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Intraplate Seismicity of the Southern Part of the Pacific Plate

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We present a compilation of all the known seismicity of the part of the Pacific plate located south of the equator. This includes historical earthquakes dating back to 1937. In addition to the Gilbert Islands swarm of 1981-1983 we identify 60 events at 32 epicenters in the magnitude range 4.3-6.9. Our results indicate that the level of seismicity of the southern Pacific plate is comparable to that of other plates, including continent-bearing ones. Most of the seismic activity is concentrated in the youngest parts of the plate, but a few large events are present in Cretaceous lithosphere. With the exception of events from the Gilbert Islands swarm, earthquakes located in lithosphere 35 Ma and older are usually compatible with the release of horizontal compressional stress due to the gravitational process known as ridge push; in younger areas the situation is much less clear-cut. The final geometry of the focal mechanisms is controlled by local tectonics; in particular, due to the Miocene reorientation of the spreading at the East Pacific Rise, the bulk of the Pacific plate is particularly vulnerable to ridge push, and this explains the numerous strike-slip events and the absence of magnitude 6 events in most of the plate. Correlations of epicenters with zones of weakness or bathymetric features are frequent but not universal. Finally, a number of areas with high seismic release are identified as potential targets for future exploration.

INTRODUCTION

Ever since *Sykes and Sbar* [1974] started investigating the focal mechanisms of intraplate earthquakes, these events have provided exclusive information on the regime of stress inside oceanic plates, thereby constraining the nature of forces driving their system [*Richardson et al.*, 1979; *Bergman and Solomon*, 1980; *Okal*, 1983]. Additionally, oceanic intraplate earthquakes have been found confined to a relatively shallow maximum depth, which correlates well with the flexural elastic thickness of the lithosphere; this observation suggests a dry olivine rheology for the plate [*Wiens and Stein*, 1983].

However, few of the previous studies of intraplate seismicity have relied on complete data sets including historical earthquakes. Among the most recent ones, *Bergman and Solomon*'s [1980] catalogue includes only three pre-1963 events, while *Wiens and Stein* [1983] use a homogeneous data set based on records of the World Wide Standard Seismograph Network (WWSSN) spanning the years 1964-1978. Furthermore, little systematic attention has been given to the Pacific plate: while *Bergman and Solomon* [1980] included 57 Pacific plate events in their study, only seven had at least some focal mechanism information; *Sykes and Sbar* [1974] studied five events, *Richardson et al.* [1979] used only two incomplete Pacific plate focal solutions, and *Wiens and Stein* [1983] used four Pacific plate epicenters. An exception to this pattern is the detailed study of the seismicity detected at lower magnitudes in the south-central Pacific, through the operation of the French Polynesian network. The reader is referred to *Talandier and Kuster* [1976] and *Talandier and Okal* [1984], who investigated local swarms of volcanic character, and to *Okal et al.* [1980] for a review of regional seismicity and a discussion of its tectonic implications over an area larger than 10^7 km^2.

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Because of the relatively low level of intraplate seismicity and its usually large recurrence times, there is an obvious trade-off between completeness and homogeneity in the data set. A quantitative study such as Wiens and Stein's requires a homogeneous detection level both in time and space and is thus limited to the lifetime of the standardized network. However, a period of investigation of only 15 years may not be a representative sampling of the long-term seismic behavior of an intraplate area. For example, no magnitude 6 event took place in the whole Pacific plate between 1955 and 1970, and conclusions on its seismicity drawn from this relatively short interval would obviously be erroneous. The systematic compilation of all earthquakes, including historical events, is therefore a necessary step of the careful investigation of intraplate seismicity and can provide a wealth of additional information. This approach was used for the Antarctic plate, serving proof that its level of seismicity is comparable to that of other plates and allowing an estimate of the seismic recurrence time and of the strain rate inside the plate [*Okal*, 1981]. The purpose of the present paper is to compile similarly all known seismicity in the southern part of the Pacific plate and to discuss it in the framework of recent advances in lithospheric studies, such as *Wiens and Stein*'s [1983; 1984*a*]. Preliminary results discussed below show that the seismicity of the northern portion of the plate has a generally lower level than south of the equator (except for Hawaiian events clearly associated with the hot spot's activity). Also, most recent significant earthquakes in the northern Pacific plate have been individually studied [e.g., *Stein*, 1979; *Bergman and Solomon*, 1980]. Thus we restrict ourselves to the southern portion of the plate, shown on Figure 1, the equator being used as a voluntary and somewhat arbitrary boundary of our study area to the north.

The goals pursued in the present study are several: Apart from the general improvement of our knowledge of the seismicity which results from compilation and careful relocation, magnitudes measured from available records are used to compute cumulative seismic figures, found to match closely results obtained worldwide or in other plates. By

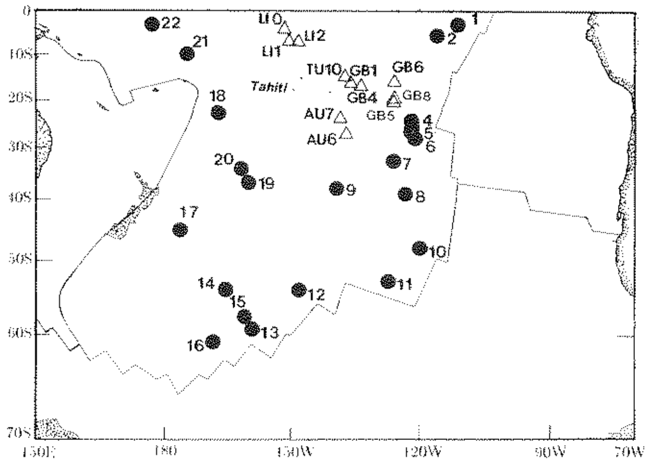

Fig. 1. Map of the southern part of the Pacific with epicenters listed in Table 1. This is a standard Mercator projection. The continuous lines are the plate boundaries. Solid dots are IP sites, identified by their number. Open triangles, with codes, are seismic locations identified by the French Polynesian network, having had seismicity at a magnitude level ≥ 4.7, or discussed in the text.

using all available (and often incomplete) bathymetric information, we attempt, with a highly variable degree of success, to correlate the location of seismic epicenters with active or fossil tectonic features, evocative of potential weakness in the plate [*Sykes,* 1978]. Finally, a few new focal mechanisms are discussed; some, but not all, confirm the predominance of gravity-induced forces as agents responsible for the plate's internal stresses.

A particularly intriguing feature of the seismicity of the Pacific plate is the notable paucity of magnitude 6 events: in particular, *Bergman and Solomon* [1980] observed that b value regressions of their worldwide data sets single out the Pacific plate as having higher than normal b values, and they suggested that it may be due to a larger contribution from volcanic earthquakes. With the help of new magnitude determinations we confirm that magnitude 6 events are found in the Pacific plate only along its fringes and in conjunction with the Hawaiian hot spot; we propose to explain this feature as a consequence of the spreading reorientation which took place in the south Pacific during the Miocene, resulting in a more vulnerable geometry for the bulk of the plate, which was generated at the now defunct Farallon Ridge.

Source Of Data And Methods

As in the case of our previous work in the Antarctic plate [*Okal,* 1981], we obtained a preliminary catalogue through a computer search of the National Oceanic and Atmospheric Administration (NOAA) epicentral tape (1920 to May 1981), completed by recent Preliminary Determination of Earthquakes (PDE) monthly and weekly bulletins (to December 31, 1983). The intraplate character of an event was assessed originally by requiring that it lie at least 2° away from the closest plate boundary, digitized from various bathymetric maps. Relocations occasionally sent an epicenter back to a plate boundary, resulting in deletion from the catalogue.

Since the purpose of this study is to emphasize large and old events, we chose not to include the entire low-level seismicity detected by the regional network in French Polynesia, but rather to retain only events with $M_L \geq 4.7$. The reader is referred to *Okal et al.* [1980] for the full data set covering the period 1965-1979. We also delete from the catalogue events suspected of being nuclear explosions, discriminated on the basis of their epicentral location and origin time, a depth reported in the PDE bulletin as "0, assigned by geophysicist", and strong m_b:M_s anomalies. Additionally, we delete two events improperly recorded on the NOAA tape and included in *Talandier and Kuster*'s [1976] Table 1: The January 12, 1942, earthquake reportedly at 8°S, 156.5°W, is actually at 156.5°E, in the Solomon Islands, and the December 23, 1942, event, reportedly at 9°S, 161°W, is actually at 4.5°S, 152°E, in the area of New Ireland. These errors are obvious from the International Seismological Summary (ISS) data, but they (and others of a similar nature in different areas) emphasize the need to treat the NOAA listings with caution.

Copies of original records from worldwide stations were obtained for 14 events predating the WWSSN, and together with times reported in the ISS (later ISC), form the backbone of the further study of individual events.

Relocations

Relocations were carried out for all events for which detailed P arrival times are available in the ISS. This includes major events, both located or simply listed in the ISS, prior to 1957. During the period 1957-1963, the ISS does not list data from earthquakes which it does not locate. After 1963, and with the implementation of the WWSSN, NOAA and ISC epicenters become highly accurate, and relocations are usually not warranted. The relocation program is interactive and uses an iterative least squares regression. As in the case of Antarctica [*Okal,* 1981] and given the general paucity of the data, we chose to constrain the hypocentral depths to 10 km during the relocations. This average value is in agreement with the general observation of the shallowness of oceanic intraplate seismicity [*Bergman and Solomon,* 1980; *Wiens and Stein,* 1983]. It is also justified by a few observations at regional distances of high-frequency Rayleigh-type interface waves controlled by sediments, which *Harkrider and Okal* [1982] have used to derive a maximum depth of a few kilometers into the hard rock for event 10 at site AU-6. Finally, the 10-km depth is now taken as the standard default for oceanic earthquakes for which the PDE hypocentral solution fails to converge. At any rate, and since most of our relocations use only distant stations, any change in the fixed hypocentral depth would only trade off with origin time and would not affect the epicentral location. Root mean square residuals (σ in Table 1) following relocation were usually on the order of 1 s. It is difficult to estimate the precision of times listed in the ISS bulletins and thus to conduct a formal variance investigation of the resulting relocation parameters; if the ISS times are given uncorrelated errors of ± 1 s, standard deviations on the epicentral parameters vary from 15 to 50 km.

Focal Mechanisms

Because of the paucity of first-hand data as well as of the large epicentral distances involved, very few focal mecha-

TABLE 1. Intraplate Earthquakes in the Southern Pacific Plate

Site	Event	Date	Original Epicenter						Relocation			
			°S	°W	Time	Ref	m_b	M_s	°S	°W	Time	σ, s
LI-0	1	Jan. 25, 1962	4.6	152.6	1003:06.8	a	4.6					
LI-1	2	April 13, 1967	7.0	151.0	1426:49.5	a	5.2					
LI-2	3	July 29, 1968	7.50	148.27	0245:44.5	a	4.9					
	4	Dec. 28, 1968	7.60	148.60	2023:04.7	a	4.7					
	5	Aug. 6, 1969	7.36	148.38	1715:39.7	a	5.1					
	6	April 19, 1970	7.4	148.0	0147:22	a	4.7					
	7	Jan. 19, 1973	7.37	148.28	0736:32.9	a	4.9					
	8	Jan. 19, 1973	7.27	148.35	1526:26.1	a	4.9					
TU-10	9	April 16, 1982	15.37	138.14	0252:52.1	d	4.8					
AU-6	10	Nov. 20, 1979	26.57	138.84	0645:04.4	a	5.3					
	11	Sept. 20, 1982	26.6	138.8	0829:44	b						
AU-7	12	Oct. 6, 1982	23.56	140.11	0747:32.5	b	4.2					
GB-1	13	May 14, 1975	17.43	136.05	1149:42.3	a	5.0					
GB-4	14	March 6, 1965	18.40	132.90	1110:53.1	a	5.5					
	15	Sept. 18, 1966	18.48	132.81	0640:35.5	a	5.0					
	16	Aug. 18, 1968	18.4	132.8	1447:59	a	4.7					
	17	May 25, 1975	18.46	132.79	1416:32.8	a	5.0					
GB-5	18	Oct. 31, 1977	20.75	126.89	0819:13.5	a	5.1					
	19	Jan. 5, 1978	20.89	126.93	0323:24.9	a	5.5					
	20	Feb. 19, 1978	20.83	126.70	2357:59.6	c	4.7					
	21	March 11, 1978	20.89	126.96	0332:10.1	c	4.8					
	22	July 13, 1978	20.74	127.00	1804:19.6	a	5.1					
	23	July 25, 1978	20.83	126.94	0754:07.5	a	5.4					
	24	Oct. 18, 1978	20.97	126.85	0203:56.3	a	5.1					
	25	Feb. 5, 1979	20.75	127.26	1735:52.9	a	4.8					
	26	Feb. 26, 1979	20.8	126.7	0631:53	a	5.1					
	27	May 21, 1979	20.8	126.7	1534:02	a	4.8					
GB-8	28	July 31, 1983	20.14	126.87	1026:00.2	d	6.0	5.3				
	29	July 31, 1983	20.12	127.01	1157:50.0	d	5.4					
GB-6	30	Jan. 30, 1978	16.23	126.94	1834:19.6	a	5.0					
IP-1	31	Sept. 30, 1981	4.80	112.02	2303:47.6	d	6.1	5.3				
	32	Oct. 4, 1981	4.71	111.95	1710:22.6	d	5.1					
IP-2	33	Jan. 30, 1980	5.33	115.98	0031:14.0	f	5.3	4.2				
IP-3	34	Nov. 11, 1949	9	119	1057:36	e		6.0	9.0	109.8	1057:29	*
IP-4	35	Nov. 22, 1955	24.5	123	0324:00	e	6.8	6.2	24.27	122.77	0324:05	1.48
IP-5	36	Sept. 29, 1951	26.6	122	1815:00	g	5.9		26.55	121.95	1815:00	0.5
IP-6	37	Aug. 3, 1951	28	121	1920:15	g	5.6					
IP-7	38	Sept. 14, 1963	33.6	126.7	1616:51.8	e	4.9		33.46	126.56	1616:48.0	1.15
IP-8	39	July 31, 1976	38.7	123.0	0511:21	f	4.3					
IP-9	40	March 29, 1975	37.8	138.9	1502:29.6	f	5.0					
IP-10	41	July 30, 1958	48	120	1510:12	e						
	42	Sept. 28, 1972	47.71	119.88	0657:35.6	f	5.1					
IP-11	43	July 14, 1951	52	128	0621:14	e		5.5	52.94	126.57	0621:19.7	1.32
IP-12	44	Sept. 5, 1938	56	147	1442:26	g		6.0	54.74	148.35	1442:31.0	1.25
IP-13	45	Dec. 15, 1947	59.4	159.7	1920:24	g		6.8	58.76	159.20	1920:26.3	1.79
	46	April 10, 1950	58	160	0606:46	g		6.4	58.70	159.11	0606:43.7	0.97
	47	Sept. 17, 1949	35	154	2246:25	e		6.2	58.8	157.7	2246:34	
IP-14	48	Oct. 17, 1959	54	165	0123:00	e						
IP-15	49	Oct. 17, 1959	57.33	160.95	0835:00	g		5.5	57.36	161.00	0835:02.5	1.28
	50	Oct. 21, 1976	57.37	161.1	0356:28	f	5.4					
IP-16	51	July 26, 1958	60.5	168.5	0835:10	e						
IP-17	52	Aug. 1, 1953	44.8	175.8	1336:28	g	5.3					
IP-18	53	Oct. 4, 1937	26.8	163.1	0740:30	e		6.2	22.01	166.99	0740:34.8	1.20
IP-19	54	Jan. 13, 1938	31	155	0315:18	e	5.6		36.67	160.39	0315:22.5	
IP-20	55	Jan. 4, 1940	34	162	0110:18	e		5.9	33.84	162.47	0110:21.4	1.3
IP-21	56	April 24, 1937	12	178	0458:35	e		6.2	10.14	176.00	0458:30	1.64
	57	April 25, 1940	8.5	176.5	1018:42	e						
	58	June 26, 1956	10	173.5	1123:09	e						
	59	Oct. 3, 1957	10	179	1344:30	e						
	60	March 5, 1964	6.1	172.5	0550:15.2	e						
IP-22	G1	Jan. 7, 1982	3.4	-177.5	0842:52.7	h	5.8	5.4				
	G2	Feb. 15, 1982	3.5	-177.5	0550:12.2	h	5.7	5.6				
	G3	March 16, 1982	3.3	-177.5	0020:43.9	h	5.6	5.6				
	G4	May 23, 1982	3.4	-177.4	2132:35.1	h	5.9	5.7				

References for initial locations are a, *Okal et al.* [1980]; b, J. Talandier (personal communication, 1983); c, *Jordan and Sverdrup* [1981]; d, PDE bulletins; e, ISS; f, ISC; g, NOAA epicentral tape; h, *Lay and Okal* [1983]. An additional 213 events were recorded at IP-22.

*Event relocated as interplate.

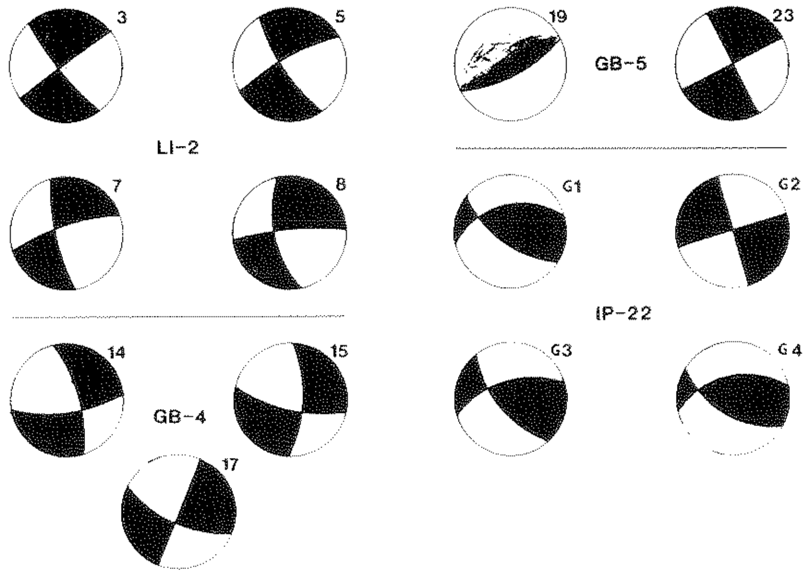

Fig. 2*a*. Focal mechanisms obtained by *Okal et al.* [1980] in Polynesia and *Lay and Okal* [1983] at IP-22. Event numbers referring to Tables 1 and 2 are given at upper right for each event. The shading of event 19 illustrates the poor constraints on the second plane.

nisms could be studied, and *P* wave first-motions usually did not suffice to constrain them. Our limited *P*-wave first motion data sets were complemented by occasional observation of *SV/SH* ratios, Love/Rayleigh ratios, and relative amplitudes of *S, SS, PP,* and *PS*. This information was then used to eliminate solutions in a computer search of all focal mechanisms compatible with the *P* data. As a safeguard, only conservative constraints were used (e.g., if an *SV/SH* ratio is observed to be larger than 2, we would simply eliminate all solutions for which it is smaller than 1). This rather inelegant approach turned out to be surprisingly effective in constraining the mechanisms, as will be discussed in the appendix for specific cases. Focal solutions obtained both in this study and previous ones are sketched on Figures 2-4.

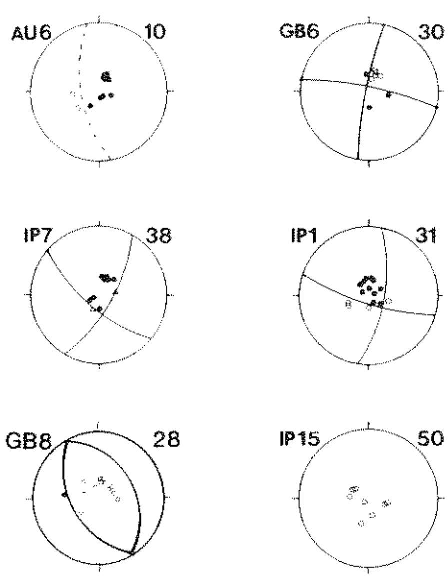

Fig. 2*b*. Focal mechanisms obtained in the present study (lower hemisphere stereographic projection of *P* wave first arrivals). Solid circles are compressions, open ones dilatations. Open triangles identify emergent arrivals on high-gain ($\geq$ 50,000) WWSSN stations. The code of the site is at upper left, the event number at upper right. See the appendix for discussion of constraints on focal planes.

Correlation With Bathymetry

For each identified seismic site the local bathymetry was investigated, and an evaluation was made of a possible spatial correlation of seismicity with bathymetric features. We used *Mammerickx et al.*'s [1975] maps of the ocean floor and more recent and detailed charts in specific areas (e.g., *Mammerickx and Smith* [1978] in the eastern part of our study area). The extreme variability of shiptrack coverage considerably hampers this effort.

The use of Seasat radar altimeter data helps alleviate the poor charting of the southern seas by providing geoid signatures of long-wavelength uncompensated bathymetric features. In particular, it may be used to discover large seamounts and fracture zones or confirm their presence [*Sailor and Okal,* 1983]; it cannot, however, substitute for a detailed mapping of the bathymetry, especially in the presence of compensated features, such as island chains gener-

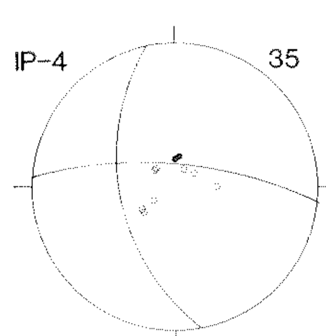

Fig. 3. Same as Figure 2*b* for the large 1955 event at IP-4 (number 35). See the appendix for discussion of constraints on planes.

ated on-ridge. In addition to individual Seasat tracks we used a variety of maps of the geoid [*Haxby,* 1982; *Francheteau,* 1983; *Sandwell,* 1984].

INDIVIDUAL SITES AND EVENTS

The appendix gives a detailed description of the seismic sites and events identified in the study area. New sites not previously identified were given codes starting with the letters "IP" (for intra-Pacific); their order follows loosely the boundary of the Pacific plate from east to south to southwest. Epicentral data are compiled in Table 1, and Figure 1 is a seismicity map.

We refer to *Okal et al.* [1980] for a complete description of the seismicity of the sites identified with the help of the French Polynesian network. We take, however, this opportunity to update their seismic history above the magnitude threshold $M_L = 4.7$. We also include two new sites (AU-7, TU-10) which were active above $M_L = 4$ and whose properties make them particularly interesting to the present discussion.

In discussing the possible correlation of epicenters with potential zones of weakness we occasionally refer to the "J" line defined by *Okal and Bergeal* [1983]. This line separates, on the present Pacific plate, lithosphere generated at the Farallon Ridge before the Miocene ridge jump from newer lithosphere generated at the East Pacific Rise. It is a line of age discontinuity and may represent a zone of weakness in the plate. Additionally, and because the jump was accompanied by a change in the azimuth of spreading, the J line separates two portions of lithosphere with differing major tectonic orientations; this is particularly well seen for fracture zones which trend 250° in the older lithosphere and 290° east of the J line. Because the jump took place during a period of relatively few magnetic reversals, its mapping is difficult, and *Okal and Bergeal* [1983] could only give "error bands" where it has to lie.

We list below data for the sites exhibiting the most interesting properties. All other technicalities justifying the further discussion can be found in the appendix.

GB-5 and GB-8

The long swarm at GB-5 (also called "region C" by *Okal et al.* [1980]) ended in June 1979, as abruptly as it had started 3 years earlier, and no seismicity was recorded at the magnitude 4 level in this area until events 28 ($m_b = 6.0$) and 29 ($m_b = 5.4$) took place on July 31, 1983. These two earthquakes are located at a new site (GB-8), approximately 65 km north of the previous cluster of epicenters, which *Jordan and Sverdrup* [1981] had found to be no larger than 15 km, with confidence ellipses generally oriented ESE-WNW. *Okal and Bergeal* [1983] have shown that region C is in the immediate vicinity of the "J" line, and on the basis of shipboard and satellite bathymetry, *Cazenave and Okal* [1983] have proposed that the Miocene jump has involved rift propagation in the immediate vicinity of region C. This would explain some rugged small-scale bathymetry observed at the site by J. Francheteau (personal communication, 1981) and the concentration of the seismicity at what would be a zone of weakness in the plate. The obvious distinction between the swarmlike behavior of the 1976-1979 activity at

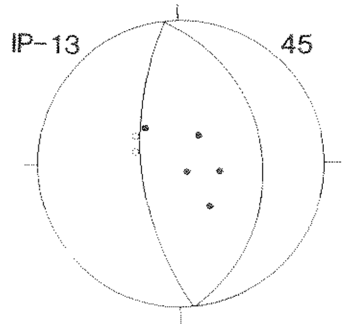

Fig. 4. Same as Figure 2*b* for the large 1947 event at IP-13 (number 45). See the appendix for discussion of constraints on planes.

GB-5 and the two isolated events at GB-8 in 1983 suggests different regimes of fracture for the rock at the two sites, possibly because of a more complex tectonic history at the southern location, resulting in lower yield stresses and a larger number of smaller events.

Focal mechanisms for event 28 were recently published in the monthly PDE bulletin. The U.S. Geological Survey solution involves a strike $\phi = 128°$, a dip $\delta = 36°$, and a slip $\lambda = 277°$; the Harvard solution is somewhat different: $\phi = 159°$, $\delta = 22°$, and $\lambda = 286°$. On the basis of body wave modeling of the Seismic Research Observatory data set, completed by available WWSSN and Polynesian records, we prefer $\phi = 150°$, $\delta = 40°$, and $\lambda = 270°$. The best modeled depth is between 15 and 20 km. The focal mechanism is included in Figure 2*b*. The most striking feature of this mechanism is a complete departure from the solutions at GB-5 obtained by *Okal et al.* [1980]. Event 28 is pure normal faulting with a horizontal tensional axis oriented N60°E. The compressional axis is quasi-vertical. In addition, body wave modeling requires a substantially greater depth than at GB-5. Mantle hypocentral velocities are clearly required to fit body wave constraints in Polynesia.

IP-4

The event of November 22, 1955, is one the largest normal faulting earthquakes known in young oceanic lithosphere ($m_b = 6.8$; $M_s = 6.2$). According to *Mammerickx et al.*'s [1975] map the epicenter lies on the northwestern flank of a major seamount about 100 km in diameter and having two summits less than 1200 m below sea level (see Figure 5). This feature is confirmed on individual shiptracks (J. Mammerickx, personal communication, 1983) and from Seasat data (see Figure 6).

IP-13

With a surface wave magnitude of 6.9 the 1947 event is the largest ever recorded in the Pacific plate (with the exception of the 1975 Kalapana, Hawaii, earthquake, which resulted from a major volcanic intrusion), and thus a detailed study is warranted. As shown on Figure 7, relocation displaces the earthquake approximately 80 km to the northeast, an epicenter shared by event 46 and probably also by event 47. This makes this region the most active seismic zone in our study area. The thrusting nature of the

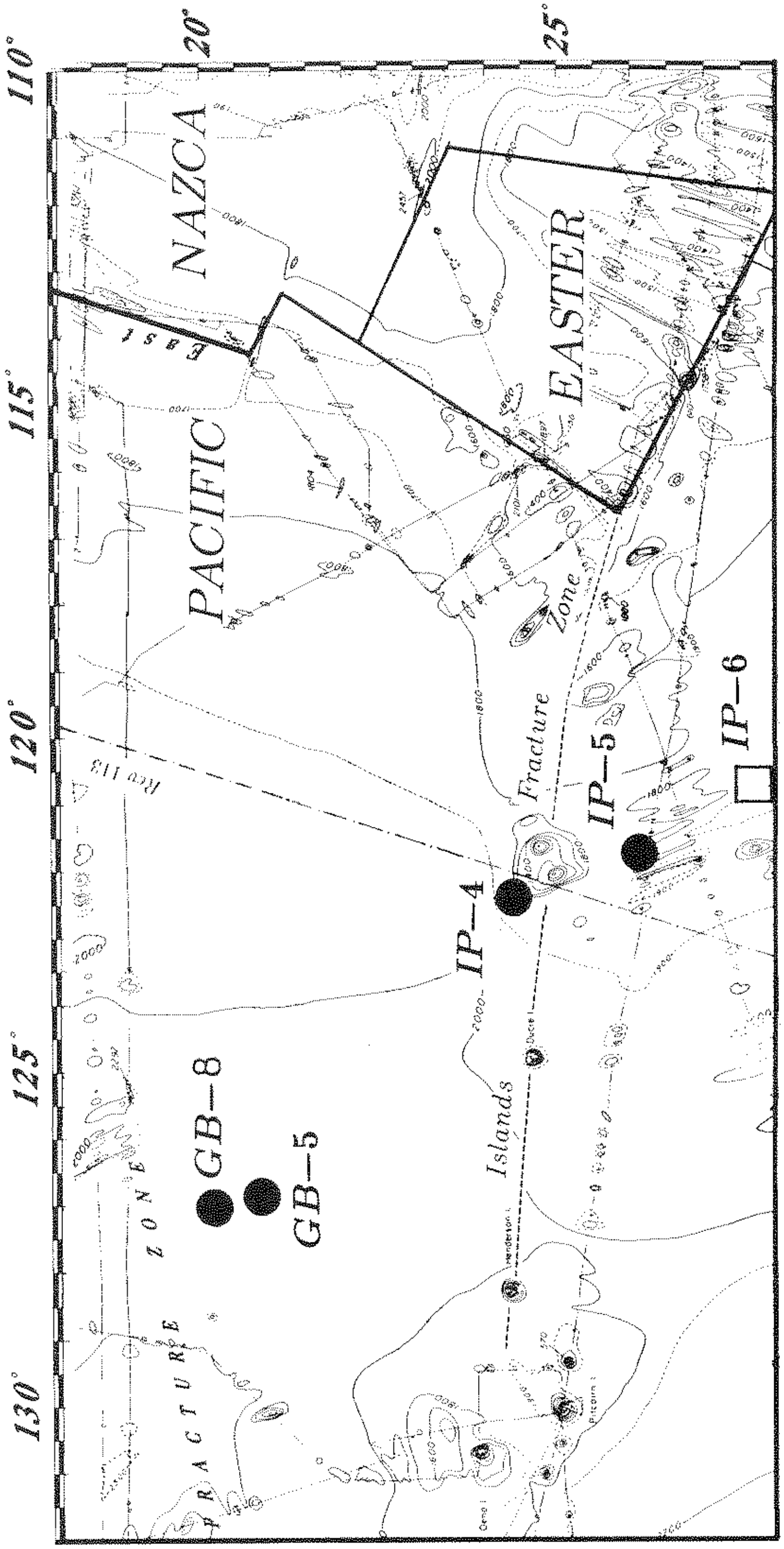

Fig. 5. Bathymetry of epicentral region IP-4 from *Mammerickx et al.* [1975]. Thick lines show plate boundaries. Dots are well-located seismic sites defined in this study. The open square is the unrelocated epicenter IP-6. The dashed line shows fracture zone FZ2, identified by *Cazenave and Okal* [1983], and the dash-dot line is the path of Seasat rev 113, whose profile is shown on Figure 6.

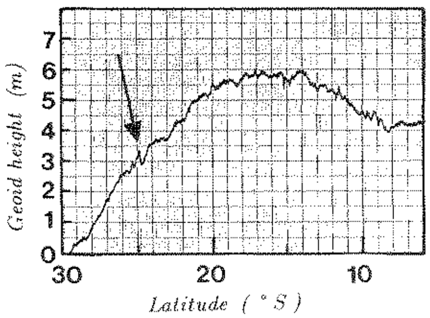

Fig. 6. Seasat radar altimeter data along rev 113 (see path on Figure 5). Relative geoid height is in meters; arrow identifies seamount next to epicenter IP-4.

mechanism is well constrained, as documented, for example, by the strong impulsive compressional P wave at La Paz (see details in the appendix). The axis of maximum compression, however, is oriented as much as 68° away from expected ridge push. To the west, IP-15 is the site of normal faulting (event 50).

IP-21: North of Samoa

The region extending between the Ellice, Tokelau and Samoa Islands is the locus of diffuse historical seismicity. Our computer search located eight events in this area. ISS data allowed only one (event 56, April 24, 1937) to be relocated (see the appendix) to the vicinity of the Robbie Ridge, a more or less continuous plateau ranging from the Tokelau Islands to Robbie Bank and the Wallis-Ellice lineament (see Figure 8).

If we assume that the other seven events are also intraplate, and on the basis of their NOAA locations, then the area north of Samoa appears as the site of substantial deformation of the Pacific plate. We include in Table 1 and Figure 8 only those events whose preliminary location was north of 10°S and which are thus most probably intraplate.

IP-22: Gilbert Islands

During 1981-1983 a seismic swarm involving 217 detected events with maximum magnitudes $m_b = 5.8$; $M_s = 6.0$ took place at the southeastern tip of the Gilbert Islands chain. No previous seismicity was known in the area. A description of this remarkable sequence, is given by *Lay and Okal* [1983], to which the reader is referred. Focal mechanisms for four of the largest events are reproduced in Figure 2*a*. The cumulative moment is estimated to be 1×10^{26} dyn cm, about 20% of the total moment for the whole study area (see below). Although it is unlikely that the origin of this swarm was volcanic, no interpretation of the events' mechanisms could be found in the context of simple theories; in particular, their common compressional stress axis is oriented horizontally but perpendicular to ridge push.

Preliminary Data in the Northern Pacific Plate

Preliminary data on the seismicity of the northern part of the Pacific plate (Hawaii excepted) reveal only two earthquakes reported with at least one magnitude ≥ 6: a 1970 thrust event $M_s = 6.8$ in the immediate vicinity of the East Pacific Ridge and discussed by *Bergman and Solomon* [1980] and a 1945 earthquake, with $M_{PAS} = 6\,¾$, at 17°N, 116°W, about 750 km from the ridge. As discussed below, we emphasize that these events take place in young lithosphere and once again that magnitude 6 seismicity not directly connected with a major volcanic edifice is absent from most of the plate.

DISCUSSION

Cumulative Seismicity

In order to compute an estimate of the cumulative seismic release in the southern part of the Pacific plate during the period of our study we converted available magnitudes to seismic moment using the formula

$$\log M_0 = 1.5M + 16.1 \quad (1)$$

In order to convert seismic magnitudes to the additive observable seismic moment it is necessary to use a nonsaturating magnitude scale, such as the energy magnitude M_w defined by *Kanamori* [1977]. For the range of magnitudes of our events this is equivalent to the surface wave magnitude M_s. In cases when only a body wave magnitude is available we use m_b. For a few earthquakes, dating back several decades, no magnitude is reported and we could not obtain records; we assign these events a magnitude of 5, representing the detection threshold in this part of the world before the development of the WWSSN. In view of these sources of error, the figure obtained, 5×10^{26} dyn cm, should be interpreted only as an order of magnitude of the total seismic moment release, and certainly no more than one digit can be significant. In this respect, the use of *Wiens and Stein*'s [1983] somewhat different relation between moment and magnitude

$$\log M_0 = 1.07M_s + 18.8 \quad (2)$$

would not alter our results significantly. Although the oldest events involved took place in 1937, our experience in other remote areas [*Okal,* 1981; *Stewart and Okal,* 1983] as well as *Gutenberg and Richter*'s [1954] compilation suggest that events of magnitude 6 would have been detected as early as 1920 and that the area was probably quiescent at this level during 1920-1937, as it was during 1955-1970. The average moment release rate during 1920-1983 is then 3×10^{10} W, and the average moment release rate per unit area of plate 7×10^{-4} kg/s^3. In more practical, non-SI units this represents 2×10^{17} dyn cm/(yr km^2), a figure in perfect agreement with *Wiens and Stein*'s [1984*a*] estimate of the average intraplate activity of the world's oceans. It is important, however, to realize that the principal contribution to this cumulative figure comes from a few large earthquakes. In particular, events 35 and 45 contribute nearly half of the moment release. A study based on the years 1957-1981 would yield only 4×10^{16} dyn cm/(yr km^2), a figure significantly lower than the world's average because it would miss both the 1955 event at IP-4 and the Gilbert Islands swarm, while still spanning an apparently safe 25 years. Similarly, a frequency-magnitude regression of our

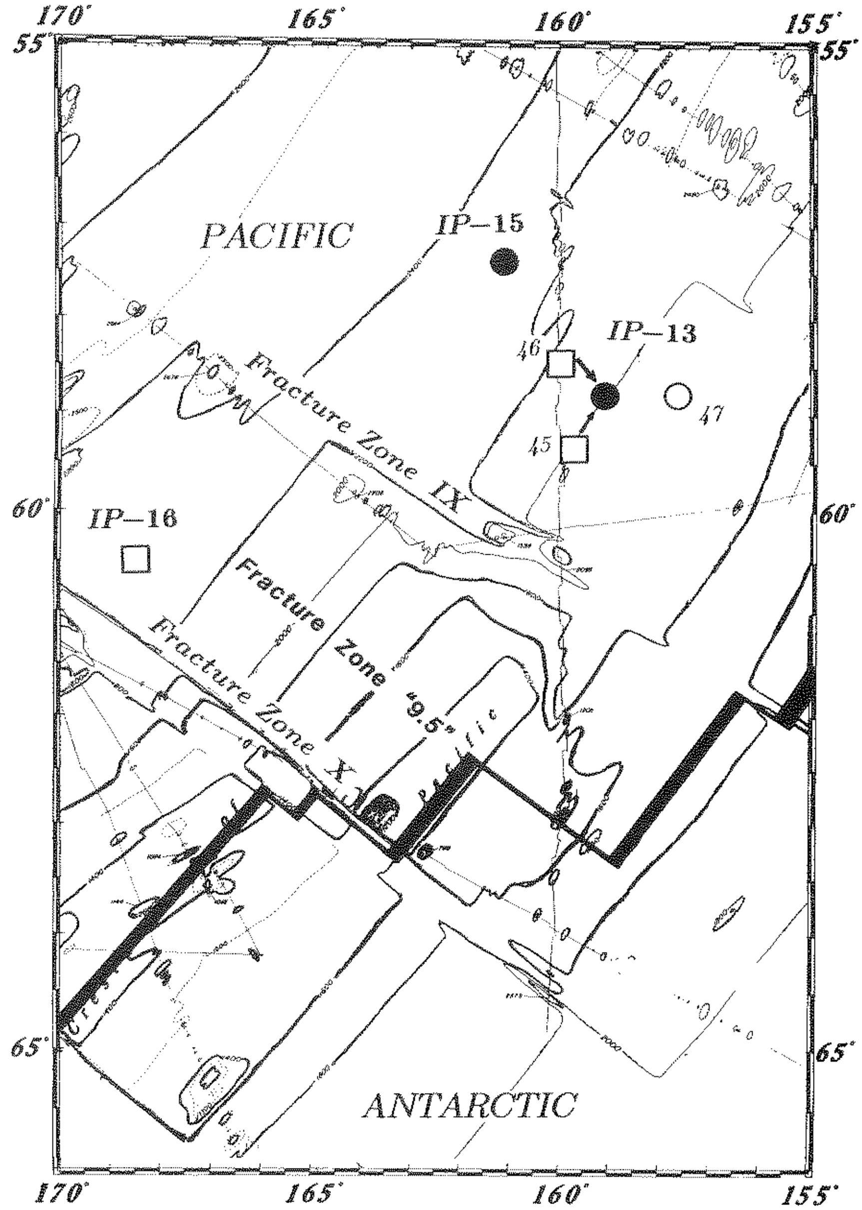

Fig. 7. Map of epicentral regions IP-13 and IP-15. The bathymetry (in particular fracture zones IX and X) is *Mammerickx et al.*'s [1975], based on *Molnar et al.*'s [1975] data. On the basis of seismicity and Seasat data we prefer to map the plate boundary with an intermediate offset (fracture zone "9.5"), as shown by the thick line. Solid dots are well-located epicenters. The open squares are the initial locations of events 45 and 46 converging on IP-13. The open circle is the proposed location of event 47. Also shown is the unrelocated event 51 at IP-16 (possibly on FZ 9.5).

whole data set (not including the older events for which no magnitude is available; conceivably this biases the results slightly toward low b values) yields $b = 0.6$, while the 1957-1981 sample requires $b = 1.25$, a figure comparable to *Bergman and Solomon*'s [1980] results for their entire Pacific plate catalogue ($b = 1.32$). It is clear that the shorter time interval does not give an proper picture of the seismic properties of the interior of the plate.

The use of *Gutenberg and Richter*'s [1954] formula

$$\log E_s = 1.5M + 11.8 \tag{3}$$

yields an average seismic energy output of 3×10^{-8} kg/s^3, or 1×10^{13} ergs/(yr km^2), a figure comparable to our results for both the Antarctic and African plates.

The dependence of seismicity upon lithospheric age is not a clear-cut one: in general, our results uphold *Wiens and Stein*'s [1983] observation that most of the seismic moment release takes place at ages less than 30 Ma. As expected from our experience with the French Polynesian network at a lower magnitude level, diffuse activity does, however, exist consistently throughout the plate (e.g., at sites IP-9, IP-19, and IP-20). Furthermore, the older parts of the plate (such as sites IP-18 and IP-20, where the lithosphere is more than 80 Ma old) are not immune to large earthquakes, a situation similar to that of other plates (see, for example, the $M_s = 6.1$ Bermuda earthquake of 1978 in lithosphere 117 Ma old [*Stewart and Helmberger*, 1981]).

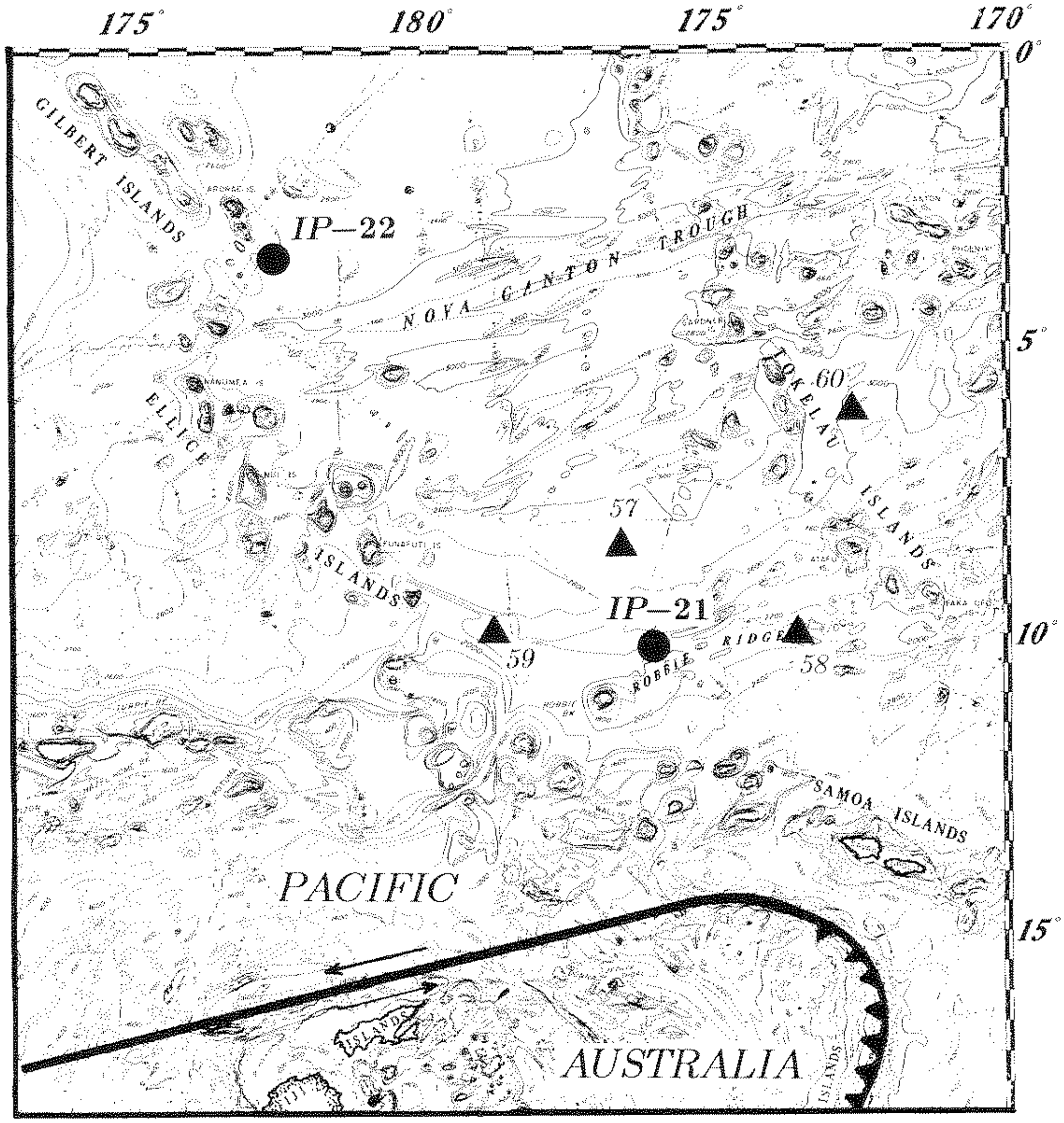

Fig. 8. Bathymetry of the epicentral areas IP-21 and IP-22 [after *Mammerickx et al.* 1975]. The boundary between the Pacific and Australian plates is shown as the thick line. The solid dots are the well-located sites IP-21 and IP-22. The triangles are the unrelocated events whose initial latitudes are at least 10°S. The whole area south of 12°S is the site of diffuse shallow seismicity.

Correlation With Bathymetry

Among potential lines of weakness of the plate, healed portions of fracture zones have been recognized early on [*Sykes,* 1978]. In addition, *Okal* [1981] proposed to interpret as such a line of maximum age in the panhandle of the Antarctica plate and *Okal and Bergeal* [1983] later hinted that a third type of weak line may be the J line separating portions of the Pacific plate generated during different spreading regimes. On the basis of these several types of weak lines we correlate epicenters LI-1, GB-5, GB-6, GB-8, IP-2, IP-7, IP-8, IP-12, and, possibly, IP-22 with zones of weakness of the plate.

Furthermore, epicenters LI-0, IP-4, IP-5 (and possibly IP-6), IP-17 and IP-21 are associated with major bathymetric features (seamounts or ridges). Site IP-20 lies in an uncharted area but is aligned with a linear trough documented 150 km to the northwest. Epicenter IP-14 is so poorly constrained that it would be futile to say anything about its bathymetry; IP-16 is very poorly located but a major fracture zone runs less than a 100 km away; IP-11, IP-18 and IP-19 are too poorly charted for a definite conclusion. Satellite data at IP-13 and IP-15 clearly show the lack of a major bathymetric feature. For LI-2, IP-9, and IP-10 there is sufficiently good bathymetric data to reject correlation with a major seamount, ridge, or fracture zone. As a whole, we can associate 15 out of 34 sites with either a zone of weakness or a major bathymetric expression. This proportion generally agrees with *Bergman and Solomon*'s [1980] results. However, and as proposed by *Sverdrup and Jordan* [1979] for LI-2 which they surveyed in detail, not all Pacific intraplate earthquakes develop on zones of evident weakness; our data set confirms this observation, since we reject a correlation at five sites. Also, we do not find any

TABLE 2. Focal Mechanism Data

		Magnitudes			Focal Mechanism										
					Plane 1		Plane 2		*P* Axis		*T* Axis		Plate		
Site	Number	m_b	M_s	Type	Az.	Dip	Az.	Dip	Dip	Az.	Dip	Az.	Age, Ma	AM1-2 Az.	Ref.
LI-2	3	4.9		SS	232	85	141	80	11	97	4	6	73	298	a
LI-2	5	5.1		SS	241	75	146	73	23	104	1	13	73	298	a
LI-2	7	4.9		SS	253	75	159	75	21	116	0	26	73	298	a
LI-2	8	4.9		SS	265	80	169	60	28	131	13	34	73	298	a
AU-6	10	5.3		T	?	?	?	?					41	296	b
GB-4	14	5.5		SS	84	69	345	68	31	305	1	214	34	294	a
GB-4	15	5.0		SS	103	69	3	66	33	324	2	232	34	294	a
GB-4	17	5.0		SS	115	63	22	84	23	335	14	71	34	294	a
GB-5	19	5.5		T	62	67	?	?	<20				20	292	a
GB-5	23	5.4		SS	62	90	152	90	0	107	0	17	20	292	a
GB-8	28	6.0	5.3	N	150	50	330	40	85	60	5	240	20	292	b
GB-6	30	5.0		SS	190	85	281	80	3	236	11	145	25	292	b
IP-1	31	6.1		SS	10	66	107	75	6	237	28	330	15	287	b
IP-4	35	6.8	6.2	N	162	50	275	65	49	136	9	35	10	290	b
IP-7	38	4.9		SS	29	67	130	65	2	80	35	149	24	292	b
IP-13	45		6.8	T	335	30	155	60	15	245	75	65	12	313	b
IP-15	50	5.4		N	?	?	?	?					18	313	b
IP-22	G1	5.8	5.4	T	125	60	256	41	10	194	62	84	112	300	c
IP-22	G2	5.7	5.6	SS	163	90	253	86	3	208	3	118	112	300	c
IP-22	G3	5.6	5.6	T	140	60	254	55	3	198	50	104	112	300	c
IP-22	G4	5.9	5.7	T	120	54	256	46	5	189	66	89	112	300	c

All angles in degrees. References are a, *Okal et al.* [1980]; b, this study; and c, *Lay and Okal* [1983]. Types of mechanism are T, thrust; N, normal faulting; and SS, strike slip. AM1-2 azimuth refers to the plate's absolute motion [*Minster and Jordan*, 1978].

systematic influence of earthquake magnitude on the correlation (or lack of it) between epicenters and bathymetric features.

Finally, it is interesting to note that the new site TU-10 falls right on the seismic line running from TU-9 to GB-4 and identified by *Okal et al.* [1980] along the northeastern flank of the Tuamotu archipelago. The origin of this lineament of seismic epicenters is obscure, especially since this branch of the Northern Tuamotus was generated on-ridge [*Pilger and Handschumacher*, 1981], is fully compensated [*Talandier*, 1982; *Cazenave and Okal*, 1983], and therefore should not be loading the plate as do, for example, the Hawaiian Islands. *Artyushkov*'s [1973] model of a compensated plate with varying thickness does predict local stress variations of a few hundred bars. However neither the pressure nor tension axes of the focal solutions observed at GB-4 can be reconciled with this model, which would require compressional stress perpendicular to the margin. Thus the seismic line is simply a weak line, along which release of the larger scale ridge push stress is preferentially located. In order to explain the absence of a similar line of seismicity along the southern flank of the archipelago one could invoke an asymmetric structure or ongoing tilting of the island chain about its axis.

Focal Mechanisms in Relation to Age

Table 2 presents focal mechanism data for 20 events discussed in this paper. The ages of the lithospheric plate in the epicentral areas were estimated from *Pitman et al.*'s [1974] map, completed by *Molnar et al.*'s [1975] data in the southernmost Pacific. These ages are approximate, since in some instances they had to be extrapolated to uncharted or magnetically quiet areas. Question marks indicate that the nature of the mechanism (thrust, normal or strike-slip) was inferred, but that no constraints could be put on the focal planes. As a whole, we obtained 12 solutions with dominant strike slip, six thrust events, and three earthquakes requiring normal faulting. The seven events at LI-2 and GB-4 were part of *Wiens and Stein*'s [1984*a*] data set. Figure 9 is comparable to their Figure 1, and shows a histogram of focal mechanisms as a function of age, both in terms of number of events and seismic moment release.

One of the most intriguing findings is that the mechanism of faulting can occasionally vary among earthquakes at the same epicenter and over a period of only a few months. Dramatic examples include the four Gilbert Islands earthquakes investigated by *Lay and Okal* [1983], and the two GB-5 events studied by *Okal et al.* [1980]. In both cases however, and significantly, the differing focal solutions share a horizontal compressional axis. This indicates that very local conditions involving small-scale tectonics control the final geometry of the rupture, even though the focal mechanism may remain representative of the large-scale stress field in the plate. *Lay and Okal*'s [1983] detailed study at the Gilbert Islands site also suggests that the isolated strike-slip event is 5 km deeper than its thrust counterparts. This could be due to the increased vertical compressional stress created by the weight of the overlying crust, which would inhibit the release of vertical tensional stress, resulting in a strike-slip geometry to achieve the release of the horizontal compressional stress.

It would then be expected to find no clear-cut evolution of focal mechanism geometry with age. On the basis of three new earthquakes we do confirm that normal faulting is found only in young material, but the repartition of thrust and strike-slip faulting is more random. Indeed, *Wiens and Stein*'s [1984*b*] data show that except in the Indian Ocean, one does not observe a regular change from normal faulting to strike slip and thrust as a simple function of age. A distance of only 65 km separates sites GB-5 and GB-8, where all three types of mechanisms are represented. Our results at sites IP-13 and IP-15 even point to a reversed situation: the younger lithosphere is the site of the large thrust event at IP-13, while the older site at IP-15 features normal faulting.

The orientation of the principal axes is more difficult to discuss, since even fewer events had their focal planes fully constrained. *Okal et al.* [1980] had indicated that their solutions at LI-2, GB-4, and GB-5 were, in general, compatible with the release of horizontal compressional stress oriented parallel to the motion of the plate away from the East Pacific Ridge and stemming from the gravitational forces known as "ridge push", a conclusion generally upheld for ages greater than 35 Ma by *Wiens and Stein* [1983]. We have no new fully constrained data confirming this, but the thrust mechanism at AU-6 is in principle compatible with this trend. On the other hand, the mechanisms at the Gilbert Islands site clearly violate it.

In younger lithosphere we do find compressional axes oriented at an angle from the direction of plate motion away from the ridge varying between only 5° at GB-5 (event 23) and 68° at IP-13 (see Table 2), without any simple dependence on age. The GB-8 and IP-15 mechanisms are normal faulting. It is not clear at this point whether the stress governing the mechanisms is still the compressional gravitational ridge push, adjusted locally to the particular geometry of the weakness of the epicentral zone (this would be particularly unlikely in the case of IP-13), a large-scale tensional stress (also horizontal in most cases, except IP-13) of a different physical origin (e.g., thermal contraction [*Turcotte and Oxburgh,* 1973]), or the result of purely local conditions. However, one common feature of the whole set of new data is the general absence of tensional stresses in the direction of spreading, confirming the fact that the extensional rifting regime is limited to the immediate vicinity of the spreading ridge [*Sleep and Rosendahl,* 1979; *Wiens and Stein,* 1984*a,b*].

Finally, we turn to the intriguing absence of magnitude 6 earthquakes in the bulk of the Pacific plate (apart from Hawaiian events, clearly deriving from local conditions). Assuming that ridge push is the main contributor to intraplate seismicity in older lithosphere, we propose to explain this phenomenon as resulting from the increased vulnerability of the plate, due to the reorientation of spreading which took place in the East Pacific in the Miocene: *Herron* [1972] presented detailed evidence for the deactivation of the Farallon Ridge and the development of the East Pacific Rise sometime between 22 and 18 Ma. As a result, most of the Pacific plate generated at the defunct Farallon Ridge is characterized by tectonic directions oriented about 250° and 340°. Fracture zones are obvious examples of such features on a large scale, but investigations on a scale of 10 km have confirmed these preferred tectonic azimuths for abyssal blocks [e.g., *Sverdrup,* 1981]. Ridge push in the older parts

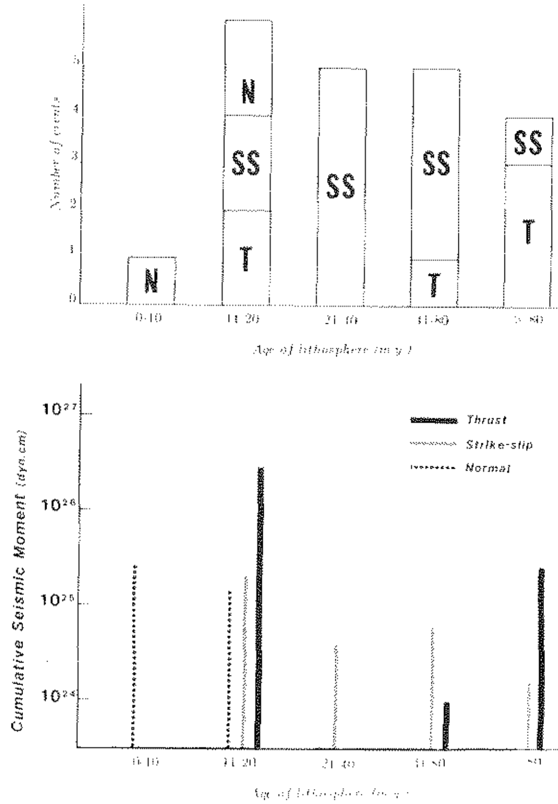

Fig. 9. (top) Histogram of the different types of focal mechanisms observed in the southern Pacific plate as a function of age. T, thrust; N, normal; SS, strike slip. (bottom) Cumulative seismic moment released as a function of age for the different types of mechanisms. This figure includes only events listed in Table 2: Other large earthquakes for which no mechanism could be constrained (e.g., event 18) could affect the observed patterns.

of the plate will be composed of the gravitational sliding of the old lithosphere, oriented parallel to its fracture zones, and of the transmitted forces from the new lithosphere, oriented parallel to the present direction of spreading, 290°. The direction of the resulting compressional stress will probably be age-dependent but grossly oriented between 260° and 290°. The faults along the old blocks of ocean floor at 250° and 340° will be oriented to yield preferentially in strike slip and at low levels of loading against such stress [*Raleigh et al.,* 1972; *Byerlee,* 1978]. On the opposite hand, in a more classic ocean where no reorientation has taken place the stresses will face the old abyssal blocks head-on, resulting in more difficult thrust faulting and therefore larger magnitudes (see Figure 10); magnitude 6 intraplate events are indeed current in all other oceanic plates [*Okal,* 1983]. We think that this mechanism is responsible for the large number of strike-slip solutions in the Pacific plate and for the absence of magnitude 6 events in the bulk of it. The only exceptions would be the parts of the plate which are young (then ridge push may not be the dominant stress [*Wiens and Stein,* 1984*b*]), or which did not undergo the reorientation, either because they postdate the jump or because they were not generated at the Farallon Ridge (e.g., the extreme southern portion of the plate generated at the Pacific-Antarctic or Pacific-Phoenix ridges). These areas are mapped on Figure 11, showing that they are (we believe

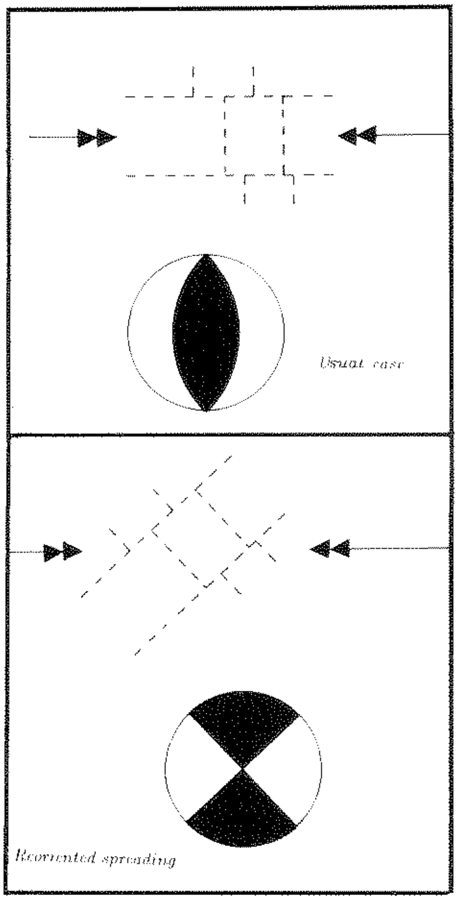

Fig. 10. This figure illustrates the increased vulnerability of the bulk of the Pacific plate, generated at the Farallon Ridge, to ridge push stresses. Dashed lines are fossil tectonic features, such as fracture zones and isochrones defining abyssal blocks; the double arrows represent ridge push stress. In the usual situation (top) not involving reorientation it is parallel to fracture zones, resulting in thrust faulting along the isochrones; this situation is found, for example, in the Nazca and Antarctica plates [*Mendiguren*, 1971; *Forsyth*, 1973]. In the portion of the Pacific plate generated at the Farallon Ridge (bottom) the fossil features are oriented close to 45° from ridge push, resulting in strike-slip events at lower magnitudes; this is the situation at LI-2.

not fortuitiously) the only sites of magnitude 6 seismicity in the plate. Thus we think that the intriguing absence of magnitude 6 events in most of the Pacific is due to the change in geometry of the accretion mechanism in the Miocene and does not necessarily involve the volcanic contributions invoked by *Bergman and Solomon* [1980] to account for their high b value of 1.32. In fact, b values obtained by *Okal et al.* [1980] at sites LI-2, GB-4, and GB-5 in the magnitude range 2.7 – 5.5 were if anything lower than the world's average; high b values were found by *Talandier and Kuster* [1976] and *Talandier and Okal* [1984] exclusively during swarms of activity at volcanic sites, and then only for specific episodes of the swarms, which totally escaped teleseismic detection.

A FEW INTRIGUING SITES: SUGGESTIONS FOR FUTURE STUDY

In this section we will try to interpret the characteristics of three of the major seismic areas identified in this study:

IP-4

Epicenter IP-4 is located in the immediate vicinity of an important seamount. This feature is conspicuously aligned with the islands of Oeno, Henderson, and Ducie (see Figure 5). Its signature on the geoid, about 40 cm, reflects partial compensation, because of the weakness of the young lithosphere on which it was generated. *Cazenave and Okal* [1983] also identified a fracture zone (FZ2) running through the islands of Henderson and Ducie and through the seamount as well as additional seamounts along the chain. East of IP-4, this system bends southward and probably branches into the southern boundary of the Easter platelet. These authors also confirmed the existence of the "region C" fracture zone identified by *Sailor and Okal* [1983], 300 km to the north, and with the opposite polarity. The common azimuth of these two fracture zones, 275°, suggests that a regime of spreading intermediate in direction between the old and new systems existed for a few million years in this area. An interpretation of the relationship of the islands, the plate, and the fracture zone, is given by *Cazenave and Okal* [1983] in the framework of the Miocene Farallon Ridge jump. It shows that IP-4 is a young seamount with a probable age of 4 ±1 Ma, which may still feature active volcanism comparable to the posterosional series observed in Hawaiian chains.

Two avenues are then provided for the interpretation of the normal faulting at IP-4: in a purely tectonic model the origin of the axis of tensional stress of event 35 at IP-4 (trending about N35°E) could be attributed to causes such as thermal stresses due to the shrinking of the plate as it cools away from the ridge [*Turcotte and Oxburgh*, 1973]. FZ2 could then be the weak line along which the rupture takes place preferentially. There are several problems with such a model: the fracture zone is actually 50 km south of the epicenter, and this distance is too large to associate the event with it; there is some indication that this event is deeper than usual (from the need of a mantle velocity to fit the focal mechanism constraints); and finally, the size of the event (m_b = 6.8; M_s = 6.2) makes it about an order of magnitude larger than all the adjoining known seismicity and the biggest normal fault earthquake in young oceanic lithosphere. Similarly, it is difficult to draw a direct parallel with the earthquake on the Emperor Trough interpreted by *Stein* [1979] as continued moderate activity on a supposedly fossil feature because of both the size of the event and the distance to the fracture zone. The same problems arise when trying to compare it with the Chagos events [*Stein*, 1978] whose focal mechanisms also involve normal faulting.

Another possibility, which presently remains speculative, would be to associate this event with a caldera collapse or other tectonic readjustment, following a volcanic eruption at the presumed volcano IP-4. Similar normal faulting has been documented at Deception Island along the Antarctica Peninsula [*Farmer et al.*, 1982], where a magnitude 6.9 earthquake followed an eruption by about 6 months, and on a smaller scale at Fernandina, Galápagos [*Kaufman and Burdick*, 1980]. In the case of the Deception event, both the magnitude of the earthquake and its distance from the summit would be comparable to the present case; the general orientation of the tensional axis would fit the geometry of the edifice. Also, earthquakes associated with intraplate volcanic edifices have been found to extend deeper than

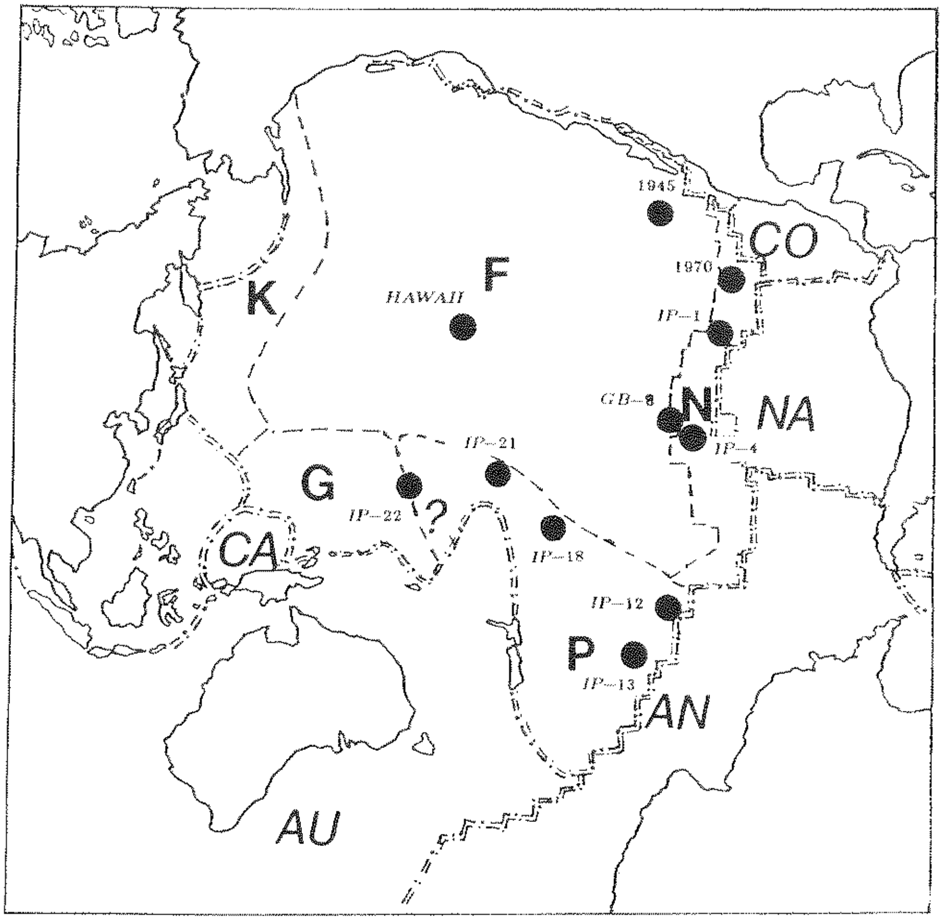

Fig. 11. Map of Pacific intraplate epicenters with at least one confirmed magnitude ≥ 6. This is a Mercator projection about the pole of rotation of the Pacific plate with respect to the Nazca plate; thus all transforms of the Pacific-Nazca system (and in practice, also, of the Pacific-Antarctic system) are horizontal lines. Double dashed lines are plate boundaries, with two-letter code for adjoining plates: CO (Coco), NA (Nazca), AN (Antarctica), AU (Australia), CA (Caroline). Single dashed lines delimit portions of the Pacific plate generated at various ridges with one-letter code: F (Farallon Ridge), K (Kula Ridge), P (Phoenix Ridge), N (New, postjump East Pacific Rise), and G (Gondwana-Pacific rise). The J line is, of course, the boundary between Regions F and N. Note that the only magnitude 6 or greater seismicity in "F" portion of the plate not associated with Hawaii is in very young material.

other oceanic events [*Eaton and Murata,* 1962; *Butler,* 1982]. On the basis of our experience in French Polynesia with Macdonald volcano and the Tahiti-Mehetia area [*Talandier and Okal,* 1982, 1984] the volcanic eruption which is likely in the proposed interpretation could have gone totally unsuspected in the early 1950's in that part of the world. Pushing the speculation one step further, event 36 at IP-5 (and possibly 37 at IP-6) could have a similar origin, although at a reduced magnitude.

We propose to name the seamount south of IP-4 after Sherman Thomas Crough in tribute to his powerful insight into the intricate relationship between plate and volcano. It is clear that a systematic and multidisciplinary exploration of this site would greatly enhance our understanding of stresses in this and other young lithosphere, an adventure in science which he would have loved to pursue.

IP-13

Epicenter IP-13 was the site of the $M_s = 6.9$ thrust faulting event 45 in 1947, of event 46, and presumably also of event 47 (see Figure 7). As such, it has been the most active seismic area in the Pacific plate, with the exception of the Hawaiian hot spot, for the past 60 years. Shipboard bathymetry in this area is extremely sparse; *Molnar et al.* [1975] estimated the position of fracture zones IX and X from three shiptracks, but it is clear from their magnetic anomaly map that the exact location of the necessary offset can be moved ~200 km between IX and X. The recent map of the *American Association of Petroleum Geologists* [1981] plots a large transform fault about half way between IX and X, in agreement with the seismicity data. On this basis, Figure 7 shows the boundary substantially offset along this feature which we name "fracture zone 9.5". This model of the plate boundary is also preferred on the basis of Seasat data, made abundant by the proximity of the 71°S turning point [*Sandwell,* 1984]. The age of the lithosphere at IP-13 is about 12 Ma, as estimated from the profiles of *Molnar et al.* [1975]; IP-15 is less well charted, but an assumption of symmetric spreading in this area leads to an age of 17 Ma. Seasat fails to identify notable features in the geoid, as does *Anderson et al.*'s [1973] map of gravity anomalies. We must conclude that whatever tectonic features are responsible for the local stress, they must be fully compensated. The whole area is evidently another excellent target for more detailed exploration.

The Gilbert-Samoa Region: IP-21 and IP-22

As discussed by *Lay and Okal* [1983], there is no interpretation of the compressional stresses released during the Gilbert Islands earthquakes at IP-22 in the context of simple plate tectonics theory. Given the large cumulative seismic moment, the constancy of the compressional axis, and the absence of any large anomaly in the geoid and gravity data of the area, a volcanic origin for the swarm is unlikely. The compressional stress released could perhaps be related to the indentation of the Pacific plate by the Australian one at the northern corner of the Vanuatu trench. If a Gilbert-Samoa line of seismicity could be confirmed, deformation in this part of the Pacific plate would indicate a definite non-rigid behavior, possibly in the form of the early stages of the carving out of a new platelet, reminiscent of the Caroline plate farther west [*Weissel and Anderson,* 1978]. A program of long baseline geodetic measurements, taking advantage of the numerous islands in the region, would shed some light on these problems.

CONCLUSIONS

1. We have compiled all known historical seismicity in the southern part of the Pacific plate, defining 32 epicenters, at which about 60 events took place above a magnitude threshold of 4.7, during the period 1937-1983 (excluding the Gilbert Islands swarm). In addition, some sites have been considerably more active at lower magnitudes.

2. When compiled over 60 years, the cumulative level of seismicity is comparable to the worldwide averages of *Wiens and Stein* [1983] in terms of seismic moment and also to our previous results in other, continent-bearing plates in terms of seismic energy release. Magnitude 6 and greater earthquakes are confined to the fringes of the plate, a situation which we think is due to the more vulnerable geometry of old tectonic features achieved in the plate by the reorientation of spreading at the East Pacific Rise in the Miocene.

3. The falloff of seismicity with age noticed by *Wiens and Stein* [1983] is confirmed by this extended data set but does not preclude existence of occasional large earthquakes in older lithosphere. Correlation of epicenters with zones of weakness and/or large bathymetric features is frequently observed but not universal, some areas with adequate bathymetric coverage failing to show any correlation.

4. Local features play an important role in determining the exact geometry of focal mechanisms. No simple interpretation of the stresses released in the youngest parts of the plate is forthcoming.

5. We identify three areas having released 2/3 of the cumulative seismic moment; mechanisms of their seismicity are not fully understood and they should make excellent targets for exploration and future investigation.

APPENDIX: DETAILED DESCRIPTION OF SITES AND EVENTS

We provide here the details of the relocations and focal mechanism constraints justifying the data sets given in Tables 1 and 2 as well as comments on some of the less active sites.

Polynesian Sites

TU-10. This is a new seismic location at the northeastern end of the Tuamotu archipelago. A single event occurred on August 16, 1982, but was not recorded teleseismically. The location is taken from J. Talandier (personal communication, 1983). It is remarkable that the site fits together with TU-9, GB-1, GB-2, GB-3, and GB-4 (and, indeed, also GB-5) within $\pm 1°$ of a small circle about the Nazca-Pacific pole. See main text for a discussion of a possible origin of this lineament.

GB-6. This remains a single event epicenter. We were able to read nine first motions and have two emergent ones on high-gain instruments; this suggests the strike-slip mechanism shown on Figure 2*b*, with the compressional axis oriented 245°, about 65° away from the direction of ridge push. The site is located within the range of the J line [*Okal and Bergeal,* 1983].

AU-6. Only two earthquakes were recorded at this site. No swarm comparable to the activity at AU-7 was recorded. Available first motions for event 10 (November 20, 1979) clearly require a strong component of thrust faulting but cannot resolve the fault planes. Seasat tracks in the area fail to reveal any substantial bathymetric feature.

AU-7. This is a new epicenter, 400 km north of AU-6. In addition to the single event listed in Table 1 it was the site of at least seven other earthquakes at the magnitude $M_L = 2$ level in a swarm lasting 2 months in 1981. Events at this site do not generate high-frequency Rayleigh waves of the type observed for AU-6, suggesting a somewhat deeper epicenter at AU-7: a depth increased by just a few kilometers would be sufficient to quench their excitation [*Harkrider and Okal,* 1982].

Other Sites

IP-1: Events 31 and 32. The first of the two events, on September 30, 1981, was well recorded at the WWSSN stations. First motions read on high-gain WWSSN records provide some constraints on the mechanism: most stations are compressional, with dilatations in Polynesia, at SPA and NNA, and the stations LPB, LPS, and AFI appearing nodal, despite operation at a gain of 50,000 for the former two. Thus the mechanism approaches strike slip, with a compressional axis oriented about 240°. The second event, on October 4, 1981, was too small for further study. The site is located about 880 km east of the Gallego Fossil Ridge, identified in the bathymetry by *Mammerickx et al.* [1980], and probably also east of the J line, that is on lithosphere generated at the new ridge system.

IP-2: Event 33, January 30, 1980. We use the ISC epicenter for this earthquake. Its location is in the immediate vicinity of the Marquesas Fracture Zone, and *Okal and Bergeal* [1983] have proposed that it may lie on the J line. This $m_b = 5.3$ event was poorly recorded. Polarities of first motions could be read only at SPA and DUG (both compressional).

IP-3: Event 34, November 11, 1949. No individual times are listed in the ISS; however, we obtained records with good body waves at LPB and PAS. The NOAA epicenter gives far too large a distance to LPB. On the basis of $S-P$ at LPB, and P at PAS, we obtain the preferred epicenter 9.0°S, 109.8°W, which places the earthquake on the Wilkes transform fault of the East Pacific Rise system. This makes this event interplate, and we delete it from Figure 1 and the further discussion. The expected mechanism, left-lateral strike slip, is consistent with emergent P waves at PAS and TUC, a small compression at LPB, and strong SH and Love waves at LPB.

IP-4: Event 35, November 22, 1955. This event, which is one of the largest tensional earthquakes ever recorded, is well located by the ISS and the NOAA tape, at 24.4°S, 122.6°W, and 24.5°S, 123°W, respectively. On the basis of 19 arrivals listed as impulsive in the ISS we obtain a preferred solution at 24.21°S, 122.77°W, with a standard deviation $\sigma = 1.48$ s. This solution is also compatible with a T wave arrival at Uwekahuna, Hawaii. Revised figures for body and surface wave magnitudes are $m_b = 6.8$; $M_s = 6.2$.

The mechanism of this event was found to be tensional by *Sykes and Sbar* [1974], who could not constrain it further. We obtained copies of original records at 12 stations and present a lower hemisphere focal mechanism on Figure 3. We first attempted to fit all the first motion data with a shallow focus, using a crustal source velocity, and obtained one solution which fits the polarity of all arrivals but would predict extremely low amplitudes at HON and USZ, which are not observed. On this basis we prefer a deeper source; using a mantle velocity, we obtain the mechanism shown on Figure 3. One plane (striking at 275° and dipping 65°) is well constrained by the compressional arrivals at PAS and BKS. The slip angle λ could in principle vary from 215° to 320°. A computer optimization of the focal mechanism yields $\lambda = 225°$ on the basis of large SV/SH ratios at CHR and PAS, large $PS/(SS)_{SV}$ at FLO, large PS at HON, and substantial PP at DBN. The solution also fits Rayleigh/Love ratios much larger than 1 at CHR. The resulting tensional axis dips only 9° at an azimuth of N35°E.

IP-5: Event 36, September 29, 1951. The ISS epicenter is confirmed by a relocation using 12 arrivals, and achieving $\sigma = 0.50$ s. *Sandwell* [1984] has identified in the Seasat geoid an uncharted seamount centered at 26.76°S, 121.63°W, only about 50 km away, a figure comparable to his estimated horizontal accuracy. We obtained only a few records, which cannot yield any focal mechanism. Given the low magnitude of the event, the large number of PKP reports in Europe would suggest a strong component of either normal or thrust faulting. As in the case of event 35, PAS and BKS have compressional first motions; additional data listed in the ISS could make the mechanism compatible with that of event 35.

IP-6: Event 37, August 3, 1951. This event is not listed in the ISS, and we obtained only one record (PAS short-period). Given the spatial and temporal relation of this earthquake with the previous one, it is possible that it was mislocated and was actually at IP-5. In particular, intervals $PcP - P$ from the two events at PAS are indistinguishable.

IP-7: Event 38, September 14, 1963. Relocation of this event using 12 times picked on WWSSN and southern California records yields an epicenter about 15 km northeast of the NOAA location. A clear short-period record at Woody, California, shows a phase following P by 1.2 s, suggesting a depth of only 3-4 km if interpreted as pP. All first arrivals which could be picked are compressional, including a PKP_2 at SHL. SBA and LPB may be nodal, suggesting the strike-slip mechanism shown on Figure 2*b*. This event falls in the range of uncertainty of the J line.

IP-8: Event 39, July 31, 1976. This isolated and small earthquake ($m_b = 4.3$) is located in the immediate vicinity of the Agassiz Fracture Zone and may also be related to the J line. It was rather poorly recorded, and no focal mechanism information could be gathered.

IP-9: Event 40, March 29, 1975. Despite an m_b of 5.0, first motions from this well-recorded event could not, in general, be read conclusively. A strong dilatation at SPA, and unverified but consistent reports of dilatational PKP in Europe would suggest a component of normal faulting.

IP-10: Events 41, July 30, 1958, and 42, September 28, 1972. These two events have listed epicenters whose difference is not significant: the 1972 earthquake is given a precision of ±10 km by the ISC, about 50 km north of the Menard Fracture Zone, 600 km east of the ridge. No records were obtained from the 1958 earthquake, and it could not be relocated, in the absence of arrival time data. The 1972 earthquake was too small to yield any valuable focal mechanism information.

IP-11: Event 43, July 14, 1951. The location of this shock is given as 52°S, 128°W by the NOAA tape; the ISS does not give a location; on the basis of eight reported P times, we relocate the event 143 km to the southeast, at 52.94°S, 126.57°W. The earthquake cannot be moved to the interplate Eltanin system without incurring root mean square (rms) residuals of 17 s or more. Numerous reports of PKP in Europe and of an impulsive P_{diff} at COL would argue against strike-slip motion, but this could not be checked or constrained further by the limited data which we obtained. A surface wave magnitude of 5.5 is obtained from New Zealand records.

IP-12: Event 44, September 5, 1938. The NOAA and ISS locations of this earthquake differ by about 270 km. Our relocation moves the epicenter significantly northward, to 54.74°S, 148.35°W, right on the healed portion of the Udintsev Fracture Zone. Interpreting this event as interplate would lead to an rms residual of more than 15 s, a figure clearly unacceptable. Additionally, the necessary strike-slip mechanism on the Udintsev transform fault would violate the large Rayleigh-to-Love ratios observed at PAS and TUC. The mechanism could not be constrained further.

IP-13: Event 45, December 15, 1947. The preferred epicenter (obtained from 13 ISS impulsive times and achieving $\sigma = 1.8$ s) lies about 200 km from the nearest plate boundary, assuming *Molnar et al.*'s [1975] model; constraining the event to be interplate on transform fault IX would result in an unacceptable $\sigma = 6$ s. Even higher residuals would be necessary to make the event interplate in the more realistic model involving fracture zone 9.5. Furthermore, as will be seen below, its thrust mechanism would be in violation of plate motions, both along the ridge, and on any major nearby transform fault.

Figure 4 shows a lower focal hemisphere representation of its mechanism. The thrusting nature of the mechanism is apparent from the excellent impulsive compressional records of $PKIKP$ at European stations. Only the nearby stations WEL and RIV have dilatational first arrivals. Note at this point that both a left-lateral strike-slip mechanism on a nearby transform and normal faulting along the ridge

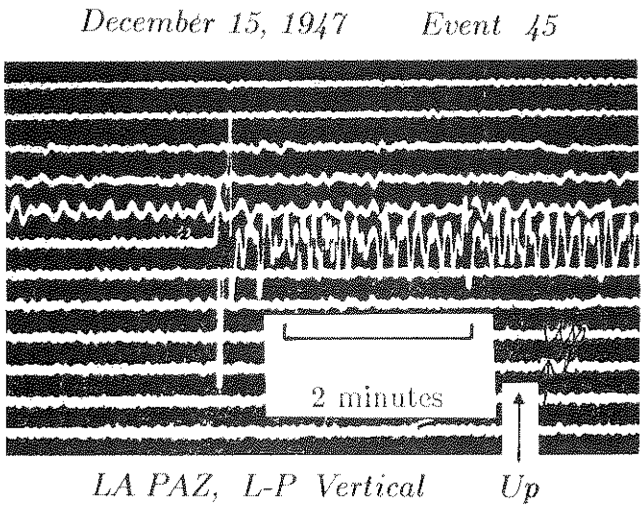

Fig. A1. Long-period P wave recorded at La Paz, Bolivia, from event 45 at IP-13. The excellent impulsive compression proves by itself that the event cannot be on any nearby transform fault (see the appendix for details).

(hardly common along the mid-oceanic ridge system at this magnitude, anyway), would lead to dilatational arrivals at LPB, in evident violation of the observed record (see Figure A1). One fault plane is constrained by the WEL, RIV and AUC arrivals. The other plane is less well constrained. Additional observations of strong *PP* and *PS* at TUC, strong *PP* at ANC, low *PP* and a large *SV/SH* ratio both for *S* and *ScS* at LPB as well as the virtual absence of any surface waves (both Love and Rayleigh) at HON, all favor a slip angle close to 90°. It is on this basis that the second plane is drawn on Figure 4, but the solution could be adjusted somewhat, and the range of variation of the pressure axis compatible with the above observations is 243° – 277° in azimuth and 11° – 19° in dip. The possible mechanisms generally express the release of horizontal compressional stress in the plate, although the azimuth of the pressure axis on Figure 4 is oriented about 68° away from the expected direction of ridge push.

Event 46, April 10, 1950. This event was not located by the ISS, but a few times are available. On their basis we relocate it to 58.70°S, 159.11°W, about 90 km southeast of its NOAA location, and at the exact 1947 epicenter, with a standard deviation of only $\sigma = 0.97$ s. Its mechanism could not be constrained, although the three identifiable first motions would be compatible with the 1947 mechanism.

Event 47, September 17, 1949. The location of this event is extremely poorly known. However, the NOAA epicenter, at 34°S, 154°W is clearly incompatible with many observations; since it is uncommon to relocate an earthquake 2700 km away from its listed location, we will detail the evidence on which we favor an epicenter close to IP-13.

First, the NOAA epicenter is grossly erroneous. For that epicenter, CHR, WEL, AUC, and SUV would be at 27°, 25°, 25°, and 29°, respectively. It is clear, however, from Figure A2 that the epicentral distance increases regularly and substantially from WEL to SUV as the recording station moves north along New Zealand and the Tonga trench. In particular, the S phase at SUV requires a distance of 40° – 44°, depending on the interpretation of the previous phase as *PP* or *PPP*. Additionally, the NOAA epicenter would predict back azimuths of 83° at CHR and 107° at RIV. This is in violation of the observation of Love waves with comparable amplitudes on both N-S and E-W at RIV and prominently on N-S at CHR. Finally, the pattern of *PKP* times at KSA and STR would also be violated.

Second, the travel times listed in the ISS are mutually incompatible. We could not achieve a relocation matching all impulsive P, let alone S times listed in the ISS. In particular, and on the basis of the RIV records which we obtained, the first arrival at this station (listed as *iZ* in the ISS) is not P but probably *PP*. Similarly, simultaneous use of the BRS and LPB P times cannot lead to a stable solution. Since we did not obtain any BRS records but where able to confirm many characteristics of the LPB ones, we chose to ignore the reported time at BRS.

Third, the earthquake is most probably intraplate. The abundance and coherence of *PKP* reports from stations in Europe and the Mediterranean area strongly advocate against placing the event along an active transform fault segment, where a strike-slip mechanism would be required. Normal faulting along ridge segments is not known in the southern Pacific at the $M_s = 6.2$ level of this event. Thus it is difficult to account for the core waves from this event, if it is allowed to be on the ridge, but the evidence is somewhat weaker than for the 1947 event. Nor can the earthquake be moved to the Macquarie Ridge, where large thrust events are known, without grossly violating the RIV epicentral distance, which must be about 42° on the basis of Love and S travel times; the Tonga-Kermadec trench would not fare any better, on the basis of the distance to SUV.

Finally, IP-13 is an excellent solution. A three-station relocation using only WEL, CHR, and LPB P times converges on 58.8°S, 157.7°W. This epicenter also fits S times at CHR, WEL, SUV, and RIV; Love times at RIV; the *iZ* time at RIV if interpreted as *PP;* Love and Rayleigh times at LPB; Rayleigh waves at TUC (all read from available records); and the reported impulsive *PKP* times at many stations in Europe, North Africa, and the Middle East; it is also compatible with back-azimuth constraints from Love and Rayleigh records. Only BRS is clearly violated (see above). This epicenter also puts the Southwestern U.S. stations (PAS, TUC, TIN, RVR) at distances close to or in the shadow zone, explaining the absence of body wave records at these sites.

On this basis we propose to associate this large earthquake with epicentral area IP-13, although this location keeps a somewhat tentative character. We could not gather enough data to constrain further its focal mechanism.

IP-14: Event 48, October 17, 1959. Two events are listed on the NOAA tape for this date. From the few New Zealand records available we cannot confirm the location of the first one, given by the BCIS as 54°S; 165°W.

IP-15: Event 49, October 17, 1959. Our relocation of the second event does not differ significantly from the ISS epicenter. The CHR records suggest a surface wave magnitude of $M_s = 5.5$; LPB clearly shows a dilatational arrival. Differences in the available data clearly indicate that the two events on this day cannot be in a foreshock-aftershock relationship.

Event 50, October 21, 1976. This earthquake is well located by the ISC, with a precision given as ±8 km; this makes its epicenter the same as event 49's. Thanks to its magnitude of $m_b = 5.4$, we were able to observe eight

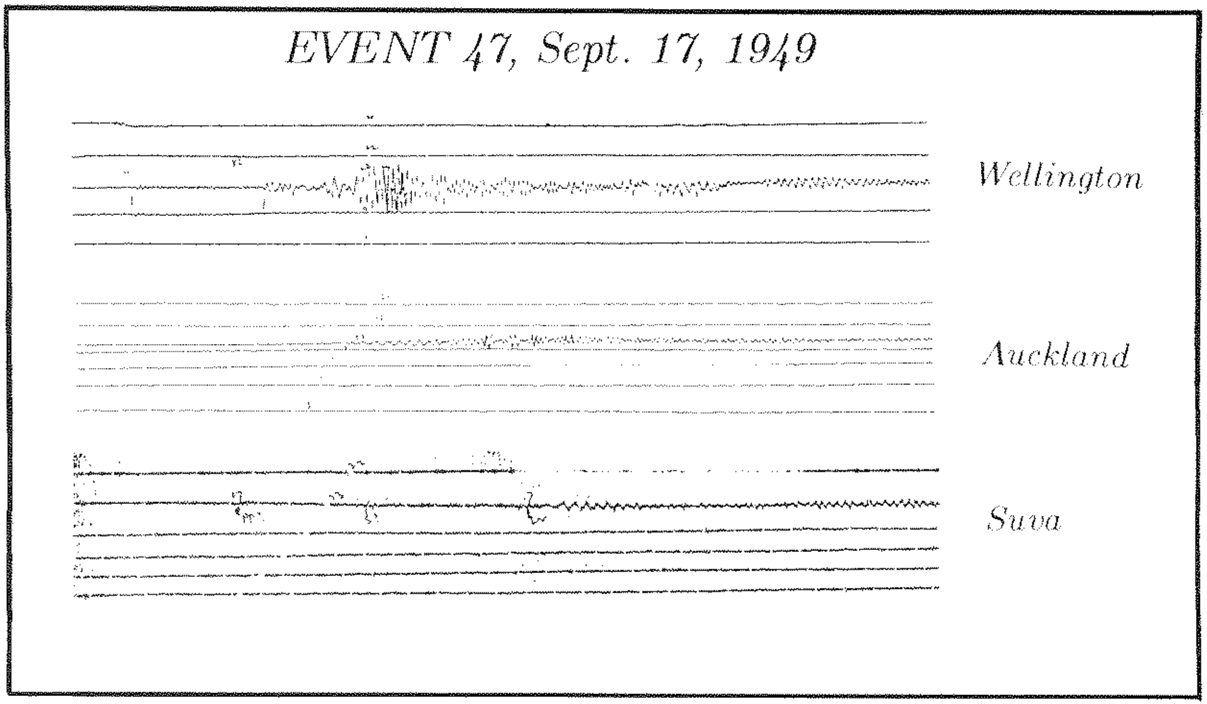

Fig. A2. Milne-Shaw North-South long-period records of event 47 at Wellington, Auckland, and Suva, Fiji (magnifications used are 250, 150, 150). Note delay in the whole seismogram, and interval between phases increasing progressively as station is moved north. These records require an epicenter located south of these three stations and are incompatible with the NOAA location.

WWSSN first motions, all dilatational, including *PKIKP* at EIL. This earthquake had to involve normal faulting, but no further constraint could be worked out (see Figure 2*b*).

The common epicenter of events 49 and 50 is an extremely poorly charted area, approximately 200 km northeast of IP-13. A systematic search of Seasat altimeter data has failed to evidence any substantial tectonic feature apparent in the geoid.

IP-16: Event 51, July 26, 1958. No additional information or data are available for this event. If the NOAA epicenter is correct, it would lie about 60 km NE of fracture zone "X" [*Molnar et al.*, 1975], and 250 km from the nearest ridge segment. It could be much closer to the more significant fracture zone "9.5".

IP-17: Event 52, August 1, 1953. This event was recorded only in New Zealand and located in the immediate vicinity of Chatham Island, a feature believed to be of continental origin [*Fleming*, 1970]. No additional data are available.

IP-18: Event 53, October 4, 1937. The original NOAA location of this earthquake is at 26.8°S, 163.1°W. The ISS did not compute an epicenter. *P* wave arrivals listed in the ISS do not lead to a stable relocation. In particular, the CHR arrival listed as *"PE?"* is several minutes early. Our inspection of the CHR records uphold a bulletin remark that it may really be microseismic activity. On the basis of four other stations it would be possible to relocate this earthquake in the Tonga Trench, at 23.1°S, 175.5°W; however, this solution violates several impulsive arrivals (e.g., MEL), and the later arrivals at CHR cannot be explained either. Rather, we prefer a relocation using only *P* arrivals listed as impulsive, and the CHR *"iSZ"* arrival interpreted as *P*; it places the epicenter at 22°S, 167°W. This solution is also compatible with an *eP* arrival at Wellington, identified on the N-S record which we obtained. Thus we think that this event is truly intraplate, with a poorly constrained epicenter about 300 km southeast of Niue. The magnitude of this earthquake is estimated to be $M_s = 5.8$ on the basis of the CHR records. Additional information on the epicentral area and background low-level seismicity at this site can be expected from the results of the Ngendie campaign during the Spring of 1983 [*Jordan et al.*, 1983].

IP-19: Event 54, January 13, 1938. Arrival times are not listed in the ISS bulletin. From records and station bulletins obtained, we used *P* arrivals at PAS, TUC, and CHR and *S* arrivals at HON and CHR to relocate this event several hundred kilometers southwest of the NOAA epicenter; it does not appear possible to send the epicenter to the Tonga trench without grossly violating the excellent arrivals at the southwestern U.S. stations. CHR records give an M_s of 5.8. No information on its focal mechanism could be gathered.

IP-20: Event 55, January 4, 1940. This event is listed by the ISS at 34°S, 162°W. Relocation using five stations confirms this epicenter with a standard deviation of only 1.3 s.

IP-21: North of Samoa. Only event 56, April 24, 1937 could be relocated. On the basis of seven stations we obtain a preferred epicenter at 10.14°S, 176.00°W, with $\sigma = 1.2$ s. It was not possible to put the earthquake on the plate boundary (defined as the zone of intense seismicity around 15°S; see *Chase* [1971] for details) without incurring a standard deviation of more than 20 s, even if a deep focus was allowed.

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