Dynamics of orogenic wedges at Pachmarhi

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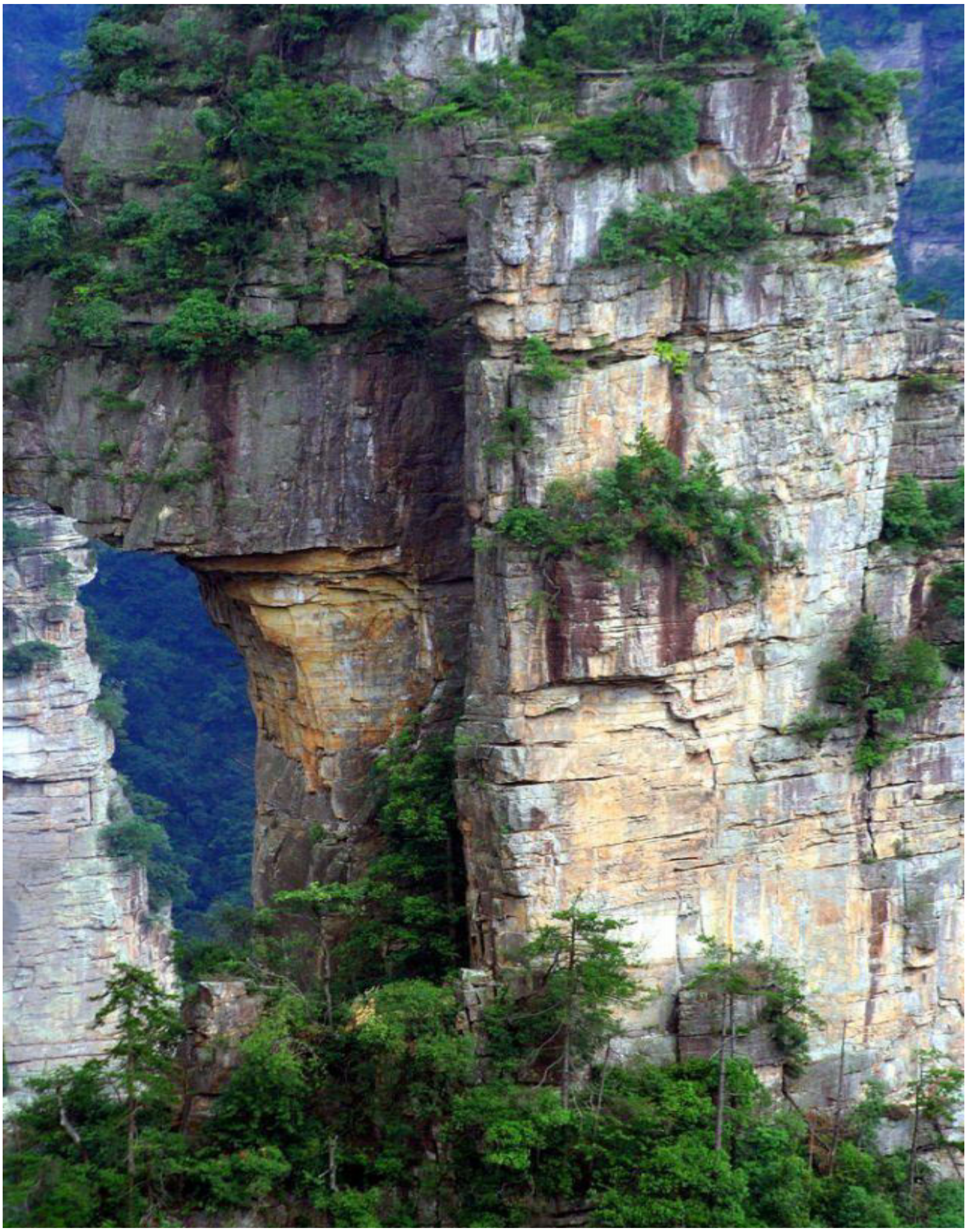

ABSTRACT - Subduction-accretion complexes can be approximated as wedge-shaped continua with a rigid buttress behind and a subducting litho- spheric slab beneath. Thick wedges undergoing prograde metamorphism have negligible long-term yield strength and are likely to exhibit a complex nonlinear viscous rheology. Such a wedge will tend to deform internally until it reaches a stable configuration, in which the gravitational forces generated by the wedge geometry balance the

traction exerted on its underside by the subducting slab. Accretion of material at the wedge front will lengthen the wedge and cause it to shorten internally to regain the stable geometry. This shortening will be expressed as late (out-of-sequence) thrusting, back-thrusting, and folding. Conversely, underplating of sediment or crustal slices will thicken the wedge, which may need to extend internally to regain stability. Extension will cause listric normal faults that may merge downward into zones of ductile extension. Continued under-plating at depth and compensating extension above provides a mechanism for bringing high-P/low-T metamorphic rocks to upper levels in the rear of the wedge, where they are commonly observed. Many major tectonic boundaries in convergent orogens, such as the Dhupgarh, Mahadeva, Chauragarh Complex, major nappe contacts in the Patalkot, and the contact between the Tamia.Kalapathar and Chandimai Higher complexes in the Pachmarhi show abrupt increases in metamorphic grade downward across them. This is consistent with their origin or reactivation as uplift-related, extensional structures.

Introduction

Uplift and exposure at the Earth's surface of high-pressure low- temperature metamorphic rocks in Phanerozoic orogenic belts has been a long-standing geological problem. Difficulty in reconciling the tectonic setting of large ($10-1{,}000\ km^2$) $high-P/low-T$ terranes with the experimental evidence that they have been buried to depths of $30-50\ km$ led to the idea of tectonic overpressure (Blake and others1967; Brothers, 1970; de Roever, 1972). Tectonic overpressure is limited by the mechanical strength of rocks to $\sim 1\ kb$ or less (Brace and others, 1970; Ernst, 1971 a) and is not now regarded as a significant factor in contributing to metamorphic pressures. The problem still remains, however; we must explain how the Pachmarhi terranes that have been buried to depths equal to or greater than the thickness of normal continental crust are brought back to the surface, and how their $high-P/low-T$ assemblages escape overprinting by thermal re-equilibration during uplift (England, 1978; Draper and Bone, 1980).

The initial process of deep burial can be explained in terms of plate convergence: material on the subducting plate will be subjected to progressively greater load pressures as it is carried beneath the upper plate. If the material is then detached at depth, it will be incorporated into a growing accretionary wedge and will be potentially available for subsequent exhumation. Compressional deformation alone, however, cannot bring rocks closer to the topographic surface. An additional process is needed, either to bring the high-P rocks upward relative to their surroundings or to remove material from above them. Possibilities include the buoyant rise of subducted slabs of crustal material relative to higher density mantle rocks (Ernst, 1971 b, 1975, 1984; Piatt, 1986); entrainment of blocks of $high-P$ rocks in relatively low-density diapirs (England and Holland, 1979; Carlson, 1981; Moore, 1984) or in flowing mud-matrix melanges (Cloos, 1982); or regional uplift driven by under-plating of material at the base of the accretionary wedge, coupled with erosion at the surface (Piatt, 1975; Rubie, 1984). None of these processes appears to adequately explain the tectonic setting and timing of uplift of regionally coherent metamorphic terranes that are many tens of square kilometers in extent. The difficulties can be outlined briefly as follows :

1. Many $high-P/low-T$ terranes, which have specific gravities in the range of $2.8-3.0$, form nappes or thrust sheets near the structural top of an orogenic complex consisting largely of crustal rocks with similar or lower specific gravity (Figure 1 through 3). They are unlikely to have been emplaced there by buoyancy forces.
2. Stratigraphic and geochronological evidence suggests that the major part of the uplift of these terranes occurred while plate convergence was continuing. This is also supported by

the lack of thermal overprinting, which indicates that thermal gradients were kept low by subduction during the uplift (Rubie, 1984). Uplift does not appear to have been primarily a result of continental collision or post-convergence wrench-faulting.

3. The same evidence indicates that uplift largely predated extensive emergence and erosional unroofing. Erosion, therefore, was not the principal mechanism by which the overburden was removed.
4. The $high-P/low-T$ terranes shown in Figures 1 through 3 are overlain by allocnthons that lack $high-P$ mineral assemblages and are insufficiently thick to explain the metamorphic pressures in the rocks beneath them. Material has therefore been removed from *within* the structural pile, rather than from off the top. This suggests a tectonic mechanism for unroofing, rather than erosion.

The Dhupgarh, Mahadeva, Chauragarh Complex, major nappe contacts in the Patalkot, and the contact between the Tamia, Kalapathar and Chandimai Higher complexes in the Pachmarhiare discussed in detail in later sections, and the statements above are documented. The similarities in the tectonic setting arid uplift history of the $high-P/low-T$ terranes in these otherwise dissimilar convergent orogens suggest that a common process has operated. The purpose of this paper is to examine the dynamics of convergent orogenic wedges at Pachmarhi undergoing prograde metamorphism, and to show that a kinematic pattern can be predicted that satisfactorily accounts for the observed relationships.

Dynamics of orogenic wedges review of current concepts

Convergent orogens, including both accretionary complexes forming above subduction zones and foreland thrust belts, can be approximated in terms of a wedge-shaped prism resting on a rigid slab that is sliding beneath them, with a buttress behind the prism (Figure 5). The buttress may not be absolutely rigid, but it deforms much more slowly than the material in the wedge. The wedge itself is composed largely of material scraped off the undersliding slab (Figure 6).

A fundamental concept in the analysis of such a wedge is that it behaves as a single, mechanically continuous, dynamic unit. This was argued explicitly by Price (1973), who suggested that the wedge as a whole is macroscopically ductile, in that it does not lose its over-all mechanical integrity during deformation. Price's discussion was in the context of a particular model, an analogy between the "flow" of an orogenic wedge and that of a glacier or ice-sheet, driven by the gravitational potential associated with a surface slope. Elliott (1976) elaborated this concept and showed that if the material in the wedge is relatively weak, gravity produces a horizontal shear stress r given by

$$\tau = \rho\, gha, \tag{1}$$

where$p = density$,$g = \ gravitational\ acceleration$, $h = \ depth\ below\ surface$and $a = \ surface\ slope\ of\ wedge$, using a small-angle approximation.

This is the stress that drives both the internal flow and basal sliding of a glacier (Paterson, 1981), and it cannot produce a net longitudinal shortening in the system.

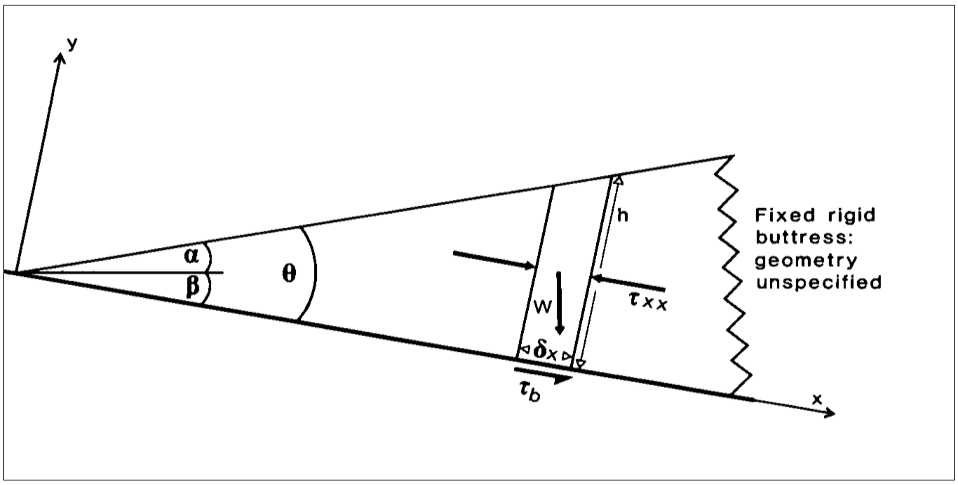

Figure5. Simplified model of an accretionary wedge used for the analysis, α is the surface slope, β the basal slope, relative to the horizontal. A segment of length δx and height h is subject to a body force$W$, to a basal traction$\tau_b \delta x$, and to "push" forces produced by the longitudinal deviatoric normal stress τ_{xx}.

This mechanism cannot, therefore, explain the evidence for large horizontal shortening in convergent orogenic belts. The next step was the landmark paper by Chappie (1978), who treated an orogenic wedge as a continuum with perfectly plastic properties, and who showed that the wedge-shaped geometry is a consequence of its internal dynamics. If a force applied at the rear of the wedge is progressively counteracted along its length by the resistance to sliding on its base, a greater cross-sectional area is needed toward the rear to prevent the longitudinal stress from exceeding the strength of the material. If the wedge is insufficiently tapered, it shortens internally and thickens at the rear until it is able to transmit the push-force forward. A side effect of the resulting taper is the generation of a surface slope in the direction of motion; this is therefore a consequence rather than a cause of the deformation. Chappie derived an expression for the dynamic balance of a yielding plastic wedge, based on the assumption that the wedge could slide on a weak basal layer with plastic yield strength not more than one-third of the strength of the wedge itself. This expression, which I have slightly modified here, gives the basal shear stress

$$\tau_b = p\ gha + 2K, \qquad (2)$$

The first term on the right side of the equation is the gravitational stress associated with the surface slope, as discussed above. The second term is caused by the longitudinal compressive stress, or "push"; it depends on the yield strength of the wedge (K) and the angle of taper(θ). Chappie criticized Elliott's (1976) analysis fairly sharply, but much of the difference arises from different assumptions about the material properties of the wedge. Elliott assumed that the wedge would be unable to sustain deviatoric stresses of more than 200 bars ($20\ MPa$) on a geological time scale. Significant longitudinal stress would not be transmitted through such a system, and Elliott therefore neglected it. Chappie, conversely, assumed yield stresses of $1-2\ kb$ but relatively low shear stresses at the base, a system that favors transmission of longitudinal stress. The two analyses are, in fact, simply special cases of a more general dynamic law.

Davis and others (1983) came to similar conclusions as those of Chappie, but based their findings on a different Theological assumption: that of a Coulomb fracture criterion modified for the effects of pore-fluid pressure. They introduced the term "critical taper" to describe the angle of taper for a wedge that is in a state of yield throughout, such that the

Figure1. Highest Point of Dhupgarh, consisting Basaltic Rock. Note that $high - P$ rock accured near the top of the Dhupgarh.

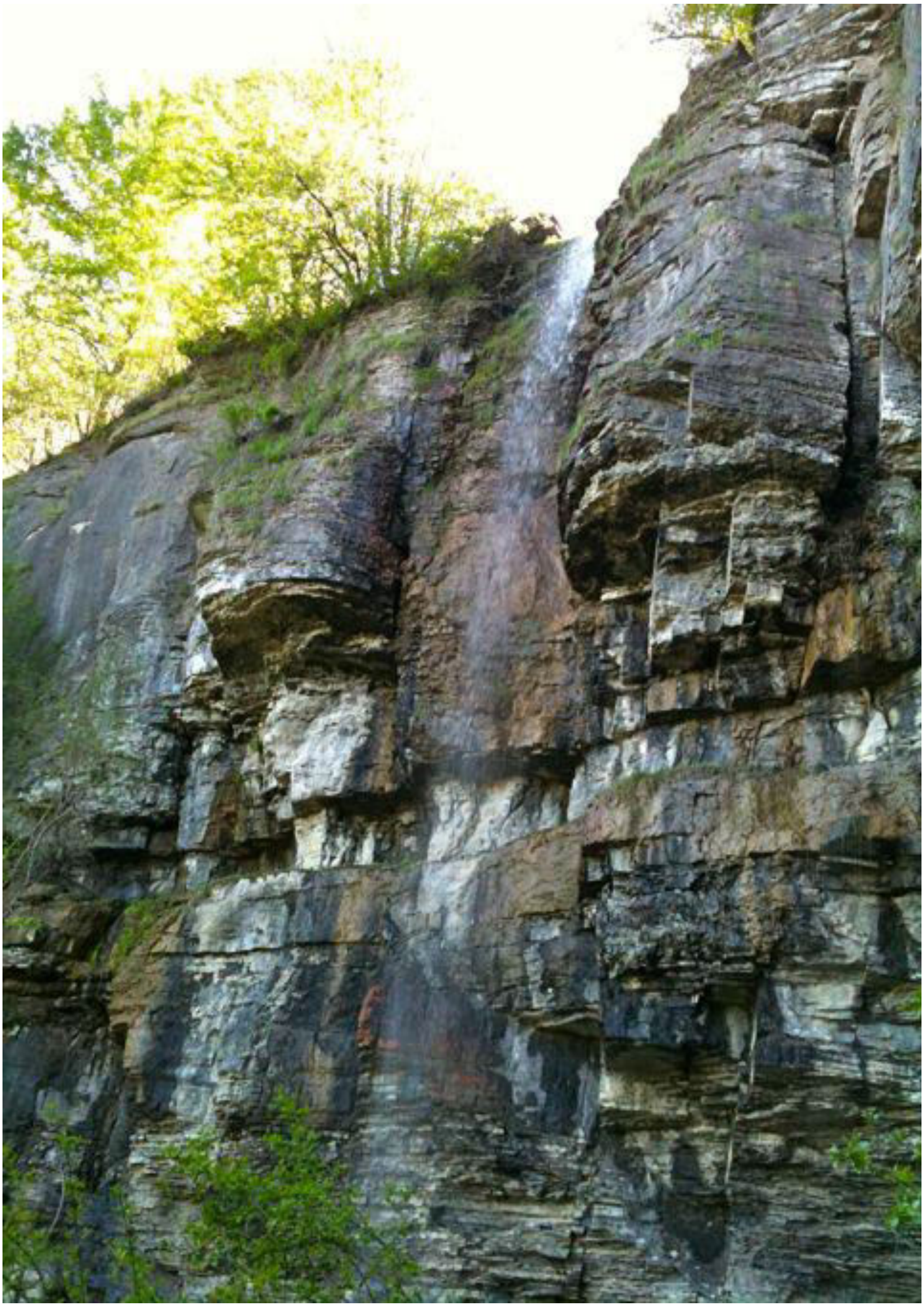

Figure 2 . The Mahadeva ranges , younger, $lower - P$ rocks occur at depth. Present tectonic contacs postdate the $high - P$ peaks.

Figure 3. High pressure assemblages in the Chauragarh have been strongly overprinted by the basaltic rocks. Chauragarh are poorly constrained. Note that $high - P$ rocks occurred at high structural levels bounded by contact that postdate the $high - P$ Peak

Figure 4. Patalkot -Tamia Scarpment, through P rock occured at the top tectonic contacts postdate the $high - P$ metamorphism

Compressive and gravitational forces are just adequate to overcome the resistance to sliding at the base their expression for the basal shear stress, slightly simplified, is

$$\tau_b = p\ gha + (\ 1 - \lambda\ )K_p\ gh\theta, \tag{3}$$

The two terms on the right side of the equation have the same significance as those in Chappie's criterion. The second term on the right side is related to material properties of the wedge: λ is the ratio of pore-fluid to lithostatic pressure, and K is a function of the internal strength of the wedge, and an inverse function of the coefficient of friction at the base. Davis and others (1983) point out that for submarine wedges, the term ρ must be replaced by $\rho' = \rho - \rho_w$ to compensate for the effect of the water column with density ρ_w. they have since modified their analysis to allow for the effect of cohesion (Dahlen and others, 1984).

Thick orogenic wedges: rheology

Orogenic wedges that are thick enough to induce metamorphism are unlikely to exhibit either plastic or Coulomb-type rheologies, so that the analyses of Chappie (1978) and Davis and others (1983) are not directly applicable. Near-surface and frontal parts of the wedge may behave in a brittle fashion, but ductile deformation will be of increasing importance at depth.

Sediments undergoing low-grade metamorphism (at temperatures between 150 and500°C) commonly deform by a process loosely known as pressure-solution, involving the directed diffusional transfer of material in such a way as to allow the rock to change shape (Rutter, 1983). This process probably obeys an approximately linear-viscous constitutive flow law (Rutter, 1976, 1983), at least at low stresses. More highly deformed rocks, particularly dry crystalline basement materials, show evidence of in tracrystalline deformation by dislocation glide or creep (White, 1976). This process probably leads to a power-law relationship $\epsilon^{\cdot} = f(\sigma^n)$ between the strain-rate,$\epsilon^{\cdot}$and the stress intensity, σ , but the situation is complicated by the fact that continuing crystal-plastic deformation leads to profound microstructural changes that affect the rock rheology. In particular, grain- size reduction caused by dynamic recrystallization can lead to a switch in the dominant deformational mechanism to a grain-flow process in which unrestricted grain-boundary sliding is accommodated by some form of diffusional or dislocation creep (Bouiller and Gueguen, 1975; White and others, 1980)This is probably accompanied by substantial softening, so that deformation is progressively concentrated into weak zones of ultra- mylonite with high rates of deformation. The bulk-rock rheology will therefore depart from that derived for crystal-plastic deformation; it will change with time and may involve a stress-drop (ductile yield).

1969); it may deflect propagating fractures (Donath, 1961); and if a plane of anisotropy is rotated with respect to the stress field by folding or imbrication, the effective rheology changes (Cobbold, 1976). Mechanical in homogeneities (such as a thrust sheet of strong crystalline rock) can cause local amplifications in stress and reorientations of stress trajectories.

The complexities discussed above effectively frustrate the complete description of a thick orogenic wedge in terms of a bulk rheology, and we are not yet in a position to set up useful quantitative models. The following basic rheological statements can be justified, however, and provide the basis for the simple qualitative analysis in the next section.

1. Below $\sim 10 - 15\ km$, the material of the wedge is probably capable of deforming at very low stress intensities, given sufficient time (Pavlis and Bruhn, 1983). The reasons for believing this are (a) the importance of pressure-solution processes in deforming metasedimentary rocks, and (b) the fact that several powerful softening mechanisms are

likely to act cooperatively at depth in an orogenic wedge. Increasing temperature with depth reduces effective viscosities for all ductile deformation mechanisms. Metamorphic dehydration reactions in clay-rich sediments yield abundant water. A small amount of water reduces plastic yield strength in quartz by an order of magnitude (Blacic, 1975), facilitates diffusional creep (Rutter, 1983), favors subcritical corrosion cracking (Atkinson, 1980), and lowers the stress needed for brittle fracture (Etheridge, 1983). The evidence suggests that rocks undergoing prograde metamorphism at temperatures in the range of $300 - 500$℃ can deform at significant rates (that is, faster than 10^{-14}/sec) under differential stresses of ~ 100 bar (Rutter,1976).

2. If the wedge is subjected to higher stresses, deformation will not become unconstrained (as in a perfect plastic or Coulomb material) but will simply occur at a higher rate. This can be described as general (nonlinear) viscoelastic behavior, for which the concept of a discrete, long-term yield-strength is meaningless.The material will deform, although slowly, under any deviatoric stress, however low. On the other hand, the material can "support" high stresses because of its viscous resistance to a high strain-rate.

Variables that must be taken into account in assessing rock rheology include mineral assemblage, grain size, temperature, and the availability and activity of water. Anisotropy of material properties, caused by bedding or deformational fabrics, has particularly disconcerting effects: it causes a divergence between the axes of stress and instantaneous strain (Malvern,

Thick orogenic wedges: a stability criterion

Conditions for the mechanical equilibrium of a thick orogenic wedge can be described using the simplified geometry for a two−dimensional wedge shown in Figure 5. The forces acting in the x −direction on a segment of length δx are as follows. (1) Longitudinal normal forces act in opposite directions on the rear and front faces of the segment, arising from the deviatoric normal stresses τ_{xx}within the wedge. (2) The resistance to sliding or shear flow (τ_B) at the base of the wedge leads to a net resistive force in the x −direction. (3) A component of the weight (W) of the segment acts down the dip of the basal surface. We can relate these terms to one another using the stress equilibrium equations for creeping flow, appropriate to a material with a bulk viscous rheology (Malvern, 1969, p. 215). In the x −direction (using the engineering sign convention),

$$\frac{\partial \sigma_{\mathrm{xx}}}{\partial \mathrm{x}} + \frac{\partial \tau_{yx}}{\partial y} + \rho\, g \sin\beta = 0; \tag{3}$$

And in the y −direction,

$$\frac{\partial \sigma_{\mathrm{yy}}}{\partial \mathrm{y}} + \frac{\partial \tau_{xy}}{\partial x} + p\, g \cos\beta = 0, \tag{4}$$

In a basin wedge, the densityρ should be corrected for the density of the surrounding water, as noted above. The longitudinal gradient in horizontal shear stress,$\partial \tau_{xy}/\partial x$ is likely to be negligible compared with the vertical gradient in normal stress ,$\partial \sigma_{yy}/\partial y,$ so that equation 4 reduces to

$$\frac{\partial \sigma_{\mathrm{yy}}}{\partial \mathrm{y}} = \rho\, g \cos\beta, \tag{5}$$

Integrating equation 5 with respect to y,

$$\sigma_{yy} = pg(y - h)\cos\beta, \tag{6}$$

Figure 6. Patalkot-Tamia, fixed rigid Buttress.

Stresses can be treated as the sum of a hydrostatic component, p. (the mean stress) and a deviatoric component; that is

$$\sigma_{jj} = \mathrm{p} + \tau_{\mathrm{ij}}, \tag{7}$$

Hence,

$$\sigma_{\mathrm{xx}} = p + \tau_{\mathrm{xx}}, \tag{8}$$

and from equation 6,

$$\mathrm{p} = pg(y - h)\cos\beta - \tau_{yy}, \tag{9}$$

Because for plane stress,

$$\tau_{yy} \;=\; -\tau_{xx},$$

$$\sigma_{xx} = \rho g(y - h)\cos\beta + 2\tau_{xx},, \tag{10}$$

σ_{xx}can therefore be treated as the sum of the local lithostatic load and an additional longitudinal normal stress component, $2\tau_{xx}$ which may be either positive or negative. The significance of$\tau_{xx}$is that it alone determines whether the wedge segment will shorten or extend in the x-direction. Substituting equation 10 in equation 3,

$$\frac{\partial}{\partial x}[p\ g\ (y - h)\cos\beta + 2\tau_{xx.}] + \frac{\partial\tau_{yx}}{\partial y} + \rho\ g\sin\beta = 0, \tag{11}$$

Integrating with respect to y,

$$-\frac{\partial}{\partial x}(1/2\,\rho g h^2\cos\beta) + 2\frac{\partial}{\partial \mathrm{x}}\int_0^h \tau_{xx}dy + \tau_b + \rho g h\sin\beta = 0, \tag{12}$$

For small angles$\frac{dh}{dx} = \tan\theta$, and taking a depth-averaged value for τ_{xx},

$$-{}^{1}\!/_{2}\,\rho\ gh\tan\theta\cos\beta + 2\frac{\partial}{\partial \mathrm{x}}(\tau_{xx}h) + \tau_b + \rho\ g\mathrm{h}\sin\upbeta = 0, \tag{13}$$

Using small-angle approximations and noting that$\theta = a + \beta,$

$$\tau_b = \rho\ gh\ \alpha - 2\tau_{xx}\theta - 2\frac{d\tau_{xx}}{\partial x}h, \tag{14}$$

This equation states that the traction exerted on the wedge by the slab sliding down beneath it is balanced by three terms. The first is the familiar gravitational term produced by the surface slope. The second is the result of the longitudinal stress,τ_{xx} acting in a tapered wedge; because of the progressive change in cross-sectional area of the wedge, this produces a net longitudinal force. The third results from the gradient $\partial\tau_{xy}/\partial x$in the longitudinal stress, which also produces a net force in the x −direction.

Equation 14 cannot be solved without further information on the rheology, which would allow limits to be placed on$\tau_{xx.}$, and $\partial\tau_{xy}/\partial x$, but a criterion for stability can be derived from it without any additional assumptions. I use "stability" to mean a condition in which the wedge, or a finite section of the wedge, is not shortening or extending internally, so that $a,\ \beta,$ andθ are constant. In this condition, deformation will be limited to sliding or distributed shear strain parallel to the base of the wedge (the effect of accretion is discussed below).

The condition for stability is that the longitudinal stretching rate$\dot{\epsilon_{xx}} = 0$.For a viscous material $\dot{\epsilon_{xx}} = \mathrm{f}(\tau_{xx})$, and so for stability $\tau_{xx} = 0$ and $\partial\tau_{xy}/\partial x = 0$. Equation 14 then reduces to a stability criterion:

$$a = \frac{\tau_b}{\rho \mathrm{gh}}, \tag{15}$$

This is identical to the glacier sliding term already discussed. It is not, however, a general expression for the dynamics of the wedge; it simply states that the special case of a wedge that is neither shortening nor extending has a surface slope α such that the gravity sliding stress just balances the shearing stress at the base. In this condition, the only stresses on the rear buttress are lithostatic. Equation 15 does not, therefore, explain or apply to the deformation of material in the wedge. As discussed below, deformation is a consequence of processes such as accretion, which change the shape of the wedge and throw it out of stability. The equilibrium condition 14 then applies, and the resulting deformation is caused by the longitudinal stress terms.

The significance of the stability criterion can be examined by considering the effect of changing the surface slope α so that the wedge is unstable. If α is too small, the gravity term in equation 14 will be too low, so that either or both terms in τ_{xx} must be negative. Bearing in mind that τ_{xx} is likely to be zero or negative (compressional) at the front of the wedge (depending on the strength of the incoming material) a negative $\partial\tau_{xy}/\partial x$ is likely to lead to a negative τ_{xx} in the wedge. A value of that is too low will hence lead to horizontal compression and shortening in the wedge. This will thicken the wedge, leading to an increase in $\alpha$and a return to stability. Conversely, if α is too large, the equilibrium equation 14 requires that one or both terms in τ_{xx} be positive. τ_{xx} at the wedge front is limited by the strength of the incoming material, and so positive values of $\partial\tau_{xy}/\partial x$will tend to lead to positive values of τ_{xx} in the wedge. This will result in longitudinal extension, thereby reducing a. It is this latter conclusion that is of particular significance to the problem of the uplift history of $high - P/low - T$ metamorphic rocks. Note that if the wedge approaches isostatic compensation, $\beta \longrightarrow \alpha\rho_c/(\rho_m - \rho_c)$ and $\theta \longrightarrow \alpha\rho_m/(\rho_m - \rho_c)$where ρ_m is the mantle density at the level of compensation, and ρ_c is the density of the wedge. For a submerged wedge, these should be $\beta \longrightarrow \alpha(\rho_c - \rho_w)/(\rho_m - \rho_c)$and $\theta \longrightarrow \alpha(\rho_m - \rho_w)/(\rho_m - \rho_c)$ This means that to produce a given change in α, the change in θ will have to be 3 to 6 times greater, and the required change in thickness of the wedge will be correspondingly much larger.

The stability criterion allows us to make a generalized prediction about the surface profile of an accretionary wedge. If, for example,τ_B is constant, in a stable (nondeforming) wedge, α will increase toward the wedge front as h decreases, giving a convex-upward profile. This is comparable to the profiles predicted by Emerman and Turcotte (1983), who used a linear viscous model and lubrication theory, and by Stockmal (1983), who used a plastic model. Many accretionary prisms and orogenic wedges show this sort of profile (Emerman and Turcotte, 1983), although it cannot be assumed that they have all reached a stable configuration.

Effect of accretion on stability

The significance of the stability criterion derived above is that it allows us to predict patterns of deformation in the accretionary wedge resulting from externally imposed changes in its geometry. The most important cause of change is accretion, and it is useful to distinguish two principal types.

1. *Frontal accretion* refers to the accumulation of material at the tip of the wedge; it is detached, shortened, and incorporated into the wedge at a deformation front propagating through the incoming material. Because the accreted material tends to lengthen the wedge, α will be low in the frontal region, which will therefore be in compression (Figure 7A). In this area, the wedge may exhibit Coulomb behavior (Davis and others, 1983; Dahlen and others, 1984) with finite yield strength. If the longitudinal stresses are large enough, internal shortening in the wedge is likely. This can lead to renewed (out of sequence) thrusting, which may create thrusts and backthrusts cutting across the earlier accretionary structures (Figure 7A). The importance of backthrusting as a mechanism for internal thickening of the wedge is beautifully illustrated by the sandbox experiments of Davis and others (1983). Movement on each successive "thrust" is almost immediately followed by the formation of a backthrust, and movement on the backthrusts in particular continues as the wedge grows forward, allowing it to thicken and maintain its surface slope. These effects would be greatly increased in an isostatically compensated wedge. Large-scale backthrusting accompanying and following forward thrusting has been documented in the Pachmarhi (Figure 8, Figure 9, Figure 10, Figure 11,) Late-stage (out-of-sequence) forward thrusting occurs for the same reason.

2. *Underplating* is the process by which material is accreted to the underside of the wedge, after it has traveled down with the underthrust slab a certain distance (Figure 7A). Underthrusting is probably the only way in which sedimentary materials can reach the depths required for blueschist-facies metamorphism (Ernst, 1971b; Piatt, 1975). Underplating has been demonstrated experimentally by Cowan and Silling (1978), and seismic evidence makes it clear that sediment is being thrust for tens of kilometres beneath several active subduction complexes (Westbrook and others, 1982; Aoki and others, 1982; White and Louden, 1982; Leggett and others, 1985). Mass-balance and structural arguments have also been invoked in favor of large-scale underthrusting and underplating of sediment inboard of the Lower Mahadeva of Satpura, The process by which underthrust material is eventually accreted is poorly understood, but it may involve detachment and shortening either by folding, or by imbrication to form duplexes (Silver and others, 1985; von Huene, 1984; Piatt and others, 1985). It may be triggered by physical changes caused by increasing pressure and temperature, and buoyancy forces may inhibit subduction of sediment or continental basement into the mantle (Molnar and Gray, 1979). Underplating beneath the rear of the orogenic wedge will thicken it and cause α and θ to increase. The longitudinal stress τ_{xx}, may therefore become tensional and cause the wedge to extend (Figure 7B). The prediction, therefore, is that underplating at the base of the wedge will be accompanied by extension higher up, which provides a possible solution to the problem of the uplift of $high - P/low - T$ metamorphic rocks.

Other controls on stability

The other major factors that are likely to influence the shape of orogenic wedges are erosion, sedimentation, and changes in the basal shear stress, τ_b. Erosion at the rear of the wedge decreases α and θ, and may lead to eventual shortening in the wedge (Chappie, 1978). Conversely, sedimentation toward the rear of submarine wedges increases α and θ overall and will favor extension and further sedimentation. This may explain the formation of constructional fore-arc basins at the rear of some accretionary complexes. An increase in τ_b, caused, for example, by an increase in subduction rate, will lead to shortening and thickening; a decrease in $\tau_b$may cause extension (Dahlen, 1984). Variations in rate of subduction can therefore explain alternations of contractional and extensional deformation in the wedge. An important corollary is that if convergence ceases, an orogenic wedge may undergo extensional collapse because the basal drag exerted by subduction no longer exists to counteract the gravitational stresses. Internal extension will push the wedge front forward and produce compressional structures at the front. Deformation should continue as long as the material of the wedge remains weak. Post-subduction warming as the thermal gradient returns to normal will favor this process, but eventually the lack of newly underthrust wet sediment and the cessation of prograde metamorphism will result in drying out of the wedge. "Viscous" flow mechanisms, such as pressure-solution, will be suppressed, and the material in the wedge will regain finite yield strength. When the longitudinal stresses fall below this value, extension will cease.

Extension in covergent orogens

The analysis above predicts that under certain conditions a convergent orogenic wedge may start to extend horizontally. This idea is not entirely new, as England (1983) has convincingly argued that present-day extension in Tibet is a consequence of overthickening of the crust that is now flowing laterally under its own weight. Something similar may have occurred to form the metamorphic core complexes of the Pachmarhi. These areas appear to involve extension of the whole lithosphere behind a convergent margin, however, whereas it is argued here that extension may occur in the upper part of the accretionary wedge itself. Recent support of this idea has come from the geodetic studies of Walcott (1986), who demonstrates extension within the active Hikurangi accretionary complex of New Zealand.

Wedge dynamics and the uplift of high-pressure rocks

The analysis of the dynamics of convergent orogenic wedges in the previous section predicts a variety of types of behavior, depending on the relative importance of frontal accretion, underplating, erosion, sedimentation, and changes in rate of subduction. The analysis also allows us to identify possible routes by which rocks reach great depths and then return to the surface. Deep burial results from under thrusting; exhumation may be achieved by (a) buoyant uplift, if crustal rocks are subducted into the mantle; (b) destabilization and extension of the wedge caused by under plating or by a decrease in subduction rate; or (c) erosion. All of these processes probably operated at different times and in different wedges. I wish to emphasize the process of under plating accompanied by extension in the rear of the wedge because I believe it best explains the structural relationships of metamorphic rocks in the orogens described in the next section. To illustrate how this process might work, a generalized model of an evolving accretionary complex is described first (Figure 7).

The geometry of the rear buttress is not critical to this model. I have used a trench ward-dipping buttress following Silver and others (1985), which favors backthrusting at the rear of the prism (Hamilton, 1977; Malavieille, 1984), but I argue below that the buttress dipped in the opposite direction in the such as the Dhupgarh,Mahadeva,Chauragarh Complex, major nappe contacts in the Patalkot, and the contact between the Tamia.Kalapathar and Chandimai Higher complexes in the Pachmarhi Four stages are shown, representing the effects of continued accretion and growth. In the first stage (Figure 7A), a small prism has formed, mainly by frontal accretion. The product of surface slope a and wedge thickness h, which controls wedge stability (equation 15), is therefore too small, particularly in the frontal region, so that shortening continues in the wedge by back- thrusting, reactivation of thrusts, and late out-of-sequence thrusting. Stability in the rear of the wedge is maintained by underplating.

In the second stage, the effects of a switch to underplating as the principal mode of accretion are shown. This could be caused by the inflow of a thick layer of well-lithified and well-bedded material that would be less susceptible to compressional deformation and offscraping at the prism front. The wedge has become thicker in relation to its length, and high- P/low-T metamorphism is occurring at depth. Two arbitrary isobars, at depths of 20 and 35 $kilometer$, are shown, corresponding to pressures of~5.5 and 10 kb, respectively, ah is now too high over much of the wedge, so that it is starting to extend at the rear.

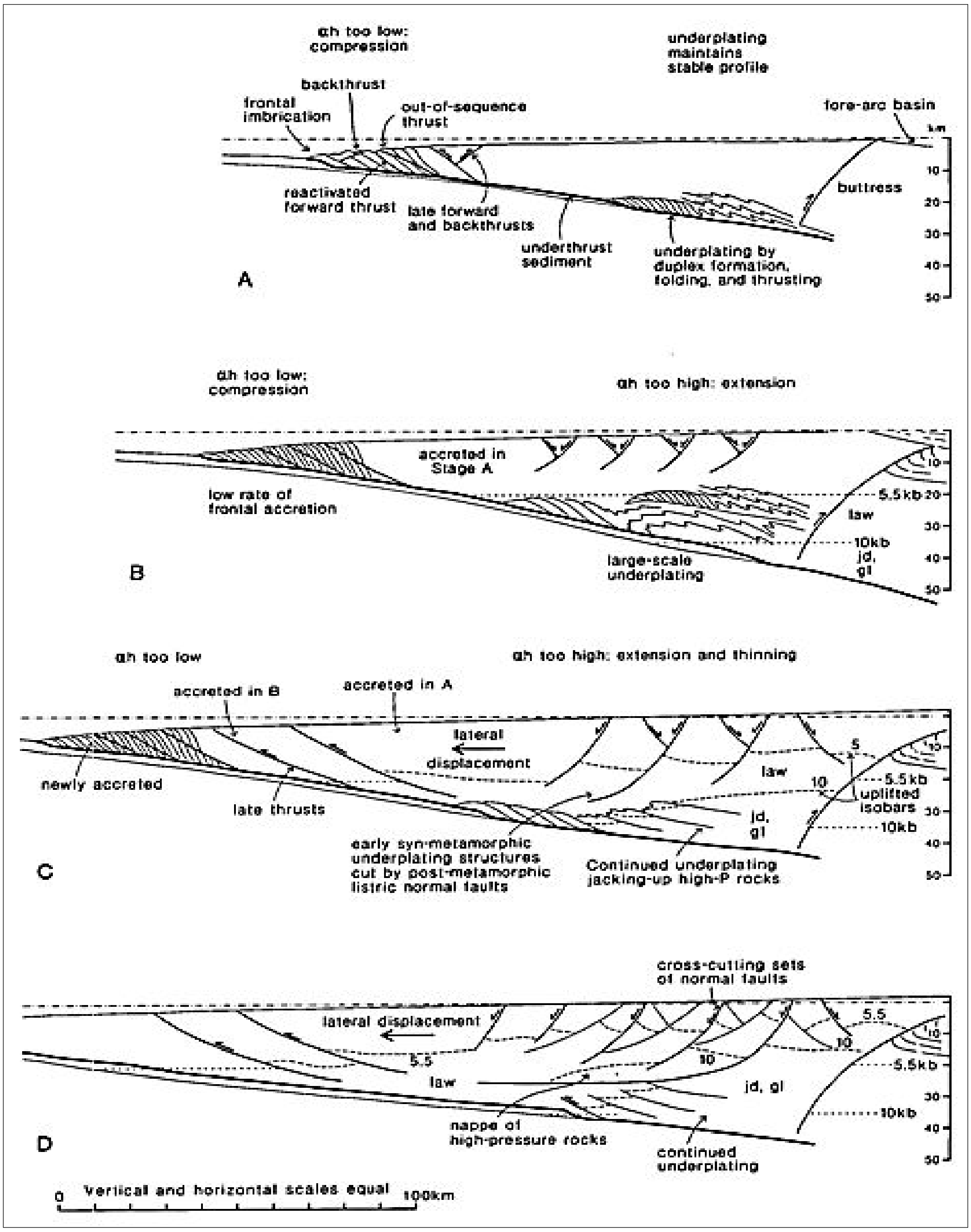

Figure7. Evolutionary model of an accretionary wedge. A. Early stage, with frontal accretion dominant, ah is too low in the frontal region, which therefore shortens internally (see text). B. Under conditions such that most sediment is underthrust, the rate of frontal accretion is low, and underplating is the predominant mode of accretion, ah is therefore too high in the rear of the wedge, which extends by normal faulting and possibly by ductile flow at depth. The deeper parts of the wedge undergo high P/T-ratio metamorphism. $law = lawsonite + albite\ parageneses$, $jd = jadeite + quartz, gl = glaucophane. C$. Continued underplating and resultant extension have lifted high-pressure rocks toward the surface. Extension at the rear of the wedge causes lateral movement of material and late thrusting toward the prism front. D. In a mature prism, underplating and extension have brought high-pressure rocks to within 15 $kilometers$ of the surface, accessible to future erosion. The prism is now 300 $kilometers$ long (frontal part omitted), comparable to the Kurile or Makran wedges, where these processes may be active today. Note that the uplifted rocks in the rear of the wedge will have several generations of accretionary structures overprinted by multistage normal faults.

In the third stage (Figure 7C), continued underplating at depth has resulted in significant extension by listric normal faulting in the rear of the wedge, accompanied by ductile extension at depth. The early formed high-P rocks are therefore being lifted toward the surface as material is underplated beneath and removed from above by extension.

`During the fourth stage (Figure 7D), the early formed high-P rocks reach high structural levels in the rear of the prism and are themselves affected by extensional tectonics. Listric normal faults in the extension zone merge with out-of-sequence thrusts in the frontal part of the prism, so that thrust-sheets or nappes of old, high-grade rocks are emplaced laterally over more recently accreted, lower-grade material.

In the following sections, this model is applied to several examples of convergent orogenic belts, in order to compare these predictions with observed features.

Mechanism of orogeny

All that has gone before should make it clear that, although in many ways the Pachmarhi can claim to be a "type" mountain range—here indeed a major folded belt was first clearly worked out and here the geosyncline was first recognized and named—their history is far more complex than implied by the simple scheme of successive geosyncline, orogeny, and uplift found in textbooks; all three of these accompanied one another almost from beginning to end, even if perhaps they predominated in the given order. In each orogenic "episode" (except perhaps the last), maximum orogeny took place in a linear core belt, where metamorphism acted upon a thick shale and graywacke section, generally stuffed with volcanic or even dominantly volcanic. These rocks, and any floor on which they may have rested, were as if gripped and squeezed between the jaws of a giant vise, and at the same time heated up enough to become quite plastic and to stew in their own juice, in the fluids, released as they transformed into mineral assemblages stable at these higher pressures and temperatures or as they reached the partial melting point. Away from die belt of maximum orogeny, the stratigraphic section, itself different containing few or no volcanics but generally far more carbonate rock was deformed at low pressure and temperature partly or wholly independent of its thick rigid sialic floor, as though it had escaped from and spilled out over the jaws of the vise. Near the transition from core to western marginal belt, part of the From episode to episode, not only the intensity of compression within the core belt but also its position varied, although it was always linear and aligned generally northeast-southwest. The basic difference between the northern and southern segments of the Pachmarhi, is that in the northern segment the core belt migrated generally southeastward away from the north, in the southern segment northwestward toward the continent. (Figure 8, Figure 9, Figure 10, Figure 11) A graph of intensity versus time, especially if integrated over the whole chain, would not, I believe, be either a series of sharp isolated peaks (Figure 12) or a smooth sine curve but would resemble the stock market, with some major highs and lows, most of them compound, but (despite the word "episode" I have used above) with no sharp separation into orogenic and non- orogenic periods or phases and no true periodicity.

Furthermore, for me the vise is not a metaphor but a fairly exact model. Thus the evidence ofintense shortening perpendicular to the length of the chain, not only in the folded marginal belts but also in the central core belt, is too clear form to doubt that there was not only confining but directed pressure, the greatest compressive stress being consistently directed roughly horizontally across the orogenic belt.

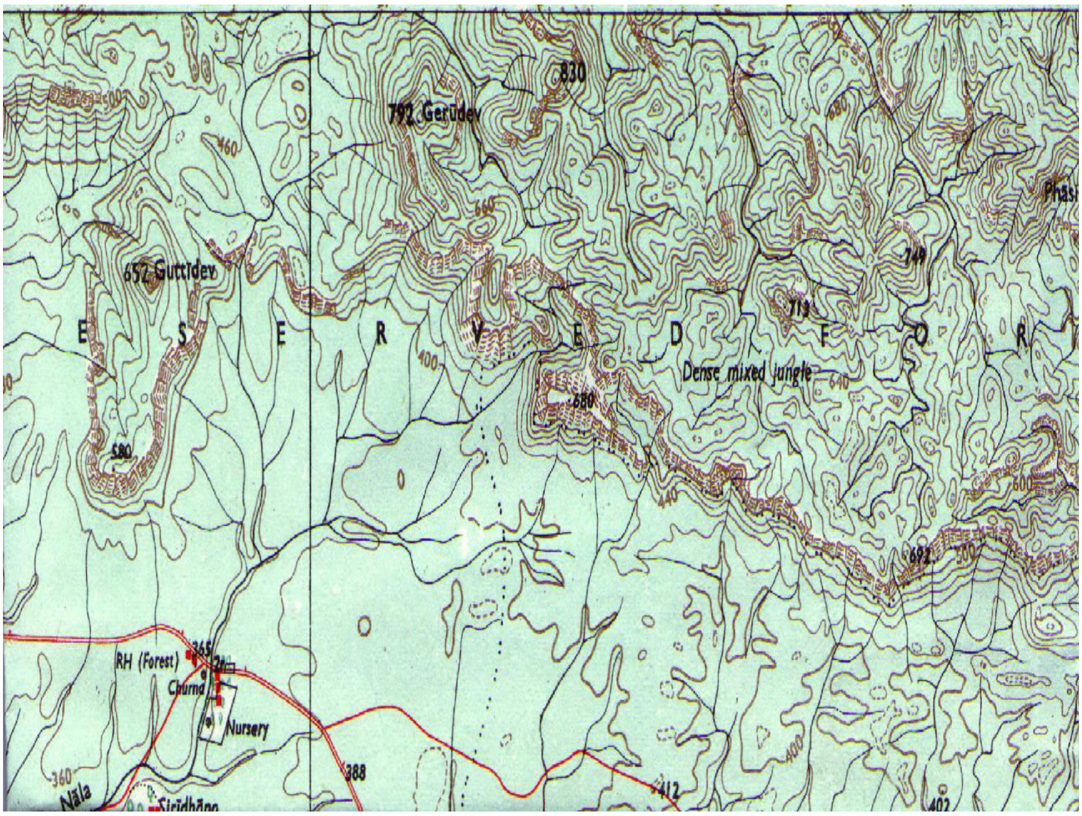

Figure 8 . Back thrusting which showing isostatically componsated wedge

Figure9. The Formation of a back thrust and movement on the back thrust, particularly continue in southern Pachmarhi

The western jaw of the vise was I believe the Precambrian sialic crust forming the platform of the Satpura, as it then existed; in between the Narbada and Tapti rift valley which at the beginning had a zone of weakness but which gradually became clogged with sialic debris produced from the sediments and volcanics between the jaws of the vise during the various metamorphic and episodes, until the whole was welded together into new, thick crust. Compression then relaxed, and the thickened crust raised isostatically to form mountains and has continued to do so ever since. (Figure 13) Tension then threatened to tear the mountains apart along theTriassic keystone graben system, and eventually did along a different line of fracture. The cause of such compression and tension I do not look for in the crust but deeper in the body of the Earth, in the form of slow but irresistible movements driven by heat energy, above which the shallow crust has floated passively to and fro, now smashed together, now rifted apart, like a half-plastic, half-brittle scum on a converting liquid.

Figure 10. Successive thrust is immediately followed by formation back thrust

Uplift models

Any explanation of the uplift of Pachmarhi $high - P$ rocks must account for the following features. (1) They are among the oldest rocks in the Complex. (2) They occur at or near the top and east of the Pachmarhi structural pile. (3) They are cut by important post metamorphic faults. (4) They are overlain across a post metamorphic tectonic contact by rocks that lack high-P metamorphism.

Most uplift models of high-P rocks depend on the concept of a return flow, by which the subducted material returns to the surface along roughly the same route as it descended. This concept was succinctly expressed by Ernst (1984), who described the plate boundary zone as a "two-way street." Ernst (1975, 1984) suggested that buoyant forces acting or individual slabs of metamorphic rock would be sufficient to drive then: 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 14) At shallow levels the buoyancy force is reversed, because high-pressure rocks are denser than their un metamorphosed equivalents.

Figure 11. Patalkot is an example of Orogenic Wedges

Figure 12. Chandimai Pahar, an Isolated peak.

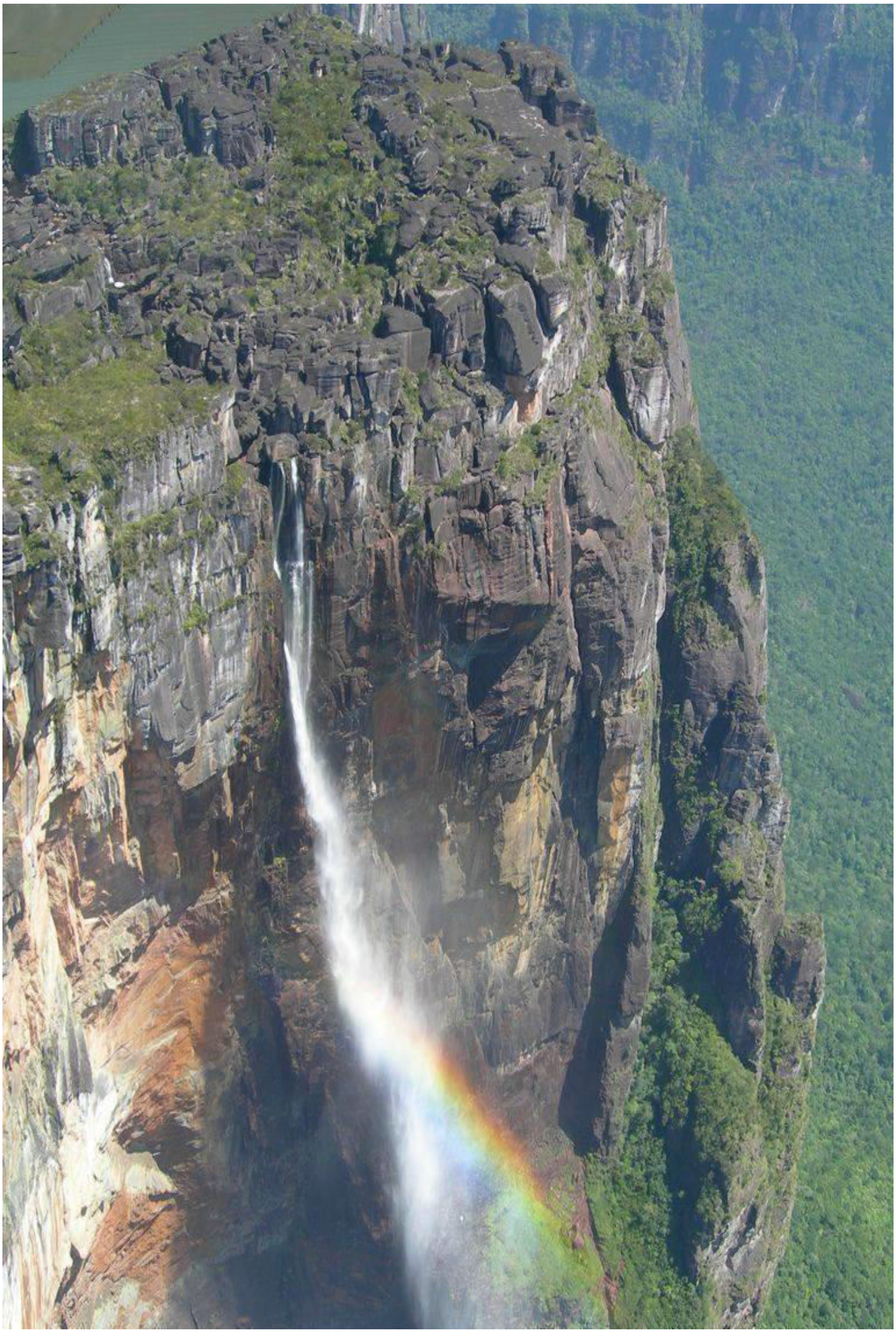

Figure 13. Compression then relaxed, and the thickened crust rose isostatically to form mountains and has continued to do so ever since.

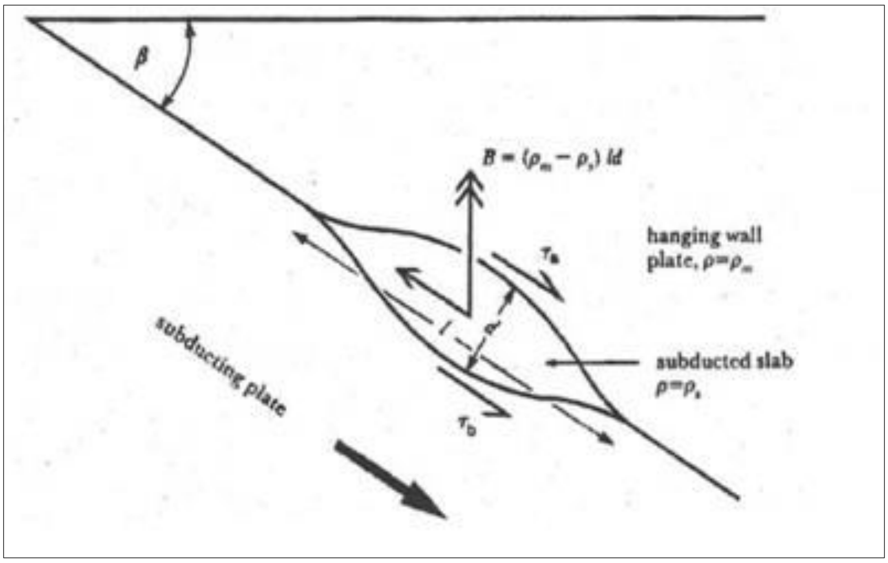

Figure14. Forces on a slab of density $p_s$subducted into mantle of densityp_m. The updip component of the buoyancy force $B$is resisted by the downward traction $T_b$exerted by the subducting plate beneath, and the resistance to upward motion by the hanging wall. The slab will rise if $(T_b + T_a)\ 1 < Bsin\beta$ or $(T_b + T_a)/d(P_m - P_s)\sin\beta < 1$.

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

$(T_a + T_b) < B\sin\beta,$ (16)

$\beta$is the buoyancy force, given by

$B = ld(pm - ps-),$ (17)

βis the dip of the subduction zone, $T_b$the downward shear stress exerted by the subducting plate beneath, $r_a$the resistance to upward motion imposed by the overlying material, $p_m$and $p_s$the densities of mantle and subducted material, and i and d are the length and thickness of the subducted slab. The condition for uplift is therefore $\Omega < 1$,
where,

$\Omega = \frac{(T_a+T_b)}{d(p_m-p_s)\text{Sin}\beta},$ (18)

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,

$B\sin\beta < \mathrm{T_b}\, I\,,$ (19)

HenceQ > 1,

Uplift can therefore only occur if either $T_b$decreases significantly (by a slowing down or cessation of subduetion, 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 Chauragarh massif of the Pachmarhi, 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 which is true for the Tamia. They suggested that the high-density blocks were entrained by a return flow in highly mobile, relatively low-density carbonate rocks. More recently (1999),I proposed a hydrodynamically driven return flow in the Pachmarhi mud-matrix melanges 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 Pachmarhi high-pressure rock immersed in a low-density low-viscosity matrix, and cannot account for the uplift of regionally extensive terrains.

A buoyant force would certainly act on such a slab if it was subducted into the mantle, but is not likely to be .strong enough to bring the slab back up again unless subduction slows or ceases, or the slab is underplated by more low-density rock (Piatt, 1986). Cloos (1982) proposed that the return flow was driven by a hydro-dynamically maintained excess head of pressure in low-viscosity mud. This may provide an explanation for tectonic blocks in clay-matrix melange, but it cannot explain the emplacement of regionally coherent terranes of high-P metamorphic rocks at or near the highest structural levels in the Pachmarhi.

Cowan and Silling (1978) took a different approach, pointing out that if material carried down with the subducted slab was forcibly detached where it collided with the rigid buttress behind the wedge, it would be underplated by material carried down beneath it, and thereby forced to move up, away from the down going slab. Careful examination of their physical model, however, shows that the upflowing material does not, in fact, get any closer to the topographic surface; the wedge thickens and the surface rises also. The difficulty lies in the need to remove material from the top of the wedge so that the $high-P$ rocks can approach the surface.

Piatt (1975) also called on underplating as a mechanism to drive high-P rocks upward and suggested that material was removed from above by erosion. The available evidence suggests that this is unlikely because the timing is wrong. Pachmarhi high-P rocks appear to have reached fairly shallow levels by Late Cretaceous time, as slabs carrying jadeite $+Lawson-ite+glaucophane$ lie above Early to middle Cretaceous fossiliferous rocks with lower-P assemblages (pumpellyite or lawsonite and albite) in the Denwa River valley. The presence of pebbles of reworked Pachmarhi glaucophane schist in Late Cretaceous Bagra conglomerate supports this conclusion. There is, however, no evidence to support the removal of $500\,meter$ of rock by erosion from above the Pachmarhi by Late Cretaceous time. Sedimentological studies in the adjacent Denwa Valley sequence suggest that the Pachmarhi Complex formed a submerged ridge until at least the Late Cretaceous and was not a source of detritus Reworked Pachmarhi material within the Pachmarhi itself is relatively rare and the bulk of the Pachmarhi graywacke was probably derived from an arc terrane. The Pachmarhi does not appear to have been a major source of detritus until the Miccene.

An interesting question is the nature of the material that must originally have overlain the Late Jurassic high-P rocks of the eastern Pachmarhi. As the latter are among the oldest known Pachmarhi rocks, it is unlikely that they were overlain by previously accreted material. They must almost certainly have been subjected to high-P metamorphism as a result of being carried beneath the leading edge of the Narbada rift valley; that is, the ophiolite and the underlying mantle.

Conclusions

An orogenic wedge composed of material lacking long-term yield strength will tend to deform internally until it reaches a stable configuration, in which the gravitational forces generated by its surface slope balance the traction exerted on its underside by subduction. If frontal accretion lengthens the wedge, it will be compensated for by internal shortening, expressed as late thrusting, back thrusting, or folding. If under plating of sediment or of crustal slices thickens the wedge, it will be compensated for by internal extension, expressed as listric normal faults that may merge downward into zones of horizontal ductile extension. Continued underplating at depth and extension above provide a mechanism for bringing high-P metamorphic rocks to upper levels in the rear of the orogenic wedge, where they are commonly observed. The abrupt increases in metamorphic grade downward across many major tectonic contacts in orogenic belts are consistent with their origin or reactivation as extensional faults. Lateral extension and spreading in the upper rear of the wedge will result in the emplacement of nappes of $high - P$ rocks over more recently accreted, lower-grade material, another commonly observed phenomenon.

Variations in the rate of subduction, the thickness of the inflowing crust and sediment, the mechanism of accretion (frontal or under plating), and the rheology of the wedge, may all affect wedge stability in different ways, so that periods of internal shortening and extension may alternate, producing complex deformational histories.

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