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MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Active Continental Deformation and Regional Metamorphism [and Discussion]

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Active continental deformation and regional metamorphism

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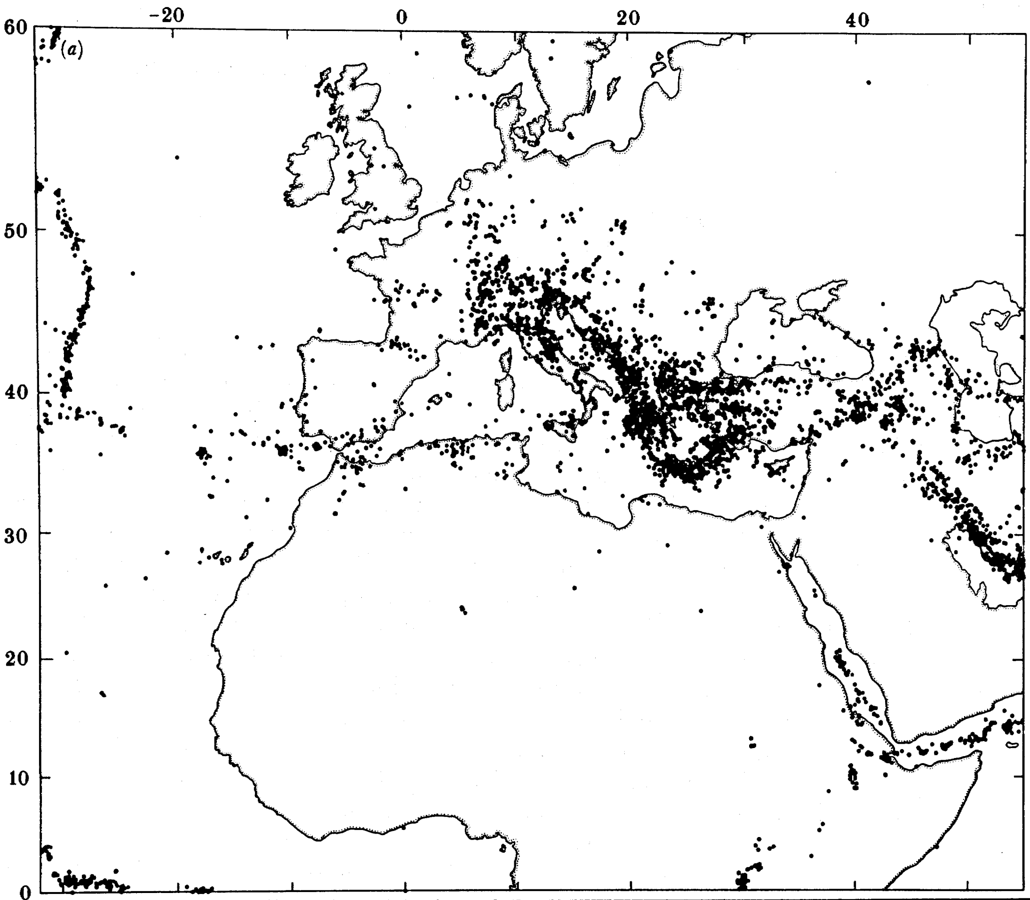
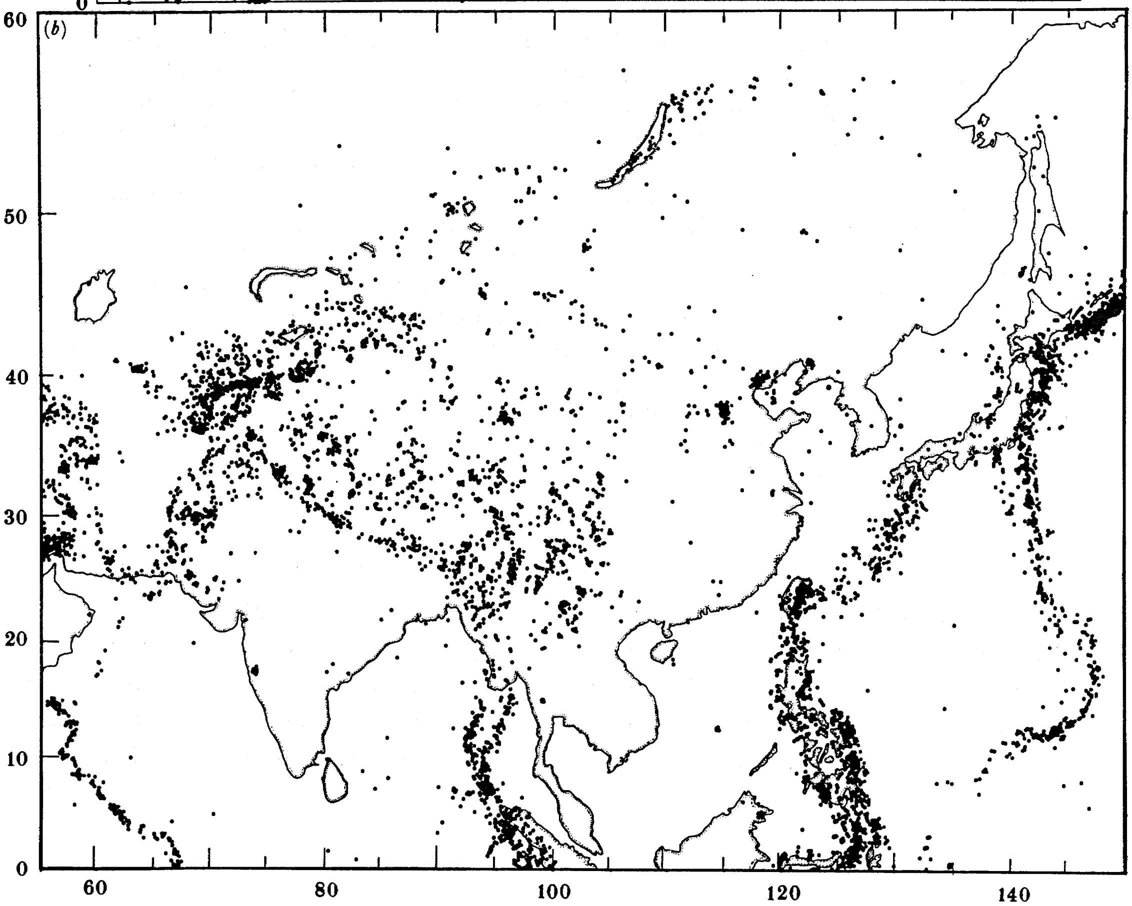
The vertical and horizontal distribution of present-day continental deformation is examined to see how tectonic movements may be related to large wavelength perturbations to the temperature and pressure experienced by rocks in the crust. Earthquakes are generally restricted to the upper part of the continental crust. The lower crust is usually aseismic and assumed to be weaker. The uppermost mantle beneath continental regions has minor seismic activity that does not account for much deformation, but probably indicates an important strength contrast between the lower continental crust and the upper mantle. The maximum focal depth of earthquakes in any region appears to be limited by temperature, with most restricted to material colder than 350 ± 100 °C in the crust and colder than 700 ± 100 °C in the mantle. At length scales long compared with the thickness of the brittle upper crust, the deformation in regions of continental extension or shortening appears to be continuous, even though, in reality, discontinuous movement on faults occurs. This probably indicates that the deformation is dominated by distributed flow in the ductile portion of the lithosphere and not by the behaviour of the thin brittle upper crust. The distribution of seismicity, elevation contrasts and vertical movements at the surface suggests that there is little spatial separation between the brittle deformation in the upper crust and the ductile deformation below on length scales larger than the lithosphere thickness. For this reason, and because of the short thermal time constant of the crust, long-wavelength perturbations to the thermal régime are more influenced by the behaviour of the lithosphere as a whole than by the precise geometry of deformation in the crust. Large-scale regional metamorphism in zones of shortening may result from the re-establishment of the initial geotherm in thickened crust when the lower part of the lithosphere detaches and falls into the asthenosphere. In regions of extension, an increased geothermal gradient is unlikely to result in regional metamorphism unless magmatic augmentation to the heat supply is important. However, if the stretched region is covered by thick sediments, the basement may experience a small increase in temperature and remain significantly hotter than it would be if there were no sediment cover. While unlikely to account for significant metamorphism, this effect may strongly influence the rheological behaviour of the lithosphere in extending regions. The rapid vertical movements associated with syn- or post-orogenic normal faulting in regions of large-scale crustal thickening are probably at least as important in exhuming mid-crustal metamorphosed rocks, and in disrupting patterns of isograds, as those associated with erosion.

1. Introduction: where is regional metamorphism occurring today?

Regional metamorphism requires a perturbation to the temperature and pressure experienced by rocks over a wide area. One way in which this is probably achieved is by the deformation of the crust or lithosphere during the tectonic movements that create most of the large-scale topographic features on the Earth's surface. Exhumed regionally metamorphosed terranes are often intensely deformed, and this common, although not always well understood, association





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between regional metamorphism and deformation suggests that it is sensible to begin an inquiry into where regional metamorphism is occurring today by examining areas of presently active continental tectonics. The most obvious manifestation of active tectonics is seismicity. The occurrence of an earthquake indicates unequivocally the presence of an active fault and hence of deformation. Unfortunately, as we shall see, a lack of earthquakes on the continents does not necessarily indicate an absence of deformation, as it is known that much movement is taken up aseismically. None the less, because they are common, and can be studied at great distances from their epicentral regions, large earthquakes are the most accessible source of information on active continental tectonics, and, until recently, provided most of our understanding of this subject.

The last decade has seen a steady improvement in the ability of seismological techniques to extract information about earthquake source processes. In particular, the orientation and depth of seismogenic faults can now be resolved with sufficient accuracy to provide useful constraints on many problems in structual geology. As a consequence, seismogenic fault geometries in actively deforming belts are now much better known, and clear patterns are starting to emerge.

This paper reviews some of the recent advances in our understanding of active continental tectonics that are relevant to possible settings of regional metamorphism. Particular emphasis is placed on the geometry of the deformation observed in the upper crust and on the consequent implications for the deformation of the lower crust and mantle part of the lithosphere. Likely perturbations to the regional geotherm are also briefly discussed. Most of this paper is concerned with observations from the Alpine–Himalayan Belt (figure 1): the main active continental collision zone today, within which occurs a great variety of deformation including shortening, extensional and strike-slip motions. Since this review is aimed at those not already familiar with active continental tectonics, most of the discussion is a rapid, and sometimes generalized, treatment of much detailed and careful observational work by other people. Consequently, I would urge those developing a new interest in this subject to read the primary sources referenced here, and assess the quality of the observations themselves.

2. The vertical distribution of seismicity: focal depths

Earthquake foci on continents are not uniformly distributed with depth. Such a statement could not have been made with confidence ten years ago, because reliable focal depths are deceptively difficult to obtain by any routine method. Depths computed from travel times on a routine basis by agencies such as the U.S. Geological Survey and the International Seismological Centre may be in error by 50 km or more and cannot be accepted at face value. However, either when data from dense local seismograph networks are used, or when the observed teleseismic waveforms from earthquakes are modelled synthetically, focal depths can be estimated to within 3 or 4 km (see e.g. Chen & Molnar 1983). As a result of these advances in seismological techniques we now have a good knowledge of the depth distribution of seismicity on the continents, and the following generalizations can be made. (i) Most continental earthquakes nucleate in the upper 25 km of the crust. The maximum focal depth in the crust varies regionally and is taken, through common usage, to define the base of the seismogenic or 'brittle' upper crust. (ii) Some continental earthquakes nucleate in the uppermost upper mantle. These appear to be restricted spatially to those areas where the upper crust is also seismically active. (iii) The lower crust appears to be almost completely aseismic. (iv) Focal depths greater than 70 km are very rare outside oceanic subduction zones.

The data supporting this distribution of focal depths are summarized by Sibson (1982), Meissner & Strehlau (1982) and Chen & Molnar (1983), among others. The data span a wide range in earthquake magnitude, from microearthquakes located by dense local seismograph networks to large earthquakes whose teleseismic waveforms were modelled synthetically. The focal depth distribution is usually clearer, with a more sharply defined cutoff at the base of the upper crust, for the larger earthquakes.

It has been thought for some time that the maximum focal depth of earthquakes in any region is limited by temperature. McKenzie (1969) suggested that focal depths in the Tonga–Kermadec subducting slab were confined to mantle regions cooler than about 700 °C, and Brace & Byerlee (1970) suggested that the lack of seismicity deeper than 12 to 15 km on the San Andreas fault zone is caused by the cessation of brittle failure above a limiting maximum temperature. With more abundant and more accurate focal depth observations, particularly within California, the maximum depth at which seismicity occurs in the crust and mantle appears to be related to regional heat flow (see, for example, Sibson 1982; Meissener & Strehlau 1982). In spite of the difficulties in estimating continental geotherms (see Sclater *et al.* 1980) it appears that seismicity in the continental crust is confined to regions colder than about 350 ± 100 °C (Chen & Molnar 1983). In the mantle, the maximum temperature at which seismicity occurs is best estimated from observations in the oceans, where geotherms can be estimated more reliably than on the continents (Parsons & Sclater 1977). Intraplate earthquakes in the oceanic mantle appear to be confined to regions colder than about 700 ± 100 °C (Chen & Molnar 1983; Wiens & Stein 1983; Grimson & Chen 1986). Before continuing this discussion of focal depth distribution it is important to emphasize a point made by Chen & Molnar (1983); that although temperature, stress and rheology in the lithosphere must all be estimated or extrapolated from laboratory or surface observations, the earthquake focal depths themselves are real *in situ* measurements under geological conditions, and are not extrapolations. For this reason the vertical distribution of seismicity provides strong constraints on acceptable models of lithosphere deformation and rheology.

Qualitatively, the interpretation of the observed focal depth distribution is usually as follows. Both the brittle and ductile strengths of materials are known to depend on several physical parameters, such as temperature, confining pressure, pore pressure, strain rate and the material itself. At low temperatures, brittle frictional sliding dominates the deformation and the coefficient of friction appears to be about the same for a wide variety of rock and mineral types (Byerlee 1978). The brittle strength τ_B, is thus expected to be roughly proportional to depth, z.

$$\tau_B = Bz, \tag{1}$$

where B is determined by experiment.

At high temperatures the most likely mode of deformation for geological materials is some form of plastic creep or flow. Several different dislocation creep mechanisms are known, which depend on strain rate ($\dot{\epsilon}$), temperature (T) and differential stress (τ_D) (see, for example, Stocker & Ashby 1973) with relations of the form:

$$\tau_D = (\dot{\epsilon}/A)^{1/n} \exp{[Q/nRT]}, \tag{2}$$

where R is the gas constant and A, Q, and n are material constants determined by experiment.

At any particular depth the maximum sustainable differential stress is expected to be the lesser of τ_B or τ_D, with failure in the brittle régime generating earthquakes and failure by creep being aseismic. Thus the vertical distribution of seismicity is seen as the result of two competing effects: the increase of frictional strength with confining pressure and depth, and the decrease of creep strength as temperature increases with depth. The difference in composition between the crust and mantle is also likely to be important since laboratory experiments show that, at temperatures in the range 300 to 1000 °C, the creep strength of mantle materials, such as olivine, is likely to be much greater than that of crustal materials, such as quartz and diabase (Brace & Kohlstedt 1980).

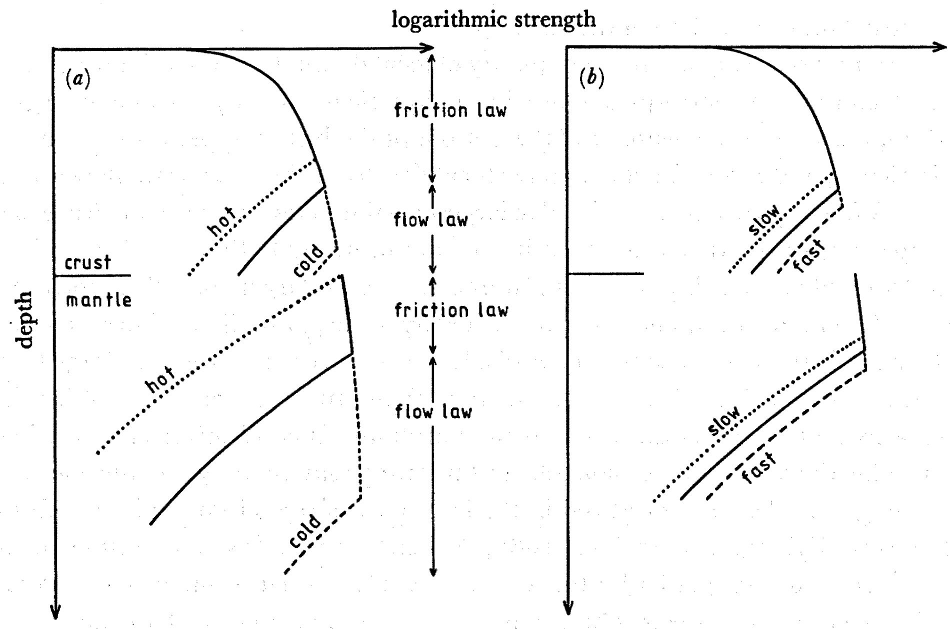

FIGURE 2. Schematic profiles of strength against depth, for the continental crust and uppermost mantle (after Chen & Molnar 1983). The strength at any depth is taken to be the smaller of τ_B (friction law: equation 1) and τ_D (flow law: equation 2). In (*a*), perturbations to the curve for a 'normal' geotherm (solid) are shown for hotter (dotted) and colder (broken) geotherms, corresponding to differences in thermal gradient of ± 5 K km^{-1}. All three curves are shown for the same strain rate. In (*b*), curves are shown for fast, 'normal', and slow strain rates (differing by a factor of 10) under the same geothermal gradient. Following Chen & Molnar (1983), the scales have been omitted to emphasize the uncertainty in actual values of strength. In reality, changes in strength at the transition from the friction to flow law and at the Moho are likely to be less abrupt than shown.

A quantitative interpretation of the distribution of focal depths in terms of material strengths estimated from laboratory experiments is difficult for a number of reasons. Most laboratory experiments are carried out on monomineralic or relatively homogeneous materials, whereas earthquakes occur on pre-existing faults where reduced grain size, retrogressive mineral assemblages and anisotropic crystallographic fabrics are likely to make the fault (or deeper down, the mylonite zone) weaker than its surrounding rock (see, for example, White 1976; Etheridge 1986). In spite of these difficulties, the correlation between estimated subsurface temperatures and maximum focal depths implies that the responses of frictional and creep strengths to increased temperature and pressure with depth that are suggested above are qualitatively correct.

Figure 2 (which is based on figure 3 of Chen & Molnar 1983) shows schematic profiles of strength as a function of depth for a mantle overlain by crust. The curves represent the relations in (1) and (2) calculated by using laboratory estimates of material constants for diabase (representing the crust) and olivine (representing the mantle), but, following the example of Chen & Molnar (1983), the numerical scales are omitted to emphasize how uncertain the actual stresses are. Figure 2 shows the main effects suggested by Chen & Molnar (1983) to account for the observed distribution of focal depths: (i) brittle strength increases with depth below the surface to a maximum within the crust, below which stresses are relieved by ductile flow and earthquakes are not generated, and (ii) a large increase in strength at or near the Moho caused by the change from crustal to mantle composition. Under suitable temperature or strain rate conditions this increase in strength may be sufficient for the upper mantle to deform by brittle failure rather than ductile flow.

With the recent increase in amount and quality of focal depth data other details of the depth distribution of continental earthquakes are becoming clear. In any particular region, the largest earthquakes seem to nucleate near the bottom of the brittle upper crust (as defined by the local seismicity), and most aftershocks have focal depths the same as or shallower than that of the mainshock (see e.g. Strehlau 1986). The largest earthquakes appear to nucleate at depths where the upper crust is predicted to reach its maximum strength. Since this depth is usually in the range 10 to 20 km, earthquakes with source dimension larger than this (corresponding to magnitude $M_s > ca.$ 6.0 or moment $M_0 > ca.$ 10^{26} dyne cm†) usually lead to surface faulting as rupture extends from the hypocentre upwards. However, because of the very large transitory increase in strain rate during seismic faulting, from about 10^{-13} to about $10^{-4}\ s^{-1}$, it is likely that rupture also continues downwards into the normally 'ductile' lower crust during large earthquakes. The downward extension of rupture propagation may be the cause of the anomalously long-period signals observed in the later part of the seismic radiation from some large earthquakes (Eyidogan & Jackson 1985; Nabelek 1985). Discrete faulting below the brittle upper crust is also suggested by the asymmetry of the vertical movements observed at the surface in extensional grabens whose fault geometry is symmetrical to the base of the seismogenic upper crust (see e.g. Jackson & McKenzie 1983; King *et al.* 1985; Jackson 1986).

Thus there appears to be a transition zone at the base of the brittle upper crust where the mode of deformation is dependent on strain rate; during the low strain rates between large earthquakes it may deform by ductile flow, which could be either distributed over a large volume or concentrated on mylonite shear zones, but during large earthquakes it may fail in a brittle fashion. This effect may be responsible for the reworking of pseudotachylites by mylonites described from some exhumed mid-crustal shear zones (Sibson 1980). We have no information on how far into the normally ductile lower crust such earthquakes faulting may penetrate.

In continental regions, earthquakes in the uppermost mantle away from subduction zones are much rarer than crustal earthquakes and tend also to be smaller (Chen & Molnar 1983). The largest known of such earthquakes, that might have occurred in the uppermost mantle, is that of 13 November 1965 in the Tien Shan (43.8° N 87.7° E), at a depth of 45 km and with a magnitude (m_b) of about 6.3 (Chen & Molnar 1977). This low level of mantle seismicity presumably implies that even when the uppermost mantle is potentially seismic, a greater fraction of the deformation occurs aseismically in the upper mantle than in the crust (Chen & Molnar 1983). The relatively small magnitudes of these upper mantle earthquakes may imply

† 1 dyne = 10^{-5} N.

that the vertical extent of potentially 'brittle' upper mantle is not large enough to sustain earthquakes with a source dimension of more than a few kilometres.

Although figure 2*b* implies a dependence of focal depth on strain rate, except in the special case of downward rupture propagation during large earthquakes mentioned above (and even then only indirectly), this has proved difficult to demonstrate with observations, largely because of uncertainties in estimating regional strain rates. Regional variations in geothermal gradient are, anyway, likely to have a larger effect and are discussed in a later section.

In summary, continental earthquakes are largely confined to the upper crust (whose base, in this sense, they define) and uppermost mantle. The lower crust is largely aseismic and assumed to be weak. The largest continental earthquakes nucleate at or near the base of the brittle upper crust and rupture to the surface. Rupture is also likely to continue downwards into the lower crust, at least during the transitory high strain rates associated with brittle fracture of the upper crust during large earthquakes, but how far into the lower crust rupture extends under these conditions is not known.

3. The horizontal distribution of seismicity and deformation

The distribution of earthquake epicentres on the continents is quite different from that in the oceans (figure 1). Oceanic epicentres are confined to narrow belts that define plate boundaries, whereas continental epicentres are dispersed over deforming belts several hundred kilometres wide. This is not an effect caused by errors in epicentre location, which are about the same size as the dots in figure 1, but is real, and has been known for some time. This difference between oceanic and continental deformation is usually attributed to the combined effects of (a) the buoyancy of continental crust, which tends to inhibit subduction (McKenzie 1969), (b) the reactivation of old lines of weakness on continents, and (c) the weakness of continental lithosphere relative to oceanic lithosphere (McKenzie 1972; Chen & Molnar 1983).

Within the Alpine–Himalayan region are several large blocks that are relatively aseismic, such as Central Turkey, central Iran, southwest Afghanistan, the Tarim Basin and the Ordos Plateau. While it may be useful to describe the motions of these blocks relative to each other and relative to India or Eurasia by rotations about Euler poles, this plate tectonic description is little help in understanding the complicated motions within the wide deforming belts surrounding the blocks (see, for example, McKenzie 1972, 1978*a*; Jackson & McKenzie 1984 for the region in figure 1*a*; Molnar & Tapponnier 1975, 1978 and Tapponnier & Molnar 1977, 1979 for figure 1*b*).

An alternative approach is to abandon the attempt to describe the deformation by the interaction between rigid blocks, and to use a continuum description for the deformation within the wide active belts that bound large aseismic blocks and plates. Such descriptions usually model the deforming lithosphere as though it were a fluid, with smoothly varying strain rates and no strain discontinuities (Bird & Piper 1980; England & McKenzie 1982; Houseman & England 1986; England & Houseman 1985, 1986; McKenzie & Jackson 1983). At first sight such descriptions are absurd; earthquakes show that brittle deformation is important in the Earth's crust, and strain discontinuities, in the form of faults, are common. Yet continuum descriptions have been surprisingly successful in modelling the long wavelength topography and the spatial variation of deformation type in Asia (England & Houseman 1985, 1986). Recent palaeomagnetic work in southern California (Luyendyk *et al.* 1980, 1985), Greece

(Kissel *et al.* 1985, 1986), Israel (Ron *et al.* 1984) and New Zealand (Walcott 1984) shows that large rotations (up to 90° in 10^7 years) of palaeomagnetic declination about a vertical axis are common in regions of distributed continental deformation. If the crust is considered to be composed of blocks floating on a continuously deforming lower lithosphere, then McKenzie & Jackson (1983) show that the floating blocks will rotate with the angular velocity of the underlying fluid (figure 3), and suggest this as the cause of the observed palaeomagnetic

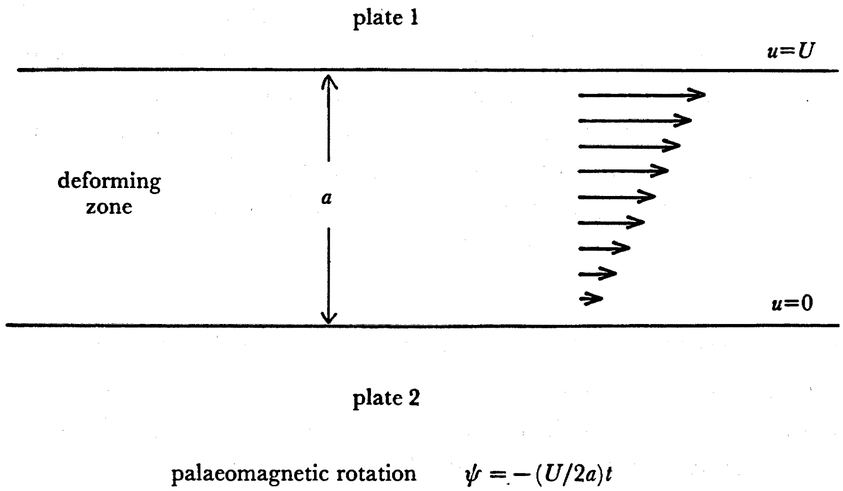

FIGURE 3. Plan view of a zone of continuous distributed simple shear between two rigid plates. If the velocity gradient is U/a across the zone, a rigid disc within the zone rotates at a rate $-U/2a$ (clockwise in this case). McKenzie & Jackson (1983) suggest that such rotations are analogous to the motions producing rotations of palaeomagnetic declination observed in rocks within deforming zones.

rotations. When this description is applied to southern California, the predicted rotations are in the same sense and are roughly the same amount as those seen palaeomagnetically by Luyendyk *et al.* (1980 and 1985). In New Zealand, Walcott (1984), by using geodetic data, was able to show that, on a scale of 40 km, the deformation across the actively deforming zone appears continuous. At a length scale of 40 km, not only does the short term (100 years) strain vary smoothly across the zone but it also agrees well in orientation and magnitude with that observed in the longer term (a few million years) from crustal uplift and palaeomagnetism. This was a surprising result because 100 years is short compared with the recurrence time for earthquakes on faults (a few hundred or thousand years). Thus although movement on faults does occur, and contributes to the long term strain, on a scale of 40 km the deformation in New Zealand can be modelled as though the faults do not exist.

There is therefore an apparent paradox: at long wavelengths the dilatational and rotational strains in regions of distributed continental deformation can be adequately described as continuous, yet we know that, in reality, the strain is discontinuous and concentrated on faults. This suggests that the overall deformation of the lithosphere is controlled by the thick lower ductile part, and that the thin brittle seismogenic layer on top accommodates the imposed strain by motion on many distributed faults (McKenzie & Jackson 1986). Then, on a length scale that is large compared with the thickness of the brittle layer, the surface deformation may be expected to look continuous.

While the deformation of the brittle upper crust appears to reflect that of the lower lithosphere, not all this deformation is taken up seismically. Whereas in Asia most of the motion of India northwards is probably accommodated seismically (Chen & Molnar 1977; Molnar & Deng 1984), further west, only a small fraction (*ca.* 5%) of the northwards motion of Arabia with respect to Eurasia is seen seismically (North 1974; Ambraseys & Melville 1982) and the rest must occur by creep. How much of the creep is concentrated on the faults that move seismically and how much is distributed within the blocks either side is unknown. Unless all the creep is concentrated on the seismogenic faults it is unlikely that all the movement in the lower crust is concentrated on the mylonite zones that represent the downward continuation of upper crustal faults.

In summary, present evidence suggests that motions in the brittle upper crust are controlled by the distributed deformation in the lower crust and mantle parts of the lithosphere. The next section summarizes what is known about large-scale motions in regions of strike-slip, compressional and extensional tectonics, and how these motions are likely to affect the geotherm on a scale comparable with the lithosphere thickness.

4. Tectonic influences on regional continental geotherms

(*a*) *Large-scale strike-slip motion*

Where the relative motion between large aseismic continental blocks is predominantly strike-slip, earthquake epicentres, geomorphology and geodetic measurements show that deformation is usually confined to a narrow zone a few kilometres or tens of kilometres in width (McKenzie 1972; Savage 1983), with large earthquakes often restricted to a single large strike-slip fault, such as the North Anatolian Fault in Turkey (Ambraseys 1970). By contrast, where crustal shortening or extension is predominant, deformation is usually spread over belts several hundred kilometres wide. The restriction of major strike-slip motion to relatively narrow belts is predicted by dynamic models of continuum deformation (England *et al.* 1985) and required by kinematic models (McKenzie & Jackson 1983), in which distributed strike-slip faulting is shown to be unstable because of consequent rotation about a vertical axis. On such major strike-slip systems the vast majority of earthquakes are confined to the upper 15 km of the crust, with no seismicity in the mantle (Chen & Molnar 1983). The absence of large-scale vertical movement, the lack of detectable frictional heating (at least on the San Andreas Fault; Lachenbruch & Sass 1980) and the narrowness of the deforming belt itself, make it unlikely that perturbations to the geotherm on a regional scale occur as a result of such strike-slip motions.

(*b*) *Large-scale crustal shortening*

Active continental shortening is usually distributed, with thrust faulting earthquakes and their accompanying topography spread over belts several hundred kilometres wide. Most of the largest earthquakes occur close to the edge of the highest topography (Jackson & McKenzie 1984), a result predicted by continuum models that calculate the differential stresses caused by buoyancy forces acting on elevation contrasts (England & McKenzie 1982). The gradient of such stresses is greatest where the product of the elevation and the gradient of the topography is a maximum. In the Middle East it is known that a substantial proportion of the shortening in the upper crust occurs aseismically (North 1974; Ambraseys & Melville 1982). Because regions of low seismicity, such as central Iran, are relatively flat and have low elevation compared with the elevated seismically active belts, it is thought that the aseismic deformation is restricted to those regions that are seismically active; although whether the aseismic deformation is also restricted to creep on faults or whether it is distributed, for example by folding, is not known. This large-scale separation between elevated seismic belts and topographically low aseismic regions suggests that, if long wavelength isostatic equilibrium is to be maintained, then the lower crust shortens in roughly the same regions as the brittle upper crust, at least over distances large compared with the lithosphere thickness.

Except along the southern, frontal edge of the Himalaya, where large thrusting earthquakes occur on faults dipping 0–20° northwards, most large dip-slip earthquakes in areas of crustal shortening involve high-angle reverse faulting within the basement on planes dipping 30–60° (see Molnar & Chen 1982, for a review). The predominance of high-angle reverse faulting in earthquakes does not necessarily imply that the crust shortens homogeneously in a concertina-like fashion. Low-angle thrust faults may move aseismically either above the high angle basement faults, perhaps on a weak layer at the basement–sediment interface (Jackson 1980), or within the lower 'ductile' crust beneath the seismogenic faults, at a level that is weak because of its temperature rather than its composition (see figure 2 and Chen & Molnar 1983). Whatever the precise geometry of crustal shortening, which probably varies considerably (Molnar & Chen 1982), crustal thickening is likely to depress the geothermal gradient, at least initially. It is thus not surprising that crustal earthquakes in regions of continental shortening occur to greater focal depths than those where extensional or strike-slip motion is occurring, with depths of 20–25 km not uncommon. It is in regions of crustal shortening that most of the upper mantle earthquakes reported by Chen & Molnar (1983) are located, although, as elsewhere, they are less common, and generally smaller, than earthquakes in the upper crust.

The time constant for the decay of a thermal perturbation of wavelength L is approximately $L^2/\pi^2 K$ where K, the diffusivity, is about $10^{-6}\ \mathrm{m^2\ s^{-1}}$ in the crust and uppermost mantle. If the crust is fragmented on a scale of 25 km, then perturbations due to a lateral change in thermal structure will decay with a time constant of only 2 Ma. For this reason, the precise geometry of the deformation in the upper crust has less influence on the thermal evolution (on large length scales) of the region undergoing shortening than the geometry of the lithosphere as a whole.

After shortening, the geothermal gradient is reduced, but, as heat continues to be supplied to the base of, and generated within, the lithosphere, the geothermal gradient will eventually become as great as, or greater than, it was originally. If re-establishment of the geotherm occurred more quickly than the removal of excess material by erosion and isostatic uplift, the lower crust would become much hotter than normal and might undergo widespread regional metamorphism and melting (Oxburgh & Turcotte 1974; England & Richardson 1977). However, re-establishing the initial geotherm purely by conductive heating of a thickened lithosphere is a slow process; for example, a lithosphere thickened to 240 km has a time constant of 180 Ma. During this time, erosion is likely to remove most of the excess crust so that melting and regional amphibolite facies metamorphism are not expected (Houseman *et al.* 1981; England & Thompson 1984). However, Houseman *et al.* (1981) show that the lower part of the cold, dense, thickened lithosphere beneath thickened crust will probably detach itself and fall as a blob into the asthenosphere. The asthenosphere then wells up to replace the detached boundary layer, bringing hot material that transfers heat into the thickened crust much more quickly than if it were conducted through the entire thickened lithosphere. The lower

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Table 1

(Four earthquakes within the East Africa Rift system whose focal depths are constrained to the range 25–30 km by bodywave modelling (Shudofsky 1985). These include the two largest earthquakes in the East Africa Rift system during the period 1964–1978. All data are from Shudofsky (1985).)

date	lat.	long.	depth km	m_b	moment, *Mo* 10^{24} dyn cm	mechanism	region
7 May 1965	3.90° S	35.10° E	28	6.4	70.8	strike-slip	Tanzania
20 Mar. 1966	0.81° N	29.90° E	29	6.0	85.1	normal	Ruenzori
15 May 1968	15.91° S	26.16° E	27	5.7	4.5	normal	Zambia
19 Sep. 1976	11.08° S	32.84° E	25	5.7	2.2	normal	Luanga Rift

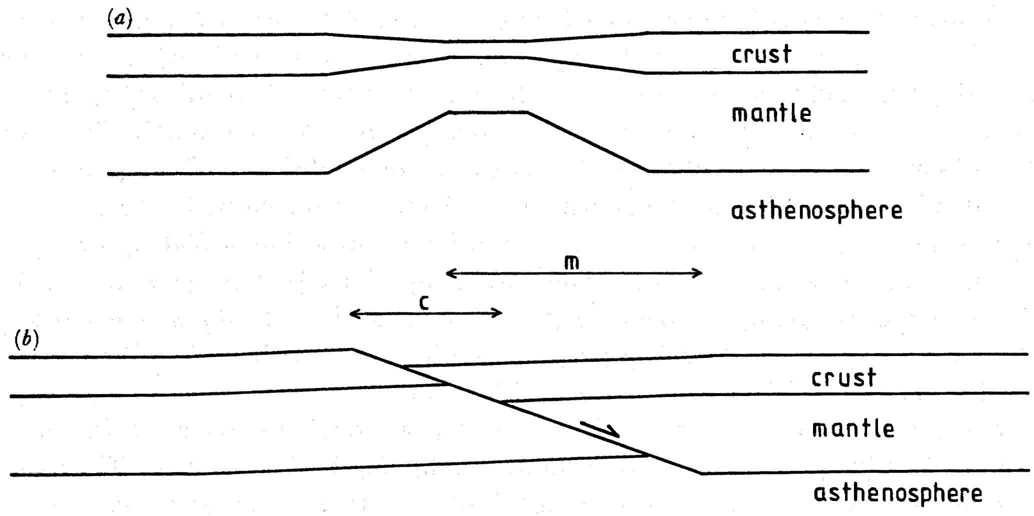

Figure 4. (*a*) Cross section to illustrate the uniform extension model of McKenzie (1978*a*, *b*). Although β, the extensional strain, varies across the basin, the crust and mantle parts of the lithosphere have stretched by the same amount in any place. (*b*) Non-uniform extension of the lithosphere, in which there is a spatial separation between the areas where crust is thinned, c, and those where the mantle part of the lithosphere is thinned, m. See, for example, Wernicke (1985).

lithosphere can detach and sink quickly (in less than 10 Ma), on a time scale short compared with the duration of the shortening episode. Under these circumstances, regional metamorphism may quickly follow shortening or even be contemporaneous with it. The experiments of Houseman *et al.* (1981) show that detachment of the lower lithosphere will not lead to a temperature gradient exceeding that before lithospheric thickening. Regional metamorphism under these conditions is thus seen as a response to crustal thickening rather than to increased temperature gradient at depth. Dewey & Burke (1973) have suggested that the volcanism in Tibet, where the crust is about 70 km thick (Chen & Molnar 1981), may be a consequence of such lower crustal melting. Pressure–temperature–time paths for material in thickened crust but with normal lithosphere thickness are discussed by England & Thompson (1984).

(*c*) *Crustal extension*

Active crustal extension, like active shortening, is also distributed over wide areas, such as Tibet (Molnar & Tapponnier 1978), the Aegean Sea and western Turkey (McKenzie 1978*a*), and the Basin and Range Province of the western USA (see, for example, Zoback *et al.* 1981). It was the correlation between high surface heat flow, crustal thinning and active normal faulting in the Aegean region that led McKenzie (1978*a*, *b*) to propose an extensional model for the origin of some continental sedimentary basins. In this model, extension of the whole lithosphere leads to subsidence and an increase in the thermal gradient during stretching. After stretching has ceased, subsidence continues at an exponentially decreasing rate as the thermal anomaly decays. It is now clear that extension has played a major part in the evolution of many continental sedimentary basins and, because of the expected increase in geothermal gradient, it is perhaps not surprising that focal depths in regions of active extension are, in general, somewhat shallower than those where shortening is taking place. Most crustal earthquakes in extensional regions have focal depths shallower than 15 km (Chen & Molnar 1983), with some larger events as shallow as 6 km (see, for example, Langston & Butler 1976; Soufleris & Stewart 1981; Eyidogan & Jackson 1985); although in the Basin and Range Province several recent large earthquakes had focal depths of 15 km (Doser 1985; Doser & Smith 1985). In general, upper-mantle events are rare in extensional regions (Chen & Molnar 1983). This pattern of focal depths appears less straightforward in Africa, where Shudofsky (1985) reports four large earthquakes in the East African rift system whose depths are shown by bodywave modelling to be in the range 25–30 km. These events are sufficiently unusual to be highlighted separately (table 1). Available seismic evidence (summarized by Shudofsky) suggests that these focal depths are within the crust, although whether the crust at these depths is anomalously cold, is not known. Chapman & Pollack (1977) report a heat flow of about 70 mW m^{-2}, which is higher than normal, near the epicentre of 15 May 1968.

There is currently a controversy over the geometry of extension in the crust since nearly all large normal faulting earthquakes occur on planes dipping in the range 30–60°, yet normal faults with much shallower dips, in the range 5–10°, are seen at the surface in some regions than are no longer actively extending (Jackson 1986). Because of the short thermal time constant for the upper crust this controversy is not relevant here unless the very low-angle normal faulting extends right through the lithosphere, as proposed by Wernicke (1985) and shown diagrammatically in figure 4 (*b*). This represents a modification of the uniform extension model of McKenzie (1978*b*), in which the crust and mantle part of the lithosphere are thinned in the same place (figure 4*a*). The geometry of figure 4*b*, in its most extreme form, leads to a lateral separation between the stretched crust and the stretched mantle that greatly affects the vertical movements in the region. In figure 4*b* the stretched crust sinks only during the stretching, and the subsidence caused by thermal relaxation after stretching has ceased occurs in a different place, above the stretched mantle lithosphere in a region that experiences no faulting at the surface. Thus the evidence for such a geometry must be sought in the horizontal distribution and the amount of subsidence relative to some datum (e.g. sea level), both before and after faulting has ceased. At present, the only reported data able to test such a geometry are from the North Sea, where Barton & Wood (1984) show that the amount of syn- and post-faulting subsidence in a series of wells across the Central Graben agree with the crustal thinning seen in a seismic refraction line. These data suggest that, on a length scale comparable with the lithosphere thickness, there is no spatial separation between crustal and mantle thinning. Thus, at least in the North Sea there is no reason to depart from the simple uniform lithospheric stretching model proposed by McKenzie (1978*b*).

The thermal consequences of uniform lithospheric stretching are discussed by Jarvis &

McKenzie (1980) and McKenzie (1981). If stretching is achieved in a time short compared to $60/\beta^2$ Ma (where β is the ratio of surface area after stretching to that before stretching) then the process is essentially isothermal; although the geothermal gradient increases each material particle remains at the same temperature. Thereafter, if the basin is not filled with sediments, cooling occurs at all depths (figure 5*a*). The situation is very different if the basin is filled with sediments of low conductivity. Under these circumstances material in the basement may increase in temperature for a period after stretching has ceased (figure 5*b*). Although this increase in temperature is not great, the temperature–time history of any particle of rock is greatly changed, and the temperature difference between particles initially at the same depth in figure 5*a*, *b* can be considerable.

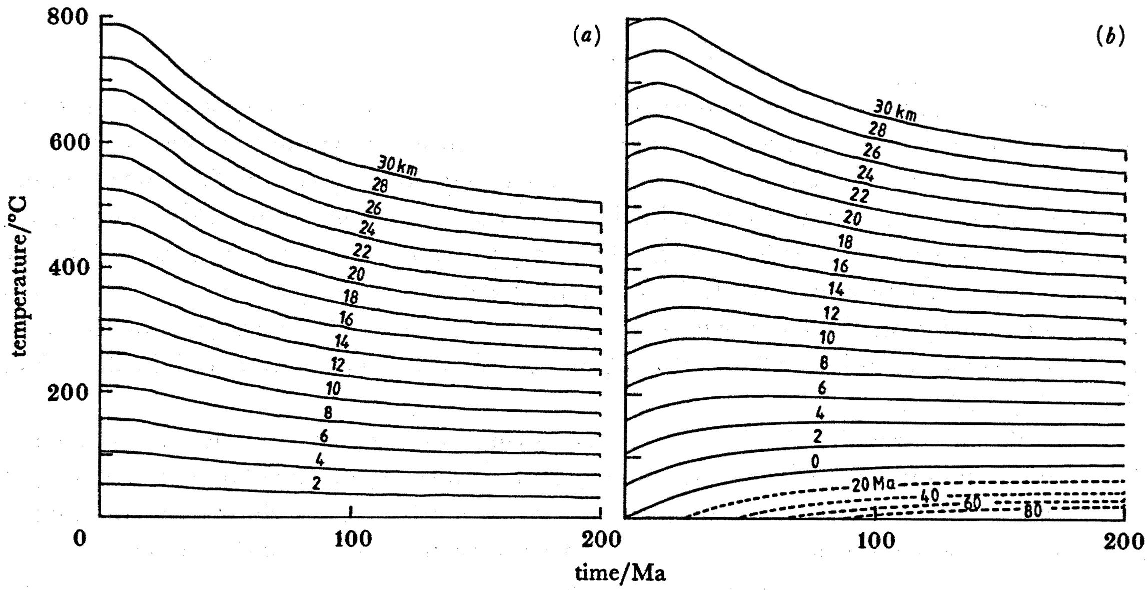

FIGURE 5. Temperature as a function of time for particles of rock at depths of 0–30 km (at 2 km intervals) after instantaneous stretching by a factor (β) of 2 at time $t = 0$. For example, the curve for 10 km is that for a particle initially at 20 km, which is brought to a depth of 10 km instantaneously at $t = 0$. In (*a*), curves are shown for the case in which no sediment is deposited in the subsiding basin. In (*b*), the basin is always filled to sea level with sediments and curves are also shown for sediment deposited at sea level at times of 20, 40, 60 and 80 Ma after extension. In both (*a*) and (*b*), the downward ticks at the end of the curves represent the asymptotic values at $t = \infty$. The curves were generated using the full expressions in McKenzie (1978*b*, 1981) with a uniform conductivity of 0.008 cal K^{-1} cm^{-1} s^{-1}, and with no lithospheric radioactive contribution to surface heat flux.

Because of the strong influence of temperature on strength, figure 5 suggests that the rheological behaviour of the basement in regions of extensional tectonics will be largely controlled by the presence or absence of sedimentary cover. Thus, metamorphism involving increased temperatures in regions of extension (Wickham & Oxburgh 1985) is likely to be mild, and to depend more on the sedimentation than on the extension itself; if the basin is starved of sediments each particle of rock becomes colder even though the thermal gradient increases. Under continental margins or intracontinental basins such as the North Sea, where the basement is blanketed by thick sediment, mild regional metamorphism involving increased temperatures is more likely. If large volumes of magma are generated by the extension process, the situation is clearly different. Under these circumstances very large increases in temperature may be experienced, but only for a short time, as the thermal anomaly associated with a few kilometres of magma will decay with a time constant of less than a million years.

Finally, it is clear from studies of recent and active normal faulting that large scale normal faults are capable of rapidly bringing material from mid-crustal depths to the surface, although the geometry by which they do so is not well understood (Davis 1983; Spencer 1984; Parry & Bruhn 1986; Spencer & Welty 1986). Normal faulting is also common in regions like Tibet whose high elevation is probably limited by their material strength (Molnar & Tapponnier 1978; England & McKenzie 1982; England & Houseman 1986; Houseman & England 1986). Thus the speed with which mid-crustal metamorphosed rocks are raised to the surface in orogenic belts may be influenced as much by normal faulting as by erosion (England & Richardson 1977).

5. Conclusions

In regions of active continental shortening or extension most seismicity is confined to the upper 15–20 km of the crust. Some earthquakes occur in the uppermost mantle beneath the deforming upper crust but these are rarer, smaller and do not account for much motion. However, they are important as probable indicators of a relatively strong uppermost mantle beneath a relatively weak lower crust, which is predominantly aseismic. Where data are available, the maximum focal depth in a region appears to be correlated to the surface heat flow or estimated geothermal gradient. Earthquakes in continental crust are generally confined to material colder than about 350 ± 100 °C, and those in the mantle to material colder than about 700 ± 100 °C.

Thus, although earthquakes are the most obvious manifestation of active continental deformation, they are largely confined to the upper crust. At length scales large compared with the thickness of this brittle layer, continental deformation is best described by a continuum, rather than by a plate tectonic, approach. As a consequence, discontinuous motion on faults in the thin, brittle, upper crust is seen as a response to the continuous deformation of the much thicker, ductile lithosphere below. The distribution of seismicity, elevation contrasts and vertical movements at the surface suggest that, on a length scale larger than the lithospheric thickness, there is no spatial separation between the brittle surface deformation and the ductile deformation at depth. For this reason, and because of the short thermal time constant of the upper crust, long-wavelength perturbations to the geothermal gradient are less sensitive to the precise geometry of shortening or extension in the crust than to the behaviour of the lithosphere as a whole.

The most likely sites of regional metamorphism at large length scales are in regions of shortening or extension. Where large-scale thickening occurs, detachment of the lower part of the cold, dense lithosphere can lead to the re-establishment of a near-initial geothermal gradient in thickened crust, with resultant widespread melting of the lower crust. In such environments, regional metamorphism is a consequence of crustal thickening rather than increased geothermal gradients, which are never likely to be much greater than they were initially, except when perturbed by the movement of melts. By contrast, in extensional environments lithospheric thinning will lead to an increase in geothermal gradient but, in the absence of a sedimentary cover or magmatism, each material particle actually becomes colder and no high temperature metamorphism should occur. However, where there is a thick blanket of low-conductivity sediments, small increases in temperature may be experienced by the basement beneath continental extensional basins and passive margins, and temperatures are likely to reach a peak after stretching has ceased. Although the magnitude of such temperature increases is small, they may have a significant effect on the rheology of the crust and upper mantle under extending regions. Thus, mild high-temperature regional metamorphism in extensional environments is controlled more by the sedimentary history than by the extension itself. The differential vertical movements during the normal faulting that accompanies the post- or syn-orogenic extension in regions of large-scale crustal thickening, such as Tibet, may play an important role in rapidly bringing mid-crustal metamorphosed rocks to the surface, thereby accelerating a process usually attributed solely to erosion. Such normal faulting will also disrupt the patterns of metamorphism established during thickening.

It is a pleasure to thank W.-P. Chen, P. England, D. P. McKenzie, F.R.S., and P. Molnar for many discussions that have helped shape my views of continental tectonics. I am grateful to them, and to the others listed in the references, for the careful observations and experiments, without which a review of this sort could not be written. R. Bruhn, W.-P. Chen, C. Kissel, J. Nabelek, G. Shudofsky and J. Strehlau kindly made available preprints of their work prior to publication, and Dan McKenzie helped produce figure 5. I thank P. England and E. R. Oxburgh, F.R.S., for detailed and helpful reviews of this paper. This paper is contribution No. 743 of Cambridge University Earth Sciences Department, and was supported by a grant from the Natural Environment Research Council.

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Discussion

E. H. Rutter (*Geology Department, Imperial College, London*). In the early part of his contribution, Dr Jackson drew attention to the widely held view that the truncation at about 15 km depth of the upper crustal seismogenic layer is attributable to a change in the predominant mode of rock failure from cataclastic faulting to distributed plastic flow. The perpetuation through force of repetition of this possibly restricted view leads to confusion between a fact of observation (in this case the existence of a depth limit to upper crustal seismicity) and its interpretation (the idea that the cut-off is due to a particular deformation mechanism change). While this particular interpretation is a very reasonable one, and may well be correct in many cases, we should remember that it is based on some theoretical extrapolations from a very limited number of experimental results, and that rocks can display a wide range of mode-of-failure transitions than simply this one, all of which can lead to a depth limit to seismicity.

Many rocks in the laboratory show a régime of distributed, work-hardening cataclastic (brittle and involving dilatancy) flow intermediate between cataclastic faulting and distributed plastic flow. Alternatively, seismicity may be suppressed by a transition to creep by a range of diffusive mass-transfer processes in fault zones. There are also possible combined processes involving subcritical brittle cracking and partial accommodation of dilatancy by diffusive mass transfer that have hardly begun to be explored.

It is widely assumed that the lower crust will be pervasively ductile by intracrystalline plasticity. It is alternatively possible that most lower crustal strain is accommodated in highly localized zones, which may display a wide range of types of rheological behaviour depending on the mineralogy and microstructural state of the rocks in such zones. Cataclasis may even be important in the lower crust if *ca.* 1% (by volume) free water in flat-lying planar zones really is responsible for the anomalous seismic reflection and electrical conduction properties that have been observed in some surveys.

At our present state of knowledge regarding all the possible mechanical properties of rocks in the mid- and lower continental crust, it is surely unwise to overemphasize one particular interpretation of the depth limit of upper-crustal seismicity, despite its obvious attractiveness.

J. A. Jackson. I agree with the spirit of Dr Rutter's remarks, but would make the following comments. A depth limit to upper crustal seismicity is widely observed. Faulting in the largest earthquakes appears to nucleate near the base of this seismogenic layer and to break to the surface. There is thus no doubt that within the seismogenic upper crust substantial strain is accommodated discontinuously on large faults. It is most unlikely that strain is accommodated in the same discontinuous fashion within the convecting asthenosphere. Therefore, at some depth below the nucleation depth of large earthquakes (perhaps within the lower crust, perhaps within the mantle lithosphere) discontinuous strain on faults or shear zones passes into continuously distributed strain. The question is: at what depth does this happen? This is an important question because the depth to which localized discontinuous strain occurs will control the lengthscale over which associated vertical motions are observed at the surface. Certainly we believe, for a variety of reasons mentioned in this paper and in Jackson (1986), that localized discontinuous strain on shear zones does occur in the uppermost part of the non-seismogenic lower crust, particularly at high strain rates. But we do not, in my opinion, know to what depth such localized strain continues, nor whether it continues throughout the lower crust.

The lack of earthquakes, but not of deformation, below a particular depth does indeed suggest a change to a mechanism of deformation different from that of simple brittle behaviour. Furthermore, the correlation of this limit to the depth of seismicity with regional heat flow and predicted temperature, particularly in the oceans, suggests that some temperature-dependence of strength, of the type suggested by the flow law in equation (2), is appropriate below this depth limit. I agree with Dr Rutter that the precise mechanism of deformation at such depths is uncertain and will definitely affect the values of Q, n and A in (2). In common with Chen & Molnar (1983), I yield to such uncertainty by not putting scales on figure 2 and by not using extrapolations of laboratory experiments to estimate actual values of stress in the crust. Figure 2 is intended to be used qualitatively to develop a feeling for the likely effect of *changes* in temperature gradient and strain rate on the depth limit of seismicity. Failure to limit its use to such purposes is clearly unwise, but it is probably equally unwise to take the view that real rock mechanics is so complicated that we can do nothing at all.

E. V. Artyushkov (*Institute of Physics of the Earth, Moscow, U.S.S.R.*). In this paper, high seismicity within the mountain belts (Caucasus and others) is considered as resulting from the crustal shortening now developing in these regions. In most present mountain belts an intense shortening, however, terminated long ago, while the present high relief was formed quite recently. For instance, an intense compression terminated 17 Ma ago in the Alps, *ca.* 30 Ma ago in the Dinarides and 11 Ma ago in the Carpathians. The now-existing mountain ranges in these regions began to form 3 Ma ago, and occurred without significant compression.

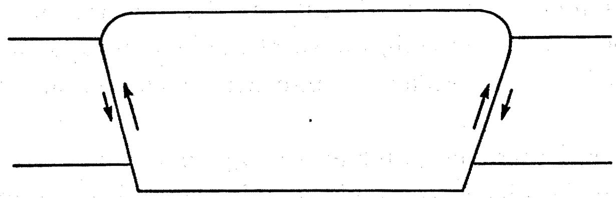

Figure D1

This indicates that crustal shortening and mountain building represent independent phenomena. The former process results from collision of lithospheric plates and commonly stops after the crustal surface emerges by several hundred metres above sea level. High mountains commonly form by vertical crustal movements considerably later. In some cases, the time gap between the end of shortening and the formation of mountains in the same region can be quite large. For example, in the Southern Alps compression terminated 85 Ma ago, while the first mountains of a moderate height originated 28–25 Ma ago. In the Northern Tien-Shan folding terminated 220 Ma ago. A considerable crustal uplift began *ca.* 30 Ma ago and high mountains were formed only *ca.* 3–5 Ma ago. Shortening of the crust and folding are now proceeding in some sedimentary basins near the mountain ranges. For example, this occurs in the Kura Depression between the Great and Lesser Caucasus. No intense crustal shortening, however, takes place in the Caucasus itself.

In the above mountain regions, most large earthquakes occur on large thrust faults that separate highly elevated regions from the adjacent lowlands (figure D1), or strike-slip faults that cross the ranges. For instance, the Tien-Shan is bounded by normal faults with a dip angle of *ca.* 60–70° on the north and south. A relative displacement along these faults between the elevating mountains and the adjacent subsiding basins is of *ca.* 5–7 km. Large earthquakes with a magnitude $M \approx 7.0$–7.5 (sometimes $M \approx 8$) occur on these faults. At the same time, highly elevated regions (*ca.* 4 km) are characterized by much lower seismicity. This is because the crustal blocks now uplift vertically without significant compressive deformations within the blocks.

J. A. Jackson. Professor Artyushkov's comments ignore both the distribution of the large earthquakes and also their focal mechanisms. The fault plane solutions in Tapponnier & Molnar (1979), Molnar & Chen (1982) and Jackson & McKenzie (1984) show that very few of the steep nodal planes in the fault plane solutions of thrust or reverse faulting earthquakes in the Caucasus or Tien-Shan are, in fact vertical, and that most dip at 70° or less. This observation is well constrained by the data as any steep nodal plane passes near the centre of the focal sphere, which is the part best sampled by teleseismic first motion observations. Thus

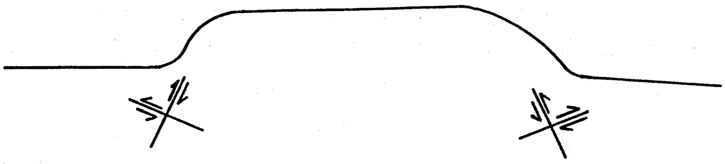

FIGURE D2

even if the steep nodal plane was the fault plane, some shortening would result: the faults are *not* vertical.

Secondly, the location of the large earthquakes in the Caucasus and the Tien-Shan is such that the shallow dipping nodal plane dips towards the highest topography, not away from it. Thus figure D1 in Professor Artyushkov's comments is incorrect, and should be drawn as in figure D2.

In these circumstances Tapponnier, Molnar, Chen, Jackson and McKenzie (not surprisingly) favour the low-angle nodal planes as the likely fault planes, because motion on them is able to produce the high topography. Motion on the steep nodal planes will have the reverse effect.

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