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AGE DEPENDENCE OF OCEANIC INTRAPLATE SEISMICITY AND IMPLICATIONS FOR LITHOSPHERIC EVOLUTION

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Abstract. We have determined depths for 16 oceanic intraplate earthquakes using body wave modeling. A data set composed of these and 11 other well-constrained depths shows that the maximum depth of seismicity deepens with increasing lithospheric age and appears to be bounded by a 700^{o}-800^{o}C isotherm. The maximum depth to which intraplate earthquakes occur is approximately equal to the flexural elastic thickness but much less than the seismic thickness from surface wave dispersion. The maximum faulting depth is consistent with the predictions of rapid weakening at high temperatures for dry olivine rheologies determined by experimental rock mechanics. Oceanic intraplate earthquakes can occur at greater depths and temperatures than has previously been reported for continental crustal events, presumably because oceanic lithosphere is stronger at high temperatures. Both the number of earthquakes and the cumulative seismic moment decrease with lithospheric age. The cause of this trend is unclear: possibilities include nonrandom location of preexisting weak zones with respect to the age provinces, fault strengthening with age, or a decrease in the stress level. A decrease in intraplate stress with age is opposite that expected from "ridge push" stresses, which increase with age, but may result either from thermoelastic stresses which are greatest in young lithosphere or near-ridge stress concentrations due to local heterogeneities in the spreading process.

Introduction

In recent years, considerable attention has been focused on the evolution of the mechanical properties of the oceanic lithosphere, the upper layer of strong material which thickens as the plate cools. Lithospheric thickness is not directly measured but is inferred from a variety of techniques which yield different results. Surface wave dispersion studies [Leeds et al., 1974; Forsyth, 1975] measure the "seismic thickness," the thickness of the region above the low velocity zone. The effective elastic thickness of the lithosphere is determined from the flexure below midplate loads [Watts et al., 1975, 1980; Detrick and Watts, 1979] and at oceanic trenches [Watts and Talwani, 1974; McAdoo et al., 1978; Bodine and Watts, 1979; Chapple and Forsyth, 1979]. This flexural thickness inferred from long-term loads is much less than the thickness measured with surface waves (which can be considered shorter-term loads), but both thicknesses increase with age as the lithosphere cools. Olivine rheologies determined by experimental rock mechanics [Goetze and Evans, 1979; Brace and Kohlstedt, 1980] predict increasing ductility and decreasing strength at depth due to high temperatures, in accord with the thickness measurements. Decreased seismic wave attenuation [Canas and Mitchell, 1978, 1981] and lower rates of intraplate volcanism with increasing age [Batiza, 1981, 1982] have also been proposed as results of the thickening of the oceanic lithosphere.

Earthquake focal mechanisms provide important constraints on the state of stress within oceanic plates [Sykes and Sbar, 1974; Richardson and Solomon, 1977; Bergman and Solomon, 1980; Okal, 1983]. Most events far from plate boundaries indicate compressive stress; this stress is thought to result from the gravitational force due to lithospheric thickening with age ("ridge push"), which is considered a fundamental plate driving force [Hales, 1969; Lister, 1975; Forsyth and Uyeda, 1975; Richardson et al., 1979; Parsons and Richter, 1980; Dahlen, 1981]. In contrast, there is no direct way to measure variations in the magnitude of intraplate stresses from earthquake data. One possible way to infer such changes would be to look for variations in the level of seismicity. Since the magnitude of stress is predicted by driving force calculations for various assumptions, even an approximate method of estimating stress levels would be useful. Such data may also reveal whether stress sources other than plate driving forces are responsible for significant seismicity.

The depths of intraplate earthquakes provide important constraints on the rheology of lithosphere. Studies of the depths of earthquakes in continental regions [Brace and Byerlee, 1970; Sibson, 1982; Meissner and Strehlau, 1982; Chen and Molnar, 1983] show that except for regions of recent continental convergence, a close inverse correlation exists between the maximum depth of earthquakes and the geothermal gradient. Seismicity seems restricted to shallow depths cool enough that rocks can sustain stresses necessary for seismic fracture. Experimental olivine flow laws [Goetze and Evans, 1979] suggest that similar behavior should be expected in oceanic plates since the oceanic lithosphere is probably incapable of supporting the necessary stresses at temperatures greater than 700^{o}-800^{o}C [Brace and Kohlstedt, 1980; Kirby, 1980]. The maximum depth to which oceanic intraplate seismicity occurs thus should reflect both the mechanical and thermal properties of the lithosphere. In particular, plate cooling models imply that this limiting depth should increase with lithospheric age in a predictable fashion.

One limitation of studies of intraplate earthquakes is that the depths of most events are poorly known. Sykes and Sbar [1974] speculate that all oceanic intraplate seismicity may be concentrated in the upper few kilometers of the oceanic lithosphere, and thus stress orientations

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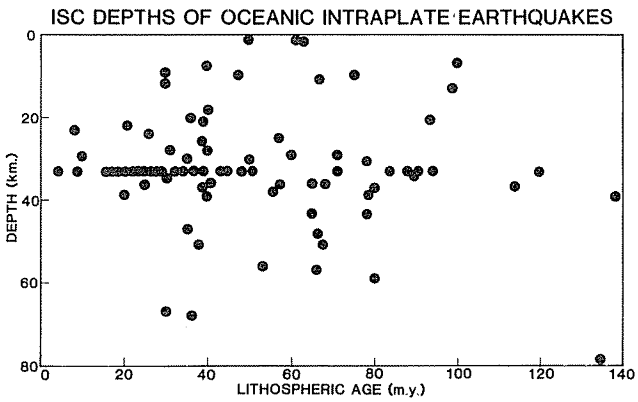

Fig. 1. ISC depths for oceanic intraplate earthquakes from the Bergman and Solomon [1980] compilation with $m_b \geq 5.0$ plotted against lithospheric age. The data are too scattered to show any trend, especially since events for which the program fails to converge are assigned a 33-km nominal depth.

derived from focal mechanisms may represent the state of stress only in the uppermost lithosphere. Therefore, in addition to providing constraints on rheology, better focal depth resolution should provide a better understanding of the distribution of stress within the oceanic lithosphere.

Depth Determination

The only focal depth data currently available for most oceanic intraplate earthquakes are derived from teleseismic location programs, commonly those of the International Seismological Centre (ISC). The depths are often substantially less accurate than their formal precision implies, since stations are generally far from the event. Figure 1 shows ISC depths for a data set of oceanic intraplate earthquakes selected by using all events in Bergman and Solomon's [1980] compilation with body wave magnitude $m_b \geq 5.0$. These depths show no correlation with lithospheric age at the epicenter; many are assigned the 33-km default depth used when the location program fails to converge.

In this study we have determined focal depths for as many oceanic intraplate earthquakes as possible by using standard body wave modeling techniques [Fukao, 1970; Langston and Helmberger, 1975; Kanamori and Stewart, 1976; Helmberger and Burdick, 1979]. Long-period P waveforms are quite sensitive to focal depth and thus provide reasonably accurate data for depth determination. Figure 2 shows examples of this process for four of the events studied; the model depth which produces synthetic seismograms that best fit the data is often significantly different from the ISC depth, while synthetics computed for the ISC depth do not match the data well.

Waveforms of body waves also depend on source parameters other than focal depth. Fortunately, the time separation between major arrivals on the seismogram, which provides most of the depth information, is reasonably insensitive to focal mechanism and a realistic range of source-time functions. In particular, because many of the events studied are dip slip faults for which most teleseismic arrivals leave the focal sphere far from nodal planes, uncertainties in focal mechanism do not pose significant difficulties.

We studied 16 events, listed in Table 1, using published focal mechanisms, or slight variations thereof, for all but five events. Four events (October 31, 1965, Indian Ocean; November 20, 1974, South Atlantic; August 30, 1976, Caroline Basin; and August 3, 1978, Indian Ocean) do not have published focal mechanisms. Figure 3 shows mechanisms we constructed for these events based on first motions and waveform modeling. Also shown is a new strike slip mechanism for the September 3, 1968, Puerto Rico event; a previously published thrust mechanism [Sykes and Sbar, 1974] does not fit the data as well.

Approximately 10 records from stations at distances of 30^o–90^o were modeled for each event, with 5-km depth increments. The depths for which synthetics best fit the data are given in Table 1. All events but one were modeled as single point sources, using a 3.5-s trapezoidal time function appropriate to these moderate ($M_s < 6.5$) size events. This choice has little effect on our depth results since the P waveform is relatively insensitive to time function variations as long as the total time function length is less than about 6 s. One earthquake, the May 9, 1971, Southeast Pacific event [Forsyth, 1973] was modeled as a triple shock (Figure 4). The timing of the three shocks was determined by the three compressive pulses at the beginning of the long-period P waveform, the largest of which can be seen as a distinct later arrival on the short-period records. The depth constraint comes from the time separation between the three initial pulses and the later broad up pulse.

Except for an event in the thickly sedimented Bay of Bengal, all synthetics were calculated using a water layer of depth appropriate for each event overlying a half space. This model successfully reproduces the major features of the data: inclusion of crustal layering has no real effect on the depth determination in most oceanic areas since for long-period P waves the major reflectors are the free surface and the sea bottom [Stein and Kroeger, 1980]. For example, the depth of the October 7, 1965, South China Sea event determined with the water half space model is 5 km [Wang et al., 1979]; detailed modeling using crustal layering produces no change in this depth (G. Kroeger and R. Geller, unpublished manuscript, 1982). Similarly, the depth of the September 12, 1965, Chagos Bank event was estimated as 15 km by Stein [1978] using a half space structure and at 12±2 km by G. Kroeger and R. Geller (unpublished manuscript, 1982) using a water-crust-mantle structure. One exception to this is when events occur under thick sediments. The August 30, 1973, Bay of Bengal event was modeled using several crustal layers to simulate the 4.5 km of sediment over the epicenter.

Though crustal layering may often be neglected, near-source velocities have significant effects since depths were estimated largely from the time separation between the direct and various reflected arrivals. Since velocity structure is not well known in most oceanic areas, we chose a standard half-space velocity of 7.4 km/s for most events, corresponding to a

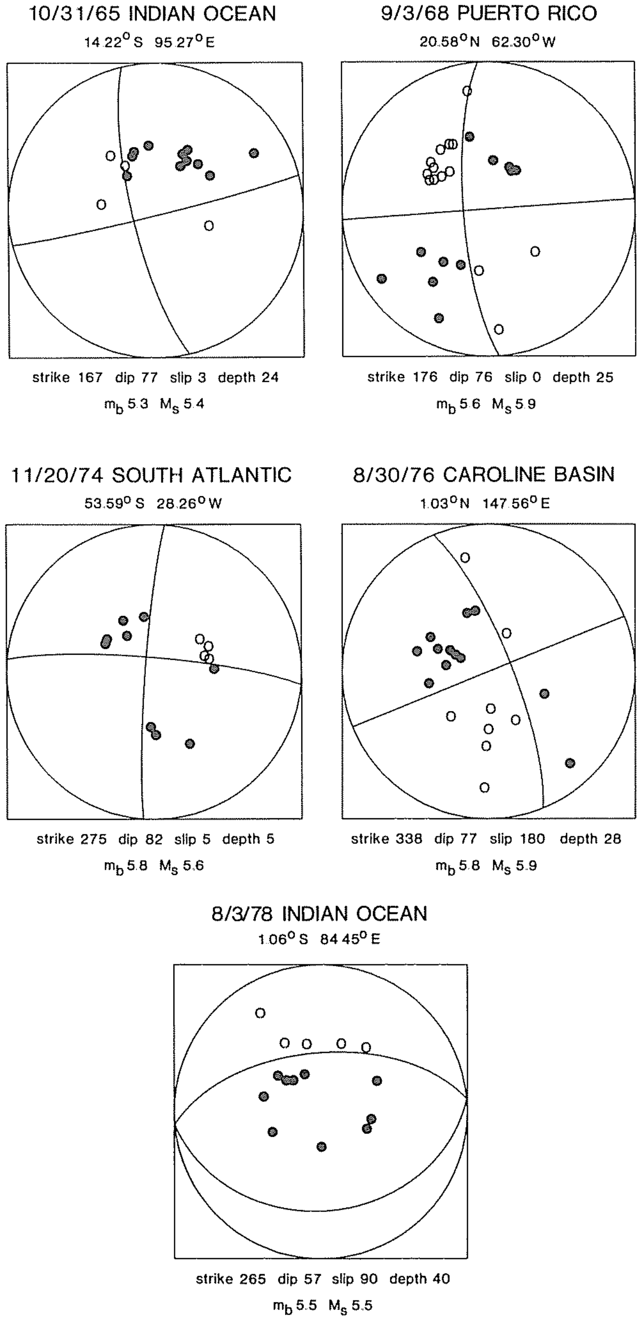

Fig. 3. Focal mechanisms of oceanic intraplate earthquakes determined in this study from first motions and body wave modeling.

four events had no modelable long-period body waves; their depths are thus more uncertain than those derived by modeling.

The 15-km depths we obtained for the May 2, 1973, Kerguelen and February 5, 1977, Antarctic (Bellingshausen Sea) events are significantly shallower than those previously found by fitting surface wave spectra. Since this procedure has often proved successful [Tsai and Aki, 1970; Trehu et al., 1981], we suspect that the difficulties in these cases result from biases caused by heterogeneities in phase velocity [Romanowicz, 1981] to the remote locations; biases due to uncertainties in near-source structure are also possible but may be less significant [Patton and Aki, 1979]. Because path phase velocity information is generally not available for oceanic intraplate events and because the uncertainty of previous depth determinations is difficult to assess, we have not included surface wave depth determinations in our data set.

In selecting the events used, we attempted to ensure that all were both intraplate and within oceanic lithosphere. Bergman and Solomon's [1980] criteria were used to separate intraplate from plate boundary events. We also excluded continental margin earthquakes such as those in Baffin Bay [Stein et al., 1979; Reid and Falconer, 1982; Stein et al., 1982] and the Beaufort Sea [Hasegawa et al., 1979; Chen and Molnar, 1983] because these events occur in a transitional region between continental and oceanic lithosphere.

A number of events occurred in the general vicinity of subduction zones. In most cases the events were sufficiently removed from zones of bending and deformation associated with the plate boundary that they may be considered intraplate in nature. For instance, the October 23, 1964, and December 13, 1977, events occurred 600 km from the Lesser Antilles trench [Stein et al., 1982], and the June 26, 1971, event occurred 380 km from the Java trench; it is thought that seismicity associated with bending plates is usually within 100 km of the trench axis [Chapple and Forsyth, 1979]. All three events show thrust faulting with the greatest principal stress axis approximately parallel to the strike of the trench, while the maximum or minimum stress axes for plate bending events should be perpendicular to the trench. Four events do occur rather close to possible plate boundaries; two of these events (September 3, 1968, and November 20, 1974) are strike slip events located about 150 km north of the Puerto Rico trench and the South Sandwich trench, respectively. Motion on both plate boundaries is mostly oblique; large plate bending stresses are not likely. The South Sandwich event could be caused by a diffuse shear zone extending from the oblique plate boundary. This model is unlikely for the Puerto Rico event because the direction of motion inferred from the mechanism (Figure 3) is opposite the direction of motion on the plate boundary [Jordan, 1975]. Two other events in the Caroline Basin (August 20, 1968 and August 30, 1976) may be related to the proposed underthrusting of the Pacific Plate by the Caroline Plate [Weissel and Anderson, 1978]; however, the August 30, 1976, event is strike slip (Figure 3), and the proposed boundary does not display typical plate boundary seismicity patterns. It is difficult to arrive at any formal criterion for defining whether regions such as the Caroline Basin or the Ninetyeast Ridge [Stein and Okal, 1978] are zones of intraplate deformation or very slow, diffuse plate boundaries; for the purposes of this study we have considered them to be intraplate. In any case, removal of individual events would not change the conclusions of this study.

Earthquake Depths and Lithospheric Properties

Figure 7 shows our results for oceanic intraplate earthquake depths as a function of lithos-

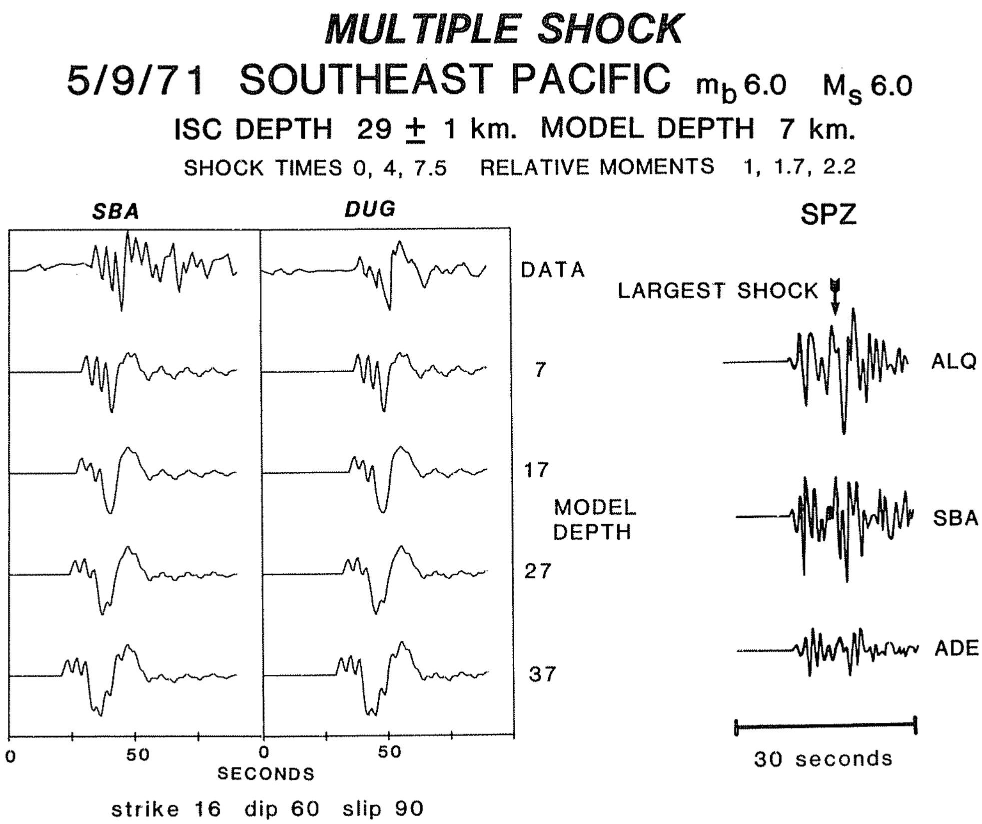

Fig. 4. Modeling of the May 9, 1971, southeast Pacific earthquake. The event is modeled as a triple shock; the first and third shocks are clearly visible on the short-period vertical records. The width of the downgoing pulse constrains the depth to about 7 km.

pheric age. The maximum depth of seismicity appears to increase with age. Also plotted on the figure are theoretical isotherms derived from a standard plate cooling model [Parsons and Sclater, 1977]. It appears that the depth of seismicity is temperature controlled and that the limiting isotherm is approximately 700°-800°C. This result is similar to that obtained by other studies; Molnar et al. [1979] found a limiting temperature of 600°C for seismicity in shallow subducting slabs, and Chen and Molnar [1983] found a limiting temperature of 600°-800°C for oceanic intraplate events.

It is interesting to compare the thickness in which earthquakes occur with other measures of lithospheric thickness. Figure 7 also shows the

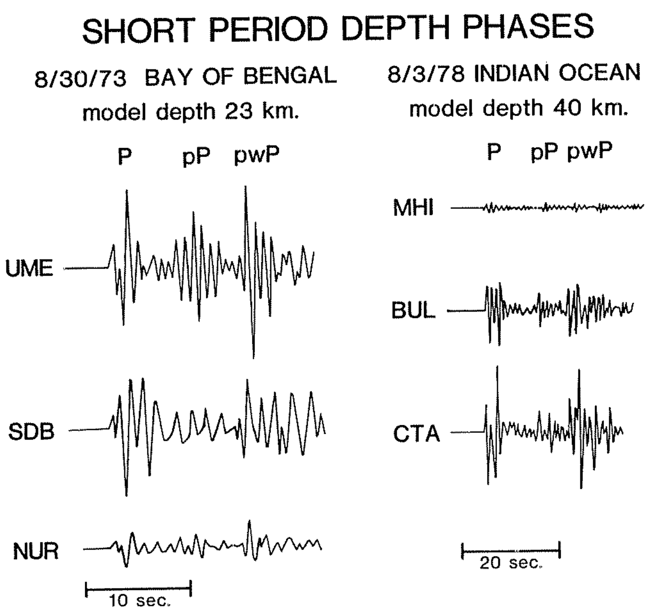

Fig. 5. Short-period depth phase identification for two Indian Ocean events. The timing of the depth phases give depths in agreement with modeling results.

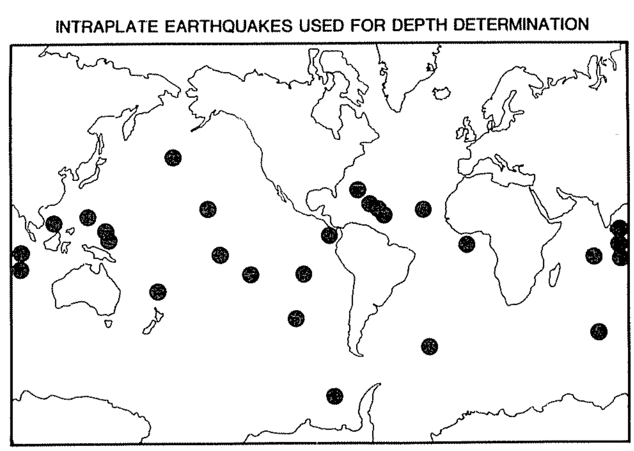

Fig. 6. Locations of oceanic intraplate earthquakes used in this study.

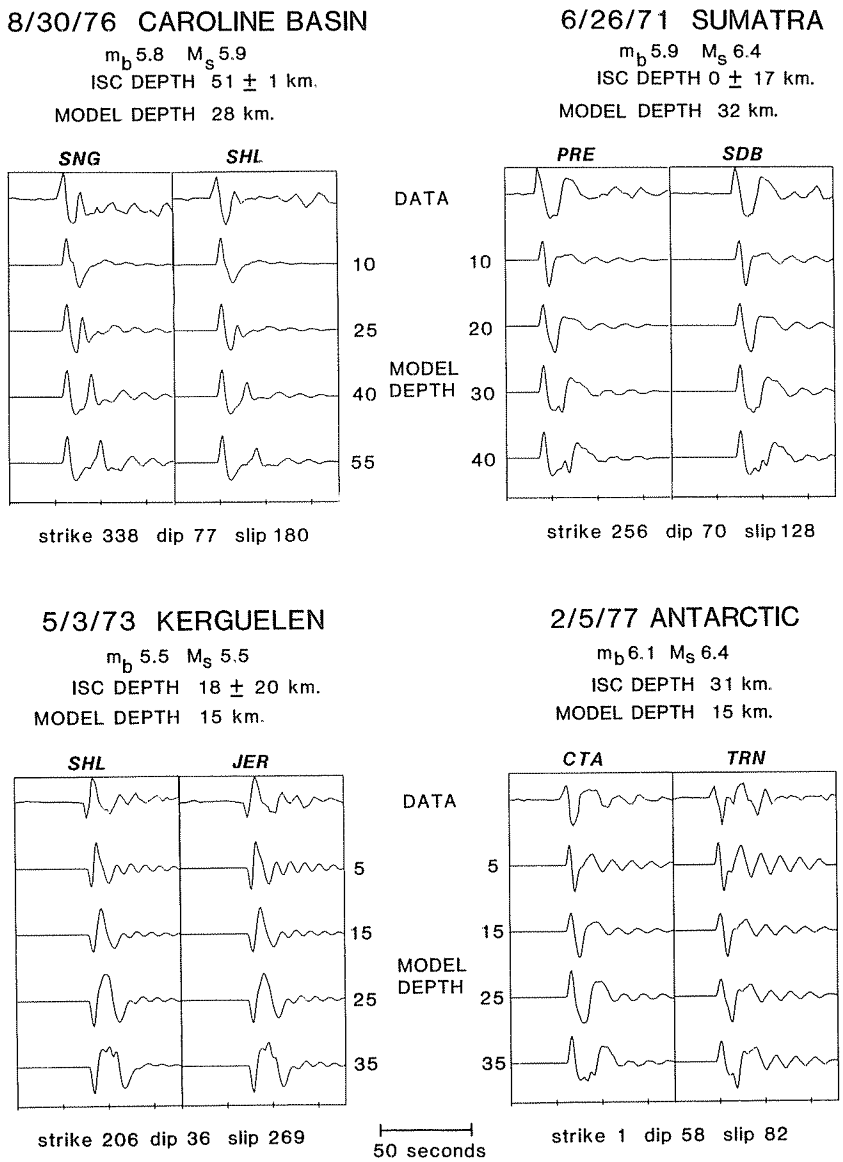

Fig. 2. Four examples of events modeled in this study showing the large discrepancies between modeled depth and ISC depth. In the actual modeling process, approximately 10 records per event were used with 5-km depth increments for the synthetics.

hypothetical event 20 km deep at a site with a 7-km-thick crust (P wave velocity 6.5 km/s) and a mantle velocity of 8.0 km/s. Very shallow events were modeled with velocities more typical of oceanic crust (6-6.5 km/s). Analysis for a range of conventional oceanic structures [Spudich and Orcutt, 1980] showed resulting errors were limited to 2-3 km in depth. This velocity uncertainty, and the precision limitation on matching waveforms, were the major sources of uncertainty in our depth determinations. The uncertainty was difficult to determine formally; we estimated it as ±5 km. For example, we propose a depth of 15 km for the May 2, 1973, Kerguelen event (Figure 2); Bergman and Solomon [1982a] model the same event at 20 km below sea level, a depth of 16.5 km.

Depth phases observed on short-period records also can help constrain the depth. Figure 5 shows depth phase identification for two events in the Indian Ocean. In both cases the depths determined from short-period records were very close to the modeled depths. In this study we did not use short-period depth phases to determine depths for events too small to model because of the likelihood of misidentified phases and possible source complexity [Forsyth, 1982].

We also include in our depth data set 11 events from other studies, listed in Table 1. The locations of all 27 events are shown in Figure 6. All but four depth determinations are from waveform modeling; when several depths are available, the one listed first was used. Two depths, the April 28, 1968, Emperor Trough event [Stein, 1979] and the November 25, 1965, Nazca plate event [Mendiguren, 1971], come from short-period depth phase identifications. Two other depths are for groups of small events in the South Pacific whose depth phases indicated foci shallower than 5 km [Okal et al., 1980]. These

TABLE 1. Oceanic Intraplate Earthquake Depths

Depth Beneath Ocean Floor (±5 km), km	Lithospheric Age, m.y.	Date	m_b	M_s	Location	Latitude °N	Longitude °E	Mechanism Reference	Depth Reference
17	55	May 25, 1964	5.7	6.0	Ninetyeast Ridge	-9.1	89.9	SY	WS
30	70	Oct. 23, 1964	6.2	6.3	Lesser Antilles	19.8	-56.1	S3,L,FI	S3,L,FI
7	10	Sept. 9, 1965	5.8	6.2	Cocos Ridge	6.5	-84.4	MS	WS
15	35	Sept. 12, 1965	6.1	6.0	Chagos Bank	-6.5	70.8	S1	S1
5	25?	Oct. 7, 1965	5.8	5.6	South China Sea	12.5	114.4	W	W
24	63	Oct. 31, 1965	5.3	5.4	Indian Ocean	-14.2	95.3	WS	WS
13	20	Nov. 25, 1965	5.3	4.6	Nazca Plate	-17.1	-100.2	M	M
10	75	April 28, 1968	5.5	5.2	Emperor Trough	44.8	174.6	S2	S2
10	33	Aug. 20, 1968	5.6	5.0	Caroline Basin	5.4	147.1	BS1	WS
27	85	Sept. 3, 1968	5.6	5.9	Puerto Rico	20.6	-62.3	WS	WS
8	10	May 9, 1971	6.0	6.0	SE Pacific	-39.8	-104.9	FO	WS
32	65	June 26, 1971	5.9	6.4	Sumatra	-5.2	96.9	SS	WS
13	50	Sept. 30, 1971	6.0	5.5	East Atlantic	-0.5	-4.9	L	L
10	30?	May 21, 1972	5.6	4.9	Fiji Basin	-27.1	175.0	BS1	WS
20	90	Oct. 20, 1972	5.7	5.8	North Atlantic	20.6	-29.7	R	WS
48	95	April 26, 1973	5.9	6.1	Hawaii	20.0	-155.2	U,B	U,B
15	25	May 3, 1973	5.5	5.5	Kerguelen	-46.1	73.2	O3	WS,BS2
23	95	Aug. 30, 1973	5.8	5.2	Bay of Bengal	7.1	84.3	BS1	WS
7	30?	April 12, 1974	5.5	4.9	Philippine Sea	14.3	134.4	BS1	WS
8	60	Nov. 20, 1974	5.8	5.6	South Atlantic	-53.6	-28.3	WS	WS
28	38?	Aug. 30, 1976	5.8	5.9	Caroline Basin	1.0	147.6	WS	WS
15	50	Feb. 5, 1977	6.1	6.4	Antarctic	-66.5	-82.4	O1	WS
25	68	Dec. 13, 1977	5.7	6.9	Lesser Antilles	17.4	-54.8	S3,BS1	S3
11	130	March 24, 1978	6.1	6.1	Bermuda	29.8	-67.4	SH	SH
40	80	Aug. 3, 1978	5.5	5.5	Indian Ocean	-16.3	92.9	WS	WS,BS3
5	80	1968-1976 (swarm)			South Pacific (Okal area A)	-71.4	-148.3	O2	O2
5	33	1965-1969			South Pacific (Okal area B)	-18.4	-132.8	O2	O2

Lithospheric ages are either from Pitman et al. [1974] with anomaly ages from Harland et al. [1983] or from Sclater et al. [1981]. Where several depth references are given, the one listed first is used. Question marks after age values denote marginal basin anomaly identifications which are somewhat more uncertain.

References:

B, Butler [1982].
BS1, Bergman and Solomon [1980].
BS2, Bergman and Solomon [1982a].
BS3, Bergman and Solomon [1982b].
FI, Fitch et al. [1980].
FO, Forsyth [1973].
K, G. Kroeger and R. Geller (unpublished manuscript, 1982).
L, Liu and Kanamori [1980].
M, Mendiguren [1981].
MS, Molnar and Sykes [1969].
O1, Okal [1980].
O2, Okal et al. [1980].
O3, Okal [1981].
R, Richardson and Solomon [1977].
S1, Stein [1978].
S2, Stein [1979].
S3, Stein et al. [1982].
SH, Steward and Helmberger [1981].
SS, Sykes and Sbar [1974].
SY, Sykes [1970].
U, Unger and Ward [1979].
W, Wang et al. [1979].
WS, this study.

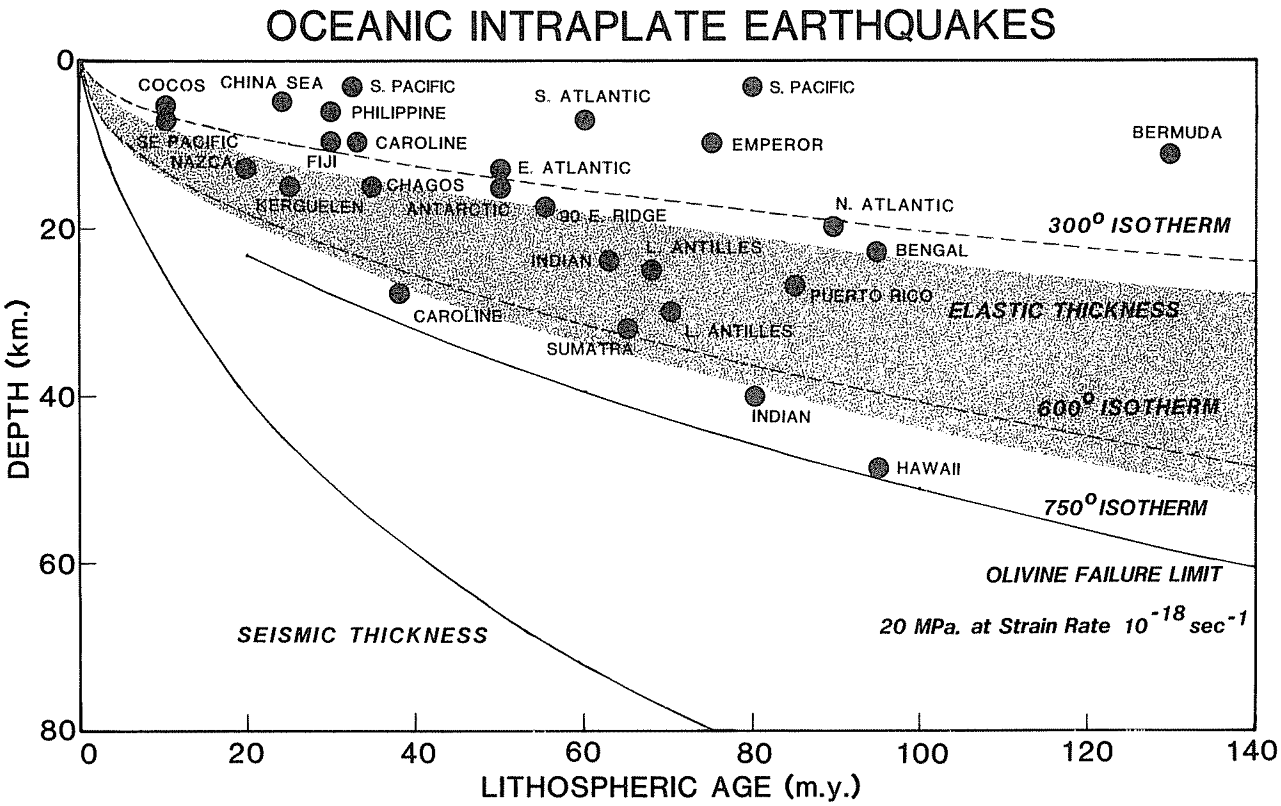

Fig. 7. Well-constrained intraplate earthquake depths shown on a depth-age plot of the ocean lithosphere. Isotherms (in degrees Celsius) shown are calculated from a lithospheric cooling model [Parsons and Sclater, 1977]. Stippled region denotes range of estimates of the flexural elastic thickness [Watts et al., 1980]. The seismic thickness is taken from Rayleigh wave dispersion data by Leeds et al. [1974]. The failure limit is the lower limit at which 20-MPa (200 bars) deviatoric stress can be sustained, calculated for a dry olivine rheology [Goetze and Evans, 1979] and a strain rate of 10^{-18} s^{-1}.

range of measurements of the elastic thickness [Watts et al., 1980] from studies of oceanic flexure and the seismic thickness [Leeds et al., 1974] from studies of Rayleigh wave dispersion. The elastic thickness is derived from the response of the lithosphere to long-term loads, whereas the seismic thickness is taken from its response to short-term loads. It is clear from the figure that the thickness of the lithosphere that is seismically active is equal to or slightly greater than the elastic thickness but is only about one half the seismic (surface wave) thickness. Thus intraplate earthquakes occur primarily in areas of the lithosphere which can sustain long-term loads as identified by studies of flexure [Watts et al., 1975, 1980; Detrick and Watts, 1979].

The data are too sparse to identify any lithospheric thinning due to reheating by hot spots [Detrick and Crough, 1978; Crough, 1978]. The parameter in question is the maximum depth of seismicity, so that the existence of shallow events does not necessarily indicate thin lithosphere, though in this argument, deep seismicity shows thick lithosphere. The April 26, 1973, Hawaii event which occurred at 48-km depth does not necessarily conflict with lithospheric reheating. Current models of thermal rejuvenation due to the Hawaiian hot spot are constructed to fit the observations that the flexural rigidity of the lithosphere is unaffected beneath Hawaii and is reduced beneath the older islands. Rejuvenation is modeled as the instantaneous heating of the lithosphere below about 45 km. Instantaneous heating is an acceptable modeling simplification but does not attempt to describe the details of the heating event, which presumably requires a finite time interval [Menard and McNutt, 1982; Von Herzen et al., 1982]. The lithosphere at the focus of the April 26, 1973, event, like the flexural rigidity at Hawaii, may still be unaffected by the reheating.

The observation that the maximum depth of oceanic intraplate seismicity seems bounded by the 700°-800°C isotherm suggests that above this temperature the lithosphere cannot support the stresses needed for seismic failure. It is generally thought that the controlling effect is the variation of strength (the maximum difference between principal stresses that rock can support) with depth in the lithosphere. Rock mechanics experiments, reviewed by Goetze and Evans [1979], Brace and Kohlstedt [1980], and Kirby [1980], show that strength depends on temperature, pressure, pore pressure, and strain rate.

At shallow depths, where brittle fracture occurs, strength increases linearly according to Byerlee's [1968] law for frictional sliding on faulted rock. Strength appears not to depend on rock type but does depend on pore pressure

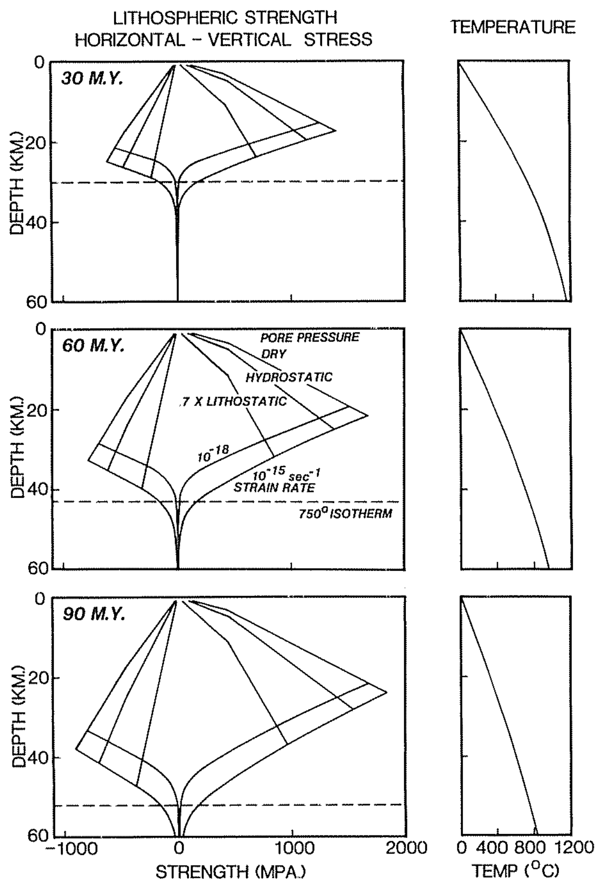

Fig. 8. Strength envelopes showing maximum stress difference as a function of depth calculated using temperatures in degrees Celsius from the Parsons and Sclater [1977] plate cooling model. At shallow depth, strength is controlled by Byerlee's Law [1968]; at greater depth, dry olivine rheologies [Goetze and Evans, 1979] predict rapid weakening. The 750° isotherm (dashed line) is the approximate lower bound for intraplate seismicity.

because the relevant variables are effective stresses (stress minus pore pressure). At greater depths, strength decreases with depth according to ductile flow laws, which vary with mineral type.

Specifically, in Brace and Kohlstedt's [1980] formulation, the maximum ($\bar{\sigma}_1$) and minimum ($\bar{\sigma}_3$) effective stresses are related by

$$\bar{\sigma}_1 \simeq 5\bar{\sigma}_3 \qquad \bar{\sigma}_3 < 110 \text{ MPa}$$

$$\bar{\sigma}_1 \simeq 3.1\bar{\sigma}_3 + 210 \qquad \bar{\sigma}_3 > 110 \text{ MPa}$$

in the brittle failure region. In the ductile flow region, for a dry olivine rheology [Goetze, 1978; Goetze and Evans, 1979], the maximum stress difference supportable depends on strain rate, $\dot{\varepsilon}$ in s^{-1} and temperature T in degrees Kelvin:

$$\dot{\varepsilon} = 7\times10^{4}(\sigma_1-\sigma_3)^3 \exp\left(\frac{-0.52 \text{ MJ/mol}}{RT}\right)$$

$$\sigma_1-\sigma_3 < 200 \text{ MPa}$$

$$\dot{\varepsilon} = 5.7\times10^{11} \exp\left(\frac{-0.54 \text{ MJ/mol}}{RT}\left(1 - \frac{\sigma_1-\sigma_3}{8500}\right)^2\right)$$

$$\sigma_1-\sigma_3 > 200 \text{ MPa}$$

Using this formulation and the temperature model of Parsons and Sclater [1977] , we calculated strength envelopes for the oceanic lithosphere at ages of 30, 60, and 90 m.y. (Figure 8). These envelopes show that the strong portion of the lithosphere deepens as the plate cools, similar to results shown by Kirby [1980]. Crucial parameters for this calculation are the assumed strain rate and pore pressure. The pore pressure controls the slope of the strength line and thus the depth at which the brittle-ductile transition occurs. The figure shows three different pore pressures: zero (dry), hydrostatic, and 70% lithostatic. At higher pore pressure, material is weaker, and the brittle-ductile transition occurs at greater depth.

The strain rates appropriate for intraplate deformation of the oceanic lithosphere are unknown. Higher strain rates correspond to stronger material and a deeper brittle-ductile transition. The figure shows strength curves for two possible cases, 10^{-18} and 10^{-15} s^{-1}. To derive these strain rates, we used the seismic moment release rate of oceanic intraplate earthquakes, discussed in a later section, to derive order of magnitude estimates for the displacement per cubic kilometers per second and thus strain rate. The lower value was found using an average moment release of 5×10^8 N m/km^3 (5×10^{15} dyne cm/km^3) per year. Division by a rigidity of 3×10^{10} Pa. (3×10^{11} dyne/cm^2) yielded a strain rate of 5×10^{-19} s^{-1}. The upper value was found from the same technique applied to seismic moment data for a very active zone of intraplate deformation, the Ninetyeast Ridge area [Stein and Okal, 1978]. Assuming that the deformation zone is 1100 km long, 400 km wide, and 40 km deep, the strain rate is 10^{-17} s^{-1} for the last 16 years or 10^{-15}

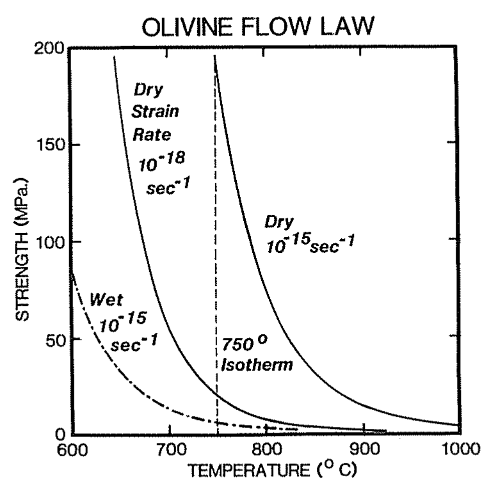

Fig. 9. Strength as a function of temperature for different rheologies and strain rates. The 750° isotherm is the approximate lower bound for intraplate seismicity. This limiting temperature corresponds to a range of strengths depending on strain rate. A wet rheology [Post, 1977] requires higher strain rates for the same strength.

s^{-1} for the last 70 years. The latter figures show one difficulty with this approach: the sampling time affects the results significantly since a few very large earthquakes provide most of the moment. Also, the calculated strain rates include only seismic strain; the amount of additional aseismic strain is unknown. Despite these uncertainties we feel that the intraplate strain rates are reasonable: the upper value, 10^{-15} s^{-1} corresponds to the strain rate used by Goetze and Evans [1979] for deformation in a bending plate at a subduction zone, and the lower value, 10^{-18} s^{-1}, is similar to Morgan's [1981] theoretical estimate of 10^{-17} s^{-1} for the average intraplate strain rate. The dominant effect in the rheology is the exponential temperature dependence; only an order of magnitude estimate is necessary for the strain rate.

The relationship between rock mechanics results derived (and extrapolated) from the laboratory and the actual behavior of materials in the earth is a complex and controversial one. In particular, the strength of rock and thus maximum sustainable stresses implied by the laboratory results are much higher than generally thought by seismologists to exist in the lithosphere, but there is no clear cause for this discrepancy [Kirby, 1980; Raleigh and Evernden, 1981]. Our results provide no new insight on these questions. We observe a temperature dependence of the maximum depth of seismicity, which can be related to the strength of rock by making certain assumptions. Figure 9 shows that the 750°C isotherm can be treated as a limiting strength by assuming a rheology (in this case, dry olivine as shown before) and a strain rate. For the lower strain rate, 10^{-18} s^{-1}, the limiting strength is 20 MPa; for the higher strain rate the limiting strength is 190 MPa. The advantage of the flow laws is that they can be calculated for a given thermal and mechanical situation and can be fit to observations by making certain assumptions. Whether the actual rock strengths predicted are correct is an important question but not essential for our purposes.

Assuming that the rheology is correct, the maximum depth of seismicity, shown by the 750°C isotherm, occurs in the ductile region unless extraordinarily high pore pressures or strain rates are assumed. For example, with a pore pressure of 70% lithostatic, a strain rate of approximately 10^{-10} s^{-1} would be required to place all seismicity in the brittle region. (Similarly, deep-focus earthquakes are difficult to explain given a conventional rheology without near lithostatic pore pressure.) Weaker rheologies, such as a wet olivine rheology [Post, 1977], require much higher strain rates to provide comparable strength, placing the seismicity further into the ductile region. For example, using the Post wet rheology, a strain rate of 10^{-13} s^{-1} is required to give approximately the same strength as the dry rheology shown before at a 10^{-18} s^{-1} strain rate. Goetze and Evans [1979] consider such a rheology inadequate to explain the flexural data at subduction zones, but McNutt and Menard [1982] argue that a weakened rheology is favored by the flexure data.

An interesting aspect of the distribution of earthquake depths is that the three events which occur at temperatures greater than 600°C (Hawaii, Indian Ocean, and Caroline Basin) are located in areas undergoing considerable deformation and where high strain rates might be expected. Thus unusually deep earthquakes such as those in Hawaii may be indicative of unusually high strain rates, whereas most intraplate areas may have lower strain rates and thus shallower seismicity. Uncertainties of this kind prevent us from identifying one isotherm at which seismicity ceases.

Our results imply that earthquakes can occur deeper and at higher temperatures in oceanic lithosphere than in continental crust. Sibson [1982] and Meissner and Strehlau [1982] correlate the depths of earthquakes beneath continental regions with strength-depth curves for a quartz rheology and note that earthquakes in continental areas rarely occur deeper than 15 km except in regions of low heat flow or active compressional tectonics. Meissner and Strehlau also propose that the limiting temperature for seismicity in the continental crust is 370°-500°C (J. Strehlau and R. Meissner, unpublished manuscript, 1982). The greater depths and temperatures at which oceanic earthquakes occur presumably result from the relative strengths of oceanic and continental lithosphere. Brace and Kohlstedt [1980] discuss the relative strength of olivine and quartz rheologies and propose that the high strength region is much shallower in continental lithosphere, assuming that the quartz rheology is appropriate. Chen and Molnar [1983] using earthquake depth data reach a similar conclusion and further suggest that the continental mantle may also be stronger than lower continental crust.

An interesting derivative question is whether earthquake mechanisms or intraplate stresses vary significantly with lithospheric depth in oceanic intraplate regions. Such changes are known in the oceanic lithosphere near trenches [Chen and Forsyth, 1978; Forsyth, 1982]. Additionally, J. Strehlau and R. Meissner (unpublished manuscript, 1982) suggest that zones of normal and strike slip faulting extend to greater depths in the continental crust than thrust faulting. Figure 10 shows a histogram of the depths of earthquakes in this study and their mechanisms; this limited

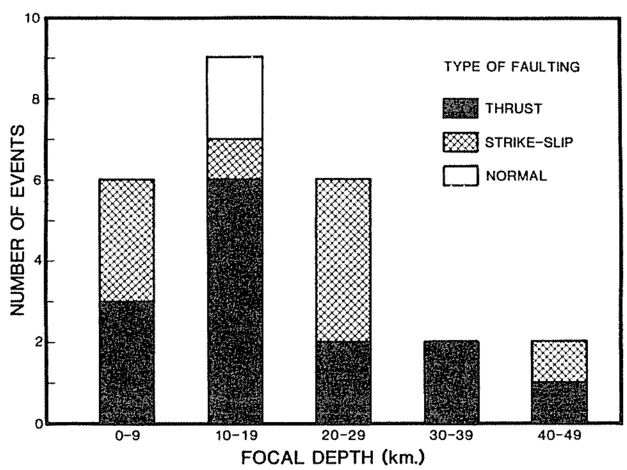

Fig. 10. Histogram showing the depth and mechanism type for earthquakes in the table. This limited data set shows no change in type of faulting with depth.

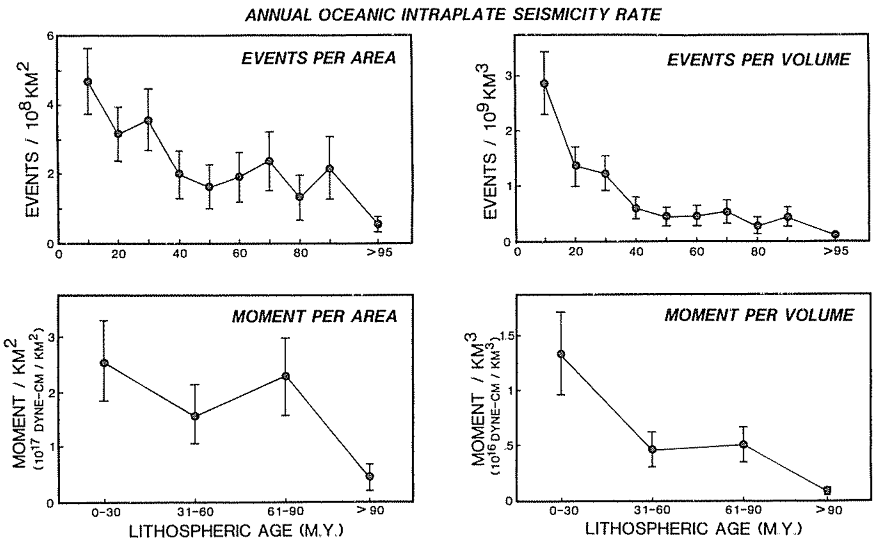

Fig. 11. (top) Number of oceanic intraplate earthquakes with $m_b \geq 5.0$ per surface area (left) and volume of strong lithosphere (right) as a function of lithospheric age. The level of seismicity shows a significant decrease with age; the cumulative moment (bottom) also seems to show a decrease with age. The area/age relations are from Parsons [1982], the volume of lithosphere is defined as the volume above the 750° isotherm, and the earthquake data are from the years 1964-1979.

data set shows no significant depth effect. Ideally, to detect such an effect, a large number of events from similar age provinces should be used. The data set used in Figure 10 consists of a small number of events from different age provinces; a better data set would be needed to determine conclusively whether the effect exists.

Age Dependence of Seismicity

Many physical properties of the lithosphere including surface wave velocity and attenuation, flexural thickness, heat flow, depth of earthquakes, and frequency of intraplate volcanism seem to be age dependent. Does the rate of seismicity show a similar dependence? Figure 11 (top left) shows the number of oceanic intraplate events with $m_b \geq 5$ per unit area of a given age. The area-age relation used was from Parsons [1982]. The Bergman and Solomon [1980] compilation was used for lithosphere older than 30 m.y. and augmented for younger regions, since normal faulting events in young lithosphere were not included in their compilation. The earthquake occurrence rate seems relatively constant between 40 and 90 m.y. but is much higher for young lithosphere (5-35 m.y.) and much lower for old lithosphere (>95 m.y.). Error bars, calculated treating earthquake occurrence as a Poisson distribution (like radioactive decay) [Bevington, 1969] suggest that this phenomenon is statistically significant. Since earthquakes occur to a depth which varies with age, the lithospheric volume susceptible to faulting may be more significant than the area. In Figure 11 (top right) the number of events in a particular age range is divided by the volume above the 750°C isotherm. The intraplate earthquake rate per unit volume shows a continuous decrease as the lithosphere ages. The high rate of seismicity in young lithosphere (5-15 m.y.) is probably not a result of mislocated ridge events; events closer than 100 km to ridges and events with few stations reporting were excluded.

Rather than an actual decrease in the seismic energy release, the smaller number of events in old lithosphere might result from fewer and larger events. The latter would imply that cumulative moment would not decrease with age even though the number of events does. To calculate the cumulative moment, a scaling relation was needed to estimate the moment from M_s for intraplate events without measured moments. Figure 12 shows measured moments and surface wave magnitudes for 27 intraplate events between magnitudes 5 and 7. The events are those used by Nuttli [1983] plus the others listed in Table 1 with measured moments (two events for the Antilles and Chagos and one for the Bermuda, Bellingshausen Sea, and South China Sea). In this magnitude range, scaling laws [Geller, 1976; Nuttli, 1983] suitable for a much larger magnitude range predict moments generally lower than those measured. We used a least squares fit to the data

$$\log M_o = 18.8 + 1.07 M_s$$

for events whose moments have not been measured. Figure 11 (bottom) shows cumulative moment as a function of the age of the lithosphere. The uncertainties of the cumulative moments are difficult to assess. A serious question is whether our sample size is large enough to be meaningful; it is possible that the time window is too small compared to the repeat times of large intraplate events. The error bars for the number of events attempt to deal with this effect, but the uncertainty in the number of events cannot be converted to the corresponding uncertainty in moment release without an assumption of the moment distribution of the earthquakes. The moment error bars shown are measurement errors estimated by

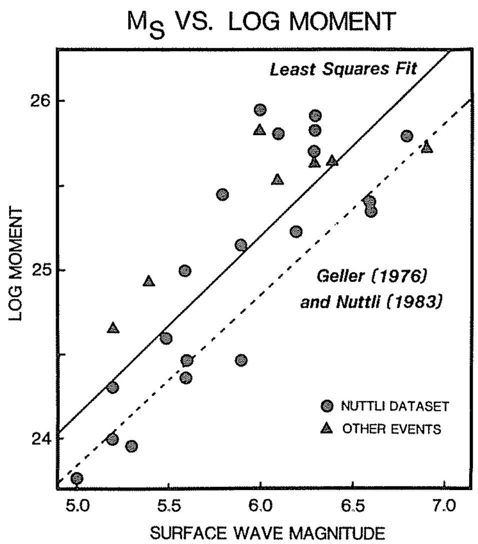

Fig. 12. Plot of log moment against M_s for intraplate earthquakes. Published scaling laws (dashed line) derived over a larger range of magnitudes [Geller, 1976, Nuttli, 1983] predict moments lower than those generally measured; a least squares fit (solid line) was used in this study. Dots are events in the Nuttli [1983] data set; triangles are events from Table 1 with measured moments.

comparing measured moments with those calculated from M_s. The error bars suggest that the decrease in moment release is real for the observed population of events, but this conclusion is significantly weaker than for the number of events. For example, the large pre-WWSSN earthquakes in the Ninetyeast Ridge area would distort the moment release data but not the number of events plot. Pre-WWSSN data were not included in this study due to uncertainties in locations and nonuniformity of coverage.

The cause of the decrease in seismicity is unclear. Given that a large fraction of intraplate earthquakes result from the reactivation of preexisting weak zones [Stein, 1979; Bergman and Solomon, 1980], the seismicity may reflect a fortuitous distribution of such features. Alternatively, the strength of such zones may increase with age, thus reducing the level of seismicity. This is hard to reconcile with the decrease in cumulative moment with age; fault strengthening should result in larger earthquakes in older lithosphere.

An alternative is that the variation in seismicity reflects differences in stress level. Several different mechanisms give rise to stresses in the oceanic lithosphere. The "ridge-push" stresses resulting from topography increase approximately as the square root of age because the force increases linearly while the lithosphere thickens as the square root of age [Lister, 1975; Hager, 1978; Parsons and Richter, 1980; Dahlen, 1981]. This stress alone, then, predicts the opposite of what we observe. There are also mechanisms which concentrate stress in younger lithosphere. One is the thermoelastic stress due to the cooling and contraction of the lithosphere, whose amplitude decays with age. Thermoelastic stresses may be especially significant in causing seismicity along age offsets such as fracture zones or old ridge jumps [Okal and Bergeal, 1983] which are already zones of weakness in the lithosphere. A second possibility is that localized concentrations of stress near the ridge due to irregularities in the spreading process give rise to increased seismicity. The true stress field in the lithosphere results from all of these effects as well as localized topographic and structural features. Detailed investigation would be required to establish the relative contribution of these effects. If the decay in seismicity results from stress level variation, it can provide a new constraint on the nature of oceanic intraplate stresses.

Conclusions

1. Oceanic intraplate earthquake depths show a clear deepening of the maximum focal depth with increasing lithospheric age. The maximum depth of seismicity is approximately bounded by a 700°–800° isotherm, in agreement with a dry olivine rheology determined by experimental rock mechanics, which predicts rapid weakening at these temperatures.
2. The thickness of the lithosphere in which oceanic intraplate earthquakes occur is approximately equal to or sightly greater than the flexural elastic thickness but is much less than the seismic thickness determined from surface wave dispersion.
3. Oceanic intraplate earthquakes occur at greater depths and temperatures than has been previously observed for continental crustal events, presumably because continental rocks weaken at lower temperatures.
4. The number of intraplate events per volume and the cumulative moment per volume decrease with age. The decrease may be an artifact of the location of weak zones with respect to age provinces or represent an increase in the strength of weak zones with age. Alternatively, this may indicate a decrease in the level of stress in the lithosphere due to thermoelastic stresses or stresses related to the spreading process.

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