Pageoph, Vol. 115 (1977), Birkhäuser Verlag, Basel

Earthquake Stress Drops, Ambient Tectonic Stresses and Stresses That Drive Plate Motions

By THOMAS C. HANKS[1])

Abstract – A variety of geophysical observations suggests that the upper portion of the lithosphere, herein referred to as the elastic plate, has long-term material properties and frictional strength significantly greater than the lower lithosphere. If the average frictional stress along the non-ridge margin of the elastic plate is of the order of a kilobar, as suggested by the many observations of the frictional strength of rocks at mid-crustal conditions of pressure and temperature, the only viable mechanism for driving the motion of the elastic plate is a basal shear stress of several tens of bars. Kilobars of tectonic stress are then an ambient, steady condition of the earth's crust and uppermost mantle. The approximate equality of the basal shear stress and the average crustal earthquake stress drop, the localization of strain release for major plate margin earthquakes, and the rough equivalence of plate margin slip rates and gross plate motion rates suggest that the stress drops of major plate margin earthquakes are controlled by the elastic release of the basal shear stress in the vicinity of the plate margin, despite the existence of kilobars of tectonic stress existing across vertical planes parallel to the plate margin. If the stress differences available to be released at the time of faulting are distributed in a random, white fashion with a mean-square value determined by the average earthquake stress drop, the frequency of occurrence of constant stress drop earthquakes will be proportional to reciprocal faulting area, in accordance with empirically known frequency of occurrence statistics.

Key words: Earthquake stress drops; Plate tectonics; Stress in lithosphere.

1. Introduction

Since CHINNERY [1964] first concluded that 'the true stress relieved by [crustal earthquakes] is probably of the order of 10^7 dynes/cm^2' and noted that these stress drops were considerably less than the strength of the earth's crust and uppermost mantle as inferred by JEFFREYS [1959], the relationship of crustal earthquake stress drops to the ambient tectonic stress field giving rise to the earthquake in the first place has been a recurring theme in both observational and theoretical studies of the earthquake mechanism and in controlled laboratory experiments of rock failure. The advent of plate tectonics has only heightened geophysical interest in earthquake stress drops and ambient tectonic stresses, for it is generally agreed that the stress field that drives the motion of plates at the earth's surface should be related to earthquake stress drops along plate margins, the ambient tectonic stress across plate margins, or both. There is little agreement, however, on what such a relationship should be, chiefly because there is little agreement on the magnitude of tectonic stresses across

[1]) U.S. Geological Survey, Menlo Park, California 94025, USA.

plate margins or on the nature of the stress field driving plate motions across the earth's surface.

One point of view, first espoused by CHINNERY [1964], is that active crustal faults are throughgoing zones of weakness, with frictional or breaking strengths two orders of magnitude less than that for intact crustal rocks. Thus, while kilobars of deviatoric stress may exist in the crust beneath certain mountainous regions and perhaps locally on active crustal fault zones as well, the average tectonic stress in the earth's crust and uppermost mantle is probably not greater than 100 bars. This idea has been reinforced by BRUNE *et al.* [1969], who argued on the basis of the absence of a heat-flow anomaly localized to the San Andreas fault that the average frictional stress along the San Andreas fault could not exceed several hundred bars. In this framework, earthquake stress drops represent by and large a complete drop of the causative tectonic stresses, at least in the source region.

Laboratory experiments of rock failure have not, however, validated the fundamental proposition of this line of reasoning that crustal fault zones have strengths two orders of magnitude less than intact rocks. Across a wide range of rock types and surface preparations, kilobars of deviatoric stress are required to offset differentially rock masses subjected to mid-crustal conditions of pressure and temperature. Thus, if crustal faulting is governed by frictional processes, as it is generally agreed, kilobars of tectonic stress are required to induce fracture or frictional sliding of crustal materials through the seismogenic zone, unless fluid pressures essentially negate the lithostatic pressure.

The consequences of frictional stresses of kilobars on active crustal fault zones are far-reaching. It follows in a straightforward way that if kilobars of frictional stress resist plate motions across the upper several tens of kilometers of the plate margin, the only viable mechanism for driving plate motions at the earth's surface is a basal shear stress driving the motion of the plate in a direction parallel to itself, in a manner similar to the ductile model of LACHENBRUCH and SASS [1973]. Furthermore, this condition implies that tectonic stresses of kilobars across vertical planes parallel to the plate margin are an ambient, steady condition of the earth's crust and uppermost mantle. Less clear, however, is what relationship earthquake stress drops of tens of bars have to tectonic stresses of kilobars.

There are, nevertheless, several characteristics of earthquake stress drops in particular and faulting at plate margins in general that point to one such relationship. The first of these is the constancy of earthquake stress drops, in their average value, across the entire range of earthquakes for which instrumental recordings are available. More specifically, earthquakes along plate margins with fault lengths both much greater and much less than the seismogenic depth of approximately 15 km not only possess stress drops independent of source dimension but share the same average constant value of several tens of bars. A second characteristic is that the stress release accompanying even the largest earthquakes along plate margins is distinctly localized. It is well known from conventional geodetic observations that faulting displacements decay rapidly away from the fault surface, and both theory and observation are in accord that this decay is controlled by the depth or width of the fault. A third observation, probably related to the second, is that a variety of geophysical observations suggests that the frictional resistance to plate motions along their margins is not uniform across a 100 km thickness of the lithosphere but is probably concentrated in the upper several tens of kilometers.

The purpose of this study is to bring these observations, together with the proposition that kilobars of frictional stress exist on active crustal fault zones, into a framework which jointly relates them all, and several other matters of geophysical interest as well. Because it is not a consensus view of the geophysical community that an average frictional stress of kilobars exists on active crustal fault zones (e.g., CHINNERY 1964, BRUNE *et al.* 1969, FORSYTH and UYEDA 1975, CARTER 1976), the evidence for and against his proposition is presented. It is the point of view of this paper that the evidence supporting this proposition is, at the present time, persuasive although not irresistible. As such, we summarize the principal conclusions of this study in juxtaposition to the parallel consequences of the alternate (low-frictional stress) hypothesis at the end of this paper.

2. *Earthquake stress drops*

Figure 1 presents determinations of seismic moment (M_0), source dimension (r), and stress drop ($\Delta\sigma$) for 390 earthquakes, principally crustal earthquakes of the Southern California Region. These include 167 San Fernando aftershocks [TUCKER and BRUNE 1973], 138 southern California and northern Mexico earthquakes [THATCHER and HANKS 1973], 28 earthquakes in northern Baja California and the northern Gulf of California [THATCHER 1972], the 1968 Borrego Mountain [HANKS and WYSS 1972], the 1969 Coyote Mountain [THATCHER and HAMILTON 1973], the 1971 San Fernando [WYSS and HANKS 1972] and 1973 Pt. Mugu [ELLSWORTH *et al.* 1973] earthquakes. In addition, Fig. 1 includes 34 shallow focus earthquakes along the Tonga–Kermadec island arc [MOLNAR and WYSS 1972] and 19 intermediate and deep focus earthquakes in the Benioff zone beneath and behind the Tonga–Kermadec island arc [WYSS and MOLNAR 1972].

The distribution of stress drops, which is approximately logarithmically normal about the average value near 10 bars with two standard deviations corresponding to about an order of magnitude in stress drop, is generally but not completely independent of source strength. There are no stress drops less than 1 bar for earthquakes for which $M_0 \geq 10^{24}$ dyne-cm. THATCHER and HANKS [1973] suggested that r was likely to be overestimated near the lower range of M_0 for the earthquakes they studied (note the lower limit of $r \simeq 0.5$ km across two orders of magnitude in M_0 for their data in Fig. 1). A similar bias affects the stress drops of the smaller M_0 earthquakes considered by THATCHER [1972] and perhaps those of TUCKER and BRUNE

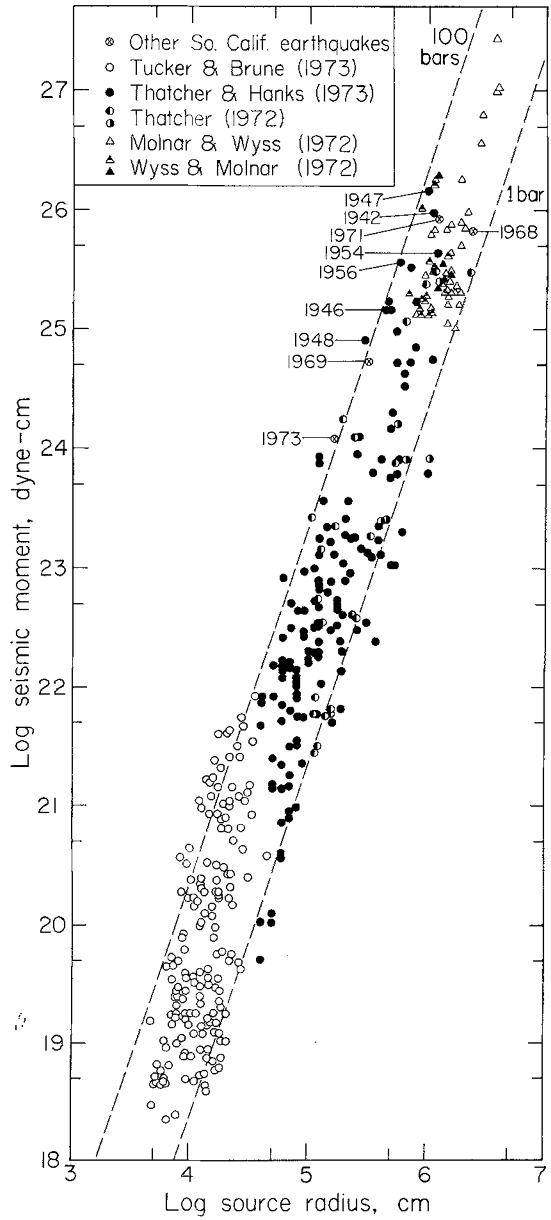

Figure 1

Earthquake stress drops. Open circles: San Fernando aftershocks; solid circles: earthquakes of the Southern California Region; half-filled circles: earthquakes in Baja California (left side filled) and in the Gulf of California (right side filled); triangles: earthquakes along the Tonga-Kermadec Island arc at shallow (open), intermediate (top filled), and deep (solid) depths. The dashed lines of slope 3 are lines of constant stress drop (1 bar and 100 bars as indicated).

[1973] as well. With allowance for these systematic effects, the average stress drop for the data in Fig. 1 would be slightly higher, perhaps 20–30 bars, with a corresponding reduction in the logarithmic standard deviation.

Because all of the data in Fig. 1 have been obtained by scaling spectral parameters of far-field body waves with the source model of BRUNE [1970, 1971], it is important to compare them with similar data for which source parameter estimates are available from geodetic data, field observations of fault length and offset, and aftershock distributions. With respect to the average earthquake stress drop, Fig. 1, quite remarkably, does little to change CHINNERY's [1964] conclusion, reproduced in the first sentence of this paper, reached on the basis of geodetic observations of faulting displacements for five major crustal earthquakes. Similarly, KANAMORI and ANDERSON [1975] (see also AKI 1972) have presented source parameter data for 41 moderate to great earthquakes ($8.3 \times 10^{24} \leq M_0 \leq 2 \times 10^{30}$ dyne-cm) for which source dimension data are principally determined by fault length observations and aftershock distributions. The resulting stress drops are similar to those for the larger earthquakes of Fig. 1, although KANAMORI and ANDERSON [1975] find that most of the values are between 10 and 100 bars.

It is worth noting that each of the stress drops in Fig. 1 and in KANAMORI and ANDERSON [1975] represents a value averaged over the entire faulting surface. They do not preclude the existence of larger, perhaps transient, stress differences across localized regions of the fault surface, as discussed for the San Fernando earthquake by HANKS [1974], the Borrego Mountain earthquake by BURDICK and MELLMAN [1976] and for a recent interpretation of peak accelerations [HANKS and JOHNSON, 1976]. Neither do these data preclude regional variations of stress drops within the observed range nor magnitude-dependent stress drops within a limited range of M_0 and r. But none of these considerations alter the principal conclusion to be extracted from the large number of observations presently available that crustal earthquake stress drops, in their average value $\overline{\Delta\sigma}$, are several tens of bars and that this average value is independent of source strength across twelve orders of magnitude in M_0.

It is curious that this is so, for the stress release of crustal earthquakes for which r is significantly less than the depth or width of faulting h is quite plainly a three-dimensional problem, whereas for earthquakes for which the fault length is significantly greater than h the stress release can be modelled well by two-dimensional geometry along most of the fault length [KASAHARA 1957, BYERLY and DENOYER 1958, CHINNERY 1961, THATCHER 1975, among many such studies]. In the following two sections, we explore the implications of this coincidence separately for earthquakes of small and large faulting dimensions.

3. *Stress drops of small dimension earthquakes*

An interesting way to explore the geophysical significance of the stress drops of small dimension earthquakes is in terms of the frequency of occurrence of constant stress drop events. The relations between the frequency of occurrence N of earthquake magnitude M

$$\log N = a - bM, \tag{1}$$

between M_0 and M

$$\log M_0 = cM + d, \tag{2}$$

and between M_0, r, and $\Delta\sigma$

$$M_0 = k\Delta\sigma r^3 \tag{3}$$

can be algebraically combined to obtain

$$\log N = \left(a + \frac{bd}{c}\right) - \frac{b}{c}\log(k\Delta\sigma r^3). \tag{4}$$

In these relations, a is a constant defined by the choice of region and time interval in which earthquakes are counted; d is an empirically determined constant, and k is a constant equal to 16/7 for a circular fault surface of radius r. It is empirically known that b is generally, but not always, very nearly equal to 1, irrespective of the choice of region and time interval in which earthquakes are counted. Also, c is empirically known to be 1.5 whether local magnitude [THATCHER and HANKS 1973] or surface wave magnitude [KANAMORI and ANDERSON 1975] is used in (2).

With $b = 1$ and $c = 1.5$, (4) reduces to

$$N = \frac{\text{const}}{(\Delta\sigma)^{2/3}r^2}. \tag{5}$$

If the earthquakes of the counted sample share the same $\Delta\sigma$, as they do on the average for all samples for which $\Delta\sigma$ has been determined, earthquake magnitude-frequency of occurrence statistics reduce to a simple matter of geometrical scaling in terms of the reciprocal faulting area.

This result can be interpreted in terms of a two-dimensional stress drop potential function on the fault surface, the functional form of which is specified by a mean-square value determined by $\overline{\Delta\sigma}$ and a spectral composition with constant amplitudes at all wavelengths $\leq h$; the stress drop potential in a region of incipient faulting is realized as the earthquake stress drop at the time of faulting. On the average, such a stress drop potential function will produce earthquakes with stress drop $\overline{\Delta\sigma}$, and their frequency of occurrence will scale as $1/r^2 (r \leq h)$, due to the constant spectral amplitudes at all wavelengths $\leq h$ in two dimensions; earthquakes with $\Delta\sigma$ both higher and lower than $\overline{\Delta\sigma}$ will occur, however, with certain probabilities. Whatever the origin of the stress differences recoverable in crustal faulting may be, then, it is more or less distributed as random, white noise on crustal fault zones. The stress drop potential function, moreover, retains its mean-square value and random, white characteristics until such time as the region of interest is faulted by a major earthquake, inasmuch as earthquakes with $r \ll h$ do not materially reduce the net stress on the fault surface.

4. *Stress drops of major plate margin earthquakes*

In the case of major plate margin earthquakes, a well-defined stress drop exists along most of the fault length through the seismogenic zone. Inasmuch as this represents a reduction of stresses opposing plate motions along the faulted portion of the plate margin, it must be balanced by a corresponding increase of resisting stresses elsewhere along the plate margin or a corresponding decrease in the stress-field driving the relative motion of the adjacent plates. There are, clearly, many options in placing this stress increment, but available geodetic observations of coseismic and postseismic displacements restrict the range of possibilities.

Geodetically measured surface displacements taken shortly (as much as years) after many major crustal earthquakes invariably reveal a rapid decay of earthquake-induced displacements away from the fault surface. When the fault length is much greater than the fault depth, such data are reasonably well-approximated by a two-dimensional elastostatic model of faulting [KASAHARA 1957, BYERLY and DENOYER 1958, CHINNERY 1961]. For an infinitely long, vertical fault in a uniform half space with constant offset U across $0 \leq z \leq h$ (Fig. 2a), *surface* displacements $u_y\,(x, z = 0)$ parallel to the fault are given by

$$u_y(x, z = 0) = \frac{U}{\pi} \tan^{-1}\left(\frac{h}{x}\right) \tag{6}$$

and the induced horizontal shear stress increment across any vertical plane $x =$ constant is

$$\sigma_{xy}(x, z = 0) = \mu \frac{U}{\pi h}\left(1 + \frac{x^2}{h^2}\right)^{-1} \tag{7}$$

where μ is the shear modulus.

The changes in σ_{xy} are thus strongly localized to the neighborhood of the fault, and the horizontal gradient of σ_{xy} is even more so. It requires little further analysis to conclude that $\partial\sigma_{xy}/\partial x$ can only be accompanied by corresponding gradients of stresses in the vertical direction across horizontal planes $z =$ constant. This requirement is, according to the equations of equilibrium in this two-dimensional problem,

$$\frac{\partial\sigma_{xy}}{\partial x} + \frac{\partial\sigma_{zy}}{\partial z} = 0. \tag{8}$$

In the geometry of Fig. 2b, we approximate $\partial\sigma_{xy}/\partial x$ in the region $0 \leq x \leq h$, $0 \leq z \leq h$ with

$$\overline{\frac{\partial\sigma_{xy}}{\partial x}} = -\frac{\Delta\sigma}{2h} \tag{9}$$

by taking the difference in σ_{xy} at $x = 0$ and $x = h$ along $z = 0$. In the same region, and again in an average sense without regard to localized stress changes induced by irregular faulting displacements, $\overline{\partial\sigma_{zy}/\partial z}$ must be a value comparable to (9). That is, the average stress difference ($\Delta\sigma$) imposed on the fault ($x = 0$, $0 \leq z \leq h$) in the course of the earthquake is accompanied by a stress change of comparable magnitude ($\frac{1}{2}\Delta\sigma$) and the same sign along $z = h$, $0 \leq x \leq h$.

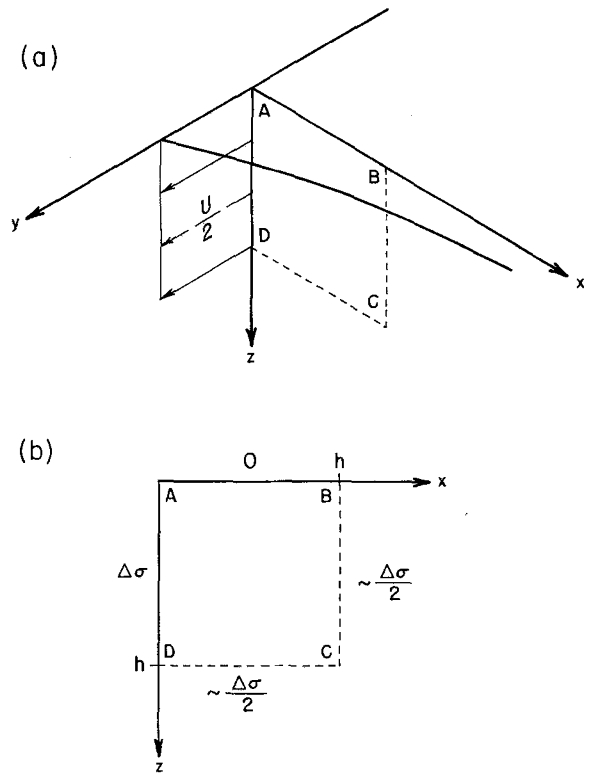

Figure 2
(a) Geometry of throughgoing faulting at a transform plate margin; the x-coordinate is into the plate interior, the y-coordinate is along the plate margin, and the z-coordinate measures depth. The curved line in the x–y plane is $u_y(x, z = 0) = U/\pi \tan^{-1}(h/x)$ which arises in a uniform elastic half space when a displacement $U/2$ is applied on (each side of) the fault through $0 \leq z \leq h$. (b) The approximate stress increments on the sides of ABCD in the x–z plane of (a), given that the faulting displacement $U/2$ through $0 \leq z \leq h$ corresponds to an average stress increment $\Delta\sigma$ through $0 \leq z \leq h$ on the fault plane ($x = 0$).

To what extent the stress increments along $x = 0$, $z \leq h$ and $z = h$, $x \leq h$ cause postseismic displacements at greater depths on the fault is largely unknown. THATCHER [1975] has presented evidence that postseismic displacements in the several decades following the 1906 San Francisco earthquake between 10 and 30 km were of the same order as the coseismic displacements for $z < 10$ km. While the conventional geodetic observations are not well disposed to resolving relaxation over greater dimensions, there is no evidence within these data that significant displacement occurred at distances greater than several tens of kilometers from the fault.

The point of interest here is that the stress release following major plate margin earthquakes is distinctly localized within several fault depths (widths) of the corner of the lithospheric plate. Because of this, the stress change along the fault to depth h (or perhaps several h including postseismic relaxation) is accompanied by a comparable stress change of the same sign along horizontal planes at $z = h$ (or several h) within $x = h$ (or several h) of the plate margin. An interesting consequence of all of this, and the principal issue to be pursued in the remainder of this paper, is that the stress drops of earthquakes along plate margins *may* be controlled by recoverable stress differences of several tens of bars that exist along horizontal planes at a depth of an h or so and within an h or so of the plate margin, whatever the stresses on planes parallel to the plate margin may be.

5. *Strain accumulation and the basal shear stress*

The process of strain accumulation leading to major earthquakes along plate margins is not well understood, but a process that provides a roughly equal and opposite localization of strain energy to the manner in which it is released seems reasonable. Such a model, again for an infinitely long transform fault, is described by SAVAGE and BURFORD [1973] and is reproduced in Fig. 3a. Below the depth h, there is a uniform displacement $U/2$, giving a relative offset U at the fault surface for $z > h$, but for $z \leq h$ the fault is locked. The surface displacements parallel to the fault are then

$$u_y(x, z = 0) = \frac{U}{\pi} \tan^{-1}\left(\frac{x}{h}\right) \tag{10}$$

Following throughgoing faulting, we imagine the block motion indicated in Fig. 3b to ensue, as the addition of the coseismic surface displacements (6) to (10) suggests.

In this model of strain accumulation, it is reasonable to associate a shear stress across horizontal planes at depth $\sim h$ with the displacements $U/2$ below h, in a manner analogous to the ductile model of LACHENBRUCH and SASS [1973]. This possibility is envisioned in the simplest possible way: a single driving force, a basal shear stress driving the overlying plate in a direction parallel to itself, is in equilibrium with a single resisting force, the frictional stresses along the plate margin (Fig. 4). For a circular plate of radius R and thickness H, the condition of static equilibrium becomes

$$\sigma_b = 2\sigma_f \frac{H}{R} \tag{11}$$

where σ_b is the basal shear stress acting at depth H and σ_f is the frictional resisting stress along the plate margin.

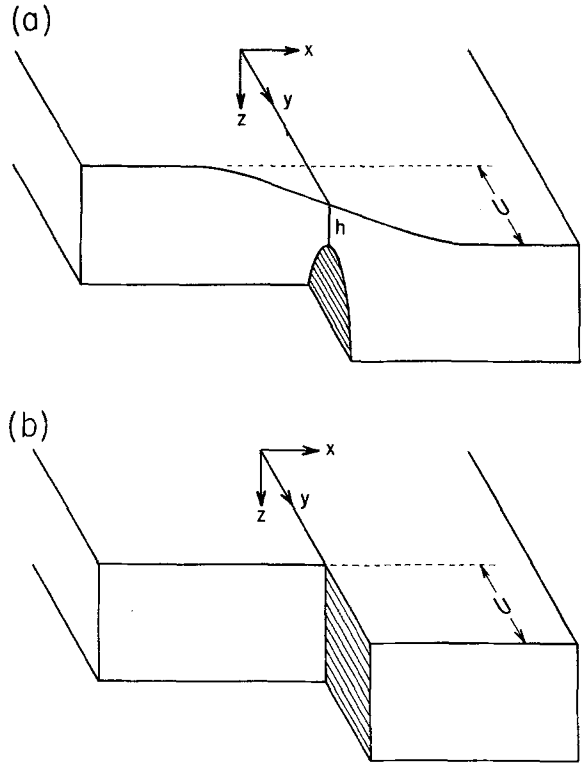

Figure 3
The SAVAGE and BURFORD (1973) model of strain accumulation along a transform plate margin; coordinate axes are the same as in Fig. 2. (a) Prior to throughgoing faulting, the seismogenic zone is essentially locked; strain accumulation in the vicinity of the plate margin arises from uniform, relative displacement U at depths greater than h. (b) After throughgoing faulting, slip on the seismogenic zone additive to aseismic displacements at greater depth results in an essentially undeformed block motion with relative offset U.

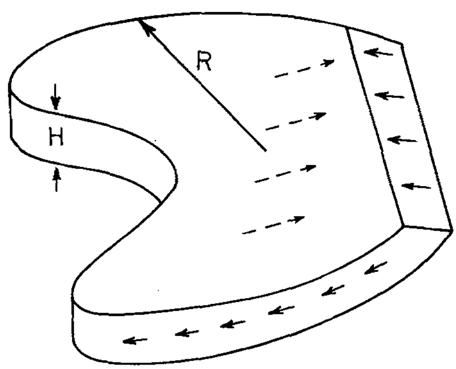

Figure 4
Schematic illustration of the mechanism driving the motion of the elastic plate: a single driving force, the shear stress acting at the base of the elastic plate (dashed arrows) is equilibrated by a single resisting force, the frictional stresses along the plate margin (solid arrows).

It is important to emphasize that the principal assumption of the model summarized in (11) lies not in the postulated existence and action of σ_b but in the values to be prescribed for σ_f, H, and R, particularly σ_f. This is simply because, if σ_f is of the order of a kilobar for H equal to several tens of kilometers or greater, the only viable mechanism for driving plate motions at the earth's surface is the basal shear stress, arising from thermal convection in the upper mantle. Neither the ridge push nor the net trench pull (for a discussion of the general classes of driving forces, see FORSYTH and UYEDA 1975) are sufficiently strong to offset resistance to plate motions of this magnitude (e.g., MCKENZIE 1969, ANDREWS 1972, SOLOMON *et al.* 1975, FORSYTH and UYEDA 1975).

We take this point of view here, postponing to the discussion a more complete statement of the evidence for and against the existence of kilobars of frictional stress on active crustal fault zones. On the basis of the many observations summarized in BRACE and BYERLEE [1970], BRACE [1972], STESKY and BRACE [1973], and ZOBACK and BYERLEE [1976], it seems likely that the maximum shear strength along active crustal fault zones is several kilobars near 15 km with a ductile strength decreasing below this depth due to increasing temperature and with a brittle strength decreasing above this depth due to decreasing confining pressure. The temperature dependence of the strength of rocks as well as the depth of earthquakes along plate margins suggest that the frictional resistance to plate motions along their margins is not uniform across 100 km thickness but is concentrated in the upper 20 to 30 km. These considerations motivate an important distinction made in this study between the *elastic plate thickness* of perhaps 20 to 30 km, at the base of which σ_b is presumed to act, and the lithospheric plate thickness, generally estimated to be 50 to 125 km on the basis of heat flow (e.g., MCKENZIE 1967, SCLATER and FRANCHETEAU 1970) and seismic (e.g., KANAMORI and PRESS 1970, FORSYTH 1977) studies. Additional support for the distinction of an elastic plate thickness considerably less than 100 km may be found in the model calculations of the deformation arising in the oceanic lithosphere offshore of many oceanic trenches [HANKS 1971, WATTS and TALWANI 1974].

With an average value of 1 kb for σ_f across $H = 30$ km along the plate margin and for $R = 2000$ km, $\sigma_b = 30$ bars. Since the chosen values of σ_f, H, and R are plainly uncertain to or variable over a factor of 2 or so, we conclude that σ_b is of the order of several tens of bars. We must, however, exclude those plates with R significantly less than 1000 km, and we do so on the grounds that their motions are more likely to be governed by forces transmitted through their boundaries rather than through their basal areas.

Depending on the scale of dimension and time, the basal shear stress plainly has two distinguishable effects. First, it induces a deformation field far from the plate margin that is unaffected by faulting at the margins and may well be steady on time scales of uniform plate motions, perhaps millions of years. This steady deformation, which is not shown in Fig. 3, is not measurable by any geodetic technique and gives rise to kilobars of tectonic stresses across vertical planes. In this case, it seem reasonable to associate σ_b with the creep strength of mantle materials at depths of several tens of kilometers, and several tens of bars is certainly a reasonable number [KIRBY 1977]. Additive to this steady deformation field, of course, is the simple translation (rotation on a spherical surface) of the plate determined by the plate velocity.

Within several h of the plate margin and on a time scale of minutes to years, σ_b plays a much different role: it is a recoverable, elastic stress available to be released at the time of faulting. In this context, the stress increment on the plate margin, to be released at the time of faulting, is equilibrated by σ_b within an h or so of the plate margin, also released at the time of faulting. If, at the time of faulting, the deformation deficit relative to $U/2$ away from the margin is fully recovered, the stress drop across horizontal planes within several h of the plate margin will be complete, and the earthquake stress drop will be approximately equal to σ_b. To assume that this deformation deficit is fully recovered is equivalent to assuming that, on time scales comparable to the repeat time of major plate margin earthquakes, all points in the elastic plate move the same amount. Given the approximate equality between slip rates at plate margins estimated from the cumulative seismic moments of earthquakes along them with gross plate motion velocities determined from magnetic anomalies [BRUNE 1968, DAVIES and BRUNE 1971], this is a reasonable enough assumption.

Thus, earthquake stress drops need not be controlled, and quite possibly are not controlled, by the shear stress existing across vertical planes parallel to the plate margins but instead are controlled by the tens of bars of basal shear stress across horizontal planes, which is released within an h or so of the plate margin at the time of throughgoing faulting. Before this occurs, we imagine the equilibrating stress increment on the elastic plate margin to be distributed in a random, white fashion, which leads to constant stress drop earthquakes at all dimensions $r \lesssim h$ for which the frequency of occurrence follows the geometric scaling described previously.

6. Discussion

In the preceding two sections, the stress drops of major plate margin earthquakes, the accumulation of strain energy leading to such earthquakes, and the connection between σ_b and $\overline{\Delta\sigma}$ have been discussed implicitly for the transform-type plate margin. The same ideas hold for a converging plate margin, although the problem is even simpler because the geometry is essentially one-dimensional. Figure 5 is a cross-section of the elastic plate in the vicinity of and normal to an oceanic trench. At the time of a major earthquake along the Benioff zone at shallow depth, the stress drop along the faulted region is simply balanced by a corresponding drop of σ_b to zero at the base of the elastic plate beneath the faulted region. The gross static equilibrium of the plate in this case, however, is maintained by a nearly horizontal compressive stress of several kilobars, acting normal to the trench axis and arising from frictional resistance to motion of the elastic plate along the Benioff zone at shallow depth.

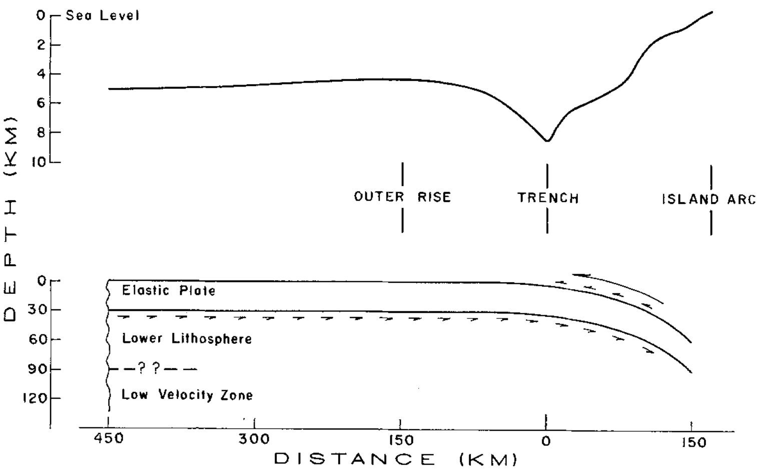

Figure 5
Cross-section of the oceanic lithosphere in the vicinity of and normal to a subducting plate margin. The top part of the figure displays typical bathymetry, at vertical exaggeration of 15:1. σ_b is denoted by small arrows at the bottom of the elastic plate, driving it to the right. On the upper surface of the subducted oceanic lithosphere between the trench and island arc (the crustal portion of the Benioff zone) the long arrow pointing to the left represents a frictional stress of kilobars, which maintains the gross static equilibrium of the plate and induces kilobars of nearly horizontal compression to deform the lithosphere offshore of the trench. The short arrows pointing to the left represent the potential $\Delta\sigma$ of a major earthquake in this region, which will be equilibrated by a drop of σ_b to zero at the base of the elastic plate in the same region.

Such a horizontal compressive stress has been inferred independently on the basis of the deformation of the oceanic lithosphere suggested by the bathymetry of the oceanic trench-outer rise systems by HANKS [1971] and WATTS and TALWANI [1974]. A bending moment applied onshore of the trench axis, however, provides much the same deflection as the horizontal load does (PARSON and MOLNAR 1976, CALDWELL *et al.* 1976). HANKS [1971] noted that horizontal compression is the more plausible loading mechanism in that it approximately negates the enormous tensional strains ($\sim 10^{-2}$) that develop on the surface of the oceanic lithosphere in the case of pure bending. Moreover, NAKAMURA and JACOB [1976] have found evidence in the alignment of flank eruptions on volcanoes along the Aleutian arc that the axis of maximum compression is parallel to the relative velocity vector of the converging Pacific plate. In any event, the gross difference between the predicted stress fields at the surface of the oceanic lithosphere deformed under these two loading conditions should allow them to be distinguished on the basis of hydrofracture measurements.

In either case, however, the deformed lithosphere in these regions must be capable of sustaining kilobars of deviatoric stress, if the upper several tens of kilometers behave in an elastic manner. If deformation in this region is principally anelastic, it must also be largely aseismic, for the oceanic lithosphere offshore of the trench is notably although not completely aseismic; major ($M \gtrsim 7$) earthquakes occur offshore of oceanic trenches only rarely. On the other hand, it seems to be of little value to invoke such an aseismic, anelastic mechanism for deformation of the oceanic lithosphere in its oldest, coldest, thickest and presumably most brittle condition known at the earth's surface when it is nevertheless capable of producing $M \simeq 7$ earthquakes at oceanic ridges in its youngest, hottest, thinnest and presumably most ductile condition. This matter returns us to the principal tenet of this study, that an average frictional stress of kilobars exists on active crustal fault zones and, by consequence, that tectonic stresses of kilobars are an ambient steady condition of the earth's crust and uppermost mantle. What further evidence is there for, and against, this proposition?

Certainly, this proposition is not a new one. JEFFREYS [1959] had concluded at an early date that 'stress differences of at least 1.5×10^9 dyne-cm^2 must exist within the outermost 50 km of the earth' in order to support high, uncompensated mountain ranges. CHINNERY [1964], explicitly assuming that crustal earthquake stress drops represented the breaking strength of crustal fault zones, was forced to conclude that crustal fault zones were throughgoing zones of weakness with breaking strengths approximately two orders of magnitude less than that for intact rock specimens. Laboratory experiments, however, are uniformly inconsistent with Chinnery's assumption. Across a wide range of rock types and surface preparations, kilobars of shear stress are required to offset differentially two rock masses at mid-crustal conditions of confining pressure and temperature (e.g., BRACE and BYERLEE 1970, BRACE 1972, STESKY and BRACE 1973, ZOBACK and BYERLEE 1976).

If internal fluid pressures are approximately equal to the confining pressures, however, the effective pressure and therefore the frictional strength are significantly reduced. The difference in density between rock and water, assuming water to be the pressurized fluid, requires the fluid pressure to be well above hydrostatic at all depths along the seismogenic zone, if this mechanism is to reduce frictional strength to a value significantly less than a kilobar. While superhydrostatic fluid pressures are known to occur in sedimentary basins, there is no evidence that this a prevailing condition across the seismogenic portion of active crustal fault zones, although our knowledge in this regard is hardly sufficient to rule this possibility out. In any event, holes drilled to even modest depth (~ 5 km) in active crustal fault zones should resolve this issue.

BRUNE *et al.* [1969] have argued that the absence of a heat flow anomaly localized to the San Andreas fault precluded the existence of frictional stresses along it in excess of several hundred bars. Their steady-state model, however, made no allowance for the heat required for transient temperature rises, the energy absorbed to create new fractures in fresh rock, or mechanisms of heat transport other than ordinary thermal conduction in the rock mass adjacent to the fault.

In view of the role water could play in reducing the frictional strength of active crustal fault zones, it is interesting to speculate on the role water could play in masking the heat flow anomaly estimated by BRUNE *et al.* [1969]. Although it is not localized to the fault zone, a substantial heat flow anomaly does exist in the Coast Ranges of central California (LACHENBRUCH and SASS 1973), but whether this excess heat flow is in fact due to the San Andreas fault, as opposed to residual thermal effects of late Cenozoic metamorphism, is unknown. The excess heat of this anomaly to ± 50 km away from the San Andreas fault, however, is enough to account for the heat generated by kilobars of frictional stress on the fault; the only problem is transporting this heat so far from the fault zone. A particularly efficient way of doing this would be to vaporize water in the fault zone with the frictional heat generated at the time of a major earthquake, convect it laterally away from the fault zone under the influence of thermally-induced pressure increases, and there release the heat carried in the latent heat of vaporization.

It is the point of view of this study that the argument of JEFFREYS [1959], the deformation of the lithosphere offshore of oceanic trenches, and especially the laboratory results of rock failure provide substantial evidence for the existence of ambient tectonic stresses of kilobars in the earth's crust and uppermost mantle and of frictional stresses of kilobars along active crustal faults; and that neither the possibility of nearly lithostatic fluid pressures to mid-crustal depths nor the absence of a heat flow anomaly localized to the San Andreas fault constitutes a compelling argument against these conditions. In view of the uncertainties associated with any line of evidence as well as the existence of contradictory possibilities and arguments, however, it is probably inappropriate to press this point too hard. As such, we summarize below the principal findings of this study, conditioned to the existence of frictional stresses of kilobars along active crustal fault zones, in juxtaposition with the parallel consequences of the alternate (low frictional stress) hypothesis.

7. *Summary and conclusions*

If kilobars of frictional stress resist plate motions along the upper several tens of kilometers of their non-ridge margins, it follows in a straightforward way that the only viable mechanism for driving the motion of the elastic plate is a basal shear stress of several tens of bars. Furthermore, tectonic stresses of kilobars are an ambient, steady condition of the earth's crust and uppermost mantle. The approximate equality of the basal shear stress and the average earthquake stress drop, the localization of strain release for major plate margin earthquakes, and the rough equivalence of plate margin slip rates and gross plate motion rates suggest that earthquake stress drops are governed by the elastic release of σ_b near the plate margin at the time of throughgoing faulting, despite the existence of kilobars of deviatoric stress existing across vertical planes parallel to the plate margin. Prior to throughgoing faulting, the stress increment $\overline{\Delta\sigma}$ is distributed in a random, white fashion along the elastic plate margin in order to provide the empirically known frequency of occurrence distribution of constant stress drop earthquakes.

σ_b must have its origin in some form of thermal convection beneath the elastic plate; its magnitude is several tens of bars, and its action is to drive the overlying plate in a direction parallel to itself. On a time scale of uniform plate motions, σ_b is presumably related to the creep strength of upper mantle materials at the appropriate pressure and temperature. On the much shorter time scales of throughgoing crustal faulting, however, σ_b is simply a recoverable elastic stress available to be released in the vicinity of the plate margin. On time scales between hundreds of seconds and hundreds of years, σ_b must be restored preparatory to the next major earthquake by some deformation mechanism largely confined to the neighborhood of the plate margin beneath the elastic plate. Whether or not σ_b in fact possesses these attributes is the major unresolved issue of this study, for little additional information is available with which this problem might be further constrained.

If the frictional stress on active crustal fault zones is of the order of 100 bars or less, there is no compelling reason to appeal to σ_b or to general thermal convection in the upper mantle as the principal mechanism for driving plate motions at the earth's surface. In this case, as SOLOMON *et al.* [1975] and FORSYTH and UYEDA [1975] have concluded, plate motions are probably governed by weak ridge pushes and trench pulls equilibrated by very weak basal drag forces. Ambient tectonic stresses will be comparable to the frictional stresses, and earthquake stress drops represent by and large a complete reduction of deviatoric stresses, at least in the source region. The most important consequence of this possibility, however, is that laboratory observations of rock failure at mid-crustal conditions of pressure and temperature have little if any relevance to the stress conditions governing crustal faulting, unless nearly lithostatic fluid pressures are a pervasive condition along crustal fault zones. The explanation of constant stress drop earthquakes for events with $r \lesssim h$ is unaffected by the choice of high or low frictional stress hypothesis.

The potential value of holes drilled in and near active crustal fault zones to depths of 5–10 km for resolving whether or not kilobars of frictional stress exist on active crustal fault zones cannot be overestimated. Measurements of the amount and fluid pressure of H_2O, the frictional strength of core samples, the permeability of the fault zone, the *in situ* stress, $\partial u_y/\partial z$, and the temperature field at these depths all bear directly on this central issue.

Acknowledgments

I have enjoyed the critical comments of many individuals in the development of the ideas presented in this manuscript, but those of D. J. Andrews, A. H. Lachenbruch, C. B. Raleigh, J. C. Savage, and W. R. Thatcher have been especially valuable.

REFERENCES

AKI, K. (1972), *Earthquake mechanism*, Tectonophysics *13*, 423–446.

ANDREWS, D. J. (1972), *Numerical simulation of sea-floor spreading*, J. Geophys. Res. *77*, 6470–6481.

BRACE, W. F. (1972), *Laboratory studies of stick-slip and their application to earthquakes*, Tectonophysics *14*, 189–200.

BRACE, W. F. and BYERLEE, J. D. (1970), *California earthquakes: why only shallow focus?* Science *168*, 1573–1575.

BRUNE, J. N. (1968), *Seismic moment, seismicity, and rate of slip along major fault zones*, J. Geophys. Res. *73*, 777–784.

BRUNE, J. N. (1970), *Tectonic stress and the spectra of seismic shear waves*, J. Geophys. Res. *75*, 4997–5009.

BRUNE, J. N. (1971), *Correction*, J. Geophys. Res. *76*, 5002.

BRUNE, J. N., HENYEY, T. L. and ROY, R. F. (1969), *Heat flow, stress, and rate of slip along the San Andreas fault, California*, J. Geophys. Res. *74*, 3821–3827.

BURDICK, L. J. and MELLMAN, G. R. (1976), *Inversion of the body waves from the Borrego Mountain earthquake to the source mechanism*, Bull. Seismol. Soc. Amer. *66*, 1485–1499.

BYERLY, P. and DE NOYER, J. *Energy in earthquakes as computed from geodetic observations* in *Contributions in Geophysics in Honor of Beno Gutenberg* (Pergamon Press, New York, 1958).

CALDWELL, J., HIXBY, W. F., KARIG, D. E. and TURCOTTE, D. L. (1976), *On the applicability of a universal elastic trench profile*, Earth Plan. Sci. Lett. *31*, 239–246.

CARTER, N. L. (1976), *Steady state flow of rocks*, Rev. Geophys. Space Phys. *14*, 301–360.

CHINNERY, M. A. (1961). *The deformation of the ground around surface faults*, Bull. Seismol. Soc. Amer. *51*, 355–372.

CHINNERY, M. A. (1964), *The strength of the Earth's crust under horizontal shear stress*, J. Geophys. Res. *69*, 2085–2089.

DAVIES, G. F. and BRUNE, J. N. (1971), *Regional and global fault slip rates from seismicity*, Nature *229*, 101–107.

ELLSWORTH, W. L., CAMPBELL, R. H., HILL, D. P., PAGE, R. A., ALEWINE, R. W., HANKS, T. C., HEATON, T. H., HILEMAN, J. A., KANAMORI, H., MINSTER, B. and WHITCOMB, J. H. (1973), *Point Mugu, California, earthquake of 21 February 1973 and its aftershocks*, Science *182*, 1127–1129.

FORSYTH, D. W. (1977), *The evolution of the upper mantle beneath mid-ocean ridges*, Tectonophysics, in press.

FORSYTH, D. and UYEDA, S. (1975), *On the relative importance of the driving forces of plate motion*, Geophys. J. Roy. Astron. Soc. *43*, 163–200.

HANKS, T. C. (1971), *The Kuril trench-Hokkaido rise system: large shallow earthquakes and simple models of deformation*, Geophys. J. Roy. Astron. Soc. *23*, 173–190.

HANKS, T. C. (1974), *The faulting mechanism of the San Fernando earthquake*, J. Geophys. Res. *79*, 1215–1229.

HANKS, T. C. and JOHNSON, D. A. (1976), *Geophysical assessment of peak accelerations*, Bull. Seismol. Soc. Amer. *66*, 959–968.

HANKS, T. C. and WYSS, M. (1972), *The use of body-wave spectra in the determination of seismic source parameters*, Bull. Seismol. Soc. Amer. *62*, 561–590.

JEFFREYS, H., *The Earth* (Cambridge University Press, Cambridge, 1959).

KANAMORI, H. and ANDERSON, D. L. (1975), *Theoretical basis of some empirical relations in seismology*, Bull. Seismol. Soc. Amer. *65*, 1073–1096.

KANAMORI, H. and PRESS, F. (1970), *How thick is the lithosphere?* Nature *226*, 330–331.

KASAHARA, K. (1957), *The nature of seismic origins as inferred from seismological and geodetic observations (1)*, Bull. Earthquake Res. Inst. Tokyo Univ. *35*, 475–532.

KIRBY, S. (1977), *State of stress in the lithosphere: inferences from the flow laws of olivine*, Pure appl. Phys. *115*, 245–258.

LACHENBRUCH, A. H. and SASS, J. H., *Thermo-mechanical aspects of the San Andreas fault system* in *Proceedings of the Conference on Tectonic Problems of the San Andreas fault system* (eds. R. L. Kovach and A. Nur) (Stanford University Publications, Stanford, 1973), pp. 192–205.

MCKENZIE, D. P. (1967), *Some remarks on heat flow and gravity anomalies*, J. Geophys. Res. *72*, 6261–6273.

MCKENZIE, D. P. (1969), *Speculations on the consequences and causes of plate motions*, Geophys. J. Roy. Astron. Soc. *18*, 1–32.

MOLNAR, P. and WYSS, M. (1972), *Moments, source dimensions and stress drops of shallow-focus earthquakes in the Tonga–Kermadec arc*, Phys. Earth Planet. Interiors *6*, 263–278.

NAKAMURA, K. and JACOB, K. (1976), *Volcanoes as indicators of tectonic stress orientation – Aleutians and Alaska* (abs.), EOS *57*, 1006.

PARSONS, B. and MOLNAR, P. (1976), *The origin of outer topographic rises associated with trenches*, Geophys. J. Roy. Astron. Soc. *45*, 707–712.

SAVAGE, J. C. and BURFORD, R. O. (1973), *Geodetic determination of relative plate motion in central California*, J. Geophys. Res. *78*, 832–845.

SCLATER, J. G. and FRANCHETEAU, J. (1970), *The implications of terrestrial heat flow observations on current tectonic and geochemical models of the crust and upper mantle of the earth*, Geophys. J. Roy. Astron. Soc. *20*, 509–542.

SOLOMON, S. C., SLEEP, N. H. and RICHARDSON, R. M. (1975), *On the forces driving plate tectonics: inferences from absolute plate velocities and intraplate stress*, Geophys. J. Roy. Astron. Soc. *42*, 769–801.

STESKY, R. M. and BRACE, W. F., *Estimation of frictional stress on the San Andreas fault from laboratory measurements* in *Proceedings of the Conference on Tectonic Problems of the San Andreas fault system* (eds. R. L. Kovach and A. Nur) (Stanford University Publications, Stanford, 1973), pp. 206–214.

THATCHER, W. (1972), *Regional variations of seismic source parameters in the northern Baja California area*, J. Geophys. Res. *77*, 1549–1565.

THATCHER, W. (1975), *Strain accumulation and release mechanism of the 1906 San Francisco earthquake*, J. Geophys. Res. *80*, 4862–4872.

THATCHER, W. and HAMILTON, R. M. (1973), *Aftershocks and source characteristics of the 1969 Coyote Mountain earthquake, San Jacinto fault zone, California*, Bull. Seismol. Soc. Amer. *63*, 647–661.

THATCHER, W. and HANKS, T. C. (1973), *Source parameters of southern California earthquakes*, J. Geophys. Res. *78*, 8547–8576.

TUCKER, B. E. and BRUNE, J. N., *Seismograms, S-wave spectra, and source parameters for aftershocks of San Fernando earthquake* in *San Fernando, California Earthquake of February 9, 1971*, Vol. III (Geological and Geophysical Studies, 69–122, U.S. Dept. of Commerce, 1973).

WATTS, A. B. and TALWANI, M. (1974), *Gravity anomalies seaward of deep-sea trenches and their tectonic implications*, Geophys. J. Roy. Astron. Soc. *36*, 57–90.

WYSS, M. and HANKS, T. C. (1972), *The source parameters of the San Fernando earthquake inferred from teleseismic body waves*, Bull. Seismol. Soc. Amer. *62*, 591–602.

WYSS, M. and MOLNAR, P. (1972), *Source parameters of intermediate and deep focus earthquakes in the Tonga arc*, Phys. Earth Planet. Interiors *6*, 279–292.

ZOBACK, M. D. and BYERLEE, J. D. (1976), *A note on the deformational behavior and permeability of crushed granite*, Int. J. Rock Mech. Min. Sci. Geomech. Abstr. *13*, 291–294.

(Received 14th January 1977)