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

# Inferences of Deviatoric Stress in Actively Deforming Belts from Simple Physical Models

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*Phil. Trans. R. Soc. Lond. A* 1991 **337**, 151-164
doi: 10.1098/rsta.1991.0113

# Inferences of deviatoric stress in actively deforming belts from simple physical models

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Observations of temperatures and heat flux near major thrust zones indicate that in their deep levels shear stresses may exceed 50–100 MPa. Within strike–slip zones shear stresses in the lower lithosphere may also approach 50–100 MPa, though shear stresses in the upper crust of those regions are probably much lower. The relationship of tectonic style to surface elevation in the Andes and Tibet yields an estimate of about $5 \times 10^{12}$ N m$^{-1}$ for the force per unit length required to deform the lithosphere of these regions. This force per unit length is equivalent to an average shear stress of about 25 MPa through a lithosphere 100 km thick. The width-to-length ratios of active belts are consistent with deformation determined by the creep of the lower lithosphere rather than by friction on faults. The patterns of rotation of crustal blocks in western North America suggest that these blocks passively follow the deformation of a continuous substrate. The observations of deformation and the estimates of stress derived from them, both suggest that the upper continental crust is weak, relative to the lower parts of the lithosphere the deformation of which it follows passively. If this is so, determinations of stress in the upper crust may have only limited relevance to the deformation of the lithosphere as a whole.

## 1. Introduction

In contrast to the oceanic portions of the Earth's surface, which move as essentially rigid bodies, the continents exhibit distributed deformation over horizontal scales of hundreds to thousands of kilometres (Molnar & Tapponnier 1975; Molnar 1988*a*; England & Jackson 1989). A key problem in continental dynamics, therefore, is to relate the velocity gradients inferred from geodetic, seismological, palaeomagnetic, and structural data to the stresses acting within the lithosphere. The term 'lithosphere', as used here, refers to the crust and that part of the upper mantle through which heat is transferred primarily by conduction.

Orientations of principal stresses, and sometimes their magnitude can, in principle, be constrained by direct observation in boreholes in the uppermost crust (Brereton & Müller 1991; Fuchs *et al.* 1991; Zoback & Magee 1991; Zoback 1991). Such observations are usually expensive to make and, with a very few exceptions, are limited to depths of a few kilometres or less. The observation that strains and velocity gradients in the crust are organized on a horizontal lengthscale that far exceeds the thickness of the crust has led several authors to suggest that the upper crust passively follows the deformation of the lower lithosphere (Ekström & England 1989; England & Jackson 1989; England & McKenzie 1982; McKenzie & Jackson

*Phil. Trans. R. Soc. Lond.* A (1991) **337**, 151–164
*Printed in Great Britain*

1983; Jackson & McKenzie 1988). If this suggestion is correct, then the stress in the upper crust is not directly related to the state of stress in the deeper lithosphere (Houseman & England 1986, Appendix A).

The moment tensors of earthquakes give indications of strain (Kostrov 1974) through ranges of depths that are inaccessible to routine boring. These measurements of strain may, with much care, be used to supplement determinations of stress. But, because most of our information about active strain in the continents has come from the measurement of strain using the moment tensors of earthquakes (Ekström & England 1989; Jackson & McKenzie 1984, 1988; McKenzie 1978; Molnar & Deng 1984; Molnar & Lyon-Caen 1989; Molnar & Tapponnier 1975), any information on stress that might come from earthquake studies could not be used in a dynamical study without circularity.

Laboratory studies of the strain of rocks have been a major source of information on deformation processes in the inaccessible parts of the crust and mantle. The problems of extrapolating the results of these studies to geological conditions are too well known to need rehearsing here. Even in the absence of uncertainty in material properties determined in the laboratory, uncertainty in temperature and lithology at any given level within the lithosphere is sufficiently large to make a prediction of deviatoric stress from laboratory flow laws uncertain by an order of magnitude at least (see, for example, Molnar 1991).

The purpose of this paper is to review estimates of the magnitude of deviatoric stress in the deeper portions of the lithosphere that do not depend upon downwards extrapolation of observations made near the surface, nor upon using laboratory determinations of rheology. In the absence of a direct observation of stress in the deep lithosphere, these estimates must come from comparing the results of theory with observations of other quantities. The observations involved are heat flux and temperature near major shear zones, and the distributions of strain rate and topography in zones of active continental deformation. The theories, befitting our nascent understanding of the problem, are simple; they contain, in the first case, a single governing parameter and, in the second case, two.

## 2. Estimates of stresses from temperatures in major fault zones

Most of the energy dissipated by friction on faults is likely to be released as heat (see discussions by Lachenbruch & Sass (1980)). In aseismic shear zones, the rate of release of heat by dissipation is (neglecting the expenditure of energy in creating new surfaces) equal to the deviatoric stress multiplied by the strain rate. Observations of the thermal state of large faults and shear zones offer, therefore, the possibility of determining the magnitude of deviatoric stress within them.

### (a) *Thrust faults*

Molnar & England (1990) gave simple formulae that permit estimates to be made of the deviatoric stress on thrust faults, either from surface heat flux measurements made above active thrust faults, or from temperatures inferred from exhumed thrust faults. The surface heat flux, $Q$, above a thrust fault that is slipping at a speed $V$ is related to the shear stress, $\tau$, on the fault and the heat flux, $Q_0$, that would be observed in steady state, without slip on the fault, by:

$$Q = (Q_0 + \tau V)/S \tag{1}$$

Table 1. *Estimates of shear stress in active subduction zones*

($z_f$ is the depth of the subduction zone below the site of heat flux measurement. $V$ is the rate of subduction and $\tau$ is the shear stress calculated from equation (1) on the assumption that the shear stress is independent of depth; calculations assuming that shear stress increases proportionately to depth yield stresses approximately 30% higher (Kao & Chen 1991). The range in shear stress reflects uncertainties in the quantities entering equation (1).)

| | $z_f$ km | $V/(\text{mm a}^{-1})$ | $\tau$/MPa |
|---|---|---|---|
| Japan[a] | $70\pm10$ | $105\pm10$ | 65–100 |
| | $60\pm10$ | $105\pm10$ | 55–90 |
| S. Ryukyu–Kyushu[b] | $25\pm3$ | $52\pm13$ | 35–50 |
| N. Ryukyu–Kyushu[b] | $10\pm3$ | $34\pm12$ | 25–50 |
| | $19\pm5$ | $34\pm12$ | 35–75 |
| | $30\pm8$ | $34\pm12$ | 80–115 |
| Peru[a] | $30\pm5$ | $80\pm5$ | 30–80 |
| Himalaya[a] | 15–18 | $15\pm5$ | 70–140 |

[a] Molnar & England (1990).
[b] Kao & Chen (1991).

(Molnar & England 1990, eq. (28)) where

$$S \approx 1+\surd(Vz_f\sin\delta/\kappa) \tag{2}$$

and $\delta$ is the dip of the fault, $z_f$ is the depth of the fault at the point of interest and $\kappa$ is the thermal diffusivity of the rocks.

Molnar & England (1990) and Kao & Chen (1991) applied this analysis to estimate shear stresses in active subduction zones. The estimates lie in the range 25–140 MPa, and generally increase with the depth of the fault (table 1). Molnar & England (1990) used related arguments to estimate the shear stress on the active thrust fault that bounds the south of the Himalayas and obtained an estimate of 70–140 MPa for the shear stress at 15–18 km on that fault.

It appears from these estimates that shear stress in the upper portions of major active thrust faults is several tens to *ca.* 100 MPa. These stresses, resolved horizontally, contribute to the support of mountain ranges (Molnar & Lyon-Caen 1988). The range of shear stresses listed in table 1 for subduction zones implies a compressional force per unit length of convergent boundary of about $2.5\times10^{12}\ \text{N m}^{-1}$. The equivalent quantity calculated for the seismic part of the active intracontinental thrust fault beneath the Himalayas is 0.5 to $1\times10^{12}\ \text{N m}^{-1}$.

Estimates of driving forces based on shear stresses determined from the seismically active portions of faults are lower limits to the driving forces associated with major thrust faults, for they contain no contribution from stress on the deeper, aseismic, portions of those shear zones. Several authors have attempted to estimate shear stresses from temperatures inferred from metamorphic mineral assemblages generated during the slip of large thrust faults. In the studies summarized below, with the exception of that of Barton & England (1979), these mineral assemblages formed at temperatures above 450 °C and therefore above the assumed maximum temperature at which moderately sized-to-large earthquakes occur within continental crust (Chen & Molnar 1983). Analysis of such exhumed thrust faults can, therefore, provide information complementary to that derived from active fault zones.





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The observation of an inverted metamorphic zonation associated with major thrust faults led to the suggestion that dissipative heating in the fault zones might be an important source of heat from metamorphism (Le Fort 1975). Early estimates of shear stress, from geological observations made on such faults, were around 100 MPa (Barton & England 1979; Graham & England 1976; Scholz 1980). These estimates were all based upon the steepness of the preserved metamorphic gradient near the fault, which was assumed to give an indication of the thermal gradient, and hence the rate of dissipative heating, near the fault (Graham & England 1976). The recognition that many zones showing inverted metamorphic zonation have been tectonically thinned after the metamorphism was imprinted (see, for example, England *et al.* 1991) removes the quantitative basis for those estimates. It is none the less clear from qualitative considerations that dissipative heating plays an important role in the thermal evolution of at least some thrust fault zones. In the absence of dissipative heating, the temperature at a particular point on the hanging wall of a thrust fault must drop once slip on the fault begins, because for most geologically relevant rates of slip, the temperature on a thrust fault that slips without dissipating appreciable heat is no higher than the average of the pre-slip temperatures of the rocks juxtaposed by slip on the fault. Because, in the cases discussed below, slip on the faults brought rocks originally at or close to the Earth's surface to the point of interest on the fault, and because the syn-slip temperatures recorded in the metamorphic assemblages approach or exceed the likely pre-slip temperatures in the hanging walls, some heat source – presumably dissipative heating – must have been operating at the faults.

Re-examination of the earlier estimates of shear stress from the Pelona Schist (Graham & England 1976) and from the Olympos Thrust (Barton & England 1979) using Molnar & England's (1990) approach, and taking account of post-metamorphic thinning of the inverted metamorphic sequence, yields estimates of shear stress that are somewhat lower than those obtained in the earlier studies, but still in the range 100–200 MPa (England & Molnar 1990).

From an analysis of the temperatures and pressures recorded by mineral assemblages generated during slip on the Main Central Thrust in the Annapurna–Manaslu region of central Nepal, England *et al.* (1991) concluded that shear stresses in the range 50–100 MPa were required to generate the inferred syntectonic temperatures on the Main Central Thrust. A similar magnitude of shear stress is obtained from Hubbard's (1989) inverted metamorphic sequence beneath the Main Central Thrust in the Mt Everest region (England & Molnar 1990).

If we combine the estimates of shear stress from the deeper, now inactive, Main Central Thrust and the estimate, made above on the basis of present day seismicity, of shear stress on the active thrust fault bounding the Himalayas, we may conclude that shear stress as high as 100 MPa acted to a depth of 35 km or more on the Main Central Thrust whilst it was active and, by extension, that a similar stress distribution is associated with the present thrust fault beneath the Himalayas.

Furthermore it seems reasonable to conclude, from estimates of shear stress at other major thrust faults, that the Main Central Thrust is not unusual in having supported shear stress of about 100 MPa during its slip. Shear stresses of this magnitude, resolved horizontally, would provide a driving force per unit length of about $4 \times 10^{12}$ N m$^{-1}$, if the shear stress were independent of depth, or about $2 \times 10^{12}$ N m$^{-1}$ if it rose linearly from zero at the Earth's surface to about 100 MPa at a depth of 40 km.

### (*b*) *Strike–slip zones*

The absence of a measurable heat flux anomaly associated with the San Andreas fault has been taken as evidence for generally low shear stresses on the fault (Brune *et al*. 1969; Lachenbruch & Sass 1980). In contrast, Scholz *et al*. (1979) concluded, from the distribution of metamorphism associated with the Alpine fault of the South Island of New Zealand – another major transcurrent zone – that this fault must have experienced shear stresses in excess of 100 MPa during a substantial portion of its slip.

Lachenbruch & Sass (1980) summarized over 90 heat flux determinations near the San Andreas fault and showed that there is no measurable heat flux anomaly associated with the fault itself and that, therefore, the shear stress on the fault is probably less than 10 MPa. On the other hand heat flux measurements in central California show an anomaly approximately 80 km wide, which covers the southern Coastal Ranges. Figure 1 shows the heat flux observations of Lachenbruch & Sass's table 1, for the latitude range 35–40° N, where the deformation is predominantly transcurrent.

As Lachenbruch & Sass (1980) point out, the heat flux anomaly is far too broad to be explicable by dissipative heating on the seismogenic portion (upper 10–15 km) of the San Andreas fault. Nevertheless, the relatively sharp decrease in heat flux at the edges of the anomaly argues for a shallow source. This line of reasoning led Lachenbruch & Sass (1980) to consider dissipative heating within a broad shear zone as one plausible cause for the heat flux anomaly.

Lachenbruch & Sass's mechanical analysis of the heat flux anomaly postulated distributed shear on horizontal planes within a weak zone which decoupled, to a greater or lesser degree, the upper crust in the neighbourhood of the fault from the surrounding plates. An alternative view, taken by Molnar (1991) and in the next section, is to take upper crust as being *well* coupled to a relatively stronger viscous lower lithosphere, whose deformation it follows. According to that view, the deformation accommodating the relative motion of the Pacific and North America plates is expected to be spread out over some tens or hundreds of kilometres, in a fashion dictated by the rheology of the lithosphere as a whole (see discussion surrounding equation (4) below), and dissipative heating near the San Andreas fault is the result of shear on vertical planes in the strong lower lithosphere.

Let us take the intensity of the heat flux anomaly in central California to be 30 mW $m^{-2}$, spread over about 80 km width (figure 1). If shear on vertical planes over this same width accommodates the relative motion of the Pacific and North America plates, then the shear strain rate is *ca*. $2\times10^{-14}$ $s^{-1}$. In a steady-state situation, the anomalous surface heat flux produced by dissipation is the product of a force per unit length, that is the integral from the surface to the base of the lithosphere of the associated deviatoric stress, and the strain rate, and these values of strain rate and anomalous heat flux imply that the integral of the shear stress through the lithosphere in the southern Coast Ranges is $1.5\times10^{12}$ N $m^{-1}$. To obtain rough estimates of the shear stress, let us assume that the strong portion of the lithosphere is only 15 (30) km thick, and that stresses elsewhere in the lithosphere are much lower. The average shear stress in the strong layer is, on this assumption, 100 (50) MPa.

Molnar (1991) made a similar calculation but assumed a width of 200 km, instead of 80 km, for the deforming zone. Accordingly, he deduced a proportionally larger





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Figure 1

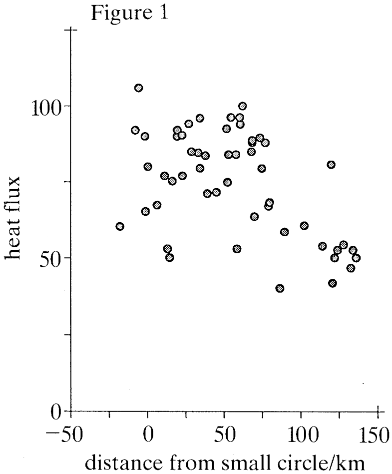


Figure 2

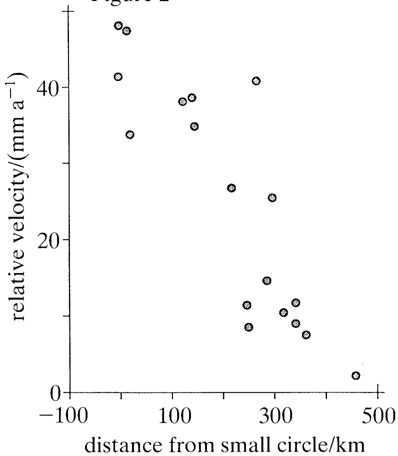


Figure 1. Observations of surface heat flow in California between 35° and 40° N, from Lachenbruch & Sass (1980, table 1). Observations from south of 35° N are excluded, where vertical advection of heat associated with crustal shortening in the Transverse Ranges, and with extension further south, presumably influences surface heat flux. Observations are plotted against distance from a small circle about the Euler pole describing the relative motion of the Pacific and North America plates (De Mets *et al.* 1990). The small circle passes through 38.1° N, 122.9° W (Point Reyes VLBI station).

Figure 2. Observed velocities of VLBI sites within California and Arizona. Velocities are observed in a frame that minimizes the velocities of VLBI sites within 'stable' North America (Ward 1990, tables 2 and 3, and fig. 3). The magnitudes of the velocities are plotted against distance from the same small circle used in figure 1. These data come predominantly from outside the region to which figure 1 refers, and do not provide an estimate of the shear strain rate within that region.

shear stress. Note that in southern California, where geodetic measurements suggest that the deformation is distributed over a width of 200 km or more (figure 2), there is no detectable heat flux anomaly associated with the San Andreas fault zone (Lachenbruch & Sass 1980, fig. 19). With the same shear stresses estimated above for central California, but with a shear strain rate three times smaller, the corresponding heat flux anomaly would be $10\ \mathrm{mW\ m^{-2}}$. If, as seems likely, the shear stress diminishes with strain rate, this anomaly would be still lower.

The shear stresses estimated above for the lower lithosphere of the San Andreas fault system are similar in magnitude to the shear stresses estimated by Scholz *et al.* (1979) for mid-to-lower crustal levels near the Alpine fault of New Zealand. Thus, the apparently discrepant estimates of shear stress near major strike–slip faults that were cited at the beginning of this section may be reconciled by postulating that deformation of the uppermost crust of such regions is driven by tractions applied to its base by a stronger substrate that is deforming in a distributed fashion (Molnar 1991). This stronger substrate may lie in the lower crust or in the upper mantle, or in both.

## 3. Inferences of lithospheric strength from continuum calculations of continental deformation

Some success has been achieved in investigating the largest scale of continental deformation (hundreds to thousands of kilometres) by treating the continental





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lithosphere as a continuous medium (Bird & Piper 1980; England & McKenzie 1982, 1983; England *et al.* 1985; Houseman & England 1986; Vilotte *et al.* 1982, 1984, 1986). At the core of this approach is the recognition that most major zones of active continental deformation are much greater in their horizontal extents than they are thick (typically they are several hundreds to thousands of kilometres wide, whereas the lithosphere is about 100 km thick). This configuration, coupled with the assumption that *shear* tractions on the base of the continental lithosphere are generally negligible compared with those within it, yields the result that the deformation of the continental lithosphere depends principally upon the *vertical averages* of the stresses acting within it and on its rheology. The observations of strain rate and of topography that are summarized below are made over a horizontal scale of several hundreds of kilometres. It must be remembered that they are, therefore, incapable of directly yielding information on the variation with depth of stress or rheology.

At its simplest the continuum approach treats the lithosphere as a power law fluid whose rheology is characterized by two dimensionless parameters only (England & McKenzie 1982, 1983). The first of these parameters is the exponent in a power law that describes the rheology of the lithosphere (equation (3)). The second parameter (called the Argand number by England & McKenzie (1982)) expresses the ratio of the stresses associated with isostatically compensated density contrasts within the lithosphere (referred to here as 'buoyancy forces') to the stresses required to deform the lithosphere at a characteristic strain rate. These two parameters influence the deformation in different ways. The influence of the power law exponent in the rheology is seen most clearly in the way that it determines the shapes of deforming zones (England & McKenzie 1982, 1983; England *et al.* 1985), whereas the Argand number principally influences the magnitude of crustal thickness contrasts that can be supported by a given boundary condition (England & McKenzie 1982, 1983; Houseman & England 1986; Vilotte *et al.* 1986).

### (*a*) *Distribution of strain within deforming regions*

The representation of the vertical average of the rheology of continental lithosphere by a power law is partly a matter of convenience. A relationship between strain rate, $\dot{\epsilon}$, and deviatoric stress $\tau$ of the form,

$$\dot{\epsilon} \propto \tau^n, \tag{3}$$

can represent, with the variation of a single parameter, a wide range of flow laws, from that of a newtonian fluid with $n = 1$, to a purely plastic material as $n \rightarrow \infty$. Equation (3) is not merely convenient, however. Sonder & England (1986) showed it to be a close approximation to the vertical average of the rheology of lithosphere failing by slip on faults in its upper crust and by a combination of creep and brittle failure in the lower lithosphere, according to the strain rate and temperature there. (The success of this approximation derives from the ease with which one can obtain a straight line by plotting the logarithm of one quantity against the logarithm of another, not from any single fundamental physical process.) Sonder & England (1986) point out that the exponent $n$ in equation (3) does not represent any individual deformation mechanism. Rather, it represents the vertical averaging of deformation involving slip on faults (for which stress is essentially independent of strain rate, and therefore $n \approx \infty$) with deformation by creep (commonly with $n \approx 3$).

If continental lithosphere shows deformation that is consistent with its vertically averaged rheology having $n \approx 3$, then it may be assumed that a relatively small fraction of the total strength of the lithosphere is supported by friction on faults, and that most of the strength resides in the creeping lower lithosphere; in contrast, values of $n$ higher than 10 indicate that a substantial fraction of the strength may be supported by friction on faults (Sonder & England 1986).

If continental lithosphere does behave as a thin sheet of fluid that obeys equation (3), then, in the absence of buoyancy forces, deformation associated with a boundary of length, $D$, ought to die out exponentially with distance from that boundary over a characteristic distance that is proportional to $D$ and is inversely proportional to the square root of $n$ in the vertically averaged rheology (equation (3)) (England *et al.* 1985). Furthermore, the across-strike width of the deforming region should be four times smaller for a transcurrent boundary than for a compressional or extensional boundary of the same length. Because the aspect ratios of the major active zones of continental deformation are in qualitative agreement with this prediction (England & Jackson 1989), it seems worthwhile to try to extract, from the aspect ratios of actively deforming regions, estimates of the exponent, $n$, that characterize the vertically averaged rheology of the lithosphere of those regions. Because compressional or extensional deformation leads inevitably to changes in crustal thickness, and because lateral variation in crustal thickness is a major contributor to buoyancy forces within continental lithosphere, the discussion below is restricted to inferences of $n$ from the aspect ratios of zones of transcurrent deformation, where buoyancy forces may be neglected. In particular, we concentrate upon the Pacific–North-America plate-boundary zone in California, upon the South Island of New Zealand, and upon the Pacific Northwest of North America.

The characteristic across-strike dimension, $l$, of a zone of transcurrent deformation, of along-strike length $D$, in lithosphere whose vertically averaged rheology is given by equation (3), is:

$$l \approx D/\pi\sqrt{n} \tag{4}$$

(Sonder *et al.* 1986).

Analysis of VLBI baseline lengths between stations on the Pacific and North America plates showed that the rate of displacement between these stations and stable North America increases smoothly with distance from the Pacific–North-America pole over a zone that is about 300 km wide (figure 2) (Sauber 1989; Ward 1990). It is clear from each study that velocities within this deforming region are nearly parallel to the overall Pacific–North-America relative motion. The distance over which the relative velocity falls from the full Pacific–North-America velocity of about 50 mm $a^{-1}$ to $1/e$ of that value is about 250 km (figure 2).

It is, of course, very difficult to distinguish, on the basis of geodetic measurements spanning a decade, or even a century, between elastic strain and permanent deformation. Geodetic observations cannot, therefore, be taken as directly relating to the deformation of the lower lithosphere. In southern California, however, the uniform-sense rotation of crustal blocks shown by palaeomagnetic data (Luyendyk *et al.* 1985) suggests that the region underwent, since middle Miocene time, distributed permanent deformation over approximately the same lengthscale as is shown by the geodetic observations of strain. McKenzie & Jackson (1983) pointed out that such rotations of crustal blocks could be explained if the upper crust were regarded as a set of small rigid blocks floating on a viscous substrate whose deformation they followed. In particular the rate of rotation, $\dot{\theta}$, of roughly circular blocks about a vertical axis would be:

$$\dot{\theta} = \tfrac{1}{2}(\partial u_x/\partial y - \partial u_y/\partial x), \tag{5}$$

where $u_x$ and $u_y$ are the $x$- and $y$-components of the velocity, and $z$ is vertical. If the geodetic observations of figure 2 are taken to yield a gradient of velocity, say $\partial u_x/\partial y$ with $x$ parallel to Pacific–North-America relative motion, and $y$ perpendicular, then the rate of rotation of crustal blocks perfectly coupled to a fluid with such a velocity gradient would, at the boundary, be:

$$\dot{\theta} \approx \tfrac{1}{2} \times 55 \text{ mm a}^{-1}/250 \text{ km}, \tag{6}$$

or about 6° per million years, dropping to $1/e$ of that value at about 250 km from the boundary.

Average rotation rates about vertical axes determined from late Neogene rocks in southern California are a few degrees per million years (see Jackson & Molnar 1990, table 2). These observations support qualitatively the notion that the deformation of the upper crust of California follows passively the velocity gradient field of a continuous substrate. Sonder *et al.* (1986) also reached this conclusion by using a continuum calculation of the deformation associated with the past 29 Ma plate interactions in the region to analyse the palaeomagnetic data. Jackson & Molnar (1990) show that the active faulting and block rotation of the western Transverse Ranges are consistent with the motion of crustal blocks driven by tractions applied to their bases by a viscous substrate.

Making use of equation (4) with $l = 250$ km and taking a plate-boundary length of 1200 km (Salton Sea to Cape Mendocino), yields an estimate of $n \approx 3$ for the vertically averaged rheology of the lithosphere in California.

The Pacific Northwest of North America has experienced transcurrent deformation as a result of the relative motion of the North America and Juan de Fuca plates. Observations of palaeomagnetically determined rotations of rocks of the Columbia River Basalt Group allowed England & Wells (1991) to estimate the across-strike lengthscale of that deformation. The rotations die out to about $1/e$ of their values near the coast over a distance of about 150 km. The along-strike length of the plate boundary is about 1200 km which, from equation (4), gives an estimate of $n \approx 4.5$.

Determinations of strain rate in the South Island of New Zealand, based on retriangulation spanning about a century, and summarized and interpreted by Walcott (1984), show that the transcurrent deformation associated with the relative motion of the Pacific and Australia plates is distributed through a zone approximately 150 km wide in which the shear strain rate diminishes with distance east of the Alpine fault. The along-strike length of this zone is about 800 km. These dimensions yield a ratio $D/l \approx 5.5$, implying $n \approx 3$ if, as for California, the geodetic shear strain reflects distributed deformation of the lower lithosphere and not elastic strain of the upper crust.

### (*b*) *Topographic contrasts supported by continental lithosphere*

The distribution of strain rates within active compressional belts may be used to infer levels of vertically averaged deviatoric stress in lithosphere of these regions. The presence of large plateaux in the Andes, Iran and Tibet implies that the stresses required to drive deformation of the lithosphere in these regions are in approximate





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balance with the buoyancy forces associated with the thickened crust underlying the plateaux (Molnar & Tapponnier 1978; Dalmayrac & Molnar 1981; England & McKenzie 1983; Houseman & England 1986; Vilotte *et al.* 1986; England & Houseman 1988, fig. 3; Molnar & Lyon-Caen 1988). In particular, both the Andes and Tibet exhibit a transition in tectonic style that is strongly correlated with surface elevation. Each region is in an overall convergent setting, and is surrounded by thrust faulting, yet in its highest part, each is undergoing normal faulting and crustal extension. The transition between extensional and thrust faulting earthquakes in the region of Tibet occurs at an elevation of about 4 km (England & Molnar, unpublished data), and in the Andes Quaternary normal faulting (almost seismically inactive during the instrumental period) is found above about 3–4 km, although the transition is less clear-cut than in Tibet (Dalmayrac & Molnar 1981; Sébrier *et al.* 1985).

These observations allow us to identify the surface height in each region at which the vertically averaged vertical stress acting on horizontal planes, $\bar{\sigma}_{zz}$, is equal to the vertically averaged horizontal stress acting on vertical planes, $\bar{\sigma}_{xx}$ (Dalmayrac & Molnar 1981). Because $\bar{\sigma}_{xx}$ varies slowly with distance, estimating $\bar{\sigma}_{zz}$ at the transition between normal and thrust faulting yields an estimate of the horizontal driving stress supporting the entire mountain range (Dalmayrac & Molnar 1981).

Unfortunately, determining $\bar{\sigma}_{zz}$ associated with a given surface height is not a simple matter. England & Houseman (1989, Appendix A), show that the magnitude of $\bar{\sigma}_{zz}$ is so dependent upon assumptions about the density distribution within the lithosphere, that a quantitative estimate of $\bar{\sigma}_{zz}$ based upon surface height alone is impossible. Lateral variations in $\bar{\sigma}_{zz}$ are directly related to variations in geoid height (Turcotte & Schubert 1982), but the geoid height is generally poorly determined over mountainous regions.

Molnar & Lyon-Caen (1988) estimate that the difference between $\bar{\sigma}_{zz}$ for the 5 km high Tibetan plateau and its surrounding lowlands is 70–80 MPa, or $7\text{–}8 \times 10^{12}$ N m$^{-1}$ over a lithosphere 100 km thick. This difference is proportional to the difference between the squares of the crustal thicknesses of the two regions (England & McKenzie 1982). Let us take 35 km to be the thickness of continental crust whose surface lies at sea level, and 65 km for the thickness of crust whose surface lies at 5 km above sea level. The difference $\bar{\sigma}_{zz} - \bar{\sigma}_{xx}$ for crust whose surface height is at 4 km above sea level is, then, $5\text{–}6 \times 10^{12}$ N m$^{-1}$. The transition to normal faulting in the Andes takes place at a somewhat lower elevation, about 3 km (Sébrier *et al.* 1985; Mercier 1991). By the same kind of calculation, the driving force supporting the Andes at present is about $3\text{–}4 \times 10^{12}$ N m$^{-1}$. These driving forces are too large to be attributed to the driving force from the mid-ocean ridges, but descending and negatively buoyant lithosphere like the slab beneath the Andes can add appreciably to the driving force (Bott 1990; Bott *et al.* 1989; Molnar 1988*b*; Lyon-Caen & Molnar 1985).

In their present tectonic régimes the plateaux show an approximate balance between buoyancy forces arising from crustal thickness contrasts and the driving forces applied to their edges by the convergence of the blocks surrounding them. In the past, however, the plateaux were lower, and the driving forces were resisted mainly by the shear stresses required to deform the lithosphere (England & Houseman 1988, fig. 3; England & McKenzie 1983; Houseman & England 1986). Furthermore, each plateau is bordered by lower-lying regions that are now shortening in response to these same driving forces. Consequently, we may presume that the continental lithosphere of these plateaux, and their surroundings, can support driving forces of about $3–6 \times 10^{12}$ N m$^{-1}$ at geological strain rates.

## 4. Conclusion

The heat fluxes observed, and the temperatures inferred from metamorphic petrology, in major thrust zones imply that the shear stresses in such zones are generally between 50 and 150 MPa (§2*a*). Shear stresses of this magnitude persist to depths of 40 km or greater, and can contribute $2–4 \times 10^{12}$ N m$^{-1}$ (or more if such stresses are maintained at depths greater than 40 km) to the support of mountain ranges bordering these thrust zones. The relations between the style of active faulting and surface elevation in the Andes and Tibet (§3*b*) imply that the continental lithosphere of these regions requires a driving force of $3–6 \times 10^{12}$ N m$^{-1}$ to overcome resistance to deformation. These driving forces are equivalent to vertically-averaged shear stresses of 15–30 MPa over a lithosphere 100 km thick, but if supported by, a layer for example, 30 km thick within the lithosphere, they would require shear stresses of 50–100 MPa.

The aspect ratios of the major active zones of transcurrent deformation (§3*a*) suggest that the vertically averaged rheology of the lithosphere is described by a power law (equation (3)) with $n \approx 3$. This observation implies that deformation by high-temperature creep is more important in determining the strength of the lithosphere of these regions than is friction on faults. The clockwise rotations recorded in Neogene rocks of western North America show that the lithosphere of the region has undergone distributed deformation in response to strike–slip boundary conditions, in a fashion consistent with the deformation of a viscous fluid (§3*a*). This implication is supported, but not proved, by the geodetic determinations of strain summarized in figure 2. Estimates of the shear stress acting below the seismogenic upper crust of the San Andreas fault system, and in amphibolite-grade rocks exposed near the Alpine fault of New Zealand (§2*a*), lie in the range 50–100 MPa. Note that these estimates are not at variance with the conclusions of Mount & Suppe (1987) and Zoback *et al.* (1987) that the shear stress in the upper crust on the San Andreas fault is low.

Several other lines of evidence also suggest that the upper crust does not control the regional distribution of deformation in tectonically active parts of the continents. The distributions of strain and strain rate in central and southern Asia agree well with those calculated assuming that the lithosphere behaves as a continuous, power-law, fluid (England & Houseman 1986). Even such apparently dominant features as the major strike–slip faults in eastern Tibet are probably rotating, in response to the continuous deformation of their substrate (England & Molnar 1990). Similarly, the distributions of active strain rate and total extensional strain in the Aegean region appear to be consistent with those of a temperature-dependent fluid such as olivine, rather than with the strain of lithosphere controlled by friction on faults (Sonder & England 1989).

The observations of distributed strain in active belts, and the inferences of shear stress magnitudes within these belts, support the idea that the uppermost, seismically active, part of the continental lithosphere is generally much weaker than a layer beneath it that deforms in a continuous fashion. This layer probably contains the uppermost mantle, and may include the lower crust. It seems likely that the motions of, and the stresses within, the upper crust reflect the response of this heterogeneous,





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anisotropic, and discontinuous layer to tractions applied to its base by distributed deformation of the stronger lower lithosphere. Information on the state of stress in the lower lithosphere is sparse and indirect. Perhaps the most valuable measurement that could be made to constrain better the state of stress in deforming belts would be an improved determination of the gravity field of the continents.

We thank M. H. P. Bott for a helpful review. This work was supported by NERC grant GR3/7032 and NASA grant NAG5-795.

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