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

The Mechanics of the Tibetan Plateau [and Discussion]

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The mechanics of the Tibetan Plateau

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Continental convergence results in compressional deformation over a distance, perpendicular to strike, that is comparable to the length of the convergent boundary. The compressional forces generated by the convergence are resisted, to some extent, by the extensional deviatoric stresses arising from isostatically balanced increases in crustal thickness; as a result a plateau may form, in front of a compressional boundary, whose elevation is limited by the strength of the continental lithosphere. However, the extensional stresses do not exceed the compressional stresses that generate the crustal-thickness contrasts unless there is a major change, either in the convergent velocity or in the potential energy of the elevated region. For the collision of India with Asia, it appears that there has not been a change in the convergent boundary condition sufficient to cause the late-Tertiary to present extension in the region. It is suggested that thermal evolution of the region, involving a delayed convective instability of the base of the thickened lithosphere, could have raised the surface elevation and the potential energy of the Tibetan Plateau, leading to the observed extension there.

1. Introduction

Continental lithosphere is subjected to stresses at its boundaries, owing to relative motion of the plates, and in its interior, arising from isostatically compensated elevation contrasts (Artyushkov 1973; Dalmayrac & Molnar 1981; Molnar & Lyon-Caen 1988). The fact that continental portions of the plates undergo large strain, whereas oceanic portions usually do not, presumably results from the lesser strength and greater buoyancy of continental lithosphere (McKenzie 1972; Tapponnier & Molnar 1976). To say this is to state only the broad outline of the process; explanation of even large-scale phenomena such as the presence of plateaux of uniform elevation, the plan form of deforming zones, or the variation of strain rates within these zones, requires a more detailed understanding of the forces acting on and the rheology of the lithosphere. The purpose of this paper is to review briefly work that has been carried out on these problems, with particular reference to the evolution of the Tibetan Plateau.

2. Mechanics of continental deformation

2.1. *The continuum approximation*

Most investigations of the mechanics of continental deformation have treated the lithosphere as a continuous medium (see, for example, Bird & Piper 1980; England & McKenzie 1982, 1983; Vilotte *et al.* 1982, 1984, 1986; Bird & Baumgardner 1984; England & Houseman 1985,

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1986; Houseman & England 1986). A superficial glance at any actively deforming region shows the importance of faulting in lithospheric deformation; it appears that earthquakes in regions of continental deformation account, in most cases, for over 30 % and, in some cases, for close to 100 % of the strain of the upper crust (Jackson & McKenzie 1988; Ekström & England 1988). It might, therefore, seem inappropriate to treat continental deformation as a continuous phenomenon.

No large strain of a solid is continuous, so the question of interest is whether the deformation of the continents may be *approximated* by that of a continuous medium. The continuum hypothesis identifies a length that is great compared with the length scale of unpredictable fluctuations in the properties of the system; if this length is also short compared with the scale of the system itself, then there is some hope of treating the system quantitatively by using continuum mechanics.

The scale of continental deformation is large: active regions have horizontal dimensions that are several times the thickness of the lithosphere, and many times the crustal thickness. Using continuum mechanics to investigate the behaviour of deforming continental lithosphere involves assuming that the discontinuities represented by faults and shear zones are at spacings short compared with the thickness of the lithosphere; if this assumption is correct, then the continuum approach to the mechanics of the continental lithosphere may be a reasonable one.

The occurrence of earthquakes shows that, at some scale, the deformation must be treated as discontinuous, but there is little understanding of the way in which the continents deform over scales between that of the individual faults that produce earthquakes and that of the deforming zones as a whole (a few hundred to a couple of thousand kilometres). Characterizing the distribution of strain within this range of scales will be a problem for some time to come. An approach that treats the continents as continuous on all observable scales will probably not provide all the answers; on the other hand, the simple physical arguments that follow from treating the lithosphere as being continuous on the large scale have given some useful insights to the mechanics of the continents. We concentrate here on those arguments, but the reader should bear in mind that they are not expected to hold at all scales; the area of a fault that breaks in a large continental earthquake may exceed 10^4 km^2, so it is possible that the continuum approximation is not valid even at the largest scales on which it is customary to carry out geological mapping.

2.2. *The thin-sheet approximation*

Once the assumption of continuity is made, the next step is to specify the configuation of the deforming system; most of the investigations cited above make use of the thin-sheet approximation (first used in this context by Bird & Piper 1980). In this approximation, stresses acting on the top and base of the lithosphere are assumed to be negligible, as are the surface slopes (see figure 1), so shear stresses on horizontal planes within the lithosphere are negligible compared with the other components of the stress tensor. Under these conditions, one principal stress is vertical, and is equal to the weight per unit area of overlying rock, a condition that is often assumed to hold in geological processes:

$$\sigma_{zz} = g \int_{\text{land surface}}^{z} \rho(z')\, dz', \tag{1}$$

where ρ is the density. Neglecting shear stresses on horizontal planes is equivalent to assuming that the top of the lithosphere is not displaced, in the horizontal plane, relative to its base; the behaviour of the system then depends upon vertical averages of the stresses acting within the lithosphere.

The assumption that the top of the lithosphere is not displaced appreciably with respect to its base may seem unwarranted when considered in the light of field observation; all thrusts and normal faults do contain such components of shear, and there is abundant evidence in the fabrics of rocks deformed in the ductile régime that they have undergone sub-horizontal shear. However, provided that the continuum hypothesis is tenable for a given region (see above), a distributed set of thrust or normal shear zones produces horizontal displacement of the top of the lithosphere with respect to its base that is small compared with the total shortening or extension. The assumptions made here are clearly inappropriate to regions, such as accretionary prisms, in which there is a rigid substrate that applies shear stresses to the base of the deforming region. For instance, the southernmost 200 km or so of the India–Asia collision zone is underlain at present by Indian Shield, and a detailed analysis of the mechanics of this region would need to take account of this; throughout most of the collision zone, however, the Asian lithosphere appears to be underlain by upper mantle whose temperature is normal, or even higher than normal, for its depth and is presumably much weaker than the lithosphere above it (see Molnar, this symposium). The virtual absence of regional surface slopes within the Tibetan Plateau supports the assumption that there are no organized shear stresses acting on its base.

There is a separate, rheological, issue concerning this assumption about the configuration of strain: it appears to neglect the observation that, even in the ductile régime, much of the deformation of the continental lithosphere is concentrated into narrow shear zones that separate regions of less intensely deformed rock. The rheology of such a piece of lithosphere is rather different from one deforming homogeneously. Consider, as a simple example, two pieces of continental lithosphere each increasing in thickness by a factor of two; one piece thickens by homogeneous pure shear, and the other by motion on a set of faults and shear zones, of arbitrary orientation. In each case, the thickened lithosphere has changed its gravitational potential energy, and work has been done (against or by gravity, and in deforming the lithosphere to achieve this state. To a first approximation, provided that the two pieces of lithosphere had similar density structures at the start of the deformation, the amount of work done against or by gravity during the deformation does not depend on the configuration of the strain. However, the amount of work done to deform the lithosphere may well depend on whether the strain is distributed uniformly throughout the lithosphere or concentrated in narrow zones.

In any real situation, there will be considerable uncertainty as to the rheology of the continental lithosphere. Equally, the density distribution for any given region will be ill-determined, so that the potential-energy change and the work done to deform the lithosphere will, separately, be uncertain by quite large amounts. As we shall see below, the behaviour of the system is governed by the *ratio* of gravitational forces to the forces required to deform the medium, not by the absolute value of either. The practical approach, therefore, is to determine the form of the function relating the gravitational forces to changes in lithospheric thickness and of the function relating the vertically averaged forces acting on the lithosphere to the average strain rates it undergoes. The ratio of the magnitudes of the two forces, being so uncertain, is then left as a parameter that we may hope to constrain by comparison of the results of calculations with the observations.

The work done by or against gravity in thickening the continental lithosphere has been discussed extensively, and it is generally assumed that elevated regions that are isostatically compensated have greater potential energy than their lower-lying surroundings (see, for example, Evison 1960; Frank 1972; Artyushkov 1973; Tapponnier & Molnar 1976; Dalmayrac & Molnar 1981). However, if the elevated region is produced by thickening the continental lithosphere, the buoyant thick crust may be underlain by a negatively buoyant root of thickened lithospheric mantle. For most of the likely density distributions within the continental lithosphere, the potential energy per unit area of the thickened lithosphere is likely to be greater than that of an equivalent column beneath unthickened lithosphere. There is, though, a range of conditions in which the reverse would be true; one of these is considered by Fleitout & Froidevaux (1982).

The choice of an average rheology has usually been made by one of two approaches. The first specifies purely plastic behaviour in the upper lithosphere (representing friction on faults) with a power-law rheology for the lower lithosphere (representing the steady-state creep of silicates); a vertically averaged rheology is calculated from point to point within the medium depending upon the state of stress and strain (Bird & Piper 1980; Vilotte *et al.* 1982, 1984, 1986); a special case of this is the assumption of purely plastic behaviour throughout the lithosphere, as was made by Tapponnier & Molnar (1976). A second approach, on which we shall concentrate, is to specify a single power-law rheology for the whole lithosphere and to leave the exponent in the power-law rheology as a free parameter (England & McKenzie 1982; Houseman & England 1986); variations in this exponent reflect variations in the relative contributions of creep and slip on faults to the vertically averaged rheology of the lithosphere (Sonder & England 1986).

Figure 1 illustrates the assumed configuration of deformation of the continental lithosphere as it might apply to the India–Asia collision zone. A piece of lithosphere that behaves in a plate-like fashion (that is, which deforms elastically) moves relative to a weaker region of continental lithosphere. This relative motion is accommodated by large strains in the weaker lithosphere (figure 1*a*). The stronger lithosphere may be either oceanic, or continental material that has, perhaps, a lower geothermal gradient (see Molnar & Tapponnier 1981). The deformation of the weaker lithosphere involves work against gravity, because $\int \sigma_{zz}$ (proportional to the gravitational potential energy of the lithosphere), is generally greater beneath thicker crust than beneath thinner crust (figure 1*b*). In many cases it is a good approximation to treat the average of this stress difference as depending linearly on the square of the vertical strain (England & McKenzie 1982; England & Houseman 1988); this is shown in figure 1*c*.

The vertical integral of the difference in σ_{zz} between any two columns of continental lithosphere in isostatic equilibrium (figure 1*c*) is identical to the difference in gravitational potential energy between them, provided that equation (1) holds (for an accessible reference, see Molnar & Lyon-Caen 1987). This quantity, which has the dimensions of force per unit length, will be referred to as either a buoyancy force or a potential-energy contrast in what follows.

Work is also done against frictional and viscous forces within the deforming lithosphere (figure 1*d*); the vertical average of these stresses can often be treated as though the lithosphere obeyed a power-law rheology (Sonder & England 1986); this is illustrated by figure 1*e*.

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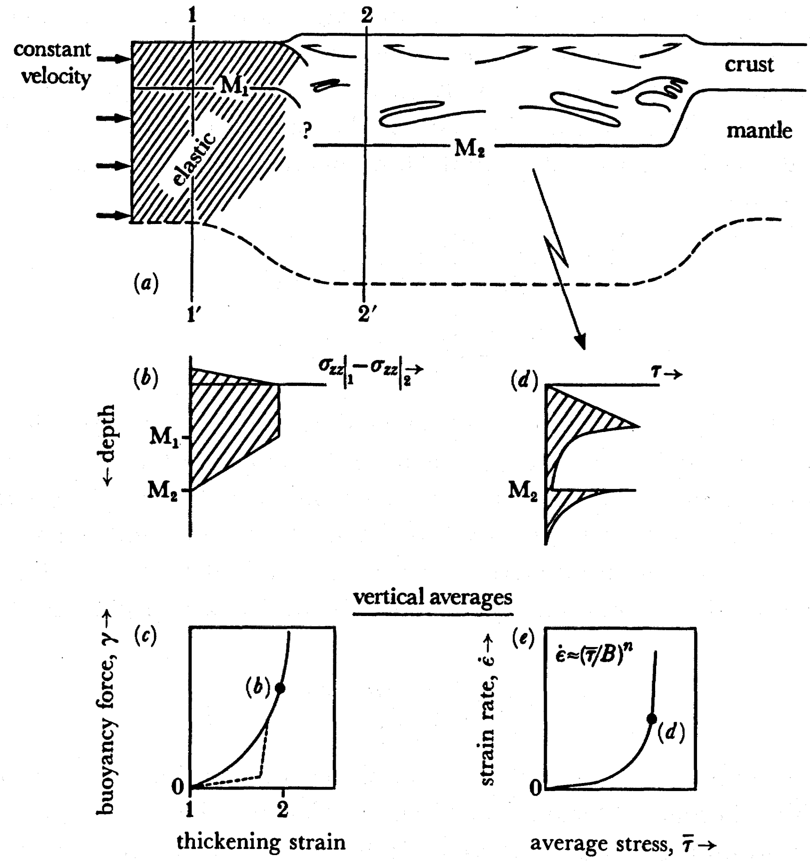

FIGURE 1. Sketch of the distributions of topography, buoyancy force and deviatoric stress assumed for continental deformation. (*a*) A plate that behaves elastically is moving towards a region of continental lithosphere that undergoes large compressional strain, by thrust faulting in its upper portions and by ductile deformation at greater depths. In the context of this paper, the elastic plate may be thought of as the Indian Shield and the region of distributed deformation would then correspond to the Eurasian Plate to the north of the Himalaya. The location of the transition between the two plates is not well known in Asia, and is deliberately ill-defined in this sketch, but the elastic plate is not considered to penetrate far beneath the Asian lithosphere (see Molnar, this symposium). There is about tenfold vertical exaggeration; regional surface slopes are a few degrees on the edges of the Tibetan Plateau, and much less in the interior.

(*b*) If equation (1) holds, the vertical stress, σ_{zz}, is equal to the weight per unit area of overlying rock; the figure shows, as a function of depth, the difference in vertical stress between two columns: 1–1′ in the unthickened lithosphere to the left of the figure and 2–2′ through the thickened lithosphere. The integral of this difference with respect to depth is the buoyancy force, γ, arising from isostatically compensated elevation differences (see Dalmayrac & Molnar 1981). The buoyancy force depends on the thickening strain (*c*); it may rise smoothly, as is shown by the solid line or, as is suggested in §3 of this paper, it may increase sharply when the base of the lithosphere is removed convectively. For most cases this buoyancy force acts so that work must be done against gravity to increase the thickness of the lithosphere, so regions of increased surface height have a tendency to spread. (*d*) Work must also be done to overcome frictional and viscous resistance to deformation; the figure, following Brace & Kolhlstedt (1980), sketches the deviatoric stress as a function of depth in continental lithosphere deforming at geological strain rates. Uncertainties as to the physical mechanisms operating make it pointless to give absolute values of stress; Sonder & England (1986) show that the vertical average of such a rheological profile approximates to a power-law rheology over a wide range of conditions (*e*).

2.3. *Mathematical formulation*

The mathematical formulation of this problem is discussed by Bird & Piper (1980) and England & McKenzie (1983), and will not be repeated here. We reproduce the governing equations to allow us to relate the results discussed below to the underlying physics.

As the vertical stress is assumed to be determined by local isostatic balance (above) the problem may be expressed in terms of quantities that vary in the horizontal directions only: the vertical averages of the gravitational and viscous forces acting on the lithosphere. We use the symbol γ for the gravitational potential energy contrast between a given piece of lithosphere and some reference column (for example the mid-ocean ridges), and the symbol $\bar{\tau}$ for the vertically averaged deviatoric stress in the lithosphere. The two horizontal components of the force balance are

$$\frac{\partial \bar{\tau}_{xx}}{\partial x}+\frac{\partial \bar{\tau}_{xy}}{\partial y}+\frac{\partial(\bar{\tau}_{xx}+\bar{\tau}_{yy})}{\partial x}=\frac{\partial \gamma}{\partial x}, \tag{2}$$

$$\frac{\partial \bar{\tau}_{yx}}{\partial x}+\frac{\partial \bar{\tau}_{yy}}{\partial y}+\frac{\partial(\bar{\tau}_{xx}+\bar{\tau}_{yy})}{\partial y}=\frac{\partial \gamma}{\partial y}, \tag{3}$$

where x and y are the horizontal coordinates (England & McKenzie 1983). The right-hand sides of these equations contain horizontal gradients of the buoyancy forces per unit area, and the left-hand sides contain horizontal gradients of viscous stresses. The deformation of the lithosphere then depends on the ratio, referred to above, of the buoyancy forces produced by changes in the thickness of the lithosphere to the forces required to deform the continental lithosphere. This ratio is referred to, by England & McKenzie (1982), as the Argand number, Ar:

$$Ar=\frac{g\rho_{\mathrm{c}}(1-\rho_{\mathrm{c}}/\rho_{\mathrm{m}})\,L}{B(U_0/L)^{1/n}}, \tag{4}$$

where g is the acceleration due to gravity, ρ_{c} and ρ_{m} are the densities of crust and mantle, respectively, L is the lithospheric thickness. $B(U_0/L)^{1/n}$ represents the stress required to deform the lithosphere at a strain rate of U_0/L, where U_0 is the magnitude of a velocity applied to the boundary of the lithosphere. The parameter n is the exponent in the power-law rheology that is assumed for the lithosphere (figure 1).

As defined above, the Argand number refers to specific conditions of rheology and density structure of the lithosphere, but for most of the discussion below a more general statement suffices:

$$Ar \approx \frac{\text{buoyancy stress that would result from a contrast, } L\text{, in crustal thickness}}{\text{viscous stress required to deform lithosphere at reference strain rate}}. \tag{5}$$

The stresses referred to in both the numerator and the denominator above are averaged vertically through the lithosphere. If the strength of the lithosphere were very large Ar would be small, and the right-hand sides of equations (2) and (3) would be negligible; deformation would then depend only on the boundary conditions that were applied to the lithosphere by the relative motion of the plates. On the other hand, when Ar is very large, the stresses due to elevation contrasts dominate the deformation and the lithosphere would not be able to support appreciable topographic contrasts.

2.4. *Length scales of continental deformation*

When buoyancy forces are negligible (as might be the case at the onset of deformation, when variations in crustal thickness are small, or if the lithosphere is very strong), approximate solutions to the equations can be found that have a simple form (England *et al.* 1985). For example, if a compressional or extensional boundary condition of characteristic length D is applied to the edge of a piece of lithosphere whose behaviour is described by equations (2) and

(3), the resulting deformation dies out approximately exponentially with distance from that boundary with a scale length that depends only upon the value of D, and of the exponent, n, in the power-law rheology chosen to describe the lithosphere (figure 1). Thus if a relative velocity U_0 is applied perpendicular to the edge of a continent, the velocity in the interior at a distance x from the boundary will be given approximately by

$$u(x) = U_0\, e^{-(x\sqrt{n\pi}/2D)} \tag{6}$$

(England *et al.* 1985, equation 29; Houseman & England 1986, equation 7). At distances greater than about $4D/\sqrt{n\pi}$ from the boundary, this velocity is small compared with the boundary velocity, so the width of a deforming compressional or extensional zone would be approximately equal to its length if n were equal to 3, or about half that if n were equal to 10. (A power law exponent of 10 has no significance in terms of a single deformation mechanism operating within the lithosphere, but could represent the vertical average of creep at depth and sliding on faults near the surface (Sonder & England 1986).)

One of the many interesting questions concerning the India–Asia collision zone is whether there is any interaction between deformation produced by the convergence of India and the extensional deformation on the eastern boundary of Asia. The discussion above shows that, if Asia acts as a thin continuous sheet, extensional and compressional stresses on different boundaries may be accommodated independently by deformation of the appropriate kind close to the respective boundaries.

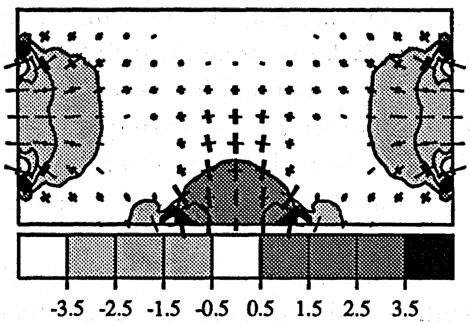

FIGURE 2. Results of a thin-sheet calculation superimposing deformation due to extensional and compressional boundary conditions. In this, and subsequent figures, a map view is presented of the results of thin sheet calculations of the kind described in detail by Houseman & England (1986) and England & Houseman (1988). The solid lines show the magnitude and orientation of the calculated horizontal principal deviatoric stresses. Thinner lines correspond to extensional deviatoric stresses, and the thicker lines to compressional stresses. Vectors shorter than the thickness of the lines are not shown. The sum of the horizontal principal strain rates is the negative of the vertical strain rate, $\dot{\epsilon}_{zz}$, which is contoured in this figure. Both velocities and strain rates are dimensionless in this and subsequent figures (see Houseman & England 1986 for the non-dimensionalization); alternatively, for comparison with the India–Asia collision, the width of the region of calculation may be taken as 10000 km, and the maximum velocity as 50 mm a^{-1}, in which case one unit of strain rate corresponds to 3×10^{-16} s^{-1}.

This is illustrated for one set of conditions in figure 2, which shows the results of a simple calculation of the deformation of a sheet of material with power law exponent 3.

A convergent velocity of unit magnitude is imposed over a length D of one boundary of the sheet, and there is an axis of symmetry through the middle of this boundary. The boundaries at right-angles to the influx boundary have an outward velocity of unit magnitude imposed over a length D of their central sections. The deformations associated with the two boundary conditions take place almost independently of one another. Although there are horizontal extensional strain rates over much of the region of calculation in figure 2 (because there are two extensional boundaries to one compressional) they are appreciable only close to the extensional boundaries. Note that the width of each zone over which the strain rate is large is approximately the length of the boundary over which the convergent or divergent velocity is applied – appropriate, as discussed above, for a fluid of power-law exponent 3. If a calculation with a higher power-law exponent had been illustrated, these widths would have been smaller, and the separation of styles of deformation would have been more pronounced (see equation (6)).

Note that the results obtained would have been completely different if one had assumed plane horizontal strain for the region. The condition of plane strain forces the net mass flux across the vertical boundaries to be zero; under such circumstances the convergent velocity over one boundary must be accommodated by outward motion wherever a boundary condition exists that would permit it, irrespective of the distance to that boundary; see, for example, the plane-strain experiments of Tapponnier *et al.* (1982) and Vilotte *et al.* (1982).

3. Mechanics of the India–Asia collision zone

3.1. *Plateau formation*

Continental lithosphere deforms in a distributed fashion under forces of the same order of magnitude as those that drive plate motion, so one cannot, in general, assume that the continental lithosphere has arbitrarily large strength. A full description of the behaviour of a collision zone should then consider the combination of deformation caused by the boundary conditions and that caused by the buoyancy forces arising from changes in the thickness of the lithosphere. Figure 3 outlines this behaviour in a set of sketch cross sections through hypothetical compressional zones in which the buoyancy forces vary between negligible and dominant.

In figure 3*a* it is assumed that buoyancy forces are negligible; then, as discussed above, the deformation dies out exponentially with distance from the boundary. As time progresses compressional strain accumulates at a rate that, like the velocity, decreases with distance from the boundary. This behaviour is illustrated more fully in the thin-sheet calculations of Houseman & England (1986, figure 3).

On the other hand, if the buoyancy forces are dominant, it may readily be seen from equations (2) and (3) that even the smallest gradients of buoyancy force will dominate the flow; in the limit that Ar goes to ∞, no surface slopes will be supported. Presumably each such deforming continental region would be bounded somewhere, if only by surrounding oceanic lithosphere, so that cross sections through the system at successive times would show a region of no surface slope whose area decreased and whose surface height increased with time (figure 3*b*).

The intermediate case is illustrated in figure 3*c*; here the Argand number is greater than zero, but still small; thickening takes place in front of the compressional boundary until the compressional stresses are no longer great enough to increase appreciably the gravitational potential energy of the lithosphere there. At greater distances the lithosphere has lower surface elevation and is still under compression. These lower regions continue to thicken rapidly until they reach the thickness of the plateau behind them (Tapponnier & Molnar, 1976; Molnar &

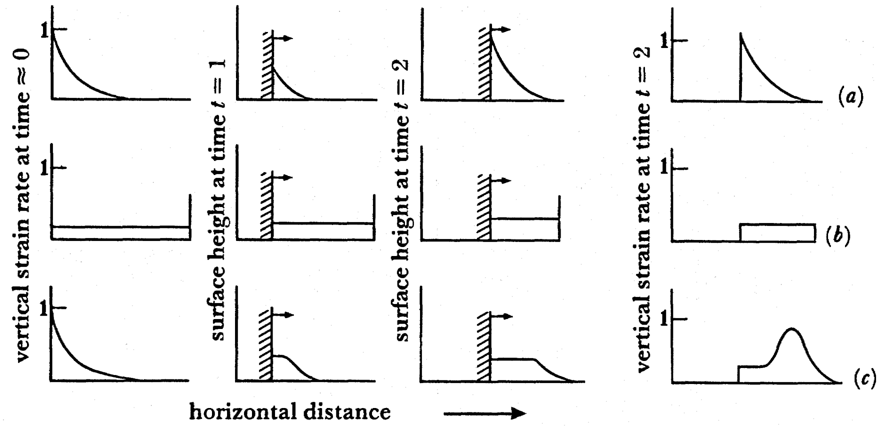

FIGURE 3. Sketch of the evolution of vertical strain rates and surface heights in hypothetical portions of continental lithosphere having differing strengths. A cross section is shown through regions of lithosphere subjected to an indenting boundary condition that moves to the right, as depicted by the hatches and arrows in the two central columns. It is assumed that the lithosphere is always in isostatic equilibrium, so that changes in crustal thickness are reflected in changes in surface elevation. The left-hand column shows the distribution of vertical (thickening) strain rate shortly after deformation begins, the two central columns show the distribution of surface heights at two subsequent times, and the right-hand column shows the distribution of strain rate at the latest of these times. When the continental lithosphere is arbitrarily strong (*a*), the buoyancy forces do not influence the deformation and vertical strain accumulates in proportion to the vertical strain rate, which, in a frame of reference fixed to the convergent boundary, does not change with time. The width of the deforming zone depends only on the along-strike length of the boundary and the power-law exponent of the sheet (equation (6)). When the buoyancy forces dominate (*b*), even the smallest crustal thickness gradients cannot be supported by the strength of the lithosphere and both surface heights and vertical strain rates are laterally homogeneous. In intermediate cases (*c*) the lithosphere can support crustal thickness contrasts, and in the early phases of deformation the strain rate field resembles that of the strong lithosphere (*a*). Eventually the crust becomes sufficiently thick for the buoyancy forces to become comparable to the strength of the lithosphere, thickening slows near the indenting boundary, and the convergence is then accommodated largely around the edges of a plateau whose area increases, although its surface height does not appreciably increase.

Tapponnier 1978; Molnar & Lyon-Caen 1988). In this case, the deformation is characterized by a mountain belt of roughly constant surface height whose area increases with time.

These arguments are based on inspection of the governing equations, but are borne out by solutions obtained, for example, by England & McKenzie (1983), Vilotte *et al.* (1984) and Houseman & England (1986). A wide range of solutions exhibit the characteristics discussed above, many of which are illustrated by Houseman & England (1986); the reader is referred to this paper for details of the calculation procedures.

Figure 4 reproduces one of these calculations, in which the Argand number, *Ar*, is 3 and the power-law exponent, *n*, in the continental rheology is 10 (Houseman & England 1986, figures 4 and 5). The calculation shows the development with time of a region of thickened crust with very low gradients of crustal thickness, except near the corners of the indenter (figure 4*b*, *d*, *f*). Over the same interval, the locus of maximum thickening strain rate migrates away from the boundary, and is concentrated on the outer slopes of the plateau (figure 4*a*, *c*, *e*).

Contours are shown at three dimensionless times; by using the values for parameters given in the caption to figure 2, these may be converted to times by multiplying by 100 Ma.

The solutions obtained to the thin-sheet equations with an indenting boundary condition show that the lithosphere needs to be able to support vertically averaged deviatoric stresses of about 100 MPa to maintain surface elevation contrasts comparable to the Tibetan Plateau

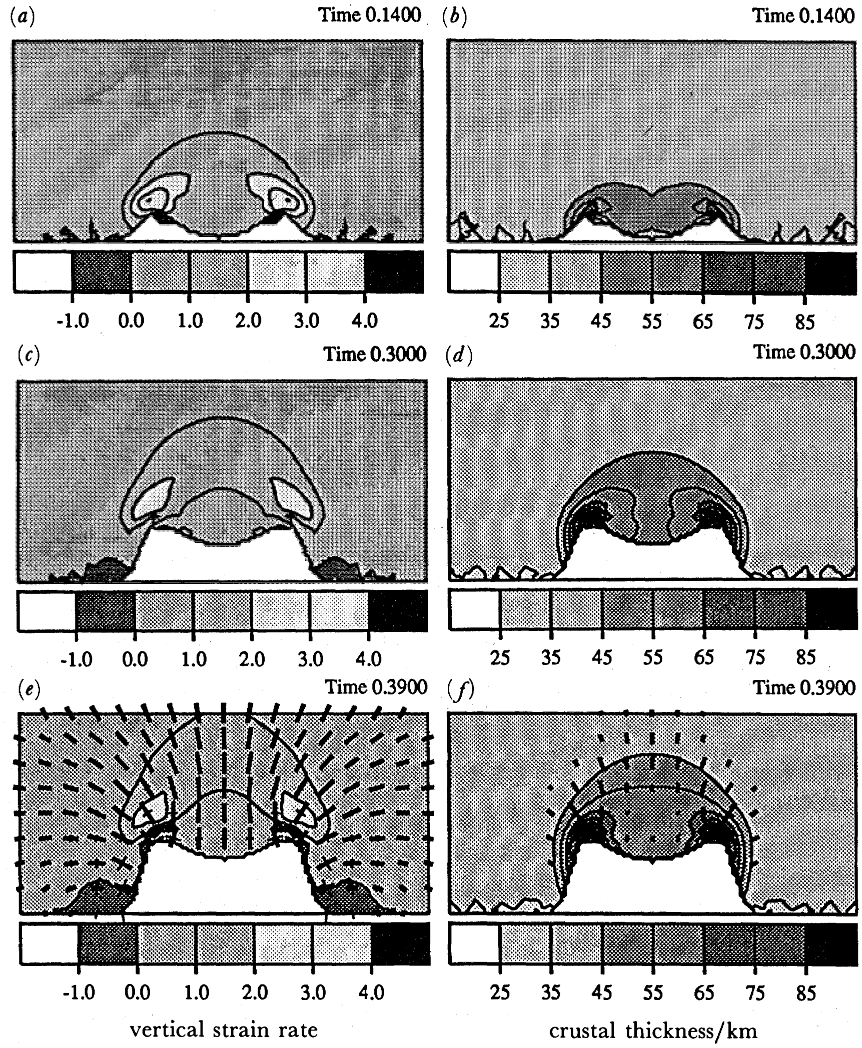

FIGURE 4. Contours of vertical strain rate and crustal thickness for a thin-sheet calculation with $n = 10$ and $Ar = 3$ (equations (1)–(5); see Houseman & England 1986 for details of calculation). Shaded regions correspond to the deforming lithosphere, the solid lines surrounding them correspond to the boundaries to the deforming material. One portion of the bottom boundary moves with time. Dimensionless times, shown above the parts of the figure, can be converted to times by multiplying by 100 Ma. As in figure 2, these are map views of the deformation. Figure 4*a*, *c*, *e* show contours of vertical strain rate. Figure 4*b*, *d*, *e* show contours of crustal thickness; before deformation starts the crustal thickness is 35 km throughout. Solid lines superimposed on (*e*) show the orientations and relative magnitudes of the calculated horizontal principal stresses; thicker lines correspond to compressional stress, the few thinner lines, around the corners of the indenter, to extensional stresses. The longest symbol corresponds to a compressional deviatoric stress of 1.0×10^8 Pa averaged vertically through the lithosphere. The solid lines on (*f*) show the calculated horizontal principal strain rates; these are parallel to the respective principal stresses, but depend on the *n*th power of the deviatoric stress (*e*). Note the concentration of compressional strain rates around the edges of the plateau. The maximum compressional strain rate is 1.4×10^{-15} s^{-1}.

(England & Houseman 1986). If this condition holds, then most of the convergent motion applied by the indenting boundary is accommodated by thickening of the lithosphere in front of that boundary; this is so for several different assumptions about the state of stress on the lateral boundaries to the lithosphere (England & McKenzie 1983; Vilotte *et al.* 1982; Houseman & England 1986; England & Houseman 1986; Vilotte *et al.* 1986).

Note that even though the outward displacement of the maximum thickening strain rates (figure 4*a*, *c*, *e*) shows that buoyancy forces are important in the deformation of the region, compressional stresses still dominate within the elevated region (figure 4*e*) and the width of the plateau is still very similar to what would be predicted by the simple arguments of § 2.4: namely about half the length of the covergent boundary condition. Compare this with a width of 1000–1500 km for the Tibetan Plateau, and a length of 2500 km for the Himalayan front. (Other comparisons between calculation and observation are given by England & Houseman 1986.) If the relative importance of the buoyancy forces is greater than in figure 4, then the width of the deforming region becomes larger, but the maximum thickening strain is correspondingly diminished for an equivalent amount of convergence, and crustal thickness contrasts as great as that between the Tibetan Plateau and its surroundings are not formed (Houseman & England 1986).

3.2. *Extension of a thickened plateau*

The results of thin-sheet calculations have been compared with observations in the India–Asia collision zone by England & Houseman (1986) and Vilotte *et al.* (1986). These calculations appear to account satisfactorily for the major features of the Tertiary deformation in Asia: the present size and shape of the plateau (above), the northward displacement of the southern margin of the region by at least 1000–2000 km (Zhu *et al.* 1977, 1981; Molnar & Chen 1978; Achache *et al.* 1984; Lin & Watts, this symposium), the distributed thrusting early in the deformation of Tibet, that is followed by strike-slip deformation (Chang *et al.* 1986) and the concentration of thickening strain around the edges of the plateau at the present day (Molnar & Deng 1984; Molnar, this symposium). The presence of low-lying and relatively aseismic areas within the deforming region, of which the most prominent is the Tarim Basin, can be accounted for by quite modest lateral variations in the strength of the lithosphere (Molnar & Tapponnier 1981; Vilotte *et al.* 1984; England & Houseman 1985). (The calculation illustrated in figure 4 is only one of a set described by Houseman & England (1986), and the reader is referred to that paper, and to England & Houseman (1986) for illustrations of deformation with different values of the parameters *n* and *Ar*.)

The most striking feature of the active strain in the Tibetan Plateau is the extension that occurs on roughly north–south normal faults (Molnar & Tapponnier 1978; Tapponnier *et al.* 1981; Armijo *et al.* 1982; 1986). The origin of this extension is explained in principle by the buoyancy forces associated with crustal thickness contrasts (Tapponnier & Molnar 1976), but this argument is a purely static one that does not take into account the compressional stresses associated with the convergence of India with Asia. Calculations that do take these stresses into account (one of which is illustrated in figure 4) do not exhibit extension in the region of thickened crust at any stage: although some stretching does occur parallel to the indenting boundary, it is always smaller in magnitude than the horizontal shortening, so that there is no net thinning of the lithosphere.

The extension in Tibet began comparatively recently, probably in the past 5 Ma (Armijo *et al.* 1982, 1986). The discrepancy between calculation and observation can be resolved if the India–Asia collision zone underwent, in this time interval, a change in the balance between the horizontal deviatoric stresses applied on the boundaries of Asia, and the buoyancy stresses in the interior of the continent. Three possible changes that could lead to extension within the plateau are:

(1) a change in the strength of the lithosphere;

(2) a change in the boundary conditions acting on Asia: either a reduction in the north–south

compressional stress applied by the convergence of India with Asia, or an increase in east–west extensional stresses owing to a condition applied on the eastern boundary of Asia.

(3) an increase in the potential energy of the Tibetan Plateau.

3.2.1. *Changes in strength of the lithosphere*

Before discussing mechanisms that can produce extension, it is worth mentioning another mechanism that has been proposed several times to the authors, which would not. The Argand number relates the strength of the continental lithosphere to the stresses arising from crustal thickness contrasts (equations (4) and (5)). It might seem at first sight that, if some mechanism diminishes the strength of the thickened lithosphere with time, any plateau that forms would eventually become sufficiently weak to flow away under its own weight. The most likely cause of a reduction in strength of the continental lithosphere during deformation is the heating of rocks by thermal relaxation of thickened lithosphere. Presumably, these changes occur first in the thickened crust near the indenting boundary and, it might be thought, would lead to exactly the imbalance between viscous stresses and buoyancy required for extension of the elevated region.

Such a change, if it affected the entire sheet, would indeed lead to extension; as the governing equations show, this is equivalent to a reduction in the convergent velocity (see equation (4)), a case that is considered below. However, we cannot consider local changes in viscosity in the same way as changes that affect the Argand number of the whole system: there is a balance between the compressional forces applied to the plateau by its surroundings (both the convergent boundary and the deforming foreland) and the work that is done within the plateau (both against gravity and to deform the material of the plateau). If the strength of the plateau is decreased while the boundary stresses are maintained, the rate of thickening in the plateau must increase.

Figure 5 illustrates this point. The topography and velocity boundary conditions are the same in each portion of the figure, and are those calculated for the set of parameters used in figure 4 (figure 5*b*, *d*, *f*), but the strain rate fields are calculated under three different assumptions. In figure 5*c*, *d* no modification to the original viscosity structure has occurred, and this calculation would carry on to yield the results shown in figure 4. In figure 5*a*, *b* the factor B in the Argand number (figure 1 and equation (4)) has been decreased by 20% for the region of the lithosphere in which the crustal thickness exceeds 55 km; this change would result in a tenfold increase in strain rate for a constant stress. The change is imposed arbitrarily and instantaneously to illustrate the influence of what would presumably be a gradual warming and weakening of the material of the thickened lithosphere. Comparison of figures 5*a* and *c* shows that the result of this change in rheology is to increase the compressional strain rates in the plateau, particularly at its edges, where the gradients of buoyancy forces are greatest, and that there is no appreciable change in the extensional strain rates parallel to the boundary of the plateau.

This result may be understood by considering a simple two-dimensional calculation, as might be appropriate close to the axis of symmetry in figure 5, where deviatoric stresses and strain rates in the east–west direction are small.

Figure 6 gives a simplified sketch of such a region, which is treated as being in two parts having lengths l_1 and l_2, gravitational potential energies γ_1 and γ_2 and vertically averaged viscosities η_1 and η_2, respectively. For simplicity we consider newtonian viscous fluids, but the

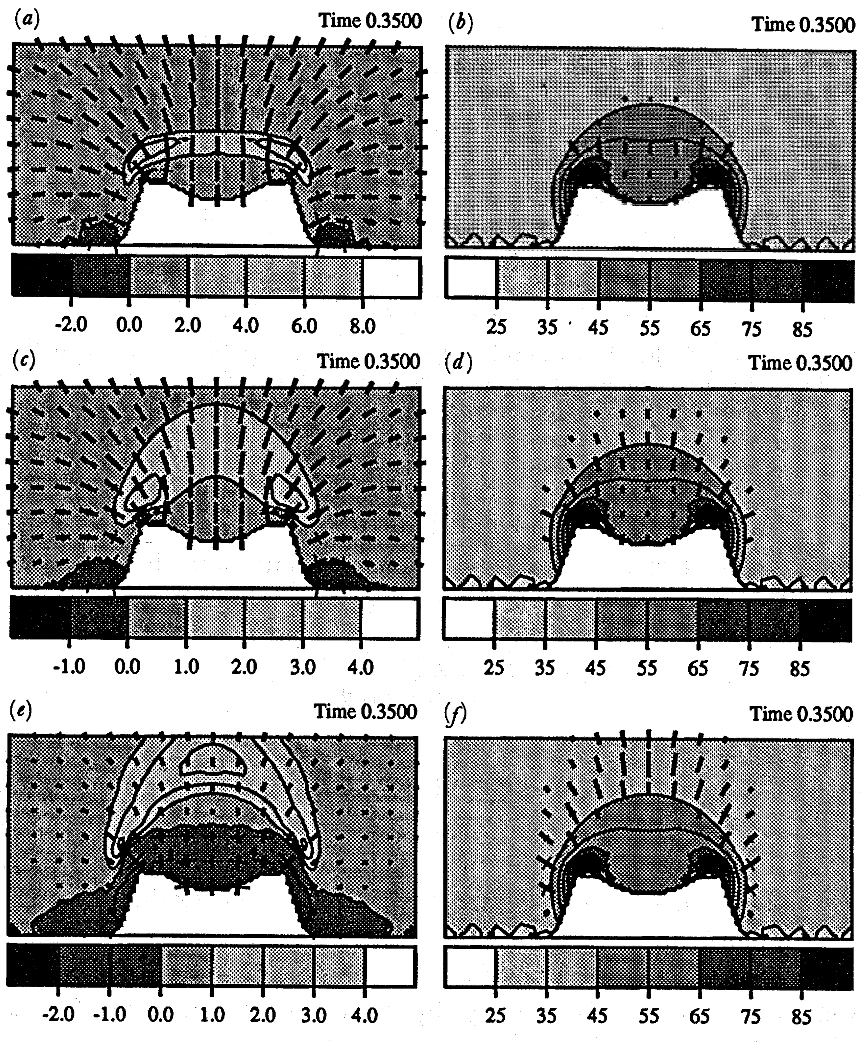

FIGURE 5. Contours of vertical strain rate and crustal thickness for calculations illustrating the influence of local changes in strength of the lithosphere. Figures 5*c*, *d* are taken from the calculation shown in figure 4, at a dimensionless time of 0.35 (35 Ma after the start of deformation). Figures 5*a*, *b* are calculated for the same distribution of topography, but with a decrease in strength in the region where the crustal thickness exceeds 55 km; figures 5*e*, *f* are also calculated with this distribution of topography, but the region in which the crustal thickness is less than 55 km has its strength reduced by a factor of three. The band of high compressional strain rates in (*a*) coincides with the edge of the region whose crustal thickness exceeds 55 km, and the equivalent region in (*f*) is seen as an area of lithospheric thinning in front of the convergent boundary. Symbols as in figure 4. Vectors indicating principal horizontal deviatoric stresses are superimposed on figures 5*a*, *c* and *e* (maximum values 8.6×10^{7}, 1.0×10^{8} and 6.3×10^{7} Pa); principal horizontal strain rates are shown on figures 5*b*, *d* and *e* (maximum values 1.8, 1.3 and $1.2\times10^{-15}\ s^{-1}$).

same arguments apply to a non-newtonian lithosphere. If deformation in the x direction is neglected, the y component of the force balance (equation (3)) reduces to

$$\frac{\partial(2\bar{\tau}_{yy}-\gamma)}{\partial y}=0 \qquad (7)$$

so that $$4\eta_1\dot{\epsilon}_1-\gamma_1=4\eta_2\dot{\epsilon}_2-\gamma_2=\text{constant},$$

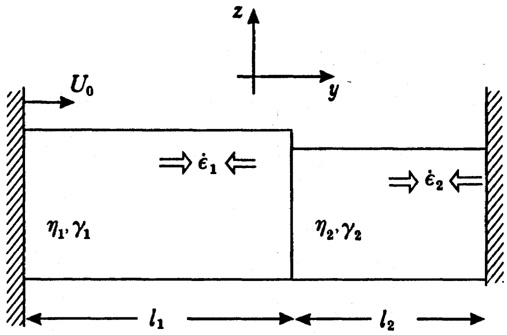

FIGURE 6. Sketch cross section of lithosphere consisting of two regions, one having viscosity η_1 and potential energy contrast γ_1 with a reference column of lithosphere, the other with viscosity η_2 and potential energy contrast γ_2. They are confined by rigid boundaries that move towards each other with velocity U_0.

where $\dot{\epsilon}_1$ and $\dot{\epsilon}_2$ are the horizontal strain rates in the two regions. The constant may be evaluated from the condition that the integral of the compressional strain across the two regions must be equal to the *y*-velocity, U_0, of the moving boundary, and these expressions may be re-arranged to yield:

$$\dot{\epsilon}_1 = -\frac{U_0 - l_2(\gamma_1 - \gamma_2)/4\eta_2}{l_1 + l_2\eta_1/\eta_2}. \tag{8}$$

As expected, when $\gamma_1 = \gamma_2$ and $\eta_1 = \eta_2$ the strain rate is constant across the region, and $\dot{\epsilon}_1 = \dot{\epsilon}_2 = -U_0/(l_1 + l_2)$; as $\eta_1 \rightarrow \infty$, $\dot{\epsilon}_1 \rightarrow 0$, and as $\eta_2 \rightarrow \infty$, $\dot{\epsilon}_1 \rightarrow -U_0/l_1$. The important result displayed in equation (8) is that if, as envisaged above, some initial combination of viscosities and buoyancy forces that support a plateau in compression is perturbed by decreasing the viscosity of the plateau (η_1 in this example), the compressional strain rate in the plateau is *increased*, as seen in figure 5.

Equation 8 shows that extensional strain rates within the plateau can be induced, under conditions of constant convergence velocity, by decreasing the strength of its surroundings, η_2; this is illustrated in figure 5*e*, *f*, where *B* has been reduced by a factor of three in the portion of the lithosphere where the crustal thickness is less than 55 km. *Ad hoc* ways might be imagined of decreasing rapidly the strength of the Asian lithosphere relative to that of the Tibetan Plateau, but they are not pursued here.

3.2.2. *Changes in boundary condition*

As can be seen from the discussion of the governing equations and the Argand number in §2.3, and from the simple case described by equation (8), changes in the convergent velocity would influence the balance between compressional stresses and buoyancy forces within the plateau. Complete cessation of the convergence of India with Asia would remove the compressional stress from the plateau, and presumably lead to widespread extension. The definition of the Argand number (equation (4)) shows that dividing the convergent velocity U_0 by a given amount is equivalent to increasing the Argand number by the $(1/n)$th power of that amount. England & Houseman (1988) show that a reduction in compressive stress by a factor of three is required before a change in boundary condition in the thin-sheet calculations leads to extension within the plateau. For a lithosphere with a power-law exponent $n = 3$, this would correspond to a decrease in India's velocity by a factor of around 30 (equation (4)) before the extension began in the Pliocene. Such a decrease can be ruled out from present-day observations of plate motion (see, for example, Minster & Jordan 1978) and from the moment release of large earthquakes in Asia (Molnar & Deng 1984), both of which suggest that India's rate of convergence with Asia is still about 50 mm a^{-1}.

The arguments of §2.4 suggest that extensional stresses on eastern or western boundaries to the deforming region (such as appear to exist in Eastern China) should die out exponentially with distance from the boundary and would, therefore need to exceed considerably the compressional stresses applied by India to the southern boundary of Asia if they were to be responsible for extension within the Tibetan Plateau. Regional stresses are difficult to determine; the extensional strain rates in Eastern China determined by Molnar & Deng (1984) are similar to those determined for the Tibetan Plateau. This, together with the observation that extensional faulting, although covering most of the plateau, is confined within the 4 km contour (Chen & Molnar 1983) suggests that conditions on the eastern boundary of Asia do not provide the explanation for extension within the plateau.

3.2.3. *Increase in potential energy of the plateau*

The removal of compressional stresses by cessation of convergence, or the weakening of the lithosphere surrounding an elevated region, are acceptable mechanisms for producing extension in elevated regions, and may have been responsible for the onset of extension in places such as the Basin and Range or the Andes. However, the discussion of the previous sections suggest that these mechanisms probably did not operate in the case of the Tibetan Plateau.

Equation (8) shows that there is a third perturbation to the force balance that could produce extension in the elevated region: increasing the potential energy of the plateau (γ_1, equation (8)) may cause extension, even under conditions of constant convergence rate.

Conductive relaxation of the lithosphere can lead to increases in its potential energy, but this takes place over tens of millions of years. An alternative mechanism is suggested by the numerical experiments of Houseman *et al.* (1981) who investigate the stability of the continental lithosphere after thickening. Parsons & McKenzie (1978) point out that the plates should be thought of as consisting of two distinct layers: each of them transfers heat predominantly by conduction but the upper layer, consisting of the crust and uppermost mantle, is mechanically strong whereas the lower layer is fluid, and represents the upper thermal boundary layer to the convecting mantle. Houseman *et al.* (1981) showed that thickening such a rheologically layered lithosphere resulted in gravitational instability of the lower of these layers and its rapid descent into the convecting mantle. Replacement of the thermal boundary layer with hot asthenosphere results in increases to the surface elevation and gravitational potential energy of the mechanical boundary layer. Figure 7 illustrates the influence of this process on the mechanical evolution of thickened lithosphere; a fuller discussion is given by England & Houseman (1988).

For discussion of the mechanics of the Tibetan Plateau, the important result of the experiments of Houseman *et al.* is that the descent of the thickened thermal boundary layer into the convecting mantle follows some time after the thickening. The duration of the delay depends strongly upon the Rayleigh number of the convecting mantle and upon the degree of thickening of the thermal boundary layer, and may be from several million years to several tens of millions of years. If this mechanism operates the conditions exist, in a thickened orogenic belt, for relatively rapid increase in potential energy of the belt at some time after the belt

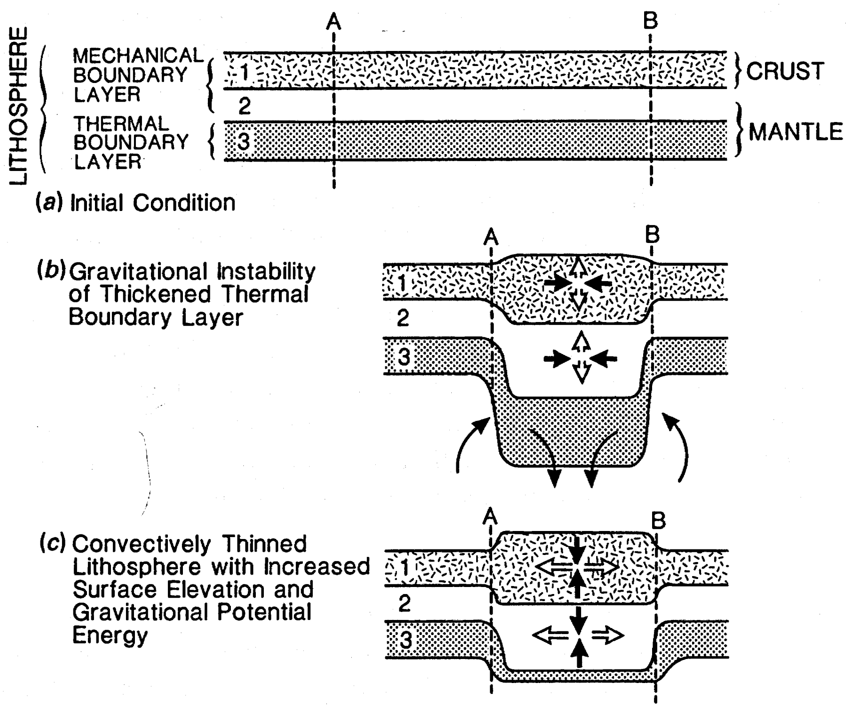

FIGURE 7. Sketch of the mechanical evolution of thickened continental lithosphere, following Houseman *et al.* (1981). (*a*) In its undisturbed state the continental lithosphere consists of (1) crust, (2) upper mantle that is sufficiently cold to be much stronger than the convecting mantle and (3) mantle that is hot and weak enough to take part in mantle convection; this layer forms the upper thermal boundary layer to the convecting upper mantle. (*b*) A portion of the continental lithosphere is thickened; horizontal thermal gradients are generated and amplified within the thermal boundary layer and, after a time that depends on the viscosity of the convecting mantle and the strain in the boundary layer, the thermal boundary layer drops off, to be replaced by hot material from below. This process increases the surface height and potential energy of the lithosphere (*c*). As sketched here, the process occurs in a uniform fashion across the deformed region; there is no reason to believe that this would be the case in reality.

forms; if the increased potential energy exceeds that supportable by the boundary conditions, extension of the orogen will occur.

Treating the full three-dimensional problem of mantle convection coupled to a deforming lithosphere is prohibitively expensive of computer time, and a simplified approach to the convective removal of the thermal boundary layer is required. England & Houseman (1988) investigate the influence of this process by treating the removal of the thermal boundary layer as an instantaneous event that occurs at a given value of the thickening strain experienced by the continental lithosphere. They also show that, for a lithosphere that does not lose its thermal boundary layer, increases in potential energy due to crustal thickening are largely offset by decreases due to thickening of dense lithospheric mantle, so that overall changes in potential energy under these conditions are small compared with those that take place once the thermal boundary layer leaves the thickened lithosphere.

England & Houseman (1988) consider a range of conditions for the relative magnitudes of buoyancy forces before and after loss of the thermal boundary layer, but we illustrate their results with a single calculation (figure 8) that neglects buoyancy forces until the loss of the thermal boundary layer. In figure 4, the Argand number (equation (4)) is kept at 3 throughout; the conditions of figure 8 are those of figure 4, except that the Argand number for

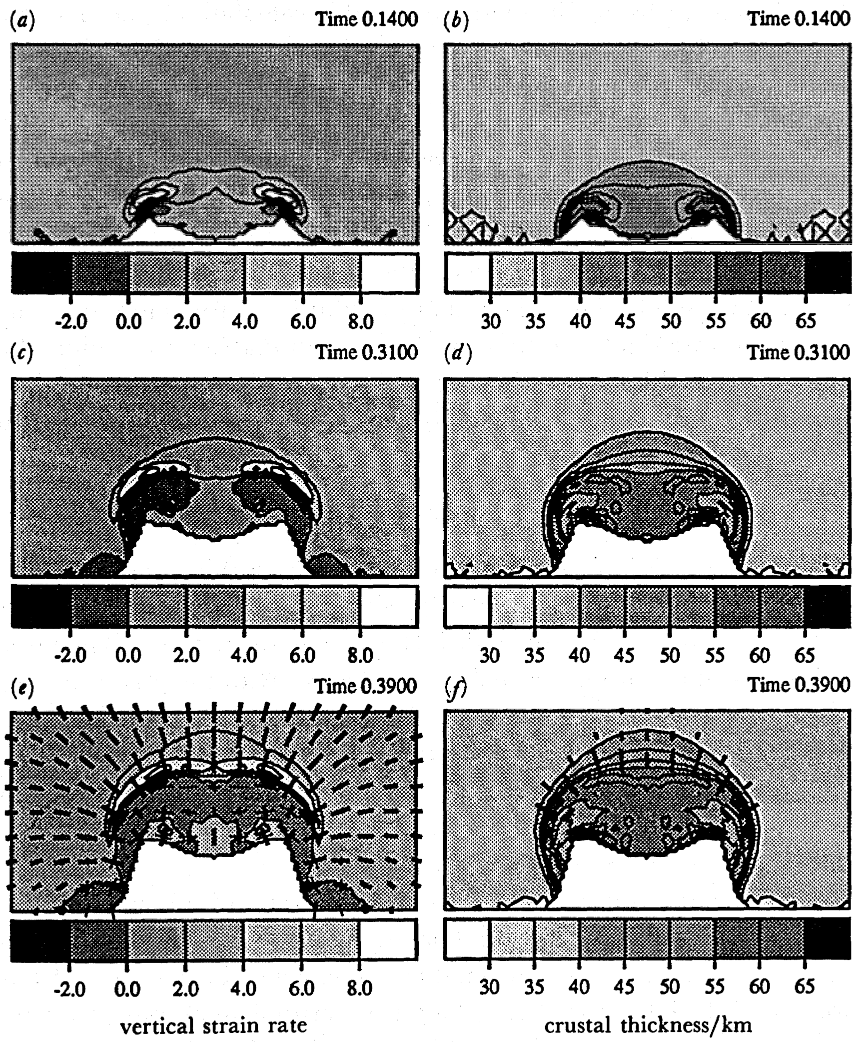

FIGURE 8. As figure 4, except that the potential energy of any piece of lithosphere increases abruptly if its thickening strain exceeds 70% (see text). Note the predominance of extensional horizontal strain rates in the regions where this has occurred (crustal thickness greater than 60 km), and that the principal extensional strain rate is perpendicular to the convergent velocity in front of the indenter. The strength of the lithosphere is assumed here to be unaffected by the convective removal of the boundary layer. Maximum deviatoric stress in (*e*) is 3.0×10^7 Pa and maximum principal strain rate in (*f*) is $1.9 \times 10^{-15}\ s^{-1}$.

a particular piece of lithosphere is kept at zero until that region has experienced 70% thickening strain, when it is set to 10. The change in Argand number corresponds to a change in the potential energy of the lithosphere, without any change in its strength; it is analagous to changing the quantity $(\gamma_1 - \gamma_2)$ in equation (8), leaving the other parameters unchanged.

The differences between the two cases may be clearly seen in the contours of strain rate (figures 4*a*, *c*, *e* and 8*a*, *c*, *e*) and crustal thickness (figures 4*b*, *d*, *f* and 8*b*, *d*, *f*): once crustal thickness exceeds 60 km in figure 8, the buoyancy forces dominate the deformation, and there is no further thickening of the lithosphere, indeed there is net thinning of the lithosphere throughout the region where this condition is achieved. Figure 8*e* shows that this region can be as extensive, relative to the indenting boundary condition, as the Tibetan Plateau is in relation to the Himalayan front. Furthermore, in this calculation, one horizontal principal stress is extensional and is perpendicular to the direction of motion of the indenting boundary; the focal mechanisms of earthquakes (Molnar & Chen 1983) indicate that this is also the case within the Tibetan Plateau.

There are four parameters to the problem discussed in this section: the value of the power-law exponent, n, the values of the Argand number before and after the thermal boundary layer leaves the lithosphere, and the value of the compressional strain at which this transition is assumed to occur. There is not space to discuss the full range of solutions here; this is done by England & Houseman (1988). They show that distributions of topography, crustal thickness and present day strain rate similar to those of the Tibetan Plateau are exhibited by a range of calculations, provided that the vertically averaged deviatoric stresses within the viscous sheet are a few times 10^7 Pa and its power-law exponent, n, is greater than or equal to 3.

4. Conclusions

Calculations that treat the Asian continental lithosphere as a thin continuous sheet appear to be able to account adequately for the distribution of Tertiary strain within Asia (see, for example, Vilotte *et al.* 1984, 1986; England & Houseman 1985, 1986). The horizontal extent of the region of appreciable crustal thickening is consistent with the predicted length scales for the deformation of such a sheet (England *et al.* 1985, and §2.4 of the present paper). The elevation of this plateau, and the concentration of present-day compressional strain rates around its margins (Molnar & Deng 1984) are consistent with calculations that treat the lithosphere as a sheet whose strength is of order 100–200 MPa at geological strain rates (England & Houseman 1986; Vilotte *et al.* 1986).

However, those calculations do not exhibit the thinning that is observed as the most recent phase of deformation on Tibetan Plateau. The arguments of §3 suggest that syn-convergent thinning of thickened crust is not the inevitable consequence of that crustal thickening: the extensional horizontal deviatoric stresses arising from crustal thickening are balanced by the compressional stresses that generate the thick crust; unless there is a change in this balance, extension will not occur in the thickened crust. One simple way in which the balance may change is by the removal of the convergent boundary stresses; this may be the explanation for the extension in the Basin and Range province of western North America. We argue in §3 that this does not seem acceptable for the present extension in the Tibetan Plateau, as convergence between India and Asia is still active, and we suggest that the balance there has been perturbed by the loss of the lower portion of the continental lithosphere, which forms the upper thermal boundary layer to the convecting mantle (Parsons & McKenzie 1978; Houseman *et al.* 1981).

The loss of the thermal boundary layer at the base of the lithosphere would have the effect of raising the surface height of the region of thickened crust by between 1 and 3 km (depending upon the density distribution within the lithosphere, and the thickness of material lost); the consequent change in the potential energy per unit surface area of lithosphere would be equivalent to an extensional driving force of several times 10^{12} N per metre length of the elevated region (England & Houseman 1988). The calculation illustrated in figure 8 shows that such a change in the potential energy of the thickened lithosphere is sufficient to effect a transition from horizontal shortening to horizontal extension within the region where the change occurs.

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References

Achache, J., Courtillot, V. & Zhou, Y. X. 1984 *J. geophys. Res.* **89**, 10311–10339.
Armijo, R., Tapponnier, P., Mercier, J. L. & Han, T. 1982 *Eos, Wash.* **63**, 1093.
Armijo, R., Tapponnier, P., Mercier, J. L. & Han, T. 1986 *J. geophys. Res.* **91**, 13803–13872.
Artyushkov, E. V. 1973 *J. geophys. Res.* **78**, 7675–7708.
Bird, P. & Baumgardner, J. 1984 *J. geophys. Res.* **89**, 1932–1944.
Bird, P. & Piper, K. 1980 *Phys. Earth planet. Inter.* **21**, 158–175.
Brace, W. F. & Kohlstedt, D. L. 1980 *J. geophys. Res.* **85**, 6248–6252.
Chang, C. *et al.* 1986 *Nature, Lond.* **323**, 501–507.
Chen, W.-P. & Molnar, P. 1983 *J. geophys. Res.* **88**, 4183.
Dalmayrac, B. & Molnar, P. 1981 *Earth planet. Sci. Lett.* **55**, 473.
Ekström, G. & England, P. 1988 *J. geophys. Res.* (Submitted.)
England, P. C. & Houseman, G. A. 1985 *Nature, Lond.* **315**, 297–301.
England, P. C. & Houseman, G. A. 1986 *J. geophys. Res.* **91**, 3664–3676.
England, P. C., Houseman, G. A. & Sonder, L. S. 1985 *J. geophys. Res.* **90**, 3551–3557.
England, P. C. & Houseman, G. A. 1988 *J. geophys. Res.* (In the press.)
England, P. C. & McKenzie, D. P. 1982 *Geophys. Jl R. astr. Soc.* **70**, 292–321.
England, P. C. & McKenzie, D. P. 1983 *Geophys. Jl R. astr. Soc.* **73**, 523–532.
Evison, F. F. 1960 *Geophys. Jl R. astr. Soc.* **3**, 155–190.
Fleitout, L. & Froidevaux, C. 1982 *Tectonics* **1**, 21–56.
Frank, F. C. 1972 In *Flow and fracture of rocks* (Geophys. Monogr. no. 16) (ed. H. C. Heard, I. Y. Burg, N. L. Carter & C. B. Rayleigh), pp. 285–292. Washington, D.C.: American Geophysical Union.
Houseman, G. A. & England, P. C. 1986 *J. Geophys. Res.* **91**, 3651–3663.
Houseman, G. A., McKenzie, D. P. & Molnar, P. 1981 *J. geophys. Res.* **86**, 6115–6132.
Jackson, J. & McKenzie, D. 1988 *Geophys. Jl R. astr. Soc.* (In the press.)
McKenzie, D. P. 1972 *Geophys. Jl R. astr. Soc.* **30**, 109–185.
Minster, B. & Jordan, T. 1978 *J. geophys Res.* **83**, 5331–5354.
Molnar, P. & Chen, W. P. 1978 *Nature, Lond.* **273**, 218–220.
Molnar, P. & Chen, W. P. 1983 *J. geophys. Res.* **88**, 1180–1196.
Molnar, P. & Deng, Q. 1984 *J. geophys. Res.* **89**, 6203–6228.
Molnar, P. & Lyon-Caen, H. 1988 *Geol. Soc. Am. Spec. Publ.* (In the press.)
Molnar, P. & Tapponnier, P. 1978 *J. geophys. Res.* **83**, 5361–5375.
Molnar, P. & Tapponnier, P. 1981 *Earth planet. Sci. Lett.* **52**, 107–114.
Molnar, P., W.-P. Chen, & Tapponnier, P. 1981 In *Geological and ecological studies of Qinghai-Xizang Plateau* vol. 1, pp. 757–762.
Molnar, P. & Tapponnier, P. 1975 *Science, Wash.* **189**, 419–428.
Molnar, P. & Tapponnier, P. 1978 *J. geophys. Res.* **83**, 5361–5375.
Parsons, B. & McKenzie, D. 1978 *J. geophys Res.* **83**, 4485–4496.
Sonder, L. J. & England, P. C. 1986 *Earth planet. Sci. Lett.* **77**, 81–90.
Tapponnier, P., & Molnar, P. 1976 *Nature, Lond.* **264**, 319–324.
Tapponnier, P. *et al.* 1981 *Nature, Lond.* **294**, 405–410.
Tapponnier, P., Pelzer, G., Le Dain, A. Y., Armijo, R. & Cobbold, P. 1982 *Geology* **10**, 611–616.
Vilotte, J. P., Daignieres, M. & Madaraiga, R. 1982 *J. geophys. Res.* **87**, 10709–10728.
Vilotte, J. P., Daignieres, M., Madaraiga, R. & Zienckiewicz, O. C. 1984 *Phys. Earth planet. Inter.* **36**, 236–259.
Vilotte, J. P., Madariaga, R., Daignieres, M. & Zienckiewicz, O. 1986 *Geophys. Jl R. astr. Soc.* **84**, 279–310.
Zhu, X., Liu, C., Ye, S. & Liu, J. 1977 *Sci. Geol. Sin.* **1**, 44–51.
Zhu, Z., Zhu, X. & Zhang, Y. 1981 In *Geological and ecological studies of Qinghai-Xizang Plateau* vol. 1, pp. 931–939. Beijing: Science Press.

Discussion

M. F. Osmaston (*The White Cottage, Woking, Surrey, U.K.*). I am concerned about the force balance across the High Himalaya that Dr England's kind of modelling would seem to require to achieve and maintain (if that is the case) its extreme elevation. From the north, how can the spreading of Tibet, which is significantly lower, provide enough force? From the south, oceanridge push, as customarily calculated at present, is quite inadequate. Work that I am doing on the shape-evolution histories of subduction interfaces makes me increasingly doubtful that slab pull is the major force it has been made out to be. I therefore suggest that the High Himalayan elevation may be being maintained thermally, by crustal reheating, at a rate that sufficiently balances its rheological tendency to spread. The migration of upthrust activity to areas of lower elevation seems consistent with this.

P. C. England. There is no appreciable difference in *average* elevation between the High Himalaya and the Tibetan Plateau (see, for example, Bird 1978).

Reference

Bird, P. 1978 *J. geophys. Res.* **83**, 4975–4987.

S. Ghosh (*Grant Institute of Geology, University of Edinburgh, U.K.*). It appears that although we know with a fair amount of confidence much about the early history of the tectonic evolution of the Himalayas and Tibet, the present-day movement scenario around the Himalayas including its crustal–subcrustal parts is still not very clear. There is hardly any doubt that the Himalayas is still an active mountain chain. Would it not be worthwhile for us now to explore the possibilities of identifying thrust areas for Himalayan research so that the important gaps in our knowledge about the geology and tectonics of the Himalayas are filled up within a comparatively short time?

P. C. England. It would perhaps be more appropriate if this reply were left to Dr Molnar, who succinctly summarizes elsewhere in this volume the major problem areas, from the Himalaya to the Qilian Shan (with many references), but this comment raises an important point. It should be emphasized that the mechanism that is proposed in this paper for the evolution of the Tibetan Plateau, although consistent with our understanding of the behaviour of very viscous fluids, and with the observations of bathymetry and heat flow in the oceans is not established as occurring within the Earth. Observations of the history of surface elevation and extensional faulting and of present-day rates of movement in the Himalaya and the Tibetan Plateau will yield important insights into the dynamics of the lithosphere and mantle in zones of continental convergence. The technology will exist within the next few years for the precise determination of strain rates on these scales, and the successful pursuit of this very important data set will depend strongly both upon international cooperation, and upon cooperation between Earth scientists using a wide range of tools, from field observations to space geodetic techniques.

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