

IMPLICATIONS OF OCEANIC INTRAPLATE SEISMICITY FOR PLATE STRESSES, DRIVING FORCES AND RHEOLOGY

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ABSTRACT

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Recent studies of earthquakes occurring within the oceanic lithosphere provide valuable information about stresses in the lithosphere, plate tectonic driving forces, and the rheology of the lithosphere, asthenosphere and mantle. Focal mechanisms indicate that oceanic lithosphere older than 35 million years is almost entirely in deviatoric compression. Both compressional and extensional events occur in younger lithosphere. In young lithosphere mechanisms generally indicate compressive stress in the spreading direction or extension oblique to the spreading direction; extension in the spreading direction is not observed. Using the constraint of compression in the spreading direction in old lithosphere, models of the stresses produced by a combination of ridge push and basal drag forces require the magnitude of the drag be less than a few bars for rapidly moving plates and a few tens of bars for slow moving plates. Assuming that basal drag results from mantle return flow, the upper limits on drag can be converted to constraints on the viscosity structure of the asthenosphere and mantle using simple two-dimensional models. If return flow occurs in a mantle of viscosity 10^{22} poise, comparable to the results from glacial rebound studies, the predicted basal drag is too high unless a thin asthenosphere with viscosity less than 10^{19} to 10^{20} poise (depending on flow depth) is present. The intraplate stress data thus are consistent with the idea of oceanic plates largely decoupled from the underlying mantle.

The strength of the lithosphere is constrained by the maximum depth of oceanic intraplate seismicity, which increases with lithospheric age and appears to be bounded by a 700° – 800° C isotherm. This limiting depth is approximately equal to the flexural elastic thickness of the lithosphere and is consistent with experimental olivine rheologies which predict rapid weakening at high temperatures. Similar phenomena are important for estimating the fraction of seismic and aseismic slip on transform faults and in determining the extent of rupture for large trench normal faulting events.

INTRODUCTION

Considerable attention has been focused on the analysis of oceanic intraplate earthquakes. In a general sense, the goal of such studies has been to gain insights

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into the mechanics of plate tectonics comparable to the insights into plate kinematics given by plate boundary earthquakes. A large body of information has been acquired (i.e. Bergman and Solomon, 1980, 1984; Okal, 1983; Wiens and Stein, 1983, 1984). From a global tectonic view, the results have been somewhat disappointing, since it is now clear that many intraplate earthquakes tend to occur on preexisting weak zones and largely reflect localized, rather than global, tectonic processes (Stein, 1979; Bergman and Solomon, 1980). These local processes are significant: for example, magnitude seven intraplate earthquakes demonstrate the intense internal deformation of the Indian plate (Stein and Okal, 1978) and the continued activity of the passive margin of eastern Canada (Stein et al., 1979). Despite their tectonic importance, from a global perspective such local phenomena are "noise" rather than signal. Nonetheless, the intraplate earthquake data set contains valuable information about stresses in the lithosphere, plate driving forces, and the rheology of the lithosphere, asthenosphere and mantle. For this symposium volume, rather than review the intraplate earthquake data, we will use the results to highlight for a nonseismological audience some of the most significant implications for plate tectonics.

SEISMOLOGICAL STRESS CONSTRAINTS

One key set of constraints on plate driving forces are derived from stresses inferred from focal mechanisms. For example, the depth variation of stress within downgoing slabs demonstrates that the large driving force due to subducting slabs, which appears to be the primary determinant of plate velocities, is balanced locally by resistive forces and has little effect on the nonsubducting portions of the plate (Toksöz et al., 1973; Forsyth and Uyeda, 1975; Chapple and Tullis, 1977; Richter and McKenzie, 1978; Richter, 1977, 1979). Similarly stresses inferred from oceanic intraplate earthquake mechanisms can be used to study the forces acting on the plates (Solomon et al., 1975; Mendiguren and Richter, 1978; Richardson et al., 1979). 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.

As previously discussed, the major difficulty in inferring stresses from mechanisms is the concentration of seismicity on pre-existing weak zones and its strong relation to local tectonics. As a result, mechanisms may represent a purely local process and convey no useful stress information for our purposes, or stress directions inferred from the mechanism may be significantly different from the actual ones (McKenzie, 1969; Stein, 1979). As the mechanisms, with all their problems, represent a practical approach to stress estimation, and given the difficulty of isolating the effects of pre-existing weaknesses, the natural approach is to examine the full

available data set to see if useful conclusions can be drawn while avoiding reliance on individual mechanisms as much as possible.

Since the thermal evolution of the oceanic lithosphere controls its mechanical properties and applied forces, seismicity can be studied as a function of age. Figure 1 shows the variation of mechanism type with lithospheric age. The events are tabulated in Wiens and Stein (1984); many mechanisms are from that study, while others are from previous compilations of intraplate seismicity (Sykes and Sbar, 1974; Bergman and Solomon, 1980; Okal, 1983) or individual event studies.

The first basic observation is that lithosphere older than about 35 Ma is in deviatoric compression, as shown by the thrust and strike-slip mechanisms. Mendiguren (1971), Forsyth (1973) and Sykes and Sbar (1974) noted that the compression was approximately in the spreading direction, and related the compression to the "ridge push" force due to the cooling of the lithosphere (Hales, 1969; Lister, 1975; Richter, 1977; Hager, 1978; Parsons and Richter, 1980; Dahlen, 1981; Fleitout and Froidevaux, 1983). The situation in younger lithosphere is more complex and is not fully understood at present. Sykes and Sbar (1971) suggested a progression from extensional stress near the ridges to compressional stress in older

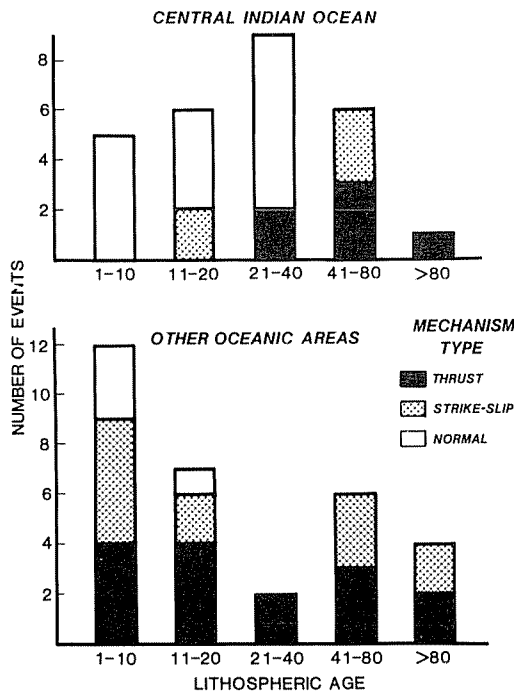


Fig. 1. Mechanism type as a function of lithospheric age for oceanic intraplate earthquakes. The events and references are tabulated in Wiens and Stein (1984). Older oceanic lithosphere is in compression, whereas younger lithosphere has both extensional and compressional mechanisms. Extensional events are located primarily in the Central Indian Ocean.

lithosphere, with the transition occurring near the 20 Ma isochron. The Fleitout and Froidevaux (1983) stress model shows a zone of ridge-normal tension in young lithosphere based on this data. More detailed recent analysis, however, suggests a more complex picture of the seismicity and stresses in young lithosphere (Wiens and Stein, 1984; Bergman and Solomon, 1984). Although, from a global view, both compression and extension are observed in young lithosphere, most of the extensional events occur in the Central Indian Ocean (Stein, 1978; Bergman et al., 1984; Wiens and Stein, 1984). In young lithosphere elsewhere, faulting is heterogeneous. Thus, while normal faulting seems to occur almost exclusively in young oceanic lithosphere, there is little evidence for a general progression from extension to compression except in the Central Indian Ocean.

Seismicity in young lithosphere results from a number of different effects. In particular, the high rate of intraplate seismicity in young lithosphere (Wiens and Stein, 1983) indicates that stress sources localized near the ridges may produce much of this seismicity, rather than plate-wide stresses. Detailed studies of the near-ridge seismicity (Wiens and Stein, 1984) suggests that these effects may include thermo-elastic stress (Turcotte and Oxburgh, 1973) and stresses associated with the geometry of ridge-transform intersections (Fujita and Sleep, 1978). Wiens and Stein (1984)

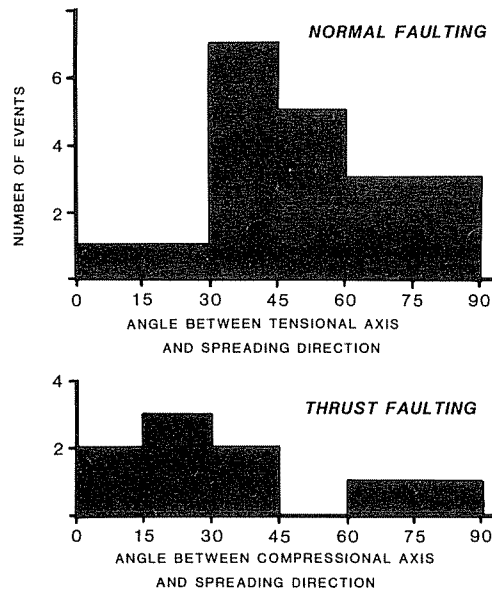


Fig. 2. Histograms of the number of intraplate events as a function of the angle between the horizontal component of principal stress and the spreading direction for normal faulting (top) and thrust faulting (bottom) intraplate earthquakes in young oceanic lithosphere. Tensional axis of normal faulting 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 preferred orientation in the spreading direction.

also found that intraplate earthquakes in young lithosphere tend to be located in small seismically active regions, further suggesting local processes are important in generating this seismicity. However, some information regarding plate-wide stresses can be extracted from the earthquake data. Figure 2 shows histograms of the angle between the horizontal component of the principal stress axis and the spreading direction for normal and thrust faulting intraplate earthquakes in lithosphere younger than 35 Ma. Tensional axes of normal faulting events are oriented at large angles to the spreading direction. Analysis of Indian Ocean normal faulting suggests the tensional axes are generally aligned parallel with the overall orientation of the spreading ridge (which is not perpendicular to the spreading direction in the Central Indian Ocean) (Wiens and Stein, 1984). Because most of the tensional events occurred in the Indian Ocean, it is not certain whether this dataset reflects the general stress field in young lithosphere or stresses in several possibly anomalous regions. However, several normal faulting events in the Eastern Pacific also show extension oblique to the spreading direction. Compressional axes of thrust faulting events show a weak preferred orientation in the spreading direction. Thus tensional stress in the spreading direction is not indicated; instead, the data suggest compressional stresses may be oriented in the spreading direction. This fact, combined with the preponderance of thrust faulting in older lithosphere (Fig. 1), indicates a general compressional stress field in the spreading direction for oceanic lithosphere of all ages. We will use this constraint in developing a simple model for stress in the oceanic lithosphere.

INTRAPLATE STRESS MODEL FOR OCEANIC LITHOSPHERE

To model stress in the oceanic lithosphere, we use an approach derived from one used by Mendiguren and Richter (1978), for the geometry shown in Fig. 3. 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 an input; 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

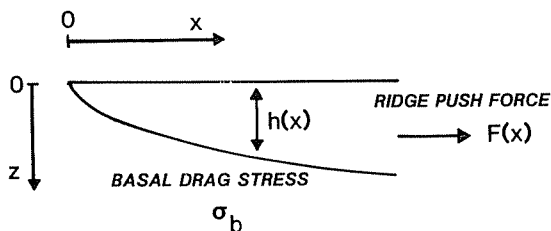


Fig. 3. Geometry assumed in the intraplate stress model.

resistive. The constraint of deviatoric compression for oceanic lithosphere of all ages, along with previously derived expressions for the ridge push force, allows us to estimate the basal drag. As discussed in the previous section, the previously proposed zone of extension to an age of 20 Ma is not supported by the data.

The equilibrium equation in the spreading (x) direction relates the deviatoric stresses and $f(x, z)$, the ridge push body force due to density changes as the lithosphere cools:

$$\frac{\partial \sigma_{xx}(x, z)}{\partial x} + \frac{\partial \sigma_{xz}(x, z)}{\partial z} + f(x, z) = 0$$

Two integrations, the first with respect to x , assuming a stress free “ridge” boundary condition ($\sigma_{xx}(x, h(x)) = 0$), and the second over the thickness of the lithosphere $h(x)$, assuming no shear stress at the top of the plate ($\sigma_{xz}(x, 0) = 0$) yield:

$$\int_0^{h(x)} \sigma_{xx}(x, z) dz + \int_0^x \sigma_{xz}(x, h(x)) dx + \int_0^x \int_0^{h(x)} f(x, z) dz dx = 0$$

This can be written in terms of the vertically averaged stress in the spreading direction:

$$\overline{\sigma_{xx}}(x) = \frac{1}{h(x)} \int_0^{h(x)} \sigma_{xx}(x, z) dz$$

the conventionally defined ridge push force:

$$F(x) = \int_0^x \int_0^{h(x)} f(x, z) dx dz$$

and a constant drag stress at the base of the plate $-\sigma_b$ as:

$$\overline{\sigma_{xx}}(x) = \frac{\sigma_b x - F(x)}{h(x)}$$

The stress free “ridge” boundary condition is a crude way of simulating a “weak” ridge (Sleep and Rosendahl, 1979), which can be used since lithostatic effects have been included in the formulation of the ridge push force. If $\sigma_{xx}(x, h(x))$ is tensional to a given depth, corresponding to a ridge with substantial tensile strength, a zone of extension parallel to the spreading direction in young lithosphere results (Wiens and Stein, 1984), which is not observed in the focal mechanism data.

For a constant half spreading rate v , the stress as a function of age, t , is

$$\overline{\sigma_{xx}}(t) = \frac{\sigma_b vt - F(t)}{h(t)}$$

A more useful form, allowing comparison between different plates, comes from the customary assumption (Solomon and Sleep, 1974; Harper, 1975; Forsyth and Uyeda, 1975) that drag equals the product of absolute velocity u and basal drag coefficient C .

$$\overline{\sigma_{xx}}(t) = \frac{Cuvt - F(t)}{h(t)}$$

Thus a drag force depending on absolute velocity is applied over an area proportional to the spreading rate. For our examples, we make the simplifying assumption that $v = u$, spreading rate equals absolute velocity (the ridge is fixed with respect to the mantle), thus making the integrated drag force at a given lithospheric age proportional to velocity squared.

Both lithospheric thickness (defined by the thickness above an isotherm T_1) and the magnitude of the ridge push force depend on the plate's thermal structure. For a cooling halfspace model (McKenzie, 1969):

$$T(z, t) = T_m \operatorname{erf}\left(\frac{z}{2\sqrt{\kappa t}}\right)$$

and:

$$F(t) = g\alpha\rho_m T_m \kappa t$$

(Lister, 1975; Parsons and Richter, 1980), where ρ_m and T_m are the mantle reference density and temperature, and κ is thermal diffusivity. Figure 4 shows intraplate stress as a function of age and drag coefficient, for a halfspace model ($T_1 = 1250^\circ\text{C}$, $T_m = 1300^\circ\text{C}$, $\kappa = 8 \times 10^{-7} \text{ m}^2/\text{s}$, $\alpha = 8.3 \times 10^{-5} \text{ }^\circ\text{C}^{-1}$, $\rho = 3.33 \text{ g/cm}^3$) for a fast moving plate (10 cm/yr) (upper left) and a slower moving plate (1 cm/yr) (lower left). If the drag coefficient is zero, the deviatoric stress resulting from ridge push alone is purely compressive ($\overline{\sigma_{xx}} < 0$) and varies as the square root of age since force

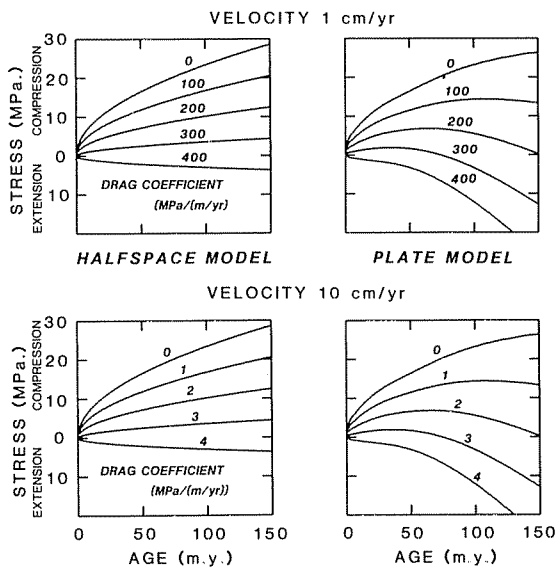


Fig. 4. Plots of intraplate stress as a function of lithospheric age and assumed basal drag coefficient for slow moving (top) and fast moving (bottom) plates. Figures on the left assume a halfspace thermal model (McKenzie, 1969); figures on the right assume a plate model (Parsons and Sclater, 1977). The observation of compressional stresses in oceanic lithosphere places an upper bound on the drag coefficient.

increases linearly while lithospheric thickness increases as the square root. A similar increase is shown by Dahlen (1981). 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 our constraint that the lithosphere is in compression a rapidly moving plate must be acted on by a drag coefficient less than about 3.5 MPa/(m/yr), corresponding to a basal stress less than 0.35 MPa (3.5 bar). Comparable values were found by Hager and O'Connell (1981) and Fleitout and Froidevaux (1983). In contrast, if drag coefficients are allowed to vary, a plate moving a factor of ten more slowly can remain in compression with a drag coefficient 100 times greater (which corresponds to a basal stress ten times greater).

If the lithosphere is modeled as a cooling plate of thickness α , temperature is given by (Parsons and Sclater, 1977):

$$T(z, x) = T_m \left[\frac{z}{a} + \frac{2}{\pi} \sum_{n=1}^{\infty} \frac{1}{n} \sin \left[\frac{n\pi z}{a} \right] \exp \left[\frac{-\beta_n x}{a} \right] \right]$$

and the ridge push force is (Parsons and Richter, 1980):

$$F(t) = \frac{g\alpha\rho_m T_m a^2}{6} \left[1 - \frac{12}{\pi^2} \sum_{n=1}^{\infty} \frac{(-1)^{n+1}}{n^2} \exp \left[\frac{-\beta_n vt}{a} \right] \right]$$

where $\beta_n = (R^2 + n^2\pi^2)^{1/2} - R$ and $R = (va/2\kappa)$. In this model, since 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 ($C = 0$), deviatoric stress becomes more compressive with age, but for larger values of the drag coefficient the stress reaches a compressional maximum, becomes less compressive, and eventually becomes extensional. Figure 4 (right) shows results for $\alpha = 125$ km, and velocities of 1 and 10 cm/yr. For the plate models, the maximum drag coefficient allowing compression is less (about half) of the value for the halfspace models.

Intraplate earthquake focal mechanisms (Fig. 1) show compression in old lithosphere, and thus place an upper bound on the basal drag. This bound (about 2–4 MPa/(m/yr)) comes from consideration of a fast moving plate, since a slower plate can be entirely in compression for greater drag values. Compression in old lithosphere indicates that the integrated ridge push force dominates drag for all ages. If the drag coefficient is constant for all plates the stress and thus the difference between the ridge push and drag forces depends heavily on plate velocity, since the integrated drag force at a given lithospheric age varies as the velocity squared. In this case, basal drag is a much less significant resisting force for slow moving plates than for fast moving ones. Alternatively, if the drag coefficient differs between plates, the drag coefficient and thus basal stresses can be much greater for slow moving plates without violating the focal mechanism data. Since at present plates with significant fractions of continental area move more slowly than oceanic plates, increased drag

under continental regions is often proposed (Solomon et al., 1975; Forsyth and Uyeda, 1975; Chapple and Tullis, 1977), although Gordon et al. (1979) suggest that in the past continental plates have moved as rapidly as oceanic ones. As discussed in the following section, some variation in drag coefficient would be expected, both due to different temperature structure and different flow directions. Whether this variation is significant or not is unknown.

The stress model we have used is an extremely simple two-dimensional one for an "average plate" in which only the overall compression in the direction of plate motion is modeled. Application to an individual plate requires specification of absolute velocity, spreading rate, boundary configuration (Solomon et al., 1975; Richardson et al., 1979) and localized stress effects. The model presented here is based on global observations; this "average oceanic plate" (summarized in Fig. 5) is deliberately schematic and does not attempt to describe the detailed structure of any individual plate.

VISCOSITY STRUCTURE

Some constraints on mantle viscosity structure can be derived from intraplate earthquake mechanisms. Clearly, earthquake mechanisms are not a primary tool for estimation of mantle viscosity: nonetheless, under certain assumptions interesting inferences can be drawn. The analysis assumes a commonly used model in which plate motions are maintained largely by thermal buoyancy forces acting on the lithosphere, and some mantle flow can be treated as the return flow required to balance the mass transport due to plate motions (Elsasser, 1969; Schubert and Turcotte, 1972; Richter, 1977; Richter and McKenzie, 1978; Schubert et al., 1978;

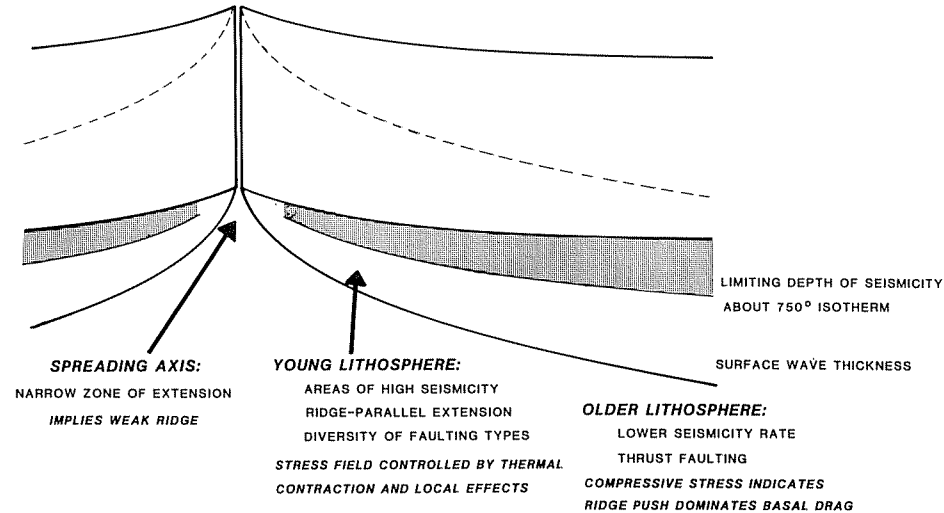


Fig. 5. Schematic diagram relating seismicity observations to the tectonics of the oceanic lithosphere.

Hager and O'Connell, 1979, 1981; Parsons and Richter, 1981). An alternative model assumes that convective mantle flow provides an active plate motion driving mechanism (McKenzie, 1969; Morgan, 1971; Davies, 1977, 1978; Peltier, 1980; Jarvis and Peltier, 1982). Discussion of the various models is beyond our scope; it is important to note that the analysis presented here assumes a given model.

In the model, since the drag on the base of a plate is due to motion over the viscous mantle, the fact that earthquake mechanisms are compressive in old lithosphere constrains the viscosity structure (Mendiguren and Richter, 1978). As in the previous section, it is useful to consider a simple two dimensional geometry where in the interior portion of a plate moving over a fluid mantle, the mass flux due to the moving plate is balanced by a return flow at depth.

The simplest case is one in which the underlying fluid has uniform Newtonian viscosity. For a plate of thickness h , overlying a fluid layer with depth d and viscosity η , the drag coefficient is:

$$C = 2\eta(3s + 2)/d$$

where $s = h/d$ is the normalized plate thickness (Schubert and Turcotte, 1972; Richter and McKenzie, 1978).

Figure 6 shows drag coefficients for various values of viscosity and flow depth and a 125 km thick lithosphere. The focal mechanism data require that the drag coefficient be less than about 2–4 MPa/(m/yr) for the fast moving plates (Fig. 4). If return flow occurs within the upper mantle, a depth of 700 km, mantle viscosity must be less than 2×10^{20} poise. (By comparison, Mendiguren and Richter's (1979) higher viscosity bound of 3×10^{21} poise for 700 km flow results from requiring only that a slow moving (1 cm/yr) plate be in compression.) The whole mantle flow, down to a depth of 2900 km, requires viscosity less than 1×10^{21} poise. Both values are

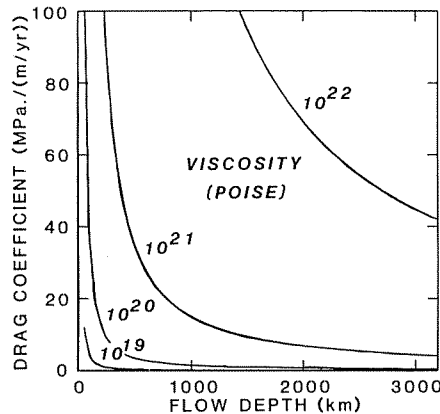


Fig. 6. Basal drag coefficients as a function of flow depth and mantle viscosity for single layer flow. Upper bounds on drag coefficients constrain the mantle viscosity.

significantly lower than the $1-5 \times 10^{22}$ poise average mantle viscosity estimated from glacial rebound (Cathles, 1975; Peltier, 1980), earth rotation (Yuen et al., 1982) and LAGEOS satellite orbits (Peltier, 1983). Richter and McKenzie's (1978) analysis of the depth and gravity anomalies, produced by the nonhydrostatic pressure gradient required to drive the return flow (Schubert and Turcotte, 1972; Schubert et al., 1978), also found a rather low upper viscosity bound of 3×10^{20} poise for flow in a 700 km thick uniform viscosity channel.

Viscosities inferred from intraplate stresses and gravity and depth anomalies can be reconciled with those inferred from glacial rebound (Cathles, 1975) by assuming that the plate is underlain by a thin low-viscosity layer, as commonly assumed from glacial rebound data (Cathles, 1975). Richter and McKenzie (1978) found that such a two-layer model allowed higher mantle viscosities while not violating the depth and gravity constraints. A similar model has been proposed by Hager and O'Connell (1979). In such models, the low-viscosity layer acts to decouple the plates, but only a fraction of the return flow occurs in the layer. We find that such a thin asthenosphere model satisfies the intraplate stress data while allowing acceptably high mantle viscosities.

For a two-layer model, with upper and lower layers of thickness d_1 and d_2 and Newtonian viscosities η_1 and η_2 , underlying a plate of thickness h , the drag coefficient is:

$$C = \frac{\left[\frac{\eta_2}{d_2} \right] \left[1 - \frac{P}{2} (Mr^2 + 2r + 1) \right]}{1 + Mr}$$

where P , the normalized pressure gradient is:

$$P = \frac{-6[2s(1 + Mr) + (1 + 2r + Mr^2)]}{1 + 4Mr + 6Mr^2 + 4Mr^3 + Mr^4}$$

and $M = \eta_1/\eta_2$, $r = d_1/d_2$, $s = h/d_2$ (Richter and McKenzie, 1978).

Figure 7 (left) shows the drag coefficient as a function of upper layer depth and viscosity, for a lower layer with viscosity 3×10^{21} poise. The lower-layer thickness is adjusted so that the combined depth of the two layers and the 100 km thick lithosphere is 700 km (top left) or 2900 km (bottom left). Acceptable values of drag coefficient (less than 2–4 MPa/(m/yr)) occur for asthenospheric (upper layer) viscosities less than about 6×10^{19} poise for upper mantle flow and 2×10^{20} poise for the whole mantle case. The precise viscosity depends on the asthenospheric thickness. For a more viscous mantle, 1×10^{22} poise (right), the maximum acceptable asthenospheric viscosities decrease. In the two layer models, the low viscosity layer significantly reduces the effect of the total flow depth such that either upper mantle or whole mantle flow are acceptable. Richter and McKenzie's model, upper mantle flow with viscosity 2.5×10^{21} poise, overlain by an 85 km thick asthenosphere with viscosity 5×10^{18} poise, falls within the acceptable range. On the other

hand, Cathles' (1975) model of a 75 km thick asthenosphere with viscosity 4×10^{20} poise over a 1×10^{22} poise mantle predicts too high a drag coefficient.

These calculations are crude and provide fairly general estimates of viscosity structure. In particular, we have assumed two dimensional return flow, an assumption reasonable in the case of large plates overlying a low-viscosity asthenosphere but less valid for smaller plates or a higher-viscosity asthenosphere (Hager and O'Connell, 1979). Even if the drag coefficient is spatially uniform, the apparent drag would vary when the return flow is not antiparallel to the plate's absolute motion. Additionally, the temperature dependence of viscosity should yield structures more complicated than simple layers, whose description requires detailed thermal and mechanical modeling of the entire plate (Schubert et al., 1978). Nonetheless, it is gratifying that the intraplate stresses derived from focal mechanisms can be explained by conventional ideas of mantle viscosity, and in particular provide further support for low basal drag due to a low-viscosity layer below the oceanic plates. Such decoupling is consistent with the lack of correlation between oceanic plate area and absolute velocity (Forsyth and Uyeda, 1975), kinematic driving force studies (Forsyth and Uyeda, 1975; Chapple and Tullis, 1977; Hager and O'Connell, 1981; Carlson et al., 1983), gravity and depth anomalies (Richter and McKenzie, 1978) thermal and mechanical models (Schubert et al., 1978) and glacial rebound (Cathles, 1975). The viscosity bounds discussed in this paper have been derived for oceanic regions and may not be comparable to those derived from continental regions.

As emphasized previously, these conclusions result from the common assumption of mantle return flow controlled largely by the motions of the plates. The alternative model, mantle flow as an active driving mechanism (in which strong coupling

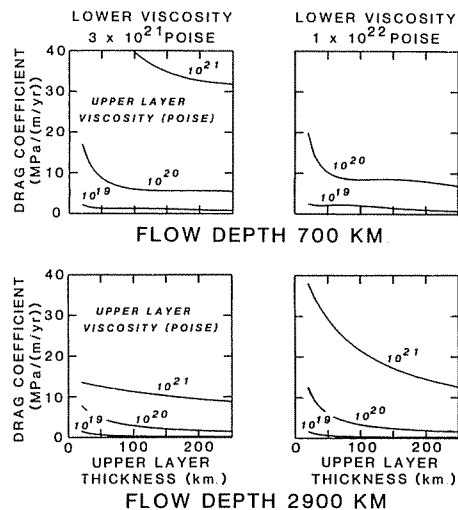


Fig. 7. Basal drag coefficients as a function of upper layer thickness and viscosity for two values of lower-layer viscosity and total flow depths of 700 km (top) and 2900 km (bottom).

between the plate and the mantle explains the lack of correlation between plate area and absolute velocity) has quite different consequences for basal stress (Hanks, 1977; Davies, 1978). The assumed kilobar basal stresses seem high, given that earthquake mechanisms (Stein et al., 1979) and other data (Raleigh and Evernden, 1981) favor intraplate stresses of a few hundred bars. However, kilobar stresses can be interpreted as more consistent with laboratory rock mechanics experiments, as discussed later.

EARTHQUAKE DEPTHS AND LITHOSPHERIC RHEOLOGIES

While the stress field in the oceanic plates helps constrain the viscosity of the mantle, the depths of intraplate earthquakes constrain the rheology of the lithosphere (Wiens and Stein, 1983; Chen and Molnar, 1983). The maximum depth of oceanic intraplate seismicity increases with lithospheric age (Fig. 8). Isotherms derived from a plate cooling model (Parsons and Sclater, 1977) suggest that the maximum depth of seismicity is temperature controlled and that the limiting isotherm is approximately 700° – 800° C. This depth is approximately equal to the elastic thickness determined from studies of lithospheric flexure (Watts et al., 1975; 1980) but much less than the depth to the low-velocity zone (“seismic thickness”) inferred from Rayleigh wave dispersion studies (Leeds et al., 1974; Forsyth, 1975). As the elastic thickness measures the response of the lithosphere to long term loads and the seismic thickness represents its response to much shorter term loads, it

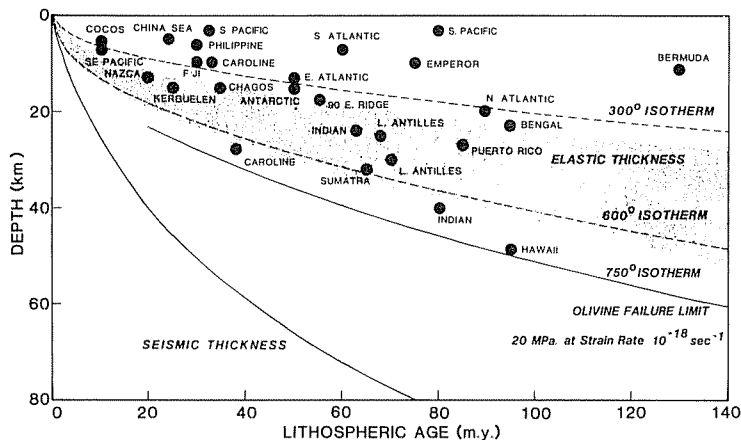


Fig. 8. Well-constrained intraplate earthquake depths tabulated in Wiens and Stein (1983) shown on a depth–age plot of the oceanic lithosphere. Isotherms (in $^{\circ}$ C) shown are calculated from a lithospheric cooling model (Parsons and Sclater, 1977). Stippled region denotes range of estimates of the flexural elastic thickness (Watts et al., 1980). The seismic thickness is from Rayleigh wave dispersion data by Leeds et al. (1974). The failure limit is the lower limit at which 20 MPa (200 bar) deviatoric stress can be sustained with a dry olivine rheology (Goetze and Evans, 1979) and a strain rate of 10^{-18} s^{-1} .

appears that intraplate earthquakes occur primarily where the lithosphere can sustain long term loads.

This observed temperature dependence of the limiting depth of seismicity can be compared to the results of laboratory experiments on the strength of rocks (Goetze and Evans, 1979; Brace and Kohlstedt, 1980). At low temperatures, corresponding to shallow depths, failure occurs by brittle fracture and strength increases linearly with depth since it depends only on effective stresses (stress minus pore pressure). At higher temperatures strength is controlled by ductile flow laws, which depend on temperature, strain rate, and mineral type. Olivine flow laws (Goetze and Evans, 1979) show rapid decreases in strength at high temperatures, suggesting that the maximum depth of seismicity occurs when temperatures are too high to allow the lithosphere to sustain earthquake causing stresses.

Fig. 9a shows strength as a function of temperature using the formulation of Brace and Kohlstedt (1980), while Fig. 9b shows the distribution of oceanic intraplate events versus temperature at the focal depth. In the brittle zone, the strength increases approximately linearly with temperature, with the slope depending on the age of the lithosphere, the pore pressure, and whether the stress is tensional or

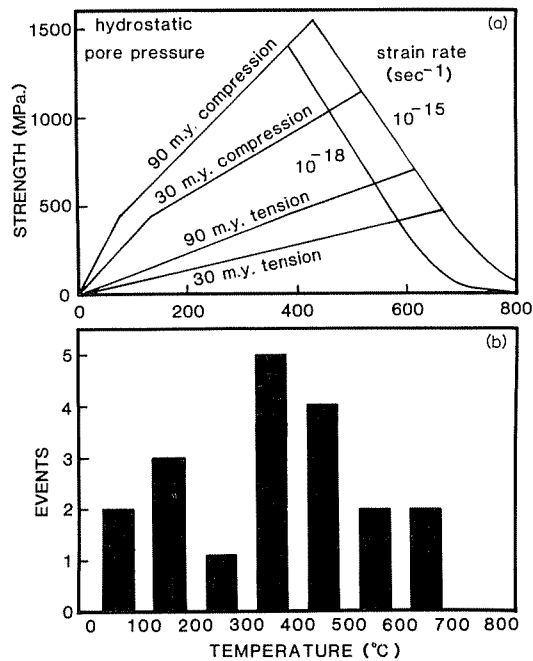


Fig. 9. a. Lithospheric strength as a function of temperature using the formulation of Brace and Kohlstedt (1980) for two ages of oceanic lithosphere and two possible strain rates. b. Number of intraplate earthquakes as a function of temperature at the focus. The earthquakes are those shown in Fig. 8 excluding marginal basin events. The highest seismicity seems to correspond to the strongest part of the lithosphere.

compressive. In the ductile region a rapid decrease in strength with temperature is observed. The highest rate of intraplate seismicity seems to coincide with the area of highest lithospheric strength. Such a relationship has been observed for a much larger data set of continental earthquakes by Meissner and Strehlau (1982). Note that (if the assumed rheology and pore pressure is appropriate) seismicity is not limited to the region of brittle deformation: it appears to be strength rather than the mode of deformation which controls the occurrence of earthquakes (Wiens and Stein, 1983).

One difficulty in applying the rheology to the earth is that the appropriate strain rate is unknown. Wiens and Stein (1983) estimated the intraplate strain rate from the seismic moment release of intraplate earthquakes. Two strain rates are shown in Fig. 9; the lower value, 10^{-18} s^{-1} was calculated for a global average release rate while the upper value, 10^{-15} s^{-1} , represents an upper bound calculated for areas of active intraplate deformation. While the method of estimation is somewhat imprecise, the dependence of the rheology on strain rate is only linear while the dependence on temperature is exponential. The observed limiting isotherm, about 750°C , corresponds to a limiting strength of 200 bars (20 MPa) for a strain rate of 10^{-18} s^{-1} .

Several other difficulties arise in applying the rheology to the earth. The presence of water significantly weakens olivine (Post, 1977). Whether a wet or dry rheology is appropriate is not known. A wet rheology would move the strain rate curves on Fig. 9 toward lower temperatures. Also, the oceanic lithosphere at depth may not be pure olivine; the presence and effect of other phases on the rheology is also not known. Finally, the fact that the rheological results come from laboratory experiments conducted at strain rates approximately ten orders of magnitude higher than for intraplate areas may lead to errors difficult to assess. This possibility is suggested by the fact that rock strengths predicted by laboratory derived rheologies are much higher than lithospheric stresses inferred from earthquake stress drops for reasons as yet unclear (Kirby, 1980; Raleigh and Evernden, 1981). Our results provide no new insight on the actual values of lithospheric stress: the fact that flow laws can be used to predict the temperature dependence of seismicity is very valuable even if the actual values of the inferred stresses are in doubt.

The observation that the depth of seismicity is limited by the temperature dependence of lithospheric strength has significance for a variety of tectonic processes, two of which are worthy of brief discussion. Several investigators have compared the cumulative seismic moment observed along transform faults with the expected cumulative moment predicted from the spreading rate and fault area (Brune, 1968; Kanamori and Stewart, 1976; Burr and Solomon, 1978; Stewart and Okal, 1983). The depth of faulting on transforms is a crucial parameter in such studies, as it and the transform length determine the fault area, which in turn is used to convert the measured seismic moment to an inferred seismic slip rate. Recently, Engeln et al. (1984) studied the depths of seismicity on Atlantic transforms and concluded that the limiting temperature of seismicity was considerably lower than

750°. This observation not only has implications for rupture area calculations but may also indicate a significant rheological difference between intraplate areas and transforms.

A temperature controlled depth of faulting also has implications for the tectonic interpretation of large tensional earthquakes seaward of trenches. Earthquakes of this type include the 1933 Sanriku (Kanamori, 1971), 1965 Rat Island (Stauder, 1968; Abe, 1972a), 1969 Lesser Antilles (Stein et al., 1982), 1970 Peru (Abe, 1972b), and 1977 Indonesian (Stewart, 1978) events. The location and large moment of these events has prompted proposals that they represent tensional failure of the subducting lithosphere due to the weight of the downgoing slab (Kanamori, 1971), possibly triggered by a major earthquake on the shallow thrust zone (Abe, 1972a). An alternative view is that the tensional events are merely large examples of flexural earthquakes resulting from plate bending at subduction zones (Chapple and Forsyth, 1979). One possible method of discriminating between these models is the depth of failure. The lack of deep aftershocks has been used to argue that faulting did not penetrate the lower lithosphere (Chapple and Forsyth, 1979; Forsyth, 1982; Stein et

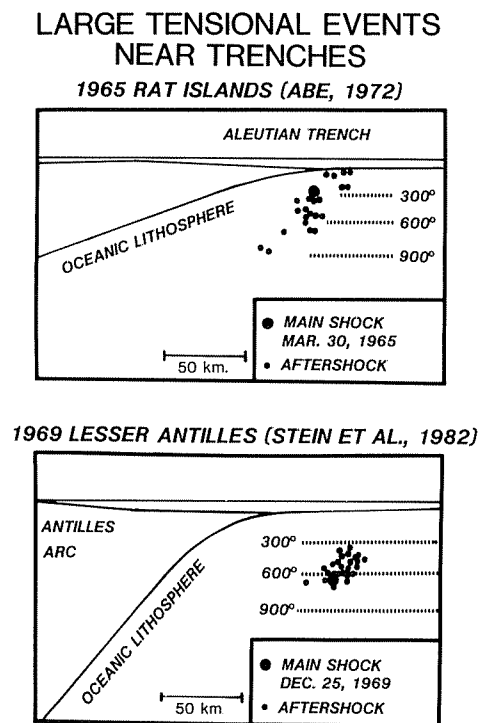


Fig. 10. The aftershock zones of two large normal-faulting events near subduction zones with isotherms from the Parsons and Sclater (1977) plate model. The depth of the aftershocks may be limited by thermal weakening; thus they may not accurately reflect the depth of dislocation of the main shocks.

al., 1982). However, the intraplate earthquake data suggests that the lower lithosphere may be too hot and thus too weak to permit seismic failure. The absence of deep seismicity may not indicate whether a decoupling event took place, if deformation in the lower lithosphere occurred largely through ductile flow.

Figure 10 shows the location of the aftershocks of the 1965 Rat Island (Abe, 1972a) and 1969 Lesser Antilles (Stein et al., 1982) earthquakes along with isotherms from a plate cooling model (Parsons and Sclater, 1977). A limiting strength argument like that used for intraplate areas, with Goetze and Evans' (1979) 10^{-15} s^{-1} estimate of the strain rate at subduction zones, suggests that seismicity should not occur below about the 850°C isotherm. The aftershock data are grossly consistent with this idea, although difficulties in obtaining accurate aftershock depths and locations may make the correlation meaningless. We do not claim to show that the aftershock zones are temperature controlled; we merely point out that even if accurate depths can be obtained for the large normal faulting events the depths may not indicate whether decoupling has taken place.

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