Seismological constraints on stress in the oceanic lithosphere

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Studies of earthquakes provide some of the basic constraints on the stress within the oceanic lithosphere. The focal mechanisms of earthquakes indicate the directions of the principal stress axes. These directions can be compared with the stresses predicted by plate driving-force models, given the caveat that they can be biased by the locations of pre-existing weak zones and local processes. One basic result, that lithosphere older than about 35 Ma is in deviatoric compression, can constrain the basal drag on plates and hence mantle viscosity. A second result, the absence of a broad zone of ridge-normal extension in younger lithosphere, suggests that the ridge axis is rheologically weak. The focal depths of earthquakes, which indicate the depth to which the lithosphere is strong enough to support seismicity, can be combined with thermal models and laboratory rock mechanics to constrain the rheology of the lithosphere, both in plate interiors and at plate boundaries. An additional data type, which could in principle yield inferences about the magnitudes of stress, consists of the seismic moments, magnitudes, and source time functions of the earthquakes. Some initial results suggest that transform fault earthquakes release seismic energy more slowly than ridge events, perhaps due to lower stress drops or source geometry.

1. Introduction

For tectonic purposes, ideally we would like to know the directions and magnitudes of the principal stress components as a function of position and depth in the lithosphere. Such data, however, can only be derived from overcoring measurements in mines or at the Earth's surface and hence are sparse. Hence much of the available information about stress in the lithosphere is derived from other techniques. Among these are observations of earthquakes, which are a manifestation of the stresses and thus a source of information about them. For example, earthquakes provide 50% of the data in a recent global compilation of stress orientations (Zoback *et al.* 1989). In oceanic lithosphere, where borehole data are sparse, earthquakes provide almost all of the available data.

Here, we review some inferences about stress in the oceanic lithosphere that can be derived from earthquake data. The oceanic lithosphere includes material within plate interiors, and that at the boundaries: the newly formed material at spreading centres, that being subducted, and that undergoing strike–slip motion at transforms. Earthquakes at the plate boundaries reflect the forces and stresses there which presumably have a large effect on the stresses, and hence earthquakes, within the plates. Here, we focus primarily on oceanic intraplate earthquakes. These earthquakes (figure 1), though few compared with plate-boundary earthquakes,

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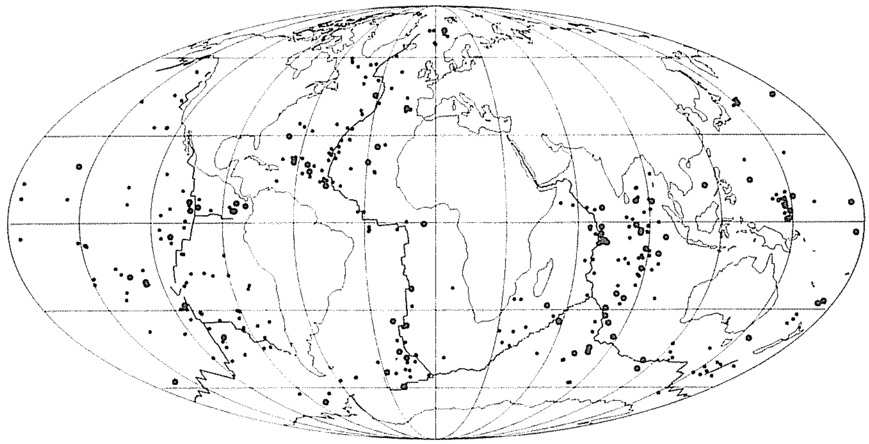

Figure 1. Locations of oceanic intraplate earthquakes between 1964 and 1983. Larger symbols show earthquakes with $m_b \geqslant 5.4$. Events of this size are generally large enough for mechanism and depth study (Bergman 1986).

Table 1. *Seismological constraints on lithospheric stress*

observable	constraints	caveats
focal mechanism	principal stress directions	can be biased by pre-existing faults
focal depth	strength against depth	interpretation requires thermal, rheological, and faulting models
source parameters (magnitudes, moment, time function)	stress drop	interpretation requires assumptions on fault geometry, rupture process, and relation to ambient stress

convey some interesting information about intraplate stresses and plate driving forces. The oceanic plates should be those most easily understood from a driving force standpoint, since they are presently the fastest moving plates and should be more uniform thermally and mechanically, and thus less affected by local stress perturbations from density inhomogeneities than the continents.

The inferences that can be drawn from oceanic intraplate earthquakes can be divided by the type of observations used. Table 1 shows the three types of observations discussed here, the constraints each provides, and some of the limitations of the resulting inferences. The locations of earthquakes and their focal mechanisms provide constraints to test and discriminate models of plate driving forces. In this application, the stresses inferred from the focal mechanisms are compared with those predicted by various models. The focal depths of earthquakes constrain the stresses by indicating depths at which the lithosphere supports stresses adequate for seismic failure. A third data type, which might yield inferences about the magnitudes of stress, consists of source parameters which describe the rupture process and may provide information on the stress drop during the earthquake.

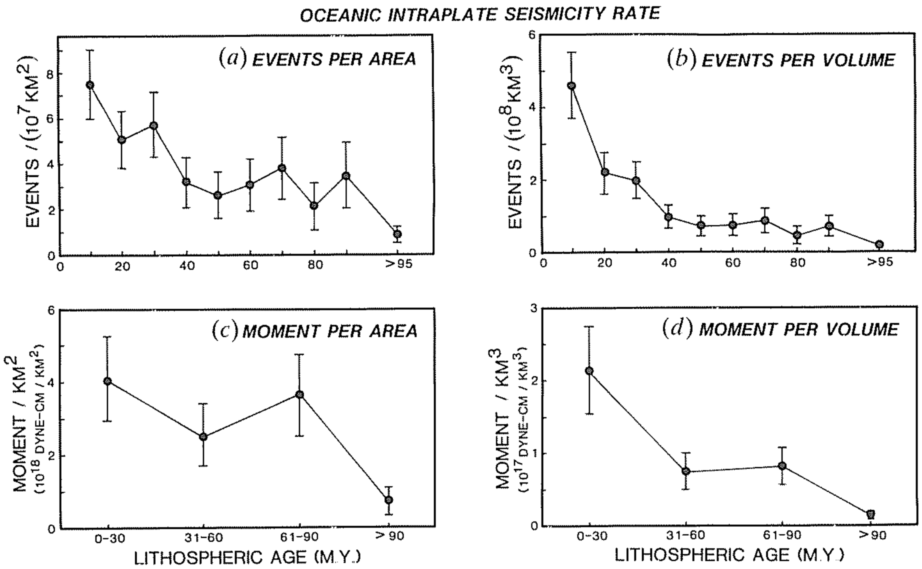

Figure 2. (*a*), (*b*). Number of oceanic intraplate earthquakes with $m_b \geqslant 5.0$ per surface area (*a*) and per volume of lithosphere above the 750 °C isotherm (*b*) as a function of lithospheric age. Both the number of earthquakes and the cumulative seismic moment (*c*, *d*) decrease with age. The area/age relations are from Parsons (1982), and the earthquake data are from 1964 to 1979 (Wiens & Stein 1983).

2. Focal mechanisms

An important application of earthquake seismology is to determine the geometry of the fault giving rise to the observed seismic radiation. For earthquakes of magnitude greater than about 5.5, it is generally possible to determine two orthogonal nodal planes, one of which is the fault plane (Aki & Richards 1980). The observed fault geometry is related to the principal stress axes via the simple Coulomb criterion (Jaeger & Cook 1976), which predicts fracture on a plane of maximum shear stress. These planes are halfway between the maximum and minimum principal stress directions, and include the intermediate principal stress direction. Since the shear stresses acting on the Earth's free surface are zero, near the surface one principal stress axis is vertical, and the other two are horizontal. Using this approach, the directions and relative magnitudes of the principal stresses are inferred from the fault geometry.

Implicit in this approach is the idea that the fault formed in response to the stress field, such that its geometry reflects the stress. Often, however, the earthquake occurred on a pre-existing weak zone or fault. In this case, inferring principal stress directions 45° from the fault is inappropriate, since a range of principal stress directions will give rise to slip on the pre-existing fault, rather than formation of a new fault (Handin 1969). Hence stress directions inferred from individual mechanisms may reflect the orientations of weak zones rather than regional tectonic stresses (McKenzie 1969*a*; Stein 1979). Ideally, the regional stress field would be estimated from a number of events in a given area showing a consistent inferred stress orientation. Unfortunately, intraplate earthquakes are rare enough that an adequate sample of earthquakes is generally not available for most regions, and inferences are drawn from single events.

The primary application of the stresses inferred from mechanisms is in the development and testing of models of the forces giving rise to regional stress fields which vary smoothly over large areas. Of special interest are the stresses expected to result from the forces driving-plate motions (Harper 1975; Forsyth & Uyeda 1975; Richter 1977; Richter & McKenzie 1978). Various studies (Solomon *et al.* 1975; Mendiguren & Richter 1978; Richardson *et al.* 1976, 1979; Richardson & Cox 1984; Wortel & Cloetingh 1985; Cloetingh & Wortel 1985, 1986) thus compare the inferred stresses to those predicted by models. A complication in this approach is the possibility that relatively localized stress sources can also give rise to seismicity, such as is observed in young oceanic lithosphere (discussed shortly) or on passive continental margins (Stein *et al.* 1989).

To develop some basic ideas, it is useful to consider first the variation of seismicity and mechanism types with lithospheric age. Comparison with age is a natural starting point, because the thermal evolution of the oceanic lithosphere is expected to give rise to the driving forces. This approach attempts to identify general patterns, while reducing the dependence on individual earthquakes, which pose the difficulties discussed earlier.

An interesting observation is that the level of seismicity, in terms of both the number of earthquakes and the seismic moment release, decreases dramatically with lithospheric age (figure 2). This effect is especially noticeable when corrected for the observation (discussed later) that the maximum depth of earthquakes appears to be controlled by temperature, so that the relevant variable is the volume of lithosphere of a given age above a limiting isotherm. This may provide a useful constraint if, as often implicitly assumed, the amount of seismicity is related approximately to the magnitude of stresses. The focal mechanisms of the earthquakes also vary with age (figure 3). Lithosphere older than about 35 Ma is in deviatoric compression, as shown by the thrust and strike–slip mechanisms. Younger lithosphere shows both extension and compression, with the former especially common in the Indian Ocean.

Models that seek to explain the seismological observations attempt to relate the stresses inferred from seismicity to those predicted by plate driving-force models. The models postulate a number of forces which contribute to the stress in any region. The stress at any location can be modelled by numerically finding the net effect of these forces for the entire plate.

Various authors have attempted to estimate the relative importance of different forces. For simplicity, we consider only three here. One, misleadingly termed 'ridge push', should arise from density gradients due to the cooling of the lithosphere. Within a region of the lithosphere, ridge push can be thought of as a body force, which is zero at the ridge and increases linearly with age, acting in the direction of the age gradient (Lister 1975; Parsons & Richter 1980). This force should give rise to deviatoric compression and hence thrust fault mechanisms. In young lithosphere the compression would be oriented approximately in the spreading direction, whereas for older lithosphere, the overall plate geometry can cause the compression direction to deviate from the local age gradient. A second major force, slab pull, should result from the density anomaly of the cold subducting lithospheric slabs (McKenzie 1969*b*). This force, which would depend on the thermal structure of the slab, varies with the age of the subducting lithosphere. A third force should arise from drag on the base of the plates, which resists their motion over the viscous mantle. At any point, this force should be proportional to, and directed opposite to, the absolute plate velocity (Solomon & Sleep 1974).

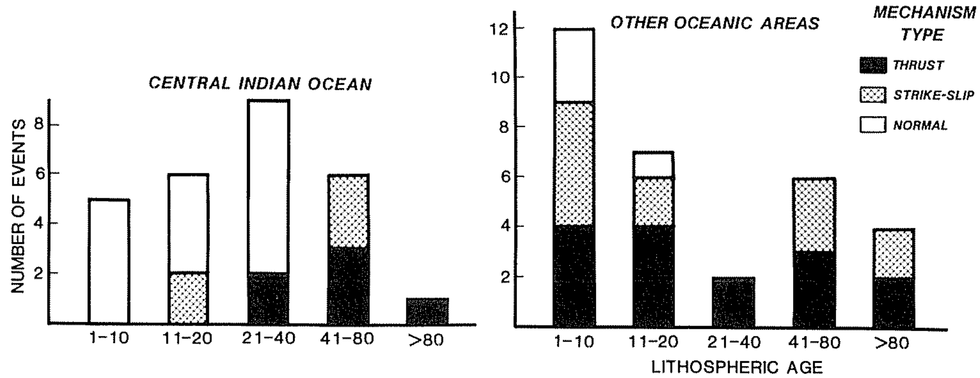

Figure 3. Mechanism type as a function of lithospheric age for oceanic intraplate earthquakes. Older oceanic lithosphere is in compression, whereas younger lithosphere has both extensional and compressional mechanisms. Extensional events are located primarily in the Central Indian Ocean (Wiens & Stein 1984).

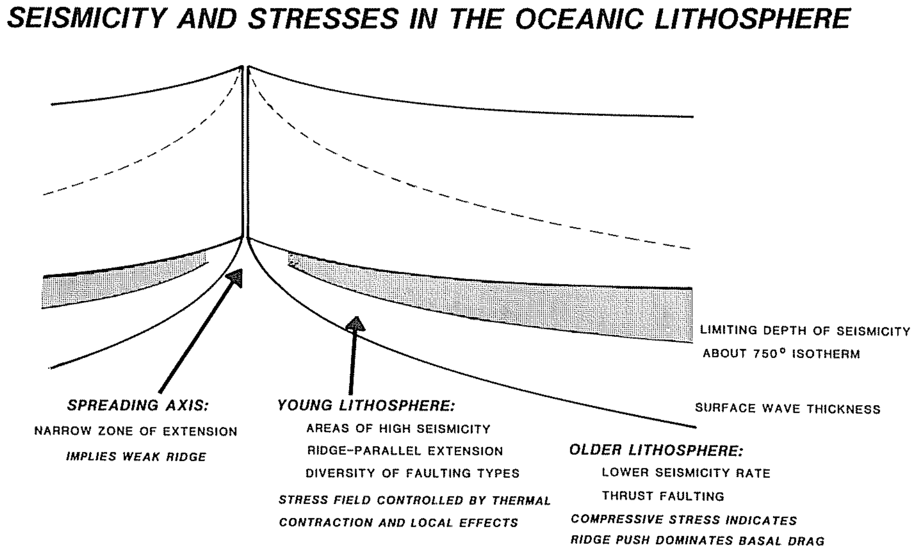

Figure 4. Schematic diagram relating seismicity observations to the tectonics of the oceanic lithosphere (Wiens & Stein 1985).

Following this approach, a number of studies have considered the relation of the focal mechanism data to the stresses predicted by different driving-force models. Since space precludes discussion of all of these, we focus on several key ideas, summarized in figure 4. Specifically, we summarize some ideas about the stresses firstly in old lithosphere, and then secondly in young lithosphere.

3. Focal mechanisms and stresses in old lithosphere

The observation of thrust faulting in older lithosphere has for some time been attributed to the compression expected from ridge push (Mendiguren 1971; Forsyth 1973; Sykes & Sbar 1974). The platewide models listed earlier generally predict this compression, although the direction at any point is not always that predicted. Given the difficulties with inferring stress directions from isolated mechanisms, a misfit

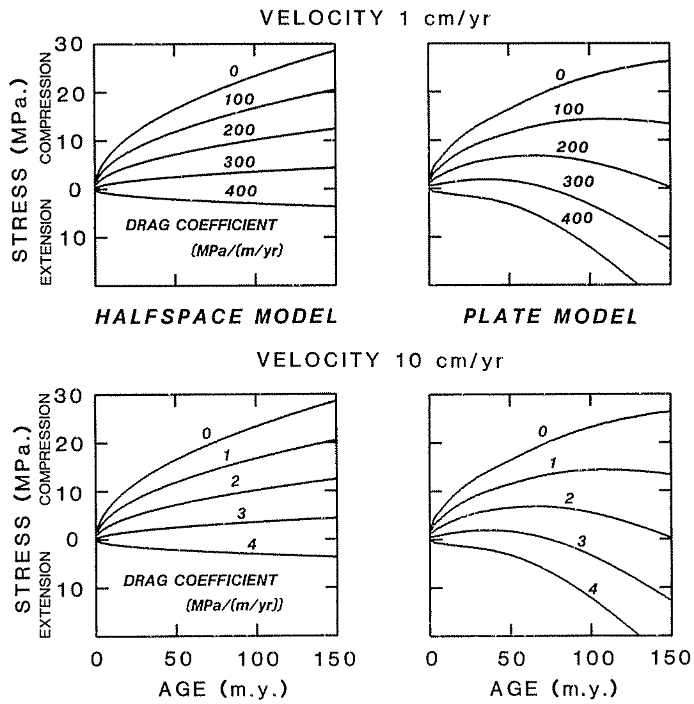

Figure 5. Intraplate stress as a function of lithospheric age and assumed basal drag coefficient for slow moving (top) and fast moving (bottom) plates. Figures on the left assume a half-space thermal model (McKenzie 1969*b*); figures on the right assume a plate model (Parsons & Sclater 1977). The observation of compressional stresses in oceanic lithosphere places an upper bound on the drag coefficient (Wiens & Stein 1985).

need not be a criticism of a model. The generalization that stress in old lithosphere is compressive can constrain the relative magnitudes of the ridge push and basal drag forces. Although the magnitude of ridge push is predicted from models of the thermal evolution of the lithosphere, the magnitude of the basal drag is unknown. An estimate of the drag is useful both for plate driving-force models, and for study of the viscosity of the asthenosphere below the plates.

A simple approach to this issue is to use a two-dimensional model (Wiens & Stein 1985). Assuming the plate is in equilibrium, the stress is determined by the distributed ridge-push body force and boundary conditions representing the effect of material below the plate and at the ridge. Stress at the 'end' boundary is a result rather than a condition; deviatoric compression in old lithosphere indicates that the net force due to the adjacent plate (which in the case of subduction includes the thrust fault contact) is resistive. Initially, the ridge axis is assumed to be rheologically weak, and hence stress free, as discussed later.

The magnitude of the intraplate stress in this model reflects the difference between the effects of ridge push and basal drag. Figure 5 shows intraplate stress as a function of age and drag coefficient, for a half-space model of slow and fast moving plates. If the drag coefficient is zero, the deviatoric stress from ridge push alone is compressive (negative) and varies as the square root of age, because ridge push increases linearly while lithospheric thickness increases as the square root. For larger drag values, the stress follows square-root curves corresponding to less compression, until the lithosphere is in extension for all ages. To satisfy the constraint that old lithosphere is in compression the drag coefficient for the rapidly moving plate must be less than about 3.5 MPa (m $a^{-1})^{-1}$, corresponding to a basal stress less than 0.35 MPa (3.5 bar). Comparable values were found by Hager & O'Connell (1981) and Fleitout & Froidevaux (1983). In contrast, a plate moving a factor of ten slower can remain in compression with a drag coefficient 100 times greater, corresponding to a basal stress ten times greater.

Similar results arise if the lithosphere is modelled as a cooling plate. In this model, because the lithosphere approaches constant thickness with age, the ridge-push force also approaches a constant. As a result, the stress need not be monotonic with age. For no drag, deviatoric stress becomes more compressive with age. For larger values of the drag coefficient the stress reaches a compressional maximum, becomes less compressive, and eventually becomes extensional. For the plate models, the maximum drag coefficient allowing compression in old lithosphere is about half of the value for the half-space models.

Such a model can also yield constraints on mantle viscosity structure (Wiens & Stein 1985) under a simple assumption that the drag on the base of a plate is due to motion over the viscous mantle. For a simple two-dimensional geometry the mass flux due to the moving plate is balanced by a return flow at depth. A drag coefficient consistent with the focal mechanisms would be expected if the plate is underlain by a thin low-viscosity layer, which decouples the plates from the higher-viscosity material below, as proposed by Richter & McKenzie (1978) and Hager & O'Connell (1979).

These two-dimensional models are simple representations of what might occur for an average oceanic plate. None the less, they demonstrate how the observation of intraplate thrust earthquakes in old lithosphere, and the consequent assumption of deviatoric compression, can be used to draw interesting inferences about driving forces. More detailed results can be derived from models in which the forces vary laterally, such that the direction of the compressive stress can be modelled.

4. Young lithosphere

The focal mechanisms for intraplate earthquakes in lithosphere younger than about 35 Ma (figure 3) show both compression and extension (Wiens & Stein 1984; Bergman & Solomon 1984). The simplest interpretation of the extensional events would be that they reflect a broad zone of extension about the extensional mid-ocean ridges. If so, the direction of extension should be normal to the spreading centres. Surprisingly, this does not appear to be the case.

Figure 6 shows that the direction of the horizontal component of the T (tensional) axes for normal faulting events do not indicate extension in the spreading direction. The T axes are oriented at large angles to the spreading direction, and may show a weak preferred orientation approximately perpendicular to the spreading direction. However, since the T axis data-set is strongly dominated by data from the Indian Ocean, it may largely reflect local processes, discussed later. The P axes of near-ridge thrust faulting events show a weak preferred orientation in the direction of spreading.

The absence of ridge-normal extensional earthquakes suggests that the ridge is rheologically weak, as assumed in the previous section. Figure 7 shows the vertically averaged stress in the spreading direction for the model used earlier (§3), with

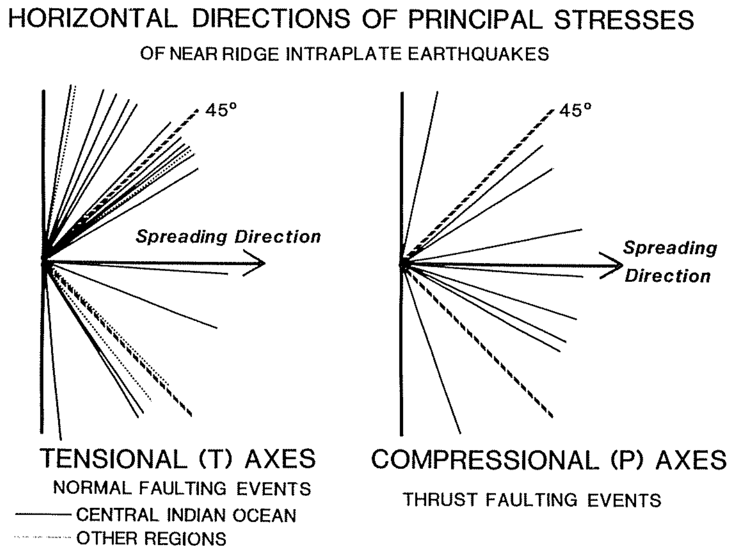

Figure 6. Histograms of the angle between the horizontal component of principal stress and the spreading direction for normal faulting (left) and thrust faulting (right) intraplate earthquakes in young oceanic lithosphere. Tensional axis of normal faulting events, most of which are located in the Central Indian Ocean, generally show extension oblique to the spreading direction, while compressional axis of thrust faulting events show a weak preferred orientation in the spreading direction (Wiens & Stein 1984).

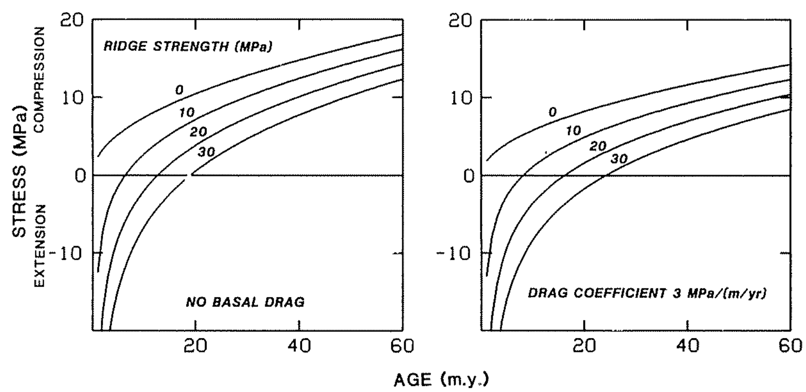

Figure 7. Vertically averaged stress in the spreading direction as a function of lithospheric age computed for a plate velocity of 5 cm a^{-1} and several values of ridge strength and basal drag. The location of the transition from ridge-normal extension to compression is very sensitive to the strength of the ridge (Wiens & Stein 1984).

different boundary conditions at the ridge. The stress field in young lithosphere, especially the location of the transition from compressive to extensional stresses, is sensitive to the strength of the ridge. Ridge axes capable of supporting greater tensional stresses produce wider zones of intraplate extension. Since this ridge-normal extension zone is not observed, the spreading-related extension seems to be restricted to a narrow (less than 100 km) zone about a weak ridge. This constraint is useful for intraplate stress and driving force modelling.

What then gives rise to the ridge-parallel extension? One possibility is thermoelastic stresses (Turcotte & Oxburgh 1973; Parmentier & Haxby 1986;

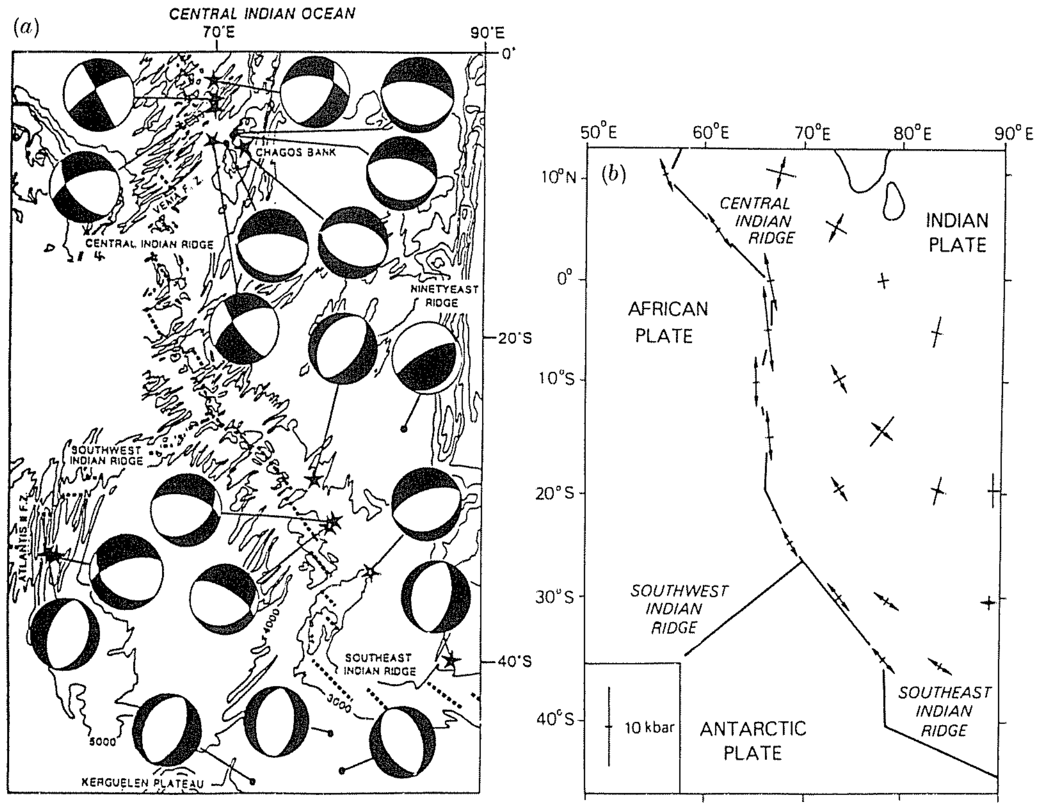

Figure 8. (*a*) Near-ridge intraplate seismicity of the Central Indian Ocean. Most events indicate a regionally consistent extensional stress field oriented parallel to the ridge axis. Dashed lines denote ridge axes. Bathymetry is shown in 1000 m intervals (Wiens & Stein 1984). (*b*) Stresses predicted by model of Cloetingh & Wortel (1986), rescaled for an elastic plate thickness corresponding with the 600 °C isotherm. Symbols (←|→) and (–|–) denote tension and compression, respectively. The predicted stresses are in good agreement with the mechanisms (Stein *et al.* 1987).

Sandwell 1986). An attractive feature of this explanation is that the thermoelastic stresses might decrease with age (Bratt *et al.* 1985), consistent with the decrease in seismicity (figure 2). A difficulty, however, is that the ridge-parallel extensional earthquakes are concentrated in the Central Indian Ocean (figure 8*a*), whereas there is no obvious reason to expect thermoelastic stresses to behave this way.

A more satisfying alternative is that the Indian Ocean extensional zone is a local tectonic situation. Cloetingh & Wortel (1985, 1986) have presented a stress model for a single Indo-Australia plate. The predicted stresses vary dramatically with position, and show a pattern consistent (figure 8*b*) with the focal mechanisms. Moreover, the stress averaged over the mechanically strong lithosphere decreases with age in a manner similar to the normal faulting seismicity (figure 9). It thus appears that the extensional zone can be modelled as a consequence of plate-driving forces. Interestingly, it appears that this situation modelled assuming a single Indo-Australia plate corresponds to the recently recognized diffuse plate-boundary zone between separate Indian and Australian plates (Wiens *et al.* 1985, 1986; Gordon *et al.* 1990). It thus appears that although thermoelastic stresses may provide a low level 'background' in all plates, the Indian Ocean extensional zone is a local phenomenon.

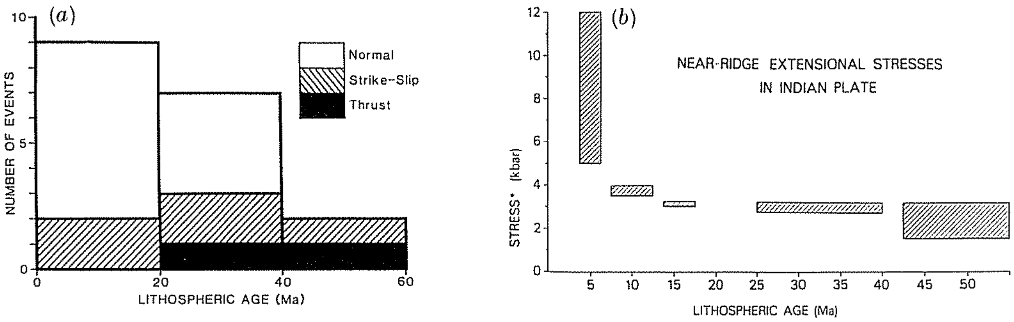

Figure 9. (*a*) Types of faulting found in intraplate earthquakes in the Central Indian Ocean, on the Indian Plate south of 0° and west of 90° E, as a function of lithospheric age. Extensional mechanisms predominate in young lithosphere. (*b*) Average stresses as a function of age from figure 8*a*. The stress decreases in a fashion similar to the number of normal fault earthquakes (Stein *et al.* 1987).

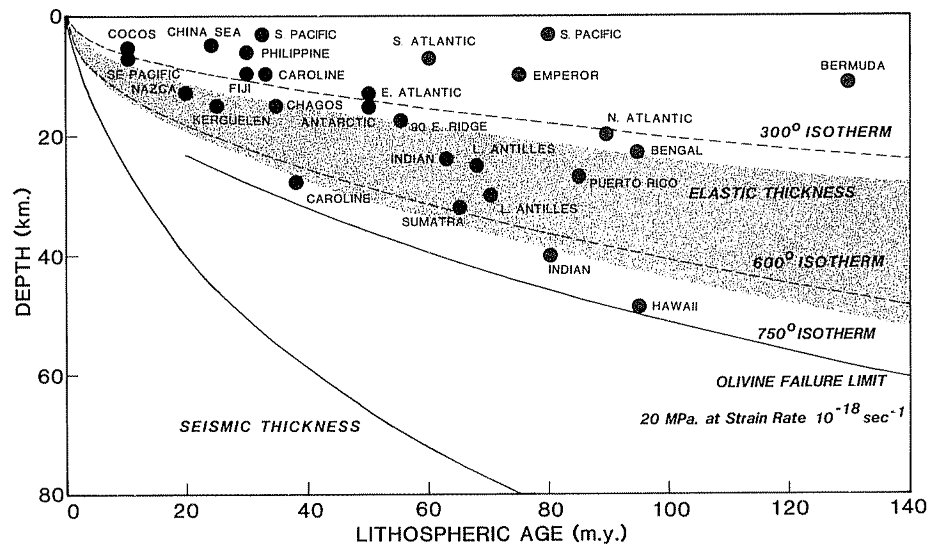

Figure 10. Oceanic earthquake depths against thermal structure of the lithosphere. Isotherms (in degrees centigrade) are for a plate cooling model (Parsons & Sclater 1977). Stippled region denotes estimates of the flexural elastic thickness (Watts *et al.* 1980). The seismic thickness is from surface-wave dispersion data (Leeds *et al.* 1974). The failure limit is the lower limit at which 20 MPa (200 bar) deviatoric stress can be sustained, calculated for a dry olivine rheology (Goetze & Evans 1979) and a strain rate of $10^{-18}\ s^{-1}$ (Wiens & Stein 1983).

5. Depths of intraplate earthquakes

The second major data-set derivable from observations of intraplate earthquakes, which has implications for lithospheric stress, is the depths of the earthquakes. In the past decade, waveform modelling techniques have made it possible to determine the focal depths of oceanic intraplate earthquakes to a precision of a few kilometres (Stein & Weins 1986). Figure 10 shows that the maximum depth of seismicity increase with ages. Comparison with the predicted thermal structure of the oceanic lithosphere for a plate cooling model suggests that the maximum depth of seismicity corresponds to a temperature of *ca.* 750 °C (Wiens & Stein 1983; Chen & Molnar

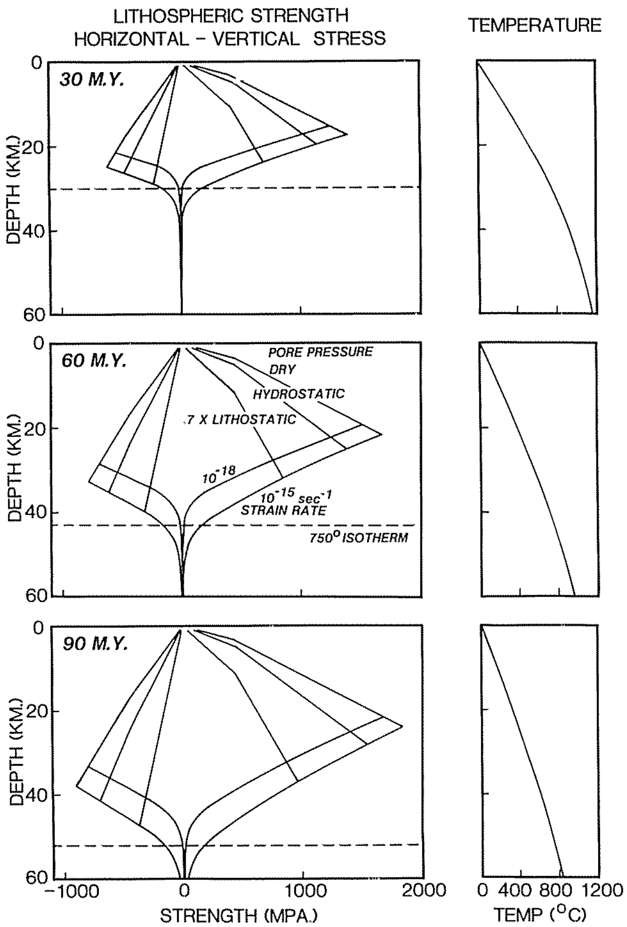

Figure 11. Strength envelopes showing maximum stress difference as a function of depth calculated using temperatures in degrees centigrade from the Parsons & Sclater (1977) plate cooling model. At shallow depth strength is controlled by brittle fracture (Byerlee 1968), at greater depth ductile flow (Goetze & Evans 1979) predicts rapid weakening. The 750 °C isotherm (dashed line) is the approximate lower bound for oceanic intraplate seismicity (Wiens & Stein 1983).

1983). This depth approximately equals the elastic thickness determined from studies of lithospheric flexure due to applied loads, but is significantly less than the depth to the low-velocity zone ('seismic thickness') found from surface-wave dispersion. As the elastic thickness measures the response of the lithosphere to long-term loads, whereas the seismic thickness measures its response to short-term loads, it appears that intraplate earthquakes occur primarily where the lithosphere can sustain long-term loads.

The focal depths suggest that above a temperature of *ca.* 750 °C the lithosphere is too weak to support the stresses needed for seismic failure. This effect can be modelled using results from laboratory rock mechanics studies (Kirby 1977, 1980; Goetze & Evans 1979; Brace & Kohlstedt 1980) to predict the variation of strength (the maximum difference between principal stresses that rock can support) with depth in the lithosphere. At low temperatures, corresponding to shallow depths, failure occurs by brittle fracture and strength increases with depth. At higher

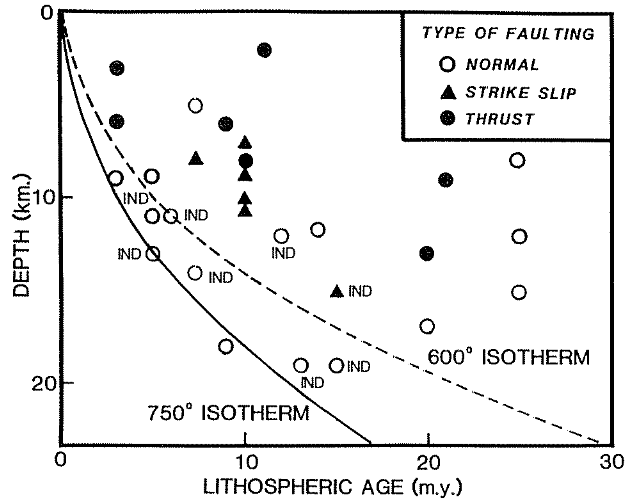

Figure 12. Depths of intraplate earthquakes in young lithosphere as a function of lithospheric age. Isotherms are from the Parsons & Sclater (1977) plate cooling model. Normal faulting events seem to be located at greater depths and temperatures than thrust and strike slip events. Most comparatively deep normal faulting occurs in the Indian Plate (events marked 'IND') (Wiens & Stein 1984).

temperatures, where rocks deform by ductile flow, strength decreases exponentially with temperature. Hence as the lithosphere cools, the portion expected to have strength adequate to support seismicity thickens (figure 11).

Comparison of the predicted strengths to the depth observations requires a specific rheology. It also requires the assumption of a strain rate, because strength in the ductile region increases with strain rate, and a pore pressure distribution, because in the brittle region strength decreases with pore pressure. None the less, the basic idea is robust. An interesting feature of this analysis is that, for the rheology and thermal structure assumed, earthquakes occur in both the brittle and ductile regions. If the parameters assumed are appropriate, it appears that the requirement for seismicity is adequate strength, rather than a brittle rheology. Another interesting implication is that normal faulting may occur deeper than thrust faulting, because the lithosphere is stronger to a greater depth in deviatoric extension than in compression. The focal depth data provide a suggestion of this phenomenon (figure 12), although the generally greater depths of the normal fault events could reflect another cause.

This type of strength envelope analysis has become a standard tool in investigating lithospheric stresses. Observations of either the maximum depth of earthquakes or the thickness of the mechanically strong portion of the lithosphere (Bodine *et al.* 1981; McNutt & Menard 1982) are combined with an assumed temperature structure to constrain stresses. This has been done to study regions of intraplate deformation (Stein & Weissel 1990), oceanic transform faulting (Engeln *et al.* 1986; Bergman & Solomon 1988; Chen 1988), and possible midplate reheating (Stein & Abbott 1991). A similar technique is also used extensively in studies of continental geology (Sibson 1982; Meissner & Strehlau 1982; Smith & Bruhn 1984; Kusznir & Karner 1985; Shudofsky *et al.* 1987).

The technique has also been applied recently to the question raised earlier (§4) about the strength of the mid-ocean ridge. Mid-ocean ridge normal fault earthquakes

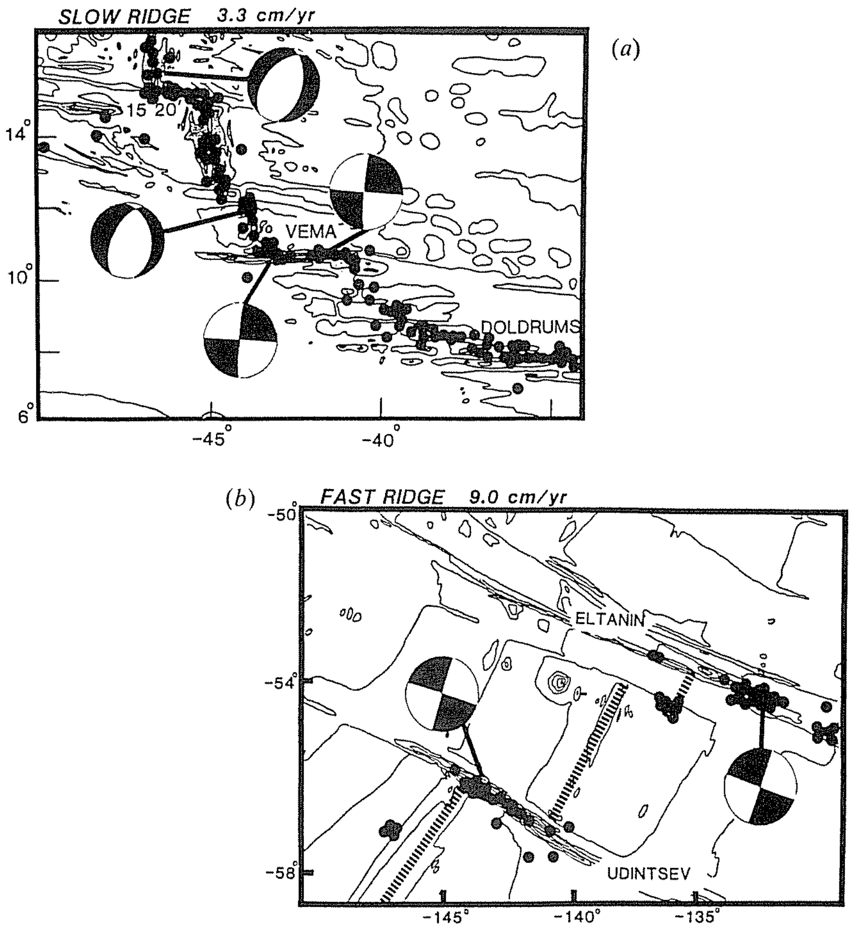

Figure 13. Comparison of earthquake mechanisms on slow and fast spreading ridges. The slow spreading Mid-Atlantic ridge (*a*) has earthquakes with normal faulting on the ridge axis, and strike–slip earthquakes on the transforms. In contrast, the fast spreading East Pacific Rise (*b*) shows no seismicity along the ridge segments, although strike–slip earthquakes occur on the transforms (Stein & Woods 1989.)

are restricted to slow spreading ridges (half rate less than 3 cm a^{-1}) (figure 13), where the axis is characterized by a median valley (Sleep & Rosendahl 1979). The maximum depth of the earthquakes decreases with spreading rate (Huang & Solomon 1988). This trend is plausible, since thermal models (Sleep 1975) predict that the ridge axis will be cooler to greater depth for slower spreading. The earthquakes extend, however, to depths below the predicted 750 °C isotherm. As a result, the ridge axis should have insufficient strength to permit this seismicity. Lin & Parmentier (1989) thus propose that the ridge is cooled sufficiently by hydrothermal circulation that seismicity can occur to these depths.

A final point worth noting is that such analyses use the earthquake depths only to infer that strength at that depth exceeds a minimum required value. A recent approach (Govers *et al.* 1991) explicitly estimates the magnitude of the tectonic stress at depth by assuming that the earthquake's occurrence indicates that stress has reached the predicted strength envelope.

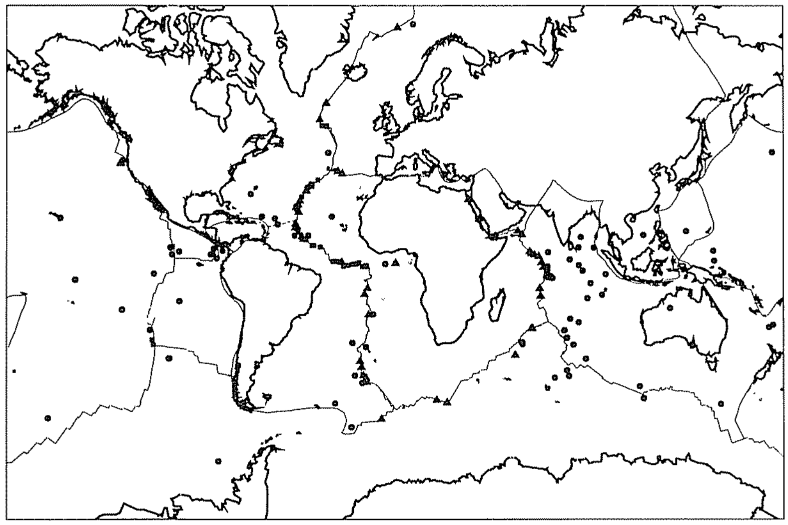

Figure 14. Locations of earthquakes used in figures 15 and 16. Intraplate events are shown by circles; transform events by squares; and mid-ocean ridge events by triangles. These events are studied by various investigators: Wiens & Stein (1984), Bergman *et al.* (1985), Bergman & Solomon (1984, 1985), and Bergman (1986) for the intraplate earthquakes; Engeln (1985), Engeln *et al.* (1986), Goff *et al.* (1987), and Bergman & Solomon (1988) for transform events; Huang *et al.* (1986), Huang & Solomon (1987, 1988) and Jemsek *et al.* (1986) for mid-ocean ridge earthquakes.

6. Source parameters

In addition to earthquake mechanisms and focal depths, seismological observations may be able to provide useful information on the magnitudes of stress via parameters characterizing the earthquake rupture. Seismic source theory (reviewed by Geller (1976) and Kanamori (1980)) provides useful insight. The seismic radiation from an earthquake depends on parameters including the seismic moment

$$M_0 = \mu \bar{D} S, \tag{6.1}$$

where μ is the rigidity and $\bar{D}$ is the average slip (or dislocation) on the fault, whose area is S. For continental or large subduction zone earthquakes, the fault area can often be inferred from aftershocks. In contrast, for oceanic intraplate events, the fault area is usually inferred from the duration of the pulse τ generated at the source. This duration depends on how long the rupture takes to propagate across the fault area, which depends on the rupture velocity, a quantity assumed to be proportional to the shear velocity β

$$\tau = cS^{\frac{1}{2}}/\beta. \tag{6.2}$$

Since the strain change associated with the earthquake is proportional to the ratio of the slip to the square root of the fault area, the corresponding stress drop is

$$\Delta\sigma = C\mu\bar{D}/S^{\frac{1}{2}} = CM_0/S^{\frac{3}{2}} \approx CM_0/(\beta\tau)^3, \tag{6.3}$$

where C depends on the fault geometry and other parameters. Hence for a given seismic moment, longer source time functions correspond to lower stress drops. Although there are further complexities, due to the finite rise time required for the dislocation to reach its full value at any point, this general concept can be used to estimate relative stress drops (Chung & Kanamori 1980). Note that this method

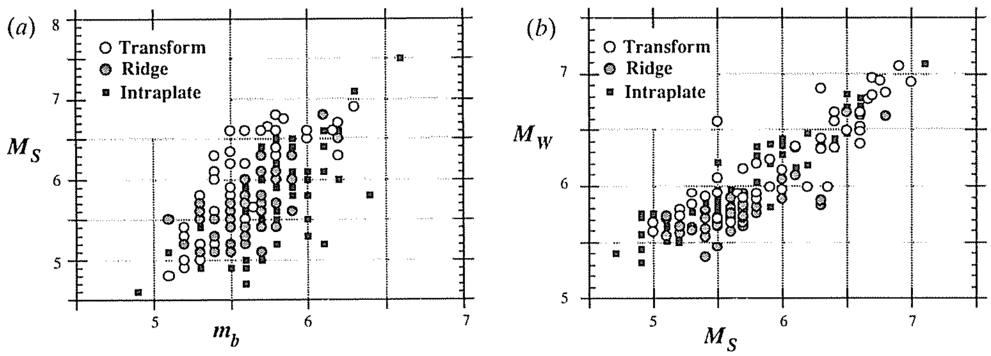

Figure 15. Comparison of body and surface wave magnitudes (*a*) and surface wave and moment magnitudes (*b*) for oceanic ridge, transform and intraplate earthquakes. Both comparisons show that the transform events release more energy at longer periods than ridge events, whereas intraplate events overlap both.

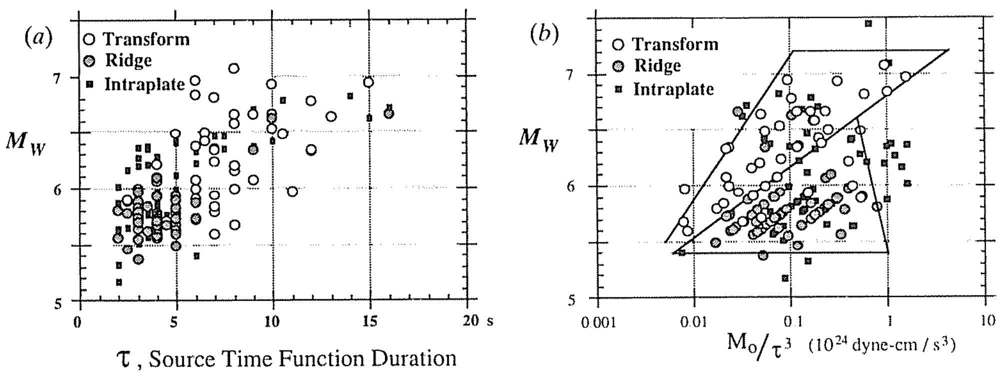

Figure 16. Comparison of time function duration τ and M_0/τ^3 as a function of moment magnitude M_w. For a given M_w, transform earthquakes have longer time functions and hence smaller M_0/τ^3 than ridge events. This effect may indicate that ridge events may have higher stress drops than transform events.

examines only stress drops, rather than the ambient tectonic stress. The relation between the two depends on assumption about the rupture process.

To date, studies of oceanic earthquakes have used a related concept, comparison of the magnitudes determined at different periods, to estimate how energy release varies with period. For this purpose, the body wave magnitude determined at a period of 1 s (m_b), the surface wave magnitude determined at a period of 20 s (M_s), and the moment magnitude determined from the longest period data possible (M_w) (Kanamori 1977) can be compared. Kanamori & Stewart (1976) and Okal & Stewart (1982) suggest that some oceanic transform earthquakes have large M_s relative to m_b, suggesting a slow energy release, perhaps qualitatively indicative of low stress.

We have begun (Pelayo & Stein 1991) to apply this idea to a data-set of oceanic ridge, transform and intraplate earthquakes (figure 14). This data-set has the advantage that all the earthquakes occur in oceanic lithosphere, so we compare events in material of presumably similar mechanical properties, rather than compare them with global data-sets that contain many continental earthquakes. The preliminary results show several interesting patterns (figures 15, 16). For a given m_b, transform events generally have higher M_s than ridge events, suggesting that more energy is released at longer periods. Intraplate events overlap both the ridge and transform events. A similar pattern emerges for M_s and M_w. A consistent pattern can be seen in another way, because for a given M_w, transform earthquakes have longer time functions and hence smaller M_0/τ^3 than ridge events. Thus with respect to the population of intraplate events, ridge earthquakes are generally 'fast' whereas transform earthquakes are generally 'slow'. A variety of factors may contribute to these variations. Ridge events may have higher stress drops than transform events, perhaps due to higher tectonic stress. Alternatively, these differences may reflect different fault geometries or rupture properties. We hope to determine which of the possibilities is most likely.

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References

Aki, K. & Richards, P. G. 1980 *Quantitative seismology: theory and methods*. San Francisco: W. H. Freeman.

Bergman, E. A. 1986 Intraplate earthquakes and the state of stress in oceanic lithosphere. *Tectonophys.* **132**, 1–35.

Bergman, E. A. & Solomon, S. C. 1984 Source mechanisms of earthquakes near mid-ocean ridges from body waveform inversion: implications for the early evolution of oceanic lithosphere. *J. geophys. Res.* **89**, 11415–11441.

Bergman, E. A. & Solomon, S. C. 1985 Earthquake source mechanisms from body waveform inversion and intraplate tectonics in the northern Indian Ocean. *Phys. Earth planet. Interiors* **40**, 1–23.

Bergman, E. A. & Solomon, S. C. 1988 Transform fault earthquakes in the North Atlantic: source mechanisms and depth of faulting. *J. geophys. Res.* **93**, 9027–9057.

Bergman, E. A., Nabelek, J. L. & Solomon, S. C. 1985 Correction to 'An extensive region of off ridge normal-faulting earthquakes in the southern Indian Ocean'. *J. geophys. Res.* **90**, 2066.

Bodine, J. H., Steckler, M. S. & Watts, A. B. 1981 Observations of flexure and the rheology of the oceanic lithosphere. *J. geophys. Res.* **86**, 3695–3707.

Brace, W. F. & Kohlstedt, D. L. 1980 Limits on lithospheric stress imposed by laboratory experiments. *J. geophys. Res.* **85**, 6248–6252.

Byerlee, J. D. 1968 Brittle–ductile transition in rocks. *J. geophys. Res.* **73**, 4741–4750.

Bratt, S. R., Bergman, E. A. & Solomon, S. C. 1985 Thermoelastic stress: how important as a cause of earthquakes in young oceanic lithosphere? *J. geophys. Res.* **90**, 10249–10260.

Chen, W.-P. & Molnar, P. 1983 Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere. *J. geophys. Res.* **88**, 4183–4214.

Chen, Y. 1988 Thermal model of oceanic transform faults. *J. geophys. Res.* **93**, 8839–8851.

Chung, W.-Y. & Kanamori, H. 1980 Variation of seismic source parameters and stress drops within a descending slab and its implications in plate mechanics. *Phys. Earth planet. Interiors* **23**, 134–159.

Cloetingh, S. & Wortel, R. 1985 Regional stress field of the Indian plate. *Geophys. Res. Lett.* **12**, 77–80.

Cloetingh, S. & Wortel, R 1986 Stress in the Indo-Australian plate. *Tectonophys.* **132**, 49–67.

Engeln, J. F. 1985 Seismological studies of the tectonics of divergent plate boundaries. Ph.D. thesis, Northwestern University, Evanston, Illinois.

Engeln, J. F., Wiens, D. A. & Stein, S. 1986 Mechanisms and depths of Atlantic transform earthquakes. *J. geophys. Res.* **91**, 548–577.

Fleitout, L. & Froidevaux, C. 1983 Tectonic stresses in the lithosphere. *Tectonics* **2**, 315–324.

Forsyth, D. W. 1973 Compressive stress between two mid-ocean ridges. *Nature, Lond.* **243**, 78–79.

Forsyth, D. W. & Uyeda, S. 1975 On the relative importance of the driving forces of plate motion. *Geophys. Jl R. astr. Soc.* **43**, 163–200.

Geller, R. J. 1976 Scaling relations for earthquake source parameters and magnitudes. *Bull. seismol. Soc. Am.* **66**, 1501–1523.

Goetze, C. & Evans, B. 1979 Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics. *Geophys. Jl R. astr. Soc.* **59**, 463–478.

Goff, J. A., Bergman, E. A. & Solomon, S. C. 1987 Earthquake source mechanisms and transform fault tectonics in the Gulf of California. *J. geophys. Res.* **92**, 10485–10510.

Gordon, R. G., DeMets, C. & Argus, D. F. 1990 Kinematic constraints on distributed lithospheric deformation in the equatorial Indian Ocean from present motion between the Australian and Indian plates. *Tectonics* **9**, 409–422.

Govers, R., Wortel, R., Cloetingh, S. & Stein, C. 1991 Stress magnitude estimates from earthquakes in oceanic plate interiors. *J. geophys. Res.* (Submitted.)

Hager, B. H. & O'Connell, R. J. 1979 Kinematic models of large-scale flow in the Earth's mantle. *J. geophys. Res.* **84**, 1031–1048.

Hager, B. H. & O'Connell, R. J. 1981 A simple global model of plate dynamics and mantle convection. *J. geophys. Res.* **86**, 4843–4867.

Handin, J. 1969 On the Coulomb–Mohr failure criterion. *J. geophys. Res.* **74**, 5343–5348.

Harper, J. F. 1975 On the driving forces of plate tectonics. *Geophys. Jl R. astr. Soc.* **40**, 465–474.

Huang, P. Y. & Solomon, S. C. 1987 Centroid depths and mechanisms of ridge-axis earthquakes in the Indian Ocean, Gulf of Aden, and Red Sea. *J. geophys. Res.* **92**, 1361–1382.

Huang, P. Y. & Solomon, S. C. 1988 Centroid depths of mid-ocean ridge earthquakes: dependence on spreading rate. *J. geophys. Res.* **93**, 13445–13477.

Huang, P. Y., Solomon, S. C., Bergman, E. A. & Nabelek, J. L. 1986 Focal depths and mechanisms of Mid-Atlantic Ridge earthquakes from body waveform inversion. *J. geophys. Res.* **91**, 579–598.

Jaeger, J. C. & Cook, N. G. W. 1976 *Fundamentals of rock mechanics*. London: Chapman & Hall.

Jemsek, J. P., Bergman, E. A., Nabelek, J. L. & Solomon, S. C. 1986 Focal depths and mechanisms of earthquakes on the Arctic mid-ocean ridge system. *J. geophys. Res.* **91**, 13993–14005.

Kanamori, H. 1977 Seismic and aseismic slip along subduction zones and their tectonic implications. In *Island arcs, deep-sea trenches and back-arc basins, Maurice Ewing Ser.*, vol. 1 (ed. M. Talwani & W. C. Pitman III), pp. 163–174. Washington, D.C.: AGU.

Kanamori, H. 1980 State of stress in the earth's lithosphere. In *Physics of the Earth's interior, Proc. Int. Sch. Phys. 'Enrico Fermi'* (ed. A. M. Dziewonski & E. Boschi), pp. 531–552. Amsterdam: North Holland.

Kanamori, H. & Stewart, G. S. 1976 Mode of the strain release along the Gibbs Fracture Zone, Mid-Atlantic Ridge. *Phys. Earth planet. Interiors* **11**, 312–332.

Kirby, S. H. 1977 State of stress in the lithosphere: Inferences from the flow laws of olivine. *Pure appl. Geophys.* **115**, 245–258.

Kirby, S. H. 1980 Tectonic stresses in the lithosphere: Constraints provided by the experimental deformation of rocks. *J. geophys. Res.* **85**, 6353–6363.

Kusznir, N. & Karner, G. 1985 Dependence of the flexural rigidity of the continental lithosphere on rheology and temperature. *Nature, Lond.* **316**, 138–142.

Leeds, A. R., Knopoff, L. & Kausel, E. G. 1974 Variations of upper mantle structure under the Pacific Ocean. *Science, Wash.* **186**, 141–143.

Lin, J. & Parmentier, E. M. 1989 Mechanisms of lithospheric extension at mid-ocean ridges. *Geophys. J.* **96**, 1–22.

Lister, C. R. B. 1975 Gravitational drive on oceanic plates caused by thermal contraction. *Nature, Lond.* **257**, 663–665.

McKenzie, D. P. 1969*a* The relation between fault plane solutions for earthquakes and the directions of the principal stresses. *Bull. seismol. Soc. Am.* **59**, 591–601.

McKenzie, D. P. 1969*b* Speculations on the consequences and causes of plate motions. *Geophys. Jl R. astr. Soc.* **18**, 1–32.

McNutt, M. K. & Menard, H. W. 1982 Constraints on yield strength in the oceanic lithosphere derived from observations of flexure. *Geophys. Jl R. astr. Soc.* **71**, 363–394.

Meissner, R. & Strehlau, J. 1982 Limits of stresses in continental crusts and their relation to the depth-frequency distribution of shallow earthquakes. *Tectonics* **1**, 73–89.

Mendiguren, J. A. 1971 Focal mechanism of a shock in the middle of the Nazca plate. *J. geophys. Res.* **76**, 3861–3879.

Mendiguren, J. A. & Richter, F. M. 1978 On the origin of compressional intraplate stresses in South America. *Phys. Earth planet. Interiors* **16**, 318–326.

Okal, E. A. & Stewart, L. M. 1982 Slow earthquakes along oceanic fracture zones: evidence for asthenospheric flow away from hot spots. *Earth planet. Sci. Lett.* **57**, 75–87.

Parmentier, E. M. & Haxby, W. F. 1986 Thermal stresses in the oceanic lithosphere: evidence from geoid anomalies at fracture zones. *J. geophys. Res.* **91**, 7193–7204.

Parsons, B. 1982 Causes and consequences of the relation between area and age of the ocean floor. *J. geophys. Res.* **87**, 289–302.

Parsons, B. & Richter, F. M. 1980 A relation between the driving force and geoid anomaly associated with mid-ocean ridges. *Earth planet. Sci. Lett.* **51**, 445–450.

Parsons, B. & Sclater, J. G. 1977 An analysis of the variation of ocean floor bathymetry and heat flow with age. *J. geophys. Res.* **82**, 803–827.

Pelayo, A. M. & Stein, S. 1991 Variation in source parameters for oceanic ridge, transform and intraplate earthquakes (abstract). *Eos Trans. Am. Geophys. Union* (*Spring Meeting Issue*), p. 202.

Richardson, R. M. & Cox, B. L. 1984 Evolution of oceanic lithosphere: a driving force study of the Nazca plate. *J. geophys. Res.* **89**, 10043–10052.

Richardson, R. M., Solomon, S. C. & Sleep, N. H. 1976 Intraplate stress as an indicator of plate tectonic driving forces. *J. geophys. Res.* **81**, 1847–1856.

Richardson, R. M., Solomon, S. C. & Sleep, N. H. 1979 Tectonic stress in the plates. *Rev. Geophys. Space Phys.* **17**, 981–1019.

Richter, F. M. 1977 On the driving mechanism of plate tectonics. *Tectonophys.* **38**, 61–88.

Richter, F. & McKenzie, D. 1978 Simple plate models of mantle convection. *J. Geophys.* **44**, 441–471.

Sandwell, D. T. 1986 Thermal stress and the spacings of transform faults. *J. geophys. Res.* **91**, 6405–6418.

Shudofsky, G., Cloetingh, S., Stein, S. & Wortel, R. 1987 Unusually deep earthquakes in East Africa: constraints on the thermo-mechanical structure of a continental rift system. *Geophys. Res. Lett.* **14**, 741–744.

Sibson, R. H. 1982 Fault zone models, heat flow, and the depth distribution of earthquakes in the continental crust of the United States. *Bull. seismol. Soc. Am.* **72**, 151–163.

Sleep, N. H. 1975 Formation of oceanic crust: some thermal constraints. *J. geophys. Res.* **80**, 4037–4042.

Sleep, N. H. & Rosendahl, B. R. 1979 Topography and tectonics of mid-oceanic ridge axes. *J. geophys. Res.* **84**, 6831–6839.

Smith, R. B. & Bruhn, R. L. 1984 Intraplate extensional tectonics on the eastern Basin-Range: inferences of structural style from seismic reflection data, regional tectonics, and thermal-mechanical models of brittle–ductile deformation. *J. geophys. Res.* **89**, 5733–5762.

Solomon, S. C. & Sleep, N. H. 1974 Some simple physical models for absolute plate motions. *J. geophys. Res.* **79**, 2557–2567.

Solomon, S. C., Sleep, N. H. & Richardson, R. M. 1975 On the forces driving plate tectonics: inferences from absolute plate velocities and intraplate stress. *Geophys. Jl R. astr. Soc.* **42**, 769–801.

Stein, S. 1979 Intraplate seismicity on bathymetric features: the 1968 Emperor Trough earthquake. *J. geophys. Res.* **84**, 4763–4768.

Stein, C. A. & Abbott, D. H. 1991 Heat flow constraints on the South Pacific superswell. *J. geophys. Res.* (In the press.)

Stein, C. A. & Weissel, J. K. 1990 Constraints on the Central Indian Basin thermal structure from heat flow, seismicity and bathymetry. *Tectonophys.* **176**, 315–332.

Stein, S. & Wiens, D. 1986 Depth determination for shallow teleseismic earthquakes: methods and results. *Rev. Geophys. Space Phys.* **24**, 806–832.

Stein, S. & Woods, D. F. 1989 Seismicity: midocean ridge. In *The encyclopedia of solid Earth geophysics* (ed. D. E. James), pp. 1050–1054. New York: Van Nostrand Reinhold.

Stein, S., Cloetingh, S., Sleep, N. & Wortel, R. 1989 Passive margin earthquakes, stresses, and rheology. In *Earthquakes at North-Atlantic passive margins: neotectonics and postglacial rebound* (ed. S. Gregerson & P. Basham), pp. 231–260. Reidel.

Stein, S., Cloetingh, S., Wiens, D. & Wortel, R. 1987 Why does near-ridge extensional seismicity occur primarily in the Indian Ocean? *Earth planet. Sci. Lett.* **82**, 107–113.

Sykes, L. R. & Sbar, M. L. 1974 Focal mechanism solutions of intraplate earthquakes and stresses in the lithosphere. In *Geodynamics of Iceland and the North Atlantic area* (ed. L. Kristjansson), pp. 207–224. Dordrecht: Reidel.

Turcotte, D. L. & Oxburgh, E. R. 1973 Mid-plate tectonics. *Nature, Lond.* **244**, 337–339.

Watts, A. B., Bodine, J. H. & Steckler, M. S. 1980 Observations of flexure and the state of stress in the oceanic lithosphere. *J. geophys. Res.* **85**, 6369–6376.

Wiens, D. A. & Stein, S. 1983 Age dependence of oceanic seismicity and implications for lithospheric evolution. *J. geophys. Res.* **88**, 6455–6468.

Wiens, D. A. & Stein, S. 1984 Intraplate seismicity and stresses in young oceanic lithosphere. *J. geophys. Res.* **89**, 11442–11464.

Wiens, D. A. & Stein, S. 1985 Implications of oceanic intraplate seismicity for plate stresses, driving forces and rheology. *Tectonophys.* **116**, 143–162.

Wiens, D. A., DeMets, C., Gordon, R. G., Stein, S., Argus, D., Engeln, J. F., Lundgren, P., Quible, D., Stein, C., Weinstein, S. & Woods, D. F. 1985 A diffuse plate boundary model for Indian Ocean tectonics. *Geophys. Res. Lett.* **12**, 429–432.

Wiens, D., Stein, S., DeMets, C., Gordon, R. & Stein, C. 1986 Plate tectonic models for Indian Ocean 'intraplate' deformation. *Tectonophys.* **132**, 37–48.

Wortel, M. J. R. & Cloetingh, S. A. P. L. 1985 Accretion and lateral variations in tectonic structure along the Peru–Chile Trench. *Tectonophys.* **112**, 443–462.

Zoback, M. L., *et al.* 1989 Global patterns of tectonic stress. *Nature, Lond.* **341**, 291–298.

Discussion

R. Wortel (*Utrecht University, The Netherlands*). Is it not valid to use focal mechanisms as a constraint on modelling basal shear stresses only for a cylindrical earth? Is not the generally adopted idea that old oceanic lithosphere is in compression based on a very limited number of data points?

S. Stein. Although the number of focal mechanisms available for old oceanic lithosphere is quite limited, they indicate compression. Whether these sparse points are an adequate sample is unknown. Given the sparse data, we used a simple two-dimensional model to deal with average properties versus age. Certainly a three-dimensional model would be more appropriate.

L. Fleitout (*École Normale Supérieure, Paris, France*). Could thermal stresses be the cause of the compression in the ocean for all ages and the focal mechanisms that Dr Stein sees do not reflect tectonic mechanisms? Thermal stresses should increase with age and the Pacific Plate can support seamounts for tens of millions of years so I do not see why the oceanic lithosphere should relieve its thermal stresses.

S. Stein. I agree that thermal stress may well contribute. Some features of the focal mechanism data, notably ridge-parallel extension in young lithosphere, are suggestive of some thermal stress contribution. A difficulty, however, is that the ridge-parallel extensional earthquakes are concentrated in the Central Indian Ocean whereas there is no obvious reason to expect thermoelastic stresses to behave this way. For the old lithosphere, the fact that the compressive stresses can be explained without thermal stress does not preclude thermal stress being a contributor.

S. Murrell (*University College, London, U.K.*). Is there any memory in the oceanic lithosphere of the original extension?

S. Stein. The lithosphere preserves evidence of the ridge-normal extension at the axis, via ridge-parallel tectonic fabric which can be used to model the extension. I know of no preserved evidence of the ridge-parallel extension in young lithosphere.

J. Bull (*Southampton University, U.K.*). In the central Indian Ocean Basin there is very clear evidence of the reactivation of pre-existing structures but there is also a problem. On seismic reflection profiles you can see that there are reactivated features going down to 5 or 6 km. So either the faulting has propagated downwards or else there has been nucleation of earthquakes at the brittle–ductile transition and then propagation upwards and reactivation of the pre-existing structures.

S. Stein. I agree that there is often reactivation of pre-existing structures, although we know little about the details.

E. J. W. Jones (*University College, London, U.K.*). I think one has to be very careful about using earthquakes alone to infer the stress régime in the oceans because if high-resolution seismic reflection profiles from the North Atlantic are examined, one finds that in abyssal plains near continental margins there is a great deal of very recent deformation in areas in which there isn't apparently any clear seismicity.

S. Stein. The fact that we see deformation in regions where we haven't observed seismicity may not be a problem, given the very short time interval over which we have instrumental records.

M. Zoback (*Stanford University, U.S.A.*). I am concerned about the big discrepancy between what is normally considered to be the brittle–ductile transition, on the basis of the rheological models, and the maximum depths of the earthquakes. This seems to be almost a factor of two.

S. Stein. The intraplate oceanic earthquakes' focal depths extend well below the brittle–ductile transition predicted from the laboratory rheology. It is possible that the rheology is inappropriate. If this is not the case, it appears that the depth of seismicity is limited not by the brittle–ductile transition, but by the depth to which the lithosphere has adequate strength. In the continents, these two depths are similar, assuming that a quartz rheology is appropriate, so it is hard to tell the difference. The idea of earthquakes below the brittle–ductile transition seems plausible, given that in subduction zones there are earthquakes below the transition.