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On The Initiation of Subduction

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Estimates of shear resistance associated with lithospheric thrusting and convergence represent lower bounds on the force necessary to promote trench formation. Three environments proposed as preferential sites of incipient subduction are investigated: passive continental margins, transform faults/fracture zones, and extinct ridges. None of these are predicted to convert into subduction zones simply by the accumulation of local gravitational stresses. Subduction cannot initiate through the foundering of dense oceanic lithosphere immediately adjacent to passive continental margins. The attempted subduction of buoyant material at a mature trench can result in large compressional forces in both subducting and overriding plates. This is the only tectonic force sufficient to trigger the nucleation of a new subduction zone. The ubiquitous distribution of transform faults and fracture zones, combined with the common proximity of these features to mature subduction complexes, suggests that they may represent the most likely sites of trench formation if they are even marginally weaker than normal oceanic lithosphere.

INTRODUCTION

Plate tectonic theory has expanded considerably from its original, essentially descriptive, form to incorporate a greater emphasis on understanding the associated dynamics as well as the nature in which plate boundaries evolve. The initiation of subduction represents a fundamental issue in both of these areas. Because plate motions are induced primarily by the negative buoyancy of subducting oceanic lithosphere, and because convergent margins occasionally evolve into nonconvergent margins, trench formation must represent a critical element in the dynamics of plate tectonics.

The most accepted model of trench formation, referred to as passive margin failure, envisions the nucleation of subduction as a direct consequence of the cooling of oceanic lithosphere. Because the oldest (and presumably densest) seafloor is adjacent to Atlantic-type, or passive, continental margins, the concept of a "foundering" passive margin is often considered to be intuitive. Factors commonly cited in support of passive margin failure are (1) availability of negatively buoyant oceanic lithosphere, (2) massive sediment loading typical of many of these margins, (3) potential compression in the adjacent oceanic lithosphere due to elevational discontinuities at the ocean-continent boundary, (4) highly faulted nature inherited from the earlier rift phase of passive margin evolution, and (5) the abundance of Andean-type margins.

Yet consider the following.

1. The existence of negatively buoyant Mesozoic seafloor [e.g., *Oxburgh and Parmentier*, 1977] demonstrates that buoyancy considerations alone cannot adequately characterize the process of trench formation. In addition, oceanic lithosphere probably does not evolve toward greater thickness and density after 100 m.y. [e.g., *Parsons and Sclater*, 1977]. Thus Atlantic-type margins may be no more buoyantly unstable than many other oceanic regions of the world.

2. Sediment loading merely induces bending stresses, which are compressive in the upper lithosphere, tensional in the lower lithosphere, and integrate to zero over the plate thickness. Incipient lithospheric thrusting and convergence requires net horizontal compression, which cannot be generated from local vertical loads.

3. Oceanic lithosphere adjacent to passive margins would be subjected to compression associated with ocean-continent elevational discontinuities only if continental masses are elastically non-self-supporting and push against the nearby seafloor during viscous relaxation. It is not clear that this is the case [e.g., *Turcotte et al.*, 1977].

4. Other tectonic settings, such as transform faults and fracture zones, may be considerably more fractured than passive margin interfaces (which may anneal subsequent to rifting).

5. The abundance of Andean-type margins may only reflect the stability of these margins. Once established, only a continent-continent collision or the attempted subduction of a ridge might eliminate such a boundary.

6. There is no obvious example in the geologic record of the direct transformation of an Atlantic-type margin into an Andean-type margin.

7. The existence of intraoceanic trenches, such as the Mariana, Aleutian, and Scotia arcs, that clearly did not migrate to their present positions from any continental margin, indicates that even if passive margin failure is a legitimate geologic phenomenon, it cannot represent a complete paradigm for trench formation. Additional, intraoceanic, nucleation sites are necessary.

Subduction nucleation along transform faults and fracture zones has been proposed on the basis of (1) geophysical studies of backarc basins [*Cooper et al.*, 1977; *Casey and Dewey*, 1984; *Uyeda and Ben-Avraham*, 1972], (2) plate motion reconstructions [*Hilde et al.*, 1977], (3) oblique convergence exhibited by young subduction complexes [*Fitch*, 1972], (4) contemporary examples of (potential) trench formation [*Fukao*, 1973; *Purdy*, 1975; *Hegarty et al.*, 1982; *Ruff et al.*, 1989], (5) ophiolite studies [*Casey and Dewey*, 1984], and (6) the proposal that trench formation along a former transform fault leads to the creation of boninites, a rock type generally found among the youngest trench-related volcanics [*Hawkins et al.*, 1984]. Spreading ridges have also been proposed as sites of trench formation [*Jones*, 1971; *Casey and Dewey*, 1984].

Using a model of driving versus resisting forces applied to a lithospheric plate [e.g., *McKenzie*, 1977], we establish lower bounds on the force necessary to promote trench formation in each of the proposed nucleation environments, as well as in normal oceanic lithosphere. These lower bounds are contrasted to magnitude estimates of tectonic forces proposed to trigger the nucleation of subduction. We conclude that regardless of nucleation

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environment, inplane compressional forces resulting from ridge push, ocean-continent elevational discontinuities, and lithospheric basal shear are all insufficient in magnitude for trench formation. The only adequate inplane compressional forces are those that result when trench congestion occurs in response to the attempted subduction of buoyant material at a mature subduction complex. Theoretical considerations, as well as deformational modeling of Central Indian Ocean lithosphere, indicate that the necessary stress levels can result from trench congestion. Due to their common proximity to mature subduction complexes, transform faults and fracture zones would represent preferential sites of trench formation if they are even marginally weaker than normal oceanic lithosphere. We conclude this paper with a review of the relevant geologic and marine geophysical literature necessary to test our conclusions.

Mechanics of Trench Formation

In a quantitative investigation of passive margin failure, *Cloetingh et al.* [1989, and references therein to their previous work] concluded that cooling passive margins increase in strength sufficiently to prevent foundering in response to increasing density. They proposed that unless failure occurs within the first 20 m.y. of margin evolution, it will not occur at all. These studies, however, did not explicitly address passive margin failure from a perspective of driving versus resisting forces [e.g., *McKenzie*, 1977]. "Failure" in the Cloetingh et al. models occurs in response to bending stresses induced by sediment loading and cannot represent a precursor to the incipient convergence necessary for trench formation. The latter requires that the entire lithosphere be in a state of compression. "Failure" in the models of Cloetingh et al. represents the development of a "plastic hinge" [e.g., *Turcotte and Schubert*, 1982] and should not be confused with trench formation.

McKenzie [1977] emphasized that buoyancy considerations alone are insufficient to characterize trench formation. He proposed that negative buoyancy is compensated by shear resistance along the potential thrust plane and flexural resistance to lithospheric bending into this plane and that an additional, externally applied (i.e., nonbody) compressional force is required to trigger subduction. McKenzie tentatively proposed that ocean-continent elevational discontinuities at passive margins might play a role. If the resulting underthrusting is rapid enough to postpone thermal assimilation to depths of a few hundred kilometers, enough negative buoyancy accumulates to sustain convergence, and a mature subduction zone develops [*McKenzie*, 1977]. In addition to the inhibiting forces considered by McKenzie, resistance to trench formation might also include the necessity of overcoming the fracture strength of rocks in establishing a fault plane, a "plowing" resistance to the "leading edge" of the underthrusting plate, and, if relatively young lithosphere is involved, positive buoyancy.

Driving Forces

Ridge push forces arise from the negative buoyancy associated with cooling of the oceanic lithosphere and are commonly assumed to trigger passive margin failure. Calculations based on the cooling plate thermal model produce ridge push estimates that asymptotically approach a value of about 3×10^{12} N/m [*Parsons and Richter*, 1980]. Estimates based on the semi-infinite half-space cooling model increase indefinitely with age in a linear fashion, from zero at the ridge to a value of approximately 7×10^{12} N/m for 200 m.y. old lithosphere [*Parsons and Richter*, 1980]. For lithospheric ages less than 80 m.y., both thermal models predict the same ridge push forces.

McKenzie [1977] postulated that stresses arising from differences in elevation between continents and oceans might promote trench formation. The effectiveness of elevational discontinuities in inducing compressional stresses within the oceanic lithosphere would reflect the degree to which the continental (i.e., elevated) landmass is elastically non-self-supporting; the continent will push on the ocean only during attempted viscous relaxation [e.g., *McKenzie*, 1977]. Under optimal circumstances the magnitude of this force would not exceed 3×10^{12} N/m [*Turcotte and Schubert*, 1982]. Thus it would not exceed the compression induced if old, thermally mature oceanic lithosphere existed in the same location.

Lithospheric basal shear, which results from relative motion between plates and the underlying mantle, can potentially induce inplane compressional forces into the lithosphere [e.g., *Phillips*, 1990]. These "drag" forces are unlikely to be very large for the Earth, however, due to the sublithospheric low-viscosity asthenosphere. Both *Forsyth and Uyeda* [1975] and *Chapple and Tullis* [1977] concluded that plate motions are not significantly affected by basal shear. Earthquake studies, as well as lithospheric thermal modeling, indicate that this stress is less than 1 MPa [*Melosh*, 1977; *Wiens and Stein*, 1985]. Inplane forces induced by basal shear stresses increase linearly with distance of application. A basal shear stress of 1 MPa potentially results in an inplane compressional force of 1×10^{12} N/m per 1000 km of application.

Resisting Forces

Flexural resistance represents the force that must be applied to oceanic lithosphere if the deflection necessary to establish a trench is to be maintained. *McKenzie* [1977] estimated that this resistance would reach 8×10^{12} N/m before the necessary deformation was accomplished. This value was based on an assumed elastic lithosphere 80 km thick, whereas more recent studies indicate that a thickness of 40 km is probably more appropriate [*McNutt and Menard*, 1982; *Wiens and Stein*, 1983], implying that the flexural rigidity incorporated by McKenzie may have been excessive by nearly an order of magnitude.

Davies [1980] calculated that even if a subducting slab presents a blunt leading edge as it penetrates the mantle, the resistance concentrated at the slab tip would only be about 2×10^{12} N/m. In the trench formation process the leading edge of underthrusting oceanic lithosphere should be tapered (e.g., Figure 1), and, initially, only the low-viscosity asthenosphere is encountered, suggesting that any "plowing" resistance to the leading edge would be even smaller than Davies' estimate.

Although it might seem that the most significant resistance to trench formation would be the necessity to "break" the lithosphere, it is commonly assumed in lithospheric modeling that the intrinsic strength of near-surface rock is negligible and the strength of the upper lithosphere is limited by frictional sliding properties [e.g., *Brace and Kohlstedt*, 1980; *McNutt and Menard*, 1982]. The justification for this is the argument that, given a large enough area, suitably oriented preexisting fractures are likely to exist, limiting the regional strength. The rapid cooling of young oceanic lithosphere in the near-ridge environment might result in an adequate distribution of fractures provided that annealing does not occur. The important point here is that the events responsible for fracturing and weakening the lithosphere are not necessarily associated with, nor do they immediately precede, the process of trench formation.

We constrain the minimum resistance to trench formation by considering only the shear resistance encountered by incipient motion along a through-going lithospheric fault. We assume that

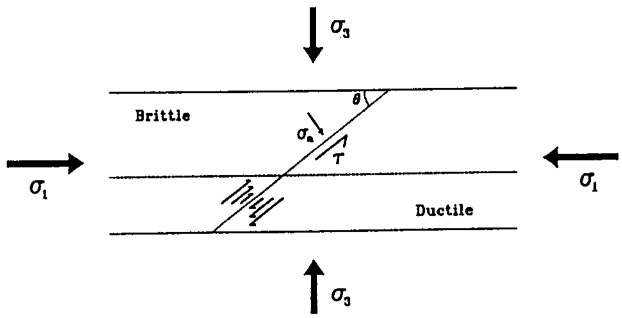

Fig. 1. Illustration of the geometry involved in determining shear resistance along an incipient fault plane within normal oceanic lithosphere. Principal tectonic stresses are represented by σ_1 and σ_3, the resultant normal and shear stresses as resolved along the fault plane by σ_n and τ, and the fault plane dip by θ. The half-arrows in the ductile portion of the lithosphere indicate that relative motion is accommodated by distributed deformation.

the plate is already "broken" and that the other resistive forces described above are negligible. Because shear stresses probably contribute the greatest resistance to trench formation, our methodology should provide useful lower bounds on forces involved in this process (near-ridge environments represent an important exception, see discussion below). Consideration of the potential for anomalous weakening uniquely characteristic of each proposed nucleation environment results in further minimization. Total resistance to lithospheric thrusting and convergence is determined by integrating the compressional shear resistance as a function of depth along the fault plane. This integral represents a lower bound on the force necessary to initiate subduction. Initially, we estimate the shear resistance of normal oceanic lithosphere, both to illustrate methodology and to provide a point of reference for further analysis.

Normal Oceanic Lithosphere

If the resistance to thrusting in the upper, brittle portion of the lithosphere is governed solely by frictional sliding behavior, shear resistance (τ) is of the general form

$$\tau = \tau_0 + \mu(\sigma_n - P_f) \tag{1}$$

where σ_n is the normal stress acting across the fault plane, P_f is the ambient pore pressure, μ is the coefficient of friction, and τ_0 is a constant. *Byerlee* [1978] has determined that with few exceptions, nearly all rock types obey the following frictional sliding relationship

$$\begin{aligned} \tau &= 0.85(\sigma_n - P_f) && \sigma_n < 200 \text{ MPa} \\ \tau &= 50 \text{ MPa} + 0.6(\sigma_n - P_f) && \sigma_n > 200 \text{ MPa} \end{aligned} \tag{2}$$

In terms of the principal stresses σ_1 and σ_3 (horizontal and vertical, respectively, Figure 1)

$$\sigma_n = \frac{\sigma_1 + \sigma_3}{2} - \frac{\sigma_1 - \sigma_3}{2}\cos 2\theta \tag{3}$$

and

$$\tau = \frac{1}{2}(\sigma_1 - \sigma_3)\sin 2\theta \tag{4}$$

where θ is the dip of the fault plane (Figure 1). Motion along an established fault plane requires that

$$\sigma_1 = \frac{2\tau_0 - 2\mu P_f + [\sin 2\theta + \mu(\cos 2\theta + 1)]\sigma_3}{\sin 2\theta + \mu(\cos 2\theta - 1)} \tag{5}$$

If σ_3 is specified as the lithostatic overburden pressure [e.g., *Brace and Kohlstedt*, 1980], then the shear resistance along a fault plane can be determined using (4) and (5) if the variation of pore pressure with depth is known.

If pore pressures approach lithostatic values, the capacity to resist motion along a fault plane can be virtually zero. This is demonstrated by the dynamics associated with the shallow portions of mature trenches. Efforts to determine the magnitude of shear coupling between subducting and overriding plates in mature subduction zones have produced surprisingly small values. Earthquake stress drop calculations [*Kanamori*, 1977; *Lay et al.*, 1982] and force balance models [*Forsyth and Uyeda*, 1975; *Bird*, 1978; *Froidevaux and Isacks*, 1984], as well as geological and geophysical observations [*Miyashiro*, 1973; *Rayleigh and Evernden*, 1981; *Davis and von Huene*, 1987; *Byrne et al.*, 1988], indicate very low shear stresses in shallow portions of trenches. The most straightforward explanation invokes elevated pore fluid pressures that effectively reduce the ambient normal stresses, facilitating fracture as well as sliding [*Davis et al.*, 1983; *von Huene and Lee*, 1983; *Moore and Silver*, 1987].

Although overpressures exist in many mature trenches, this is unlikely to be the case for oceanic lithosphere in general. High pore pressures are a consequence of the rapid subduction or burial of wet sediments [*Daines*, 1982; *Shi and Wang*, 1985, 1986, 1988; *Bethke*, 1986]. Deep-sea drilling observations strongly support the existence of near-hydrostatic conditions within the igneous portions of the oceanic crust. Drill holes that fully penetrate the sedimentary veneer rarely become the sites of upwelling flow due to overpressurization within the igneous basement [e.g., *Hyndman et al.*, 1987]. Thus we assume that pore pressures are hydrostatic.

At sufficiently high temperatures, relative motion is accommodated through distributed deformation rather than slip concentrated along a localized fault (Figure 1). The distance over which this deformation is distributed will affect the strain rate, which, along with the viscosity of the material involved, determines the magnitude of the resisting shear stress. Ductile shear stresses associated with incipient subduction can be estimated from compositionally appropriate flow laws, provided that ambient temperature and strain rate can be reasonably constrained [e.g., *Fleitout and Froidevaux*, 1980]. These flow laws are of the general form

$$\dot{\epsilon} = A\tau^n exp(-H^*/RT) \tag{6}$$

where τ is the shear stress, $\dot{\epsilon}$ is the strain rate, R is the universal gas constant, T is the temperature (K), and A, n, and H^* are constants that depend on composition.

For incipient subduction we model the thermal structure of oceanic lithosphere with a semi-infinite half-space cooling model (parameters in Table 1). There is considerable debate concerning the application of this model to seafloor older than 80–100 m.y. Thermally mature oceanic lithosphere may be more accurately characterized by the cooling plate model [e.g., *Parsons and Sclater*, 1977]. Both models are essentially in agreement for oceanic lithosphere younger than 80 m.y., beyond which the cooling plate model deviates from a linear relationship between depth and $\sqrt{t}$, and ocean depths begin to asymptotically approach 6.4 km [*Parsons and Sclater*, 1977]. This is the depth of 100 m.y. ocean floor predicted by the semi-infinite half-space cooling model (see, for example, *Turcotte and Schubert* [1982] and parameters

TABLE 1. Parameters and Values for Semi-infinite Half-Space Cooling Model

	Value
Gravitational acceleration g	9.8 m/s^2
Lithospheric thermal conductivity k_l	4.0 J/m s K
Sediment thermal conductivity k_s	3.0 J/m s K
Universal gas constant R	8.314 J/mol K
Asthenospheric temperature T_a	1650 K
Surface temperature T_s	270 K
Volumetric thermal expansion α	3 x 10^{-5} K^{-1}
Strain rate $\dot{\epsilon}$	10^{-15} s^{-1}
Thermal diffusivity κ	1.2 x 10^{-6} m^2/s
Lithospheric density ρ_l	3000 kg/m^3
Mantle/asthenosphere density ρ_m	3300 kg/m^3
Sediment density ρ_s	2500 kg/m^3
Ocean water density ρ_w	1000 kg/m^3

from our Table 1; also compare theoretical curves of *Parsons and Sclater* [1977]). If the oceanic lithosphere does approach a steady state thermal structure, this structure is adequately approximated (for our purposes) by the semi-infinite half-space thermal model with an age of 100 m.y.

Lithospheric processes involving the highest strain rate should trigger the deepest intraplate seismicity. The observation that earthquake depth determinations often reflect thermoelastic creep therefore implies that other processes normally associated with lithospheric evolution involve smaller strain rates than those characteristic of thermoelastic creep. The initiation of subduction, however, can hardly be considered within the context of normal lithospheric evolution. In fact, regions subjected to active intraplate deformation (e.g., Caroline Basin, Indian Ocean) exhibit the deepest intraplate seismicity on record [*Wiens and Stein*, 1983]. Because the initiation of subduction must involve intraplate deformation, it should have a characteristic strain rate greater than that of thermoelastic creep. We incorporate a dynamic argument to establish a lower bound on the strain rate associated with trench formation. A lower bound is pursued because this will minimize estimates of ductile shear resistance. A convergent velocity of at least 1 cm/yr is required if an underthrusting slab is to avoid thermal assimilation prior to accumulating the negative buoyancy necessary for evolution to maturity [*Purdy*, 1975; *McKenzie*, 1977]. Strain rate represents a velocity gradient perpendicular to a ductile shear zone (Figure 1), thus

$$\dot{\epsilon}_{min} = \frac{\Delta v_{min}}{\Delta x_{max}} \tag{7}$$

Equating Δx_{max} with the total thickness of oceanic lithosphere (75 km [*Craig and McKenzie*, 1986]), and setting Δv_{min} = 1 cm/yr [e.g., *Purdy*, 1975; *McKenzie*, 1977], implies a minimum strain rate of 4 x 10^{-15} s^{-1} for the process of trench formation. A strain rate of 10^{-15} s^{-1} was estimated by *Wiens and Stein* [1983] to be characteristic of oceanic regions subjected to active intraplate deformation, and, presumably, the initiation of subduction would occur in a dynamically similar environment. Thus we adopt a strain rate of 10^{-15} s^{-1}.

Shear resistance is determined as follows. Along a hypothetical fault plane, brittle and ductile shear stresses are determined at a particular depth, and the lesser of the two is assumed to represent the local shear stress resisting motion. This stress is integrated along the fault plane over a lithospheric thickness of 75 km, and the value of this integral represents the force (per unit distance) necessary to overcome the total shear resistance. (Preliminary calculations indicated that at 75 km depth shear stresses were always negligible.) In determining the ductile resistance, a diabase crust 10 km thick is incorporated [*Caristan*, 1982], and the subcrustal material is modeled with a wet dunite flow law [*Chopra and Paterson*, 1981]. Because hydrated dunite is more ductile than anhydrous dunite, and diabase is more ductile than dunite, these rheologies are adopted to insure a lower bound on the estimate of ductile shear resistance to trench formation (also, our somewhat excessive choice of 10 km for crustal thickness reflects a similar minimization). Flow parameters are summarized in Table 2. For normal seafloor the diabase crust remains within the brittle deformational regime in all cases, and the composition of the crust is not relevant to strength determinations. Significant accumulations of thermally insulating sediments near passive margins, however, can elevate crustal temperatures to the point that compositional effects become nonnegligible.

Shear resistance as a function of depth and lithospheric age is illustrated in Figure 2 for a fault plane dip of 30°. Because shear resistance in the brittle regime is independent of thermal structure, all curves share the same linear stress versus depth dependence. Figure 3 depicts the integrated shear stress along the fault plane, which represents the minimum compressional inplane force (per unit distance) that must be available if lithospheric thrusting and convergence is to occur. Because the process of trench formation may involve the reactivation of preexisting faults, we consider a range of fault plane dips. At small angles the minimum force decreases with increasing dip because the distance over which the integration is performed (75 km/sin θ) is decreasing. At larger angles, however, the maximum principal stress σ_1 begins to contribute significantly to the normal stress acting across the fault plane. This results in more efficient coupling across the fault and increased resistance to motion with increasing dip angle. Younger lithosphere is hotter, possessing a shallower brittle-ductile transition, and because ductile shear resistance is not affected by fault plane angle, the curves representing younger ages do not exhibit as pronounced a dependency on this angle. Also, note that whereas the thickness of the "mechanical" lithosphere (defined as the depth to a particular isotherm [e.g., *McNutt and Menard*, 1982]) increases as the square root of the age when determined on the basis of the semi-infinite half-space thermal model, the compressional resistance of the lithosphere does not. Lithospheric strength is not directly proportional to mechanical thickness because as the brittle-ductile transition deepens, the maximum sustainable stress difference (determined by the intersection of the brittle and ductile curves, e.g., Figure 2) increases. The increase in lithospheric strength with age is roughly linear (e.g., Figure 3). If the strain rate is reduced by an order of magnitude, the reduction in integrated shear resistance is about 1 x 10^{12} N/m for 20 m.y. oceanic lithosphere and 8 x 10^{12} N/m for 200 m.y. oceanic lithosphere.

As discussed above, if the cooling plate model is a more appropriate description of the thermal evolution of oceanic lithosphere than the semi-infinite half-space model, then the shear resistance would approach a limit approximated by the 100 m.y. curve.

TABLE 2. Flow Law Parameters

	A, MPa^{-n} s^{-1}	H^*, J/mol	n (dimensionless)
Wet olivine	1.06×10^5	4.44×10^5	3.4
Diabase	5.07×10^{-1}	2.76×10^5	3.05
Granite	7.43×10^{-4}	1.37×10^5	1.9

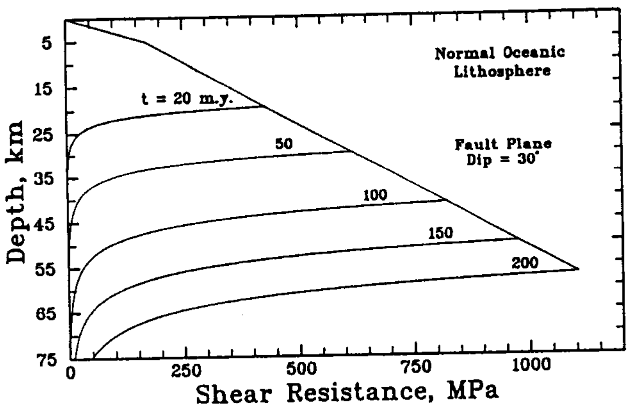

Fig. 2. Compressional shear resistance as a function of depth for normal oceanic lithosphere assuming hydrostatic pore pressure, a semi-infinite half-space cooling model (age = t), and a fault plane dip of 30°. Note that despite the superficial resemblance, these diagrams do not represent lithospheric "yield envelopes."

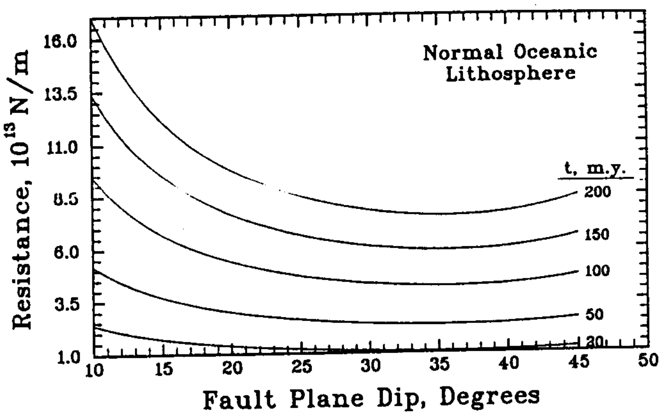

Fig. 3. Minimum force (per unit distance) required to promote incipient lithospheric convergence within normal oceanic lithosphere as a function of fault plane angle. Lithosphere of all ages (t) is capable of withstanding compressional stresses associated with ridge push. This is why seafloor does not spontaneously founder upon achieving negative buoyancy.

Lithosphere younger than 20 m.y. of age is addressed in a following section concerning trench formation in near-ridge environments. Estimates of ridge push forces corresponding to lithosphere obeying the semi-infinite half-space cooling model slightly exceed 7×10^{11} N/m and 7×10^{12} N/m for 20 m.y. old and 200 m.y. old oceanic lithosphere, respectively [*Parsons and Richter*, 1980]. Figure 3 indicates that normal oceanic lithosphere is therefore quite capable of withstanding ridge push forces (i.e., negative buoyancy). This is why oceanic lithosphere does not spontaneously founder upon achieving negative buoyancy. Seafloor of all ages is quite capable of withstanding the stresses associated with negative buoyancy.

Passive Margins

Passive margin sediments may decrease the strength of the adjacent seafloor because they are thermally insulating. Simply extrapolating near-surface thermal gradients will overestimate the temperature at the bottom of a thick sediment column because thermal conductivity increases dramatically with compaction and solidity. For this reason we perform a simple thermal evolution calculation. The precise thermal effects of passive margin sediments are dependent upon the history of sediment deposition. We assume that the sediments have increased in thickness in a $\sqrt{t}$ fashion, reaching their final value, h_s, at the current age of the margin. This assumption is primarily one of analytical convenience, because the thermal structure of sedimentary basins and their basements has already been determined subject to this constraint [*Turcotte and Ahern*, 1977]. Such an assumption, however, is quite common in passive margin evolutionary studies because it implies that the rate of sediment deposition is proportional to the thermal subsidence of the basement [e.g., *Watts and Steckler*, 1979; *Steckler and Watts*, 1982]. On the basis of this thermal model, the temperature distribution in the oceanic basement adjacent to a passive margin is predicted to be [*Turcotte and Ahern*, 1977]

$$T(z,t) = T_a - \frac{(T_a - T_s)\operatorname{erfc}\left(z/2\sqrt{\kappa t}\right)}{1 + (k_l/k_s)\operatorname{erf}\left(h_s/2\sqrt{\kappa t}\right)} \tag{8}$$

where t is time and z is depth measured from the bottom of the sediment column (i.e., the top of the oceanic basement); all other quantities are defined (and their values provided) in Table 1. This expression assumes thermal equilibrium at all times, negligible heat loss by hydrothermal circulation, and semi-infinite half-space cooling model boundary conditions.

We adopt the sediment thermal conductivity value used by *Bodri and Jessop* [1989] in their study of Canadian passive margins. Sediment thicknesses as great as 15 km are considered, which is almost certainly an overestimate for sediments deposited on oceanic lithosphere adjacent to passive margins. Although a thickness of 18 km has been observed in the Baltimore Canyon Trough, this particular sequence rests on rifted continental basement. The inferred upper limit of sediment accumulations around the passive margins of the proto-Atlantic prior to basin closure is 15 km [*Hatcher and Viele*, 1982]. In mature passive margins, sediments eventually begin building out rather than continuing to accumulate vertically (e.g., the Gulf Coast), and thus a future 300 m.y. passive margin would not be expected to have a maximum sediment thickness significantly greater than those currently observed. The sediments are assumed to be strengthless and are not included as part of the 75 km thick lithosphere.

The transition from continental crust to oceanic crust is generally not abrupt; rather, it involves a distributed thinning of continental crust to near-oceanic crustal thicknesses [e.g., *Avedik et al.*, 1982]. To further minimize estimates of trench formation resistance, we replace the diabase flow law characterizing the 10 km thick crust with a weaker wet granite rheology (Table 2) after *Hansen and Carter* [1982]. If abrupt transitions from continental to oceanic crust are common, and oceanic lithosphere were thrust against 35 km of continental crust, our estimates of passive margin trench formation resistance would be excessive. The relatively few inferences of abrupt transitions are primarily based on gravity and/or subsidence, rather than seismic, studies and cannot be regarded as highly reliable [*Tréhu et al.*, 1989]. It is also important to note that the thermal structure of oceanic lithosphere immediately adjacent to nonattenuated continental crust would be displaced to reduced temperatures due to nonnegligible lateral heat conduction into the cold craton. Finally, and perhaps most significantly, geophysical observations reveal that the point of rupture associated with the establishment of an Andean-type subduction zone often occurs distinctly seaward of the ocean-continent transition, or, in other words, entirely within oceanic lithosphere [e.g., *Dickinson and Seely*, 1979]; this will be discussed in greater detail below.

Figure 4 depicts the integrated shear resistance as a function of fault plane angle for 100 m.y. and 200 m.y. old passive margins with various sediment accumulations. Even under 15 km of insulating sediments, these margins are capable of resisting nearly 3 x 10^{13} N/m and 6 x 10^{13} N/m of compressional force, respectively. In contrast, when determined on the basis of the cooling plate model, ridge push forces do not exceed about 3 x 10^{12} N/m [*Parsons and Richter*, 1980], and semi-infinite half-space thermal models imply a ridge push force of approximately 7 x 10^{12} N/m for 200 m.y. old oceanic lithosphere [*Parsons and Richter*, 1980]. The nearly order of magnitude discrepancy between available ridge push forces and shear resistance clearly indicates that passive margin failure does not occur in response to negative buoyancy. The additional effects of basal shear stress and stress concentrations at ocean/continent boundaries would still not indicate the likelihood of trench formation at passive continental margins.

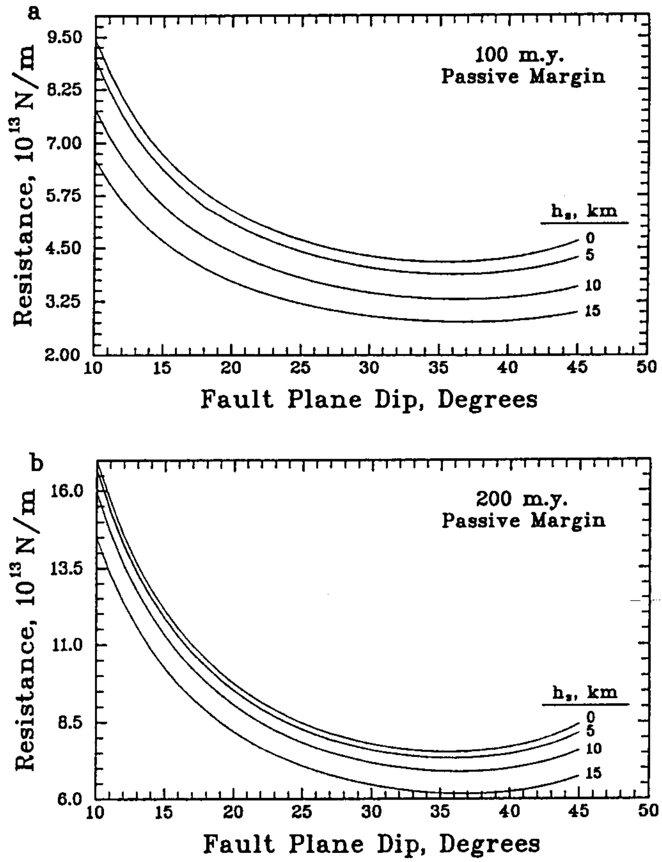

Fig. 4. Minimum force required to promote the initiation of subduction at (a) 100 m.y. and (b) 200 m.y. passive continental margins. Although greater sediment thicknesses, h_s, more efficiently insulate, and therefore weaken, the underlying lithosphere, passive margin failure does not occur. The 100 m.y. old model approximates the steady state condition achieved if the cooling plate model is a more accurate representation of cooling lithosphere.

Transform Faults and Fracture Zones

Whether or not transform faults and fracture zones represent inherent zones of lithospheric weakness is difficult to assess. A number of observations indicate that the transform portions of ridge-transform-ridge boundaries are weaker in shear than the ridge portions are in tension [*Lachenbruch and Thompson*, 1972; *Oldenburg and Brune*, 1975; *Morgan and Parmentier*, 1984; *Tamsett and Searle*, 1988]. Because transform faults are subjected to relatively high strain rates [$\simeq 10^{-13}$ s^{-1}, *Engeln et al.*, 1986; *Chen*, 1988], they might be expected to exhibit seismicity at greater depths (i.e., higher temperatures) than normal oceanic lithosphere. *Engeln et al.* [1986] concluded that the opposite is actually observed and that transform faults must therefore be weaker than normal oceanic crust. *Bergman and Solomon* [1988] questioned the centroid depth determinations of *Engeln et al.* [1986] and concluded that seismicity does occur at elevated temperatures. The ambiguity in data interpretation is demonstrated by the fact that both groups investigated many of the same events.

Fracture zones might be considered even more likely than transform faults to represent sites of trench formation due to their greater abundance as well as their more common occurrence in older (i.e., negatively buoyant) lithosphere. Unfortunately, observations concerning the relative weakness of fracture zones are also ambiguous. Correlations of intraplate seismicity with fracture zones have been proposed to exist [*Bergman and Solomon*, 1980; *Wiens and Stein*, 1984; *Okal et al.*, 1986; *Stein et al.*, 1989] and not to exist [*Bergman*, 1986; *Solomon et al.*, 1989]. *Sandwell* [1984] argued that the progressive relative subsidence that should characterize fracture zones separating oceanic lithosphere of significantly different ages is not routinely observed, indicating substantial coupling across these features; similar arguments were offered by *Haxby and Parmentier* [1988]. In contrast, *Bonatti* [1978] has documented substantial vertical tectonism in a number of oceanic fracture zones, and some Atlantic equatorial fracture zones are apparently isostatic [e.g, *Sibuet and Veyrat-Peinet*, 1980; *Diament et al.*, 1986]. Both observations suggest tectonic adjustment accommodated across a zone of weakness. *Lowrie et al.* [1986], in a review of this issue, propose that volcanic activity associated with many fracture zones is indicative of relative weakness.

The strength of transform faults and fracture zones relative to normal seafloor remains a controversial topic in marine geophysics. Because our purpose is to estimate minimum resistance to trench formation, we simply assume that weakening occurs and recognize three potential mechanisms: (1) extremely pervasive fracturing, (2) ductile weakening associated with shear heating, and (3) brittle weakening associated with serpentinite observed in these environments. If transform faults and fracture zones are weaker than normal seafloor simply because they are more pervasively fractured, then lithospheric strength estimates based on application of Byerlee's Law would provide a more accurate characterization of these regions than of oceanic lithosphere in general. Because our goal is to pursue a lower bound on the magnitude of the force involved in the trench formation process, we can proceed without resolving this issue. Extreme fracturing is already incorporated into our basic model as we use Byerlee's Law, rather than rock strength measurements, to estimate brittle shear resistance.

Significant shear heating requires negligible heat loss through hydrothermal circulation within the transform domain; a condition we view as inconsistent with the ubiquitous serpentinite commonly observed in these environments [see review by *Bonatti and Hamlyn*, 1981]. The presence of this hydrated peridotite implies substantial fluid circulation [*Francis*, 1981; *MacDonald and Fyfe*, 1985]. Seismic velocity profiles beneath active transform faults, as well as their inactive fracture zone extensions, indicate that hydrothermal alteration extends well beyond the depths sampled by surface dredging products [*White et al.*, 1984; *Calvert and Potts*, 1985]. In addition, heat flow measurements along the San Andreas fault indicate minimal shear heating [*Lachenbruch and Sass*, 1980]. For these reasons we dismiss shear heating as a significant weakening mechanism for transform faults (such a mechanism would not apply to fracture zones regardless).

The frictional sliding behavior of serpentinite may represent the most reasonable weakening mechanism. *Dengo and Logan* [1981] demonstrated that different varieties of serpentinite exhibit different sliding behavior. Whereas serpentinites exhibiting a "flare texture" were determined to possess a coefficient of frictional sliding only slightly less than materials obeying Byerlee's Law, "mesh-textured" serpentinites were determined to be considerably weaker. Coefficients of sliding friction for the latter ranged from 0.50 to 0.56 (in contrast to 0.85 for Byerlee's Law). Mesh-textured serpentinites are characteristic of the lizardite polymorph [*Dengo and Logan*, 1981], which is commonly associated with oceanic transform faults and fracture zones [*Bonatti and Hamlyn*, 1981].

We estimate a lower bound on the resistance to trench formation along oceanic transform faults and fracture zones by incorporating the frictional sliding relationship of *Dengo and Logan* [1981] in determining shear resistance to lithospheric thrusting. Thermal structure is assumed to be that of normal seafloor with an age of the younger side. In reality, temperatures will be reduced from these values due to lateral heat conduction into the colder side, but because we are pursuing a lower bound on the shear resistance, our approximation is satisfactory. Shear resistance as a function of depth is determined as above, except that a new frictional sliding relationship is incorporated for the brittle regime

$$\tau = 0.50(\sigma_n - P_f) \tag{9}$$

Figure 5 illustrates integrated shear resistance as a function of fault plane dip. Note that the resistance associated with young transform faults and fracture zones approaches that of normal seafloor of similar age. This is because younger lithosphere possesses a shallower brittle-ductile transition, and the anomalous brittle properties of the plate become less relevant to the net resistance. Even when the younger side is 20 m.y. old, lithospheric convergence requires a compressional force of at least 1 x 10^{13} N/m. These regions are thus capable of withstanding any of the potential trench-forming forces discussed above.

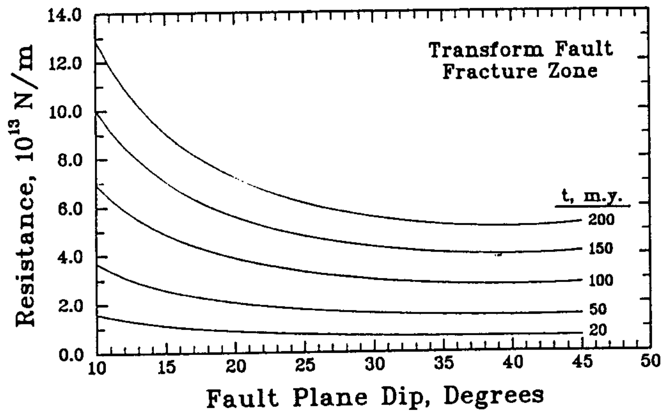

Fig. 5. Minimum force required to promote the initiation of subduction along a transform fault or fracture zone. The age t refers to the younger side. An anomalously weak frictional sliding relationship (characteristic of serpentinite) is incorporated into the calculations of resistance.

Oceanic Spreading Centers

Ridges may be hot enough that deformation occurs in an extremely ductile manner even in the near-surface environment. If this is indeed the case, then the shear resistance to lithospheric thrusting, as determined above, may become negligible. Simple thermal models, however, may not accurately describe the near-ridge environment due to the effect of vigorous hydrothermal circulation on heat transport. A review of this process is given by *Bratt et al.* [1985]. We model the thermal structure of a spreading center in the same manner as *Bratt et al.* [1985], by allowing the surface temperature to extend to various depths. This is intended to mimic efficient heat transport via hydrothermal circulation to these depths, which we assume do not extend beyond the crust-mantle interface ($\simeq$ 10 km). Other than the incorporation of modified thermal profiles, the procedure for determining shear resistance is identical to that outlined in the section on normal oceanic lithosphere.

Because trench formation in a near-ridge environment involves young lithosphere, positive buoyancy introduces an additional, nonnegligible resisting force. The force necessary to overcome positive buoyancy can be estimated from the "density defect thickness" (δ) as defined by *Oxburgh and Parmentier* [1977]

$$\delta = \int_0^T \left(\frac{\rho_m - \rho(z)}{\rho_m}\right) dz \tag{10}$$

where T is the thickness of the lithosphere, ρ_m is the density of the mantle/asthenosphere, and z is depth. At the ridge this integral is greater than zero, indicating positive buoyancy due to compositional differences with respect to the underlying mantle/asthenosphere. At some point, however, this compositional buoyancy is compensated by density increases associated with cooling, and the density defect thickness becomes negative. On the basis of component densities cited by *Oxburgh and Parmentier* [1977], the compositional contribution to the density defect thickness (δ_c) is independent of age and equals approximately 1.26 km. (Because this compositional difference exists for all oceanic crust, it does not contribute to lateral stress variations associated with ridge push; it does, however, contribute to positive buoyancy.)

The age-dependent thermal contribution (δ_{th}) can be determined from seafloor subsidence [*Oxburgh and Parmentier*, 1977]

$$\delta_{th} = \left(\frac{\rho_w - \rho_m}{\rho_m}\right)(d - d_0) \tag{11}$$

where ρ_w is the density of seawater and $(d - d_0)$ is the amount of subsidence with respect to the ridge. Using the depth versus age parameterization of young oceanic lithosphere from *Parsons and Sclater* [1977] and densities from *Oxburgh and Parmentier* [1977], the density defect thickness becomes

$$\delta_{total} = \delta_c + \delta_{th} = \delta_c - d'\sqrt{t} \tag{12}$$

where t is lithospheric age in m.y. and $d' = 0.245$ km/m.y.$^{1/2}$. This equals zero when t = 26.4 m.y.; this is younger than the 40 m.y. estimate of Oxburgh and Parmentier, probably because we used a revised depth versus age relationship. Equating t with x/u_0, where x is distance from the ridge and u_0 is spreading velocity, the buoyant force (per unit distance) that must be overcome to subduct a ridge is then

$$F = g\rho_m \int_0^{x_0} \left[\delta_c - d'\sqrt{x/u_0}\right] dx \tag{13}$$

where g is gravitational acceleration and x_0 is the distance (from the ridge) at which neutral buoyancy is achieved. For a spreading velocity of 1 cm/yr, representative of plates with no attached subducting slab, x_0 = 264 km, and this force is approximately 3.6 x 10^{12} N/m.

Figure 6 depicts the minimum resistance (including positive buoyancy) to lithospheric underthrusting, as a function of thrust plane dip, for surface boundary temperatures extending to depths (d) of 0, 5, and 10 km. Ridges subjected to a greater degree of cooling due to hydrothermal circulation possess a thicker regime of brittle deformation, which accounts for the greater dependency of total resistance on fault plane angle. If heat loss through near-ridge hydrothermal circulation is negligible (an unlikely prospect [e.g., *Lewis*, 1983; *Bratt et al.*, 1985]), and there is no near-surface isothermal zone (d = 0), then the only significant resistance is the 3.6 x 10^{12} N/m associated with positive buoyancy. Other considerations, however, suggest that mid-oceanic ridges are unlikely to be transformed into subduction zones.

Despite extinction, large remnant tensional stresses will remain as long as the region is elevated. In addition to the 3.6 x 10^{12} N/m of compressional force necessary to overcome the minimum resistance associated with positive buoyancy, enough compression must first be supplied to convert the recently spreading ridge into a zone of compression. "Shutting off the ridge" in this manner would presumably require a force equal in magnitude, but opposite in direction, to ridge push. Thus the necessary minimum force is actually in the range 7 x 10^{12} N/m to 1 x 10^{13} N/m. Such compression could not be provided by a nearby parallel ridge, because the associated gravitational forces would not be transmitted up adjacent ridge crests. Also, because plate motions are always directed away from ridges, near-ridge compressional stresses cannot be induced by lithospheric basal shear. The only potential source of compressional stress thus far discussed might be elevational discontinuities at ocean-continent boundaries. Yet even if these stresses were transmitted from the continental margins into the mid-oceanic ridges, the magnitude of this force is insufficient to induce trench formation [only 3 x 10^{12} N/m available; *Turcotte and Schubert*, 1982]. Additional complications exist as well. It is not sufficient to merely "shut off" the ridge; the transition from an tensional stress regime to a compressional stress regime must be abrupt. This is necessary because once the spreading center becomes extinct, it begins to cool (and strengthen) at approximately the same rate as normal oceanic lithosphere [e.g., *Sclater et al.*, 1971], thereby forfeiting its extreme ductility. Also, the above calculations indicate that approximately 264 km of lithospheric convergence must occur before even neutrally buoyant material is involved. If the spreading rate of a typical Atlantic-type ridge (1–1.5 cm/yr) were instantaneously reversed, consumption of the approximately 264 km of lithosphere necessary to achieve neutral buoyancy would require about 20 m.y. Unless significant amounts of shear heating accompany this process of convergence, an additional resistance of at least 10^{13} N/m would be encountered after this amount of time (Figure 3). For these reasons we conclude that mid-oceanic ridges are unlikely to represent subduction nucleation sites.

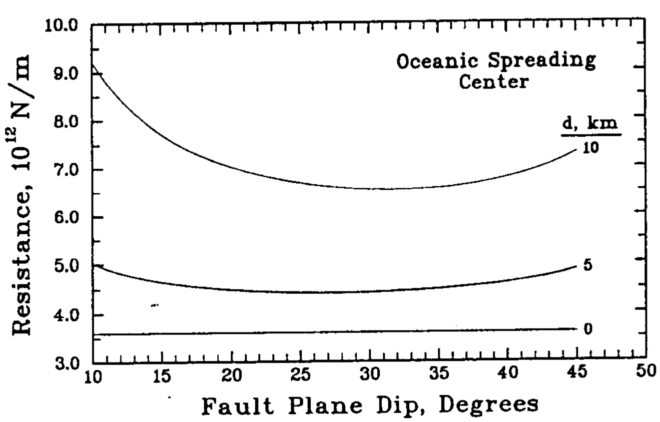

Fig. 6. Minimum resistance to the initiation of subduction for ridges. In addition to shear resistance, the effects of positive buoyancy are included. Enhanced cooling due to hydrothermal circulation is modeled by extending the surface boundary condition to depth d. Although the resistance depicted can be small, other considerations strongly suggest that ridges are unlikely sites of trench formation.

Mechanism for the Initial Stage of Trench Formation

The greatest tectonic stresses are those associated with the negative buoyancy of subducting slabs in mature trenches. These "slab pull" forces may exceed 10^{14} N/m [*Davies*, 1983] and typically are about 5 x 10^{13} N/m [*Turcotte and Schubert*, 1982; *Davies*, 1980, 1983]. There are two ways in which slab pull might induce large compressional stresses in surface portions of the lithosphere. First, along-strike variation in either the magnitude of slab pull forces or the degree of mechanical coupling between converging plates of a subduction zone may induce large compressional stresses seaward of the trench (i.e., within the subducting plate) [e.g., *Cloetingh and Wortel*, 1986; *Christensen and Ruff*, 1988]. Second, sufficient degrees of mechanical coupling may also result in large compressional stresses within the overriding plate [e.g., *Collot et al.*, 1985; *Uyeda*, 1987; *McCaffrey*, 1988].

Congestion at the Himalayan collision zone, combined with unimpeded convergence at the Sunda Arc, has resulted in compressional stress levels within the Indian plate that are an order of magnitude greater than those associated with ridge push [*Cloetingh and Wortel*, 1986; *McAdoo and Sandwell*, 1985; *Zuber*, 1987]. Thus stresses generated in this manner are capable of exceeding the lower bounds that we have determined necessary for trench formation. This is consistent with recent speculation that congestion in the eastern extremity of the Nankai Trench (by the presumably buoyant Izu-Bonin Ridge), combined with unimpeded convergence to the west, may be responsible for intraoceanic lithospheric thrusting inferred along the southeast flank of the Zenisu Ridge [*Chamot-Rooke and Le Pichon*, 1989]. This is believed to represent an incipient jump in the position of the trench. *Christensen and Ruff* [1988] have demonstrated a correlation between outer rise seismicity and variations in mechanical coupling at the trench. Outer rises normally exhibit only extensional seismicity in response to flexure at the trench (the complementary compressional stresses presumably exist at depths where deformation occurs aseismically). Seaward of locally congested trenches, however, compressional outer rise events are common, reflecting a significant level of compressional stress in response to continued subduction in adjacent portions of the convergence zone [*Christensen and Ruff*, 1988].

Compression in overriding plates as a consequence of congestion at trenches has been proposed to account for the common association of island arc cusps with the arrival of buoyant, aseismic ridges at the subduction complex [*Vogt*, 1973]. Significant levels of inplane compressional stresses on the portion of the (overriding) Pacific plate directly opposite the D'Entrecasteaux collision zone in the New Hebrides Trench have been inferred on the basis of both geologic and seismic considerations [*Collot et al.*, 1985]. The elevation of compressional stresses in overriding plates involved in collision-related congestion is also supported by forearc tectonism and geochemical studies [*Uyeda*, 1987]. Subduction polarity reversals are almost certainly the result of trench congestion inducing large compressional stresses within the overriding plate. An example of this process is represented by the tectonic evolution of the Timor region, where the Australian continental shelf is attempting to subduct into the Java Trench/Timor Trough. As the relatively buoyant continental material clogs up the trench and allows the slab pull stresses from the subducted oceanic slab (attached to the Australian plate) to be transmitted into the overriding Pacific plate, the earliest indication of subduction polarity reversal is evident in newly forming thrust zones within the backarc basin [*Silver et al.*, 1983; *McCaffrey and Nábělek*, 1984; *McCaffrey et al.*, 1985; *McCaffrey*, 1988].

If the compressional stress levels required for trench formation are indeed a consequence of the dynamics of existing subduction zones, implications concerning probable sites of incipient subduction follow. Regardless of the relative weakness of candidate sites, the common proximity of transform faults and fracture zones to mature trenches suggests that they may represent the most likely sites of trench formation if they are even marginally weaker than normal oceanic lithosphere. We are not proposing that subduction nucleates exclusively along transform faults and fracture zones, only that these sites may be preferred.

OBSERVATIONS RELATING TO TRENCH FORMATION

Insight into the subduction nucleation process may be found in the structure of modern arc-trench systems, plate motion reconstructions, the geologic record of oceanic basin closure, and sites of contemporary trench formation. The latter (if they indeed exist) might be identified on the basis of seafloor structures, anomalous plate boundary seismicity, "intraplate" seismicity and deformation, and nonclosure of plate velocity vectors. Here we review evidence that subduction nucleates intraoceanically (not at passive margins), sometimes along transform faults and fracture zones, and that the driving forces are associated with trench congestion at mature subduction complexes.

Modern Arc-Trench Systems

The structure of contemporary subduction zones provides critical information regarding trench formation. In particular, the nature of the forearc region, which lies between the trench axis and the volcanic arc massif, is extremely important in reconstructing the earliest stages of arc evolution [e.g., *Dickinson and Seely*, 1979; *Karig*, 1982; *Casey and Dewey*, 1984]. A common feature is the presence of a forearc basin, often filled with mildly deformed sediments, which is bounded by the structural highs of the volcanic arc massif and the accretionary wedge. Although some portion of the forearc basement usually consists of the tectonic melange of the accretionary wedge as well as the deeper roots of the arc massif, there is often an intervening segment representing the original, pretrench basement [e.g., *Dickinson and Seely*, 1979]. The presence of oceanic forearc basement is commonly interpreted as an indication that the initial convergence leading to trench formation was accommodated intraoceanically. In this case the basement material would then represent oceanic crust trapped between the zone of initial convergence and the developing volcanic arc massif.

Sampling from the forearc regions of the Izu-Bonin-Mariana-Yap subduction complex reveals basement consisting of metamorphosed mafic and ultramafic protoliths [*Shiraki*, 1971; *Ogawa and Naka*, 1984], indicating that these subduction zones formed entirely within oceanic crust and have not migrated to their present positions from nearby continental margins. An equivalent conclusion can be drawn from the Tonga subduction zone [*Fisher and Engel*, 1969]. Most importantly, even the forearc regions of subduction zones adjacent to continental margins may possess foundations of oceanic crust (e.g., Sunda [*Curray et al.*, 1977], Central America [*Ibrahim et al.*, 1979; *Aubouin et al.*, 1982], Colombia/Ecuador [*Henderson*, 1979; *Feininger and Seguin*, 1983]), suggesting that the evolution of these regions may be more complicated than simple models of passive margin failure allow (i.e., it becomes necessary to account for failure within oceanic lithosphere at some distance from the ocean-continent transition; see *Dickinson and Seely* [1979] for an interesting example of an attempt to reconcile the presence of oceanic forearc basement in Andean-type subduction zones with models of passive margin failure). An absence of oceanic forearc basement along an Andean-type subduction zone does not necessarily imply that the original basement involved in the trench formation process was not oceanic in nature. This is because the destruction of forearc regions through tectonic erosion and strike-slip displacement accompanying oblique subduction is an important element in the evolution of many subduction zones [e.g., *von Huene*, 1984].

Backarc (or marginal) basins may provide useful clues in understanding the trench formation process as well. Although many of these basins can be attributed to rifting and spreading of the overriding plate, others did not originate in this manner. Entrapment of preexisting seafloor behind an intraoceanic subduction zone that nucleated along a transform fault has been proposed to account for magnetic lineations in the Bering Sea marginal basin that trend approximately perpendicular to the strike of the Aleutian Trench [*Hilde et al.*, 1977]. The evolution of a transform boundary into a trench has also been proposed by *Uyeda and Ben-Avraham* [1972] to account for the near-orthogonality of an extinct spreading center in the West Philippine backarc basin to nearby subduction zones.

Appalachian Geology and Ophiolite Emplacement

The destruction of the proto-Atlantic Ocean almost certainly involved episodes of trench formation, because presumably significant amounts of lithospheric convergence were accommodated at this time. Geologic observations indicate that the earliest developing subduction zones in the proto-Atlantic were intraoceanic in nature (i.e., not the product of passive margin failure). For example, throughout both the New England and Canadian portions of the Appalachians the first departure from a typical passive margin evolution involved rapid subsidence immediately followed by ophiolite obduction [e.g, *Williams*, 1979; *Hatcher and Viele*, 1982; *Suppe*, 1985]. Because there is no evidence of landward arc-associated volcanism prior to obduction [*Currie et al.*, 1980], it must be concluded that the initial subduction zones dipped away from the continent and were therefore intraoceanic. East dipping Paleozoic subduction in Newfoundland has also been inferred geophysically [*Haworth et al.*, 1978], geochemically [*Stevens et al.*, 1974; *Strong et al.*, 1974], and sedimentologically [*Jacobi*, 1981]. Furthermore, ophiolite obduction was followed by the suturing of island arc material possessing both intraoceanic geochemical characteristics [*Strong*, 1977] and an unusually large proportion of endemic brachiopod faunas unlike contemporary North American or European varieties [*Neuman*, 1984]. This indicates that the creation of the earliest arc-related constructs was well removed from any continental margin.

A number of ophiolite studies favor emplacement mechanisms that require an initial episode of intraoceanic trench formation (see review by *Moores* [1982, and references therein]). Essentially, these models require that subduction initially dip away from the nearest passive margin, with the continent residing on the subducting plate (e.g., Timor). As the continent attempts to subduct, the island arc complex is thrust onto the continental margin. Because subduction of continental lithosphere cannot continue indefinitely, either convergence must cease or a polarity reversal must occur. In the latter case, material between the leading edge of the initially overriding plate and the point of lithospheric failure accompanying polarity reversal will be transferred onto the former passive margin as an ophiolitic/volcanic arc assemblage. The most likely place for failure to occur would be the weak, thermally immature lithosphere of the backarc basin, thus accounting for the similarities of many ophiolites to backarc basin crust. Such a model implies, however, that at least some ophiolitic assemblages must represent former forearc basement. In fact, the presence of transform-related features along the leading edge of some ophiolitic thrust sheets has been interpreted as evidence supporting the conversion of oceanic crust immediately adjacent to a transform fault into forearc basement through the nucleation of subduction along the transform fault [*Casey and Dewey*, 1984; *Ogawa and Naka*, 1984].

East Pacific Margin Evolution

As emphasized by *Casey and Dewey* [1984], the evolution of the eastern margin of the Pacific basin cautions against simplistic scenarios of passive margin failure. Although the presence of oceanic forearc basement along actively converging portions of the East Pacific border [e.g., *Henderson*, 1979; *Ibrahim et al.*, 1979; *Aubouin et al.*, 1982; *Feininger and Seguin*, 1983] as well as the accretion of allochthonous terranes (including intraoceanic island arcs) along the entirety of this border [e.g., *Coney et al.*, 1980; *Aspden and McCourt*, 1986; *Feininger*, 1987] are not necessarily inconsistent with a foundering passive margin, additional evolutionary "events" must be introduced to account for these observations [e.g., *Lebras et al.*, 1987]. Alternatively, the establishment of Andean-type subduction by means of a polarity reversal following the attempted subduction of a passive continental margin into an (originally) intraoceanic island arc, offers a single, unifying explanation for these "events". In this case, present-day oceanic forearc basement would represent former backarc basin seafloor within which renewed convergence (of opposite polarity) was initiated. The former island arc complex might represent a variety of accreted terranes. Early episodes of reverse polarity, with subduction dipping away from the continent, have been proposed for Canada [*Tempelman-Kluit*, 1979; *Price and Hatcher*, 1983; *Hansen*, 1990], California [*Dickinson*, 1981; *Schweickert and Snyder*, 1981; *Speed and Sleep*, 1982; *Dilek et al.*, 1988], and Ecuador [*Feininger and Bristow*, 1980]. Paleomagnetic data indicating an allochthonous origin for oceanic forearc basement along the Ecuadorian portion of the Andes subduction complex [*Roperch et al.*, 1987], combined with the presence of a sutured intraoceanic island arc immediately to the east [*Henderson*, 1979], are clearly consistent with an arc-continent collisional origin for allochthonous terranes and the oceanic nature of forearc basement in this region.

Contemporary Examples of Trench Formation

Anomalous seismicity has been observed within the eastern portion of the Azores-Gibraltar transform boundary (in the Atlantic near the Mediterranean). Focal mechanisms indicate a consistent north-northwest compression along the eastern end of this boundary [*McKenzie*, 1972; *Fukao*, 1973], where simple kinematic models of plate tectonics predict only east-west strike-slip motion. Seafloor structures, as well as gravity anomalies, also indicate a component of motion orthogonal to the strike of the plate boundary [*Purdy*, 1975; *Karner et al.*, 1985]. Although *Grimison and Chen* [1986] have argued that the anomalous seismicity is not restricted to the transform boundary, *Moreira* [1985] and *Buforn et al.* [1988] have noted that anomalous seismicity not directly associated with this boundary occurs along nearby former transform faults that extend onto the Portuguese mainland. Because the convergent component of motion, which probably results from subduction in the nearby Mediterranean Sea (it clearly is not associated with regional ridge push forces), is no greater than 1.5 cm/yr, thermal assimilation will occur at relatively shallow depths, and it is not clear that a mature subduction zone (i.e., one driven primarily by slab pull forces) will ever develop along this boundary [*Purdy*, 1975; *McKenzie*, 1977].

A similar evolution has been inferred for the Macquarie Ridge Complex south of New Zealand. Oblique convergence is indicated by the occurrence of thrust-related seismicity along a zone that primarily functions as a transform boundary [*Ruff et al.*, 1989]. This convergence, which is also supported by seafloor structures, gravity anomalies, and plate reconstructions [*Ruff and Cazenave*, 1985; *Ruff et al.*, 1989; *Williamson et al.*, 1989], is probably induced by subduction zones located to the north. As with the Azores-Gibraltar region, it is not clear that the Macquarie Ridge Complex will eventually reach maturity due to a low convergence rate.

Perhaps the most carefully documented example of potential incipient trench formation is from the East Caroline Basin in the western equatorial Pacific. Seafloor structures, gravity anomalies, and seismicity suggest that approximately 10 km of Caroline "plate" has been thrust beneath Pacific plate, apparently along a former fracture zone, during the last one million years [*Hegarty et al.*, 1982]. The compressional stress regime responsible for lithospheric thrusting may be a consequence of buoyant features on the

Caroline "plate" congesting parts of the Yap-Mariana trench to the west [e.g., *Vogt*, 1973]. Adjacent, noncongested portions of the subduction zone continue to vigorously subduct the Pacific plate northwestward; if this motion is sufficiently inhibited by buoyant material of the Caroline "plate", large compressional stresses will develop within this "plate". A causal relationship between compressional stresses and regional dynamics is consistent with the observation that the convergence velocity where the Caroline Rise enters the Yap-Mariana Trench is an order of magnitude less than at the nearby Mariana Trench [*American Association of Petroleum Geologists*, 1981].

To account for Pacific intraplate seismicity, *Kroenke and Walker* [1986] and *Okal et al.* [1986] proposed the existence of an incipient zone of subduction extending from Western Samoa to the Caroline Ridge. Because most of this proposed boundary is seismically quiescent, its existence has been inferred by the connection of five relatively isolated regions of intraplate seismicity. Attempted convergence across this band is suggested by the consistent direction of the (horizontally oriented) axis of compression determined for a number of earthquakes near the Gilbert Islands [*Okal et al.*, 1986]. The direction of maximum compression indicates that these thrust events do not result from ridge push. *Kroenke and Walker* [1986] contend that seafloor structures, allegedly connecting two of these localities, are reminiscent of trench-forearc morphology and indicate underthrusting of the Pacific plate beneath the Indian-Australian plate. Seismic reflection profiles [*Kroenke and Walker*, 1986, Figure 3], however, reveal that the amplitude of the inferred structures is an order of magnitude less than those associated with documented convergence along the Azores-Gibraltar zone discussed above. A more conservative conclusion, determined in part on the basis of kinematic considerations, was offered by *Okal et al.* [1986], who concluded that significant convergence has not yet occurred anywhere along this zone. Two of the loci of active seismicity are believed to coincide with fracture zones [*Okal et al.*, 1986], although no obvious potential zone of weakness can be identified along the entire length of the proposed boundary. Also, both *Kroenke and Walker* [1986] and *Okal et al.* [1986] concluded that the regional compression results from congestion at the New Hebrides, Santa Cruz, and Solomon trenches. In these trenches the Australian plate is subducting beneath the Pacific plate; thus this proposed incipient plate boundary might represent the genesis of a major subduction zone within an overriding plate in response to congestion at nearby trenches.

For at least two decades, intraplate seismicity in the northeastern Indian Ocean has generated speculation concerning incipient trench formation [e.g., *Sykes*, 1970]. The common orientation of compressional axes associated with earthquakes between the Ninetyeast and Chagos-Laccadive Ridges indicates significant north-south compression [*Stein and Okal*, 1978; *Bergman and Solomon*, 1985; *Wiens*, 1986], consistent with the stress regime expected from congestion at the Himalayan collision zone accompanied by continued subduction at the adjacent Sunda Arc [*Cloetingh and Wortel*, 1986]. Seismic reflection profiles indicate east-west striking ridges and troughs with wavelengths of 100–300 km and as much as 3 km of relative relief [*Weissel et al.*, 1980; *Geller et al.*, 1983]. Because imaging of fault plane surfaces, as well as geoid undulations, indicate that this buckling or folding involves the underlying igneous basement and does not merely reflect sedimentary structures [*Weissel et al.*, 1980; *McAdoo and Sandwell*, 1985], a significant level of inplane compressional force is implied [*McAdoo and Sandwell*, 1985; *Zuber*, 1987]. Recent kinematic models of plate motions also support the existence of a diffuse zone of "intraplate" convergence (north–south at 0.5–1.5 cm/yr) in the equatorial Indian Ocean [*Minster and Jordan*, 1978; *Wiens et al.*, 1985, 1986; *DeMets et al.*, 1988]. It is interesting to note that if trench formation is occurring, it is within the ocean basin and not at the nearby passive continental margin; despite the fact that this margin is quite old, well sedimented, and subjected to at least the same level of compressional stress as the regions exhibiting intraplate seismicity [*Cloetingh and Wortel*, 1986].

Summary of Observations

Collectively, the observations described in this section suggest that (1) subduction zones nucleate intraoceanically, (2) that this often occurs in response to trench congestion, and (3) nucleation sometimes occurs along oceanic transform faults or fracture zones. In contrast, the geologic record, to the best of our knowledge, does not reveal a single example of an Atlantic-type margin evolving into an Andean-type margin without an intervening arc-continent collision and subduction polarity reversal.

REMAINING QUESTIONS: TRENCH FORMATION IN A PRISTINE OCEANIC BASIN

Our model obviously cannot account for the creation of the earliest subduction complex(es), nor does it immediately reveal how the first subduction zone(s) might develop within an Atlantic-type basin (i.e., one that is initially devoid of trenches). Questions concerning the genesis of the earliest subduction complexes cannot be answered presently and must be included among a larger category of issues concerning the inception of plate tectonics. Establishment of the earliest subduction zones within an Atlantic-type basin, however, is an issue that can be addressed with some confidence, because the present-day Atlantic has apparently already entered this stage of evolution.

Although quite small in comparison to the extensive subduction complexes of the Pacific and Indian oceans, the Lesser Antilles and Scotia arcs potentially represent precursors to a system of convergence zones that might ultimately result in the destruction of the Atlantic Ocean. Whereas the tectonic evolution of the remote Scotia Arc remains poorly understood, evolutionary constraints in the Caribbean region allow the process by which the first subduction zones are established within a "pristine" oceanic basin to be characterized to some degree. It is now generally accepted [e.g., *Burke*, 1988] that the Lesser Antilles Arc originated at the Pacific margin of the proto-Caribbean after rifting between North and South America generated oceanic crust in immediate contact with both the Atlantic and Pacific basins. From this intraoceanic boundary (possibly a former transform fault [*Pindell and Dewey*, 1982]) the arc migrated eastward, overriding 1000–2000 km of Caribbean/Atlantic seafloor, to arrive at its present position [*Pindell and Dewey*, 1982]. Only the central portion of the original arc survived this journey, however. The northern and southern extremities became involved in arc-continent collisions with promontory members of North and South America, leaving a trail of accreted ophiolitic and island arc terranes in their wake [e.g., *Burke*, 1988]. A consequence of these collisions may have been the establishment of a new subduction zone at the boundary between the Pacific and Caribbean (i.e., where the first trench originated), isolating the latter from the former [*Pindell and Dewey*, 1982; *Ross and Scotese*, 1988]. This arc system is currently represented by the southern portion of the Middle America Trench.

Although the Lesser Antilles Arc is apparently a consequence of Pacific seafloor paleostresses, these stresses were not transmitted through continental landmasses into an isolated Atlantic Basin.

Rather, this arc originated at a time when the Pacific Basin was in direct contact with the Atlantic Basin; a circumstance that presumably facilitated stress transmission from basin to basin. If it is assumed that this example is representative, the introduction of subduction zones into a previously pristine ocean basin might be viewed, figuratively, as a process in which an "infected" ocean basin comes into contact with, and infects, an uninfected basin. This is a particularly imaginative analogy when it is considered that, once the first subduction complexes have invaded the basin, the potential to multiply and eradicate the entire basin clearly exists (e.g., the proto-Atlantic).

CONCLUSION

It has been demonstrated that the only inplane compressional forces sufficient for trench formation are those associated with the congestion of mature subduction complexes. A consequence of this may be that transform faults and fracture zones represent favored sites of trench formation. Simple models of passive margin failure, despite their popularity, cannot be quantitatively substantiated and are not evident in the geologic record.

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REFERENCES

American Association of Petroleum Geologists, Plate-Tectonic Map of the Circum-Pacific Region, Southwest Quadrant, scale 1:10,000,000, Tulsa, Oklahoma, 1981.

Aspden, J. A., and W. J. McCourt, Mesozoic oceanic terrane in the central Andes of Colombia, *Geology, 14*, 415–418, 1986.

Aubouin, J., R. von Huene, M. Baltuck, R. Arnott, J. Bourgois, M. Filewicz, K. Kvenvolden, B. Leinert, T. McDonald, K. McDougall, Y. Ogawa, E. Taylor, and B. Winsborough, Leg 84 of the Deep Sea Drilling Project, Subduction without accretion: Middle America Trench off Guatemala, *Nature, 297*, 458–460, 1982.

Avedik, F., A. L. Camus, A. Ginsburg, L. Montadert, D. G. Roberts, and R. B. Whitmarsh, A seismic refraction and reflexion study of the continent-ocean transition beneath the north Biscay margin, *Philos. Trans. R. Soc. London, Ser. A, 305*, 5–25, 1982.

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

Bergman, E. A., and S. C. Solomon, Oceanic intraplate earthquakes: Implications for local and regional intraplate stress, *J. Geophys. Res., 85*, 5389–5410, 1980.

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

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

Bethke, C. M., Inverse hydrologic analysis of the distribution and origin of Gulf Coast-type geopressured zones, *J. Geophys. Res., 91*, 6535–6545, 1986.

Bird, P., Stress and temperature in subduction shear zones: Tonga and Mariana, *Geophys. J. R. Astron. Soc., 55*, 411–434, 1978.

Bodri, B., and A. M. Jessop, Geothermal model of the continental margins of eastern Canada, *Tectonophysics, 164*, 139–150, 1989.

Bonatti, E., Vertical tectonism in oceanic fracture zones, *Earth Planet. Sci. Lett., 37*, 369–379, 1978.

Bonatti, E., and P. R. Hamlyn, Oceanic ultramafic rocks, in *The Sea*, vol. 7, *The Oceanic Lithosphere*, edited by C. Emiliani, pp. 241–283, John Wiley, New York, 1981.

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

Bratt, S. R., E. A. Bergman, and S. C. Solomon, Thermoelastic stress: How important as a cause of earthquakes in young oceanic lithosphere?, *J. Geophys. Res., 90*, 10,249–10,260, 1985.

Buforn, E., A. Udías, and M. A. Colombás, Seismicity, source mechanism and tectonics of the Azores-Gibraltar plate boundary, *Tectonophysics, 152*, 89–118, 1988.

Burke, K., Tectonic evolution of the Caribbean, *Annu. Rev. Earth Planet. Sci., 16*, 201–230, 1988.

Byerlee, J., Friction of rocks, *Pure Appl. Geophys., 116*, 615–626, 1978.

Byrne, D. E., D. M. Davis, and L. R. Sykes, Loci and maximum size of thrust earthquakes and the mechanics of the shallow regions of subduction zones, *Tectonics, 7*, 833–857, 1988.

Calvert, A. J., and C. G. Potts, Seismic evidence for hydrothermally altered mantle beneath old crust in the Tydeman fracture zone, *Earth Planet. Sci. Lett., 75*, 439–449, 1985.

Caristan, Y., The transition from high temperature creep to fracture in Maryland Diabase, *J. Geophys. Res., 87*, 6781–6790, 1982.

Casey, J. F., and J. F. Dewey, Initiation of subduction zones along transforms and accreting plate boundaries, triple junction evolution, and forearc spreading centres: Implications for ophiolitic geology and obduction, in Ophiolites and Oceanic Lithosphere, edited by I. G. Gass, S. J. Lippard, and A. W. Shelton, *Geol. Soc. Spec. Publ. London, 13*, 269–290, 1984.

Chamot-Rooke, N., and X. Le Pichon, Zenisu Ridge: Mechanical model of formation, *Tectonophysics, 160*, 175–193, 1989.

Chapple, W. M., and T. E. Tullis, Evaluation of the forces that drive the plates, *J. Geophys. Res., 82*, 1967–1984, 1977.

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

Chopra, P. N., and M. S. Paterson, The experimental deformation of dunite, *Tectonophysics, 78*, 453–473, 1981.

Christensen, D. H., and L. J. Ruff, Seismic coupling and outer rise earthquakes, *J. Geophys. Res., 93*, 13,421–13,444, 1988.

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

Cloetingh, S., R. Wortel, and N. J. Vlaar, On the initiation of subduction zones, *Pure Appl. Geophys., 129*, 7–25, 1989.

Collot, J. Y., J. Daniel, and R. V. Burne, Recent tectonics associated with the subduction/collision of the D'Entrecasteaux Zone in the central New Hebrides, *Tectonophysics, 112*, 325–356, 1985.

Coney, P. J., D. L. Jones, and J. W. H. Monger, Cordilleran suspect terranes, *Nature, 288*, 329–333, 1980.

Cooper, A. K., M. S. Marlow, and D. W. Scholl, The Bering Sea: A multifarious marginal basin, in *Island Arcs, Deep Sea Trenches, and Back-Arc Basins, Maurice Ewing Ser.*, vol. 1, edited by M. Talwani and W. C. Pitman III, pp. 437–450, AGU, Washington, D. C., 1977.

Craig, C. H., and D. McKenzie, The existence of a thin low-viscosity layer beneath the lithosphere, *Earth Planet. Sci. Lett., 78*, 420–426, 1986.

Curray, J. R., G. G. Shor, R. W. Raitt, and M. Henry, Seismic refraction and reflection studies of crustal structure of the eastern Sunda and western Banda arcs, *J. Geophys. Res., 82*, 2479–2489, 1977.

Currie, K. L., R. K. Pickerill, and G. E. Pajari, An Early Paleozoic plate tectonic model of Newfoundland, *Earth Planet. Sci. Lett., 48*, 8–14, 1980.

Daines, S. R., Aquathermal pressuring and geopressure evalution, *Am. Assoc. Pet. Geol. Bull., 60*, 931–939, 1982.

Davies, G. F., Mechanics of subducted lithosphere, *J. Geophys. Res., 85*, 6304–6318, 1980.

Davies, G. F., Subduction zone stresses: Constraints from mechanics and from topographic and geoid anomalies, *Tectonophysics, 99*, 85–98, 1983.

Davis, D. M., and R. von Huene, Inferences on sediment strength and fault friction from structures at the Aleutian Trench, *Geology, 15*, 517–522, 1987.

Davis, D., J. Suppe, and F. A. Dahlen, Mechanics of fold-and-thrust belts and accretionary wedges, *J. Geophys. Res., 88*, 1153–1172, 1983.

DeMets, C., R. G. Gordon, and D. F. Argus, Intraplate deformation and closure of the Australia-Antarctica-Africa plate circuit, *J. Geophys. Res., 93*, 11,877–11,897, 1988.

Dengo, C. A., and J. M. Logan, Implications of the mechanical and frictional behavior of serpentinite to seismogenic faulting, *J. Geophys. Res., 86*, 10,771–10,782, 1981.

Diament, M. J.-C. Sibuet, and J. Dubois, Isostatic response of the large-offset Atlantic equatorial fracture zones, *Mar. Geophys. Res., 8*, 243–264, 1986.

Dickinson, W. R., Plate tectonics and the continental margin of California, in *The Geotectonic Development of California, Rubey Vol. 1*, edited by W. G. Ernst, pp. 1–28, Prentice-Hall, Englewood Cliffs, N. J., 1981.

Dickinson, W. R., and D. R. Seely, Structure and stratigraphy of forearc regions, *Am. Assoc. Pet. Geol. Bull.*, *63*, 2–31, 1979.

Dilik, Y., E. M. Moores, and M. C. Erskine, Ophiolitic thrust nappes in western Nevada: Implications for the Cordilleran orogen, *J. Geol. Soc. London*, *145*, 969–975, 1988.

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

Feininger, T., Allochthonous terranes in the Andes of Ecuador and northwestern Peru, *Can. J. Earth Sci.*, *24*, 266–278, 1987.

Feininger, T., and C. R. Bristow, Cretaceous and Paleogene geologic history of coastal Ecuador, *Geol. Rundsch.*, *69*, 849–874, 1980.

Feininger, T., and M. K. Seguin, Simple Bouguer gravity anomaly field and the inferred crustal structure of continental Ecuador, *Geology*, *11*, 40–44, 1983.

Fisher, R. L., and C. G. Engel, Ultramafic and basaltic rocks dredged from the nearshore flank of the Tonga Trench, *Geol. Soc. Am. Bull.*, *80*, 1373–1378, 1969.

Fitch, T. J., Plate convergence, transcurrent faults, and internal deformation adjacent to Southeast Asia and the Western Pacific, *J. Geophys. Res.*, *77*, 4432–4460, 1972.

Fleitout, L., and C. Froidevaux, Thermal and mechanical evolution of shear zones, *J. Struct. Geol.*, *2*, 159–164, 1980.

Forsyth, D., and S. Uyeda, On the relative importance of the driving forces of plate motion, *Geophys. J. R. Astron. Soc.*, *43*, 163–200, 1975.

Francis, T. J. G., Serpentinization faults and their role in the tectonics of slow spreading ridges, *J. Geophys. Res.*, *86*, 11,616–11,622, 1981.

Froidevaux, C., and B. L. Isacks, The mechanical state of the lithosphere in the Altiplano-Puna segment of the Andes, *Earth Planet. Sci. Lett.*, *71*, 305–314, 1984.

Fukao, Y., Thrust faulting at a lithospheric plate boundary: The Portugal earthquake of 1969, *Earth Planet. Sci. Lett.*, *18*, 205–216, 1973.

Geller, C. A., J. K. Weissel, and R. N. Anderson, Heat transfer and intraplate deformation in the Central Indian Ocean, *J. Geophys. Res.*, *88*, 1018–1032, 1983.

Grimison, N. L., and W.-P. Chen, The Azores-Gibraltar plate boundary: Focal mechanisms, depths of earthquakes, and their tectonic implications, *J. Geophys. Res.*, *91*, 2029–2047, 1986.

Hansen, F. D., and N. L. Carter, Creep of selected crustal rocks at 1000 MPa (abstract), *Eos Trans. AGU*, *63*, 437, 1982.

Hansen, V. L., Yukon-Tanana terrane: A partial acquittal, *Geology*, *18*, 365–369, 1990.

Hatcher, R. D., and G. W. Viele, The Appalachian/Ouachita Orogens: United States and Mexico, in Perspectives in Regional Geological Synthesis, Planning for the Geology of North America, *D-NAG Spec. Publ. 1*, edited by A. R. Palmer, pp. 67–75, Geological Society of America, Boulder, Colo., 1982.

Hawkins, J. W., S. H. Bloomer, C. A. Evans, and J. T. Melchoir, Evolution of intra-oceanic arc-trench systems, *Tectonophysics*, *102*, 175–205, 1984.

Haworth, R. T., J.-P. Lefort, and H. G. Miller, Geophysical evidence for an east-dipping Appalachian subduction zone beneath Newfoundland, *Geology*, *6*, 522–526, 1978.

Haxby, W. F., and E. M. Parmentier, Thermal contraction and the state of stress in the oceanic lithosphere, *J. Geophys. Res.*, *93*, 6419–6429, 1988.

Hegarty, K. A., J. K. Weissel, and D. E. Hayes, Convergence at the Caroline-Pacific Plate Boundary: Collision and subduction, in *The Tectonic and Geologic Evolution of Southeast Asian Seas and Islands, Part 2*, *Geophys. Monogr. Ser.*, vol. 27, edited by D. E. Hayes, pp. 326–348, AGU, Washington, D. C., 1982.

Henderson, W. G., Cretaceous to Eocene volcanic arc activity in the Andes of Northern Ecuador, *J. Geol. Soc. London*, *136*, 367–378, 1979.

Hilde, T. W. C., S. Uyeda, and L. Kroenke, Evolution of the western Pacific and its margin, *Tectonophysics*, *38*, 145–165, 1977.

Hyndman, R. D., M. G. Langseth, and R. P. von Herzen, Deep Sea Drilling Project geothermal measurements: A review, *Rev. Geophys.*, *25*, 1563–1582, 1987.

Ibrahim, A.-B. K., G. V. Latham, and J. Ladd, Seismic refraction and reflection measurements in the Middle America Trench offshore Guatemala, *J. Geophys. Res.*, *84*, 5643–5649, 1979.

Jacobi, R. D., Peripheral bulge — A causal mechanism for the Lower/Middle Ordovician unconformity along the western margin of the Northern Appalachians, *Earth Planet. Sci. Lett.*, *56*, 245–251, 1981.

Jones, J. G., Aleutian enigma: A clue to transformation in time, *Nature*, *229*, 400–403, 1971.

Kanamori, H., The energy release of great earthquakes, *J. Geophys. Res.*, *82*, 2981–2987, 1977.

Karig, D. E., Initiation of subduction zones: Implications for arc evolution and ophiolite development, in Trench-Forearc Geology, edited by J. K. Leggett, *Geol. Soc. Spec. Publ. London*, *10*, 563–576, 1982.

Karner, G. D., N. L. Grimison, W.-P. Chen, and J. K. Weissel, SEASAT derived gravity and geoid anomalies of the Azores-Gibraltar plate boundary: A diffuse zone of ocean-ocean convergence near Gibraltar (abstract), *Eos Trans. AGU*, *66*, 361, 1985.

Kroenke, L. W., and D. A. Walker, Evidence for the formation of a new trench in the Western Pacific, *Eos Trans. AGU*, *67*, 145–146, 1986.

Lachenbruch, A. H., and J. H. Sass, Heat flow and energetics of the San Andreas fault zone, *J. Geophys. Res.*, *85*, 6185–6222, 1980.

Lachenbruch, A. H., and G. A. Thompson, Oceanic ridges and transform faults: Their intersection angles and resistance to plate motion, *Earth Planet. Sci. Lett.*, *15*, 116–122, 1972.

Lay, T., H. Kanamori, and L. Ruff, The asperity model and the nature of large earthquakes, *Earthquake Predict. Res.*, *1*, 3–71, 1982.

Lebras, M., F. Mégard, C. Dupuy, and J. Dostal, Geochemistry and tectonic setting of pre-collision Cretaceous and Paleogene volcanic rocks of Ecuador, *Geol. Soc. Am. Bull.*, *99*, 569–578, 1987.

Lewis, B. T., The East Pacific Rise and the thermal model, *J. Geophys. Res.*, *88*, 3348–3354, 1983.

Lowrie, A., C. Smoot, and R. Batiza, Are oceanic fracture zones locked and strong or weak?: New evidence for volcanic activity and weakness, *Geology*, *14*, 242–245, 1986.

MacDonald, A. H., and W. S. Fyfe, Rate of serpentinization in seafloor environments, *Tectonophysics*, *116*, 123–135, 1985.

McAdoo, D. C., and D. T. Sandwell, Folding of oceanic lithosphere, *J. Geophys. Res.*, *90*, 8563–8569, 1985.

McCaffrey, R., Active tectonics of the eastern Sunda and Banda arcs, *J. Geophys. Res.*, *93*, 15,163–15,182, 1988.

McCaffrey, R., and J. Nábělek, The geometry of back arc thrusting along the Eastern Sunda Arc, Indonesia: Constraints from earthquake and gravity data, *J. Geophys. Res.*, *89*, 6171–6179, 1984.

McCaffrey, R., P. Molnar, S. W. Roecker, and Y. S. Joyodiwiryo, Microearthquake seismicity and fault plane solutions related to arc-continent collision in the Eastern Sunda Arc, Indonesia, *J. Geophys. Res.*, *90*, 4511–4528, 1985.

McKenzie, D. P., Active tectonics of the Mediterranean region, *Geophys. J. R. Astron. Soc.*, *30*, 109–185, 1972.

McKenzie, D. P., The initiation of trenches: A finite amplitude instability, in *Island Arcs, Deep Sea Trenches, and Back-Arc Basins, Maurice Ewing Ser.*, vol. 1, edited by M. Talwani and W. C. Pitman III, pp. 57–61, AGU, Washington, D. C., 1977.

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

Melosh, H. J., Shear stress on the base of a lithospheric plate, *Pure Appl. Geophys.*, *115*, 429–439, 1977.

Minster, J. B., and T. H. Jordan, Present-day plate motions, *J. Geophys. Res.*, *83*, 5331–5354, 1978.

Miyashiro, A., *Metamorphism and Metamorphic Belts*, 492 pp., Allen and Unwin, Winchester, Mass., 1973.

Moore, J. C., and E. A. Silver, Continental margin tectonics: Submarine accretionary prisms, *Rev. Geophys.*, *25*, 1305–1312, 1987.

Moores, E. M., Origin and emplacement of ophiolites, *Rev. Geophys. Space Phys.*, *20*, 735–760, 1982.

Moreira, V. S., Seismotectonics of Portugal and its adjacent area in the Atlantic, *Tectonophysics*, *117*, 85–96, 1985.

Morgan, J. P., and E. M. Parmentier, Lithospheric stress near a ridge-transform intersection, *Geophys. Res. Lett.*, *11*, 113–116, 1984.

Neuman, R. B., Geology and paleobiology of islands in the Ordovician Iapetus Ocean: Review and implications, *Geol. Soc. Am. Bull.*, *95*, 1188–1201, 1984.

Ogawa, Y., and J. Naka, Emplacement of ophiolitic rocks in forearc areas: Examples from central Japan and Izu-Mariana-Yap island arc system, in Ophiolites and Oceanic Lithosphere, edited by I. G. Gass, S. J. Lippard, and A. W. Shelton, *Geol. Soc. Spec. Publ. London*, *13*, 291–300, 1984.

Okal, E. A., D. F. Woods, and T. Lay, Intraplate deformation in the Samoa-Gilbert-Ralik area: A prelude to a change of plate boundaries in the Southwest Pacific?, *Tectonophysics*, *132*, 69–77, 1986.

Oldenburg, D. W., and J. N. Brune, An explanation for the orthogonality of ocean ridges and transform faults, *J. Geophys. Res.*, *80*, 2575–2585, 1975.

Oxburgh, E. R., and E. M. Parmentier, Compositional and density stratification in oceanic lithosphere- causes and consequences, *J. Geol. Soc. London*, *133*, 343–355, 1977.

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

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

Phillips, R. J., Convection-driven tectonics on Venus, *J. Geophys. Res.*, *95*, 1301–1316, 1990.

Pindell, J., and J. F. Dewey, Permo-Triassic reconstruction of western Pangea and the evolution of the Gulf of Mexico/Caribbean region, *Tectonics*, *1*, 179–211, 1982.

Price, R. A., and R. D. Hatcher, Tectonic significance of similarities in the evolution of the Alabama-Pennsylvania Appalachians and the Alberta-British Columbia Canadian Cordillera, in Contributions to the Tectonics and Geophysics of Mountain Chains, edited by R. D. Hatcher, H. Williams, and I. Zietz, *Mem. Geol. Soc. Am.*, *158*, 149–160, 1983.

Purdy, G. M., The eastern end of the Azores-Gibraltar plate boundary, *Geophys. J. R. Astron. Soc.*, *43*, 973–1000, 1975.

Raleigh, C. B., and J. Evernden, Case for low deviatoric stress in the lithosphere, in *Mechanical Behavior of Crustal Rocks*, *Geophys. Monogr. Ser.*, vol. 24, edited by N. L. Carter, M Friedman, J. M. Logan, and D. W. Stearns, pp. 173–186, AGU, Washington, D. C., 1981.

Roperch, P., F. Mégard, C. Laj, T. Mourier, T. M. Clube, and C. Noblet, Rotated oceanic blocks in Western Ecuador, *Geophys. Res. Lett.*, *14*, 558–561, 1987.

Ross, M. I., and C. R. Scotese, A hierarchical tectonic model of the Gulf of Mexico and Caribbean region, *Tectonophysics*, *155*, 139–168, 1988.

Ruff, L., and A. Cazenave, SEASAT geoid anomalies and the Macquarie Ridge complex, *Phys. Earth Planet. Inter.*, *38*, 59–69, 1985.

Ruff, L. J., J. W. Given, C. O. Sanders, and C. M. Sperber, Large earthquakes in the Macquarie Ridge Complex: Transitional tectonics and subduction initiation, *Pure Appl. Geophys.*, *129*, 71–129, 1989.

Sandwell, D. T., Thermomechanical evolution of oceanic fracture zones, *J. Geophys. Res.*, *89*, 11,401–11,413, 1984.

Schweickert, R. A., and W. S. Snyder, Paleozoic plate tectonics of the Sierra Nevada and adjacent regions, in *The Geotectonic Development of California*, *Rubey Vol. 1*, edited by W. G. Ernst, pp. 182–201, Prentice-Hall, Englewood Cliffs, N. J., 1981.

Sclater, J. G., Anderson, R. N., and M. L. Bell, Elevation of ridges and evolution of the central eastern Pacific, *J. Geophys. Res.*, *76*, 7888–7915, 1971.

Shi, Y., and C.-Y. Wang, High pore pressure generation in sediments in front of the Barbados Ridge complex, *Geophys. Res. Lett.*, *12*, 773–776, 1985.

Shi, Y., and C.-Y. Wang, Pore pressure generation in sedimentary basins: Overloading versus aquathermal, *J. Geophys. Res.*, *91*, 2153–2162, 1986.

Shi, Y., and C.-Y. Wang, Generation of high pore pressures in accretionary prisms: Inferences from the Barbados subduction complex, *J. Geophys. Res.*, *93*, 8893–8910, 1988.

Shiraki, K., Metamorphic basement rocks of Yap Islands, Western Pacific: Possible oceanic crust beneath an island arc, *Earth Planet. Sci. Lett.*, *13*, 167–174, 1971.

Sibuet, J.-C., and B. Veyrat-Peinet, Gravimetric model of the Atlantic equatorial fracture zones, *J. Geophys. Res.*, *85*, 943–954, 1980.

Silver, E. A., D. Reed, R. McCaffrey, and Y. Joyodiwiryo, Back arc thrusting in the Eastern Sunda Arc, Indonesia: A consequence of arc-continent collision, *J. Geophys. Res.*, *88*, 7429–7448, 1983.

Solomon, S. C., E. A. Bergman, W. S. D. Wilcock, and G. M. Purdy, On the state of stress near oceanic transforms and fracture zones (abstract), *Eos Trans. AGU*, *70*, 469, 1989.

Speed, R. C., and N. H. Sleep, Antler orogeny and foreland basin: A model, *Geol. Soc. Am. Bull.*, *93*, 815–828, 1982.

Steckler, M. S., and A. B. Watts, Subsidence history and tectonic evolution of Atlantic-type continental margins, in *Dynamics of Passive Margins*, *Geodyn. Ser.*, vol. 6, edited by R. A. Scrutton, pp. 184–196, AGU, Washington, D. C., 1982.

Stein, C. A., S. Cloetingh, and R. Wortel, SEASAT-derived gravity constraints on stress and deformation in the northeastern Indian Ocean, *Geophys. Res. Lett.*, *16*, 823–826, 1989.

Stein, S., and E. A. Okal, Seismicity and tectonics of the Ninetyeast Ridge area: Evidence for internal deformation of the Indian plate, *J. Geophys. Res.*, *83*, 2233–2245, 1978.

Stevens, R. K., D. F. Strong, and B. F. Kean, Do some eastern Appalachian ultramafic rocks represent mantle diapirs produced above a subduction zone?, *Geology*, *2*, 175–178, 1974.

Strong, D. F., Volcanic regimes of the Newfoundland Appalachians, in Volcanic Regimes in Canada, edited by W. R. A. Baragar, L. C. Coleman, and J. M. Hall, *Geol. Assoc. Can. Spec. Pap.*, *16*, 61–90, 1977.

Strong, D. F., W. L. Dickson, C. F. O'Driscoll, B. F. Kean, and R. K. Stevens, Geochemical evidence for an east-dipping Appalachian subduction zone in Newfoundland, *Nature*, *248*, 37–39, 1974.

Suppe, J., *Principles of Structural Geology*, 537 pp., Prentice-Hall, Englewood Cliffs, N. J., 1985.

Sykes, L. R., Seismicity of the Indian Ocean and a possible nascent island arc between Ceylon and Australia, *J. Geophys. Res.*, *75*, 5041–5055, 1970.

Tamsett, D., and R. C. Searle, Structure and development of the midocean ridge plate boundary in the Gulf of Aden: Evidence from GLORIA side scan sonar, *J. Geophys. Res.*, *93*, 3157–3178, 1988.

Tempelman-Kluit, D. J., Transported cataclasite, ophiolite and granodiorite in Yukon: Evidence of arc-continent collision, *Pap. Geol. Surv. Can.*, *79–14*, 1–27, 1979.

Tréhu, A. M., Klitgord, K. D., Sawyer, D. S., and Buffler, R. T., Atlantic and Gulf of Mexico continental margins, in Geophysical Framework of the Continental US, edited by L. C. Pakiser and W. D. Mooney, *Mem. Geol. Soc. Am.*, *172*, 349–382, 1989.

Turcotte, D. L., and J. L. Ahern, On the thermal and subsidence history of sedimentary basins, *J. Geophys. Res.*, *82*, 3762–3766, 1977.

Turcotte, D. L., and G. Schubert, *Geodynamics: Applications of Continuum Physics to Geological Problems*, 450 pp., John Wiley, New York, 1982.

Turcotte, D. L., J. L. Ahern, and J. M. Bird, The state of stress at continental margins, *Tectonophysics*, *42*, 1–28, 1977.

Uyeda, S., Chilean vs. Mariana type subduction zones with remarks on arc volcanism and collision tectonics, in *Circum-Pacific Orogenic Belts and Evolution of the Pacific Ocean Basin*, *Geodyn. Ser.*, vol. 18, edited by J. W. H. Monger and J. Francheteau, pp. 1–7, AGU, Washington, D. C., 1987.

Uyeda, S., and Z. Ben-Avraham, Origin and development of the Philippine Sea, *Nature*, *240*, 176–178, 1972.

Vogt, P. R., Subduction and aseismic ridges, *Nature*, *241*, 189–191, 1973.

von Huene, R., Tectonic processes along the front of modern convergent margins- Research of the past decade, *Annu. Rev. Earth Planet. Sci.*, *12*, 359–381, 1984.

von Huene, R., and H. Lee, The possible significance of pore fluid pressures in subduction zones, in Studies of Continental Margin Geology, edited by J. S. Watkins and C. L. Drake, *AAPG Mem. 34*, 781–791, 1983.

Watts, A. B., and M. S. Steckler, Subsidence and eustasy at the continental margin of eastern North America, in *Deep Drilling Results in the Atlantic Ocean: Continental Margins and Paleoenvironment*, *Maurice Ewing Ser.*, vol. 3, edited by M. Talwani, W. Hay, and W. B. F. Ryan, pp. 218–234, AGU, Washington, D. C., 1979.

Weissel, J. K., R. N. Anderson, and C. A. Geller, Deformation of the Indo-Australian plate, *Nature*, *287*, 284–291, 1980.

White, R. S., R. S. Detrick, M. C. Sinha, and M. H. Cormier, Anomalous seismic crustal structure of oceanic fracture zones, *Geophys. J. R. Astron. Soc.*, *79*, 779–798, 1984.

Wiens, D. A., Historical sesimicity near Chagos: A complex deformation zone in the equatorial Indian Ocean, *Earth Planet. Sci. Lett.*, *76*, 350–360, 1986.

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

Wiens, D. A., and S. Stein, Intraplate seismicity and stresses in young oceanic lithosphere, *J. Geophys. Res.*, *89*, 11,442–11,464, 1984.

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

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

Wiens, D. A., S. Stein, C. Demets, R. G. Gordon, and C. Stein, Plate tectonic models for Indian Ocean "intraplate" deformation, *Tectonophysics*, *132*, 37–48, 1986.

Williams, H., Appalachian orogen in Canada, *Can. J. Earth Sci.*, *16*, 792–807, 1979.

Williamson, P. E., T. D. Jones, and K. F. McCue, Evidence for crustal thickening and shortening of the overriding plate during incipient plate/plate subduction, *J. Geodyn.*, *11*, 1–15, 1989.

Zuber, M. T., Compression of oceanic lithosphere: An analysis of intraplate deformation in the Central Indian Basin, *J. Geophys. Res.*, 92, 4817–4825, 1987.

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