38.

Ann. Rev. Earth Planet. Sci. 1983. 11: 195–214

OCEANIC INTRAPLATE SEISMICITY

Emile A. Okal
Department of Geology and Geophysics, Yale University,
Box 6666, New Haven, Connecticut 06511

Introduction

The occurrence of seismicity at the surface of the globe largely along preferential lines, now recognized as mid-oceanic ridges and subduction zones, was key evidence for the development of the theory of plate tectonics in the 1960s. Indeed, many plate boundaries, especially some of the southern, less accessible ridges, were drawn, at least initially, on the basis of this evidence alone. The mere concept of intraplate seismicity may then appear as somewhat of a paradox, if not as a failure of the whole theory. However, and as shown later in this paper, the relatively low level of this seismicity warrants its consideration as a perturbation in the general framework of nearly rigid plates. Additionally, intraplate earthquakes are of great value since, in the oceanic environment, they are our only clue to the state of stress of the lithosphere, thereby providing us with critical insight into the forces responsible for its motion. Understanding parameters that control these earthquakes is made easier (with respect to a continental situation) by the generally younger age of the ocean floor and the much simplified tectonic history of any given oceanic province. On the other hand, seismic detection capabilities are greatly reduced at sea, because of both logistics and generally higher seismic attenuation; thus, our knowledge of many aspects of oceanic intraplate seismicity is still rudimentary. It is in this general framework that the present paper reviews the following points:

1. Definition of intraplate seismicity (What?)
2. The level of seismicity in the oceans (How much?)
3. Location, including depth (Where and when?)
4. Mechanism of intraplate seismicity. Inferences about stresses (Why?)
5. Preferential siting of the stress release on the plate (How?)

0084–6597/83/0515–0195$02.00

Definition of Intraplate Earthquakes (What?)

In characterizing intraplate earthquakes, we must recognize at once several types of such events. The broadest definition of an intraplate earthquake would be an event not involving displacement between two plates. As such, all stress-controlled events, even those located in the immediate vicinity of a plate boundary, would qualify as intraplate. So would all decoupling events (of the type of the 1977 Indonesian or 1933 Sanriku earthquakes), involving rupture of the lithosphere seaward of the trench, in a zone where it undergoes bending just prior to subduction. However, following Sykes & Sbar (1974), we call this type of event "boundary-related," and do not include it in our definition. This category would also include smaller events controlled by the flexure of the plate seaward of a trench, such as those described by Chen & Forsyth (1978).

Additionally, some earthquakes occurring inside oceanic plates, such as the 1975 Kalapana event on Hawaii, are clearly associated with volcanism. Such events are not representative of the usual processes associated with the evolution and cooling of the lithospheric plate, and they are excluded from the definition of intraplate earthquakes used in this paper. As discussed later, this exclusion is easily made for well-documented seismicity in the vicinity of well-known hotspots, but would be difficult to extend to low magnitudes in poorly surveyed areas.

Finally, some areas of oceanic basins not involving major plate boundaries have been found to exhibit substantial seismicity, at a level clearly incompatible with the assumption of rigidity of the lithospheric plate. Examples are the Ninetyeast Ridge area in the Indo-Australian plate, whose seismicity, compiled by Stein & Okal (1978), included several magnitude 7 earthquakes since the 1910s, and the Caroline wedge of the Pacific plate between New Guinea and the Mariana Trench (Weissel & Anderson 1978). The amount of deformation documented in these areas suggests that they involve genuine plate boundaries, leading to the concept of two independent plates (Indian and Australian) in the case of the Ninetyeast Ridge, and of a Caroline miniplate north of New Guinea. Similarly, a 1964 event located east of the Caribbean arc (Liu & Kanamori 1980), characterized by NNW-SSE compression and a relatively deep focus (23 km), may be representative of the convergence between North and South America predicted by kinematic models such as Minster & Jordan's (1978). It is clear that these situations are peculiar, and that they define a separate type of seismicity.

Thus, we adopt a very restrictive view, and define intraplate oceanic earthquakes as events not controlled in their location and mechanism by present plate boundaries or by phenomena of a clearly extraordinary nature in the morphology of the oceanic plate, such as hotspot volcanism or large-scale deformation.

Level of Intraplate Seismicity in the Oceans (How much?)

The earliest attempt at recognizing the seismicity of the so-called stable blocks, including the Pacific Ocean Basin, is found in Gutenberg & Richter's *Seismicity of the Earth* (1941). Since the recent progress in detection brought about by the World-Wide Standardized Seismic Network [WWSSN] in the 1960s, we now have documented evidence for magnitude 5 or greater earthquakes in all of the world's oceans.

The systematic study of intraplate events, both continental and oceanic, was initiated by Sykes & Sbar (1973), after focal mechanisms of interplate earthquakes provided a spectacular qualitative confirmation of plate kinematics. More recently, Bergman & Solomon (1980) compiled a catalog of 159 oceanic intraplate events covering the period 1939–79. This catalog can be used to estimate the fraction of the world's seismicity located in the interior of oceanic plates, although such an estimate will be high since these authors use a definition of intraplate events less stringent than ours. Using Kanamori's (1977) moment-magnitude relations, one finds a total seismic moment release of 1.3×10^{27} dyn-cm (or slightly less than 10^{26} dyn-cm/yr) for the portion of the Bergman & Solomon catalog corresponding to the years since 1964, when the worldwide detection capabilities were upgraded substantially. This figure is to be compared with 6×10^{30} dyn-cm (or approximately 8×10^{28} dyn-cm/yr) for earthquakes of all types (but mostly interplate) compiled from Kanamori's (1977) list for 1904–76. It is immediately apparent that oceanic intraplate seismicity is only a minor contribution to worldwide seismicity, representing small-scale deformation in otherwise rigid plates. Similarly, Okal (1981) estimated that the deformation taken up through intraplate seismicity in the northern part of the Antarctica plate was only 2% of the rate of accretion of the plate. These figures warrant treating oceanic intraplate seismicity as a small perturbation of the rigid plate framework, and actually save the plate tectonics concept.

Oceanic intraplate seismicity is a universal feature of the world's oceans, occurring in all major plates, both wholly oceanic and continent-bearing. Figure 1 is adapted from Bergman & Solomon's paper, and was obtained by removing from their catalog events clearly associated with either volcanism or large-scale intraplate deformation. The epicenters removed were mostly located in the Ninetyeast, Hawaii, and Caroline areas. Table 1 is a list of some of the most significant oceanic intraplate earthquakes. It is not intended to substitute for a complete catalog of intraplate earthquakes [for this, the reader is referred to Richardson et al (1979) or Bergman &

Solomon (1980), and reminded once again of their less stringent definition of intraplate seismicity], but merely presents data on the largest events known in each oceanic area, and on a few earthquakes of particular importance that are used in this review. These epicenters are identified in Figure 2, together with the plate boundary system. Of particular interest in the Pacific plate are the seismic clusters at the so-called Regions A and C, studied in detail by Okal et al (1980). Region A, east of the Line Islands, was the site of 86 earthquakes during 1968–76; Region C, 500 km northeast of Pitcairn, underwent a swarm of 98 events during 1976–79, and has been quiescent at the magnitude 3 level ever since. Except for event 2 in the

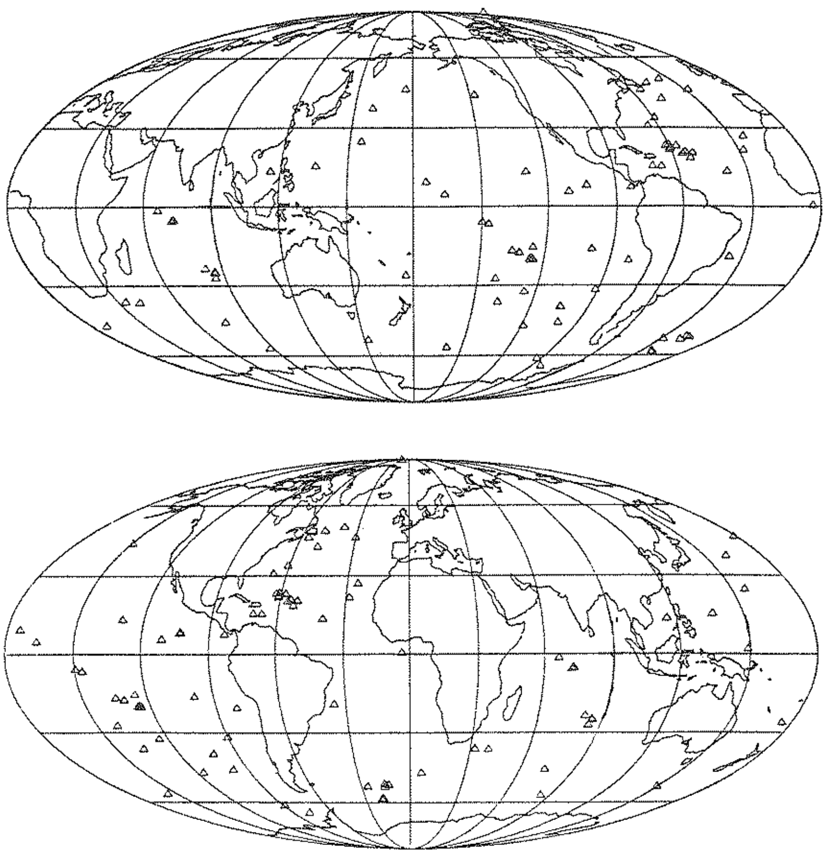

Figure 1 Oceanic intraplate seismicity of the Earth displayed on a Mollweide equal-area projection. This map is adapted from Bergman & Solomon (1980) by excluding from their catalog events associated with hotspot volcanism and large-scale internal deformation of the plates.

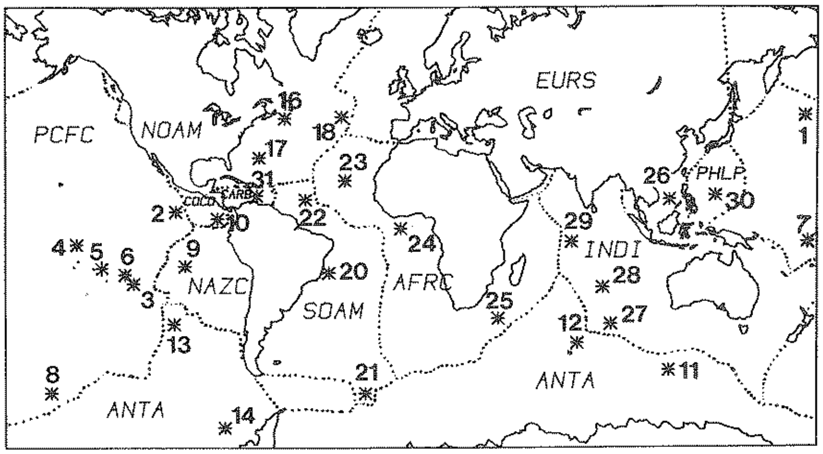

Figure 2 Mercator projection map of significant oceanic intraplate earthquakes used in the present review. Oceanic plate boundaries shown as dotted lines. Numbers refer to Table 1. Events 15 and 19, in northern part of North American plate, are not shown.

easternmost part of the plate, these two regions have contributed more than 90% of the seismic energy release in the Pacific plate during the period studied by Okal et al.

Magnitudes

The only magnitude 7 earthquakes confirmed inside oceanic plates are events 15 and 16 at the continental margins of North America, event 11 in the Indian Ocean (very little is known about this earthquake; see Okal 1981), and event 21 in the South Atlantic, a slow, complex earthquake (Creaven et al 1979), whose mechanism, not fully understood, may involve an on-going ridge jump (Farmer et al 1982). On the other hand, Table 1 shows that most plates exhibit some magnitude 6 seismicity. Of particular interest, however, is the case of the Pacific plate, by far the largest oceanic one, where the only magnitude 6 or greater seismicity is concentrated in the extreme southern tip of the plate (event 8), in the immediate vicinity of the East Pacific Rise (events 2 and 3), and at the southeastern tip of the Gilbert Island chain (event 7; Okal & Lay 1982). No earthquake with a body-wave magnitude m_b greater than 5.5 is known elsewhere in the plate.

A frequency-magnitude (b-value) investigation by Bergman & Solomon (1980) has revealed no significant departure from averages computed using both intraplate and interplate earthquakes ($b = 0.9$), except in the Pacific Ocean ($b = 1.3$). Higher b-values mean that stress release takes place through smaller, but more numerous events (a property often interpreted as weakening of the rock) that are known to be characteristic of volcanic

Table 1 Significant oceanic intraplate earthquakes

Number	Date Y M D	Epicenter °N	Epicenter °E	Magnitude	Focal type[a]	Age of plate (m.y.)	Ref.[b]	Remarks
				Pacific plate				
1	1968 04 28	44.8	174.5	$5.5m_b$	t	80–100	1	Emperor Trough
2	1970 01 21	7.0	−104.2	$6.8M_s$	t	3	2	Close to East Pacific Rise
3	1955 11 22	−24.5	−123.0	$6.5M_s$	n	10	3	Normal faulting
4	1969 08 06	−7.6	−148.3	$5.0m_b$	ss	73	4	Region A, 86 events 1965–1979
5	1965 03 06	−18.4	−132.9	$5.5m_b$	ss	34	4	Region B, 13 events 1965–1981
6	1978 07 25	−20.7	−126.8	$5.4m_b$	ss	20	4	Region C, 98 events 1976–1979
7	1982 05 23	−03.4	177.4	$6.0m_b$	t	110–120	5	Gilbert Islands
8	1947 12 15	−59.5	−159.5	$6.9M_s$	t	8	3	Largest event known in plate
				Nazca plate				
9	1965 11 25	−17.1	−100.2	$5.8m_b$	t	?	6	
				Cocos plate				
10	1976 03 29	4.0	−85.9	$6.5M_s$	ss	9	7	Slow event
				Antarctic plate				
11	1947 12 24	−54	114	$7.0M_s$		15	8	Largest event known in plate
12	1973 05 03	−46.1	73.2	$5.5m_b$	n	25	8	Off Kerguelen plateau
13	1971 05 09	−39.8	−104.9	$6.0M_s$	t	13	9	Similar event in 1925
14	1977 02 05	−66.5	−82.5	$6.4M_s$	t	56	10	Bellingshausen Sea
				North American plate				
15	1933 11 20	73.3	−70.7	$7.3M_s$	t		11	Passive margin, Baffin Bay
16	1929 11 18	44	−56	$7.2M_s$			12	Passive margin, Grand Banks
17	1978 03 24	29.9	−67.4	$6.1M_s$	t	117	13	Bermuda event
18	1964 09 17	44.5	−31.3	$5.6m_b$	t	10	2	
19	1978 01 04	85.7	−23.8	$5.0m_b$	t	10	2	Arctic Ocean

				South American plate				
20	1955 03 01	−19.9	−36.7	6.5M_s	t		14	Passive margin, Brazil
21	1977 08 26	−59.5	−20.6	7.1M_s	ss	3–4	15	Slow event
22	1968 02 20	12.4	−46.9	5.6m_b	n(?)	7	16	
				African plate				
23	1972 10 20	20.6	−29.7	5.8M_s	ss	100	17	
24	1971 09 30	−0.5	−4.8	6.0m_b	t		18	*P*-axis perpendicular to Guinea F.Z.
25	1968 09 03	−37.8	38.0	5.0m_b		80	2	
				Eurasian plate				
26	1965 10 07	12.5	114.5	5.8m_b	t	30	19	South China Sea
				Indian plate				
27	1968 10 08	−39.8	87.7	6.0m_b	n	6	16	
28	1974 06 25	−26.0	84.3	6.2M_s	t	35	20	
29	1965 09 12	−6.5	70.8	6.0M_s	n	35	21	Chagos-Laccadive Rise, swarm
				Philippine plate				
30	1974 04 12	14.3	134.4	5.5m_b	ss	37	2	
				Caribbean plate				
31	1972 08 14	14.2	−68.5	4.7m_b	ss	UK[c]	22	

[a] Focal mechanisms:
t: thrust fault
n: normal fault
ss: strike-slip

[b] References:
1. Stein (1979)
2. Bergman & Solomon (1980)
3. Okal & Greenberg, in preparation
4. Okal et al (1980)
5. Okal & Lay (1982)
6. Mendiguren (1971)
7. Okal & Stewart (1982)
8. Okal (1981)
9. Forsyth (1973)
10. Okal (1980)
11. Stein et al (1979)
12. Gutenberg & Richter (1941)
13. Stewart & Helmberger (1981)
14. Mendiguren & Richter (1978)
15. Creaven et al (1979)
16. Sykes & Sbar (1974)
17. Richardson & Solomon (1977)
18. Liu & Kanamori (1980)
19. Wang et al (1979)
20. Stein & Okal (1978)
21. Stein (1978)
22. Kafka & Weidner (1979)

[c] Age:
UK: Upper Cretaceous

seismicity (Mogi 1963). Because of the scarcity of magnitude 6 events in the Pacific plate (Bergman & Solomon did not include events 3 and 8 in their catalog), and of the extremely small range of magnitudes covered, their conclusions must be taken with caution. Okal et al (1980) conducted local b-value studies, using swarms of earthquakes at the most active locations in the southcentral Pacific, including events down to local magnitudes $M_L = 3.0$. Their results show b-values that are, if anything, lower than world averages, suggesting that Bergman & Solomon's high Pacific b-values are biased by their sampling. Okal et al then used their results ($b \leq 0.86$) to argue against a volcanic origin for the seismicity at Regions A, B (event 5 in Figure 2), and C.

Detection and Location of Oceanic Intraplate Seismicity (Where and when?)

One of the major problems encountered in assessing the seismicity of the interior of oceanic plates is the poor detection capability provided by present station coverage. At least three factors contribute significantly to this situation. First, the great majority of permanent seismic stations are still based on islands, and despite technological advances, this will probably remain a fact of life for some time: ocean-bottom seismometers (OBSs) still have recording capabilities strongly limited in time, and as such, have been used primarily in aftershock campaigns (for instance, following the 1978 Bermuda earthquake), during submersible dives usually at mid-oceanic ridges, and in the immediate vicinity of coastlines. Second, because of the swell-generated noise in the 1–5 second band, most standard oceanic stations operate at short-period gains considerably less than continental ones. Finally, higher attenuation along oceanic paths further reduces detection capabilities. A notable exception to this pattern is provided by the French Polynesian seismic network of 15 short-period stations. This network, which was described in detail by Talandier & Kuster (1976) and Okal et al (1980), uses narrow-band rejection filters to operate at gains of up to 150,000 at 1 Hz, comparable to those achieved on continents, thus allowing detection of $M_L = 4$ seismicity over an area of 1.5×10^7 km^2, and of $M_L = 3$ over 3×10^6 km^2. It has provided considerable insight into the intraplate seismicity of the southcentral Pacific. By contrast, in other oceanic areas, the absence of adequate recording facilities prevents the detection of any seismicity below the worldwide magnitude threshold of about $m_b = 4.7$.

Through the use of high-gain teleseismic records, epicentral locations for intraplate oceanic events can be achieved with reasonably good precision, usually at the level of ± 15 km for $m_b = 5.0$ earthquakes (Jordan & Sverdrup 1981). Although this precision may not be sufficient to interpret the seismicity in the context of available bathymetry, it must be considered good in view of the problems associated with hypocentral depth determination.

Because of the general paucity of arrival-time data at short distances, the trade-off between depth and origin time leads to singularity in hypocentral inversions, and very little depth information can be obtained from arrival times alone. Also, the rare arrival times at regional distances are of little if any help, since crustal structure is often poorly known in remote parts of the ocean. As a result, very little was known until recently about the depth of oceanic intraplate earthquakes. As late as 1965, event 9 in the Nazca plate was given a focal depth of 143 km by the US Geological Survey. Using surface wave spectral characteristics, Mendiguren (1971) was able to relocate this event no deeper than 13 km. A variety of techniques, including comparative surface wave excitation (Okal 1981), body-wave modeling (Wang et al 1979), identification of water-reflected phases (Okal et al 1980), and excitation of high-frequency interface waves controlled by sedimentary layering (Okal & Talandier 1981) have been used to constrain intraplate earthquake depths. All suggest that intraplate oceanic foci are located in the shallowest portion of the plate.

Using long-period body-wave modeling to constrain the depths of 18 teleseismically recorded oceanic earthquakes in the magnitude 6 range, Wiens & Stein (1983) have shown that the seismically active layer of the lithosphere thickens with age to a maximum of about 40 km. Thus, it coincides with the elastic layer, as defined by the lithosphere's response to loading over long periods of time (Watts et al 1978, McNutt & Menard 1978), and is much thinner than the 100 km or so associated with the response to faster loading, in the frequency range of seismic waves. Wiens & Stein have shown further that they can interpret this downward boundary of the occurrence of seismicity as the 600°C isotherm, corresponding to the 100 MPa olivine failure limit. Studies of smaller events, in the magnitude 5 range, have indicated even shallower foci (Harkrider & Okal 1982), located in the first 2–3 km of hard rock below sediments. It should be mentioned that some of the deeper earthquakes sampled by Wiens & Stein belong to deformed areas, such as the Caroline miniplate; this could act to bias their study slightly toward a thicker seismically active layer.

Exceptions to the pattern of very shallow oceanic intraplate seismicity are found mostly in cases involving continental margins (event 15 in Baffin Bay) or hotspot structures (a priori eliminated from this study).

Since only the uppermost part of the oceanic lithosphere is capable of rupture, and the fault width is thus strongly reduced, earthquake scaling laws (Geller 1976) are violated by oceanic intraplate earthquakes. Indeed, recent studies (Liu & Kanamori 1980) have suggested that stresses released in intraplate earthquakes are larger than their interplate counterparts; as a result, oceanic earthquakes exhibit a definite $m_b : M_s$ anomaly, their 20-s surface wave magnitude M_s being deficient by as much as 0.5 unit with respect to their 1-s body-wave magnitude m_b (Wiens & Stein 1983).

Mechanisms of Oceanic Intraplate Seismicity and Interpretation (Why?)

Sykes & Sbar (1974) were the first to gather focal mechanism data for oceanic intraplate earthquakes. Their studies were entirely based on first-motion P-wave data. In the oceanic environment, one generally faces a lack of stations at short distances; additionally, at regional distances, oceanic P_n is usually emergent and of very little use. This results in the concentration of most of the data at the center of the focal hemisphere. Under these conditions, Sykes & Sbar usually obtained only the general character of the event (normal, thrust, or strike-slip), but in many cases could not constrain the orientation of the focal planes. Their results indicated that while normal fault earthquakes are found in the youngest part of the oceans, most oceanic intraplate seismicity is of the thrust fault type once the plate has reached the age of approximately 15 m.y.

Later studies, involving the polarization of S waves (Forsyth 1973), modeling of surface waves (Mendiguren 1971) or body waves (Stewart & Helmberger 1981), or a combination of these techniques along the lines of Stein & Okal (1978), have not only confirmed the predominance of thrust fault solutions, but have allowed for the determination of the full earthquake mechanisms and of their principal stress axes. Because of greater station distances, it should be noted that the absolute threshold for reliable focal mechanisms remains around $m_b = 4.8$ for solutions relying purely on body waves; Stein (1979) has extended surface wave radiation pattern methods down to $M_s = 5.2$, below which results can be expected to be strongly affected by local (and often unknown) crustal structure. In the well documented Caribbean area, Kafka & Weidner (1979) have used event pair inversion techniques for events down to $M_s = 4.7$. Significant focal mechanisms are indicated (by type of faulting) in Table 1 and sketched in Figure 3.

A number of studies, in particular Okal et al (1980), have shown that while focal mechanisms can vary at a given seismic locality, they usually share a common direction of principal compressional stress over large areas, indicating that these earthquakes involve the release of horizontal tectonic compressional stress accumulated in the plate as a result of large-scale plate dynamics. A correlation between the azimuth of the axis of compressional stress and the direction of motion of the plate away from the ridge has been reported in the following areas: Antarctic plate panhandle

(Forsyth 1973), Bellingshausen Sea (Antarctic plate; Okal 1980), south-central Pacific (Okal et al 1980), Nazca plate (Mendiguren 1971), Canary Basin (African plate; Richardson & Solomon 1977), Brazil Basin (South American plate; Mendiguren & Richter 1978). Other events, including event 2, close to the East Pacific Rise, and event 19, in the Arctic Ocean (Bergman & Solomon 1980), have thrust fault mechanisms which, although not fully constrained, are compatible with this pattern. Many small events have thrust fault mechanisms, whose focal planes cannot be constrained (Okal et al 1980).

On continents, it has been possible to use independent in situ stress measurements to confirm that stresses derived from earthquake focal solutions are indeed representative of the ambient tectonic stress accumulated in the plate as the result of its motion over the asthenosphere (Sbar & Sykes 1973, McGarr & Gay 1978). Similar measurements cannot be taken in the oceanic environment; however, the consistency of the compressional stresses released by intraplate oceanic earthquakes over large distances is taken as a strong argument for postulating that they are indeed representative of the tectonic state of stress of the lithosphere. This is also supported by the generally good agreement found in continent-bearing plates between

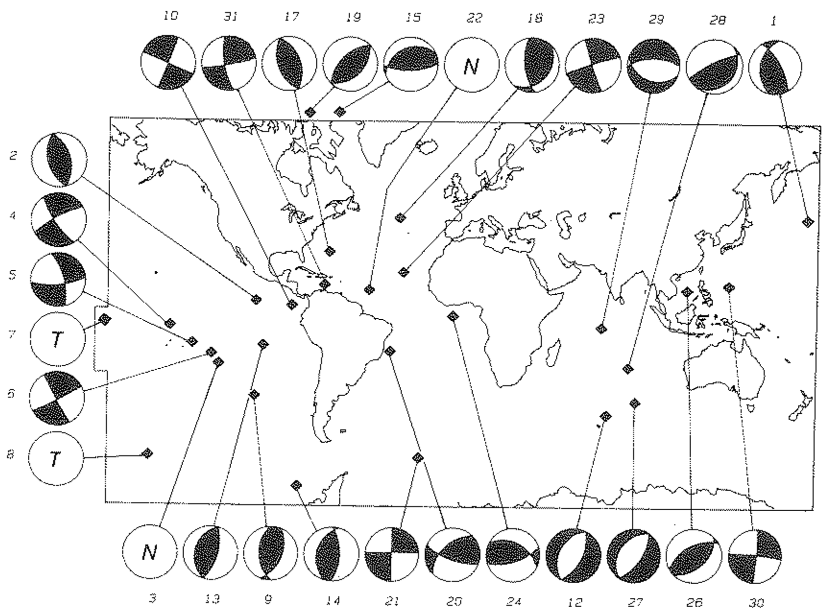

Figure 3 Focal mechanisms of oceanic intraplate earthquakes, shown by stereographic projections of lower-hemisphere focal quadrants (compressional areas in black; dilatational ones in white). N: unconstrained normal fault mechanism; T: unconstrained thrust fault mechanism. Numbers refer to Table 1 and Figure 2.

the direction of compressional stresses released during oceanic intraplate earthquakes, and of the tectonic stresses in their continental parts, as determined by either seismology or in situ measurements. Typical examples are North America (Sbar & Sykes 1977) and South America (Mendiguren & Richter 1978).

Stresses obtained from oceanic intraplate focal mechanisms may then be used to infer the relative importance of the forces involved in driving the plates. In a recent study, Richardson et al (1979) used finite element techniques to obtain estimates of the various forces acting on the plates. They were able to confirm earlier results by Forsyth & Uyeda (1975) suggesting that forces resistive to slab penetration nearly balance the strong gravitational pull of the slab at subduction zones; this leaves the state of stress in the plate controlled largely by ridge-push (a simple name for the complex process of gravitational sliding of the lithosphere as it ages away from the ridge; see Frank 1972) and by whatever drag forces exist on the lithosphere-asthenosphere boundary. Data from the Bellingshausen Sea earthquake in the Antarctic plate (which has no subduction zones and is practically fixed in the hotspot frame of reference) suggest that ridge-push is the dominant contributor (Okal 1980), a theory also upheld by the variation of the stress inside continental North America (Sbar & Sykes 1977).

However, as pointed out by McKenzie (1969), in the presence of fossil structures acting as preferential fault planes, there can be a significant deviation of the direction of released stress from the ambient one. Raleigh et al (1972) quantify this effect to a maximum $\pm 25°$. An example is believed to be event 1 (Stein 1979); another one could be the Bermuda earthquake (event 17), for which compressional stress is 23° away from the direction of ridge-push (Stewart & Helmberger 1981). Thus, the fact that a favorable morphology of the ocean floor may affect the direction of stress release must be kept in mind when interpreting earthquake mechanisms.

Another potential source of compressional stress in the plate could be asthenospheric drag in the direction of the plate's absolute motion over the mantle. Richardson et al (1979) have shown that this force is passive and has little effect on oceanic lithosphere. However, Wesnousky & Scholz (1980) have proposed that in the vicinity of a continent the craton's deeper roots distort the stress field in the oceanic lithosphere. Such a mechanism may be involved in a number of large, poorly known events at continental margins (Sykes 1978). In the case of motionless Antarctica, however, Okal (1980) has shown that the stress released in the Bellingshausen Sea earthquake, in the vicinity of the continental shelf, could be interpreted only by ridge-push.

Stresses in plates with little or no ridge boundaries must be of a different origin: these include the Caribbean and Philippine plates and the South China Sea portion of the Eurasian plate. In the Caribbean plate, event 31, studied by Kafka & Weidner (1979), shows release of a compressional stress oriented WNW-ESE, whose origin is probably related to subduction at the Lesser Antilles and convergence with South America along the Venezuelan coast. Similarly, Bergman & Solomon's (1980) solution for event 30 inside the Philippine plate involves compression along the NW-SE axis, representative of the convergence involved at the Mariana and Philippine trenches, after allowing for possible reorientation of the released stress by the Palau-Kyushu Ridge. Wang et al's (1979) focal solution in the South China Sea, an area where spreading stopped 17 m.y. ago, was also interpreted in terms of convergent tectonics.

Most earthquakes whose mechanism is not explained by ridge-push or convergence have been interpreted as phenomena associated with continental margins (e.g. Baffin Bay seismicity; Stein et al 1979), large intraplate asthenospheric flow (near Kerguelen Island; Okal 1981), or continued deformation along a major bathymetric feature (Chagos Bank; Stein 1978). This is probably also the case of the recent Gilbert Islands earthquake swarm: event 7 is located at the southeastern tip of the island chain, and its tentative focal solution is a thrust fault involving NNE-SSW compressional stress (Okal & Lay 1982).

However, some problems remain unsolved. They are basically of two types: tensional events involving normal faulting, and thrust mechanisms whose compressional axes do not fit known stress fields, even when adjusted 25° or less for preferential release along existing faults. Tensional events are usually found close to the ridges (e.g. the 1955 earthquake near Easter Island). This prompted Sykes & Sbar (1973) to propose that the plate remains under tensional stress for about 10 to 20 m.y., until it cools down sufficiently and goes to a compressional state of stress. However, as shown by the data in Table 1, thrust events are found in extremely young lithosphere (event 2 is only 200 km, or 3 m.y., west of the East Pacific Rise; the large 1947 earthquake, event 8, is 300 km, or 8 m.y., from the South Pacific Ridge; and event 18 is 100 km, or 10 m.y., from the Mid-Atlantic Ridge); in contrast, normal events are found as far as 10 m.y. away from ridges (event 3). Thus the cooling process invoked by Sykes & Sbar (1974) cannot be uniquely dependent on age, as opposed to many other cooling-controlled phenomena (Tréhu et al 1976). One cannot discard the alternative possibility that normal fault events reflect local factors, such as increased asthenospheric flow, or volcanism, potentially of "plume" origin, which may also be responsible for features such as gravity and bathymetry highs, especially in the vicinity of volcanic islands, such as Easter. As discussed by Okal & Bergeal (1983), event 3 is indeed located in the vicinity of seamounts. It is also unlikely that such tensional stresses are due to loading effects (in the absence of major island chains) or to membrane tectonics (Turcotte & Oxburgh 1973), since the lithosphere involved in the case of event 3, for example, has basically remained at the same latitude since it was created.

Some thrust events also exhibit compressional stress directions that are incompatible with the direction of motion of the plate away from its ridge: The 1971 event on the Guinea Fracture Zone, in the African plate, exhibits a compressional axis more or less perpendicular to the fracture zone (Liu & Kanamori 1980), whose origin must lie in locally created conditions, poorly understood at present.

Additionally, a few strike-slip events, whose mechanisms are not readily interpreted in the context of known tectonic stresses, add to the complexity of the picture: Event 10 in the Cocos plate north of the Galapagos Ridge (in an area about 9 m.y. old) has a compressional axis trending approximately N70°E (Okal & Stewart 1982), about 60° away from the direction of ridge-push; event 21, northeast of the South Sandwich Islands, in the South American plate, has a vertical strike-slip mechanism with compressional axis N135°E (Creaven et al 1979), 45° away from ridge-push. Both of these events exhibit "slow" mechanisms, characterized by an increase of magnitude with period and a complex source rupture process, which have led Okal & Stewart (1982) to propose that large-scale intraplate deformation, possibly related to hotspot or other magmatic activity, may be taking place in these areas.

In conclusion, we now understand most of the stresses responsible for oceanic intraplate seismicity. Ridge-push plays an apparently predominant role; convergent tectonics at the plate boundaries also contribute to horizontal compressional stresses inside the plates. The nature of occasionally observed tensional stresses, as well as the origins of some isolated cases of strike-slip faulting of an apparent local nature, remain unclear.

Finally, we should repeat that the above discussion is limited to earthquakes of magnitude $m_b \geq 4.8$. In continental areas, focal mechanism solutions have been found consistent across the magnitude scale (Sbar & Sykes 1977); however, an interpolation of the above results to lower magnitudes may not be warranted in the oceanic environment. In particular, volcanic seismicity, at a level escaping teleseismic detection except in the form of T waves, has been documented, notably at Macdonald Volcano (Johnson 1970). Despite some present insight into the characteristics of volcanic seismicity (e.g. Klein 1982), the identification of the origin of a seismic source as tectonic or volcanic remains a difficult problem in the case of low-level seismicity in unsurveyed areas. Such identification involves a study of the bathymetric characteristics of the epicentral area, and is best discussed in the framework of the next section.

Siting of Seismicity in Correlation with Bathymetry (How?)

Having described what is now considered a reasonably well understood picture of the stresses involved in major oceanic intraplate earthquakes, we must now address the question of the factors governing the siting of seismicity on the plate, in other words "How is the stress released?"

In a monumental review paper, Sykes (1978) examined the question of the preferential location of intraplate seismicity, and other forms of tectonism, onto otherwise "stable" blocks, and concluded that there exist zones of weakness, such as sutured fracture zones, representative of previous cycles of active tectonism, along which a number of phenomena, such as intraplate seismicity and kimberlites, occur preferentially. His study was, however, mainly concerned with continents. In the oceanic environment, it has also been suggested that zones of weakness, such as fracture zones outside of their active transform segments, may be areas of preferential intraplate seismicity. The Ninetyeast Ridge, one of the longest known suture zones, is indeed the preferred site of large-scale deformation induced in the Indo-Australian plate by the Himalayan collision (Stein & Okal 1978). Other areas of preferential weakness could include former plate boundaries and the traces of former hotspots, evidenced in present-day bathymetry as seamount chains. This has prompted Bergman & Solomon (1980) to investigate systematically the correlation between seismicity and bathymetry for a data set of 83 intraplate oceanic earthquakes. Their results suggest that the larger earthquakes are often associated with old fracture zones, but that the correlation with other large bathymetric features is poor; their most compelling results, however, are taken from the Ninetyeast area of the Indian Ocean, which is not fully representative of genuine intraplate oceanic seismicity. A good correlation of intraplate oceanic seismicity with major fracture zones has also been reported for events 11, 14, 17, 21, and 24, all of them major shocks. Smaller events were found by Bergman & Solomon to correlate less significantly with known bathymetry.

Okal (1981) has shown that a number of earthquakes in the northern panhandle of the Antarctic plate are aligned in the vicinity of the line of maximum age of the plate, which is the wake of the Easter Island triple junction, and as such a zone of suture and of potential weakness. More recently, Okal & Bergeal (1983) have shown that at least 6 foci in the southcentral Pacific (including Region C) are located on the boundary line of lithosphere generated at the old Farallon Ridge, before it jumped and reoriented itself along the present East Pacific Ridge sometime in the early Miocene (Herron 1972). This boundary is a line of age discontinuity on the present Pacific plate, and probably represents a zone of weakness. Epicenter 9 and two other seismic sites belong to a similar boundary in the Nazca plate. Stein (1979) has also discussed preferential occurrence of seismicity along fossil features such as the Emperor Trough, but indicated that the stress field released in their vicinity may be considerably distorted with respect to the tectonic stress in the plate. Under some circumstances, it may become difficult to distinguish between mechanisms believed to be distorted by the bathymetry and mechanisms involving local stress regimes, such as in the Chagos area.

However, old fracture zones and other weak lines are far from being the only repositories of oceanic intraplate seismicity. Most of the seismic localities identified by Okal et al (1980) in the southcentral Pacific lack a definite correlation with bathymetric features; one of the most active sites, Region A, east of the Line Islands, was the target of an on-ship survey in 1979, which failed to reveal any substantial bathymetric feature, let alone a major fracture zone. This region (the second most active in the southcentral Pacific) lies about 130 km south of the large Galápagos Fracture Zone, and some 60 km north of a much smaller one, and is not apparently associated with these features (Sverdrup 1981). Similarly, the large-scale bathymetry in Region C was explored by Sailor & Okal (1983) at long wavelengths, using satellite radar altimetry; they discovered only one fracture zone, located about 70 km to the south, too far away to permit an association with the epicenter. As discussed by Sverdrup (1981), the geomorphological features present in Region A are oriented parallel to the local regime of fracture zones, derived from the old Farallon Ridge, at approximately 45° from the tectonic stress created by the present East Pacific Ridge. This favorable situation may explain the relatively low magnitude level of the seismicity at Region A and the strike-slip focal solutions.

In particular, it is interesting to note that this discrepancy between existing large-scale tectonic directions and the orientation of present stress can exist only in a plate having undergone a change in its accretion pattern. In practice, this applies only to the bulk of the Pacific plate, which was generated at the old Farallon Ridge. Significantly, as mentioned earlier, the only magnitude 6 events known in the Pacific plate are located in areas that did not go through the reorientation process: the young, easternmost fringe of the plate, generated since the ridge jump (events 2 and 3); the region south of the Louisville Ridge (event 8); and the Gilbert Islands area, generated at the Pacific-Phoenix boundary (Hilde et al 1977). This could provide an explanation for the absence of magnitude 6 events in the bulk of the Pacific plate, and for the low b-values obtained by Bergman & Solomon (1980) in this plate.

Correlation with Volcanism?

The association of seismicity with small-scale bathymetric features such as seamounts is even more difficult to investigate because the mapping of seamounts in remote areas of the oceans is still very incomplete: Seismic swarms were responsible for the discovery of Macdonald Volcano in 1967 (Johnson 1970) and of the suspected volcanoes Rocard and Moua Pihaa in the Tahiti-Mehetia area (Talandier & Kuster 1976). Recent seismic swarms in the Society Islands also suggest volcanic activity on the flanks of Mehetia (Talandier 1981) and at Teahitia, only 65 km east of Tahiti (Talandier & Okal 1982). It is then possible, at least in principle, to suspect a similar phenomenon in the case of a sustained swarm of seismicity in an uncharted oceanic area, such as Regions A or C. Geophysicists can make use of two major tools to assert the volcanic vs tectonic origin of seismicity: knowledge of the geomorphology and bathymetry in the epicentral area, or analysis of the patterns of recorded seismicity, followed by comparison with well-documented volcanism, such as that of Kilauea and Loihi in Hawaii (Klein 1982). The former kind of information is nonexistent for most oceanic seismic foci, but large-scale bathymetry could be obtained at little cost from satellite data (Sailor & Okal 1983); the latter will require precise knowledge of the seismicity at low magnitudes, and in particular of the hypocentral depths.

The complexity of this issue can best be judged on the example of Region C: Sailor & Okal (1983) have rejected the possibility of a major volcanic edifice, a result also confirmed by an on-site survey (Francheteau, personal communication, 1981), which has, on the other hand, indicated topography on the order of 500 m and an unusually high level of low-magnitude, largely unexplained, seismic activity (Pascal, personal communication, 1981). It is clear that the concentration of seismic activity (which at the $m_b \geq 5.2$ level is compatible with release of tectonic, ridge-push type stresses) at Region C is presently not understood.

Conclusion: Our State of Ignorance, and Recommendations for Future Research Orientation

Despite the fundamental progress made over the past 12 years in our knowledge and understanding of oceanic intraplate seismicity, including the origin of the stresses released in the major events, at least one major problem remains unsolved: *What is the level of low-magnitude seismicity in the ocean basins, far away from plate boundaries, hotspot edifices, and other islands?* High-gain local networks have not been deployed permanently in the oceanic environment, with the exception of the French Polynesian array, whose existence has allowed the recognition of some 30 unsuspected epicenters. Consequently, we have no estimate of the seismicity below magnitude 4.7 for most parts of the ocean floor. For example, no seismicity other than event 11 is known east of Kerguelen in the Antarctic plate, all the way to Balleny Islands, and the question of whether this pattern of a "quiet" zone extends to lower magnitudes remains open. Instrumentation of more oceanic islands with high-gain permanent stations is thus a primordial step in furthering our knowledge of oceanic intraplate seismicity. Had not comparable networks existed on continents, the seismicity of such areas as New England would be practically undocumented.

OBSs have until now been deployed mostly in areas of major geophysical interest, such as mid-oceanic ridges, continental margins, fracture zones, or epicentral areas of major earthquakes, and always for limited periods of time. The relatively low level of intraplate seismicity is such that only a permanently deployed network can retrieve significant information about it. It would be extremely useful to develop an OBS system capable of permanent recording, and to run a prolonged OBS campaign in "average" areas on the floor of an oceanic basin, in order to get a clear picture of the level of the low-magnitude ($m_b = 2$–4) seismicity of the interior of the plate, independently of any disturbing tectonic influence. This potential seismicity presently escapes detection.

Assuming that we can improve our knowledge of oceanic intraplate seismicity down to this level of magnitudes, its interpretation will be possible only in the context of other aspects of marine geophysics, in particular of the small-scale bathymetry of the ocean floor. This is a second, and formidable, unknown; but we have seen that the large-scale bathymetry, available now at little cost from satellite data, does not always provide a clue as to the siting of the seismicity in the plate. Only detailed knowledge of the geomorphology and of the tectonic history of such areas as Region C will allow a better understanding of the relationships between oceanic seismicity, volcanism, and possibly other forms of active tectonism of a lesser intensity, which may be involved along weak lines, as suggested by the microseismicity recorded at Region C.

ACKNOWLEDGMENTS

Many of the ideas expressed in this paper evolved from several years of stimulating collaboration with Jacques Talandier and Tom Jordan. Discussions and preprint exchanges with Seth Stein are also gratefully acknowledged [even though our studies in Stein & Okal (1978) no longer qualify as intraplate in the strict sense!]. I am grateful to the staff of Hawaii Volcano Observatory for their hospitality and many discussions on volcanic seismicity, and to Sean Solomon for permission to adapt Figure 1 from Bergman & Solomon (1980). I thank Jean Francheteau and Georges Pascal for access to preliminary data obtained at Region C. This research was supported by the Office of Naval Research, under Contract N00014-79-C-0292, and also in part by the National Science Foundation, under Grant EAR-81-06106.

Literature Cited

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

Chen, T., Forsyth, D. W. 1978. A detailed study of two earthquakes seaward of the Tonga Trench: Implications for mechanical behavior of the lithosphere. *J. Geophys. Res.* 83:4995–5003

Creaven, C. T., Kanamori, H., Fujita, K. 1979. A large intraplate event near the Scotia arc. *Eos, Trans. Am. Geophys. Union* 60:894 (Abstr.)

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

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

Farmer, R. A., Fujita, K., Stein, S. 1982. Seismicity and tectonics of the Scotia arc area. *Eos, Trans. Am. Geophys. Union* 63:440 (Abstr.)

Frank, F. C. 1972. Plate tectonics, the analogy with glacier flow and isostasy. In *Flow and Fracture of Rocks*, ed. H. C. Heard. *Geophys. Monogr. Ser.* 16:285–92

Geller, R. J. 1976. Scaling relations for earthquake source parameters and magnitudes. *Bull. Seismol. Soc. Am.* 66:1501–23

Gutenberg, B., Richter, C. F. 1941. Seismicity of the Earth. *Geol. Soc. Am. Spec. Pap. 34*

Harkrider, D. G., Okal, E. A. 1982. Propagation and generation of high-frequency sediment-controlled Rayleigh modes following shallow earthquakes in the southcentral Pacific. *Eos, Trans. Am. Geophys. Union* 63:1025 (Abstr.)

Herron, E. M. 1972. Sea-floor spreading and the Cenozoic history of the eastcentral Pacific. *Geol. Soc. Am. Bull.* 83:1671–92

Hilde, T. W. C., Uyeda, S., Kroenke, L. 1977. Evolution of the western Pacific and its margin. *Tectonophysics* 38:145–65

Johnson, R. H. 1970. Active submarine volcanism in the Austral Islands. *Science* 167:977–79

Jordan, T. H., Sverdrup, K. A. 1981. Teleseismic location techniques and their application to earthquake clusters in the southcentral Pacific. *Bull. Seismol. Soc. Am.* 71:1105–30

Kafka, A. L., Weidner, D. J. 1979. The focal mechanisms and depths of small earthquakes as determined from Rayleigh-wave radiation patterns. *Bull. Seismol. Soc. Am.* 69:1379–90

Kanamori, H. 1977. The energy release in great earthquakes. *J. Geophys. Res.* 82:2981–87

Klein, F. W. 1982. Earthquakes at Loihi submarine volcano and the Hawaiian hot-spot. *J. Geophys. Res.* 87:7719–26

Liu, H.-L., Kanamori, H. 1980. Determination of source parameters of mid-plate earthquakes from the waveforms of body waves. *Bull. Seismol. Soc. Am.* 70:1989–2004

McGarr, A., Gay, N. C. 1978. State of stress in the Earth's crust. *Ann. Rev. Earth Planet. Sci.* 6:405–36

McKenzie, D. P. 1969. The relationship between fault plane solutions for earthquakes and the direction of the principal stresses. *Bull. Seismol. Soc. Am.* 59:591–601

McNutt, M. K., Menard, H. W. 1978. Lithospheric flexure and uplifted atolls. *J. Geophys. Res.* 83:1206–12

Mendiguren, J. A. 1971. Focal mechanism of a shock in the middle of the Nazca plate. *J. Geophys. Res.* 76:3861–79

Mendiguren, J. A., Richter, F. M. 1978. On the origin of compressional intraplate stress in South America. *Phys. Earth Planet. Inter.* 16:318–26

Minster, J. B., Jordan, T. H. 1978. Present-day plate motions. *J. Geophys. Res.* 83:5331–54

Mogi, K. 1963. Some discussions on aftershocks, foreshocks, and earthquake swarms, the fracture of a semi-infinite body caused by an inner stress origin and its relation to the earthquake phenomena, 3. *Bull. Earthquake Res. Inst. Tokyo Univ.* 41:615–58

Okal, E. A. 1980. The Bellingshausen Sea earthquake of February 5, 1977: evidence for ridge-generated compression in the

Antarctic plate. *Earth Planet. Sci. Lett.* 46:306–10
Okal, E. A. 1981. Intraplate seismicity of Antarctica and tectonic implications. *Earth Planet. Sci. Lett.* 54:397–406
Okal, E. A., Bergeal, J.-M. 1983. Mapping the Miocene Farallon ridge jump on the Pacific plate. *Earth Planet. Sci. Lett.* In press
Okal, E. A., Lay, T. 1982. The Gilbert Islands earthquake swarm, 1981–1982. *Eos, Trans. Am. Geophys. Union* 63:1138–39 (Abstr.)
Okal, E. A., Stewart, L. M. 1982. Slow earthquakes along oceanic fracture zones: evidence for asthenospheric flow away from hotspots? *Earth Planet. Sci. Lett.* 57:75–87
Okal, E. A., Talandier, J. 1981. Dispersion of one-second Rayleigh modes through oceanic sediments following shallow earthquakes in the southcentral Pacific. In *Bottom-Interacting Ocean Acoustics*, ed. W. A. Kuperman, F. B. Jensen, *NATO Ser. 4*, 5:345–58. New York: Plenum
Okal, E. A., Talandier, J., Sverdrup, K. A., Jordan, T. H. 1980. Seismicity and tectonic stress in the southcentral Pacific. *J. Geophys. Res.* 85:6479–95
Raleigh, C. B., Healy, H. J., Bredehoeft, D. J. 1972. Faulting and crustal stress at Rangely, Colorado. In *Flow and Fracture of Rocks*, ed. H. C. Heard. *Geophys. Monogr. Ser.* 16:275–84
Richardson, R. M., Solomon, S. C. 1977. Apparent stress and stress drop for intraplate earthquakes and tectonic stress in the plates. *Pure Appl. Geophys.* 115:317–31
Richardson, R. M., Solomon, S. C., Sleep, N. H. 1979. Tectonic stress in the plates. *Rev. Geophys. Space Phys.* 17:981–1008
Sailor, R. V., Okal, E. A. 1983. Applications of SEASAT altimeter data in seismotectonic studies of the southcentral Pacific. *J. Geophys. Res.* 88: In press
Sbar, M. L., Sykes, L. R. 1973. Contemporary compressive stress and seismicity in eastern North America: an example of intraplate tectonics. *Geol. Soc. Am. Bull.* 84:1861–82
Sbar, M. L., Sykes, L. R. 1977. Seismicity and lithospheric stress in New York and adjacent areas. *J. Geophys. Res.* 82:5771–86
Stein, S. 1978. An earthquake swarm on the Chagos-Laccadive Ridge and its tectonic implications. *Geophys. J. R. Astron. Soc.* 55:577–88
Stein, S. 1979. Intraplate seismicity on bathymetric features: The 1968 Emperor Trough earthquake. *J. Geophys. Res.* 84:4763–68
Stein, S., Okal, E. A. 1978. Seismicity and tectonics of the Ninetyeast Ridge area: evidence for internal deformation of the Indian plate. *J. Geophys. Res.* 83:2233–45
Stein, S., Sleep, N. H., Geller, R. J., Wang, S.-C., Kroeger, G. C. 1979. Earthquakes along the passive margin of eastern Canada. *Geophys. Res. Lett.* 6:537–40
Stewart, G. S., Helmberger, D. V. 1981. The Bermuda earthquake of March 24, 1978: A significant oceanic intraplate event. *J. Geophys. Res.* 86:7027–36
Sverdrup, K. A. 1981. *Seismotectonic studies in the Pacific Ocean Basin.* PhD thesis. Univ. Calif., San Diego. 436 pp.
Sykes, L. R. 1978. Intraplate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism and other tectonism postdating continental fragmentation. *Rev. Geophys. Space Phys.* 16:621–88
Sykes, L. R., Sbar, M. L. 1973. Intraplate earthquakes, lithospheric stresses and the driving mechanism of plate tectonics. *Nature* 245:298–302
Sykes, L. R., Sbar, M. L. 1974. Focal mechanism solutions of intraplate earthquakes and stresses in the lithosphere. In *Geodynamics of Iceland and the North Atlantic Area*, ed. Kristjansson, pp. 207–24. Dordrecht: Reidel
Talandier, J. 1981. 1981 Seismic crisis under Mehetia Island, French Polynesia. *Eos, Trans. Am. Geophys. Union* 62:949 (Abstr.)
Talandier, J., Kuster, G. T. 1976. Seismicity and submarine volcanic activity in French Polynesia. *J. Geophys. Res.* 81:936–48
Talandier, J., Okal, E. A. 1982. 1982 Seismic crisis at Teahitia Seamount, French Polynesia. *Eos, Trans. Am. Geophys. Union* 63:1092 (Abstr.)
Tréhu, A. M., Sclater, J. G., Nábělek, J. 1976. The depth and thickness of the ocean crust and its dependence upon age. *Bull. Soc. Géol. Fr. Sér. 7* 18:917–30
Turcotte, D. L., Oxburgh, E. R. 1973. Mid-plate tectonics. *Nature* 244:337–39
Wang, S.-C., Geller, R. J., Stein, S., Taylor, B. 1979. An intraplate thrust earthquake in the South China Sea. *J. Geophys. Res.* 84:5627–32
Watts, A. B., Bodine, J. H., Steckler, M. S. 1978. Observations of flexure and the state of stress in the oceanic lithosphere. *J. Geophys. Res.* 85:6369–76
Weissel, J. K., Anderson, R. N. 1978. Is there a Caroline plate? *Earth Planet. Sci. Lett.* 41:143–58
Wesnousky, S. G., Scholz, C. H. 1980. The craton: its effect on the distribution of seismicity and stress in North America. *Earth Planet. Sci. Lett.* 48:348–55
Wiens, D. A., Stein, S. 1983. Age dependence of oceanic intraplate seismicity and implications for lithospheric evolution. *J. Geophys. Res.* 88: In press