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APPLICATION OF MODERN TECHNIQUES TO ANALYSIS OF HISTORICAL EARTHQUAKES

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ABSTRACT

Analysis of historical earthquakes is essential for interpretations of seismicity in regions away from recognized plate boundaries, due to long earthquake recurrence periods. We have applied a number of techniques developed for WWSSN data to a variety of instrumental records dating back to the 1930's. Basic focal mechanism constraints can be obtained from P- and S-waves polarities and S-wave polarizations. Surface wave amplitudes and body wave modeling provide further mechanism constraints and allow seismic moment estimation. Body wave modeling also yields source depth and time function information. These techniques can be combined to find a solution which simultaneously fits the different data types when only a few records are available. In one approach, the moment variance reduction technique, the model space is systematically searched for the mechanism which provides the best fit to P-, SH-, Love, and Rayleigh wave amplitudes. We present three examples where significant tectonic processes were identified from analysis of historical earthquake data. Such data first demonstrated the large magnitude strike-slip "intraplate" seismicity along the "aseismic" Ninetyeast Ridge, showed the complexity and extent of the deformation in the Chagos region, and provided the key constraints for a new tectonic model in which a diffuse plate boundary runs east from the Central Indian Ridge to the Ninetyeast Ridge, and northward along the Ninetyeast Ridge to the Sumatra Trench. Similarly, historical earthquakes along the eastern Canadian passive margin demonstrate that such margins can remain tectonically active, especially in the presence of stresses due to deglaciation or rapid sediment loading. Finally, the historical seismicity of the young lithosphere in the southeast Pacific provides important information on the complex stress release mechanism in such areas.

1. Introduction

With the advent of the WWSSN in the early 1960's and of digital networks in the late 1970's, the past 25 years have seen dramatic increase in the quantity and quality of seismological data. As a result, many investigations relying on compilations of seismicity restrict themselves to the recent data, and overlook the existence and availability of useful seismograms predating 1962, despite the fact that teleseismic instrumental recording dates back to 1889. In particular, historical seismological

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data (defined here as pre-WWSSN) are essential for studies of earthquakes aimed at understanding the tectonics of areas with longer seismic recurrence intervals, such as intraplate regions or slow, diffuse plate boundaries. In this application, restriction of the analysis to recent earthquakes can lead to completely erroneous results.

The purpose of this paper is to demonstrate the application of modern analysis techniques to historical seismological data. Naturally, the results are more uncertain than for recent data largely for three reasons. First, the number of available records is much less, and thus the dataset from which results are drawn is much smaller. Second, instrument responses are much less well known. Finally, the low gain of many historical seismometers prevents study of many smaller events. In general, despite these difficulties, we can obtain reasonable and valuable results.

Our examples will be drawn from oceanic "intraplate" earthquakes, as they provide excellent examples of the need for, and value of, historical earthquake studies. Typically, in these situations, the recurrence interval is long enough that post-WWSSN data alone frequently give a misleading impression. Moreover, due to their oceanic locations, cultural accounts of older earthquakes are often not available. Thus the results of historical earthquake studies, despite the uncertainties inherent in such work, are essential precisely because they provide primary tectonic information. For example, the Ninetyeast Ridge in the Central Indian Ocean was considered "aseismic", and the eastern Canadian passive margin was considered tectonically inactive, until historical earthquakes were analyzed. This effect should not be surprising; based on instrumental seismicity alone (even including pre-WWSSN records), neither the southern San Andreas fault nor the New Madrid, Missouri area would be considered zones of intense seismic activity. Only cultural accounts of the 1857 and 1811-12 earthquakes provide the key evidence for their true tectonic environment.

Less than ten years ago historical earthquake analyses could be considered unorthodox and suspect. Nonetheless, as the number of such investigations started growing, their results were important enough to draw attention. The present IASPEI symposium, on "Historical Seismograms and Earthquakes", serves proof that these endeavors are now generally accepted as an important and valuable seismological tool. Rather than attempt to summarize the many such studies, our purpose in this paper is to offer a perspective by briefly reviewing some of the sources of information and methods of analysis we have used and some of our most significant results.

2. Sources

A number of sources of information are helpful in identifying historical earthquakes of significance and conducting further studies. Perhaps the single most valuable reference, and the starting point of any study, is Gutenberg and Richter's classic *Seismicity of the Earth* (1965). We will see that references to it would have avoided several misconceptions, notably that the Ninetyeast Ridge and the Canadian passive margin were aseismic. The 1965 edition, a re-issue of the 1954 edition, discusses earthquakes through 1952. The book divides earthquakes into various regions, and classifies them as shallow, intermediate, or deep. Gutenberg and Richter personally computed all magnitudes given in the book. Geller and Kanamori (1977) critically discussed their method, and concluded that these magnitudes are basically consistent with present day M_S. The data used for locations by Gutenberg and Richter can be found in the quarterly International Seismological Summary (ISS), an indispensable source which covers the years 1913-1963 and is the predecessor to the current *Bulletin* of the International Seismological Center (ISC). Gutenberg's annotated personal copy of the ISS, and his notepads which contain his original computations for relocation and magnitudes, are available on microfiche (Seismological Society of America, 1980). Rothé's *Seismicity of the Earth* (1969) is a bilingual continuation of Gutenberg and Richter's book, covering the years 1953–1965. During 1957–1963, the information contained in the ISS becomes more succinct, and smaller events are not included. *Starting in 1964, the ISC's Bulletin* and *Regional Catalogue of Earthquakes* provide the wealth of data generated by the full development of the networks.

Several worldwide listings of seismological stations were compiled starting in the 1930's. These are very useful in identifying which stations can be sources of data; additionally, they contain crucial information on the often murky issue of instrument types and responses. Most useful are the National Research Council's *List of Seismological Stations of the World* (McComb and West, 1931) the Royal Belgian Observatory's *Liste des Stations Séismologiques Mondiales* (Charlier and van Gils, 1953), and for the United States, the U.S. Geological Survey's *Historical Survey of U.S. Seismograph Stations* (Poppe, 1979).

3. Techniques

The methods used for analysis of abundant data for modern earthquakes can generally be used for historical earthquakes, with reasonable success. We shall see that first motions, waveform modeling for both P and S body waves, and surface wave amplitude and time domain synthesis can be very valuable. Thus, it is often possible to derive reasonably well constrained mechanisms, depths, and seismic moments.

Frequently, excellent first motions can be extracted from the data, allowing the polarities of P-, SH-, and SV-waves and S-wave polarizations to be used. Figure 1 shows a long period P-wave, recorded at La Paz, Bolivia for a 1947 $M_S = 6.9$ earthquake located about 300 km northwest of the Pacific-Antarctic Ridge studied by Okal (1984). The arrival is clearly compressional. Such first motion data constrain one nodal plane well (Figure 1). The second plane, though less well determined, fits a variety of constraints including the SV/SH ratio at one station and the low amplitudes of both Love and Rayleigh waves at another. The data thus require a large component of thrust faulting on a north-south plane. The result, despite this uncertainty, is tectonically significant, in that it excludes the simplest explanation, that of a mislocated transform fault event with the expected geometry, left-lateral strike-slip on a northwest trending plane; indeed, the record shown on Figure 1 proves the point by itself.

The International Seismological Summary (ISS) offers valuable P-wave first motion information for earthquakes from about the 1930's. Though much inferior to actual examination of the data, ISS first motions can be usefully employed when a set of adjacent stations report consistent polarities. Polarities are reported using the cryptic notation "a" or "k": "a" for "anaseismic" (away from the focus, or compressional; nowadays "C"), and "k" for "kataseismic" (backwards, to the focus, or dilatational; nowadays "D").

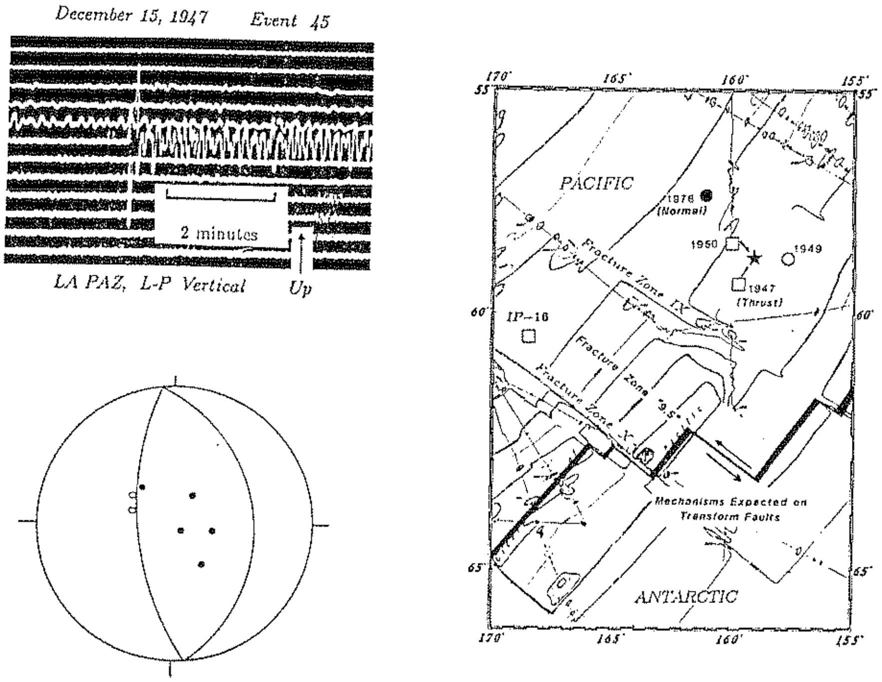

Figure 1. Sample record (top left), first motions (bottom left), and location (right) for the 1947 $M_S = 6.9$ southern Pacific earthquake [see Okal (1984) for details]. The first motion data (open circles dilatational; closed ones compressional) preclude the most likely tectonic setting, as left-lateral strike-slip would be the expected mechanism on a nearby transform fault.

Given enough data, and provided instrument responses are known, waveform modeling can be quite effective. Figure 2 shows P and SH waveforms for a 1944 M = 7.2 earthquake near the Chagos-Laccadive Ridge in the central Indian Ocean (Wiens, 1985). Synthetic body waves, generated for a thrust fault geometry derived from first motions and surface waves, match the data well and constrain the focal depth to about 12 km for P- and SH-waves at various stations. This modeling also shows the seismic moment of this event, 1.5×10^{27} dyne-cm is unusually large for an "intraplate" event.

Surface wave modeling can also be quite valuable. With abundant modern data, azimuthal coverage is generally adequate to construct an observed radiation pattern from equalized spectral amplitudes. This is hard to obtain for historical events. It is, however, often possible to employ the ratio of Love to Rayleigh waves at a small number of stations to provide a valuable constraint, especially if one nodal plane is constrained by first motions. Figure 3 shows an example for the 1939 $M_S = 7.2$ earthquake on the Ninetyeast Ridge, in the central Indian Ocean (Stein and Okal, 1978). First motions, mostly taken from the ISS, are scattered but show two clear groups which a nodal plane must bisect. Additional constraints come from surface wave ratios at two stations. Figure 3 shows a strainmeter record; after bandpass

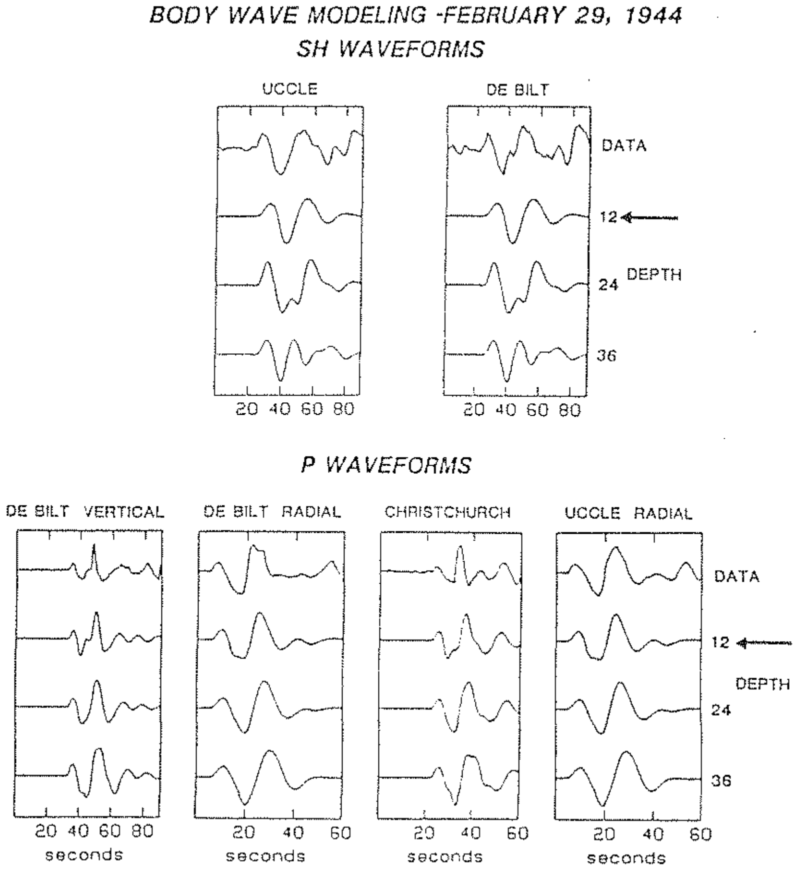

Figure 2. Synthetic body waves and data for the 1944 $M = 7.2$ Chagos-Laccadive Ridge earthquake (Wiens, 1985). The waveforms are well modeled for a focal depth of 12 km for a variety of stations and instruments.

filtering, to exclude periods below 100 s, the Love/Rayleigh ratio is about 1.5. Synthetic seismograms show that the Love/Rayleigh ratio is extremely sensitive to the slip angle on the north-south nodal plane. The best match is obtained for a slip angle of 55°, a mixture of thrust and left-lateral motion. The power of the method resides in that in many cases, the Love-to-Rayleigh ratio at a given station can span several orders of magnitude as a function of an unconstrained parameter such as slip angle, thus reducing considerably the influence of uncertainties in the gains of the various instruments used (which are inherent in the method since Love and Rayleigh waves are usually studied on different components). This is illustrated in Figure 4, for another earthquake near the Nintyeast Ridge. One fault plane is constrained by P-wave first motions, Love/Rayleigh ratios at Resolute and De Bilt uniquely determine the other nodal plane.

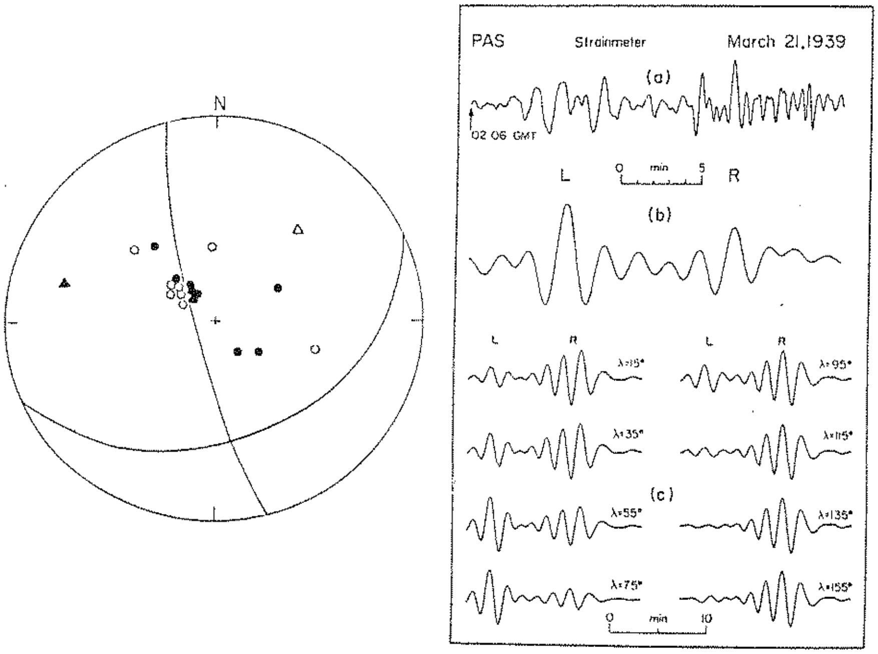

Figure 3. First motion and surface wave data for the 1939 $M_S = 7.2$ Ninetyeast Ridge earthquake (Stein and Okal, 1978). Black triangle is P axis; open triangle is T axis. Both nodal planes were obtained from surface wave data, subject to the first motion constraint (left). One surface wave datum, the Pasadena strainmeter record, is shown. The data were filtered for periods above 100 s, and modeled with various slip angles; 55 degrees yields the best fit.

Because of the limited number of records available and the often poor station distribution encountered in most studies of historical seismicity, it is desirable to analyze as many different types of data as possible in mechanism determination. Figure 5 shows an example of a particularly useful procedure, the moment variance reduction technique, applied to the 1944 Chagos earthquake mentioned previously (Wiens, 1985). The P-wave polarities are insufficient to constrain the mechanism, as all clear first motions read from actual records (large symbols) show compression. The plots on the left show the seismic moment as determined from P-, Love and Rayleigh waves at two stations of different azimuths, as a function of fault strike. The mechanism which yields the lowest variance in the seismic moments (and thus provides the best fit to the amplitudes of the various types of data), is a thrust fault with northwest striking nodal planes. (In the actual analysis, data from 4 stations were used, and the moment variance of all possible mechanisms, including those with strike-slip components, were computed. The mechanism shown produced the minimum moment variance.)

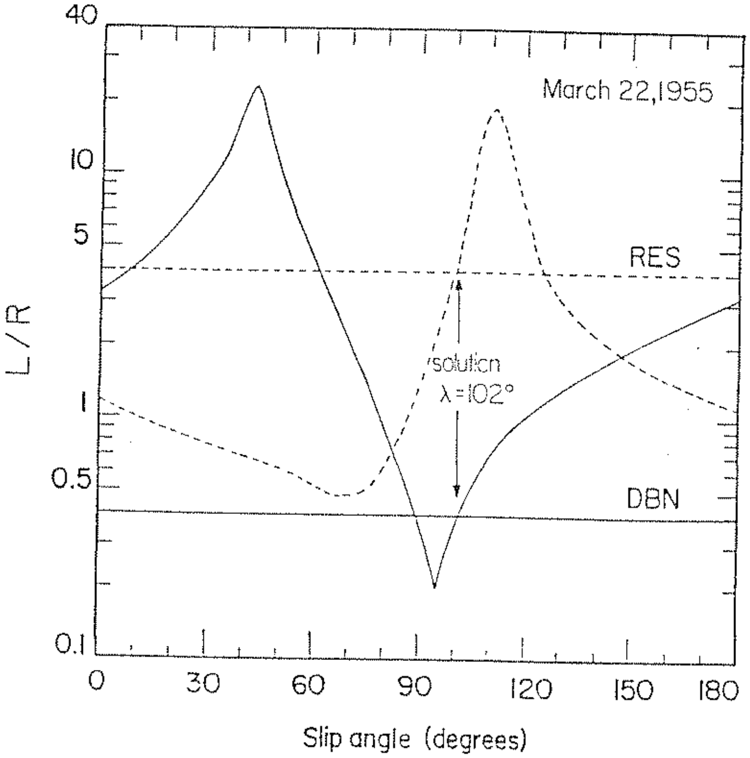

Figure 4. Surface wave data for the 1955 $M_S = 7.0$ earthquake just east of the Ninetyeast Ridge (Stein and Okal, 1978). One nodal plane is constrained by first motions. The slip angle on this plane must satisfy Love-to-Rayleigh ratios at two stations simultaneously.

4. Results

In this section, we show how seismological studies of historical earthquakes have cast decisive light on the large scale tectonic activity of three supposedly "aseismic" regions.

4.1. Indian Ocean

Tectonic models of the Indian plate are the best example of major geological results obtained from historical earthquake data. Until the late 1970's, the Indian plate was considered to be a conventional rigid plate, extending from the Central Indian Ridge to the subduction zones bordering the Pacific plate. Marine magnetic anomaly data showed that the plate was evolved from at least three previously distinct plates, now separated by linear highs known as "aseismic ridges" (Figure 6). Relative plate motion resolvable from the magnetic data along the Ninetyeast Ridge ceased 35–40 Ma ago (Sclater and Fisher, 1974; Eguchi *et al.*, 1979). Similarly, the Chagos-Laccadive Ridge was bordered by an active transform until 30 Ma ago (Fisher *et al.*, 1971).

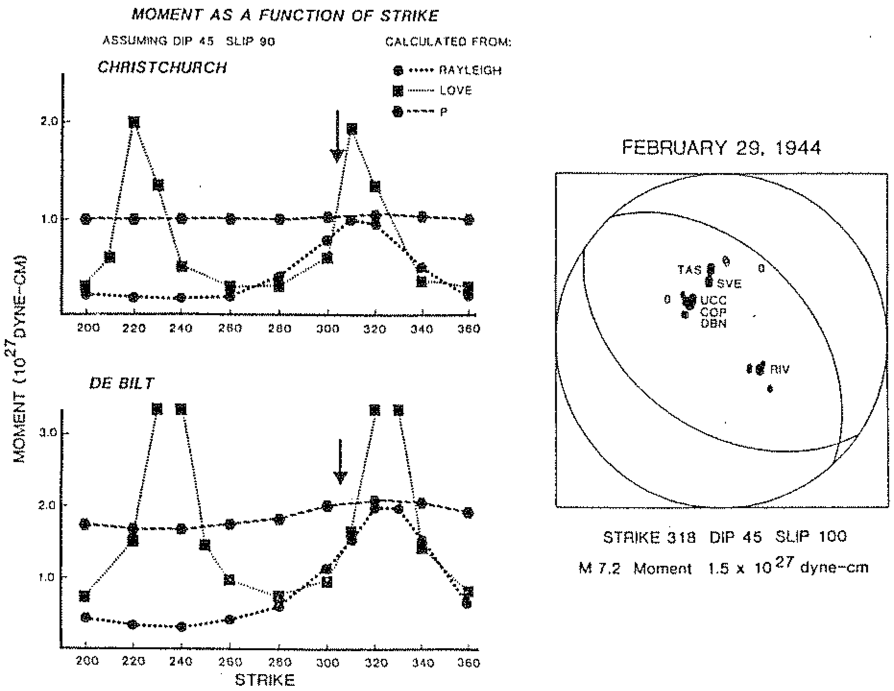

Figure 5. Mechanism determination for the 1944 $M = 7.2$ Chagos-Laccadive Ridge event (Wiens, 1985). All high quality first motions show compression; the final mechanism was constrained by minimizing the variation in seismic moments determined from body and surface waves. This process is shown for two stations.

On the other hand, until recently, present day global plate motion models (Le Pichon, 1968; Chase, 1972, 1978; Minster *et al.*, 1974) assumed a single rigid Indian plate. This assumption was reasonable given the seismicity maps in general use showing only events beginning in the 1960's (e.g., Tarr, 1974). Stein and Okal (1978) pointed out that examination of the historical seismicity yielded a very different picture. Far from being "aseismic", the Ninetyeast Ridge area is a quite active seismic zone; four magnitude seven or greater earthquakes (including one with $M_S = 7.7$) and ten magnitude six events have occurred in the general area since 1913 (Figure 6). This level of activity is much greater than along any spreading ridge and comparable to that along the southern San Andreas fault. The larger earthquakes occur along the segment of the Ninetyeast Ridge north of about 10°S. Stein and Okal (1978) pointed out that this phenomenon had, in fact, been noted by Gutenberg and Richter as early as 1941 in the first edition of *Seismicity of the Earth*:

> "A peculiarly isolated group of shocks occurs near 2°S, 89°E ... With other epicenters near 90°E north of the Equator, there is suggested a minor seismic belt following imperfectly known rises and ridges roughly north and south."

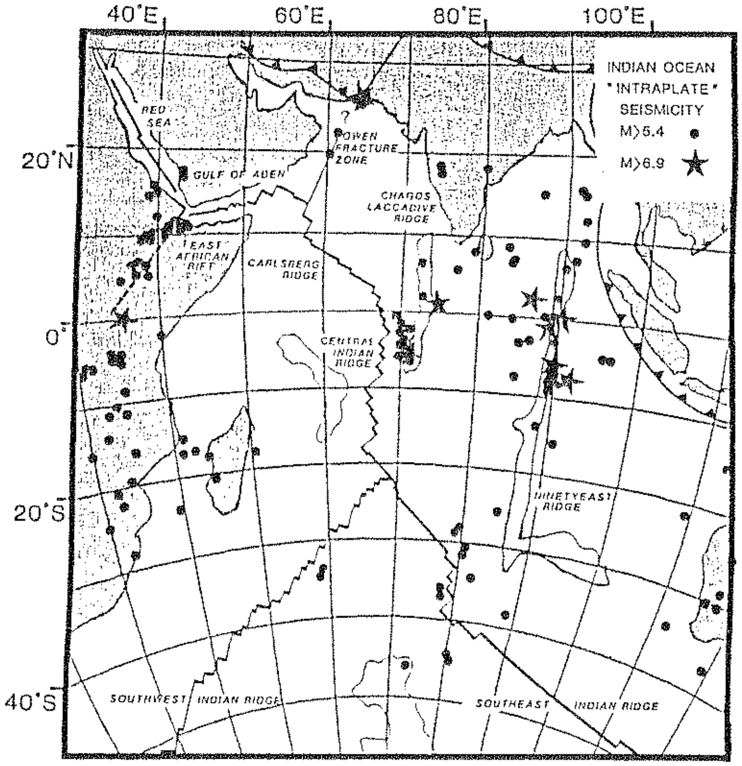

Figure 6. "Intraplate" earthquakes in the Indian Ocean region (1912-present) (Wiens *et al.*, 1985). Note the intense seismicity between the Ninetyeast and Chagos Ridges and the low level of seismicity along the Owen Fracture Zone.

Stein and Okal (1978) studied the largest historical earthquakes in the Ninetyeast Ridge area, including the 1939 $M_S = 7.2$ earthquake discussed earlier (Figure 3), and several more recent earthquakes. Based on these analyses, summarized by Figure 7, they concluded that:

> "The Ninetyeast Ridge is presently a complex zone of deformation within the Indian plate. The northern portion (3°N - 10°S) of the ridge is the active seismic zone, where both vertical and strike-slip motion occur, while further south the ridge is far less seismic ... The strike-slip motion is left lateral, which is consistent with the Indian (west) side encountering resistance due to the collision with Asia while the Australian (east) side is subducting smoothly at the Sumatra trench ... To the west the topography can be interpreted as the result of NW-SE compression which takes place largely aseismically but is observed for one large earthquake. This significant intraplate deformation may explain the difficulties that occur in attempts to close the India – Africa – Antarctica triple junction using a rigid Indian plate."

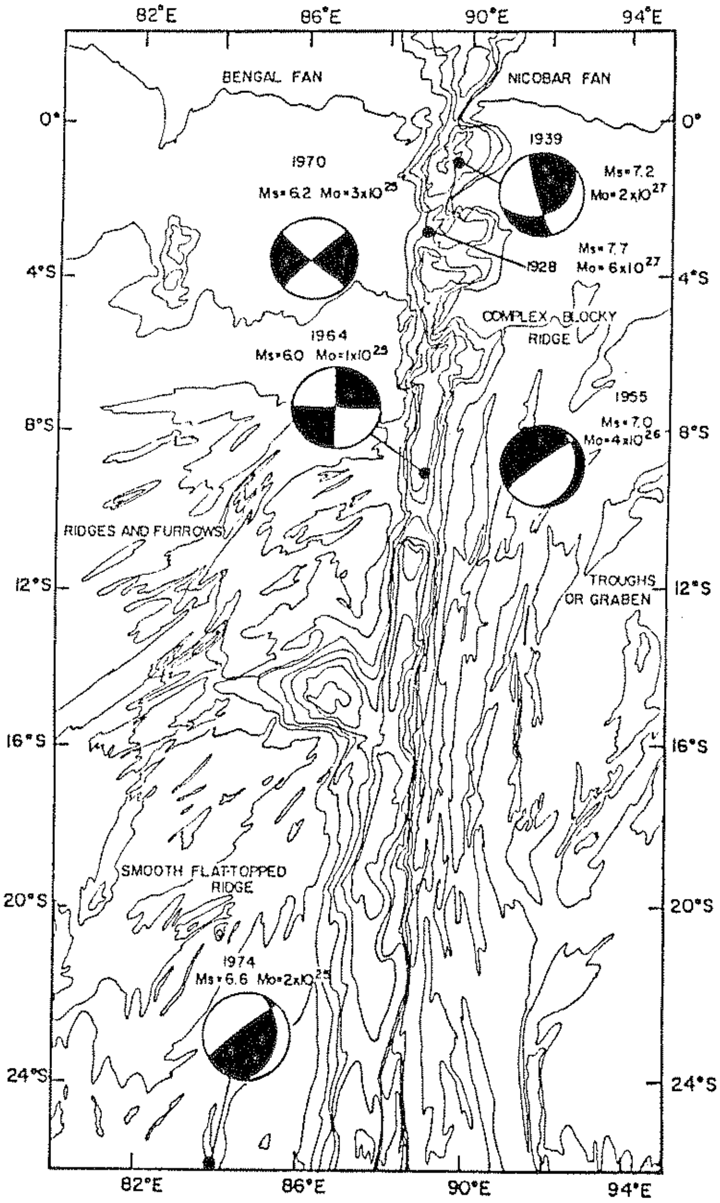

Figure 7. Results of Ninetyeast Ridge earthquake study (Stein and Okal, 1978). Left-lateral strike-slip motion occurs along the Ridge; north-south compression occurs to the west. Note that the moments of the 1928 and 1939 earthquakes are much larger than the recent events. The mechanism of the 1928 earthquake could not be reliably determined, though the data are consistent with a mechanism similar to that of the 1939 event.

They further estimated from seismic moment release that the slip rate along the Ninetyeast Ridge was about 2 cm/yr, which is greater than on some generally recognized plate boundaries. Moreover, no comparable oceanic "intraplate" seismicity occurs elsewhere. (The only other known oceanic intraplate magnitude seven earthquakes occur along passive continental margins or at sites of active volcanism like Hawaii.) The general tectonic model derived from the historical earthquakes has been confirmed by studies of recent earthquakes (Bergman and Solomon, 1985; Stein *et al.*, 1986).

An equally intriguing concentration of seismicity occurs along the Chagos-Laccadive Ridge, a parallel "aseismic" ridge to the west, where normal faulting earthquakes, including a 1965-1966 swarm (Stein, 1978) and a 1983 M_S = 7.5 event (Wiens and Stein, 1984), indicate N-S extension. These studies of recent seismicity seemed to indicate that the Chagos area was an isolated region of N-S tensional deformation, possibly similar to other normal faulting earthquakes found elsewhere in young oceanic lithosphere (Wiens and Stein, 1984; Bergman and Solomon, 1984). However, once again, the recent events are somewhat unrepresentative, as relocation of historical seismicity and analysis of the 1944 M = 7.2 thrust faulting event discussed earlier (Figures 2 and 5) show that the deformation in the Chagos area extends a considerable distance northeast of Chagos Bank and includes both thrust and normal faulting (Wiens, 1985). This finding suggests the Chagos seismicity is part of a deformation zone observed in the Central Indian Basin between the Chagos-Laccadive Ridge and the Ninetyeast Ridge, which includes unusual faulting and folding (Weissel *et al.*, 1980) and high heat flow (Geller *et al.*, 1983). Wiens (1985) proposed that the deformation observed at Chagos Bank, the Central Indian Basin, and the Ninetyeast Ridge indicates a continuous zone of deformation stretching from the Central Indian Ridge to the Sumatra Trench.

These observations provide a basis for investigating the Indian plate's internal deformation. Minster and Jordan (1978) noted that relative motion data in the Indian Ocean were poorly fit, as indicated by misclosure of the Indian Ocean triple junction, which suggests deviations from the rigid plate model used. Splitting the Indian plate along the Ninetyeast Ridge improved the fit to the data, and predicted motion in the Ninetyeast Ridge consistent with the seismological results. Stein and Gordon (1984) showed that separate Indian and Australian plates were statistically resolved, as the improved fit to plate motion data was greater than expected purely by chance, given that the new model has three additional parameters corresponding to the additional plate. Recent modeling (Cloetingh and Wortel, 1985) using the distribution of forces at the boundaries of the conventionally defined Indian plate, predicts a stress field with strong similarities to the observed seismicity, including extension in the Chagos area and large compressional stresses in the Ninetyeast area.

A recent model for Indian Ocean tectonics, which differs significantly from the conventional one, appears to explain the seismicity and plate motion data (Wiens *et al.*, 1985). The conventional model is that Australia and India are contained in a single Indian plate divided from an Arabian plate at a discrete boundary, the Owen Fracture Zone. In the new model (Figure 8), motion along the nearly aseismic Owen Fracture Zone is negligible, and Arabia and India are contained within a single Indo-Arabian plate. The Indo-Arabian plate is divided from the Australian plate by a diffuse boundary, which trends E-W from the Central Indian Ridge near

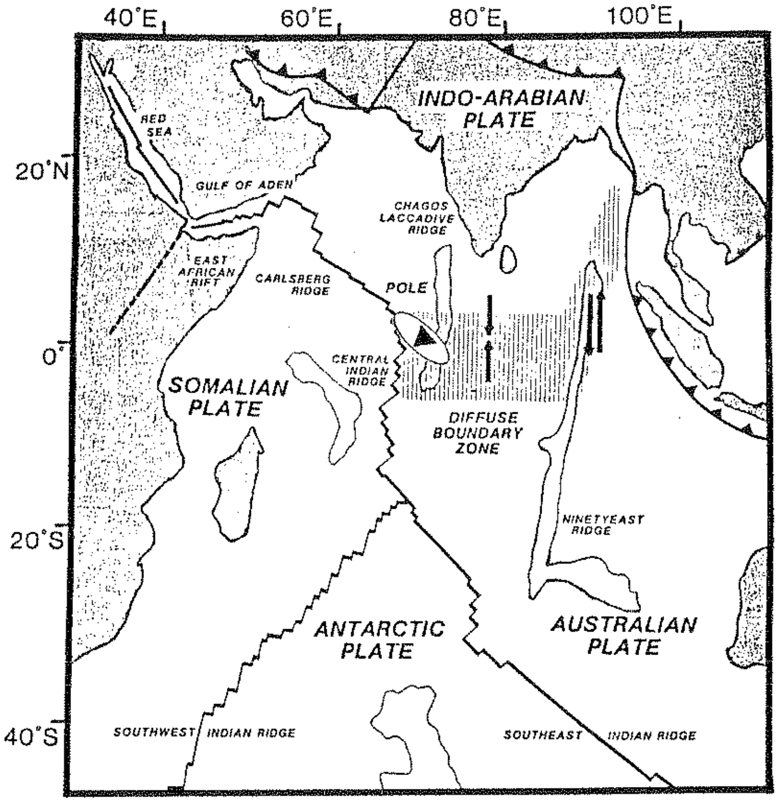

Figure 8. Schematic of the plate geometry proposed by Wiens *et al.* (1985). A diffuse boundary, with an Euler pole near the Chagos-Laccadive Ridge, separates Indo-Arabian and Australian plates. The Owen Fracture Zone is inactive. This geometry is consistent with both seismicity and plate motion data.

Chagos Bank to the Ninetyeast Ridge, and north along the Ninetyeast Ridge to the Sumatra Trench. This diffuse boundary is the zone of concentrated seismicity and deformation previously characterized as "intraplate". Relative motion data along the Carlsberg Ridge are fit significantly better by the new model than by the conventional model. The rotation vector of Australia relative to Indo-Arabia lies just east of the Central Indian Ridge, and is consistent with the about 2 cm/yr of left-lateral strike-slip observed seismologically along the Ninetyeast Ridge, with N-S compression in the Central Indian Ocean, and N-S extension near Chagos. The complexity of the deformation near Chagos is undoubtedly due to the proximity of the rotation vector. The diffuse boundary, possibly initiating in late Miocene time (Weissel *et al.*, 1980), may be related to the suturing of the Arabian plate to the Indian plate and the opening of the Gulf of Aden, or to the uplift of the Himalayas. The convergent segment of this boundary is unusual because neither a mountain range nor a subduction zone occurs along it. This segment may represent an early diffuse stage of the evolution of a convergent plate boundary; a process which may culminate with the onset of "true" subduction.

With the new model the non-closure of the Indian Ocean triple junction is reduced by 40 percent. Thus this model, though not a panacea for all problems of Indian Ocean plate kinematics, provides a simple description of motion in terms of idealized internally rigid plates, where one boundary is diffuse, not discrete. Understanding of this complex tectonic environment will improve as additional data accumulate and provide better resolution of both the diffuse boundary and motion along it. This model could not have been developed without the insight provided by historical seismicity. Consideration of only recent earthquakes underestimated the "intraplate" seismicity and resulted in plate geometry models with a single rigid Indian plate. Analysis of historical seismicity enabled the estimation of the direction and rate of motion along the Ninetyeast Ridge and established the extent and complexity of the deformation near Chagos.

4.2. Canadian Passive Margin Seismicity

A second interesting tectonic process, first observed using historical earthquake data, is intraplate seismicity at passive margins, where continental and oceanic lithosphere join. Although these areas are in general tectonically inactive, major earthquakes can occur. The type example is the active seismic zone on the eastern coast of North America (Stein *et al.*, 1979), which includes the 1929 Grand Banks (M_S = 7.2) and 1933 Baffin Bay (M_S = 7.3) earthquakes (Figure 9). Analysis of data from the 1933 earthquake, and more recent seismicity, suggested that the earthquakes were divided into thrust faulting seaward of the 1000 m isobath, and normal faulting landward. Stein *et al.* (1979) proposed that these earthquakes were associated with stresses induced by the removal of Pleistocene glacial loads extending onto the continental shelf, which reactivated the faults remaining along the continental margin from the original rifting. Simple flexure calculations predict stresses of 100-150 bars, adequate to trigger earthquakes. The only difficulty for the model, the large lithospheric thickness required by the 40 km depth estimated for the 1933 earthquake from the one available record, has been resolved by additional records which indicate a 16 km depth (Sleep *et al.*, 1986).

This general model has been accepted and extended by subsequent studies, including analysis of microseismicity in Baffin Bay (Reid and Falconer, 1982). Quinlan (1984) and Sleep *et al.* (1986) included the effect of the stress field that occurs on all passive margins due to the differing densities of continental and oceanic lithosphere (Bott and Dean, 1972; Artyushkov, 1973). This stress field is extensional in the continent and compressive seaward, in accord with the focal mechanisms. Thus, the effect of deglaciation appears to be to trigger earthquakes, as such events are observed primarily on glaciated marigins, such as Canada, Fennoscandia, and Greenland (Figure 10). This observation has obvious implications for the nature of seismic hazards along previously glaciated margins such as the northeastern United States: the largest potential for destructive earthquakes arises from the continent-ocean boundary rifting faults.

The same principal applies, to a lesser extent, on non-glaciated passive margins. Sediment loading can, in itself, generate substantial flexural stresses (Cloetingh *et al.*, 1984). Stein *et al.* (1979) suggested that, in general, these loads are much less effective at inducing earthquakes, since the sediment loads are usually in place long enough for the stress to relax. Extremely rapid sedimentation is required to induce

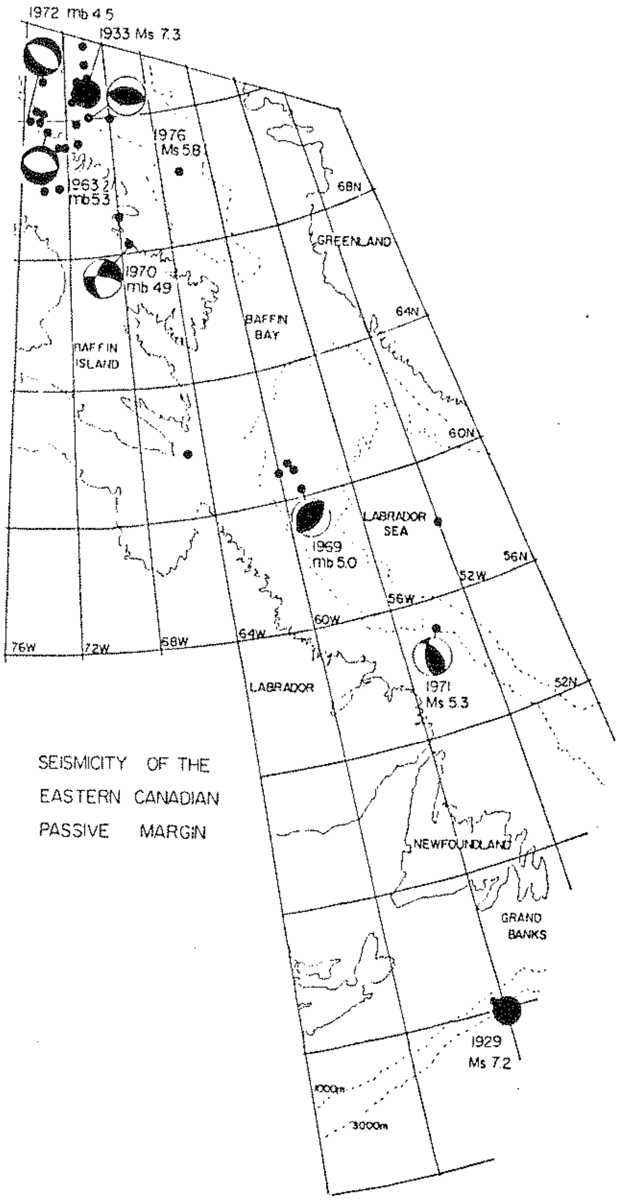

Figure 9. Seismicity and focal mechanisms for the eastern Canadian passive margin (Stein *et al.*, 1979). Earthquakes seaward of the 1000 m contour show thrust faulting; landward events show dominantly normal faulting. The mechanisms are consistent with reactivation of the continental margin rifting faults by flexure induced by deglaciation. The stress field at the margin due to differences in continent-ocean density structure also predicts the observed mechanisms.

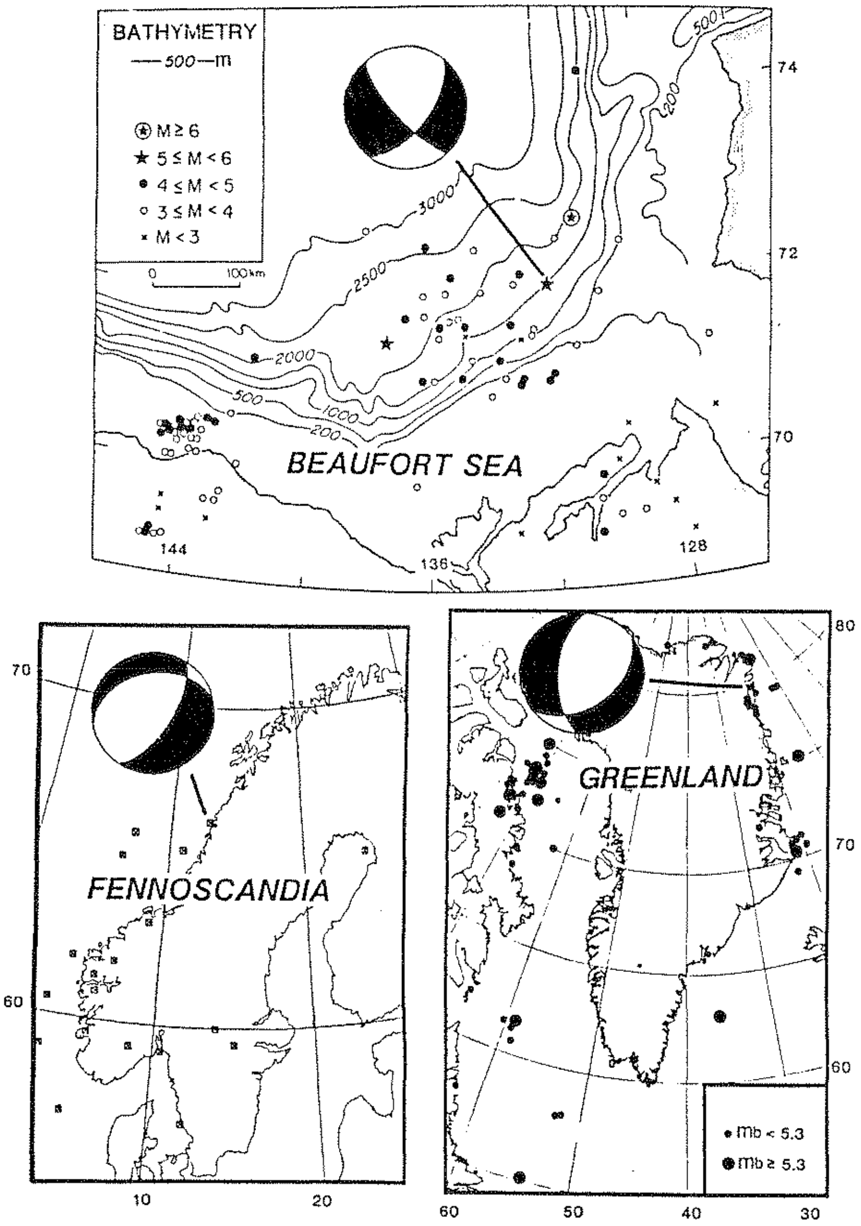

Figure 10. Seismicity of three other recently deglaciated passive margins (Sleep *et al.*, 1985); all show normal faulting inland as predicted by the eastern Canadian model. The Beaufort Sea seismicity (1920-1977) and mechanism (top) are from Hasegawa *et al.* (1979). The Fennoscandian seismicity (1497-1975) is from Husebye *et al.* (1978); the mechanism is by Bungum *et al.* (1979). The Greenland seismicity and mechanism are from Sykes (1978).

seismicity. One such case is the Gulf of Mexico; the small earthquakes in the area can be interpreted as a consequence of the rapid sediment loading history (Nunn, 1985).

As with the Indian Ocean deformation, the primary evidence for the importance of passive margin seismicity comes from the historical earthquakes. The recent seismicity is adequate to confirm its existence, but too small to convincingly demonstrate the magnitude of the effect.

4.3. Seismicity Patterns in the Eastern Part of the Pacific Plate

The easternmost part of the Pacific plate, generated at the fast-spreading East Pacific Rise in the past 20 Ma, has been the site of occasional yet intense seismic activity (see Figure 11). This seismicity is well documented by the post–WWSSN data, but only the historical earthquakes show its true dimension. As shown on Figure 11, 8 events reached or exceeded magnitude 6 between 1937 and 1962. The preferential location of these events in the young oceanic lithosphere is consistent with the observation that the rate of oceanic intraplate seismicity decreases with increasing lithospheric age (Wiens and Stein, 1983). A full study of these events is given by Okal (1984) and Okal and Wiens (1985).

At the southernmost epicenter, the 1947 earthquake (see Figure 1), the additional shock in 1950, and the probable location of the major 1949 event in its immediate vicinity, clearly reveal the area as one of preferential intraplate seismicity. The fact that the 1947 shock has a thrust faulting mechanism, as opposed to the normal faulting of a smaller event in 1976, some 200 km northwest, is a superb example of the complex regime of tectonic stress in the youngest portions of oceanic plates (Wiens and Stein, 1984; Bergman and Solomon, 1984). Once again, the historical data rules out the simplest explanation – in this case, that the lithosphere would be in a state of extension in the direction perpendicular to the ridge.

The 1955 earthquake at 25°S, 122°W, is a rare case of a large normal faulting earthquake on the flanks of a major seamount. Its similarities with shocks following documented episodes of volcanic activity suggests that the seamount may still be magmatically active.

Finally, the 1945 event further north and the recent (1984) swarm at 20°N, 116°W, further identify the fringe of the Pacific plate along the northern portion of the East Pacific Rise as a region of enhanced seismicity, particularly at the level of magnitude 6 or greater. This is in contrast to the bulk of the plate (the "F" region in Figure 11), where no magnitude 6 events are known outside of Hawaii. Okal (1984) has proposed to interpret this intriguing pattern in the framework of the reorientation which accompanied the Miocene ridge jump in the Eastern Pacific, and resulting in a geometry of particular vulnerability in relation to the intraplate tectonic stress. This picture could not have evolved without the crucial data provided by the historical earthquakes: Figure 11 shows only one event with $M_S > 6$ recorded inside the Pacific plate (Hawaii excepted) during the first 19 years of the WWSSN.

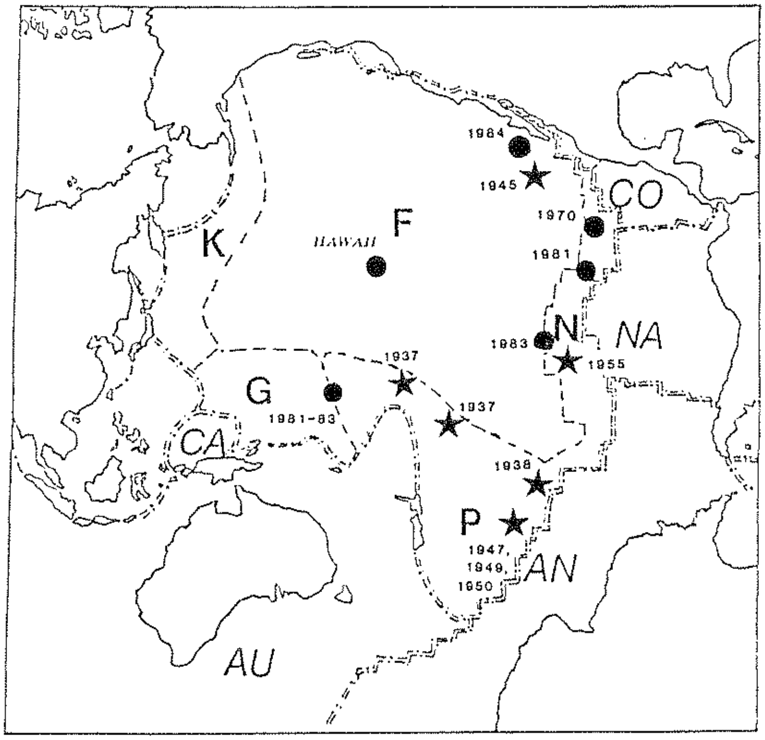

Figure 11. Map of Pacific intraplate earthquakes with at least one confirmed magnitude ≥ 6 (after Okal (1984)). Stars are historical events, dots epicenters known from post-WWSSN seismicity. This map identifies portions of the Pacific plate generated at various spreading ridges (see Okal (1984) for details). Note that magnitude 6 or greater seismicity is common outside Region "F", generated at the Farallon ridge, but that post-WWSSN seismicity gives a very incomplete picture.

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REFERENCES

Artyushkov, E. V. (1973). Stresses in the lithosphere caused by crustal thickness inhomogeneities, *J. Geophys. Res.*, **78**, 7675-7708.

Bergman, E. A. and S. C. Solomon (1984). Source Mechanisms of Earthquakes near Mid-oceanic ridges from body waveform inversion: implications for early evolution of oceanic lithosphere, *J. Geophys. Res.*, **89**, 11415-11441.

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

Bott, M. H. P. and D. S. Dean (1972). Stress systems at young continental margins, *Nature*, **235**, 23-25.

Bungum, H., B. K. Hokland, E. S. Husebye, and F. Ringdahl (1979). An exceptional intraplate earthquake sequence in Meloy, Northern Norway, *Nature*, **280**, 32-35.

Charlier, C. and J. M. van Gils (1953). Liste des Stations Séismologiques Mondiales, Observatoire Royal de Belgique, Uccle.

Chase, C. G. (1972). The N plate problem of plate tectonics, *Geophys. J. R. Astr. Soc.*, **29**, 117-122.

Chase, C. G. (1978). Plate kinematics: The Americas, East Africa, and the rest of the world, *Earth Planet. Sci. Lett.*, **37**, 355-368.

Cloetingh, S. and R. Wortel (1985). Regional stress field of the Indian plate, *Geophys. Res. Lett.*, **12**, 77-80.

Cloething, S. A. P. L., M. J. R. Wortel, and N. J. Vlaar (1984). Passive margin evolution, initiation of subduction and the Wislon cycle, *Tectonophysics*, **109**, 147-163.

Eguchi, T., S. Uyeda, and T. Maki (1979). Seismotectonics and tectonic history of the Andaman Sea, *Tectonophysics*, **57**, 35-51.

Fisher, R. L., J. G. Sclater, and D. P. McKenzie (1971). Evolution of the Central Indian Ridge, Western Indian Ocean, *Geol. Soc. Am. Bull.*, **82**, 553-562.

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

Geller, R. J. and H. Kanamori (1977). Magnitudes of great shallow earthquakes from 1904 to 1952, *Bull. Seism. Soc. Am.*, **67**, 587-598.

Gutenberg, B. and C. F. Richter (1965). *Seismicity of the Earth and Associated Phenomena*, Hafner, New York, 310 pp.

Hasegawa, H. S., C. W. Chou, and P. W. Basham (1979). Seismotectonics of the Beaufort Sea, *Can. J. Earth Sci.*, **16**, 816-830.

Husebye, E. S., H. Bungum, J. Feyn, and H. Gjøystdahl (1978). Earthquake activity in Fennoscandia between 1497 and 1975 and intraplate tectonics, *Norsk Geol. Tidsskrift*, **58**, 51-68.

Le Pichon, X. (1968). Sea-floor spreading and continental drift, *J. Geophys. Res.*, **73**, 3661-3697.

McComb, H. E. and C. J. West (1931). List of Seismological Stations of the World, *Bull. National Res. Council*, **82**, 1-119.

Minster, J.-B., T. H. Jordan, P. Molnar, and E. Haines (1974). Numerical modelling of instantaneous plate tectonics, *Geophys. J. R. Astr. Soc.*, **36**, 541-576.

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

Nunn, J. (1985). State of stress in the northern Gulf coast, *Geology*, **13**, 429-432.

Okal, E. A. (1984). Intraplate Seismicity of the southern part of the Pacific plate, *J. Geophys. Res.*, **89**, 10053-10071.

Okal, E. A. and D. A. Wiens (1985). Major normal faulting intraplate earthquakes south of Baja California, Eos, *Trans. Am. Geophys. Union*, **66**, 307 (abstract).

Poppe, B. B. (1979). Historical Survey of U.S. Seismograph Stations, *U. S. Geol. Surv. Prof. Paper 1096*, Washington, D.C., 389 pp.

Quinlan, G. (1984). Postglacial rebound and the focal mechanisms of eastern Canadian earthquakes, *Can. J. Earth Sci.*, **21**, 1018-1023.

Reid, I. and R. K. H. Falconer (1982). A seismicity survey in northern Baffin Bay, *Can. J. Earth Sci.*, **19**, 1518-1531.

Rothé, J.-P. (1969). Seismicity of the Earth / La Sismicité du Globe, 1953-1965, *UNESCO Earth Sciences Ser.*, **1**, Paris, 336 pp.

Seismological Society of America (1980). *Seismology Microfiche Publications from the Caltech Archives, Ser. II and III*, edited by J.R. Goodstein, H. Kanamori and W. H. K. Lee, Berkeley, Calif.

Sclater, J. G. and R. L. Fisher (1974). The evolution of the eastcentral Indian Ocean, with emphasis on the tectonic setting of the Ninetyeast Ridge, *Geol. Soc. Am. Bull.*, **85**, 683-702.

Sleep, N. H., S. Stein, and G. C. Kroeger (1986). Canadian margin stress field inferred from seismicity, *J. Geophys. Res.*, submitted.

Stein, C. A., Wiens, D. A., and J. K. Weissel (1986). Seismicity and thermal constraints on the deformation of the Indo-Australian plate, *Earth Planet. Sci. Lett.*, to be submitted.

Stein, S. (1985). An earthquake swarm on the Chagos-Laccadive Ridge and its tectonic implications, *Geophys. J. R. Astr. Soc.*, **55**, 577-588.

Stein, S. and R. G. Gordon (1984). Statistical tests of additional plate boundaries from plate motion inversions, *Earth Planet. Sci. Lett.*, **69**, 401-412.

Stein, S. and E. A. Okal (1978). Seismicity and tectonics of the Ninetyeast Ridge area, evidence for internal deformation of the Indian plate, *J. Geophys. Res.*, **83**, 2233-2246.

Stein, S., N. H. Sleep, R. Geller, S.-C. Wang, and G. Kroeger (1979). Earthquakes along the passive margin of eastern Canada, *Geophys. Res. Lett.*, **6**, 537-540.

Sykes, L. R. (1978). Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation, *Rev. Geophys. Space Phys.*, **16**, 621-688.

Tarr, A. (Ed.) (1974). World Seismicity Map, U.S. Geological Survey, Washington, D.C.

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

Wiens, D. A. (1985). Historical seismicity near Chagos: Deformation of the equatorial region of the Indo-Australian plate, *Earth Planet. Sci. Lett.*, **76**, 350-360.

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

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

Wiens, D. A. and S. Stein (1984). Intraplate seismicity and stresses in young oceanic lithosphere, *J. Geophys. Res.*, **89**, 11442-11464.

APPENDIX

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