which indicates a slip rate for the normal faults of <0.3 km Ma^{-1}. At the Livingstone escarpment of the northern Malawi sector of the East African rift system, the Miocene to Recent offset at the escarpment yields similar low slip rates (Ring unpublished data).

The exhumation rate during normal faulting is a function of both slip rate and fault dip. A normal fault with a 30° dip that slipped at a rate of 7–9 km Ma^{-1} would exhume rocks at a rate of 3.5–4.5 km Ma^{-1}. If this detachment had an original dip as gentle as 20°, it would exhume rocks at a rate of 2.4–3.1 km Ma^{-1}. It is important to note that most slip rates are averaged over relatively long periods of time (see cited time intervals above for the Tauern window and the central Menderes Massif). It is conceivable that normal faulting occurred in pulses at faster rates. None the less, the long-term average appears to have been relatively slow.

Ductile thinning

Penetrative deformation fabrics present in most exhumed mountain belts indicate that ductile flow is an important process. This process can either aid or hinder exhumation, depending upon whether ductile flow causes vertical thinning as associated with the formation of a subhorizontal foliation, or vertical thickening as associated with the formation of a subvertical foliation (Fig. 2b and c). The presence of a subhorizontal foliation is generally diagnostic of ductile thinning. The general observation of subhorizontal foliations in the internal zones of many orogens shows that ductile thinning commonly aids exhumation (Selverstone 1985; Wallis 1992, 1995; Wallis *et al.* 1993; Platt 1993; Mortimer 1993; Ring 1995; Krabbendam & Dewey 1998; Ring *et al.* 1999). Ductile thinning by itself cannot fully exhume rocks and an additional exhumation process is required to bring rocks back to the Earth's surface (Platt *et al.* 1998; Feehan & Brandon 1999).

In the simplest case, where exhumation is entirely controlled by ductile thinning, exhumation by ductile thinning is given by the average stretch in the vertical because this tells how much the vertical has changed in thickness. However, if exhumation occurs by additional processes as well, it is more difficult to quantify the contribution of ductile thinning to exhumation. In this case, the vertical rate at which a rock moved through its overburden and the rate of thinning of the remaining overburden at each step along the exhumation path have to be considered (Feehan & Brandon 1999). The general conclusion is that the contribution of ductile thinning to exhumation will always be less than that estimated from the vertical stretch only. Simple one-dimensional calculations show that the contribution of ductile thinning in convergent wedges from western North America is less than half that indicated by the estimated finite vertical shortening in the rock (Feehan & Brandon 1999; Ring & Brandon this volume). It follows that huge vertical strains are needed for ductile thinning to make a significant contribution to exhumation. Vertical contraction on the order of 70% has indeed been documented, for example, by Norris & Bishop (1990) and Maxelon *et al.* (1998) from the interior of the Otago accretionary wedge, exposed on the South Island of New Zealand. Dewey *et al.* (1993) also reported vertical contraction on the order of 70–80% from the Western Gneiss region. The studies by Feehan & Brandon (1999) and Ring & Brandon (this volume) show that ductile thinning operated at low rates of <0.3 km Ma^{-1}. Recent modelling by Platt *et al.* (1998) in the Betic Cordillera argues for vertical shortening of 75% and ductile thinning operating at a rate of 4.5 km Ma^{-1}.

Platt (1993) pointed out that thrusting alone can not tectonically exhume rocks. This is certainly true if thrusting is confined to a thin zone at the base of a sequence of nappes, which is commonly the case in the upper crust. However, in the internal zones of convergent orogens, nappes usually show a pervasive degree of internal deformation and generally flat-lying foliations. The foliations show that nappe stacking was associated with pronounced vertical thinning, which would ultimately contribute to exhumation of the nappes. In this regard, it is interesting to note that high-pressure belts, as a rule, occur above lower grade units indicating that the overburden of the high-pressure nappes must have been reduced prior to their emplacement above the lower pressure units. Pervasive ductile flow in the hanging wall associated with thrusting of the high-pressure nappe might have aided the exhumation of the latter. Note that vertical ductile thinning might be coupled with erosion and/or normal faulting in upper parts of the nappe pile.

***P–T–t* data and exhumation processes**

P–T paths

The shape of *P–T* paths is sometimes used to distinguish exhumation processes. The shape of a *P–T* path is typically difficult to resolve, especially the youngest part of the path, which is probably the most diagnostic of process (J. Selverstone, pers. comm. 1996). Nonetheless, fast initial exhumation rates of deeply buried rocks should generally result in near-isothermal decompression. Such *P–T* paths are known, for instance, from the garnet-oligoclase-facies rocks of the Southern Alps of New Zealand (Holm *et al.* 1989; Grapes 1995) and the Mulhacen nappe of the Betic Cordillera (e.g. Gomez-Pugnaire & Fernandez-Soler 1987) (paths 1 and 2 in Fig. 9). The Southern Alps were dominantly exhumed by erosion and the Mulhacen nappe is thought to be exhumed primarily by normal faulting. Nonetheless, the shapes of the *P–T* paths from both units are virtually indistinguishable (Fig. 9).

T–t *paths*

A number of isotopic studies have used evidence of rapid cooling to argue for rapid exhumation (e.g. Copeland *et al.* 1987; Zeck *et al.* 1992). Rapid exhumation of deep-seated rocks causes upward advection of heat and thus retards cooling and promotes isothermal decompression. The result is a relatively high thermal gradient and compressed isotherms in the upper crust (Fig. 10). In this situation, rocks can quickly move through the closely spaced isotherms without being much exhumed. The closure-temperature concept (Dodson 1973) indicates that a radiogenic isotopic system will record the time when a dated mineral cools below its effective closer temperature. However, there is no 'closure pressure'. Thus, for cases of near-isothermal decompression typical for fast exhumation (path 3 in Fig. 9), it is difficult to estimate the depth where the isotopic system closed. As a result, exchumation rates will be poorly resolved. Only hairpin *P–T*

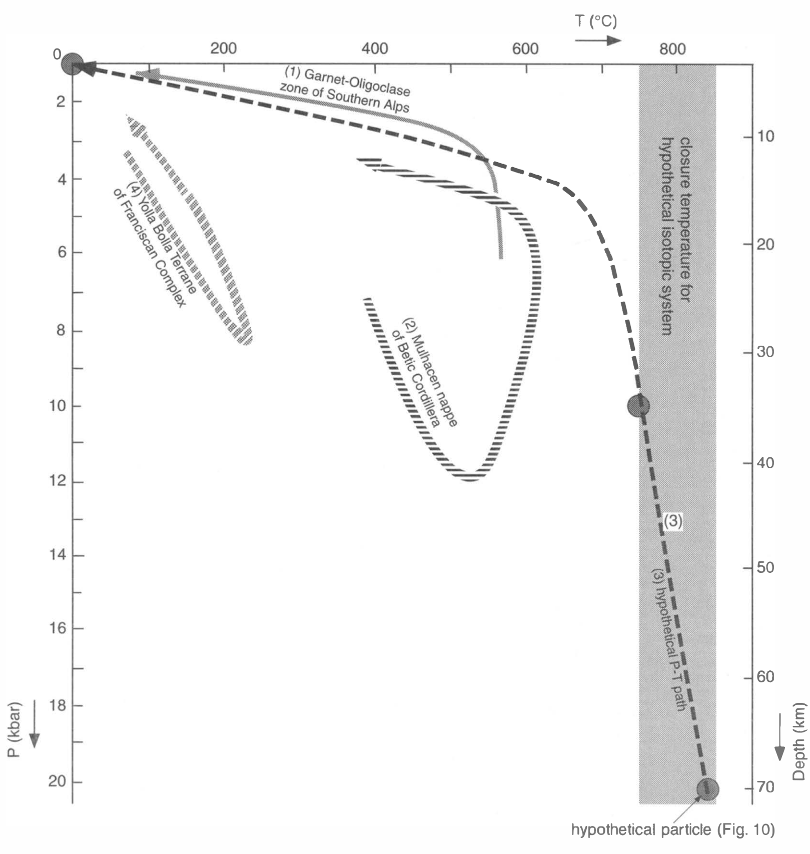

Fig. 9. Selected *P–T* paths from exhumation settings that are controlled (1) mainly by erosion, e.g. Southern Alps of New Zealand (Holm *et al.* 1989; Grapes & Wantanabe 1995), and (2) mainly by crustal extension, e.g. Mulhacen nappe of Betic Cordillera in southern Spain (Gomez-Pugnaire & Fernandez-Soler 1987). Path (3) shows a hypothetical isothermal decompression curve, which illustrates the problem of applying the 'closure-temperature concept' for estimating exhumation rates. Dark grey dot shows *P–T* evolution of the hypothetical particle shown in Fig. 10. The hairpin shape of path (4) from the Eastern Belt of the Franciscan subduction complex of California (e.g. Ernst 1993) is probably caused by slow exhumation in a subduction-zone setting.

paths like that for curve 4 in Fig. 9 from the Franciscan subduction complex have cooling rates that can be simply related to exhumation rates. This situation seems to hold only when exchumation rates are <*c.* 1 km Ma^{-1}.

P–t *data*

P–t information provides probably the most powerful tool for resolving exhumation rates. This is because metamorphic pressure can be directly related to depth by assuming an average density for the crustal column. *P–t* data are relatively easy to obtain in granitoids (e.g. magmatic pressure using the Al-in-hornblende barometer and magmatic age from U-Pb dating of zircon). In metamorphic rocks, the relation between pressure and time can only be determined by correlating barometric data to the thermal history data, as obtained from petrology and isotopic dating. Alternatively, dating of the crystallization age of minerals such as phengite, from which minimum metamorphic pressure can be obtained (Massone 1995), may provide a means for constraining a *P–t* history.

Reddy *et al.* (1999) showed that minimum pressure estimates for phengite crystallization show no correlation with age within a shear zone in the hanging wall of the Zermatt–Saas eclogites. The ages for phengite crystallization range from 36 to 45 Ma (Reddy *et al.* 1999). They concluded that the hanging wall of the shear zone was not exhumed while the shear zone was active (duration *c.* 9 Ma). These data, in conjunction with the break in pressure across the shear zone, allowed them to constrain a rate of 3.5 km Ma^{-1} for exhumation of the Zermatt–Sass eclogites by normal faulting.

Ultrahigh-pressure rocks

Surface exposures of UHP metamorphic rocks are a showcase for exhumation in the extreme. The mode of exhumation remains uncertain, but there seems to be a growing consensus that several processes, operating at different levels in the lithosphere, are involved. The evidence suggests that UHP metamorphism mainly occurs within collisional orogens, but where within the orogen? More specifically, does

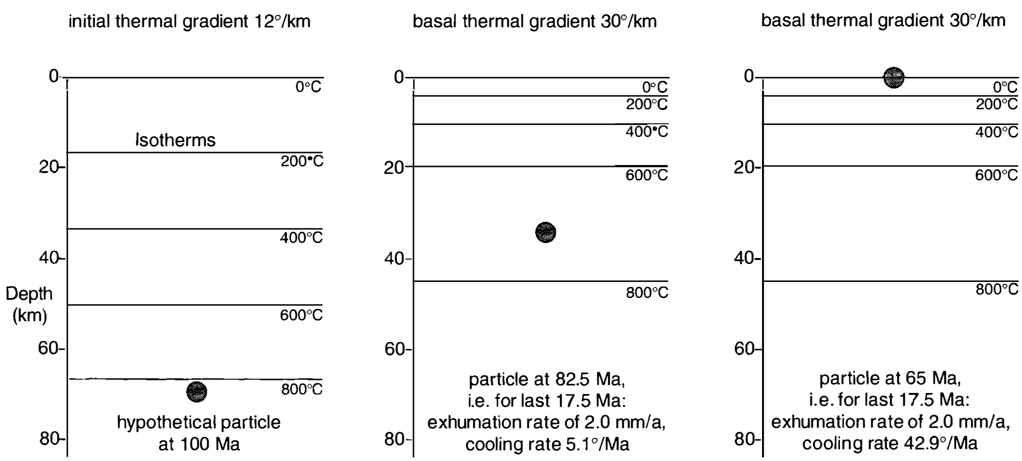

Fig. 10. Illustration of possible effects of rapid exhumation. A *P–T* path for the hypothetical particle considered in this example is shown in Fig. 9. (**a**) Depth arrangement of isotherms during rapid underthrusting of a particle to a depth of 70 km, assuming an initial thermal gradient of 12°C km^{-1}. (**b**) Accretion is followed by exhumation at a constant rate of 2 km Ma^{-1}, which initiates a new, relatively steep thermal gradient (i.e. rapid exhumation results in a very modest cooling rate of 5°C Ma^{-1} averaged over the first 17.5 Ma of exhumation). The development of the isotherms in this example has been modelled using a one-dimensional steady-state model where accretion is balanced by erosion. The model assumes that temperatures at the surface and at the base of a layer remain fixed at their set values of 0 °C and 840 °C, respectively. These set values can be viewed as describing a basal geotherm of 30°C km^{-1}. The thermal diffusivity (κ) in our example is 32 km^2 Ma^{-1}, which is typical for collisional belts such as the Himalayas (Zeitler 1985). For model details, see appendix of Brandon *et al.* (1998). (**c**) Final exhumation of particle at the same rate as in (b), but average cooling rate is now more than eight times higher as in (b) (43°C Ma^{-1}, averaged over 17.5 Ma).

metamorphism occur within the crust or the mantle?

Ernst *et al.* (1997) argues that ultrahigh-pressure metamorphism is caused by subduction of continental crust along an oceanic subduction zone. The continental crust is embedded in the oceanic lithosphere, so that slab pull is able to drag the crust down to depths of >100–125 km. The crust somehow detaches and returns to the surface, driven mainly by its strong positive buoyancy relative to the surrounding mantle.

This interpretation is certainly plausible, but it relies heavily on the early presence of an oceanic subduction zone to account for deep burial and metamophism. Early subduction of oceanic lithosphere is supported by the occurrences of ophiolitic rocks in some ultrahigh pressure metamorphic terrains (i.e. Zermatt–Sass zone described above). However, ultrahigh-pressure metamorphic rocks are generally found in Mediterranean-style continental collision zones with little to no evidence of a pre-collisional magmatic arc (Ernst *et al.* 1997). Thus, one is left to question if there was a well-organized pre-collisional oceanic subduction zone during UHP metamorphism. This problem has caused us to consider if it might be possible to form ultrahigh-pressure metamorphic rocks within the collisional orogen itself, and without appealing to an early oceanic subduction zone.

Orogenic root models

The pressures associated with UHP metamorphism seem incompatible with metamorphism in thickened continental crust of an average crustal density (2830 kg m^{-3}, Christensen & Mooney 1995). Assuming a reasonable metamorphic temperature of 800°C, this crust would have to be 100 km thick to produce the 29 kbar conditions needed to convert quartz into coesite and a thickness of >120 km, corresponding to pressures of >35 kbar, to stabilize diamond. At 100 km crustal thickness, the resulting orogenic topography would have a mean elevation of >10 km, which is twice the height of the highest orogenic topography on present Earth. It seems unlikely that such high topography could be supported for any geologically reasonable length of time (e.g. Bird 1991). Some other explanation is needed. We explore two ideas (Figs 11 and 12), both of which invoke a thicker and denser orogenic root, either due to the inclusion of lithospheric mantle (e.g. Molnar *et al.* 1993) or eclogitized lower crust (Dewey *et al.* 1993).

For the first option, the orogenic root is made up of crust and a fraction δ of the incoming lithospheric mantle (Fig. 11a). Deeper lithospheric mantle does not thicken and is thus passively depressed or subducted into the asthenosphere. UHP metamorphism is inferred to take place beneath the Moho, within the mantle part of the orogenic root. Some type of fault imbrication is required to interleave continental and oceanic crust rocks within this mantle root. This seems plausible because the lithospheric mantle would be expected to behave in a brittle fashion given the relatively low temperatures of typical UHP metamorphism (<800 to 900°C).

To explore the implications of the model, we assume the following simplified scenario. An incoming continental platform with an initial crustal thickness h_c = 40 km and a mean elevation z = 0 km (sea level), and a thickened crust H_c = 70 km within the orogen. The 70 km thickness would be representative of the Moho in a collisional orogen (Christensen & Mooney 1995). The initial thickness of the lithospheric mantle is set at h_m = 100 km (Molnar *et al.* 1993). The crust and lithospheric mantle are constrained to thicken by the same amount,

$$S_v = H_c/h_c = \delta H_m/\delta h_m = 1.75$$

assuming the representative values above for H_c and h_c. Assuming isostatic equilibrium, we can predict the maximum thickness of the orogenic root,

$$(H_c + \delta H_m) = S_v\,(h_c + \delta h_m),$$

the mean elevation of the orogen above sea level,

$$z = [((\rho_a - \rho_c)/\rho_a)\,h_c + ((\rho_a - \rho_m)/\rho_a)\,\delta h_m]\,(S_v - 1),$$

and the pressure at the base of the orogen,

$$P_{base} = (\rho_c\,h_c + \rho_m\,\delta h_m)\,S_v\,g,$$

where g is the acceleration of gravity.

These results are illustrated in Fig. 11b–d, using S_v = 1.75. About 10–40% of the incoming lithospheric mantle would have to be accreted to the orogenic root to get pressures sufficient for UHP metamorphism. The resulting mean elevation would be *c*. 3 km. For this model, the metamorphic pressure needed for UHP metamorphism is provided by the negative buoyancy of the mantle portion of the root. Note, however, that the amount of interleaved crustal rocks in the mantle part of the root must remain small to retain sufficient negative buoyancy within the root.

If this explanation is correct, we are still left to explain how the crustal rocks separate from the mantle part of the root and return to the surface. Houseman *et al.* (1981) and Molnar *et al.* (1993)

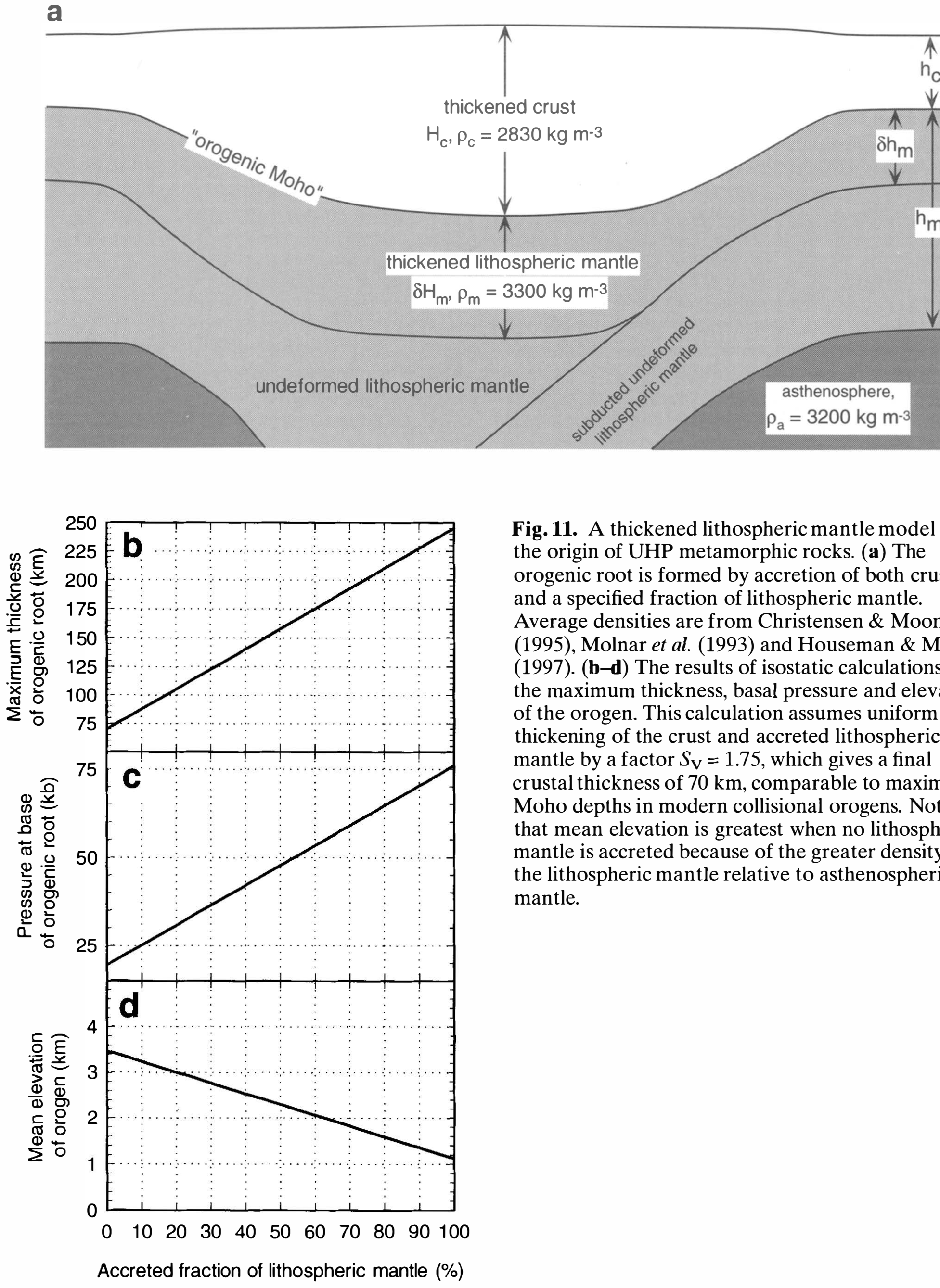

Fig. 11. A thickened lithospheric mantle model for the origin of UHP metamorphic rocks. (**a**) The orogenic root is formed by accretion of both crust and a specified fraction of lithospheric mantle. Average densities are from Christensen & Mooney (1995), Molnar *et al.* (1993) and Houseman & Molnar (1997). (**b–d**) The results of isostatic calculations for the maximum thickness, basal pressure and elevation of the orogen. This calculation assumes uniform thickening of the crust and accreted lithospheric mantle by a factor $S_V = 1.75$, which gives a final crustal thickness of 70 km, comparable to maximum Moho depths in modern collisional orogens. Note that mean elevation is greatest when no lithospheric mantle is accreted because of the greater density of the lithospheric mantle relative to asthenospheric mantle.

have argued that the gravitationally unstable lithospheric mantle in the root will ultimately detach and sink into the asthenosphere. The rise time for this detachment process depends on time constants for thermal relaxation of the root and for viscous flow associated with the Raleigh–Taylor instability (see Molnar *et al.* 1993 for a recent analysis). A similar rise time is probably associated with separation of the more buoyant UHP crustal rocks from the mantle lithosphere (Wallis *et al.* 1998). Buoyant rise or diapirism may account for how UHP rocks reach

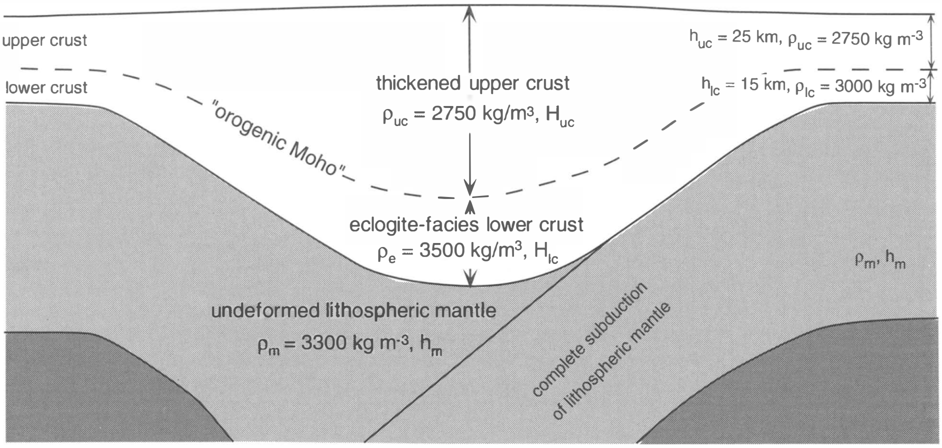

Fig. 12. An eclogitized lower crust model for the origin of UHP metamorphic rocks. See text for details.

the base of the crust, perhaps at the same time as the mantle root descends into the asthenosphere. This interpretation remains speculative but provides a stimulating view of how UHP crustal rocks might start to make their way back to the surface.

We consider a second option, that eclogization of the mafic lower crust provides the thickness and pressure needed for UHP metamorphism (Fig. 12). According to Christensen & Mooney (1995), the lower crust of the continents, from 25 to 40 km on average, is mainly mafic in composition with a density of2900–3100 kg m^{-3}. At depths >*c*. 50 km, corresponding to pressures >*c*. 14 kbar, these rocks would be converted to eclogite with an average density (ρ_e) of *c*. 3500 kg m^{-3} and P-wave velocities of *c*. 8 km s^{-1} (see Christensen & Mooney 1995, Table 4, p. 9775). In this case, the Moho would no longer mark the top of the mantle, but rather the transition into eclogite-facies mafic crustal rocks (Dewey *et al.* 1993). The high density cited above only applies to eclogite, and not to the more silicic coesite- and diamond-bearing rocks that have come to define the UHP metamorphic problem. The UHP silicic rocks are not eclogites. None the less, they belong to the higher pressure part of the eclogite facies, and can be referred to as the coesite or diamond subfacies. As in the previous interpretation, we must argue that the eclogite-rich crustal root would contain a small fraction of structurally interleaved silicic UHP metamorphic rocks.

To explore the implications of the eclogite model, we again start with a 40 km thick crust with a mean elevation at sea level, an upper crustal thickness h_{uc} = 25 km, and a lower mafic crustal thickness h_{lc} = 15 km. We examine the extreme case where the crust above the Moho is formed from the upper crust by itself. H_{uc} is set to 70 km, which we use again as our representative depth for Moho in a collisional orogen. The upper and lower crust are assumed to thicken by the same amount, which means that

$$S_v = H_{uc}/h_{uc} = H_{lc}/h_{lc} = 2.8$$

assuming the representative values for H_{uc} and h_{uc}. Isostatic balance gives the following equations for the maximum thickness of the orogenic root:

$$(H_{uc} + H_{lc}) = S_v\,(h_{uc} + h_{lc}),$$

mean elevation:

$$z = [((\rho_a - \rho_{uc}) / \rho_a)\, h_{uc} + ((\rho_a - \rho_{lc}) / \rho_a)\, h_{lc}]\,(S_v - 1) - ((\rho_e - \rho_{lc}) / \rho_a)\, S_v\, h_{lc}$$

and pressure at the base of the orogenic root:

$$P_{base} = [(h_{uc}\,\rho_{uc}) + (h_{lc}\,\rho_e)]\, S_v\, g.$$

For this model, our equations predict an orogenic root that is 112 km thick with an average elevation of only 1.5 km and a maximum pressure at the base of the root of 33 kbar. The pressure is sufficient for coesite stability but diamond would only be stable if temperatures were well below 700°C. A thicker orogen would be needed to account for the full range of pressures observed for UHP metamorphic suites but we cannot see how this model could be used to generate a thicker orogen without making the Moho deeper than 70 km.

For this model, the orogenic root is composed entirely of crust, but the ecologitized lower crust is gravitationally unstable and thus prone to detachment as described above for the thickened lithospheric mantle model. A critical problem is that the crustal root described here may be so weak that it could not persist for any significant time before detaching and sinking into the asthenosphere. Lithospheric mantle is stronger and should be able to persist in a thickened form for a longer period of time. Given this factor, plus the greater range of possible metamorphic pressures, the thickened lithospheric mantle model seems to provide a better explanation for the origin of UHP metamorphism. Note however that neither model is exclusive and that a mafic lower crust would become eclogitized if it was thickened to depths >*c.* 50 km.

Other unresolved issues

There are many additional unresolved issues regarding ultrahigh-pressure rocks, some of which are listed here.

(1) *What is the thickness of UHP units?* It has been proposed that ultrahigh-pressure rocks are typically found as relatively large (i.e. up to hundreds of km^2) internally coherent nappes, which are only a few kilometres thick (Ernst *et al.* 1997). Such a statement is certainly true for the well-mapped ultrahigh-pressure Brossasco-Isasca unit of the Dora Maira Massif which is sandwiched between lower pressure nappes (Chopin 1984; Chopin *et al.* 1991). However, in other UHP nappes the regional extent and the thickness of the nappes are poorly constrained. For example, diamond-bearing eclogites in the Erzgebirge of eastern Germany apparently represent very small lenses (< hundreds of metres in diameter) within high-pressure gneiss (Massone 1999). The average thickness of UHP nappes is crucial for understanding their thermal history and also their rheologic behaviour during exhumation.

(2) *What is the relationship of ultrahigh-pressure metamorphism to magmatism?* The high radiogenic heat production typical of continental rocks should lead to thermally-induced melting since ultrahigh-pressure metamorphism is generally at conditions above the 'wet' solidus for granitic melts (Huang & Wyllie 1975). The local presence of melt has been discussed by Schreyer *et al.* (1987), Schreyer (1995), Phillipot (1993) and Sharp *et al.* (1993) but clear evidence for melting remains scarce. Most people argue that the exhumation path of ultrahigh-pressure rocks is characterized by cooling during decompression (Roberto Compagnoni, pers. comm. 1996, 1997; Ernst *et al.* 1997). The exhumation-related cooling suggest that the ultrahigh-pressure rocks were thrust onto fairly cold foreland units, which were capable of cooling down hundreds of square kilometre-sized nappes quickly. Rapid exhumation and quick cooling might have prevented melting. However, the initial size of an UHP nappe is critical in this regard.

(3) *Is the preservation of ultrahigh-pressure assemblages in continental basement consistent with dry metamorphic conditions during exhumation?* A relatively dry granitic basement rock (e.g. Dora Maira Massif, Western Gneiss region) typically contains about 2 wt% structurally bound water. To transform dry mineral assemblages to high- and ultrahigh-pressure assemblages, one has to hydrate the rocks to stabilize minerals like talc and phengite in such rocks. The presence of a fluid phase during ultrahigh-pressure metamorphism has also been inferred from mass-transfer processes and fluid-inclusion evidence (Schreyer 1995; Harley & Carswell 1995; Phillipot *et al.* 1995). Oxygen-isotope studies in the ultrahigh-pressure rocks of Dabie Shan, China, show that the granitoids where heavily hydrothermally altered by near-surface waters before metamorphism (Rumble *et al.* 1998). This summary suggests that the continental basement is in general not dry before being converted to an ultrahigh-pressure rock. The preservation of ultrahigh-pressure assemblages during exhumation would then suggest that fluid was channelized in shear zones. Localized deformation and fluid flow may have allowed the preservation of ultrahigh-pressure assemblages in shielded blocks.

(4) *How much of the exhumation history of an ultrahigh-pressure terrain is preserved in its present structural and stratigraphic setting?* Standard structural field studies, in conjunction with *P–T* work, are capable of constraining aspects of the last 20–30 km of the exhumation path. Van der Klauw *et al.* (1997) and Stöckhert & Renner (1998) demonstrated, for instance, that quartz microfabrics in UHP rocks record greenschist-facies deformation. Structures that formed at deeper crustal levels are commonly thought to have been highly or completely obliterated. The preserved high- or even ultrahigh-pressure deformation relics are hard to relate to mapable large-scale nappe contacts or accretion-related structural discontinuities.

A remarkable feature of at least some UHP nappes is the lack of pronounced deformation. The virtually undeformed Variscan Brossasco granite of the Dora Maira Massif preserves its original igneous texture and its intrusive relationship with a surrounding metasedimentary unit (Biino & Compagnoni 1992). In Dabie Shan, some of the rocks preserve a pre-metamorphic hydrothermal alteration by near-surface meteoric waters (Rumble *et al.* 1998). In both cases, ultrahigh pressure metamorphism caused virtually no change to rock texture or isotopic composition. Stöckhert & Renner (1998) show that the undeformed Brossasco granite indicates that differential stress was too low to cause plastic flow during burial, accretion and exhumation. The only evidence for significant deformation by dislocation creep under ultrahigh-pressure conditions comes from ultrahigh-pressure eclogite from the Zermatt–Saas zone (omphacite microstructures reported by van der Klauw *et al.* 1997).

(5) *Are there transient accelerations in the rate of the processes involved in exhumation (e.g. Hill* et al. *1995)?* Many of the arguments that concern the sustainability of topography and the maintenance of steady-state geotherms, critically depend on the time constants of the processes involved. For example, it is relatively easy to produce a depressed geotherm in the overriding plate, above a subduction zone, as the result of large-scale overthrusting of a back-arc basin, providing an alternative explanation for the synchroneity of ophiolite emplacement and the formation of eclogites and blueschists (see Rawling & Lister this volume). Since relatively thick crust might exist for less than 1 Ma, rapid oscillations in tectonic mode may provide an explanation for many of the paradoxes outlined in the discussions above.

Concluding remarks – outstanding problems

In most active mountain belts, erosion and tectonism are dynamically coupled to the point where it may be difficult to separate cause and effect. This makes it difficult to distinguish between different exhumation processes. Nonetheless, exhumation typically occurs by multiple processes and there is a need to quantify the relative contributions of the different exhumation processes, using information from metamorphic petrology, isotope thermochronology, structural and kinematic analysis, synorogenic stratigraphy, geomorphology, and palaeo-elevation analysis.

To highlight some of our conclusion, we ask the following four questions.

(1) *How diagnostic are exhumation rates for distinguishing exhumation processes?* We believe that exhumation rates alone cannot distinguish between exhumation processes. Likewise, net exhumation rates do not supply much information on the rates of specific processes. The following example may help to illustrates this point: an erosion rate of 1 km Ma^{-1}, a fast slip rate of 3 km Ma^{-1} for a 20°-dipping normal fault (which equals an exhumation rate of 1 km Ma^{-1}), and a rate of 1 km Ma^{-1} for ductile thinning would combine to give a total exhumation rate of 3 km Ma^{-1}. The fast net rate is not useful in distinguishing between different exhumation processes.

Fast exhumation rates inhibits fast cooling. However, fast cooling may commonly follow rapid exhumation because of the upward advection of heat. There appears to be a serious need to constrain well-defined *T–t* and especially *P–t* paths for exhumed rocks.

(2) *How important is tectonic exhumation?* Tectonic exhumation, especially in the form of normal faulting has been recognized as a common factor in continental orogenesis. A widely held view is that early crustal thickening in convergent continental orogens will generally lead to normal faulting and crustal thinning. The first salient problem is to diagnose unequivocally horizontal crustal extension in orogens. Other problems associated with normal faulting appear to be the depth range of normal faults and the maximum throw on normal faults.

The few data on ductile thinning suggest that it is a slow exhumation process. More quantitative work on ductile thinning, especially in continental collision zones, is needed to demonstrate that this process contributes in a significant way to the exhumation of metamorphic rocks.

(3) *What is the role of erosion?* Erosion appears to be able to operate at very fast rates, perhaps as high as 15 km Ma^{-1}, given sufficient precipitation, steep terrain, and comparable uplift rates. There is no reason to indicate that these rates could not be sustained for long periods of time, as long as uplift rates continued to match erosion rates and climate conditions remained favourable for fast erosion. Alpine glaciation appears to be the most aggressive agent of erosion and one that is particularly sensitive to global climate. In this regard, the relatively high sediment production rate of the Quaternary may be the result of a cooler climate and more extensive alpine glaciation. An outstanding problem is that much of our current understanding of erosion rates is based on relatively short records. There is a serious need for better long-term estimates using sediment inventories or thermochronometry.

(4) *Where do UHP rocks form and how are they exhumed?* Ultrahigh-pressure rocks occur only in collisional belts, but they otherwise appear to form below the seismically determined Moho. Coesite-bearing ultrahigh-pressure rocks can form in the root of a highly overthickened

crust when large parts of the more mafic lower crust have been eclogitized. We have shown an example where eclogitized lower crust would be placed beneath the Moho in which case the crust might be about 110 km thick (with the Moho at a depth of *c.* 70 km). The predicted mean elevation would only be about 1.5 km. Ultrahigh-pressure rocks may also form within a mantle-rich orogenic root. The involvement of lithospheric mantle limits the mean elevation of the orogen to about 3 km. The question that remains unanswered is how do these rocks make their way back to the surface?

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