PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

The Uplift of High-Pressure--Low-Temperature Metamorphic Rocks [and Discussion]

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The uplift of high-pressure–low-temperature metamorphic rocks

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The uplift of high-*P*–low-*T* metamorphic rocks has been attributed to buoyancy, diapirism, or hydrodynamically driven return flow. Buoyancy forces can return material subducted into the mantle only if subduction slows or ceases, reducing the downward traction. The buoyancy forces will be reversed within the crust, because of the increased density of high-*P* assemblages, and therefore can not cause the subducted material to rise beyond the base of the crust. Diapirism and hydrodynamic flow processes require a low-density, low-viscosity matrix, and can only explain the emplacement of relatively small bodies of high-*P* rock entrained in the flowing material.

The tectonic setting of coherent regional high-*P*–low-*T* terrains can be explained in terms of the mechanical behaviour of an accretionary wedge with negligible yield strength, where underplating is the dominant mode of accretion. Underplating thickens the wedge from beneath and increases its surface slope. This causes the upper part of the wedge to extend horizontally, even though convergence is continuing. Continued underplating beneath and extension above can allow the oldest high-*P* rocks to rise to within reach of a moderate amount of erosion on a time scale of the order of 10 Ma. As long as subduction continues beneath the wedge, the geothermal gradient will not relax to a normal value.

This process explains (a) the evidence that high-*P*–low-*T* rocks are commonly uplifted while convergence is continuing; (b) the absence in many cases of significant overprinting by higher-*T* assemblages; (c) the position of the oldest and highest pressure rocks in the upper rear of orogenic wedges; (d) the lack of adequate tectonic thicknesses of overlying rock to explain the metamorphism; and (e) the common occurrence of post-metamorphic faults that excise parts of the metamorphic zonation.

Introduction

Most Phanerozoic orogenic belts include regions that have been affected by metamorphism under pressure–temperature conditions that indicate abnormally low geothermal gradients. This is a predictable consequence of convergent margin tectonics: the cold subducting lithospheric plate lowers the heat flow in the accretionary wedge above it (Oxburgh & Turcotte 1970), and underthrusting and shortening in the crustal rocks also temporarily depress the isotherms (England 1978). The main problem posed by these belts is the mechanism by which rocks metamorphosed at the high-pressure end of the facies series are uplifted and exposed at the surface.

The uplift of the low-pressure end of the series can reasonably be attributed to a combination of erosion and isostatic uplift consequent upon erosional unroofing. Rocks metamorphosed at, let us say, 5 kbar† (equivalent to a depth of about 18 km), in an orogenic belt initially 50 km thick, should be returned to the surface by normal rates of erosion in a few tens of millions of years at most. Assuming that the thermal gradient returns to a 'normal' continental value

† 1 bar = 10^5 Pa.

of about 20 °C km^{-1} during this uplift history, the metamorphic temperature is unlikely to exceed 300 °C at any time, and prehnite–pumpellyite or lawsonite–albite assemblages should be preserved.

This mechanism is far less credible for glaucophane–schist and eclogite–facies rocks. Regionally extensive terrains in California and the Alps, for example, have been metamorphosed at pressures in the range 7–16 kbar, requiring depths of burial between 25 and 60 km. This is comparable with the thickness of the continental crust itself in orogenic regions, so simple isostatic uplift and erosion are unlikely ever to expose such rocks. An additional problem is that if a normal geothermal gradient is established during the uplift history, high-pressure rocks are likely to pass through temperatures of 400–600 °C, so that the high-pressure assemblages should be largely overprinted by greenschist or amphibolite–facies metamorphism.

Perhaps the most crucial problem concerns the timing of the uplift history, however; and this can best be illustrated with two examples, from the European Alps and from California. In both cases, regional terrains of high-pressure rocks can be shown to have reached relatively shallow levels while plate convergence was continuing, and before the region became widely emergent or subject to rapid erosion.

In the Alps, the Sesia Zone is a large slab of continental crustal material that was subducted and metamorphosed under eclogite–facies conditions at a pressure of about 15 kbar (Compagnoni *et al.* 1977; Lardeaux *et al.* 1982) between 110–130 Ma (Hunziker 1974). Dating of lower-pressure assemblages formed during the uplift history suggests that it had reached relatively shallow depths (i.e. less than 25 km) by about 60 Ma, and less than 15 km by about 40 Ma (Rubie 1984), when it was affected by the regional meso-Alpine moderate $P:T$-ratio metamorphism. Therefore, 30–40 km of uplift occurred during the long period of subduction and accretion of oceanic crust and microcontinental fragments that predated the main Alpine collision. This began in Oligocene time, with the underthrusting of the Helvetic continental margin (Milnes 1978; Hsü 1979) and was accompanied by widespread emergence and erosion. There is no sedimentological evidence, however, for the erosional stripping of such huge thicknesses of rock from the southern side of the Alps in pre-Oligocene time, as would be required were uplift primarily a consequence of erosion.

In the Franciscan Complex of California, metabasalt and metagreywacke in the eastern (Yolla Bolly) belt of the northern Coast Ranges were affected by regional blueschist–facies metamorphism (Blake *et al.* 1967). The highest pressure rocks were metamorphosed at about 10 kbar in late Jurassic time, around 150 Ma (Suppe 1973). These rocks had returned to fairly shallow levels by late Cretaceous time, while subduction was continuing, as jadeite-bearing metagreywacke is tectonically interleaved with early to mid-Cretaceous sediments carrying lower-pressure assemblages, including pumpellyite, or lawsonite + albite (Suppe 1973; Cowan 1974; Worrall 1981). The presence of pebbles of reworked glaucophane schist in late Cretaceous Franciscan conglomerate (Cowan & Page 1975) supports this conclusion. There is, however, no evidence that 20 km or more of rock had been removed by erosion from above the Franciscan by late Cretaceous time. Sedimentological studies in the adjacent Great Valley terrain suggest that the Franciscan Complex formed a submerged ridge until at least the late Cretaceous, and was not a source of detritus until Miocene time (Ingersoll 1979).

We need, therefore, to find an uplift mechanism that does not involve extensive erosional unroofing, that can proceed during active convergence, and that protects the high-pressure rocks from the effects of a return to normal geothermal gradients.

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Uplift mechanisms

Most explanations for the uplift of high-pressure–low-temperature metamorphic rocks depend on the concept of a return flow, such that deeply underthrust material returns to the surface along roughly the same route by which it descended. Ernst (1975, 1984), for example, suggested that buoyant forces acting on individual slabs of metamorphic rock would be sufficient to drive them back up the subduction zone. A buoyant force would certainly act on such a slab if it were subducted into the mantle, but it would have to overcome both the resistance to motion on its upper surface, and the downward traction exerted by the subducting plate below it (figure 1). At shallow levels the buoyancy force is reversed, because high-pressure rocks are denser than their unmetamorphosed equivalents.

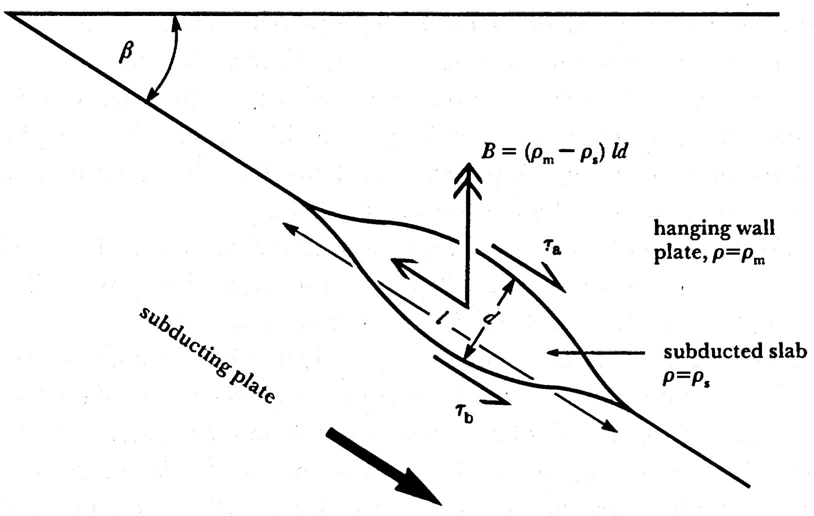

FIGURE 1. Forces on a slab of density ρ_s subducted into mantle of density ρ_m. The updip component of the buoyancy force B is resisted by the downward traction τ_b exerted by the subducting plate beneath, and the resistance τ_a to upward motion by the hanging wall. The slab will rise if $(\tau_a + \tau_b)\, l < B \sin\beta$ or $(\tau_b + \tau_a)/d(\rho_m - \rho_s) \sin\beta < 1$.

As shown in figure 1, the condition for the slab to rise is approximately

$$l(\tau_b + \tau_a) < B \sin\beta. \tag{1}$$

B is the buoyancy force, given by

$$B = ld(\rho_m - \rho_s), \tag{2}$$

β is the dip of the subduction zone, τ_b the downward shear stress exerted by the subducting plate beneath, τ_a the resistance to upward motion imposed by the overlying material, ρ_m and ρ_s the densities of mantle and subducted material, and l and d are the length and thickness of the subducted slab. The condition for uplift is therefore $\Omega < 1$, where

$$\Omega = \frac{(\tau_a + \tau_b)}{d(\rho_m - \rho_s)\sin\beta}. \tag{3}$$

This condition clearly did not hold while the slab is being carried down and we can state that, at the time it ceased to descend further,

$$B \sin\beta < \tau_b\, l, \tag{4}$$

hence $\Omega > 1$.

Uplift can therefore only occur if either τ_b decreases significantly (by a slowing down or cessation of subduction, for example), or if d or β increase. Buoyancy is probably required to explain the return to crustal levels of rocks subducted to depths of as much as 90 km in the mantle, such as the coesite + pyrope assemblage in the Dora Maira massif of the western Alps (Chopin 1984), but it cannot explain the emplacement of high-pressure rocks at high levels in the crust.

A rather different concept of buoyant diapirism was invoked by England & Holland (1979) to explain the presence of eclogite blocks ($P \approx 18$ kbar) in calc-schists in the eastern Alps. They suggested that the high-density blocks were entrained by a return flow in highly mobile, relatively low-density carbonate rocks. More recently, Cloos (1982) proposed a hydrodynamically driven return flow in Franciscan mud-matrix mélanges to explain the emplacement of tectonic blocks of eclogite and high-grade blueschist. Unlike Ernst's hypothesis, these ideas are only applicable to relatively small blocks of high-pressure rock immersed in a low-density low-viscosity matrix, and can not account for the uplift of regionally extensive terrains.

The purpose of this paper is to show how some of the recently developed ideas about the mechanics of convergent orogenic belts can explain the uplift of regional terrains of high-P–low-T metamorphic rock. Supporting evidence for this mechanism comes from some commonly observed features of these terrains, as follows.

1. High P : T-ratio terrains commonly show an overall zonation, with the oldest and highest pressure rocks at the top of the structural pile (Ernst 1975; Platt 1975). This is true of the Franciscan Complex, the Alps, and the Sanbagawa terrain of Japan.

2. The highest-pressure rocks are commonly directly overlain by nappes or thrust-sheets that lack high-pressure metamorphism, and are insufficiently thick to explain the metamorphism beneath them. At the top (east) of the Franciscan Complex, jadeitic metagreywackes are overlain by the Coast Range ophiolite. This shows only zeolite–facies metamorphism (Bailey and Jones 1973), and together with the overlying Great Valley Sequence sediments, is about 15 km thick (figure 2). In the central and eastern Alps the glaucophane–schist to eclogite–facies rocks of the Pennine and Sesia Zones, metamorphosed at pressures of 10–16 kbar, are overlain by the Austro-Alpine nappes (figure 3). These have a maximum composite structural thickness of 15–20 km (Oxburgh *et al.* 1971), and were affected by greenschist to lower amphibolite facies metamorphism at about 6 kbar near the base of the nappe pile (Hawkesworth *et al.* 1975). In the Betic Cordillera of southern Spain, glaucophanic and eclogitic rocks of the Nevado-Filabride Complex, metamorphosed at 10–12 kbar (Diaz de Federico *et al.* 1978) are overlain by the Higher Betic Nappes (figure 4). These have a present-day composite structural thickness of about 5 km, and the metamorphic pressure at the base did not exceed 6 kbar (Platt *et al.* 1983).

3. As implied by point 2, the upper boundary of the high-pressure terrains is commonly a post-metamorphic fault that has cut out a significant thickness of the original overburden. In the California Coast Ranges this is the Coast Range thrust.

4. The high-pressure terrains themselves are cut by post-metamorphic faults. These are commonly regionally subhorizontal, and define large-scale tectonic sheets of rock with differing grades, ages, and histories of metamorphism. These large-scale structures formed during the uplift history, and are likely to be causally related to it. Many major nappe contacts in the internal zones of the Alps have this character (Compagnoni *et al.* 1977); and similar faults occur in the Franciscan Complex (Suppe 1973; Platt 1975), and in the Betic Cordillera (Platt *et al.* 1983).

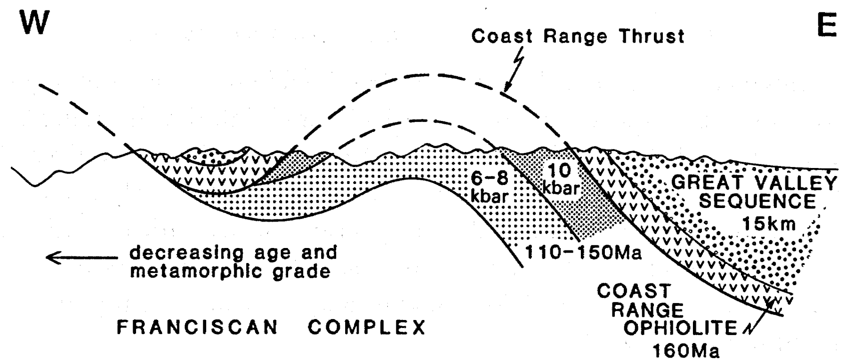

FIGURE 2. Schematic section across the eastern zone of the Franciscan Complex in the California Coast Ranges; based on relations documented by Bauder & Liou (1979), Cowan (1974), Ernst (1965, 1970, 1971) Suppe (1973) and Worrall (1981). Note the interleaving of rocks of different age and metamorphic grade across post-metamorphic faults, and the juxtaposition of high-pressure rocks with unmetamorphosed ophiolite across the Coast Range Thrust.

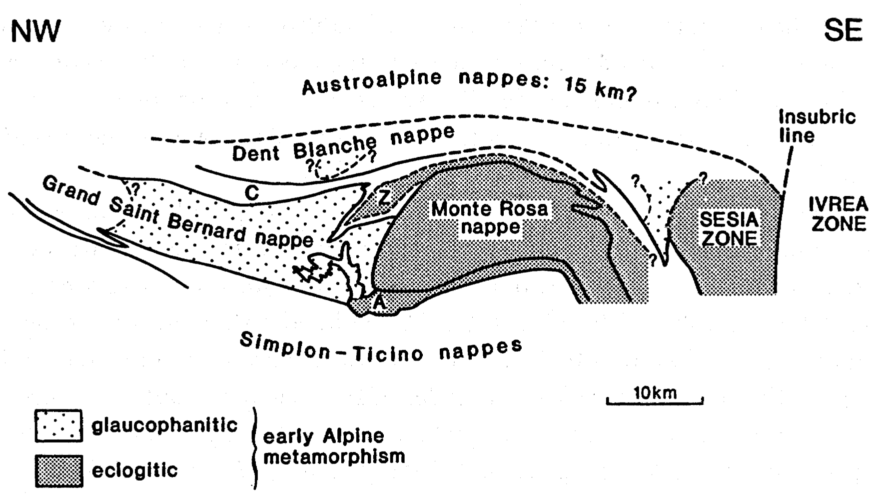

FIGURE 3. Schematic section across the internal (SE) zone of the Swiss–Italian Alps, modified after Compagnoni *et al.* (1977) and Milnes *et al.* (1981). A, Antrona ophiolite unit C, Combin unit; Z, Zermatt ophiolite unit. High-pressure assemblages in the Monte Rosa and Antrona units have been strongly overprinted by the meso-Alpine metamorphism. Facies boundaries are poorly constrained.

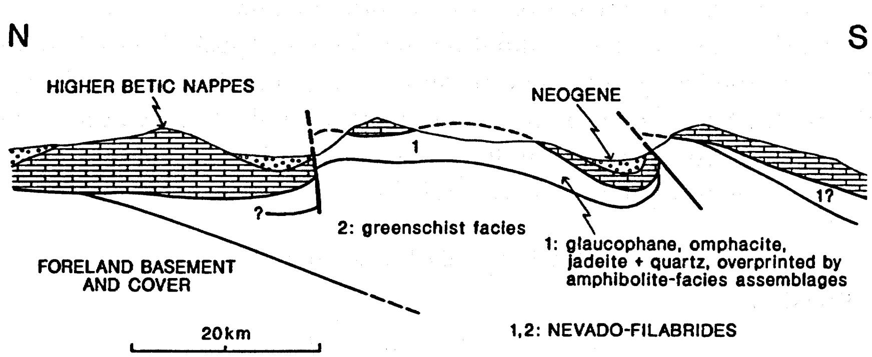

FIGURE 4. Schematic section across the Betic Zone of southern Spain, modified after Platt *et al.* (1983).

Dynamics of convergent orogenic wedges

Convergent orogenic belts can probably all be described in terms of a roughly wedge-shaped prism of deformed crustal rock, resting on a relatively rigid slab that is sliding beneath it, and supported by a buttress behind. This applies equally to accretionary wedges above oceanic subduction zones and mountain belts forming during at least the early stages of continental collision. The down-going slab consists of lithospheric mantle, with or without a veneer of crystalline oceanic or continental crust. The buttress, or hinterland, may be a continent or an island arc; it may not be absolutely rigid, but it usually deforms at a substantially lower rate than the material in the wedge.

The orogenic wedge, although it may consist of thrust sheets or slices of folded or otherwise deformed material, behaves on the large scale as a single, mechanically continuous unit. This concept has been established by a series of recent analyses of thrust belts (Price 1973; Elliott 1976*a*, *b*; Chapple 1978; Davis *et al.* 1983; Stockmal 1983). It is supported by the observational evidence that thrust faults are mechanically linked along and across strike via folds (Dahlstrom 1970; Elliott 1976*b*), branch-lines (Hossack 1983) and tip-line structures (Coward & Potts 1983), and that they initiate in a regular sequence (Dahlstrom 1969; Boyer & Elliott 1982). This concept allows us to apply continuum theory to the mechanics of the wedge, even though the deformation is discontinuous in detail. An analogy can be made with the plastic deformation of a single crystal, that proceeds at the molecular scale by the linked sequential rupture of bonds.

In orogenic wedges that are thick enough to induce regional metamorphism, ductile deformation is important, being favoured by relatively elevated temperatures, and by the chemical and mechanical effects of water released by dehydration reactions (Blacic 1975; Rutter 1983; Atkinson 1980; Etheridge 1983). Calculated strain rates for processes such as pressure-solution suggest that at 300 °C, elastic stresses of more than a few tens of bars can be relieved by ductile flow on a time scale of the order of 10^3 a (Rutter 1976). The orogenic wedge is therefore unlikely to exhibit a long-term yield stress, and analyses based on the concept of plastic (Chapple 1978; Stockmal 1983) or brittle (Davis *et al.* 1983) yield are not directly applicable. The analysis discussed here is a general one, and makes no rheological assumptions other than the absence of a long-term yield stress. For this reason it cannot be used to predict precise relations among specified mechanical variables, but it does allow us to make qualitative predictions about the overall behaviour of the wedge.

The geometrical and mechanical aspects of a simplified accretionary wedge are shown in figure 5. Its geometry is a consequence of the equilibrium among the stresses acting within it. These are produced by (1) the longitudinal normal force or 'push' exerted by the buttress, which contributes to the longitudinal deviatoric stress τ_{xx}; (2) the shear stress τ_b exerted on the base of the wedge by the downgoing slab beneath it, and (3) gravitational body forces. These terms can be related using the equilibrium equations for creeping flow, as follows.

In the *x*-direction (by using the engineering sign convention for stresses)

$$\partial\sigma_{xx}/\partial x + \partial\tau_{yx}/\partial y + \rho g \sin\beta = 0, \tag{5}$$

and in the *y*-direction,

$$\partial\sigma_{yy}/\partial y + \partial\tau_{xy}/\partial x - \rho g \cos\beta = 0. \tag{6}$$

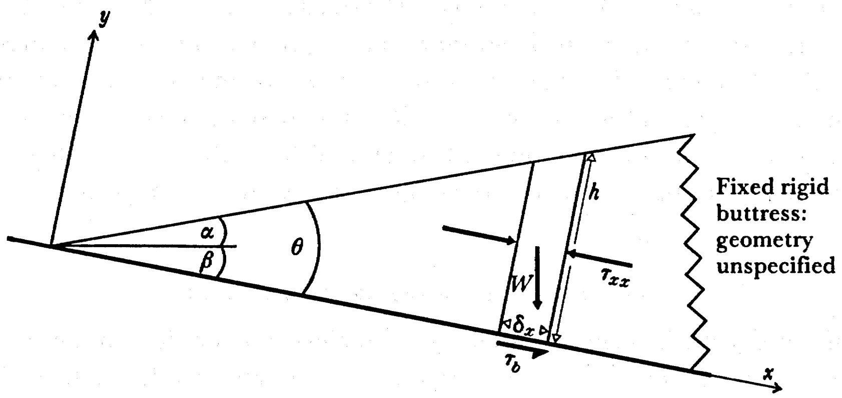

FIGURE 5. Dynamics of an accretionary wedge. See text for definition of symbols.

These expressions have to be corrected for the buoyancy effect of water if part or all of the wedge is submerged. $\partial\tau_{xy}/\partial x$ is likely to be very small compared with $\partial\sigma_{yy}/\partial y$, so (6) reduces to

$$\partial\sigma_{yy}/\partial y = \rho g \cos\beta. \tag{7}$$

Integrating with respect to y,

$$\sigma_{yy} = \rho g(y-h)\cos\beta. \tag{8}$$

Now $\sigma_{ij} = p + \tau_{ij}$, where p is the mean stress and τ_{ij} is the deviatoric component. For plane stress, $\tau_{yy} = -\tau_{xx}$, so from (8),

$$\sigma_{xx} = \rho g(y-h)\cos\beta + 2\tau_{xx}. \tag{9}$$

Substituting (9) in (5) and integrating with respect to y,

$$-\frac{\partial}{\partial x}\left(\tfrac{1}{2}\rho g h^2 \cos\beta\right) + 2\frac{\partial}{\partial x}\int_0^h \tau_{xx}\,\mathrm{d}y + \tau_\mathrm{b} + \rho g h \sin\beta = 0. \tag{10}$$

Using small angle approximations, and taking a depth-averaged value for τ_{xx}, this simplifies to

$$\tau_\mathrm{b} = \rho g h\alpha - 2\tau_{xx}\theta - 2h\partial\tau_{xx}/\partial x. \tag{11}$$

This equation states that the downward traction τ_b exerted on the wedge by the slab sliding down beneath it is balanced by three terms. The first is a gravitational term produced by the surface slope of the wedge, and is identical with the stress that drives the flow of glaciers and ice sheets. The second is produced by the longitudinal stress τ_{xx} acting in a tapered wedge, and depends on the angle of taper, θ. The third results from the gradient $\partial\tau_{xx}/\partial x$ in the longitudinal deviatoric stress.

This equation cannot be solved in the general case without specifying a constitutive flow law for the material in the wedge, which would constrain τ_{xx} and $\partial\tau_{xx}/\partial x$, but we can use it to define a potentially stable configuration, which the wedge will tend to approach. The longitudinal deviatoric stress τ_{xx} will cause the wedge to extend or shorten horizontally at a rate $\dot{\epsilon}_{xx} = f(\tau_{xx})$. If all other influences on the wedge are kept constant, it will tend to deform internally until τ_{xx} has been completely relaxed, and it will then cease to deform. The condition for this stable state is, therefore, that $\dot{\epsilon}_{xx} = 0$, and hence $\tau_{xx} = 0$ and $\partial\tau_{xx}/\partial x = 0$. Equation (11) then reduces to

$$\tau_\mathrm{b} = \rho g h\alpha. \tag{12}$$

What this means is that a wedge composed of material lacking a finite yield stress will try to relax to a configuration such that the gravitational forces generated by its surface slope just balance the basal traction. The 'push' stresses exerted by the rear buttress then drop to lithostatic. In this configuration, the wedge will tend to slope away from the direction of subduction at an angle that increases towards the tip of the wedge. The angle of taper, θ, will depend on the density of the wedge and the degree of isostatic compensation.

Uplift of high-pressure rocks

An orogenic wedge is, in general, unlikely to achieve the stable configuration described above, because accretion of material brought in by the subducting plate will continually change its geometry. The concept is useful, however, because it allows us to predict an overall pattern of deformation in the wedge that results from accretion, and that tends to restore its geometry to the stable configuration. We can examine this by looking at the effects of two end-member types of accretion.

Accumulation of material by thrust imbrication at the wedge front (*frontal accretion*) will tend to lengthen the wedge, and decrease α. This will lead to longitudinal deviatoric compression in the wedge, so that it will shorten internally and restore α to the stable value. We should, therefore, expect a reactivation of thrusts in the interior of the wedge, or the initiation of new thrusts, backthrusts, or folds.

An alternative possibility is that material may be carried down beneath the wedge and accreted to its underside (*underplating*). This is the only effective way of subjecting supracrustal materials to high-P–low-T conditions rapidly enough to explain the early onset of this type of metamorphism in accretionary wedges. Underthrusting and underplating have been well documented along currently active convergent margins, both by seismic methods (see, for example, Westbrook *et al.* 1982) and based on structural and mass-balance considerations (Watkins *et al.* 1981; Platt *et al.* 1985; Walcott, this symposium). Underthrusting is presumably a consequence of differential lithification and strong bedding-parallel anisotropy in the incoming sediment layer; the subsequent detachment and accretion (underplating) of this material may result from changes in its mechanical properties with increasing pressure and temperature. Acting alone, underplating will tend to thicken the wedge and increase α. If α exceeds the stable value, the wedge will be thrown into longitudinal deviatoric tension, and will start to extend internally (figure 6*a*). It is this conclusion that is significant to the problem of the uplift of high-pressure metamorphic rocks. As long as underplating continues, the overlying wedge will go on extending. Material underplated at an early stage will be uplifted as a result of accretion of material below and extension and thinning of the material above. Eventually, the oldest underplated material will itself be involved in extensional tectonics (figure 6*b*), and compressional structures produced by underplating will be overprinted by listric normal faults or fabrics produced by horizontal ductile extension.

This description of wedge dynamics therefore predicts a possible pattern of material flow within the wedge (figure 7) that explains the uplift of high-pressure rocks towards the surface, and within reach of a reasonable amount of erosion. It develops the concepts of Platt (1975), Cowan & Silling (1978) and Emerman & Turcotte (1983), all of whom predicted an upward flow at the rear of an accretionary wedge, but did not provide an adequate explanation for the removal of material from above the high-pressure rocks, so that they could approach the

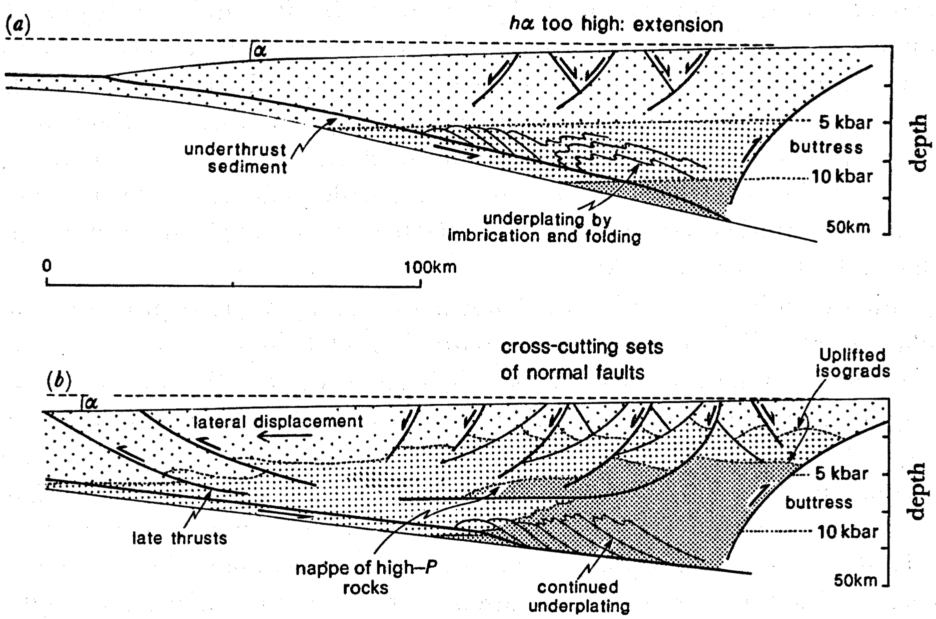

Figure 6. Evolutionary model of an accretionary wedge, assuming underplating as the dominant mode of accretion. (a) Early stage. Underplated sediment is metamorphosed at moderate to high pressure, and thickens the rear of the wedge. Extension is required to maintain a stable configuration. (b) Continued underplating at depth and extension above allows the previously underplated material to rise towards the surface. The early isograds are uplifted and disrupted by extension and transport towards the front of the wedge.

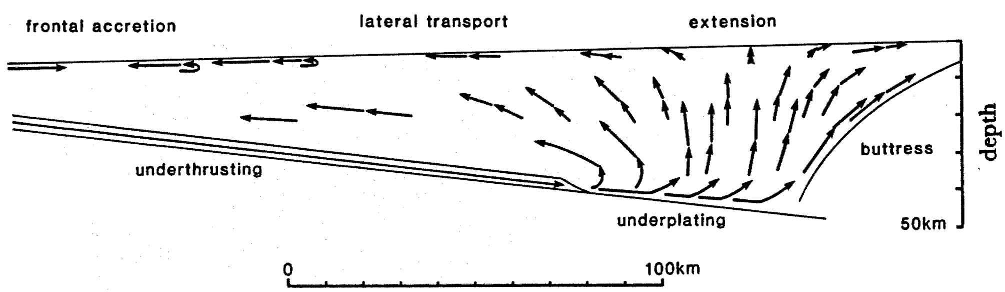

Figure 7. Qualitative prediction of the pattern of material flow in an accretionary wedge. The arrows indicate velocities relative to the buttress. Based on sequential constructions assuming underplating as the dominant mode of accretion, the maintenance of a stable profile as defined in this paper, and conservation of cross-sectional area.

topographic surface. The only assumptions are that underplating can be the principal mode of accretion for long periods of time, and that the wedge as a whole has a negligable long-term yield strength. Note that the pattern of flow in figure 7, which superficially resembles diapirism, is not driven by buoyancy contrast. It is a consequence of underplating at depth, and extension of the wedge as a whole driven by the forces acting on the whole system.

The driving forces for extension are relatively low; the upper limit is provided by the differential load associated with the vertical relief of the wedge. A submerged wedge with a relief of 6 km could produce a horizontal tensional deviatoric stress of not more than 450 bar. Whether extension could occur sufficiently rapidly to account for the uplift of high-pressure rocks will depend on the following factors: (a) the rate of underplating; (b) the basal shear stress; (c) the rheological properties of the wedge material; and (d) the surface relief. The latter will depend on the thickness of the wedge, its bulk density, and the degree of isostatic compensation. High-pressure metamorphism will increase the bulk density of the wedge and decrease the surface relief, for example. Density variations *within* the wedge will have little effect, however, as flow is not driven directly by buoyancy. The variables listed above cannot be estimated precisely enough to calculate rates of flow; but it is worth noting that at a strain rate of 10^{-14} sec^{-1}, which is reasonable for ductile deformation by pressure-solution processes under deviatoric stresses of the order of 100 bar (Rutter 1976), the cycle of uplift shown in figure 6 could take place in 10 Ma.

The hypothesis outlined above has the following advantages. (1) It provides a mechanism for the uplift of high-*P*–low-*T* rocks during continuing subduction. This is consistent with the evidence on timing summarized earlier; and it helps to explain the absence of widespread overprinting by higher-temperature assemblages. (2) It predicts the part of the wedge where these rocks are likely to be exposed (in the upper rear), and that they are likely to be among the oldest components of the wedge. (3) It predicts the existence of extensional faults above and within the high-pressure terrain. This accounts for the common occurrence of major low-angle faults in such terrains that cut out parts of the metamorphic zonation, and place lower-pressure rocks directly on high-pressure ones. Many of these faults may be the remains of early thrusts that have been reactivated along part of their trajectory by listric normal faults. The way in which this process may have acted to allow the uplift of high-pressure rocks in the Alps is illustrated in figure 8.

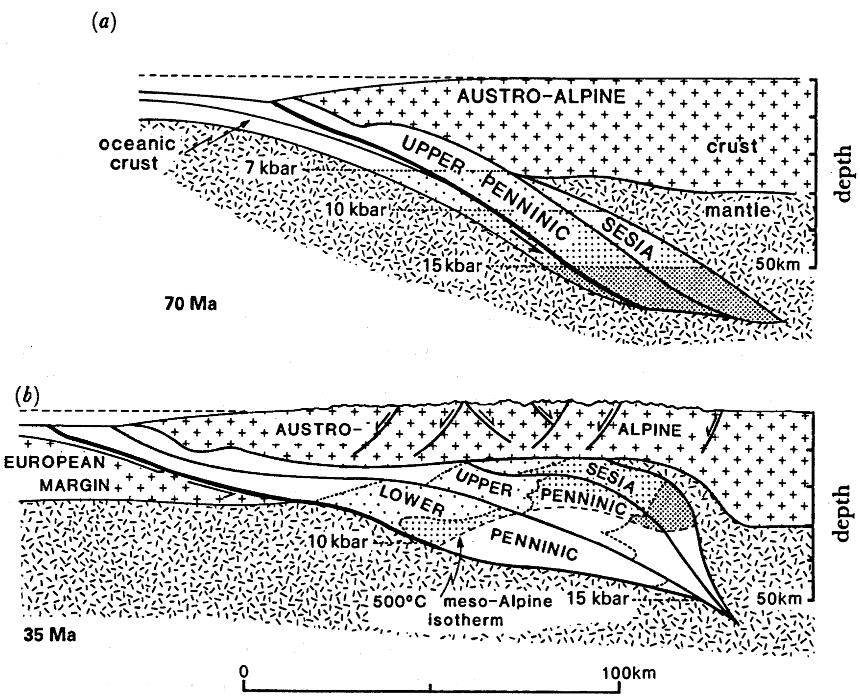

FIGURE 8. Tectonic evolution of the central Alps. (*a*) Late Cretaceous. Upper Penninic and Sesia Zone rocks have been subducted to depths of up to 70 km and metamorphosed. Subduction of oceanic crust continues beneath them, maintaining low geothermal gradients. (*b*) Early Oligocene. Continued underplating of Lower Penninic units, and extension in the overlying Austro–Alpine and Upper Penninic rocks, has allowed the high-pressure rocks to rise towards the surface. Note offset of isograds along reactivated nappe contacts. Decrease of subduction rate in early Tertiary time has allowed the temperature to rise, causing meso-Alpine metamorphism: this obliterates high-pressure assemblages above 500 °C.

Until recently there has been little documentation of extension in the rear of currently active convergent orogens. The rear portions of many active accretionary wedges are either submerged or covered by sedimentary basins. Normal faulting in the upper part of the Hikurangi accretionary prism in the North Island of New Zealand has been documented by Walcott (this symposium), however, who regards it as an effect of underplating at depth. The pattern of extension is entirely consistent with that predicted here, and underplating is strongly suggested by Walcott's mass-balance calculations. This therefore provides striking actualistic support for the description of wedge kinematics presented here. Extensional structures, possibly associated with underplating beneath the frontal part of the Nankai trough accretionary wedge, have also been documented by Leggett *et al.* (1985). Some active collisional orogens also show signs of extension: the Andes (Dalmayrac & Molnar 1981; Jordan *et al.* 1983) and the Himalaya (Burchfiel & Royden 1985) are notable examples.

Metamorphism and deformation

One of the implications of the hypothesis described above is that the uplift of high-pressure metamorphic rocks is achieved by a tectonic process involving substantial horizontal extensional strain. This extension can occur either by ductile processes or by normal faulting. Horizontal ductile extension produces a flat-lying foliation, and if imposed on steeply dipping layering produced by earlier compressional deformation, it could produce recumbent folds. Structural and metamorphic studies in the Alps, for example, have demonstrated that substantial ductile deformation, producing a flat-lying foliation and recumbent folds, occurred after the high-pressure metamorphic peak (Platt & Lister 1985*a*), and was synchronous with and related to the uplift process.

Normal faults in a metamorphic environment are likely to be listric, curving into the horizontal where they pass into a zone of distributed shear. They may not be identifiable as

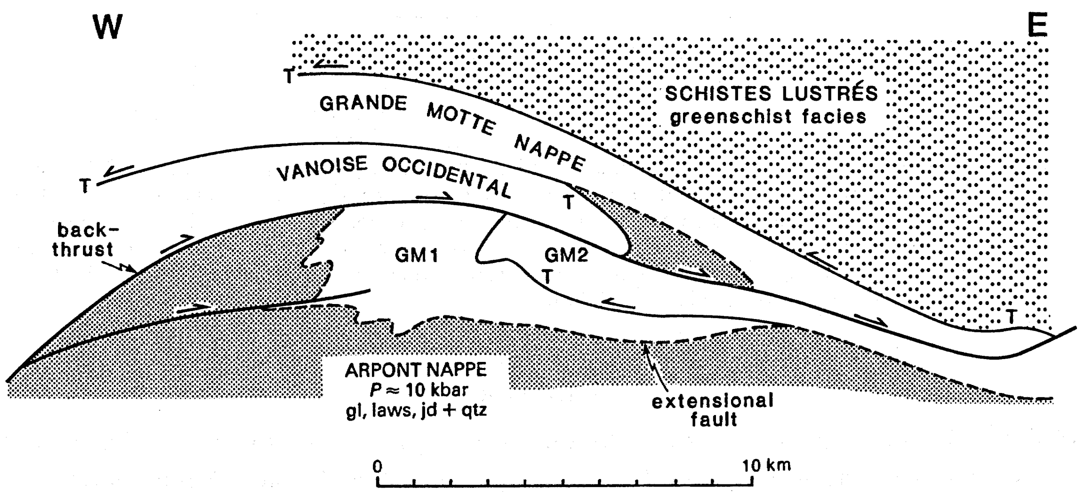

FIGURE 9. Schematic section across the southern Vanoise massif, internal French Alps. Greenschist-facies metasediments (Schistes Lustrés) lack high-pressure minerals in this area, and lie 1–3 km above the Arpont basement nappe, which carries blueschist-facies assemblages including jadeite+quartz. The early tectonic contacts (T) are thrusts in that they duplicate Mesozoic stratigraphy in the Vanoise Occidental and Grande Motte nappes, but they may have been reactivated as normal faults, cutting out parts of the metamorphic zonation. The extensional fault shown is identifiable as such because it cuts out part of the stratigraphy. GM1 and GM2 are thrust slices in the Grande Motte nappe. After Platt & Lister (1985*a*, *b*) and unpublished data.

normal faults on stratigraphic grounds, as they may cut or reactivate earlier compressional structures. They can most readily be identified by their effects on metamorphic zonation. Examples of such faults from the Alps, the Betic Cordillera, and the Franciscan Complex are illustrated in figure 9–11.

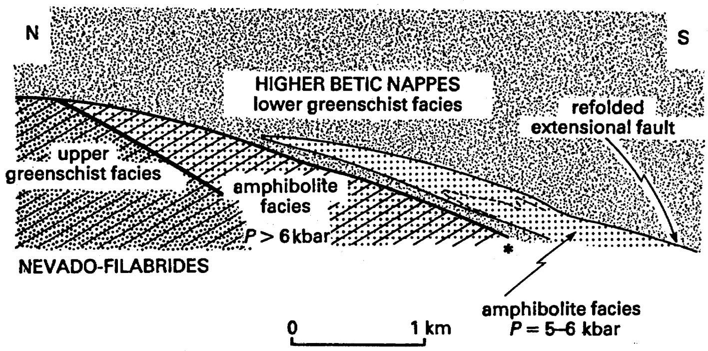

FIGURE 10. Schematic section across the Sierra Alhamilla, Betic Cordillera, SE Spain. The Higher Betic Nappes, belonging to a low to moderate P: T-ratio facies series ($P_{max} \approx 6$ kbar), lie on amphibolite-facies rocks of the Nevado–Filabride Complex, which elsewhere carries high-pressure assemblages. The contact (starred) is probably extensional. The labelled extensional fault separates lower greenschist–facies from amphibolite–facies rocks within the higher Betic Nappes, and has been refolded where it was entrained in the basal shear zone. After Platt *et al.* (1983) and Platt & Behrmann (1986).

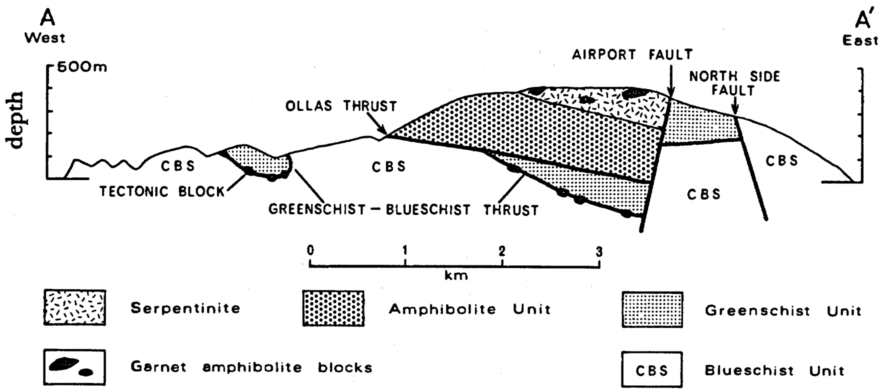

FIGURE 11. Schematic section across Santa Catalina Island, in the Franciscan Complex, southern California. The Blueschist, Greenschist, and Amphibolite Units were metamorphosed at 8–12 kbar and about 300, 450 and 600 °C respectively in an inverted metamorphic aureole beneath hot peridotite in the hangingwall of the subduction zone (Platt 1975, 1976). The present tectonic contacts are post-metamorphic and formed during uplift. They have cut out much of the original metamorphic sequence, and depending on the initial attitude of the isograds, may well be low-angle normal faults.

P–T HISTORIES

If high-P–low-T metamorphic rocks do approach the surface by the mechanism described above, distinctive patterns of P and T in time and space should result. A full analysis of this question is beyond the scope of this paper, but some qualitative statements are possible.

1. If uplift occurs concurrently with subduction and underplating, overall thermal gradients will remain low, and the temperature at a given material point will remain constant, or may even drop (Rubie 1984). Analyses in terms of a post-tectonic return to a equilibrium thermal gradient (England 1978; Oxburgh & England 1980; England & Thompson 1984; Thompson

& England 1984) are not directly applicable to this situation. P–T–time trajectories during uplift are likely to be very steep and may have positive slopes (figure 12).

2. If underplating and uplift are concurrent, the peak metamorphic pressure may remain constant or even decrease passing downwards through the high-pressure terrain, and the age of the high-pressure peak will decrease downwards (Platt 1975). The peak pressure and its age will also decrease along the erosion surface towards the toe of the wedge (figure 6).

3. Extension in the upper part of the orogenic wedge will produce apparently oversteepened pressure gradients, or jumps in metamorphic pressure with lower-grade rocks above higher-grade ones.

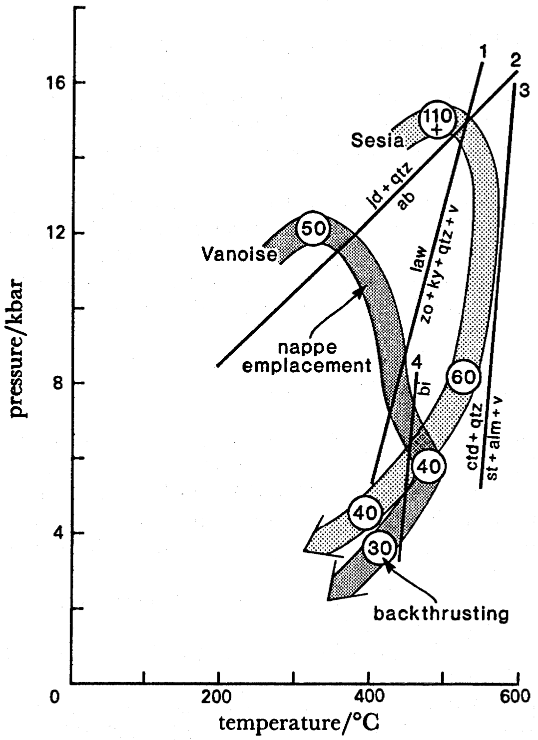

Figure 12. P–T paths of Alpine high-pressure rocks. Curve for the eclogitic mica schist unit of the Sesia Zone after Rubie (1984), for the Arpont schist of the Vanoise massif after Platt & Lister (1985*a*). Circled numbers indicate approximate ages in million years b.p.: those for the Vanoise are speculative. Abbreviations: ab, albite; alm, almandine; bi, biotite; chd, chloritoid; jd, jadeite; ky, kyanite; law, lawsonite; qtz, quartz; st, staurolite; v, vapour. Mineral stability curves: 1, Newton & Kennedy (1963); 2, Newton & Smith (1967); 3, Rao & Johannes (1979); 4, stilpnomelane + phengite→biotite + chlorite + quartz + H_2O from Nitsch, quoted in Winkler (1974, p. 202).

4. Because this process involves post-metamorphic lateral translations of several tens of kilometres, rocks from different thermal regimes may be juxtaposed across the low-angle extensional faults. This seems to be the case in the Eastern Alps (Hawkesworth *et al.* 1975) and in the Betic Cordillera (Torres-Roldán 1979), where rocks metamorphosed in low to moderate P: T-ratio environments directly overlie high P: T-ratio terrains.

Conclusions

The operation of buoyant forces, diapirism, or hydrodynamically driven return flow may explain the return to crustal depths of some high-P–low-T metamorphic rocks that have been subducted into the mantle, particularly those that have been exhumed from exceptional depths, and those that occur as exotic blocks in mélanges. These mechanisms do not, however, explain the emplacement at high tectonic levels of coherent regional high P: T-ratio terrains. The mechanical consequences of underplating beneath an accretionary wedge with negligable yield strength provides an explanation for the uplift of these terrains, and successfully predicts their tectonic position and history. The hypothesis also predicts that many currently active accretionary prisms and collisional orogens should be undergoing extension in the upper rear of the wedge – a prediction that can be tested, and that is supported by recent studies.

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Discussion

D. C. Rubie (*Department of Geology, The University, Manchester*).

1. Dr Platt's model relies on material within the orogenic wedge having a low yield strength. Although field and microstructural data from the Sesia Zone indicate a low yield strength *during* high-pressure metamorphism (see, for example, Rubie 1983), it is apparent that the eclogite–facies mineral assemblages only survived uplift back to the surface in the absence of penetrative deformation. Where the rocks were penetratively deformed under low-pressure conditions during uplift, a greenschist overprint is well developed. Thus, a large part of the Sesia Zone, now outcropping over an area of several hundred square kilometres, evidently returned to the surface as a near-rigid body suggesting that this unit had a high yield strength during uplift. Could the author comment on these observations?

2. The possible role of erosion in returning high-pressure rocks to the surface should not be underestimated. Only moderate erosion rates (averaging *ca.* 0.6 mm a^{-1}) would have been required for the entire uplift of the Sesia Zone and, as pointed out by Dr Platt, *rapid* uplift was unnecessary provided subduction and underplating operated to prevent significant temperature increases. The paucity of pre-Oligocene sediments, which would indicate significant erosion, is a problem, but is there not the possibility that large volumes of such sediment were subducted?

3. Whereas evidence from the Eastern Alps indicates an inadequate thickness of overburden to explain the high-pressure metamorphism of the Pennine Zone, in the Sesia Zone there is no trace of any overburden. The uppermost unit of the Sesia Zone, the Seconda Zone diorito-kinzigitica, developed blueschist to eclogite facies assemblages in shear zones and tectonic contacts and must, therefore, have been buried to a depth of 40–50 km with the rest of the Sesia Zone (Lardeaux *et al.* 1982). Could Dr Platt speculate on the fate of the 40–50 km of overburden that must have once overlain this unit, and of which there is now no trace?

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PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

J. P. Platt.

1. I assume that the wedge *as a whole* lacks a significant yield strength. This is justified by the fact that most orogenic wedges contain large volumes of water-rich sedimentary rocks undergoing prograde metamorphism. The influence on the bulk rheology of bodies of strong crystalline rock, such as the Sesia Zone, depends on their arrangement within the wedge. If there are enough of them, and they are in physical contact so that they can create a load-supporting element within the wedge, they could greatly influence its bulk rheology. If, on the other hand, they are physically isolated in a matrix of weaker material, they will have relatively little effect. The Sesia Zone is underlain by a group of metasedimentary rocks known collectively as Schistes Lustrés, which are among the least competent materials in the Alps, but it is less clear what lay above it. As Dr Rubie points out, parts of the Sesia zone itself were deformed and altered to assemblages rich in chlorite and mica during uplift, and were probably considerably weakened during this process. I think it is unlikely that the rigid part of the Sesia Zone would have been able to support a significant part of the total load within the Alpine wedge. It was probably carried passively into its present position as a result of the flow in the surrounding material.

2. If erosion had been the principal mechanism for unroofing the Sesia Zone, more than 50 km of rock would have had to have been eroded during its 100 Ma uplift history. This would presumably have included the entire continental crust of the leading edge of the Adriatic plate, plus 20 km or more of mantle periodotite. We should expect to see a very distinctive suite of high-pressure mafic and ultramafic minerals in the resulting sediments. Alpine flysch sequences provide evidence for the erosion of upper-crustal rock-types (Mesozoic sediments and Hercynian mica–schist, amphibolite, gneiss and granite), and ophiolitic detritus appears locally (Homewood & Caron 1983), but I am not aware of any evidence for the erosion of deep crust and mantle rocks. It would be remarkable if all sediments derived from this terrain had been preferentially subducted.

3. The tectonic position of the Sesia Zone is obscured by the late rotation of the internal margin of the Alps into a near-vertical orientation, and the associated large vertical and horizontal displacements on the Tonale–Insubric fault system. The most probable candidate for the original overburden is Southern Alpine continental crust such as that seen in the Ivrea Zone, together with a significant amount of underlying mantle. I suggest that three separate processes may have cooperated to elevate the Sesia Zone to its present position. Firstly, it may have risen through the mantle as a result of buoyancy forces. This could have brought it from its maximum depth of burial of 50 km or more to a depth of 30–40 km at the base of the Adriatic continental crust, in the rear of the Alpine orogenic wedge. Secondly, extension of the wedge by the mechanism described in my paper, in response to continued underplating of Penninic units, could have allowed the Sesia Zone to rise to within 15–20 km of the surface. A final stage of isostatic uplift and erosion in Oligocene–Recent time resulted from underthrusting of the European margin.

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