

Oceanic Intraplate Earthquakes: Implications for Local and Regional Intraplate Stress

ERIC A. BERGMAN AND SEAN C. SOLOMON

Department of Earth and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139

Focal mechanisms of intraplate earthquakes provide the only means at present by which to characterize the long-wavelength tectonic stress field in oceanic lithosphere. Stress orientations inferred from focal mechanisms may not accurately reflect the state of stress in the epicentral area, however, or the measured stresses may be dominated by local rather than regional sources. To establish a data set with which to study these possibilities, a comprehensive catalog of 159 oceanic intraplate earthquakes has been compiled for events since 1963 with m_b 4.7 or larger. Focal mechanisms are available for approximately one quarter of the events, and several new mechanisms are presented here. For a representative subset of this catalog (83 events), the bathymetry and tectonic history of the epicentral areas have been assembled, and the earthquakes have been rated according to their association with (1) a preexisting fault zone, which might decouple the P axis of the focal mechanism from the true orientation of maximum compressive stress, and (2) large bathymetric relief, which might be a source of large local stresses. Oceanic intraplate earthquakes are commonly found in association with zones of previous weakness (usually fracture zones), but they do not show any particular association with large bathymetric features. In the central Indian Ocean there are enough focal mechanisms available to establish a well-defined NW-SE orientation for P axes and presumably for the direction of greatest compressive stress. The consistency of the P axes of these widely varying mechanisms in the presence of the Ninetyeast Ridge, a site of major intraplate deformation and large bathymetric relief, is remarkable. A possible explanation is that in the presence of a large number of preexisting faults with a range of orientations, slip occurs on those faults which have large resolved shear stresses from the regional stress field. In such an instance the P axis of focal mechanisms will tend to show a consistent alignment with the true direction of maximum stress.

INTRODUCTION

One of the most direct means of investigating the causes of plate motions is through the state of stress in the lithosphere, since the stress is a response to the distribution and magnitude of forces applied to the plates [Richardson *et al.*, 1979]. The potential of intraplate stress to elucidate the driving mechanism has not yet been fully realized, however, for several reasons: (1) plate driving forces are not the only sources of stress in the plates; (2) there is a severe shortage of data on the stress field, particularly in oceanic lithosphere; and (3) there are legitimate questions concerning the interpretation of various measures of stress.

As a step toward a more complete characterization of the state of stress in oceanic areas, this paper presents a comprehensive synthesis of the distribution and tectonic setting of oceanic intraplate earthquakes and implications for the use of their focal mechanisms to determine the orientation of regional stresses in oceanic lithosphere. A variety of in situ measurement techniques may be used to determine crustal stresses on the continents [McGarr and Gay, 1978], but focal mechanism studies of intraplate earthquakes provide virtually the only data on the state of stress in oceanic lithosphere.

In order to use focal mechanisms to constrain possible driving force systems, we must first know the answer to two questions: (1) does the focal mechanism reflect the state of stress in the volume of rock sampled in an earthquake, or is the P axis only loosely related to the stress field because the rupture occurred on a preexisting fault; (2) if the local stress field is known, does it represent regional stresses (including plate driving forces), or is the earthquake focal mechanism dominated by local sources of lithospheric stress (e.g., topographic loading, thermal stresses)?

To address these questions, we have compiled a compre-

hensive catalog of oceanic intraplate earthquakes for the period 1963-1979, with selected older events. For a representative subset of the catalog (83 events), we consider the local bathymetric and tectonic setting of each epicenter and the source mechanism where one is available. On the basis of this study, the following conclusions have been reached:

1. Oceanic intraplate earthquakes are commonly associated with preexisting zones of weakness, including major faults or areas of active intraplate deformation. There is evidence that focal mechanisms for these events can nonetheless provide strong constraints on the long-wavelength stress field, even in the presence of such weak zones.

2. Oceanic intraplate earthquakes do not show a general association with areas of large bathymetric relief, which might be sources of significant local stress. However, where earthquakes do occur in proximity to large bathymetric features, their mechanisms may provide an opportunity to estimate the absolute level of regional stress in the lithosphere, if the earthquake-generating stress is contributed in roughly equal measure by the regional tectonic stress field and by well-defined local sources of stress [Solomon *et al.*, 1980].

3. There are many oceanic intraplate earthquakes for which focal mechanisms can be obtained through application of recently developed seismological techniques. Source mechanisms of individual events which are apparently free from the biasing effects of preexisting fault zones and large bathymetric relief may provide good control on the long-wavelength stress field in oceanic lithosphere, but emphasis should be placed on obtaining many independent focal mechanisms over a broad area so that regional consistency of stress orientation can be tested.

A CATALOG OF OCEANIC INTRAPLATE EARTHQUAKES

Our first step in the characterization of oceanic intraplate seismicity and its relation to the long-wavelength regional

TABLE 1. Oceanic Intraplate Earthquake Epicenters and Their Association With Zones of Weakness and Large Bathymetric Relief

Date	Origin Time	Latitude, °N	Longitude, °E	m_b^*	M_S	Focal Mechanism	Ratings†		
							I	II	III
1. March 21, 1939	0111:12	-0.9	89.5		7.2	B	3	3	A
2. March 1, 1955	0146:13.0	-19.9	-36.7	6.5		S	3	3	B
3. March 22, 1955	1405:07	-8.8	91.7		7.0	B	2	2	B
4. Aug. 21, 1963	0339:22.6	14.37	-72.51	5.4		D	2	3	A
5. Sept. 14, 1963	1616:51.8	-33.6	-126.7	4.9			1	2	C
6. May 25, 1964	1944:05.9	-9.08	89.89	5.7	6.0	C	3	3	A
7. Sept. 17, 1964	1502:01.5	44.58	-31.34	5.5		E	1	1	B
8. Oct. 11, 1964	1006:43.0	19.06	-156.63	5.2					
9. Oct. 18, 1964	2155:59.7	7.15	144.58	4.7					
10. Oct. 23, 1964	0156:05.1	19.80	-56.11	6.2	6.3	D	1	1	B
11. Oct. 23, 1964	1646:19.5	19.07	-57.5	5.0					
12. Dec. 8, 1964	0914:31.6	-21.30	-82.0	5.0			3	3	A
13. March 6, 1965	1110:52.8	-18.42	-132.82	5.4		K	2	1	C
14. May 21, 1965	0433:13	-28.40	-142.50	5.0			2	3	C
15. July 9, 1965	0041:48.0	7.38	147.4	4.7					
16. Sept. 6, 1965	2113:33.1	6.55	-84.39	5.2					
17. Sept. 9, 1965	1002:25.7	6.51	-84.44	5.8		D	3	3	A
18. Sept. 12, 1965	2202:37.7	-6.46	70.76	6.1	6.0	G	3	3	A
19. Sept. 21, 1965	0326:37.1	40.77	-50.13	5.3			3	3	B
20. Oct. 7, 1965	0336:01.4	12.46	114.45	5.8	5.6	T	1	2	B
21. Oct. 31, 1965	1724:09.5	-14.22	95.27	5.3			2	2	B
22. Nov. 25, 1965	1050:50.8	-17.07	-100.24	5.3	4.9	H	2	1	A
23. Feb. 4, 1966	2053:31	12.39	114.32	4.9	4.1				
24. March 14, 1966	2328:53	6.02	146.32	4.8					
25. March 20, 1966	1828:35.8	21.96	-58.9	5.0			1	1	C
26. March 22, 1966	0752:51	10.13	147.9	4.7					
27. Sept. 18, 1966	0640:41	-18.38	-132.87	5.0					
28. Sept. 24, 1966	0857:17	12.06	-130.82	4.9		E	2	1	C
29. Nov. 29, 1966	0921:22.4	-9.76	90.58	5.0					
30. March 4, 1967	2241:17.6	7.70	146.00	5.0					
31. April 13, 1967	1426:51.0	-6.76	-151.06	5.1		E	3	2	A
32. April 23, 1967	1501:06.2	1.60	80.22	4.9		E	2	2	B
33. April 26, 1967	1311:44.2	-1.10	89.47	4.9					
34. July 10, 1967	1943:58.6	19.30	-53.1	4.8			1	1	C
35. Nov. 10, 1967	1838:34	-6.03	71.34	5.2					
36. Nov. 11, 1967	1155:56	-6.01	71.36	5.3	5.2	G			
37. Nov. 11, 1967	1214:55	-6.03	71.34	5.6	5.8				
38. Nov. 11, 1967	1505:09	-6.12	71.27	5.0					
39. Nov. 11, 1967	1742:17	-6.16	71.45	4.9					
40. Nov. 11, 1967	1759:57	-6.10	71.32	5.4					
41. Nov. 11, 1967	2018:15	-5.99	71.34	5.1					
42. Feb. 9, 1968	2046:45.1	-14.0	82.4	4.7			3	1	B
43. March 2, 1968	2202:24.2	-6.09	71.41	5.5	5.4	G			
44. March 15, 1968	1751:56.9	-6.16	71.36	4.7					
45. March 20, 1968	2200:04	-6.06	71.22	4.8					
46. April 28, 1968	0418:15.5	44.76	174.58	5.5	5.2	EI	3	2	A
47. July 29, 1968	0245:46.1	-7.52	-148.32	4.9		E			
48. Aug. 20, 1968	1116:58.5	5.43	147.11	5.6		FK	3	2	A
49. Sept. 3, 1968	0112:28.2	-37.84	38.0	5.0			3	1	B
50. Sept. 3, 1968	1537:00.3	20.58	-62.30	5.6	5.9	E	1	1	C
51. Sept. 14, 1968	0125:18.9	-24.45	80.41	5.4		E	3	1	B
52. Nov. 26, 1968	0608:55	-3.15	86.41	4.7					
53. Dec. 2, 1968	1324:51	6.91	-104.32	4.8					
54. Dec. 14, 1968	1143:13.6	-3.03	85.53	4.9					
55. Jan. 3, 1969	0356:59.4	-18.30	88.15	5.3			3	3	B
56. Jan. 27, 1969	1440:02.3	12.43	114.37	5.0	4.8				
57. Feb. 14, 1969	0614:53.4	-17.84	87.37	4.7					
58. May 3, 1969	0822:20	8.34	-175.59	5.0		E	3	3	A
59. July 12, 1969	0557:11	-6.03	71.35	5.2					
60. July 25, 1969	2130:33.3	12.44	-40.75	4.8		E	3	1	A
61. Aug. 6, 1969	1715:40.1	-7.61	-148.14	5.0			1	1	A
62. Nov. 21, 1969	0726:36.3	8.04	147.42	4.9					
63. Jan. 11, 1970	0314:24.0	-6.20	71.33	4.9					
64. Jan. 21, 1970	1751:37.4	7.03	-104.24	6.1	6.8	LK	3	1	A
65. Jan. 22, 1970	0611:52	7.29	-104.27	4.7					
66. June 6, 1970	0614:13.3	-62.76	-93.5	4.8			3	1	C
67. Oct. 10, 1970	0853:04.5	-3.56	86.19	5.8	6.2	J	2	2	B
68. May 9, 1971	0825:01.1	-39.78	-104.87	6.0	6.0	M	2	1	B
69. May 9, 1971	0853:25.9	-39.71	-105.10	5.1					
70. May 9, 1971	1800:59.9	-39.77	-105.10	5.4					
71. May 9, 1971	1835:17	-39.72	-104.90	5.3	5.4				

TABLE 1. (continued)

Date	Origin Time	Latitude, °N	Longitude, °E	m_b^*	M_S	Focal Mechanism	Ratings†		
							I	II	III
72. May 27, 1971	2049:59	-53.85	150.03	5.3	5.7	E	3	1	A
73. June 26, 1971	1927:11	-5.18	96.90	5.9	6.4	E	2	2	B
74. Aug. 1, 1971	1852:32.9	18.02	-154.87	4.7					
75. Aug. 15, 1971	0617:15.2	47.46	-49.53	4.8					
76. Sept. 30, 1971	2124:10.8	-0.45	-4.89	6.0	5.5	E	3	1	B
77. Feb. 29, 1972	2208:23.0	19.26	-156.27	4.8					
78. April 22, 1972	0222:50.8	-52.86	-14.96	5.0			1	1	E
79. May 21, 1972	0601:54.3	-27.10	174.97	5.6	4.9	K	3	2	B
80. May 28, 1972	0838:51.8	-32.78	-92.20	4.9			1	1	D
81. Aug. 14, 1972	1045:57.0	14.20	-68.46	4.7	4.5	N	3	1	A
82. Sept. 28, 1972	0657:35.6	-47.71	-119.88	5.1			2	1	C
83. Oct. 20, 1972	0433:49.9	20.60	-29.69	5.7	5.8	O	2	1	C
84. Oct. 30, 1972	0150:35.7	22.35	-61.97	4.9	4.2		3	1	B
85. Nov. 7, 1972	1205:14.3	49.06	-39.42	5.1			1	1	C
86. Dec. 3, 1972	1758:59	-11.24	87.49	5.2			3	3	C
87. Dec. 14, 1972	2049:35.4	-1.34	89.17	5.3					
88. Dec. 23, 1972	1904:45.7	19.55	-156.10	4.8					
89. Jan. 19, 1973	0736:33.7	-7.59	-148.18	4.9	4.6				
90. Jan. 19, 1973	1526:26.4	-7.63	-148.08	4.8					
91. March 20, 1973	1813:24.7	-57.82	83.59	5.2			2	3	B
92. April 26, 1973	2026:27	20.05	-155.16	5.9	6.1	P	2	3	B
93. May 3, 1973	2311:04	-46.14	73.22	5.5		K	3	3	A
94. May 27, 1973	2117:18.2	-2.38	64.96	4.8			2	1	D
95. Aug. 30, 1973	1950:03.9	7.15	84.33	5.8	5.2	K	2	3	B
96. Sept. 15, 1973	1505:25.3	7.54	144.60	5.0					
97. Oct. 29, 1973	0707:36.5	-28.02	83.43	4.9			3	1	C
98. March 12, 1974	1511:03	8.79	151.02	5.4	4.8				
99. March 14, 1974	2322:12.8	8.72	151.06	4.8					
100. April 12, 1974	1745:18.7	14.27	134.37	5.5	4.9	K	3	3	A
101. June 25, 1974	1722:17.9	-26.02	84.30	6.1	6.6	B	3	2	A
102. June 25, 1974	1831:36.9	-25.9	84.10	5.1					
103. June 26, 1974	0559:26.4	-26.0	84.1	4.8					
104. June 26, 1974	0907:25.7	-25.95	84.11	4.7					
105. June 28, 1974	0845:43.7	-25.69	84.01	5.2					
106. Aug. 18, 1974	1524:42.2	-2.60	86.55	5.1					
107. Sept. 9, 1974	1927:52	-54.0	-17.6	4.8			1	1	E
108. Nov. 20, 1974	1321:41.6	-53.59	-28.26	5.8	5.6		1	2	E
109. Feb. 21, 1975	0038:13.5	-52.5	-17.4	4.8			1	1	E
110. March 29, 1975	1502:29.6	-37.77	-138.90	5.0			2	1	C
111. April 7, 1975	1438:07	-37.62	30.98	5.0			2	3	C
112. May 14, 1975	1149:41.9	-17.62	-136.32	5.0			2	2	C
113. May 25, 1975	1416:30.8	-18.39	-132.99	4.9	3.7				
114. Aug. 22, 1975	1940:47.7	-12.26	86.30	4.9					
115. Aug. 28, 1975	1825:45.0	-25.98	84.10	5.1					
116. Oct. 6, 1975	2221:40.9	44.59	-57.04	5.0					
117. Oct. 23, 1975	0536:36.0	-47.65	4.48	4.7			2	2	C
118. Jan. 11, 1976	2322:41.2	-46.40	-101.1	5.3			1	1	E
119. Feb. 20, 1976	1926:39	23.7	154.2	4.8			1	3	D
120. Feb. 21, 1976	0551:13.7	20.16	-156.31	5.0					
121. March 11, 1976	0045:19.9	4.7	-112.1	4.9			1	1	E
122. May 6, 1976	1442:26	36.8	158.0	5.1			2	3	B
123. May 14, 1976	0621:11	3.7	-167.6	4.8			1	1	C
124. Aug. 30, 1976	0837:54.4	1.03	147.56	5.8	5.9				
125. Oct. 21, 1976	0356:28	-57.37	-161.1	5.4			1	1	C
126. Nov. 24, 1976	2150:54.5	33.01	-61.66	5.1		K	3	3	A
127. Dec. 14, 1976	1144:37.5	7.23	89.91	4.9			3	3	C
128. Dec. 28, 1976	0257:38.5	22.1	-63.45	5.0	4.4	K	2	1	C
129. Jan. 9, 1977	0831:16.1	6.00	146.77	4.9					
130. Feb. 5, 1977	0329:19	-66.49	-82.45	6.1	6.4	KR	1	1	A
131. Feb. 16, 1977	0049:31.7	25.96	-26.21	5.4		K	3	1	A
132. Aug. 26, 1977	1950:01.4	-59.43	-20.51	6.3	7.1	UK	3	2	B
133. Aug. 26, 1977	2238:59.2	-59.48	-20.17	4.8					
134. Aug. 29, 1977	1232:39.6	-59.50	-21.30	5.2	5.4				
135. Oct. 26, 1977	1601:44.7	-20.80	-126.82	4.9					
136. Oct. 29, 1977	1852:54.2	-20.81	-126.64	4.8					
137. Oct. 31, 1977	0819:14.5	-20.78	-126.67	5.1	4.2				
138. Nov. 5, 1977	0351:24.0	-20.81	-126.72	4.9					
139. Dec. 13, 1977	0114:18.6	17.36	-54.85	5.7	6.4	K	3	2	A
140. Dec. 15, 1977	1551:35.5	-2.46	85.75	5.0					
141. Jan. 4, 1978	1452:09.3	85.74	-23.83	5.0		K	1	2	C
142. Jan. 5, 1978	0323:27.9	-20.83	-126.87	5.5	4.1	K	1	1	E

TABLE 1. (continued)

Date	Origin Time	Latitude, °N	Longitude, °E	m_b^*	M_s	Focal Mechanism	Ratings†		
							I	II	III
143. Jan. 30, 1978	1834:20.9	-16.32	-127.15	5.0	3.9				
144. March 11, 1978	0332:11.1	-20.89	-126.87	4.9					
145. March 24, 1978	0042:38.1	29.85	-67.40	6.1	6.1	A	3	2	A
146. April 12, 1978	1915:26.5	-21.20	-126.15	4.8					
147. May 4, 1978	0544:47.2	29.93	-67.52	4.7					
148. May 21, 1978	1813:13.9	-21.65	87.55	5.0			1	3	D
149. June 3, 1978	1807:21.8	-16.31	92.93	5.5			3	2	B
150. July 13, 1978	1804:19.6	-20.74	-127.00	5.1	3.7				
151. July 25, 1978	0754:09.0	-20.74	-126.93	5.2					
152. July 27, 1978	0349:29.9	-21.05	-126.50	4.7					
153. Aug. 3, 1978	0110:25.4	-1.06	84.45	5.9			2	2	C
154. Oct. 18, 1978	0203:58.9	-20.85	-127.90	5.1					
155. Dec. 6, 1978	1328:39.5	17.30	-54.72	5.7					
156. Feb. 4, 1979	2125:44.4	42.20	-149.45	5.2			1	1	D
157. Feb. 5, 1979	1735:52.9	-20.75	-127.26	4.8					
158. Feb. 26, 1979	0631:53.2	-21.23	-127.26	5.1					
159. May 9, 1979	1835:36.8	21.14	-62.13	4.9					

Epical data source, by year: pre-1963, focal mechanism reference; June 1963 to December 1963, USCGS; January 1964 to July 1977, ISC; August 1977 to August 1979, NEIS. Sources of focal mechanism: A [Stewart, 1979]; B [Stein and Okal, 1978]; C [Sykes, 1970]; D [Molnar and Sykes, 1969]; E [Sykes and Sbar, 1974]; F [Weissel and Anderson, 1978]; G [Stein, 1978]; H [Mendiguren, 1971]; I [Stein, 1979]; J [Fitch, 1972]; K (this paper, Figure 2); L [Anderson et al., 1974]; M [Forsyth, 1973]; N [Kafka and Weidner, 1979]; O [Richardson and Solomon, 1977]; P [Unger and Ward, 1979]; R [Okal, 1980]; S [Mendiguren and Richter, 1978]; T [Wang et al., 1979]; U [Creavan et al., 1979].

*Body wave magnitudes m_b are from ISC; surface wave magnitudes M_s are from U.S. Geological Survey or focal mechanism reference.

†Ratings columns: I, association with zone of weakness; II, association with large bathymetric relief; III, quality of bathymetric and tectonic data; for columns I and II, 1 is unlikely, 2 is possible, and 3 is probable association; for column III, ratings range from A (high quality, i.e., good coverage and resolution) to E (poor quality).

stress field has been to compile a catalog of 159 oceanic intraplate earthquakes (Table 1). References to individual earthquakes will include the corresponding number from Table 1. The core of the catalog comes from the compilation of Sykes and Sbar [1974], from which oceanic events have been se-

lected. Their list covers the period from 1963 to 1971 and includes only events for which they were able to obtain focal mechanisms. Table 1 includes earthquakes for which focal mechanisms are not yet available, in the hope that application of newer, more powerful techniques (e.g., moment tensor inversion from surface waves, body wave synthesis) may permit characterization of source mechanisms where simple first motion studies have failed. The catalog of oceanic intraplate earthquakes in Table 1 was updated with events from listings of the International Seismological Centre (ISC) and National Earthquake Information Service (NEIS) and from published studies on seismicity in oceanic intraplate regions [e.g., Talandier and Kuster, 1976; Stein and Okal, 1978].

Classification of an earthquake as intraplate is made after judging that the tectonic processes of the nearest recognized plate boundary are not likely to be dominant in the epicentral area. Sykes and Sbar [1974] compared the type of focal mechanism to the age of the lithosphere in which the earthquake occurred and found a change in type of mechanism between seafloor ages of 10 and 20 million years. In this age range, normal faulting events associated with ridge processes give way to thrust type events, which are generally more characteristic of intraplate seismicity. Such a transition age corresponds to a distance of tens to hundreds of kilometers from the ridge axis, depending on spreading rate.

We have used this result as an empirical rule for deciding how close an intraplate event can be to a spreading ridge. The Sykes and Sbar [1974] summary includes events which are quite close to ridges and have normal faulting mechanisms; such events are not included in this catalog. On the other hand, we have included an earthquake which is only 165 km from the East Pacific Rise (event 64) because its focal mechanism involves predominantly thrust faulting.

The intraplate stress field seaward of active subduction zones may be dominated by plate-bending stresses. Earth-

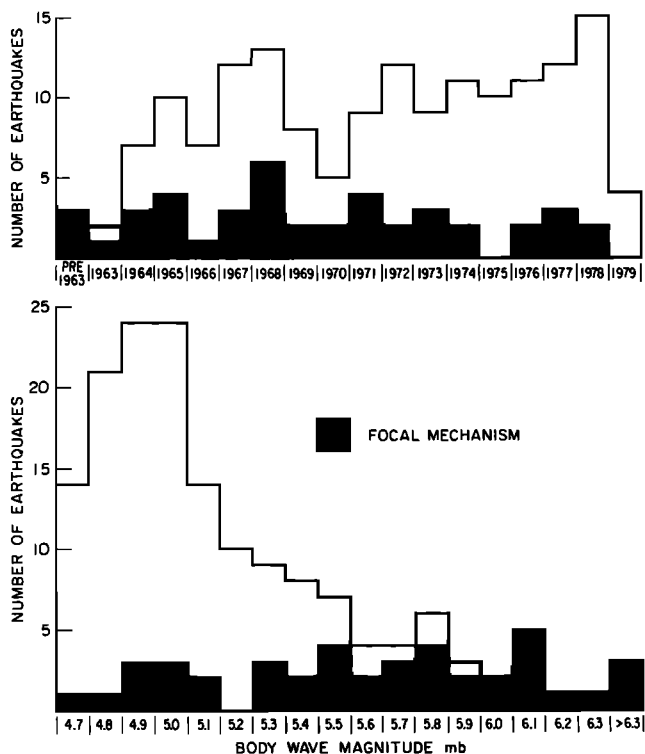


Fig. 1. Distribution of earthquakes in Table 1 by year of occurrence and by body wave magnitude m_b . Shaded portions of the histograms indicate the number of events with known focal mechanisms.

TABLE 2. Focal Mechanism Parameters for Events in Figure 2

Event Number	Date	Poles of Nodal Planes		<i>P</i> Axis Az/Pl	<i>T</i> Axis Az/Pl	Mechanism Type	Data Source
		Az/Pl	Az/Pl				
13	March 6, 1965						A
48	Aug. 20, 1968	68/26	306/48	272/12	22/56	T	A
64	Jan. 21, 1970	252/37				T	A
79	May 21, 1972						A
93	May 3, 1973	80/20				N	A
95	Aug. 30, 1973	152/42	281/36	126/5	226/60	T	A
100	April 12, 1974	94/0	184/14	135/8	228/8	SS	AB
126	Nov. 24, 1976						B
128	Dec. 28, 1976						B
130	Feb. 5, 1977					T	A
131	Feb. 16, 1977						B
132	Aug. 26, 1977						B
139	Dec. 13, 1977	43/36	146/18	359/12	106/42	T-SS	A
141	Jan. 4, 1978	132/24				T?	B
142	Jan. 5, 1978					T?	B

Az, azimuth; Pl, plunge; T, thrust; N, normal; SS, strike-slip; A, first motions picked by EAB; B, first motions reported in NEIS seismic bulletin.

quakes in this stress regime are characterized by subhorizontal *P* or *T* axes which are perpendicular to the trench axis [Chapple and Forsyth, 1979]. Most such events are within 25 km of the trench axis, but some are as far as 100 km distant. Earthquakes within 100 km of a trench are therefore excluded unless there is a focal mechanism which indicates a stress regime distinct from that expected due to plate bending. One example is event 50, which is only 60 km north of the Puerto Rico trench. The focal mechanism shows a stress direction quite different from that expected of a bending plate; Molnar and Sykes [1969] offer further evidence that the Puerto Rico trench is not behaving as a typical subduction zone at this time.

Earthquakes beneath continental shelves are excluded from the catalog because there is evidence that their mechanisms may be strongly controlled by sediment loading [Yang and Aggarwal, 1979] or, at high latitudes, by deglaciation [Stein et al., 1979].

The smallest oceanic intraplate earthquake with a published focal mechanism is a Caribbean event (81) with body wave magnitude m_b 4.7 [Kafka and Weidner, 1979]. Although focal mechanisms for teleseisms of this magnitude are rarely attempted, m_b 4.7 has been chosen as the lower limit for earthquakes of potential use in studying the tectonic stress field.

Earthquakes without published focal mechanisms are restricted in the catalog to the period since 1963, during which the World Wide Standard Seismograph Network (WWSSN) has been operating. Focal mechanisms are quite difficult to determine without this standardized and global data set and are generally subject to even greater uncertainty than usual. There are, nevertheless, several reliable mechanisms published for older intraplate events, and these earthquakes are included in Table 1. The distribution of events in Table 1 by year and magnitude is shown in Figure 1.

First-motion studies were initiated for many events if no published mechanisms were available. Several new mechanisms are presented here (Table 2 and Figure 2), but many of the events are too small, or coverage too sparse, to allow a mechanism to be well constrained by first-motion data alone.

No claim is made that the catalog in Table 1 is exhaustive. It does, however, contain the large majority of oceanic intraplate earthquakes since 1963 for which there is some hope of

obtaining a focal mechanism and of thereby constraining the oceanic intraplate stress field.

SOME CHARACTERISTICS OF OCEANIC INTRAPLATE SEISMICITY

The epicenters of the earthquakes in Table 1 are shown in Figure 3. Oceanic intraplate earthquakes are rather well dispersed, considering the difficulty of detecting and locating small events in remote oceanic areas [Evernden, 1969]. There are areas with notably higher and lower concentrations, however, such as the central Indian Ocean Basin and the South Atlantic, respectively.

A simple quantitative method for studying the distribution of oceanic intraplate seismicity is to calculate the number of earthquakes per unit area per time for the fraction of each plate consisting of oceanic lithosphere. The rate of seismicity is traditionally calculated by fitting the earthquake catalog to an equation of the form

$$\log N(M) = a - bM$$

where N is the number of events with magnitude equal to or larger than M per unit time and area. The total numbers of intraplate earthquakes of $m_b \geq 4.7$ for the oceanic portions of each plate for the period 1963 to present are given in Table 3. For the whole catalog and for the two plates with the largest number of events (Pacific and Indian), a and b values are calculated by using the above equation with $M = m_b$ and with 17 years and $40.7 \times 10^6 \text{ km}^2$ as the units of time and area, respectively. The Indian plate is seen to be more seismically active by a factor of ~ 4 for the m_b range 5 to 6 than average oceanic lithosphere. The high b value for the Pacific plate (1.3 versus 0.9 for average oceanic lithosphere) may reflect an unusually large contribution from events associated with volcanic activity.

A low level of intraplate seismicity may be explained in some cases by a relatively high detection threshold, but alternative causes should be considered. For example, certain geometries of plates and applied forces may result in smaller deviatoric stresses and lower seismicity within the plates. Sykes [1978] suggested this mechanism to explain the apparently low seismicity of the Antarctic plate, which is entirely surrounded by spreading ridges.

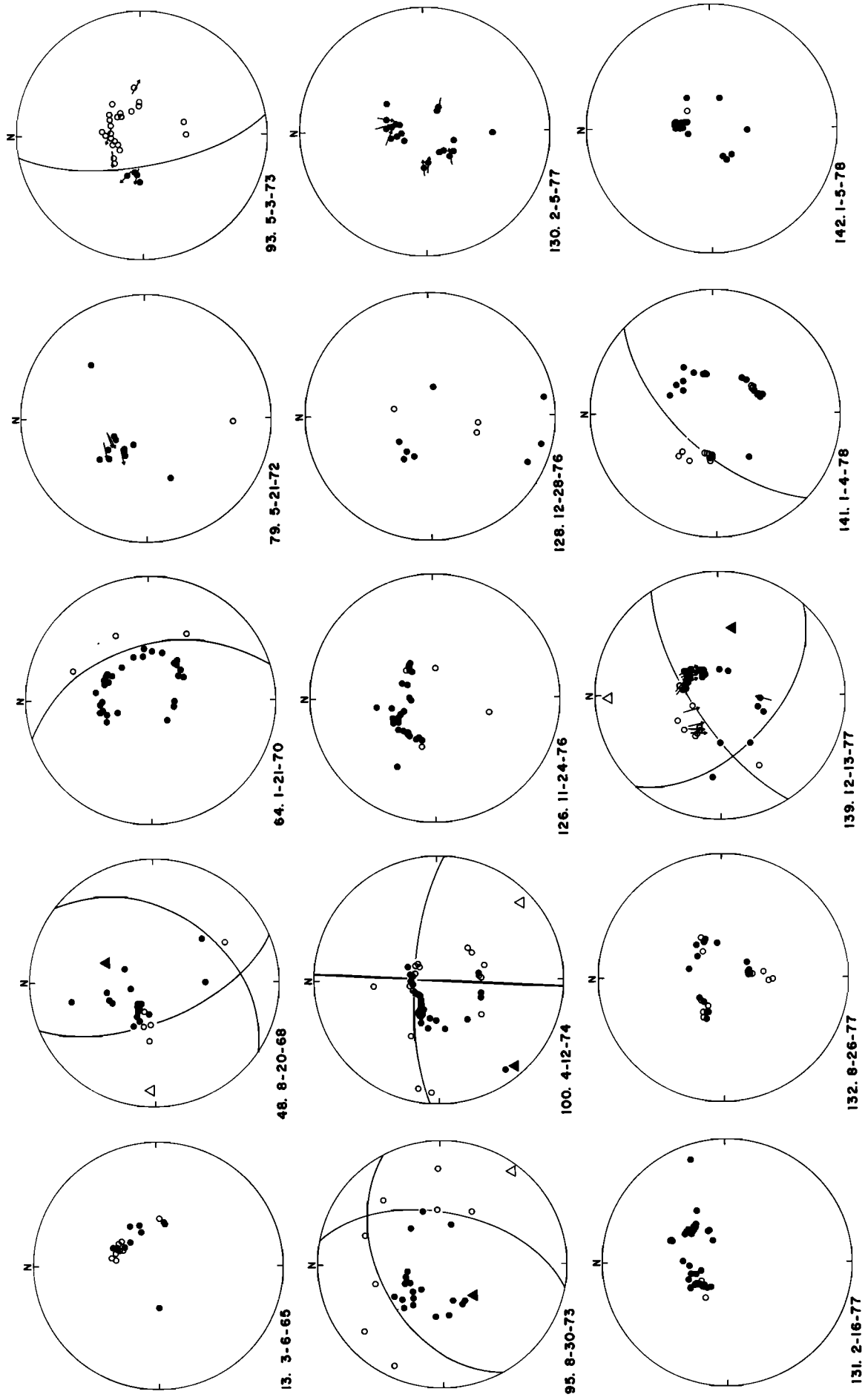


Fig. 2. New focal mechanisms and first-motion data for 15 oceanic intraplate events, plotted on the lower focal hemisphere with an equal area projection. See Table 2 for focal mechanism parameters and Table 1 for epicentral data. Solid circles and triangles are compressional first motions and *T* axes, respectively. Open circles and triangles are dilatational first motions and *P* axes, respectively. Arrows give shear wave polarizations.

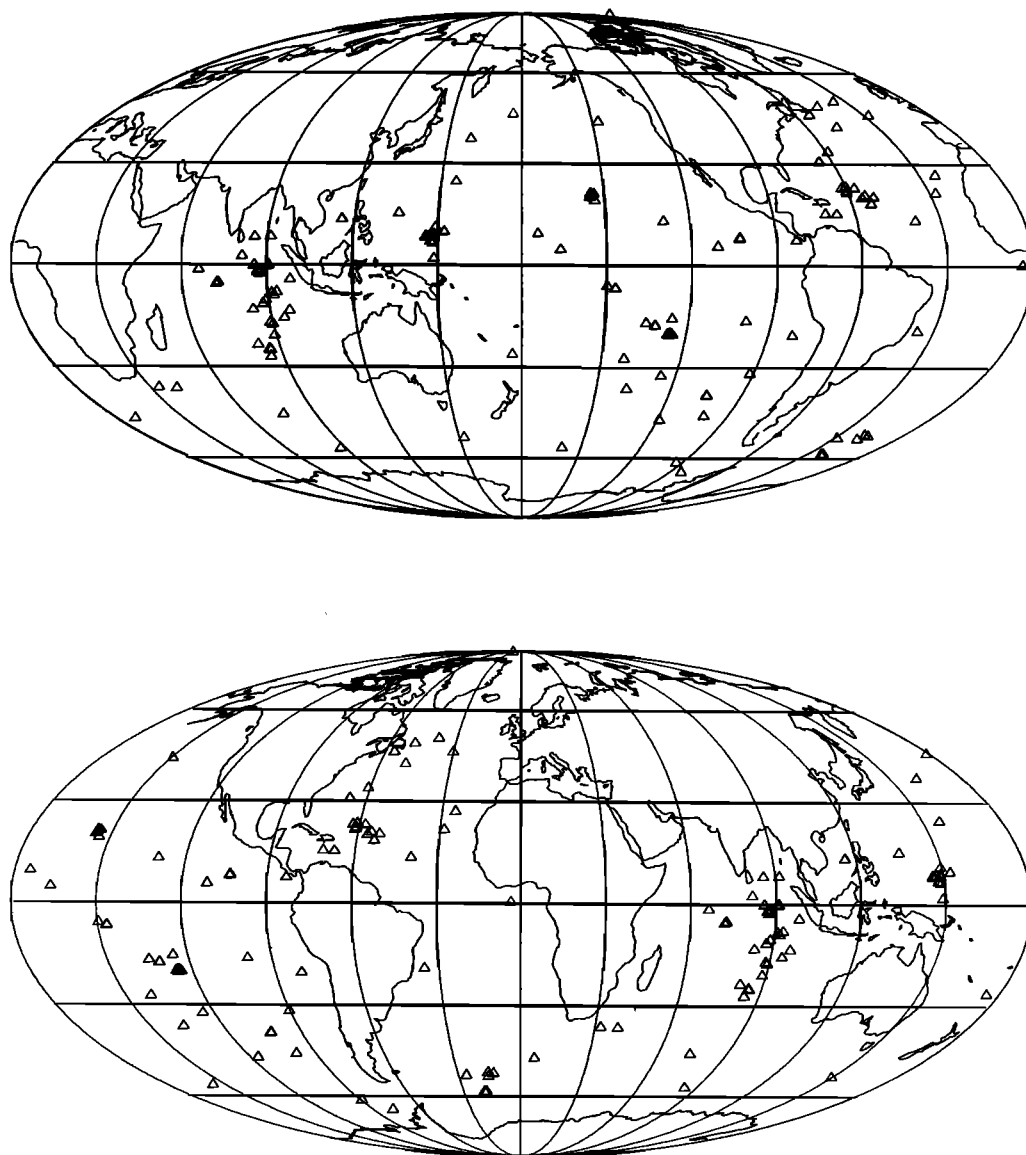


Fig. 3. Global distribution of epicenters listed in Table 1, on Mollweide elliptic equal area projections centered at 0° latitude and at 0° and 180° longitude.

In continental lithosphere, *Sbar and Sykes* [1973] and *Sykes* [1978] found a frequent association of intraplate seismicity with old zones of weakness which have been selectively reactivated by the present stress regime. Those weak zones which are not reactivated may be oriented incorrectly or may have healed in some manner, so that they are no longer preferred sites of strain release. Presumably, similar mechanisms may operate in oceanic lithosphere to influence the level of seismicity.

There are several explanations for a high level of intraplate seismicity. In some cases it clearly represents ongoing large-scale intraplate deformation. The Ninetyeast Ridge area, which has a history of seismicity as intense as many plate boundaries, is a prime example [*Stein and Okal*, 1978]. The focal mechanisms and inferences about the tectonic stress field in this region will be discussed in a later section.

High intraplate seismicity may also reveal the presence of a previously undefined plate boundary, although the transition from intraplate to interplate deformation may be ill defined. The seismicity of the Caroline Basin in the western Pacific

suggested this possibility to *Katsumata and Sykes* [1969], a view supported by more recent geophysical data [*Weissel and Anderson*, 1978; *Mammerickx*, 1978]. Strictly speaking, these events may not represent the intraplate stress field, but knowledge of the formation of new plate boundaries and relative motions across them may be equally useful for studying plate driving forces.

Another source of high intraplate seismicity is magmatism, on several scales. If plate interiors are exposed to upwelling of hot mantle material (hotspots), thermal stresses and bending stresses associated with uplift and volcanic loading may be locally larger than the regional tectonic stress field. The motions of magma in conduit systems and actual eruptions are also the source of frequent localized seismicity. Both processes are active among the linear volcanic island chains of the Pacific.

Earthquake focal mechanisms in regions of volcanism or topographic loading may not be able to constrain plate driving forces if the events are not responding primarily to the tectonic stress field corresponding to those forces, but they may be useful in estimating the absolute level of stress in the litho-

TABLE 3. Rates of Occurrence of Oceanic Intraplate Earthquakes by Plate, 1963 to Present

Plate	Area*	Number of Events	Events/ Area	a †	b †	m_b Range
Full Catalog	7.58	156	20.6	5.6 ± 0.1	0.91 ± 0.02	4.8–5.8
Pacific	2.62	57	21.8	7.6 ± 0.5	1.32 ± 0.09	4.8–6.1
African	1.06	8	7.5			
Antarctic	1.04	9	8.7			
Indian	0.97	45	46.4	6.2 ± 0.3	0.94 ± 0.06	4.9–5.9
N. American	0.54	19	35.2			
S. American	0.47	7	14.9			
Nazca	0.41	3	7.3			
Eurasian	0.21	3	14.3			
Philippine	0.14	1	7.1			
Cocos	0.08	2	25.0			
Caribbean	0.05	2	40.0			

*One unit of area equals $R_E^2 = 40.7 \times 10^6 \text{ km}^2$, where R_E is the mean earth radius.

†Errors shown are the 95% confidence interval of coefficients calculated for the least squares fit to the equation $\log N = a - bm_b$, where N is the number of events per unit area with magnitude $\geq m_b$ and one unit of area is R_E^2 , as above.

sphere if quantifiable local sources of stress and the regional stress field can be shown to be of comparable magnitude [Solomon *et al.*, 1980]. A comprehensive study of focal mechanisms is necessary to determine whether there is any regional consistency in the stress pattern in the vicinity of islands, seamounts, and other seafloor topography [Sverdrup *et al.*, 1979]. These areas may also yield information on the tectonic stress field through the geometry of volcanic constructs [Jackson and Shaw, 1975] and of volcanic fissures [Nakamura *et al.*, 1977].

There are several cases of earthquake swarms in Table 1, including the Chagos Bank earthquakes of November 1967 (events 35–41) [Stein, 1978]. Another is the series of events northeast of the Gambier Islands (21°S , 127°W), beginning in August 1976, which includes 13 events large enough to be included in Table 1. Such swarm activity can sometimes be correlated with submarine volcanism [Talandier and Kuster, 1976].

Because of the generally small size of intraplate earthquakes and the detection threshold of the global seismic network, few aftershocks are recorded. There are 12 events in the catalog (all with m_b of 5.4 or larger) which have recorded aftershocks: 6, 10, 17, 20, 46, 64, 68, 92, 98, 101, 132, and 145.

Study of repeated seismicity in the South Pacific is made more difficult by the French nuclear testing program and by the occasional lack of identification of French nuclear tests in standard earthquake catalogs. We have been careful to exclude nuclear tests from the catalog in Table 1, using Bolt [1976] and several other lists of French tests. One series of events in the catalog, however, is questionable (events 13, 27, and 113). These closely spaced events are listed as earthquakes by Talandier [1978] and Talandier and Kuster [1976], along with many smaller events with the same epicenter. Bolt [1976], however, lists event 27 as an airblast at the island of Mururoa. A first-motion study of event 13 proved inconclusive; there do appear to be both compressional and dilatational arrivals, but the record quality is quite poor (Figure 2).

Depths of oceanic intraplate earthquakes are rarely determined with any accuracy because of the lack of nearby stations. Those events for which some effort has been made to constrain the focal depth are listed in Table 4. Nearly all such earthquakes have been found to be shallow, less than 20-km

depth, but the evidence is strong that two events (93 and 130) occurred at depths of 48 and 35 km, respectively. The mantle-depth Hawaiian event (93) may be a result of plate-bending stresses [Unger and Ward, 1979], but event 130 appears to be free of any such influence [Okal, 1980].

For those events in the catalog with known seismic moment M_0 and with $M_S \geq 6.0$, apparent stress $\eta\bar{\sigma}$ was calculated by using the formula [Aki, 1966]

$$\eta\bar{\sigma} = \frac{\mu}{M_0} \exp(27.2 + 3.45 M_S)$$

where μ is the rigidity, assumed to be $3 \times 10^{11} \text{ dyn/cm}^2$. The apparent stress provides a lower bound on the level of tectonic stress in the lithosphere. The calculated apparent stresses (Table 5) lie principally in the range 2 to 15 bars; only event 101 yields a value outside this range (60 bars). Thus these results support the conclusion [Richardson and Solomon, 1977] that a distinction in the magnitude of apparent stress between intraplate earthquakes and plate boundary events [Kanamori and Anderson, 1975] is not general.

CHARACTERIZATION BY EPICENTRAL REGION

From Table 1 a subset of 83 earthquakes has been selected with which to study the bathymetric and tectonic setting of oceanic intraplate events. After choosing events from the catalog so that each site of oceanic intraplate seismicity is represented by one event, we estimate the degree of influence of local bathymetric relief and of zones of weakness or intraplate deformation on the tectonic stress field orientation which would be inferred from a known focal mechanism at that location.

Every isolated intraplate event in Table 1 is included, whether or not it has a known focal mechanism. For areas with repeated seismicity or aftershocks, the site is represented by the largest catalogued event with a focal mechanism or by the largest event if no focal mechanisms have been published. By generating a subcatalog of oceanic intraplate epicentral regions in this way, we gain an appreciation for the distribution and quality of all present 'data points' which are available to constrain the oceanic intraplate stress field. The representation of areas of repeated seismicity by a single event in the subcatalog should not obscure the desirability of studying all

TABLE 4. Focal Depths of Oceanic Intraplate Earthquakes

Event Number	Date	Depth (Below Seafloor), km	Reference
1	March 21, 1939	<10	<i>Stein and Okal [1978]</i>
18	Sept. 12, 1965	<15	<i>Stein [1978]</i>
20	Oct. 7, 1965	5	<i>Wang et al. [1979]</i>
22	Nov. 25, 1965	9	<i>Mendiguren [1971]</i>
46	April 28, 1968	<16	<i>Stein [1979]</i>
81	Aug. 14, 1972	20	<i>Kafka and Weidner [1979]</i>
93	April 26, 1973	48	<i>Unger and Ward [1979]</i>
130	Feb. 5, 1977	35	<i>Okal [1980]</i>
145	March 24, 1978	11	<i>Stewart [1979]</i>

events in each series in order to test for consistency of inferred stress orientation.

For each event in the subcatalog the published bathymetric data and tectonic history of the epicentral region have been assembled (appendix). Bathymetric data are used primarily to assess the presence of large relief near the epicenter, which could indicate either a source of large local stress or a site of intraplate deformation. For example, large volcanic loads lead to lithospheric flexure and bending stresses which may exceed a kilobar in the case of Hawaii [e.g., *Walcott, 1970; Watts and Cochran, 1974*] or for some models of seamount loading [e.g., *Cazenave and Lago, 1980*]. Even if a large bathymetric feature is isostatically compensated, lateral variations in crustal thickness may lead to horizontal stresses of hundreds of bars [*Artyushkov, 1973*]. Fracture zones, the most commonly found zones of weakness in oceanic lithosphere, are also frequently located by their bathymetric expression. Tectonic syntheses provide a context in which to study the relationship of each site to large-scale tectonic processes in the plates. Such studies frequently delineate fracture zones from magnetic anomaly patterns in areas with poorly defined bathymetry or heavy sedimentation.

Each event is rated twice on a scale from 1 to 3, according to its association with (1) bathymetric relief which might be a source of local stress and (2) zones of previous weakness (e.g., fracture zones, old spreading centers, or sites of intraplate deformation). For both categories the ratings 1, 2, and 3 correspond to 'unlikely,' 'possible,' and 'probable' association. The quality of the available bathymetric and tectonic data, which varies widely in oceanic intraplate areas, is indicated by a letter grade (from A, for high quality, to E). A focal mechanism which is clearly consistent with bathymetric data or tectonic interpretations is reflected in a higher quality rating.

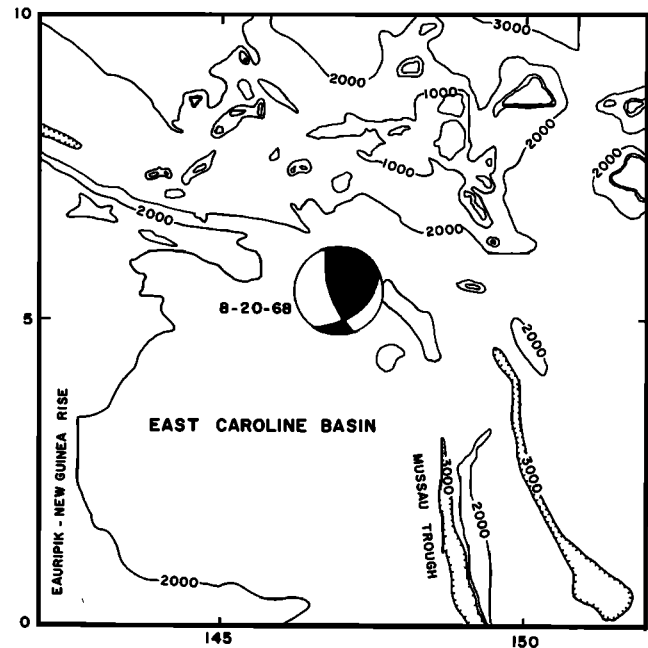


Fig. 4. Focal mechanism and epicentral setting of event 48 (August 20, 1968) in the Caroline Basin. The compressional quadrant is shaded. Bathymetry is in meters, adapted from *Chase et al. [1970]*.

In Table 1, events chosen for the subcatalog are denoted by an entry under the 'ratings' heading. Each event in the subcatalog is briefly discussed in the appendix, giving the basis for its ratings and references for bathymetric and tectonic data. Maps of the epicentral areas of eight events for which new mechanisms are presented (48, 64, 93, 95, 100, 130, 139, 141) are shown in Figures 4 through 11.

WEAK ZONES AND LOCAL SOURCES OF STRESS

A histogram of the ratings for all 83 events in the subcatalog is shown in Figure 12. The figure demonstrates that oceanic intraplate earthquakes show a marked tendency for association with zones of weakness and a lesser tendency, if any, for association with large bathymetric relief.

These ratings may be biased by the effect of poor data quality on the two classifications. The recognition of a fracture zone or other zone of weakness is much more dependent on the availability of detailed bathymetric and tectonic studies than is the identification of a large bathymetric feature. This bias can be verified by performing chi-squared tests for the dependence of the ratings on data quality, summarized as 'good' (A, B, C) and 'poor' (D, E). Table 6 shows the data nec-

TABLE 5. Seismic Moments and Apparent Stresses for Oceanic Intraplate Earthquakes

Event Number	Date	M_0 , 10^{24} dyn cm	Reference	M_S	Reference	Apparent Stress, bars
1	March 21, 1939	2000	A	7.2	A	6.0
3	March 22, 1955	400	A	7.0	A	15.
6	May 25, 1964	11	A	6.0	A	17.
10	Oct. 23, 1964	84	B	6.3	B	6.4
18	Sept. 12, 1965	35-68	C	6.0	C	2.8-5.4
67	Oct. 10, 1970	29-67	A, B	6.2	A	5.6-13.
71	May 9, 1971	90	E	6.0	D	2.1
93	April 26, 1973	64	B	6.1	D	4.2
101	June 25, 1974	25	A	6.6	A	60.
130	Feb. 5, 1977	44	F	6.2	F	8.6

References: A, *Stein and Okal [1978]*; B, *Richardson and Solomon [1977]*; C, *Stein [1978]*; D, U.S. Geological Survey; E, *Forsyth [1973]*; F, *Okal [1980]*.

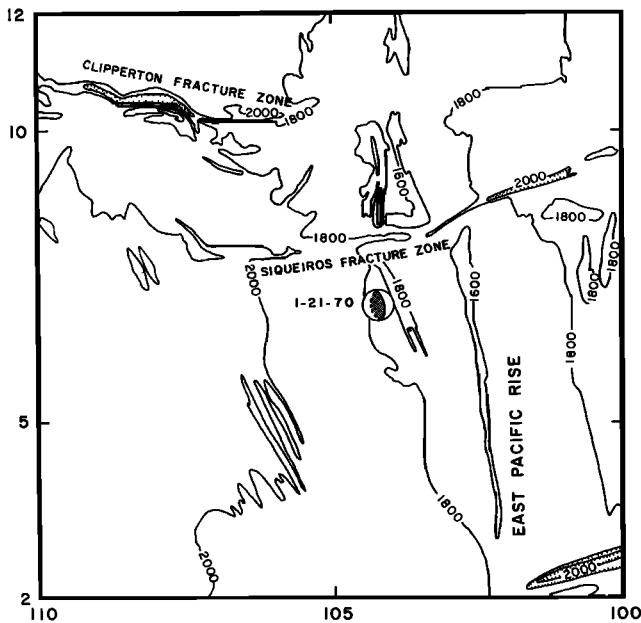


Fig. 5. Focal mechanism and epicentral area of event 64 (January 21, 1970) in the eastern Pacific. Bathymetry is in meters, adapted from Chase *et al.* [1970]. The dotted nodal plane is not constrained.

essary for computing the parameter $\chi^2 = \sum (\text{OBS} - \text{EXP})^2 / \text{EXP}$, where 'OBS' is the observed number of earthquakes with a given rating and quality and 'EXP' is the expected number of events if data quality has no effect on the distribution of ratings, so that the distribution among the three rating categories for each quality group is the same as that for the whole population.

For case A in Table 6 a χ^2 value of 4.6 indicates that at the 90% confidence level, data quality has no effect on the distribution of ratings for association with large bathymetric relief. For case B, however, a χ^2 value of 26.7 means that at greater than 99% confidence the rating distribution for association of events with weak zones is affected by the quality ratings (for

example, $\chi_{0.995}^2 = 10.6$). From Table 6 we can see that most of the contribution to χ^2 comes from the 11 poor quality events which were apparently not associated with a zone of weakness, where only 3 or 4 are expected.

A more accurate characterization of the earthquakes can therefore be obtained by removing the events with poorly characterized epicentral regions. This has been done in the histogram shown in Figure 13, where we see a clear association of oceanic intraplate earthquakes with zones of previous weakness. Correlation of earthquake epicenters with large bathymetric relief is still demonstrably less.

There are, however, a significant number of events with epicenters near probable sources of large local stress, increasing the chance of estimating the absolute level of stress in the lithosphere by the method proposed by Solomon *et al.* [1980]. If the stresses from these bathymetric features can be estimated and the focal mechanisms can be shown to be controlled by those stresses, an upper bound on the magnitude of the regional stress field can be obtained. Much more work on focal mechanisms and depths of these events is required before this approach can be widely applied, however. This endeavor may also be complicated by the fact that none of the events which are associated with large relief are clearly free of the influence of a zone of weakness (see the two-way distribution matrix in Figure 13). Since many large bathymetric features are related to hotspots or volcanic extrusion on old zones of weakness, there will be few cases when zones of weakness can be ignored.

There are 10 oceanic intraplate earthquakes (events 7, 10, 25, 34, 50, 61, 85, 123, 125, 130) which, on the basis of well-characterized local bathymetry, are not associated with a zone of weakness or large relief. An additional eight earthquakes (events 78, 80, 107, 109, 118, 121, 142, 156) also appear to be free from the effects of local structure on the basis of current data, but the bathymetric control is poor. These earthquakes deserve careful source mechanism analysis by the widest possible variety of seismological tools. A total of 10 to 20 events, however, is too small for the long-wavelength tectonic stress

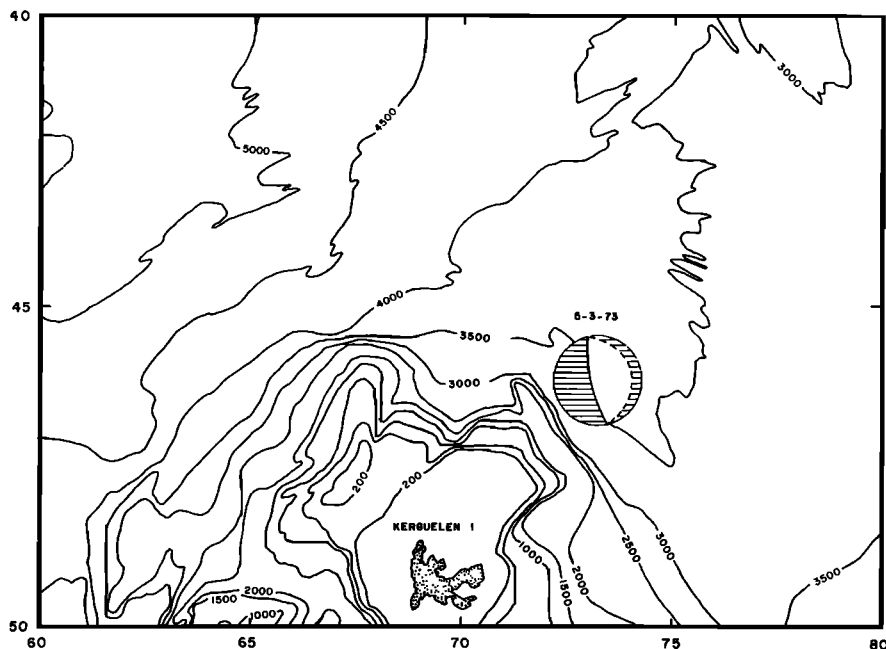


Fig. 6. Focal mechanism and epicentral setting of event 93 (May 3, 1973) near Kerguelen Plateau (South Indian Ocean). Bathymetry is in meters, adapted from Udintsev *et al.* [1975]. The dotted nodal plane is not constrained.

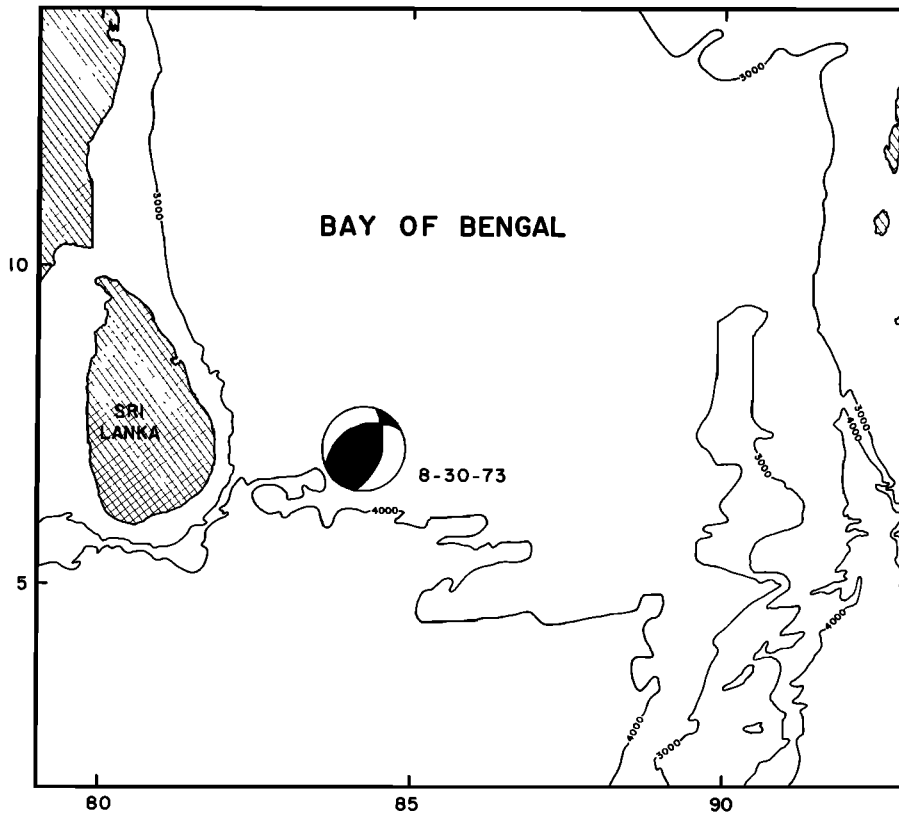


Fig. 7. Focal mechanism and epicentral area of event 95 (August 30, 1973) in the Bay of Bengal. Bathymetry is in meters, adapted from *Defense Mapping Agency* [1976].

field in the oceans to be characterized on a global basis using focal mechanisms from only these earthquakes. In particular, earthquakes on or near preexisting weak zones will have to be considered in order to have an adequate spatial distribution of events. It is therefore crucial to determine whether any useful information on the tectonic stress field orientation can be ob-

tained from focal mechanisms of events localized on weak zones.

OLD ZONES OF WEAKNESS AND THE INTRAPLATE STRESS FIELD

Stein [1979] has presented a good summary of the potential problems of using focal mechanisms of earthquakes associated with major intraplate deformation to constrain the oceanic intraplate stress field. In the presence of a preexisting fault, the orientation of the greatest principal stress is only required to lie in the dilatational quadrant of the focal mechanism [*McKenzie*, 1969].

In spite of these arguments, there are several well-documented cases where intraplate focal mechanisms over a broad region have *P* axis orientations that are consistent with each other and, in continental areas, with other data on the stress field. Two well-known examples of large-scale regional consistency are western Europe (north of the Alps) and eastern North America between the Rocky Mountains and the Appalachian Mountains [*Sbar and Sykes*, 1973; *Sykes and Sbar*, 1974]. The regional consistency of the North American stress orientation data breaks down in some areas [*Zoback and Zoback*, 1980] but remains over broad regions.

The only oceanic area with enough focal mechanisms presently available to look for regional consistency is the central Indian Ocean. Some of these epicenters are associated with the most prominent example of oceanic intraplate deformation known: the Ninetyeast Ridge [*Stein and Okal*, 1978]. In Figure 14, all known focal mechanisms in the central Indian Ocean are shown, with the horizontal projection of the *P* axis indicated. *Sykes and Sbar* [1974] published the first motion data for event 32 but did not indicate nodal planes. We feel

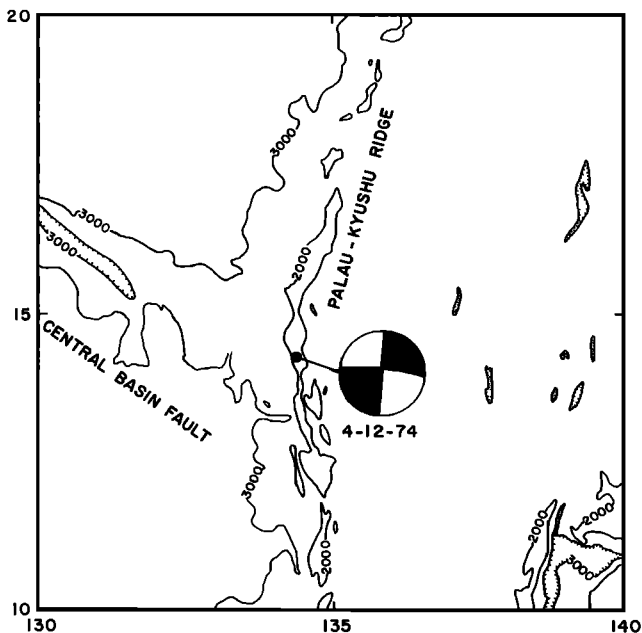


Fig. 8. Focal mechanism and epicentral setting of event 100 (April 12, 1974) in the Philippine Sea. Bathymetry is in meters, adapted from *Chase et al.* [1970].

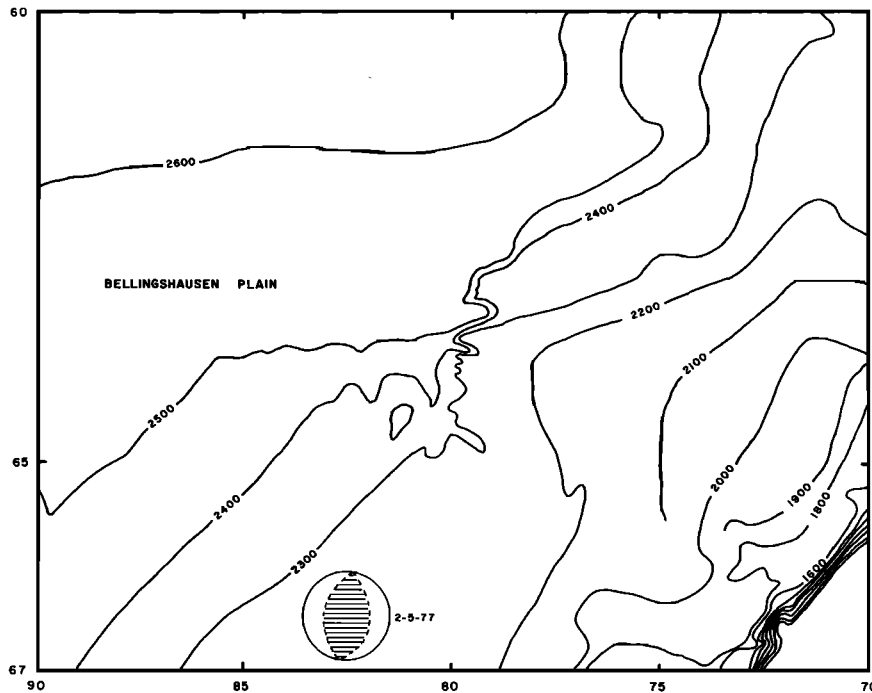


Fig. 9. Focal mechanism and epicentral area of event 130 (February 5, 1977) in the southeast Pacific Ocean. Bathymetry is in meters, adapted from *Mammerickx et al.* [1973]. Nodal planes are consistent with surface wave radiation pattern, as reported by *Okal* [1980].

there is evidence to suggest a predominantly thrust fault mechanism oriented as shown, but the mechanism is poorly constrained by the first-motion data alone. Event 3 has a weakly constrained normal faulting mechanism that is difficult to reconcile with any regional tectonic synthesis [*Stein and Okal*, 1978]. The *P* axis plunges at approximately 45° and is not plotted in Figure 14.

Fitch [1972] noted that focal mechanisms of four widely separated earthquakes suggested a regionally consistent state of stress in the central Indian Ocean, with axes of maximum and minimum compressive stress both nearly horizontal and

with the maximum compression axis oriented NW to NNW. Using a larger set of data, including in situ measurements, *Richardson et al.* [1979] concluded that the NW-SE orientation of maximum compressive stress in the Indian Ocean Basin may represent a transition between predominantly N-S compression in India and nearly E-W horizontal compression in Australia. They were able to reproduce such a stress field with models incorporating several simple driving forces on the global system of plates.

In their study of present-day plate motions, *Minster and Jordan* [1978] found that closure of the African-Indian-Antarctic triple junction required internal deformation of the Indian plate. If this deformation were occurring in the region of the Ninetyeast Ridge, NW-SE compressive stress would be predicted, in agreement with the focal mechanism data. *Weissel et al.* [1980] have recently found geological evidence for extensive deformation of the oceanic crust and sediments in the southern parts of the Bengal and Nicobar fans. The sense of the observed crustal deformation is consistent with the known stress data, leading these workers to postulate existence of a broad region of convergence between the Indian and Australian plates.

From a variety of lines of evidence, therefore, the focal mechanisms of the intraplate earthquakes in the central Indian Ocean appear to be a reliable indication of a long-wavelength feature of the lithospheric stress field. This conclusion holds despite the fact that several of the events are clearly associated with a zone of major intraplate deformation. This result may at first glance appear surprising in view of the theoretical calculations by *McKenzie* [1969], which might lead us to expect a nearly random scatter in the orientation of *P* axes for events on a major fault zone such as the Ninetyeast Ridge.

An explanation for this behavior may be found by considering an intraplate region large enough to contain many diversely oriented zones of weakness but with a relatively ho-

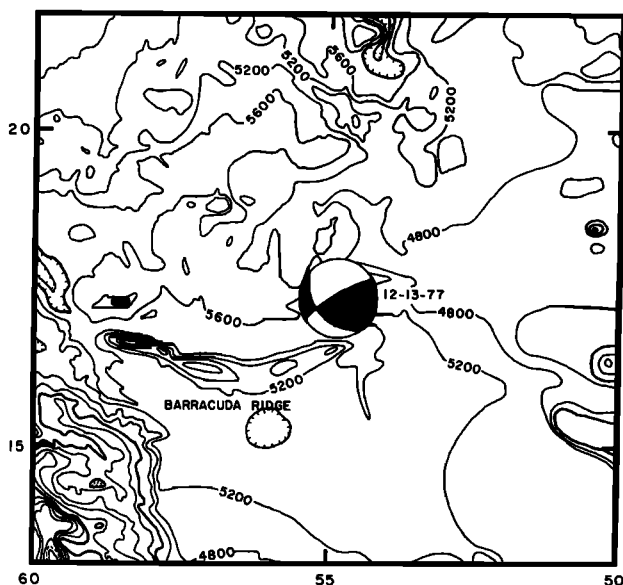


Fig. 10. Focal mechanism and epicentral setting of event 139 (December 13, 1977) in the equatorial Atlantic Ocean. Bathymetry is in meters, adapted from *Uchupi* [1971].

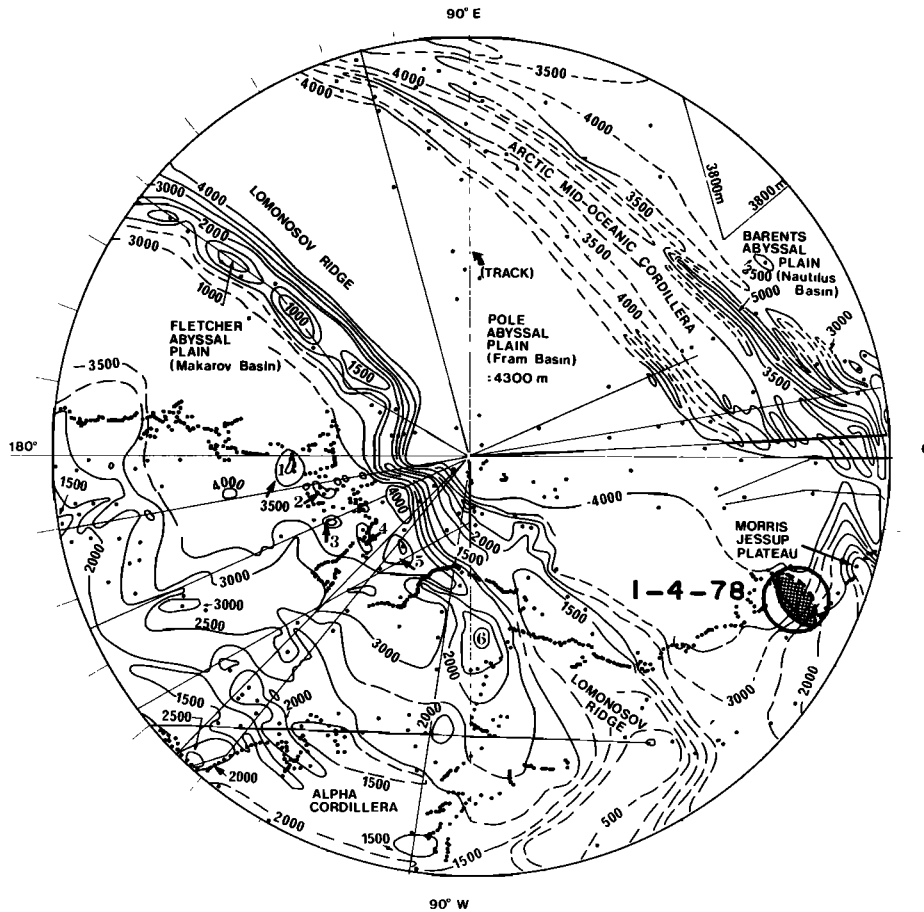


Fig. 11. Focal mechanism and epicentral area of event 141 (January 4, 1978) in the Arctic Ocean. Bathymetry is in meters from Sobczak [1977], showing individual depth stations and tracks of nuclear submarine crossings.

ogeneous tectonic stress field. If we require a particular fault to rupture under this stress regime, the *P* axis will in general have little consistent relation to the applied stress [McKenzie, 1969]. However, the faults which are most likely to rupture are

those oriented so that the shear stress resolved on the fault plane is maximized. When this occurs on a frictionless fault, the *P* axis is aligned with the axis of maximum compressive stress, 45° from the slip vector [McKenzie, 1969].

When friction on the fault is included, slip on the fault occurs most readily when the axis of maximum compressive stress is rotated toward the slip vector by an amount dependent on the coefficient of friction. Laboratory studies show that the optimum angle between the axis of maximum compressive stress and the fault plane is generally about 30° [e.g., Byerlee, 1978]. Sbar and Sykes [1973] used this result to plot stress orientations from focal mechanisms when a choice of fault plane could be made. Raleigh et al. [1972] performed field studies which showed that stress directions estimated in this way are generally within 20° of the true stress directions for rupture on preexisting faults.

For the events in Figure 14 we have not attempted to choose fault planes and plot the stress orientation according to the above scheme, although there is strong evidence for such a choice for some events. Detailed bathymetric and tectonic data, aftershock locations for larger events, and surface wave studies are all potentially useful in this regard, and a major effort in further research on intraplate earthquakes should be aimed at determining the slip vectors of the focal mechanisms. To the extent that large intraplate areas duplicate the conditions postulated (many orientations of weak zones in a homogeneous stress field), knowledge of the slip vector will allow optimum resolution of the orientation of the intraplate stress field from focal mechanisms.

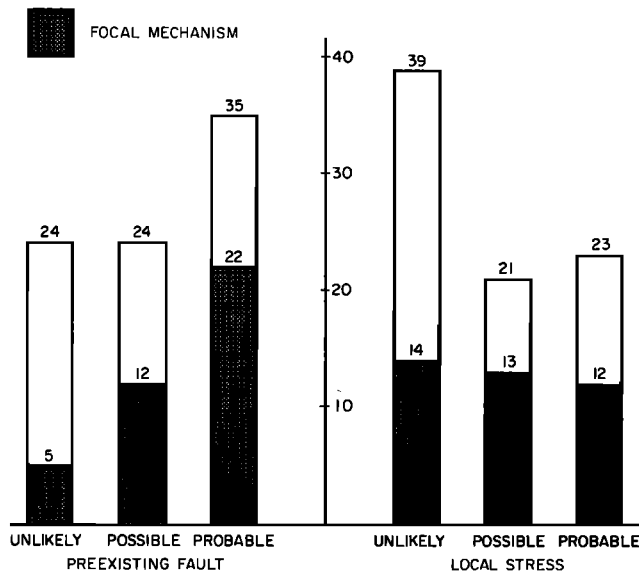


Fig. 12. Histogram of ratings for degree of association with zone of previous weakness and large bathymetric relief for 83 representative earthquakes from Table 1 (see text for method of selection). Shaded portion indicates number in each rating class with known focal mechanisms.

TABLE 6. χ^2 Tests for Dependence of Ratings on Data Quality

Ratings	Data Quality		Dependence
	Good (A, B, C)*	Poor (D, E)*	
<i>Association With Large Bathymetric Features†</i>			
1	30 (33.4)	9 (5.6)	unlikely
2	20 (18.0)	1 (3.0)	possible
3	21 (19.7)	2 (3.3)	probable
<i>Association With a Zone of Weakness‡</i>			
1	13 (20.5)	11 (3.5)	unlikely
2	23 (20.5)	1 (3.5)	possible
3	35 (30.0)	0 (5.0)	probable

See text for an explanation of ratings.

*The first number indicates the number of events in the category. The number in parentheses indicates the expected number of events in the category, assuming that data quality has no effect on the distribution of ratings.

† $\chi^2 = 4.6$ with 2 degrees of freedom.

‡ $\chi^2 = 26.7$ with 2 degrees of freedom.

Of the earthquakes which were categorized as having a close association with an old zone of weakness, relatively few are involved in large-scale intraplate deformation such as that near the Ninetyeast Ridge, Chagos Bank [Stein, 1978], or the Emperor Trough [Stein, 1979]. In most cases the event is found to lie near a fracture zone or a proposed fracture zone which has, at most, a mild bathymetric expression and no evidence of significant deformation.

This finding suggests that there may be a significant number of oceanic intraplate earthquakes which occur in regions of relatively small preexisting faults with a wide enough range of orientations so that slip occurs on that fault for which the drop in shear stress is approximately a maximum. The regional consistency of *P* axes in the central Indian Ocean is a

convincing argument for this mechanism. Thus focal mechanism information is likely to provide a strong constraint on local intraplate stress orientation for a large fraction of earthquakes which appear to occur on or near zones of previous weakness. It is obviously desirable to constrain the stress field orientation with as many focal mechanisms as possible, since the deviation of the *P* axis from the orientation of maximum stress for any single event could be substantial. This deviation can be reduced if a choice of fault plane can be made, as explained above.

CONCLUSIONS

A comprehensive catalog of 159 oceanic intraplate earthquakes has been compiled to identify virtually all events since 1963 whose focal mechanisms (if known) could contribute to knowledge of the state of stress in oceanic lithosphere. The bathymetry and tectonic setting for a subset of 83 representative event epicenters have been characterized to determine the degree to which oceanic intraplate earthquakes are associated with (1) large bathymetric features, which may be a source of large local stresses, and (2) preexisting faults.

The earthquakes in this subset, particularly the larger ones, are often associated with old zones of weakness which appear to be reactivated by the present stress field. Oceanic intraplate earthquakes are associated less often with large bathymetric features.

The regional consistency of inferred stress orientations for earthquakes in the vicinity of the Ninetyeast Ridge suggests that a mechanism is operating which allows focal mechanisms of events on preexisting faults to reflect the actual regional stress pattern more accurately than predicted by theory. One possibility is that the regional stress field preferentially reactivates those weak zones for which the resolved shear stresses

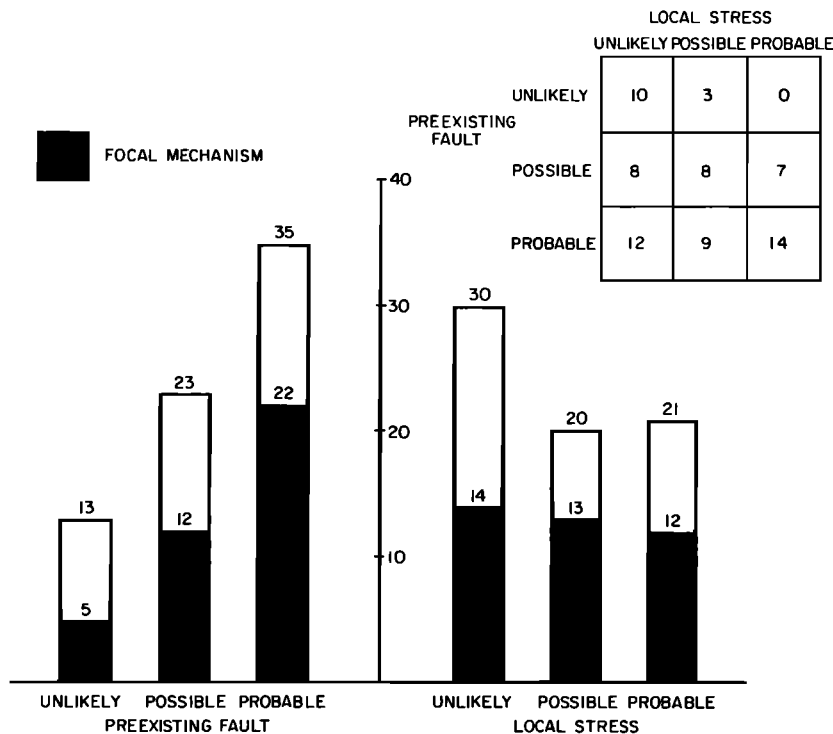


Fig. 13. Histogram of ratings of 83 representative earthquakes from Table 1 (see Figure 12) except for those with data quality ratings of D or E (12 events). Two-way table shows distribution of events among all combinations of ratings for association with zone of weakness and with large bathymetric relief. Shaded portions indicate number of events in each rating class with known focal mechanisms.

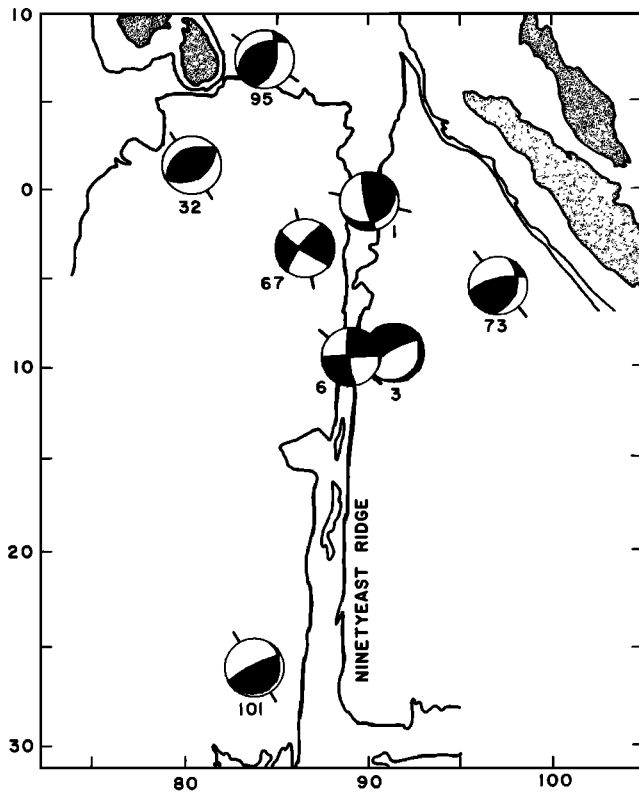


Fig. 14. All currently available focal mechanisms in the central Indian Ocean, with tick marks showing orientations of the horizontal projections of individual P axes. Mechanism of event 32 is poorly constrained. The P axis of event 3 plunges at about 45° .

are high. In such a case the direction of the observed P axis and the direction of maximum compressive stress tend to converge to within 20° – 30° . Knowledge of the fault plane of a focal mechanism allows a more accurate estimate of the local stress orientation, but the best indicator of stress field orientation is a consistent direction inferred from a number of events in a large area. Demonstration of such a regional consistency provides a powerful argument that the orientation of the regional stress field can be characterized and can be used as a constraint for models of tectonic forces acting on the lithosphere.

APPENDIX: RATINGS OF OCEANIC INTRAPLATE EPICENTRAL REGIONS

Synopses of bathymetry and tectonic setting are given for 83 representative earthquake epicentral regions, for events from Table 1 (see text for selection criteria). Each event is identified by number and by date. Parenthetical information for each event includes the rating for degree of association with a zone of previous weakness (1, unlikely; 2, possible; 3, probable association), the rating for association with large bathymetric relief, and the quality of bathymetric and tectonic data (A, high quality: good coverage and resolution; E, poor quality: virtually no data). Events are subdivided into broad geographic regions, within which they are presented chronologically.

Western Atlantic

10: October 23, 1964 (1-1-B). The epicenter lies northeast of the Lesser Antilles, in a region of poor bathymetry. One nodal plane of the mechanism [Molnar and Sykes, 1969] is

subparallel to a northwest-trending lineation evident in the bathymetry [Uchupi, 1971], but there are no corresponding magnetic anomaly offsets to indicate the presence of a fracture zone [Schouten and Klitgord, 1980]. The sense of motion of the focal mechanism (right lateral strike-slip on a fault oriented $N55^\circ W$) is consistent with relative motion between the two American plates predicted by recent global plate velocity models [Minster and Jordan, 1978]. There is no large bathymetric relief in the area. Event 11 is an aftershock of this earthquake.

25: March 20, 1966 (1-1-C). This event occurred in the Nares Abyssal Plain, where there is no major bathymetric relief discernible above the sediment cover [Uchupi, 1971]. The epicenter lies between two fracture zones inferred from magnetic anomaly offsets [Schouten and Klitgord, 1980].

34: July 10, 1967 (1-1-C). There is a bathymetrically defined lineation trending NW, which may be extrapolated through the epicentral area [Uchupi, 1971], but no corresponding magnetic anomaly offsets which would indicate a fracture zone [Schouten and Klitgord, 1980]. Bathymetric relief is subdued.

50: September 3, 1968 (1-1-C). The epicenter is about 60 km NE of the Puerto Rico trench. The mechanism [Sykes and Sbar, 1974] indicates a thrust fault with P axis oriented almost 90° from the direction expected if the event were a response to plate-bending stresses from the trench. All structural trends in the area strike NW [Vogt et al., 1971], rather than NE, as both fault planes do. Bathymetric relief is small [Uchupi, 1971]. Eleven years later, event 159 occurred very near this epicenter.

84: October 30, 1972 (3-1-B). This event lies in the Nares Abyssal Plain, on a fracture zone interpreted from offset magnetic anomalies [Schouten and Klitgord, 1980]. There is no bathymetric relief nearby [Uchupi, 1971].

126: November 24, 1976 (3-3-A). A first-motion study of this event was unable to constrain the mechanism (Figure 2). The epicenter is in an area of high relief and complex structure [Johnson and Vogt, 1971; Uchupi, 1971], on a northwest-trending fracture zone [Schouten and Klitgord, 1980].

128: December 28, 1976 (2-1-C). This event lies in the Nares Abyssal Plain, on the trace (extrapolated a short distance under the sediments) of a fracture zone [Schouten and Klitgord, 1980]. There is no large bathymetric relief in the area [Uchupi, 1971]. The fault plane solution is similarly unconstrained by available first-motion data (Figure 2).

139: December 13, 1977 (3-2-A). The epicenter lies off the northeast end of the Barracuda Ridge, between the Mid-Atlantic Ridge and the Lesser Antilles (Figure 10). The ridge was formed by normal faulting and subsequently experienced extensive horizontal shear, possibly associated with differential movement between the North American and South American plates [Birch, 1970]. The sense of motion of this shear has never been clear; the most recent model of present-day plate motions [Minster and Jordan, 1978] predicts east-west right lateral shear and a component of compression across the Barracuda Ridge area, but this type of motion is difficult to reconcile with the focal mechanism presented in Figures 2 and 10. The Barracuda Ridge is uncompensated [Birch, 1970], so topographic stresses may be significant in the epicentral region. This event was followed, 1 year later, by another earthquake in almost the same location and of equal magnitude (event 155).

145: March 24, 1978 (3-2-A). This earthquake occurred

370 km southwest of Bermuda in an area of low relief due to heavy sediment cover. Magnetic anomaly offsets may be extrapolated close to the epicenter [Schouten and Klitgord, 1980], indicating a fracture zone which trends subparallel to the nodal planes of the thrust fault mechanism [Stewart, 1979]. There may be topographic stresses from the Bermuda Platform, but the major bathymetric relief is well away from the epicenter. This earthquake had at least eight aftershocks, including one (event 147) large enough to be included in the catalog.

Caribbean

4: August 21, 1963 (2-3-A). This event occurred on the south-facing scarp of the Beata Ridge. There is a strong possibility of large local stress due to the bathymetric relief, but the strike-slip mechanism [Molnar and Sykes, 1969] does not easily fit this model. The epicenter lies approximately 100 km north of the convergent zone proposed by several authors [Jordan, 1975; Silver *et al.*, 1975]. The deformation associated with this convergence is apparently quite localized, however [Talwani *et al.*, 1977]. It has been suggested that the present Caribbean plate is unusually buoyant and has undergone extensive internal deformation because of its resistance to subduction [Burke *et al.*, 1978]. The observed sense of motion of this earthquake is exactly opposite that predicted by the modified Prandtl cell proposed by Burke and coauthors as a model for Caribbean intraplate deformation. Nevertheless, the event could be occurring along a line of weakness, associated with an earlier episode of internal deformation, in response to a different stress state.

81: August 14, 1972 (3-1-A). This earthquake is located in the Venezuelan Basin, well removed from large bathymetric relief and the convergent zone to the south [Jordan, 1975; Silver *et al.*, 1975]. The nodal planes of the strike-slip mechanism [Kafka and Weidner, 1979] are consistent with the presence of previously existing orthogonal fractures oriented NW and NE [Case and Holcombe, 1975; Burke *et al.*, 1978], but the left lateral sense of motion is opposite that predicted by the model of Burke *et al.* [1978]. This event and event 4 are consistent with a horizontal compressive stress field in the interior of the Caribbean plate oriented roughly WNW.

North Atlantic and Arctic

7: September 17, 1964 (1-1-B). The epicenter lies 300 km west of the Mid-Atlantic Ridge in an area of low bathymetric relief [Uchupi, 1971]. The epicenter is aligned with several other epicenters trending NW from the ridge [Laughton *et al.*, 1975], but there are no fracture zones revealed by magnetic anomaly patterns [Schouten and Klitgord, 1980]. Neither nodal plane of the thrust fault mechanism (strikes N7°W and N42°E) is subparallel with this trend [Sykes and Sbar, 1974].

19: September 21, 1965 (3-3-B). The epicenter lies very near a fracture zone on the south side of the Newfoundland Ridge [Schouten and Klitgord, 1980], on the continental rise off the Grand Banks [Uchupi, 1971]. The Grand Banks and Newfoundland Ridge could both be sources of local stress.

85: November 7, 1972 (1-1-C). The epicenter is about 400 km south of the Gibbs Fracture Zone, in an abyssal plain with no evidence for fracture zones or deformation from bathymetry [Uchupi, 1971]. There are very few data on magnetic lineations in this area [Vogt and Avery, 1974].

141: January 4, 1978 (1-2-C). The epicenter is in the Fram

Basin (Figure 11), near the Morris Jessup Plateau [Sobczak, 1977], a possible source of stress. There is evidence that this basin has experienced little deformation and has few fracture zones [Vogt *et al.*, 1979]. The fault plane solution indicates a predominantly thrust mechanism (Figures 2 and 11).

Eastern and Equatorial Atlantic

2: March 1, 1955 (3-3-B). The epicenter is on the continental rise of eastern Brazil, on the north flank of the Victoria-Trinidad Ridge, which has been related to a marginal fracture at the initiation of rifting in the South Atlantic [Uchupi, 1971; Francheteau and Le Pichon, 1972]. The thrust fault focal mechanism [Mendiguren and Richter, 1978] is poorly constrained, but one nodal plane could be parallel to the marginal ridge.

60: July 25, 1969 (3-1-A). This event lies several hundred kilometers east of the Mid-Atlantic Ridge, about 200 km north of the Vema Fracture Zone. Collette *et al.* [1974] identified a smaller fracture zone just north of the Vema which they extrapolate to the epicenter. The bathymetry [Uchupi, 1971] is consistent with this zone of weakness. The thrust fault solution of Sykes and Sbar [1974] is poorly constrained but could have a nodal plane parallel to this fracture zone.

76: September 30, 1971 (3-1-B). The epicenter lies in a region with numerous fracture zones and poorly understood tectonics [Sykes, 1978; Sibuet and Mascle, 1978]. There is no indication of large-scale deformation or topographic sources of stress, but a fracture zone can be traced very near the epicenter [Schouten and Klitgord, 1980]. One of the focal planes of the thrust fault mechanism [Sykes and Sbar, 1974] is subparallel with the local trend of fracture zones.

83: October 20, 1972 (2-1-C). The epicenter lies close to a proposed fracture zone, one of three which are proposed to have converged, because of a change in spreading direction of the Mid-Atlantic Ridge, to form the Cape Verde Archipelago [Dash *et al.*, 1976]. Magnetic anomaly offsets indicate a fracture zone very near the epicenter [Schouten and Klitgord, 1980]. There is no bathymetric evidence of the fracture zone or of local stress [Uchupi, 1971]. Neither nodal plane of the strike-slip focal mechanism [Richardson and Solomon, 1977] aligns with the proposed zone of weakness.

131: February 16, 1977 (3-1-A). This event lies in the Cape Verde-Madeira Abyssal Plain [Uchupi, 1971]. There is strong evidence for a fracture zone system trending WNW through the epicentral region [Harbison *et al.*, 1973; Rona and Fleming, 1973]. The epicenter is located on a topographic ridge which trends N80°E, oblique to the graben structures which characterize the fracture system. First-motion data (Figure 2) are inconclusive.

South Atlantic

78: April 22, 1972/107: September 9, 1974/109: February 21, 1975 (1-1-E). These three events occur on the west flank of the Mid-Atlantic Ridge, halfway between the ridge and the South Sandwich arc, an area of poorly known bathymetry and tectonics. There are no known offsets of the ridge at the latitudes of the events and no bathymetric evidence for large local stresses or deformation [Johnson *et al.*, 1973; Simpson, 1974].

108: November 20, 1974 (1-2-E). This epicenter lies 150 km north of the northern end of the South Sandwich trench, in an area of gentle relief which is largely unstudied. Tearing of the subducting slab from the remainder of the South Amer-

ican plate at the trench might produce large bending stresses in this region [Forsyth, 1975].

117: *October 23, 1975 (2-2-C)*. The epicenter lies between Bouvet Island and Southwest Africa, among a group of seamounts which might be a source of local stress [Johnson *et al.*, 1973; Simpson, 1974]. The Bouvet-Alfred Mertz-Meteor-Richardson seamount chain has been proposed as a hotspot track, but the epicenter is located approximately 75 km to the northwest of this trend.

132: *August 26, 1977 (3-2-B)*. This earthquake is located in a restricted intraplate area just east of the South Sandwich trench [Frankel and McCann, 1979]. The epicenter lies on the flank of a ridge which may be a source of local stress in addition to the nearby plate boundaries [Uchupi, 1971; Simpson, 1974]. The focal mechanism, nearly vertical strike-slip, and the E-W trending aftershocks suggest the event occurred on an extension of a transform fault from the nearby ridge [Creavan *et al.*, 1979]. This earthquake was followed by several aftershocks (four, according to Creavan *et al.* [1979]), two of which are in the catalog, events 133 and 134.

Central Indian Ocean Basin

1: *March 21, 1939 (3-3-A)*. This event occurred on the northern part of the Ninetyeast Ridge, which appears to be broken up into a series of en echelon blocks. The mechanism is consistent with strike-slip faulting either parallel to the ridge or to the en echelon structure, but in either case it indicates a NW-SE compressional stress regime [Stein and Okal, 1978]. There is large bathymetric relief [Sclater and Fisher, 1974]. Two small earthquakes (events 33 and 87) have occurred at this site since 1939.

3: *March 22, 1955 (2-2-B)*. The epicenter lies several hundred kilometers east of the Ninetyeast Ridge, in an area characterized by many small ridges and graben-like troughs trending roughly N-S [Bowin, 1973; Sclater and Fisher, 1974]. The mechanism is poorly constrained, but appears to represent normal faulting on a steeply dipping, NE-SW trending fault [Stein and Okal, 1978]. There is no bathymetric feature which correlates with this interpretation, but the location error may be rather large. There is no large relief in the immediate epicentral area.

6: *May 25, 1964 (3-3-A)*. This event occurred on the northern part of the Ninetyeast Ridge, south of the section characterized by large en echelon blocks [Sclater and Fisher, 1974]. One nodal plane of the purely strike-slip mechanism is parallel to the strike of the ridge [Sykes, 1970]. There is large bathymetric relief, and the ridge itself is certainly a weakened zone in the lithosphere. One other earthquake (event 29) occurred quite near this epicenter 2 years later.

32: *April 23, 1967 (2-2-B)*. This event occurred about 500 km south of Ceylon, under the Bengal Fan, a huge submarine sedimentary fan which may be a significant source of lithospheric stress [Curry and Moore, 1971]. Both nodal planes of the thrust fault mechanism [Sykes and Sbar, 1974] are almost perpendicular to a fracture zone which passes very near the epicenter [Sclater and Fisher, 1974]. The focal mechanism is not well constrained, however.

42: *February 9, 1968 (3-1-B)*. The epicenter is located on the Indira Fracture Zone of McKenzie and Sclater [1971], but Sclater and Fisher [1974] relocated the fracture zone about 100 km further west. The bathymetry shows a definite NE-SW trending structural pattern but no large relief [Sclater and Fisher, 1974].

51: *September 14, 1968 (3-1-B)*. The epicenter lies near where a fracture zone changes strike from NE-SW to N-S [Sclater and Fisher, 1974]. The bathymetry in this region is poorly characterized, but there appears to be no large relief. The epicenter is very near a narrow, deep, NNE-trending trough [Sclater and Fisher, 1974]. The focal mechanism shows thrusting, but the orientation of the focal planes cannot be determined [Sykes and Sbar, 1974].

55: *January 3, 1969 (3-3-B)*. This event occurred on one of the shallowest portions of the Ninetyeast Ridge, where it has a gentle, flat-topped expression [Sclater and Fisher, 1974]. Large local stresses are possible, as is a zone of preexisting weakness. A somewhat smaller event (57) followed this one by a month in the same area.

67: *October 10, 1970 (2-2-B)*. The focal mechanism of this event shows almost pure strike-slip faulting on nodal planes striking NW and NE [Fitch, 1972]. The epicenter is several hundred kilometers west of the Ninetyeast Ridge, in an area of heavy sediment cover which obscures any structure. Magnetic lineation offsets indicate a fracture zone parallel to the Ninetyeast Ridge which passes very near the epicenter [Sclater and Fisher, 1974], but neither nodal plane is consistent with a N-S trending fault. There is a moderately large plateau to the west which, along with the Ninetyeast Ridge to the east, may produce a local stress field with greatest compressive horizontal stress oriented E-W.

86: *December 3, 1972 (3-3-C)*. This event lies near the western margin of the Ninetyeast Ridge, near a small but very steeply sided seamount [Sclater and Fisher, 1974]. There is no bathymetric or magnetic evidence for a fracture zone on this side of the ridge, but an epicenter this close to such a major deformational feature must be considered to be in a region of lithospheric weakness. Event 114 occurred near this epicenter 3 years later.

95: *August 30, 1973 (2-3-B)*. This event is in the Bay of Bengal (Figure 7), under a thick sedimentary fan [Curry and Moore, 1971]. Ceylon is near enough on the west to be a source of stress, in addition to sediment loading. The focal mechanism (Figures 2 and 7) shows strike-slip faulting with a large component of thrusting, one plane of which lines up with the extension of a fracture zone [McKenzie and Sclater, 1971]. The fracture zone has also been located further to the west, however [Sclater and Fisher, 1974].

97: *October 29, 1973 (3-1-C)*. This event lies on the extension of a fracture zone [Sclater and Fisher, 1974]. The bathymetric data are only sketchy but show no sources of topographic stress.

101: *June 25, 1974 (3-2-A)*. The epicenter lies near the elbow of a fracture zone that changes strike, analogous to event 51 on the next fracture zone to the west [Sclater and Fisher, 1974]. The focal mechanism is a thrust fault on a steeply dipping, NE-SW trending fault, a solution consistent with control by the fracture zone [Stein and Okal, 1978]. The Ninetyeast Ridge is near enough on the east to be a possible source of stress, but the focal mechanism indicates compression nearly parallel to the ridge, rather than perpendicular to it. This event was followed by at least four aftershocks (events 102-105) and by one earthquake over a year later (event 115).

127: *December 14, 1976 (3-3-C)*. This event lies on the northern end of the Ninetyeast Ridge, where the ridge is buried under the Bengal Fan [Curry and Moore, 1971]. This end of the Ninetyeast Ridge is extensively broken up into distinct blocks [Sclater and Fisher, 1974].

148: *May 21, 1978 (1-3-D)*. This event occurred about 100 km west of a very steep scarp on the southern section of the Ninetyeast Ridge, on a broad plateau-like feature. The bathymetry in this area is poorly known [Sclater and Fisher, 1974]. Topographic stresses are likely.

149: *June 3, 1978 (3-2-B)*. The epicenter is in a region of very rough bathymetry, about 300 km east of the Ninetyeast ridge. A fracture zone passes very near the epicenter [Sclater and Fisher, 1974]. Although rough, the bathymetric relief is not large and may not be a significant source of local stresses.

153: *August 3, 1978 (2-2-C)*. This event is under the Bengal Fan, well away from Ceylon and the Ninetyeast Ridge [Curry and Moore, 1971]. There is a large bathymetric high to the south and a very narrow, NNE-trending ridge through the epicenter [Sclater and Fisher, 1974]. No fracture zones have been proposed in the epicentral region.

Remainder of the Indian Ocean

18: *September 12, 1965 (3-3-A)*. This was the first event of a swarm which occurred on the steep west scarp of the Chagos Bank [Stein, 1978]. The Vema Fracture Zone trends through the epicentral area, and there is significant bathymetric relief. [Fisher et al., 1974], but the normal faulting mechanism on an E-W fault plane is difficult to reconcile with either observation. The events may be occurring on a cross fracture remaining from the breakup of the Mascarene-Plateau-Chagos Bank [Stein, 1978]. This earthquake series includes a swarm during November 1967 (events 35-41) and several other events up to 1970 (events 43-45, 59, 63).

21: *October 31, 1965 (2-2-B)*. The epicenter is about 90 km west of a large fracture zone [Sclater and Fisher, 1974]. There is some bathymetric relief to the northeast which could be a source of stress.

49: *September 3, 1968 (3-1-B)*. The epicenter lies on the Prince Edward Fracture Zone in the Mozambique Basin [Bergh and Norton, 1976]. There is no large bathymetric relief in the epicentral region [Simpson, 1974].

73: *June 26, 1971 (2-2-B)*. The epicenter is between two fracture zones, about 300 km from the Java trench, on the west side of a broad, gentle trough which parallels the fracture zone to the east [Sclater and Fisher, 1974]. The focal mechanism shows a thrust fault with a large strike-slip component, with one nodal plane subparallel to the trough [Sykes and Sbar, 1974].

91: *March 20, 1973 (2-3-B)*. The event lies on the eastern scarp of the Kerguelen Plateau. There appear to be no fracture zones in the area [Houtz et al., 1977], but there is a strong likelihood that the lithosphere has been severely weakened during the uplift of the plateau.

93: *May 3, 1973 (3-3-A)*. The epicenter is 200 km north of the Kerguelen Plateau (Figure 6). An extensive system of grabens to the southeast may extend into the epicentral region, consistent with the normal fault mechanism (Figures 2 and 6). The plateau is considered to be uplifted oceanic crust, associated with a mantle hotspot which would now be located near the epicentral area [Houtz et al., 1977]. Balmino et al. [1979] found significant geoidal highs associated with the Kerguelen Islands, which implies lithospheric loading and possibly large bending stresses. This earthquake may therefore be a good candidate for further study with regard to constraining the absolute level of stress in the lithosphere, as suggested by Solomon et al. [1980].

94: *May 27, 1973 (2-1-D)*. This event lies near a portion of the Central Indian Ridge which is offset by many short transform faults [McKenzie and Sclater, 1971]. There is no indication of large bathymetric relief in the epicentral area [Fisher et al., 1974].

111: *April 7, 1975 (2-3-C)*. This event lies off the northeast side of the Agulhas Plateau, a likely source of stress. The tectonic history of the region is in dispute; the presence of a fracture zone in the epicentral area may depend on the origin of the plateau, i.e., whether or not it is an extinct spreading center [Scrutton, 1973; Barret, 1977]. There is no bathymetric evidence for a fracture zone [Simpson, 1974].

North Pacific

46: *April 28, 1968 (3-2-A)*. This event occurred on the Emperor Trough, and the thrust fault mechanism has nodal planes parallel to the strike of the trough [Sykes and Sbar, 1974; Stein, 1979]. The mechanism has been shown to be consistent with large-scale deformation of the Pacific plate, which is seen in the vertical offset across the trough [Stein, 1979]. The Emperor Seamounts to the west may be a source of stress [Chase et al., 1970].

156: *February 4, 1979 (1-1-D)*. This event lies midway between the Surveyor and Mendocino fracture zones, in an area of low relief [Chase et al., 1970].

Central Pacific

28: *September 24, 1966 (2-1-C)*. The epicenter lies halfway between the Clarion and Clipperton fracture zones. Bathymetry shows a poorly defined NE-trending structure 100 km to the west but no large relief [Chase et al., 1970]. One nodal plane of the combined thrust and strike-slip focal mechanism may be subparallel to the trend of the nearby structure, but the solution is poorly constrained [Sykes and Sbar, 1974].

58: *May 3, 1969 (3-3-A)*. The epicenter is on the northeast scarp of the Magellan Rise, a probable source of local stress. The Magellan Trough, 150 km to the northeast, has been shown to be an extinct spreading center with a complex history [Tamaki et al., 1979]. The focal mechanism shows thrusting with poorly constrained nodal planes, but there is some evidence that the planes strike northwest, parallel to the strong local structural trend. The *P* axis is then radial to the Magellan Rise [Sykes and Sbar, 1974].

92: *April 26, 1973 (2-3-A)*. This event occurred under the northeast coast of Hawaii, at mantle depth (48 km) [Unger and Ward, 1979]. It was not associated with any volcanic activity or known magma conduits and is therefore considered to be a tectonic event, possibly related to bending stresses associated with lithospheric flexure in response to the load of Hawaii.

Several other Hawaiian events are in Table 1: events 8, 74, 77, 88, and 120. The depths of these events are not well known, but mantle depth earthquakes are found near the epicenters of all these events [Koyanagi et al., 1976]. The large Kalapana earthquake of November 29, 1975, is probably related to magmatic deformation along the East Rift Zone of Kilauea and is not included in the catalog [Furumoto and Koyanagi, 1979].

123: *May 14, 1976 (1-1-C)*. This event occurred 500 km southwest of the Line Islands in an area whose tectonic history is quite complex [Winterer, 1976]. There is no evidence for either large local stresses or fracture zones in the epicentral area [Chase et al., 1970; Winterer, 1976].

Southwest Pacific

72: May 27, 1971 (3-1-A). This event lies in the Tasman Sea near the end of the Balleny Fracture Zone and lies near an extension of the fracture zone under the thick sediments [Hayes and Connolly, 1972]. The bathymetry is rather rough, but there is no large relief [Mammerickx et al., 1973]. The nodal planes of the thrust fault mechanism are poorly constrained, but there is some evidence that they strike NNW, parallel to the fracture zone [Sykes and Sbar, 1974].

79: May 21, 1972 (3-2-B). This event lies in the South Fiji Basin, on the southeast arm of an extinct RRR triple junction, within 100 km of the triple junction itself [Watts et al., 1977]. There is moderate bathymetric relief to the southwest [Mammerickx et al., 1973]. Available first-motion data (Figure 2) permit a wide range of possible fault plane solutions.

Southeast Pacific

12: December 8, 1964 (3-3-A). The epicenter lies on the Nazca Ridge, a possible hotspot tract which is subducted at the Peru trench [Herron, 1972]. There are probable stresses associated with the bathymetry, and the ridge may also represent a weak zone [Mammerickx et al., 1973].

22: November 25, 1965 (2-1-A). The epicenter lies near the Mendaña Fracture Zone, in the middle of the Nazca plate, on the southwest flank of the Galapagos Rise (an extinct spreading center) [Herron, 1972; Mammerickx et al., 1973]. The mechanism [Mendiguren, 1971] shows thrusting on N-S striking planes which might be aligned with weak zones associated with the old ridge. The nodal planes are nearly perpendicular to the trend of the fracture zone, however, and the relief of the Galapagos Rise is very subdued, so that topographic stresses are probably not large. The event is most likely responding to the stress field of the East Pacific Rise [Mendiguren, 1971].

68: May 9, 1971 (2-1-B). This event lies on a fracture zone (Fernandez) proposed by Heezen and Tharp [1969; Stover, 1973], but there is no bathymetric evidence for continuity of the fracture zone between the Pacific-Antarctic Ridge and the Chile Ridge [Mammerickx et al., 1973]. The nodal planes of the thrust fault mechanism are oblique to the trend of the proposed fracture zone [Forsyth, 1973]. There is no large relief in the area. The earthquake was followed by at least three aftershocks (events 69-71).

80: May 28, 1972 (1-1-D). This event is located in the southern part of the Nazca plate, in a poorly surveyed region [Mammerickx et al., 1973]. There is a dominant NNW trend to the structure but no evidence of major structure or bathymetry near the epicenter.

South Pacific

66: June 6, 1970 (3-1-C). The epicenter is in the Bellingshausen Abyssal Plain, very near one of the fracture zones of the Eltanin system [Molnar et al., 1975; Vanney and Johnson, 1976; Weissel et al., 1977]. There is no large bathymetric relief [Mammerickx et al., 1973].

82: September 28, 1972 (2-1-C). The epicenter lies about 500 km west of the East Pacific Rise, in 10-m.y.-old lithosphere. It is about 80 km north of the Menard Fracture Zone [Mammerickx et al., 1973; Molnar et al., 1975; Weissel et al., 1977]. There is no significant bathymetric relief.

118: January 11, 1976 (1-1-E). The epicenter lies in a poorly surveyed portion of the Humboldt Plain. There are no

fracture zones mapped nearby and no source of local stress [Mammerickx et al., 1973; Weissel et al., 1977].

125: October 21, 1976 (1-1-C). The epicenter lies about 100 km from fracture zone VIII of Molnar et al. [1975]. There is no significant relief nearby [Mammerickx et al., 1973].

130: February 5, 1977 (1-1-A). The epicenter is on the continental rise of Antarctica (Figure 9), beneath the Charcot Deep Sea Fan [Vanney and Johnson, 1976]. It lies between two fracture zones of the Eltanin system. The mechanism shows almost pure thrusting on a N-S fault plane (Okal [1980]; Figures 2 and 9). There is no large bathymetric relief nearby, and the *P* axis is oriented at 45° to the nearest portion of the continental shelf.

East Central Pacific

17: September 9, 1965 (3-3-A). The epicenter lies on the southeast flank of the Cocos Ridge, which formed as a hotspot track behind the Galapagos hotspot [Hey, 1977]. The ridge is probably both a source of stress and a zone of weakness. There may also be a fracture zone striking N-S which intersects the ridge near the epicenter, which would be consistent with one of the nodal planes of the strike-slip solution [Molnar and Sykes, 1969; Hey, 1977]. The epicenter is about 200 km from the axis of the Middle America trench [Chase et al., 1970]. This event was preceded by 3 days by a slightly smaller earthquake (16).

64: January 21, 1970 (3-1-A). This event lies 165 km west of the East Pacific Rise (Figure 5), in a shallow trough trending NNW [Chase et al., 1970]. A normal faulting mechanism has been published for the event [Anderson et al., 1974], but a reexamination shows it to be a thrust event on planes which are not well constrained but are consistent with the strike of the trough at the epicenter (Figures 2 and 5). This earthquake had one small aftershock (event 65) and was preceded by over a year by event 53, which was also much smaller.

121: March 11, 1976 (1-1-E). The epicentral area is poorly surveyed but shows no evidence of large relief or structure [Chase et al., 1970].

Western Pacific

20: October 7, 1965 (1-2-B). The epicenter is in the South China Basin. There appear to be no major structures near the epicenter, but there is significant bathymetric relief to the south and southeast [Mammerickx and Smith, 1976]. The mechanism shows thrusting with nodal planes striking NE [Wang et al., 1979]. Event 23 is an aftershock 4 months later, and event 56 has nearly the same epicenter, 3 years later.

48: August 20, 1968 (3-2-A). This event is located at the northern end of the East Caroline Basin (Figure 4), near the northeast corner of what may be a separate Caroline plate [Erlandson et al., 1976; Mammerickx, 1978; Weissel and Anderson, 1978]. The bathymetry is rough, and the epicenter lies in or near a shallow trough striking WNW [Chase et al., 1970]. The mechanism is poorly constrained, but horizontal compression oriented NE is indicated (Weissel and Anderson [1978]; Figures 2 and 4). The catalog contains 11 other events around the Caroline Basin, which may delineate a new plate boundary: 9, 15, 24, 26, 30, 62, 96, 98, 99, 124, and 129.

100: April 12, 1974 (3-3-A). This event occurred at the intersection of the Palau-Kyushu Ridge (an old transform fault) and the Central Basin Fault (an old spreading center) [Ben-Avraham et al., 1972; Uyeda and Ben-Avraham, 1972]. There is

significant bathymetric relief in the epicentral area (Figure 8). The focal mechanism (Figures 2 and 8) is consistent with strike-slip faulting on a plane parallel to the Palau-Kyushu ridge.

119: February 20, 1976 (1-3-D). The epicenter lies among a large group of seamounts which are probable sources of stress [Chase *et al.*, 1970]. Little is known about the tectonics of the area, which lies in the Jurassic quiet zone. Magnetic anomalies mapped to the north indicate a fracture zone which trends into the epicentral area [Hilde *et al.*, 1976; Winterer, 1976].

122: May 6, 1976 (2-3-B). This event occurred on the northwest scarp of the Shatsky Rise, a probable source of local stress. The rise probably formed by volcanic extrusion along an old transform during a reorganization of plate motions in the Mesozoic [Hilde *et al.*, 1976]. Fracture zones intersect the rise but not near the epicenter.

South Central Pacific

5: September 14, 1963 (1-2-C). The epicenter lies near several moderate-sized seamounts in an area which has a complicated tectonic history and many fracture zones, but there are no mapped fracture zones nearby [Mammerickx *et al.*, 1975].

13: March 6, 1965 (2-1-C). This is one of many events clustered near the eastern tip of the Austral Island chain [Talandier and Kuster, 1976; Talandier, 1978]. There are no fracture zones mapped in the area, but the earthquakes may be related to the tectonism which is creating the island chain [Mammerickx *et al.*, 1975]. The bathymetry in this area is subdued. Other seismicity near this epicenter includes events 27 and 113. Event 27 is listed by Bolt [1976] as a French nuclear test, but other sources list it as an earthquake (see text). First-motion data for event 13 (Figure 2) are inconclusive.

14: May 21, 1965 (2-3-C). This event is at the southeast tip of the Tubai Island chain, near large relief [Mammerickx *et al.*, 1975]. There is a fracture zone which may extend into the epicentral area [Mammerickx *et al.*, 1975].

31: April 13, 1967 (3-2-A). This is one of at least 15 earthquakes in a tight, NW-SE trending pattern which follows a small seamount chain north of Caroline Island [Talandier and Kuster, 1976]. The Galapagos Fracture Zone intersects the island chain near this event, but the pattern extends well to either side of the intersection [Mammerickx *et al.*, 1973]. The mechanism shows thrusting, but there is no control on the nodal planes [Sykes and Sbar, 1974].

61: August 6, 1969 (1-1-A). This event is one of many, located in a restricted area, which have been studied in detail by Sverdrup and Jordan [1979]. These authors concluded that the earthquakes are not occurring on an old zone of weakness and are not related to current volcanism, although incipient volcanism cannot be ruled out. There is no major bathymetric relief in the area [Mammerickx *et al.*, 1973]. Other earthquakes in Table 1 with nearly the same epicenter are events 47, 89, and 90.

110: March 29, 1975 (2-1-C). This event lies about 50 km north of a possible fracture zone [Mammerickx *et al.*, 1975]. There is no major bathymetric relief.

112: May 14, 1975 (2-2-C). The epicenter is one of many in this area, near the eastern end of the Tuamotu Archipelago [Mammerickx *et al.*, 1975]. There are no mapped fracture zones nearby, but there is significant bathymetric relief [Mammerickx *et al.*, 1975].

142: January 5, 1978 (1-1-E). This is one of a very large number of events which occurred at the same epicenter from August 1976 to July 1978 [Talandier, 1978]. The site is well east of the Tuamotu Archipelago in a poorly surveyed area [Mammerickx *et al.*, 1975]. The first motions show a thrust type mechanism but do not constrain the nodal planes (Figure 2). The seismicity of this site includes events 135, 138, 144, 146, 150-152, 154, 157, and 158.

Acknowledgments. We thank Gerardo Suárez for help in the preparation of figures, Hans Schouten for providing access to his unpublished magnetic anomaly maps for the Atlantic Ocean, and Emile Okal for preprints of his most recent work. This research was supported by the Division of Earth Sciences, National Science Foundation, grant EAR78-12936, and by an Alfred P. Sloan Research Fellowship to one of us (S.C.S.).

REFERENCES

- Aki, K., Generation and propagation of *G* waves from the Niigata earthquake of June 16, 1964, 2, Estimation of earthquake moment, released energy, and stress-strain drop from the *G* wave spectrum, *Bull. Earthquake Res. Inst. Tokyo Univ.*, 44, 73-88, 1966.
- Anderson, R. N., D. W. Forsyth, P. Molnar, and J. Mammerickx, Fault plane solutions of earthquakes on the Nazca plate boundaries and the Easter plate, *Earth Planet. Sci. Lett.*, 24, 188-202, 1974.
- Artyushkov, E. V., Stresses in the lithosphere caused by crustal thickness inhomogeneities, *J. Geophys. Res.*, 78, 7675-7708, 1973.
- Balmino, G., C. Brossier, A. Cazenave, F. Nouel, K. Dominh, and N. Vales, Geoid of the Kerguelen Islands area determined from Geos 3 altimeter data, *J. Geophys. Res.*, 84, 3827-3832, 1979.
- Barrett, D. M., Agulhas Plateau off South Africa: A geophysical study, *Geol. Soc. Amer. Bull.*, 88, 749-763, 1977.
- Ben-Avraham, Z., J. Segawa, and C. O. Bowin, An extinct spreading center in the Philippine Sea, *Nature*, 240, 453-455, 1972.
- Bergh, H. W., and I. O. Norton, Prince Edward fracture zone and the evolution of the Mozambique Basin, *J. Geophys. Res.*, 81, 5221-5239, 1976.
- Birch, F. S., The Barracuda Fault Zone in the western North Atlantic: Geological and geophysical studies, *Deep Sea Res.*, 17, 847-859, 1970.
- Bolt, B. A., *Nuclear Explosions and Earthquakes: The Parted Veil*, 309 pp., W. H. Freeman, San Francisco, Calif., 1976.
- Bowin, C. O., Origin of the Ninetyeast Ridge from studies near the equator, *J. Geophys. Res.*, 78, 6029-6043, 1973.
- Burke, K., P. J. Fox, and A. M. C. Sengör, Buoyant ocean floor and the evolution of the Caribbean, *J. Geophys. Res.*, 83, 3949-3954, 1978.
- Byerlee, J. D., Friction of rocks, *Pure Appl. Geophys.*, 116, 615-626, 1978.
- Case, J. E., and T. L. Holcombe, Preliminary geologic-tectonic and bathymetric maps of the Caribbean region, *Open File Map 75-146*, U.S. Geol. Surv., Reston, Va., 1975.
- Cazenave, A., and B. Lago, Rheological behavior of the oceanic lithosphere in response to loading (abstract), *Eos Trans. AGU*, 61, 381, 1980.
- Chapple, W. M., and D. W. Forsyth, Earthquakes and bending of plates at trenches, *J. Geophys. Res.*, 84, 6729-6749, 1979.
- Chase, T. E., H. W. Menard, and J. Mammerickx, Bathymetry of the North Pacific, *Charts 1-10*, Inst. of Mar. Resour., Scripps Inst. of Oceanogr., Univ. of Calif., San Diego, 1970.
- Collette, B. J., H. Schouten, K. Rutten, and A. P. Slootweg, Structure of the Mid-Atlantic Ridge province between 12° and 18°N, *Mar. Geophys. Res.*, 2, 143-179, 1974.
- Creavan, C. T., H. Kanamori, and K. Fujita, A large intraplate event near the Scotia Arc (abstract), *Eos Trans. AGU*, 60, 894, 1979.
- Curry, J. R., and D. G. Moore, Growth of the Bengal deep-sea fan and denudation in the Himalayas, *Geol. Soc. Amer. Bull.*, 82, 563-572, 1971.
- Dash, B. P., M. M. Ball, G. A. King, L. W. Butler, and P. A. Rona, Geophysical Investigation of the Cape Verde Archipelago, *J. Geophys. Res.*, 81, 5249-5259, 1976.
- Defense Mapping Agency, Map of the Bay of Bengal, *Int. Chart Ser. INT 706*, Hydrogr. Center, Washington, D. C., 1976.

- Erlandson, D. L., T. L. Orwig, G. Kiiilgaard, J. H. Mussells, and L. W. Kroenke, Tectonic interpretations of the East Caroline and Lyra basins from reflection profiling investigations, *Geol. Soc. Amer. Bull.*, 87, 453-462, 1976.
- Evernden, J. F., Precision of epicenters obtained by small numbers of world-wide stations, *Bull. Seismol. Soc. Amer.*, 59, 1365-1398, 1969.
- Fisher, R. L., E. T. Bunce, et al., *Initial Reports of the Deep Sea Drilling Project*, vol. 24, U.S. Government Printing Office, Washington, D. C., 1974.
- Fitch, T. J., Plate convergence, transcurrent faults, and the internal deformation adjacent to southeast Asia and the western Pacific, *J. Geophys. Res.*, 77, 4432-4460, 1972.
- Forsyth, D. W., Compressive stress between two mid-ocean ridges, *Nature*, 243, 78-79, 1973.
- Forsyth, D. W., Fault plane solutions and tectonics of the South Atlantic and Scotia Sea, *J. Geophys. Res.*, 80, 1429-1443, 1975.
- Francheteau, J., and X. Le Pichon, Marginal fracture zones as structural framework of continental margins in South Atlantic Ocean, *Amer. Ass. Petrol. Geol. Bull.*, 56, 991-1007, 1972.
- Frankel, A., and W. McCann, Moderate and large earthquakes in the South Sandwich Arc: Indicators of tectonic variations along a subduction zone, *J. Geophys. Res.*, 84, 5571-5577, 1979.
- Furumoto, A. S., and R. L. Kovach, The Kalapana earthquake of November 29, 1975: An intraplate earthquake and its relation to geothermal processes, *Phys. Earth Planet. Interiors*, 18, 197-208, 1979.
- Gutenberg, B., and C. F. Richter, Earthquake magnitude, intensity, energy and acceleration, *Bull. Seismol. Soc. Amer.*, 46, 105-145, 1956.
- Harbison, R. N., R. K. Lattimore, and P. A. Rona, Structural lineations in the Canary Basin, eastern central North Atlantic, *Mar. Geol.*, 14, 269-275, 1973.
- Hayes, D. E., and J. Connolly, Morphology of the southeast Indian Ocean, in *Antarctic Oceanology II: The Australian-New Zealand Sector*, *Antarctic Res. Ser.*, vol. 19, edited by D. E. Hayes, pp. 125-145, AGU, Washington, D. C., 1972.
- Heezen, B. C., and M. Tharp, Physiographic chart of the Pacific Ocean floor, supplement, *Nat. Geogr.*, 136, 496-499, 1969.
- Herron, E. M., Seafloor spreading and Cenozoic history of the East Central Pacific, *Geol. Soc. Amer. Bull.*, 83, 1671-1692, 1972.
- Hey, R. N., Tectonic evolution of the Cocos-Nazca spreading center, *Geol. Soc. Amer. Bull.*, 88, 1404-1420, 1977.
- Hilde, T. W. C., N. Isezaki, and J. M. Wageman, Mesozoic seafloor spreading in the North Pacific, in *Geophysics of the Pacific Ocean Basin and Its Margin*, *Geophys. Monogr. Ser.*, vol. 19, edited by G. H. Sutton, M. H. Manghnani, and R. Moberly, pp. 205-226, AGU, Washington, D. C., 1976.
- Houtz, R. E., D. E. Hayes, and R. G. Markl, Kerguelen Plateau bathymetry, sediment distribution and crustal structure, *Mar. Geol.*, 25, 95-130, 1977.
- Jackson, E. D., and H. Shaw, Stress fields in central portions of the Pacific plate: Delineated in time by linear volcanic chains, *J. Geophys. Res.*, 80, 1861-1874, 1975.
- Johnson, G. L., and P. R. Vogt, Morphology of the Bermuda Rise, *Deep Sea Res.*, 18, 605-618, 1971.
- Johnson, G. L., R. N. Hey, and A. Lowrie, Marine geology in the environs of Bouvet Island and the South Atlantic triple junction, *Mar. Geophys. Res.*, 2, 23-26, 1973.
- Jordan, T. H., The present-day motions of the Caribbean plate, *J. Geophys. Res.*, 80, 4433-4440, 1975.
- Kafka, A. L., and D. J. Weidner, The focal mechanisms and depths of small earthquakes as determined from Rayleigh-wave radiation patterns, *Bull. Seismol. Soc. Amer.*, 69, 1379-1390, 1979.
- Kanamori, H., and D. L. Anderson, Theoretical basis of some empirical relations in seismology, *Bull. Seismol. Soc. Amer.*, 65, 1073-1095, 1975.
- Katsumata, M., and L. R. Sykes, Seismicity and tectonics of the western Pacific: Izu-Mariana-Caroline and Ryukyu-Taiwan regions, *J. Geophys. Res.*, 74, 5923-5948, 1969.
- Koyanagi, R. Y., E. T. Endo, and P. L. Ward, Seismic activity on the island of Hawaii, 1970 to 1973, in *Geophysics of the Pacific Ocean Basin and Its Margin*, *Geophys. Monogr. Ser.*, vol. 19, edited by G. H. Sutton, M. H. Manghnani, and R. Moberly, pp. 169-172, AGU, Washington, D. C., 1976.
- Laughton, A. S., D. G. Roberts, and R. Graves, Bathymetry of the northeast Atlantic: Mid-Atlantic Ridge to southwest Europe, *Deep Sea Res.*, 22, 791-810, 1975.
- Mammerickx, J., Re-evaluation of some geophysical observations in the Caroline Basin, *Geol. Soc. Amer. Bull.*, 89, 192-196, 1978.
- Mammerickx, J., and S. M. Smith, Bathymetry of the southeast Pacific, *Map MC-26*, Geol. Soc. of Amer., Boulder, Colo., 1976.
- Mammerickx, J., S. M. Smith, I. L. Taylor, and T. E. Chase, Bathymetry of the South Pacific, *Charts 11-20*, Inst. of Mar. Resour., Scripps Inst. of Oceanogr., Univ. of Calif., San Diego, 1973.
- Mammerickx, J., R. N. Anderson, H. W. Menard, and S. M. Smith, Morphology and tectonic evolution of the east-central Pacific, *Geol. Soc. Amer. Bull.*, 86, 111-118, 1975.
- McGarr, A., and N. C. Gay, State of stress in the earth's crust, *Annu. Rev. Earth Planet. Sci.*, 6, 405-436, 1978.
- McKenzie, D. P., The relation between fault plane solutions for earthquakes and the directions of the principal stresses, *Bull. Seismol. Soc. Amer.*, 59, 591-601, 1969.
- McKenzie, D. P., and J. G. Sclater, The evolution of the Indian Ocean since the Late Cretaceous, *Geophys. J. Roy. Astron. Soc.*, 24, 437-528, 1971.
- Mendiguren, J. A., Focal mechanism of a shock in the middle of the Nazca plate, *J. Geophys. Res.*, 76, 3861-3879, 1971.
- Mendiguren, J. A., and F. M. Richter, On the origin of compressional intraplate stresses in South America, *Phys. Earth Planet. Interiors*, 16, 318-326, 1978.
- Minster, J. B., and T. H. Jordan, Present-day plate motions, *J. Geophys. Res.*, 83, 5331-5354, 1978.
- Molnar, P., and L. R. Sykes, Tectonics of the Caribbean and middle America regions from focal mechanisms and seismicity, *Geol. Soc. Amer. Bull.*, 80, 1639-1684, 1969.
- Molnar, P., T. Atwater, J. Mammerickx, and S. M. Smith, Magnetic anomalies, bathymetry, and tectonic evolution of the South Pacific since the Late Cretaceous, *Geophys. J. Roy. Astron. Soc.*, 40, 383-420, 1975.
- Nakamura, K., K. H. Jacobs, and J. N. Davies, Volcanoes as possible indicators of tectonic stress orientations—Aleutians and Alaska, *Pure Appl. Geophys.*, 115, 87-112, 1977.
- Okal, E. A., The Bellingshausen Sea earthquake of February 5, 1977: Evidence for ridge generated compression in the Antarctic plate, *Earth Planet. Sci. Lett.*, 46, 306-310, 1980.
- Raleigh, C. B., J. H. Healy, and J. D. Bredehoeft, Faulting and crustal stress at Rangely, Colorado, in *Flow and Fracture of Rocks*, *Geophys. Monogr. Ser.*, vol. 16, edited by H. C. Heard, I. Y. Borg, N. L. Carter, and C. B. Raleigh, pp. 275-284, AGU, Washington, D. C., 1972.
- Richardson, R. M., and S. C. Solomon, Apparent stress and stress drop for intraplate earthquakes and tectonic stress in the plates, *Pure Appl. Geophys.*, 115, 317-331, 1977.
- Richardson, R. M., S. C. Solomon, and N. H. Sleep, Tectonic stress in the plates, *Rev. Geophys. Space Phys.*, 17, 981-1020, 1979.
- Rona, P. A., and H. S. Fleming, Mesozoic plate motions in the eastern central North Atlantic, *Mar. Geol.*, 14, 239-252, 1973.
- Sbar, M. L., and L. R. Sykes, Contemporary compressive stress and seismicity in eastern North America: An example of intraplate tectonics, *Geol. Soc. Amer. Bull.*, 84, 1861-1882, 1973.
- Schouten, H., and K. D. Klitgord, Atlantic sea-floor spreading: Mesozoic magnetic anomalies, submitted to *J. Geophys. Res.*, 1980.
- Sclater, J. G., and R. L. Fisher, Evolution of the east central Indian Ocean, with emphasis on the tectonic setting of the Ninetyeast Ridge, *Geol. Soc. Amer. Bull.*, 85, 683-702, 1974.
- Scrutton, R. A., Structure and evolution of the seafloor south of South Africa, *Earth Planet. Sci. Lett.*, 19, 250-256, 1973.
- Sibuet, J. C., and J. Mascle, Plate kinematic implications of Atlantic equatorial fracture zone trends, *J. Geophys. Res.*, 83, 3401-3421, 1978.
- Silver, E. A., J. E. Case, and H. J. MacGillivray, Geophysical study of the Venezuelan borderland, *Geol. Soc. Amer. Bull.*, 86, 213-226, 1975.
- Simpson, E. S. W., Southeast Atlantic and southwest Indian oceans, *Chart 125A*, Nat. Res. Inst. of Oceanol., Johannesburg, South Africa, 1974.
- Sobczak, L. W., Bathymetry of the Arctic Ocean north of 85°N latitude, *Tectonophysics*, 42, T27-T33, 1977.
- Solomon, S. C., R. M. Richardson, and E. A. Bergman, Tectonic stress—models and magnitudes, *J. Geophys. Res.*, in press, 1980.
- Stein, S., An earthquake swarm on the Chagos-Laccadive Ridge and its tectonic implications, *Geophys. J. Roy. Astron. Soc.*, 55, 577-588, 1978.

- Stein, S., Intraplate seismicity on bathymetric features: The 1968 Emperor Trough earthquake, *J. Geophys. Res.*, **84**, 4763–4768, 1979.
- Stein, S., and E. A. Okal, Seismicity and tectonics of the Ninetyeast Ridge area: Evidence for internal deformation of the Indian Plate, *J. Geophys. Res.*, **83**, 2233–2246, 1978.
- Stein, S., N. H. Sleep, R. J. Geller, S. C. Wang, and G. C. Kroeger, Earthquakes along the passive margin of eastern Canada, *Geophys. Res. Lett.*, **6**, 537–540, 1979.
- Stewart, G. S., The Grand Banks earthquake of November 18, 1929 and the Bermuda earthquake of March 24, 1978—A comparative study in relation to their intraplate locations (abstract), *Eos Trans. AGU*, **60**, 312, 1979.
- Stover, C. W., Seismicity and tectonics of the east Pacific Ocean, *J. Geophys. Res.*, **78**, 5209–5220, 1973.
- Sverdrup, K., and T. Jordan, Bathymetric survey of seismic region A, south central Pacific Ocean (abstract), *Eos Trans. AGU*, **60**, 957, 1979.
- Sverdrup, K., T. H. Jordan, J. Talandier, and E. A. Okal, Seismicity and tectonic stress in the south central Pacific (abstract), *Eos Trans. AGU*, **60**, 311, 1979.
- Sykes, L. R., Seismicity of the Indian Ocean and a possible nascent island arc between Ceylon and Australia, *J. Geophys. Res.*, **75**, 5041–5055, 1970.
- Sykes, L. R., Intraplate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation, *Rev. Geophys. Space Phys.*, **16**, 621–688, 1978.
- Sykes, L. R., and M. L. Sbar, Focal mechanism solution of intraplate earthquakes and stresses in the lithosphere, in *Geodynamics of Iceland and the North Atlantic Area*, edited by L. Kristjansson, pp. 207–224, D. Reidel, Hingham, Mass., 1974.
- Talandier, J., Report on seismic and volcanic activity in the southern Central Pacific Ocean from 1965 to August 1978, report, Lab. de Geophys., Papeete, Tahiti, 1978.
- Talandier, J., and G. Kuster, Seismicity and submarine volcanic activity in French Polynesia, *J. Geophys. Res.*, **81**, 936–948, 1976.
- Talwani, M., C. C. Windisch, P. L. Stoffa, P. Buhl, and R. E. Houtz, Multi-channel seismic study in the Venezuelan Basin and the Curacao Ridge, in *Island Arcs, Deep Sea Trenches, and Back-Arc Basins*, *Maurice Ewing Ser.*, vol. 1, edited by M. Talwani and W. C. Pitman III, pp. 83–98, AGU, Washington, D. C., 1977.
- Tamaki, K., M. Joshima, and R. L. Larson, Remanent early Cretaceous spreading center in the Central Pacific Basin, *J. Geophys. Res.*, **84**, 4501–4510, 1979.
- Uchupi, E., Bathymetric atlas of the Atlantic, Caribbean, and Gulf of Mexico, *Ref. 71-72*, Woods Hole Oceanogr. Inst., Woods Hole, Mass., 1971.
- Udintsev, G. G., R. L. Fisher, V. F. Kanaev, A. S. Laughton, E. S. W. Simpson, and D. I. Zhiv, *Geological-Geophysical Atlas of the Indian Ocean*, USSR Academy of Sciences, Moscow, 1975.
- Unger, J. D., and P. L. Ward, A large, deep Hawaiian earthquake—The Honou, Hawaii event of April 26, 1973, *Bull. Seismol. Soc. Amer.*, **69**, 1771–1782, 1979.
- Uyeda, S., and Z. Ben-Avraham, Origin and development of the Philippine Sea, *Nature*, **240**, 176–178, 1972.
- Vanney, J. R., and G. L. Johnson, The Bellingshausen-Amundsen basins (southeastern Pacific): Major sea-floor units and problems, *Mar. Geol.*, **22**, 71–101, 1976.
- Vogt, P. R., and O. E. Avery, Detailed magnetic surveys in the northeast Atlantic and Labrador Sea, *J. Geophys. Res.*, **79**, 363–389, 1974.
- Vogt, P. R., C. N. Anderson, and D. R. Bracey, Mesozoic magnetic anomalies, sea-floor spreading, and geomagnetic reversals in the southwestern North Atlantic, *J. Geophys. Res.*, **76**, 4796–4823, 1971.
- Vogt, P. R., P. T. Taylor, L. C. Kovacs, and G. L. Johnson, Detailed aeromagnetic investigation of the Arctic Basin, *J. Geophys. Res.*, **84**, 1071–1089, 1979.
- Walcott, R. I., Flexure of the lithosphere at Hawaii, *Tectonophysics*, **9**, 435–446, 1970.
- Wang, S. C., R. J. Geller, S. Stein, and B. Taylor, An intraplate thrust earthquake in the South China Sea, *J. Geophys. Res.*, **84**, 5627–5632, 1979.
- Watts, A. B., and J. R. Cochran, Gravity anomalies and flexure of the lithosphere along the Hawaiian-Emperor seamount chain, *Geophys. J. Roy. Astron. Soc.*, **38**, 119–141, 1974.
- Watts, A. B., J. K. Weisell, and F. J. Dewey, Tectonic evolution of the South Fiji marginal basin, in *Island Arcs, Deep Sea Trenches, and Back-Arc Basins*, *Maurice Ewing Ser.*, vol. 1, edited by M. Talwani and W. C. Pitman III, pp. 419–427, AGU, Washington, D. C., 1977.
- Weissel, J. K., and R. N. Anderson, Is there a Caroline plate? *Earth Planet. Sci. Lett.*, **41**, 143–158, 1978.
- Weissel, J. K., D. E. Hayes, and E. M. Herron, Plate tectonics synthesis: The displacements between Australia, New Zealand, and Antarctica since the Late Cretaceous, *Mar. Geol.*, **25**, 231–277, 1977.
- Weissel, J. K., R. N. Anderson, and C. Geller, Deformation of the Indo-Australian plate: Observations and implications, *Nature*, in press, 1980.
- Winterer, E. L., Bathymetry and regional tectonic setting of the Line Island Chain, in *Initial Reports of the Deep Sea Drilling Project*, vol. 33, pp. 731–747, U.S. Government Printing Office, Washington, D. C., 1976.
- Yang, J. P., and Y. P. Aggarwal, Intraplate stresses near Atlantic-type passive continental margins (abstract), *Eos Trans. AGU*, **60**, 309, 1979.
- Zoback, M. L., and M. Zoback, Interpretative stress map of the coterminous United States, in *The Magnitude of Deviatoric Stresses in the Earth's Crust and Upper Mantle*, *Open File Rep. 80-625*, vol. 1, pp. 353–433, U.S. Geological Survey, Reston, Va., 1980.

(Received January 28, 1980;
revised May 16, 1980;
accepted May 19, 1980.)