SEISMICITY AND TECTONICS OF THE CENTRAL NEW HEBRIDES ISLAND ARC

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Abstract. The seismicity of the central New Hebrides convergent plate boundary is investigated with three sets of data: (1) large earthquakes (Ms > 6.9) for the past 75 years, (2) moderate-sized earthquakes (mb > 4.5 and Ms < 7.0) during the past 20 years, and (3) small earthquakes (mb = 2.5 to 4.5) located by local networks for several intervals of 1-2 months each since 1975 and continuously since mid-1978. The second set includes a nearly complete collection of focal mechanism solutions for events with Ms > 5 3/4 and new determinations of accurate focal depths based on analyses of P waveforms recorded by WWSSN long-period seismographs. On a regional scale the geometry of the Benioff zone is relatively uniform. Within the resolution of the data there are no major disruptions of the descending plate nor changes in direction of plate convergence along the length of the arc. However, the three data sets all indicate marked variations in the seismicity patterns along the strike of the arc. These variations together with major bathymetric and structural complexities of the interacting plates divide the interplate boundary of the central New Hebrides into segments of about 100 km in length. The segments delimit episodes of seismic rupture but may also differ significantly in the long term balance of seismic versus aseismic slippage. In the segments near Santo and northern Malekula islands seismic rupture of the interplate boundary occurred in complex sequences of large earthquakes during the period 1965-1974, whereas in the segments near southern Malekula and Efate islands seismic rupture of the boundary may not have occurred during the past 75 years. The interaction of a subducted ridge with a pre-existing seaward protrusion of the upper plate may result in an increased coupling of the converging plates in the Santo-Malekula segments. The orientation of horizontal compressive stress within the upper plate inferred from focal mechanism solutions and geological data is perpendicular to the arc in the recently ruptured segments but changes in the region between southern Malekula and Efate islands to a more complex and variable pattern found in the southern New Hebrides arc. In addition, the diffuse spatial distribution of small earthquakes in the southern Malekula and Efate segments contrasts with the concentration of small events along the presumed interplate boundary of the recently ruptured Santo segment. The unusual concentration of events at shallow depths within the upper plate in the southern Malekula segment may be evidence for loading of a locked segment of the plate boundary. Unusual features of seismicity suggest that in the Efate segment a significant component of creep may accommodate interplate slippage. A persistently high rate of occurrence of small and moderate-size events (Ms < 6.5) in the Efate segment contrasts with the large fluctuations in activity associated with the major events in the recently ruptured segments. The persistent nest of activity in the Efate segment also contrasts with the relative quiescence in the adjacent segments. The most recent (1978-1979) sequence of three moderate-sized shocks (Ms≈6) located in the Efate segment was caught by the local networks of seismographs and tilt measurements. Well-documented features of the temporal and spatial development of foreshocks and aftershocks include a clear migration of foreshock activity towards the epicenter of one of the mainshocks. In two cases the area of the aftershock zone expands during a period of several days to a size significantly larger than that expected for the surface wave magnitudes and body wave seismic moments of the mainshocks. For these cases tiltmeter stations on Efate Island, located at distances of 40 to 60 km from the mainshock hypocenters, did not record tilt indicative of post-seismic creep. Nevertheless, periodic releveling of a 1 km benchmark array on Efate reveals that a significant tilt of 3-4 microradians has accumulated during the four year interval between 1976 and 1980. This tilt signal could be

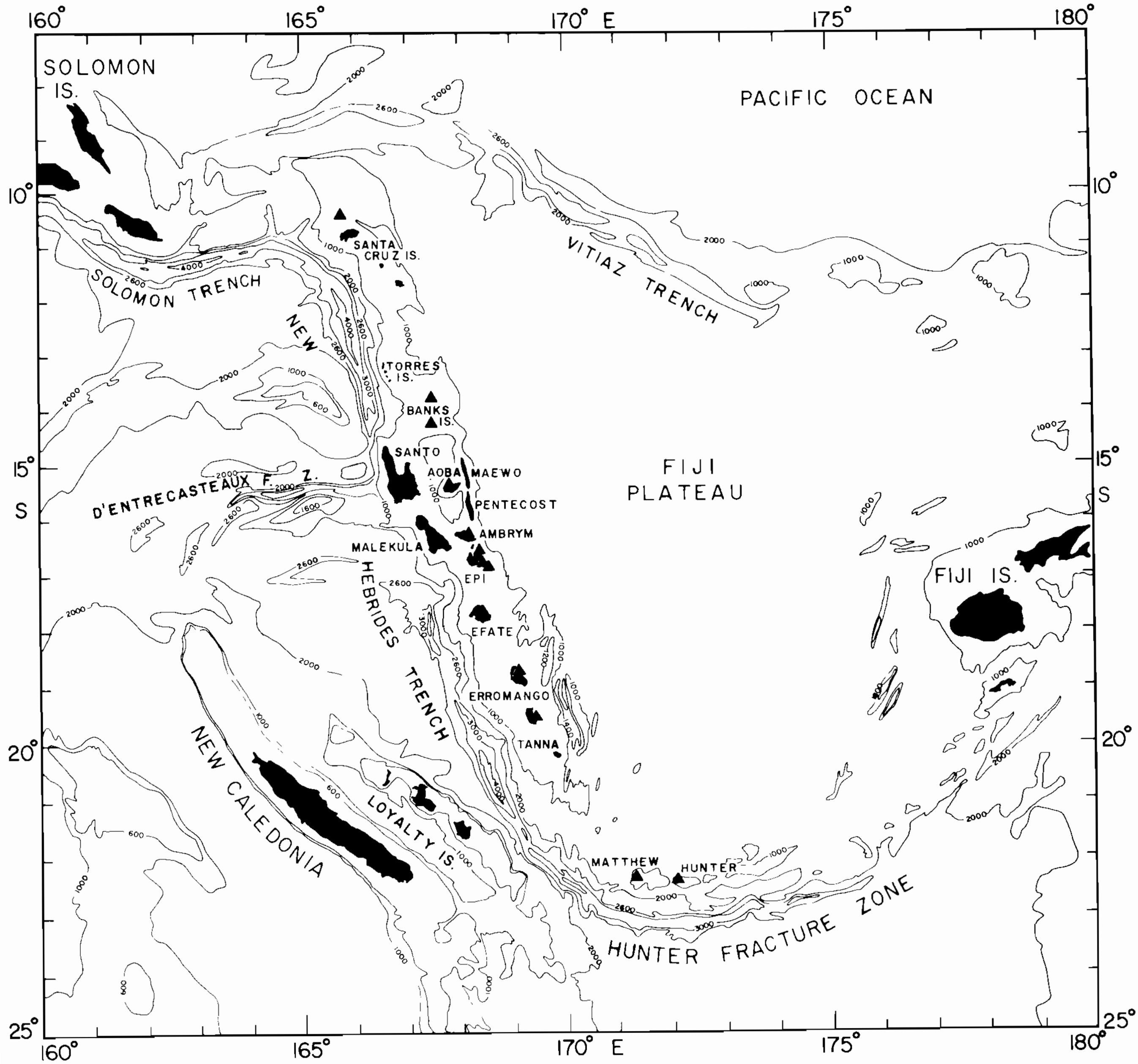


Figure 1. Bathymetric map of the New Hebrides Island arc and surrounding region, taken from Mammerickx et al., 1971. The filled triangles denote Quaternary volcanoes. Contours are in fathoms.

indicative of deformation near a transition between a creeping Efate segment and an adjacent locked segment.

Introduction

The New Hebrides island arc has a very distinctive style of seismicity. Although no earthquake has had a magnitude greater than about 8, numerous large events have occurred in remarkably clustered sequences. This style of seismicity contrasts with that of the Chilean or the Aleutian-Alaskan convergent margins where great earthquakes more or less regularly rupture well-defined segments of the plate boundary. Utsu (1974) considered similar contrasts in the seismicity in the region of Japan, and pointed out the difficulty of resolving seismic "gaps" in a region characterized by a style of seismicity like that in the New Hebrides. Major unresolved questions concern the role of aseismic creep in accommodating some fraction of interplate slippage; the nature of the loading process and its behavior over time scales less than the million year averages provided by analysis of sea-floor spreading data; and the role of structural and geometrical complexities, acting as "barriers" or "asperities" along the plate boundary, in governing the location, size and other characteristics of episodes of seismic slippage.

The inference that variations in the characteristics of seismicity among different convergent plate boundaries represent long-term properties rather than transient features depends largely on the association of seismicity characteristics with structural features of the plate boundary (e.g., Kelleher et al., 1974; Kelleher and McCann, 1976; Uyeda and Kanamori, 1979). The New Hebrides island arc offers a good opportunity to study this problem. In the class of the earth's convergent zones the New Hebrides arc is an extreme member in several respects and may thus exaggerate effects that are more difficult to see in other areas. While having an overall uniformity of configuration on a regional scale, the arc displays remarkable variations in detailed structure along strike, variations that seem to correlate with different characteristics of the seismicity.

The upper plate has a very complex structure that has resulted from the peculiar tectonic history of the region. After a Late Miocene disruption of an ancestral Solomons-New Hebrides-Fiji-Tonga subduction zone, the New Hebrides arc reversed subduction polarity, moved to its present position and left the newly created North Fiji Basin (Fiji Plateau) in its wake (see Figure 1 and Chase, 1971; Karig and Mammerickx, 1972; Gill and Gorton, 1973; Falvey, 1978; and Carney and Macfarlane, 1978). Thus, what is now the leading edge of the plate has played this role for a short time, probably less than 6-8 MY. The most outstanding feature of the New Hebrides arc is the anomalous morphology of the central region. There, the normal elements of physiography including the trench, island arc, and back arc rifts found in the northern and southern parts of the New Hebrides arc are replaced, respectively, by the island blocks of Santo and Malekula, the Aoba Basin, and the uplifted horst-like ridge upon which the islands of Maewo and Pentecost emerge (see Figures 1 and 2). The D'Entrecasteaux "Fracture Zone" (Mammerickx et al., 1971 and Daniel et al., 1977; referred to hereafter as the DFZ) intersects the arc in this region but appears to be subducted along with the rest of the oceanic plate (Isacks and Barazangi, 1977; Pascal et al., 1978; Chung and Kanamori, 1978a).

In our view the striking morphological anomalies of the central New Hebrides are best explained by (1) a Late Miocene episode of intra-arc rifting which produced a major seaward protrusion of the upper plate as the Santo and Malekula blocks moved outboard of the main arc and in so doing created the Aoba Basin (Karig and Mammerickx, 1972), and (2) a Quaternary interaction with the DFZ which affected the uplift and tilting of Santo and Malekula as well as the seismicity (Taylor et al., 1978) but which does not alone account for the main morphological anomalies as proposed by Ravenne et al. (1977) and Chung and Kanamori (1978b). Although unusual, intra-arc rifting of a type analogous to the central New Hebrides may also account for the anomalous morphology of the Bonin arc (Karig and Moore, 1975). The Quaternary subduction of a ridge beneath a pre-existing protrusion of the the upper plate is supported by the fit of the Santo and Malekula blocks into the Aoba Basin; the mismatch between the northern limits of the DFZ and the Santo block; the thickness of sediment in the basin (Ravenne et al., 1977; Luyendyk et al., 1974); and the geological history of the central New Hebrides (Mitchell and Warden, 1971; Mallick, 1973; Mallick and Greenbaum, 1977; and Carney and Macfarlane, 1978).

The complex structure of the upper plate does not appear to be associated with major disruptions or contortions of the descending plate. Pascal et al. (1978) show that on a regional scale the overall configuration of the subduction zone is relatively uniform as found for most other areas (Isacks and Barazangi, 1977). New results of this paper further support this result. The cross-sectional shape of the Benioff zone is clearly among the most sharply downbent and steep ones on earth, and the width of the arc-trench gap is among the smallest. These regional scale geometrical features imply that the width of the interplate boundary -- i.e., the width of the gently dipping thrust fault along which the major interplate earthquakes are generated -- is small compared to most other arcs, and again represents an extreme case in the spectrum of convergent zone variations.

Since 1975 Cornell University, the Office de la Recherche Scientifique et Technique Outre-Mer (ORSTOM), and the New Hebrides Mines Service have collaborated in studies of seismicity, tilting and uplift of the central New Hebrides. The program has included the operation of temporary seismograph networks, two of which included ocean bottom instruments deployed in cooperation with the University of Texas, Marine Sciences Institute (York, 1977; Stephens, 1978; Louat et al., 1979; Ibrahim et al., 1980; Coudert et al., 1981); a permanent network of 19 telemetered stations installed in 1978; measurements of tilt using bubble-level tiltmeters and relevelings of benchmark arrays (Isacks et al., 1978; Marthelot et al., 1980; Bevis and Isacks, 1981); and studies of the pattern of Late Quaternary uplift and tilting of coral terraces (Taylor et al., 1980; Jouannic et al., 1980). This study

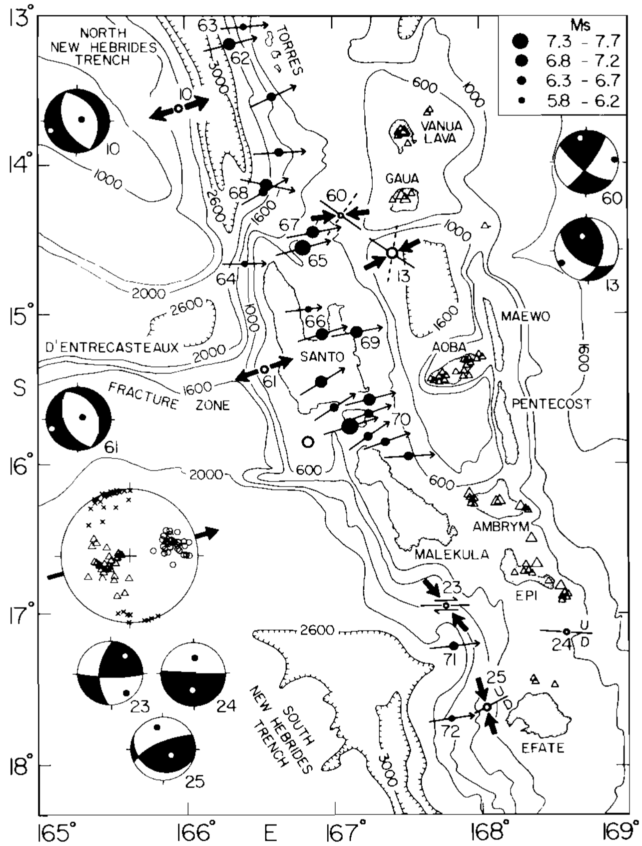


Figure 2. Focal mechanism solutions in the central New Hebrides for earthquakes during 1963 through 1976 with magnitudes (Ms>5 3/4) large enough for reliable focal mechanism determination. Intra-plate epicenters are shown as open circles and the mechanisms are shown in lower hemisphere equal area projections with quadrants of compressional first motions blackened, compressional axes as filled circles, and tensional axes as open circles. Large arrows show horizontal projections of stress axes and lines through epicenters show strike of selected nodal planes, with sense of motion indicated by strike-slip couple or by up (U) and down (D) for dip-slip motion. The arrows through the events interpreted as having interplate thrust-type mechanisms (epicenters shown by filled circles) give the horizontal projection of the slip vector. All interplate mechanisms for the arc are summarized on the larger lower hemisphere equal area plot in the lower left hand side of the figure. Open circles are poles interpreted as the slip vectors, triangles are the fault plane poles, and the X's the "B" or null axes. Numbers are from Pascal et al. (1978) or Table 1; interplate events from Pascal et al. (1978) are not numbered here. Bathymetric contours in fathoms are from Mammerickx et al., 1971. Open triangles on map show volcanoes: large triangles show recently active centers, small triangles show Quaternary centers (Geological Map of the New Hebrides Condominium, 1975).

synthesizes results from these and other published studies plus the following : (1) new focal mechanism solutions for the central New Hebrides updating the collection of Pascal et al. (1978) through 1980 for much of the central region and including new data for events that occurred between 1960 and 1962; (2) focal depths and seismic moments determined by the method of matching observed and synthetic long-period P waveforms (Chinn and Isacks, 1979); (3) preliminary results of a study of seismicity based on data in the Preliminary Determination of Epicenters (PDE) for the past 20 years (Marthelot and Isacks, 1980); and (4) results from the new seismograph network relevant to the seismicity of the region of Malekula and Efate islands. In this study, we focus on the area of Santo through Efate islands where most of the observations have been made.

The northern part of the study area experienced two major sequences of earthquakes in 1965 and 1973-1974 which accommodated seismic slippage along a segment of the convergent plate boundary about 250 km in length. In contrast, much of the plate boundary in the southern part of the study area may not have experienced major seismic slippage during the past 75 years. Several interesting features of the southern area stand out. These include (1) persistent nests and quiescent zones of seismic activity; (2) an unusually diffuse distribution of small shallow earthquakes near the inferred location of the inclined plate boundary; and (3) high rates of tilting measured by releveling of benchmark arrays during a 5 year period since 1975. One interpretation is that these features are related to a cycle of strain accumulation leading to a major episode of seismic slippage of the plate boundary similar to that which occurred recently in the northern part of the study area. Alternatively, the mode of slippage of the plate boundary may vary significantly along the arc and may be associated with specific structural features of the complex plate boundary.

Without further information on the past seismicity the choice between these alternatives remains open, but an opportunity to examine in detail an important episode of activity in the area of anomalous seismicity near Efate Island was provided by the capture of a sequence of three moderate-sized (Ms = 6) earthquakes in 1978 and 1979. In this paper we present preliminary determinations of the source parameters of the mainshocks and the space-time pattern of occurrence of the associated small shocks located by the seismograph network for a 1.5 year period. In addition, we report tilt measurements made before and after the sequence.

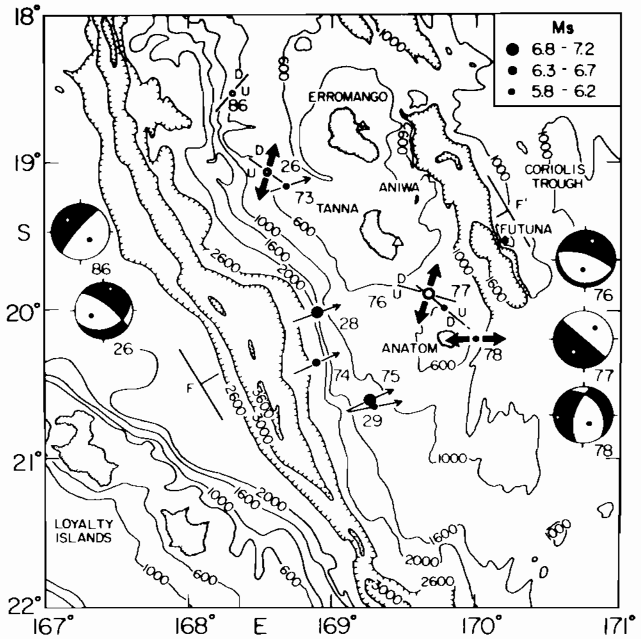


Figure 3. Same as Figure 2 but for the area south of the figure (after Coudert et al., 1981). Numbers are from Coudert et al. (1981), Pascal et al. (1978), and Table 1. Bathymetry interpolated from new results synthesized by the French Office de la Recherche Scientifique et Technique Outre-Mer (ORSTOM; Daniel, 1978) for 18°S to 20°S and Mammerickx et al. (1971) from 20°S to 22°S. X′s show locations of seismograph stations on land and on the ocean bottom operated during August and September, 1977. Section F-F′ is shown in Figure 5.

Tectonic Features of the Central New Hebrides

Complex Upper Plate

The upper plate has a major westward protrusion formed by the block-like morphological features upon which are located the islands of Santo and Malekula. The northern limit of the Santo block is clearly defined by the bathymetry. Focal mechanisms of two events (events 13 and 60, Figure 2) possibly located on the boundary between the Santo and Banks Islands (Vanua Lava and Gaua, Figure 2) blocks have large strike-slip components, but the sense of displacement is opposite that expected for the westward translation of the Santo block. The simplest interpretation is that these events represent a reactivation of the block boundary and reflect compressive stress now probably resulting from the interaction with the DFZ.

To the south the boundary of the upper plate protrusion is more complicated. The bathymetry suggests a distribution between a Santo-northern Malekula block and a southern Malekula block. The region between southern Malekula and Efate Island has a complex bathymetry and widespread distribution of volcanic centers, both features indicative of major transverse structures (Figures 2 and 4). The northern termination of the South New Hebrides Trench is adjacent to a pronounced embayment in the shape of the leading edge of the upper plate and is also located just west of the northern termination of the system of back-arc rifts (the northern continuation of the Coriolis Trough). To the south, Efate Island is part of a westward salient of the upper plate located between the major embayment to the north and the less pronounced one which separates the Efate salient from the Tanna-Erromango salient.

Focal mechanism solutions for events in the upper plate for earthquakes located south of the Santo-Malekula area are interpretable as block faulting on nearly vertical nodal planes (events 24,86,26,76,77, Figures 2 and 3). Only one mechanism indicates a significant component of horizontal extensional stress for an event (event 78) located near the western side of the Coriolis Trough. The strong transverse trends between Malekula and Efate are reflected by the east-west striking nodal planes of the solutions for events 23 and 24 (Figure 2).

Based on the pattern of horizontal compressive stress trajectories inferred from the distribution of volcanic centers (following Nakamura, 1977) and other volcanic and structural features on the islands, Roca (1978) proposes

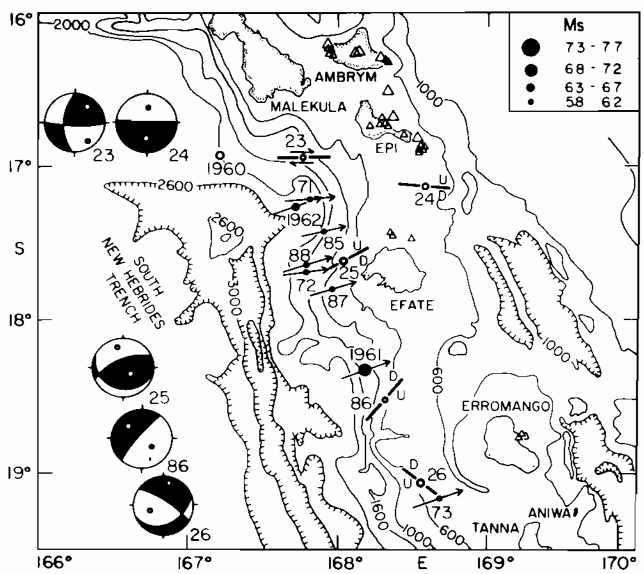


Figure 4. Earthquakes with focal mechanism solutions for the period 1960 through 1979 with symbols the same as in Figure 2 and bathymetry references of Figure 3, with the area of 16°S to 18°S taken from Mammerickx et al. (1971). Only the largest events for the period 1960-1962 are included.

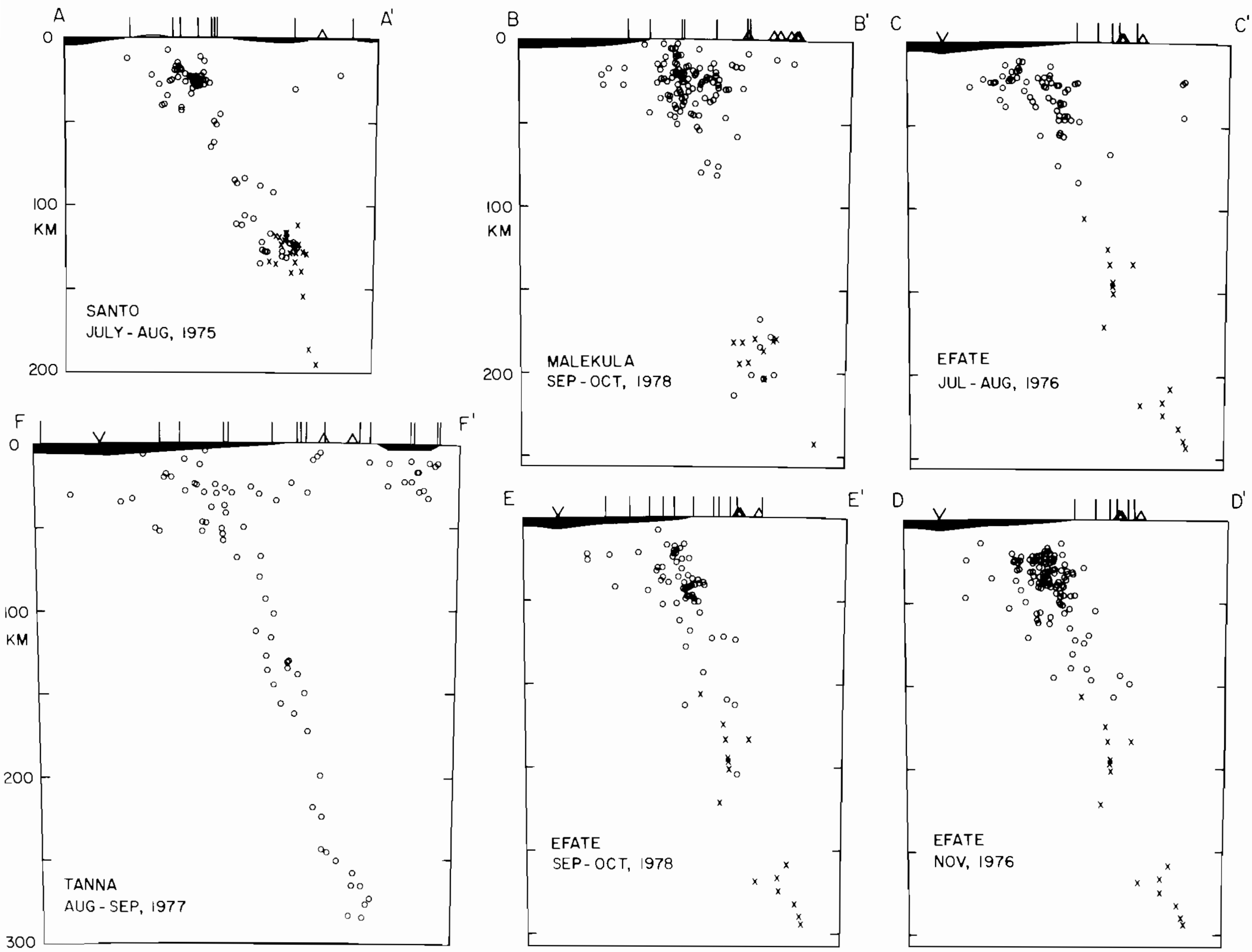


Figure 5. Cross sections of locations by temporary local networks operated for intervals of 1-2 months each. The horizontal and vertical scales are equal. Section locations are shown in Figure 3,6 and 7. The locations of sections C-C′ and D-D′ are the same as E-E′, but only epicenters for E-E′ are shown in Figure 6. The stations are shown by vertical lines on the surface, active or Quaternary volcanoes are shown by triangles, and the axis of the trench by a V. Results (open circles) are relocations based on data from Stephens (1978), York (1977) and Coudert et al. (1981) using the same flat layered model and the Hypo71 program of Lee and Lahr (1975). The model parameters include layer thicknesses (from the surface) of 15,10,175 and 100 km with P wave velocities, respectively, of 5.55, 6.5, 8.1 and 8.2 km/sec, and a Vp/Vs ratio of 1.75. The X′s are reliable ISC locations for the Efate and Malekula sections and are relocations from Pascal et al. (1978) for the Santo section (all for depths greater than 70 km).

that subduction of the DFZ acts as a concentrated source of stress within the upper plate. The main evidence is a tendency for the inferred directions of maximum compressive stress to radiate out away (in a map view) from the DFZ. This effect is most strikingly illustrated by the variations in the alignments of volcanic centers from Aoba to Ambrym, as can be seen in Figure 2. Note also that the direction of the axis of compression in the focal mechanism solution of event 23 located south of Malekula agrees with the pattern. The focal mechanism solutions for upper plate events located farther south along the arc show a rather complex and variable pattern without a clear trend in the orientation or type of the horizontal stress component.

A further argument supporting this change in upper plate stress along the strike of the arc can be made by considering the narrow ridge along which the islands of Maewo and Pentecost emerge (see Figure 2). These narrow, linear islands show clear evidence of rapid Quaternary uplift and appear to be located on horst blocks (Mallick and Neef, 1974; Luyendyck et al., 1974; Carney and Macfarlane, 1978). They occupy a location in the morphology of the arc normally occupied by the back-arc rifts found along both the northern and southern parts of the New Hebrides arc (Dubois et al., 1978). It is conceivable that a previously existing rift structure similar to that found in the northern and southern New Hebrides has been reactivated, but is now under compression rather than extension, and what were originally grabens have now become uplifted horst blocks. In a similar line of interpretation, the reactivation of faulting in a sense opposite to that involved in the original structure was suggested above as the interpretation of focal mechanisms of events in the northern part of the Aoba Basin. Mallick and Greenbaum (1977) also report reactivation of a Miocene normal fault (located in western Santo) as a reverse fault.

It is thus possible that the central New Hebrides, with its protruding upper plate interacting with major bathymetric features of the subducted plate, is the most tightly coupled part of the plate boundary and the region of major transverse features located near the northern end of the South New Hebrides Trench may be a transition zone between two different stress regimes along the arc.

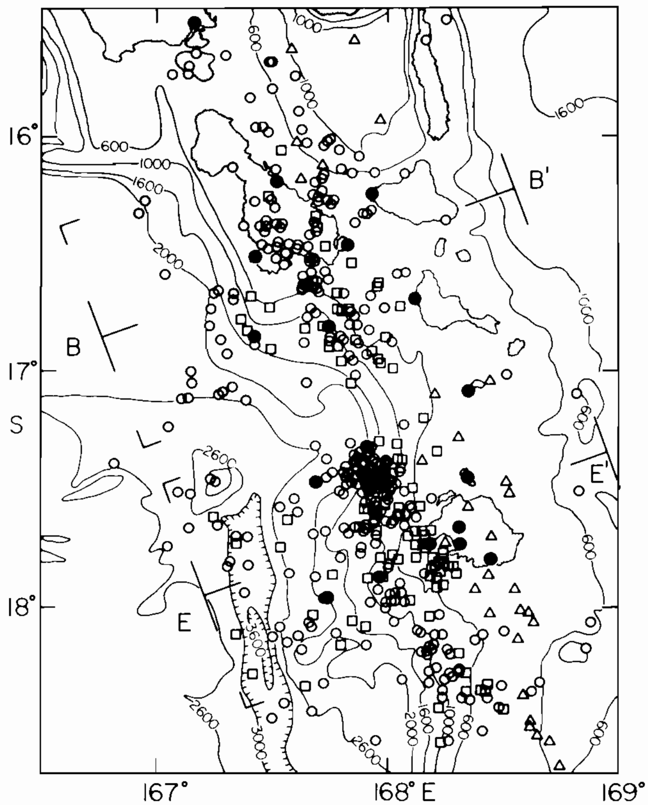


Figure 6. Epicenters of events located during September and October, 1978 by the network shown by filled circles. The ocean bottom stations operated during two weeks in September. Sections B-B′ and E-E′ are shown in Figure 5. Open circles are for depths less than 35 km, squares for depths between 36 and 69 km, and triangles for depths greater than 69 km. Bathymetry as in Figure 4.

Overall Geometry of the Subducted Plate and the Interplate Boundary

The relatively uniform configuration of the inclined zone of intermediate depth earthquakes demonstrated by Dubois (1971), Isacks and Molnar (1971) and Pascal et al. (1978) is simply related to the regional scale uniformity of the arc in map view as shown by the alignments of the North and South New Hebrides trenches and by the overall alignments of active and late Quaternary volcanic centers.

New data from local seismograph networks operated in the region support this result (Figure 5). In all sections the distribution of events with depths greater than about 50 km suggests a downbent form similar to that so well exhibited in the Tanna section. Although the activity is not uniformly distributed, comparisons of the four sections taking into account the 20-35 km thickness of the intermediate depth zone do not indicate any significant change in overall dip among the four areas. This result and the locations of events at depths between 50 and 125 km in the Santo, Malekula, and Efate sections provide fairly strong evidence against the "flap and gap" structure discussed by Choudhury et al. (1975) and Pascal et al. (1978).

It is remarkable that for three of the four areas, Efate, Malekula and Tanna, the shallow hypocenters located by the temporary networks fail to define a single thin zone that can be identified as the interplate boundary. We infer the existence of such a boundary from the abundant thrust-type focal mechanism solutions of earthquakes with epicenters in the arc-trench gap. Published and new solutions updating the results of Pascal et al. (1978) are summarized in Figures 2-4 and Table 1.

In the collections of data presented by Johnson and Molnar (1972) and Pascal et al. (1978) very few thrust-type solutions are found for events located in the southern half of the New Hebrides arc. A major result of the new data is to add thrust-type solutions for the southern half of the arc and thus largely remove the discrepancy between the two halves of the arc. As shown in Figure 8 the directions of slip plotted as a function of latitude show no significant change across the anomalous central region. The average direction of slip determined from these data, N76° ± 11°E, is essentially the same direction as that determined by Pascal et al. (1978).

The exact position and shape of the interplate boundary is difficult to determine. Relevant data include the dips of the slip vectors for the thrust-type focal mechanism solutions, the depths

TABLE 1 Focal Mechanism Solutions (see Fig. 20)

No.[1]	Date Mo/Dy/Yr	Origin Time Hr:Min	Location[2] Lat. S	Long. E	Depth, km	Pole[3] 1 Tr.	Pl.	Pole 2 Tr.	Pl.	P Axis Tr.	Pl.
60	10/09/73	07:58	14.34	167.06	27	038	20	305	10	083	07
61	12/30/73	16:39	15.37	166.54	10	070	34	250	56	070	79
62	01/23/72	21:18	13.18	166.32	33	(4)					
63	01/24/72	03:56	13.07	166.40	38	(4)					
64	01/06/73	15:53	14.66	166.41	24	090	10	280	80	271	35
65	12/28/73	13:42	14.56	166.80	13	075	46	255	44	075	01
66	12/29/73	00:19	15.13	166.92	43	073	40	253	50	253	05
67	01/10/74	08:51	14.45	166.87	36	079	45	258	45	258	00
68	01/11/74	05:37	14.19	166.54	37	060	30	240	60	240	15
69	11/20/74	04:14	15.10	167.16	62	080	60	244	30	077	15
70	04/08/73	12:41	15.81	167.24	38	055	33	218	57	227	12
71	06/05/73	03:12	17.22	167.81	5	084	26	264	64	264	19
72	05/26/74	01:32	17.69	167.80	13	080	16	260	74	260	29
85[6]	08/01/78	04:16	17.38	167.88	20	077	18	257	72	257	27
86	01/27/79	18:15	18.54	168.21	25	311	84	131	06	141	51
87	08/17/79	12:59	17.73	167.87	25	090	15	270	75	270	30
88	08/26/79	11:47	17.63	167.71	22	090	15	270	75	270	30
	03/29/60	06:30	16.93	167.22	0	(5)					
	07/23/61	21:51	18.33	168.18	0	070	20	250	70	250	25
	10/06/62	04:23	17.26	167.72	0	070	15	250	75	250	30

1 Numbers continue numbering system of Pascal et al. (1978).

2 Locations are from Bulletins of the International Seismological Centre (ISC) or from the International Seismological Summary (ISS) except for 86, which is taken from the Monthly Bulletins of the PDE, and 85,87 and 88, which are relocated with all available local and teleseismic data at the depth fixed by matching observed and synthetic long period *P* waveforms.

3 Trends measured clockwise from north; plunges measured from horizontal.

4 Solutions published by Kim and Nuttli (1975).

5 Solution not well determined but appears inconsistent with typical interplate thrust-type solutions.

6 Focal mechanism solutions for events 85-88, and the 1960-1962 events are preliminary and are not shown in Figure 20. The solutions are based on our readings of first motions recorded by the long-period seismographs of the Lamont International Geophysical Year (IGY) network plus first motions reported in the *International Seismological Summary*.

of some events, and, in the case of data from the operation of a temporary array beneath Santo Island, the location of a cluster of hypocenters whose common focal mechanism solution is of the thrust-type similar to that for larger events (see Figure 7). Better control on depth of moderate-sized events is provided by Chinn and Isacks' (1979) determination of accurate focal depths by matching observed and synthetic long period *P*, *pP*, and *sP* waveforms. The results for three areas beneath Santo, Efate, and Tanna islands are shown in Figure 9 together with data from local networks. A curve with the same shape was fitted to each of the sections as a crude approximation to the location of the plate boundary and as a reference for inter-comparison of the sections. At shallow depths the curve is an estimate of the interplate boundary, but at intermediate depths it approximates the upper envelope of the inclined seismic zone and thus

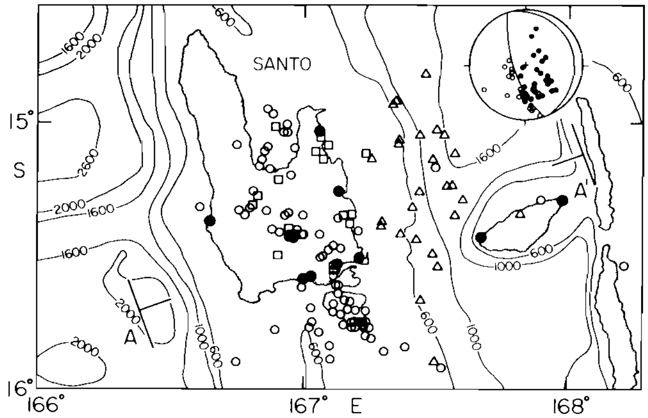


Figure 7. Epicenters of events located beneath Santo Island during July and part of August, 1975 (symbols are as in Figure 6). Not all of the stations operated at the same time. The lower hemisphere equal area plot in the upper right corner shows first motions (solid circles-compressions, open circles-dilatations) for events located in the concentrated cluster of events at depths near 20-25 km shown in section A-A′ of Figure 5. Bathymetry as in Figure 2.

may pass slightly beneath the actual upper surface of the subducting plate.

Figure 9 shows that in comparison to most other subduction zones the sharply downbent configuration of the New Hebrides subducted plate results in a very narrow interface zone with relatively steeply dipping slip vectors. The maximum depth of interplate seismic slippage obtained from the accurate focal depths is 36 km. Unfortunately, the depths of the events located farthest west and with the steepest slip vectors were not determined due to complexity of the source process, and the depth of seismic slippage along the plate boundary may in one case reach 60 km (e.g., event 69, Table 1 and Figure 2). Nevertheless, if we take a depth of 40 km as that of the down-dip edge of the most seismically active part of the plate boundary, then the Efate and Tanna sections have widths of about 90-95 km as measured to the trench axis. The seaward protrusion of the Santo-northern Malekula block in the upper plate appears to add 10-20 km to this width, as would be expected if the intermediate depth seismic zone has a uniform configuration through the central New Hebrides.

The thickness of the seismically active parts of the subducted plate beneath the shallow plate boundary as determined by well-located events is about 25-35 km, in agreement with the thickness of the intermediate depth seismic zone found by Pascal et al. (1978). No clear evidence for distinct double Benioff zones similar to those described by Hasegawa et al. (1978) have been observed.

Recent History of Seismic Plate Boundary Slippage

The 1965 and 1973-1974 Sequences of Large Earthquakes: Plate Boundary Slippage Beneath the Santo-N. Malekula Segments

Two major sequences of large earthquakes occurred in the region of Santo Island during August, 1965 and December, 1973-January, 1974 (see Figure 10). The 1965 sequence has been studied in some detail by Taylor et al. (1980) and Ebel (1980). The boundaries of the aftershock zones shown in Figure 10 are based on locations of events occurring within two days of the mainshocks. Well-located shocks are selected according to data reported in the *Bulletins of the International Seismological Centre*.

Segment Boundaries. The estimate of the southern limit of the aftershock zone of the 1965 sequence is somewhat ambiguous because possible aftershocks occurred in the region of the southern Malekula and Epi islands during the month following the main sequence. However, the immediate aftershock activity is concentrated in the region shown. Further evidence of the southern extent of rupture are provided by the determinations by Taylor et al. (1978) of the uplift associated with the 1965 sequence. The decrease in uplift southwards along the west coast of Malekula indicates that the rupture of the 1965 sequence probably did not extend beneath southern Malekula. Furthermore, the rupture boundary may coincide with the remarkable discontinuity in the tilting of the late Quaternary coral terraces described by Taylor et al. (1978). This feature is located along the

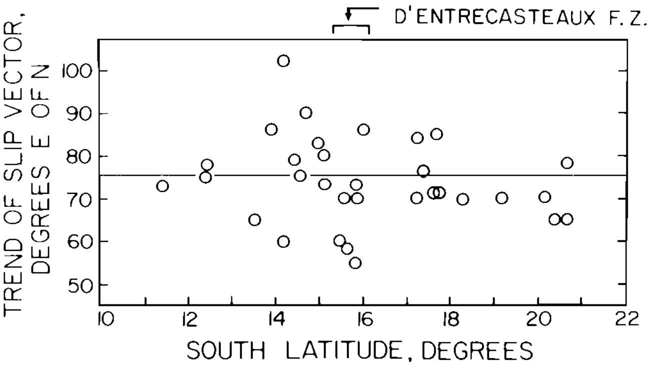


Figure 8. Trends of the nodal plane poles taken as the direction of the slip vector of all interplate thrust-type focal mechanism solutions for the New Hebrides, including data from Pascal et al. (1978), Coudert et al. (1981), and Table 1. The line gives the average azimuth of N76°E. (The trends are plotted against latitudes of the epicenters of the event.)

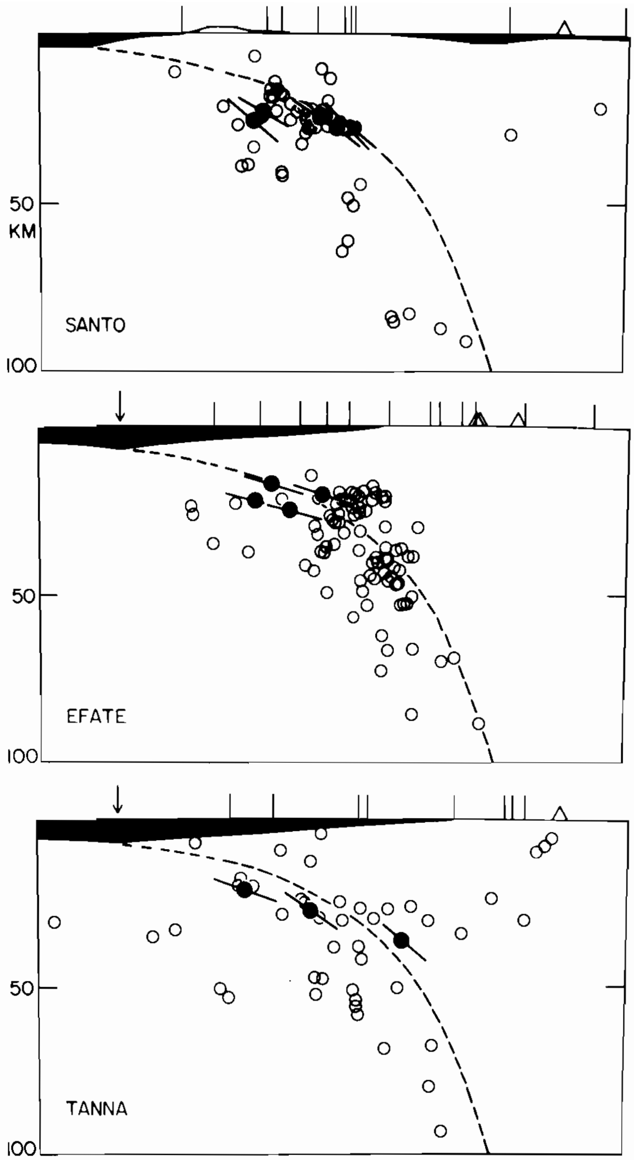


Figure 9. Detail of three of the sections, A-A′, E-E′, and F-F′ shown in Figures 5 through 7 plus earthquakes (large filled circles) with focal mechanisms and depths determined by analysis of long-period P records. The lines through the large filled circles are projections of those nodal plane poles chosen as the directions of the slip vectors. The continuous line has the same shape for the three sections and is adjusted in position to give the best fit to the locations of the interplate events and the dips of the slip vectors; at intermediate depths the line is an approximation to the upper envelope of the inclined seismic zone. The dashed part of this line is of variable length and is drawn to the assumed surface trace of the plate boundary at the trench (marked by an arrow).

eastward projection of the southern scarp and ridge of the DFZ (see Figure 2). We hypothesize that the main plate boundary rupture associated with the 1965 sequence terminates near this boundary.

The northern limits of the rupture are not very well defined. The very young uplift along the south coast of Santo (Taylor et al., 1980) could be associated with the 1965 earthquake, although the association is not clearly established. As pointed out by Pascal et al. (1978) and Chung and Kanamori (1978b) the boundary between the aftershock zones of the 1965 and 1973-1974 sequences closely corresponds to the eastward projection of the northern ridge and scarp of the DFZ. Careful examination of the well located aftershocks suggests a gap between the 1965 and 1973-1974 aftershock sequences as shown in Figure 10. It is possible that the earthquakes of October, 1971 and November, 1974 (see Figure 10) contributed to filling this gap, but the areas of rupture for these events are not known. Well-located aftershocks of the 1971 events are too few to define a clear aftershock zone and no aftershocks are reported by the ISC for the two days following the 1974 shock.

The boundary between the December 1973 and the January 1974 sequences is well defined by the aftershock zones and corresponds to the northern end of the main Santo block. The January 1974 sequence occurs beneath a triangular area of the upper plate which forms the southern part of the embayment between the Santo block and the Torres Island salient. The January sequence is further delineated by a zone of strike-slip faulting inferred from the locations and focal mechanisms of the earthquakes of October, 1973 (event 60) and May, 1965 (event 13).

In summary, the boundaries of the 1965 and 1973-1974 sequences appear to be strongly controlled by clear features in both the subducted and upper plates, including the northern boundary of the Santo block, the northern and southern scarps of the DFZ, and a boundary between the Santo-N. Malekula block and the S. Malekula block.

Development in time and space. The two-day long sequence in 1965 shows a clear southwards migration of the successive major events as shown in Figure 10. Significant slip apparently continued in the 8 years following this sequence as indicated by the events in 1966, 1971, 1972 and 1973. Similarly, the November, 1974 earthquake (no. 69) occurred nearly a year after the December, 1973-January, 1974 sequence. These "late aftershocks" occur on the eastern edges of the inferred rupture zones and could thus represent a down-dip expansion of the ruptured area. These observations could be accommodated by the model of Thatcher and Rundle (1979) in which post-seismic slippage of the interplate boundary occurs near a transition between a shallow region of stick-slip behavior and a

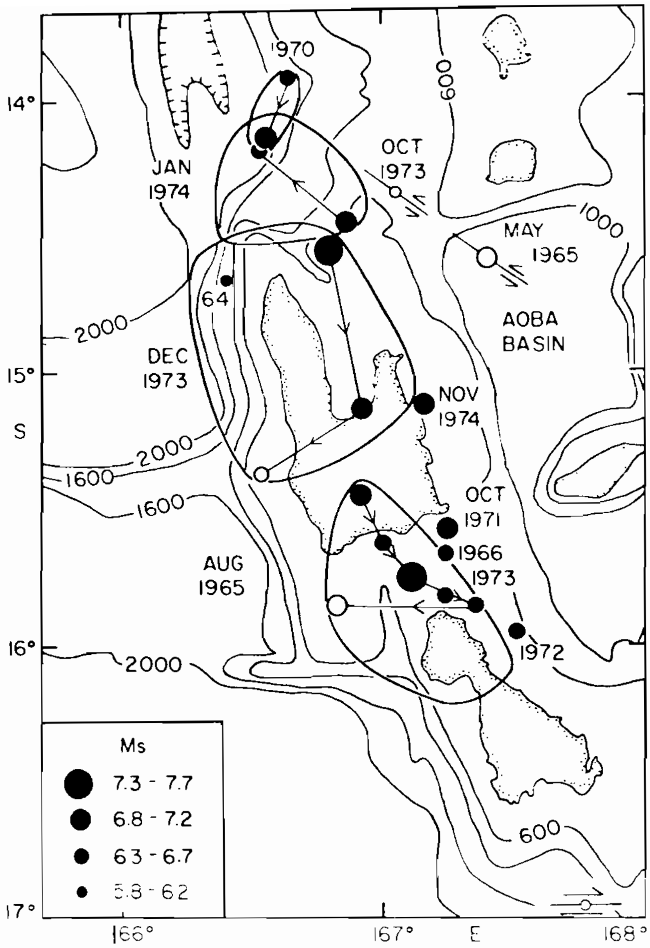


Figure 10. The temporal development of the three large sequences in 1965, 1973 and 1974 and the smaller sequence in 1970. The remaining activity is seen to be closely associated with these sequences. The irregular outlines are estimates of the boundaries of the aftershock zones associated with the four sequences, and the arrows show the sequence of events in time. The X shows the location of the leveling array on Santo Island. Earthquakes with thrusting focal mechanism solutions are shown as filled circles and other focal mechanism solutions are shown as open circles.

deeper region of creep. Further evidence for a transient effect lasting several years after the 1965 sequence is the exponential like decay in the rate of occurrence of earthquakes during the 3 year period following the sequence (see Figure 14). The 1973-1974 sequence occurs at the end of this transient, and Marthelot and Isacks (1980) describe a clear increase in the rate of occurrence of intermediate depth events during the transient.

The second sequence, a more complex one, occurs in two periods of about two days each separated by a hiatus of 10 days. In the first period, December 28-30, 1973, the rupture appears to propagate southwards as shown in Figure 10. One of the larger shocks (event 61, Figure 2) of the sequence had a normal faulting mechanism and is located within the sub-oceanic plate. A significant number of the aftershocks of the first, largest event are also located beneath and seaward of the western edge of the Santo block and probably represent deformations within the sub-oceanic plate. The second part of the sequence was initiated on January 10, 1974 by an earthquake located near the epicenter of the first mainshock in December, but the rupture then appeared to proceed northwards.

The 1965 and the 1973-1974 sequences were in each case preceded by one of the upper plate shocks along the northern boundary of the Santo-northern Malekula block. Event 13 occurred nearly 83 days prior to the 1965 sequence, while event 60 occurred 70 days prior to the beginning of the 1973-1974 sequence.

Thus, nearly all the events located near Santo in Figure 2 can be associated with the 1965 and 1973-1974 sequences. The close coordination in time and space among the inter- and intraplate events can be taken as further evidence of the degree of coupling between the convergent plates in the Santo-northern Malekula segments.

Seismic Moments. Ebel's (1980) moments for the 1965 sequence determined from amplitudes of long-period surface waves total 4.3 x 10^{27} dyne-cm for the four interplate events (excluding the probable intraplate event of August 13). A crude estimate of the fault width and slip can be obtained from the moment if the fault length is known and the stress or strain drop assumed. We take a fault length of 100 km. If the stress drop is 30 bars (Kanamori and Anderson, 1975; Kanamori, 1977; Ebel, 1980), then, with respect to simple rectangular dislocation models, the fault width would be 30-40 km and the fault slip 3-4 meters. In support of these crude estimates Taylor et al. (1980) models the co-seismic uplift with a width of 40 km and a fault slip of 5.3 meters. This is less than half the width of the boundary as measured to the estimated surface outcrop (see Figure 9). The smaller width is further supported by the concentration of interplate seismicity (Figures 2-7) along a band located at depths greater than about 10-15 km and distances of about 30-40 km down-dip from the surface trace of the plate boundary. With an estimated slip of about 3-5 meters and a rate of convergence of about 10 cm/yr (Dubois et al., 1977) the "average" repeat time for the 1965 sequence would be 30-50 years.

Moments have not been determined for the 1973-1974 sequences. Estimates based on the surface wave magnitudes (log Mo= 1.5 Ms + 16.1, taken from Kanamori and Anderson, 1975; Pucaru and Berckhemer, 1978; and Hanks and Kanamori,

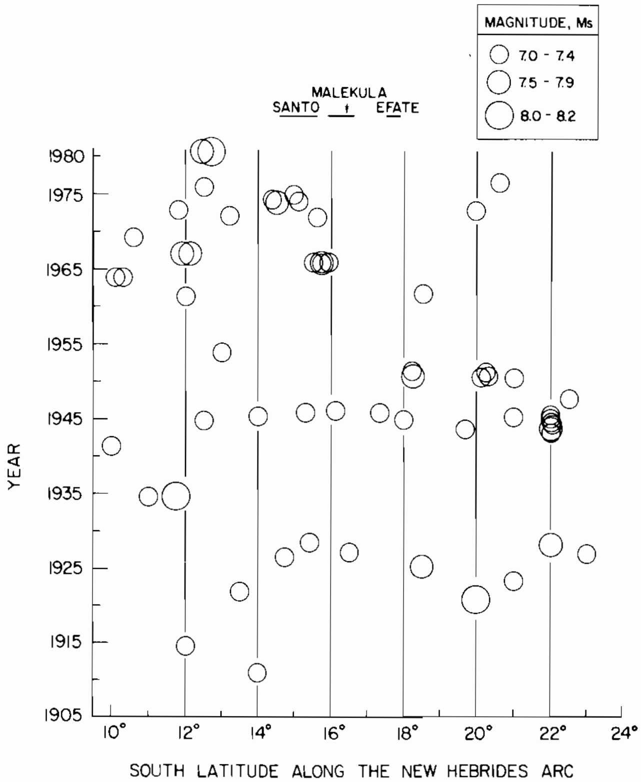


Figure 11. Large earthquakes (Ms > 6.9) plotted as a function of time and of the latitude of the epicenter. Data from Gutenberg and Richter (1954, with relocations of McCann, 1980), Rothe (1965), the "Seismological Notes" of the Bulletin of the Seismological Society of America, and the monthly listings of the Preliminary Determination of Epicenters (PDE).

1979) yield 3.7 x 10^{27}dyne-cm for the December, 1973 sequence and 1 x 10^{27} dyne-cm for the January, 1974 sequence, or a total of nearly 5 x 10^{27} dyne-cm. These data in combination with the aftershock zones indicate rupture of the plate boundary of approximately similar amount and extent to that in the 1965 sequence.

A Gap in the South Malekula-Efate Region?

In a time-space plot (Figure 11) of large earthquakes (Ms $\geq$7) occurring during the past 75 years, two nearly arc-wide periods of activation occur in the 1920's and again in the 1940's. Since then, seismicity has been most notable in the northern half of the arc. The southern Malekula-Efate region (16°-18°S) has not had any events with magnitudes as large as 7 since the 1940's. With respect to the estimated 30-50 year repeat time for the 1965 sequence, the Southern Malekula-Efate region may be considered a seismic gap. An important question is to what extent the pre-1960 earthquakes involved seismic rupture of the plate boundary comparable to that accomplished in the 1965-1974 episodes.

In Figure 12 we have used McCann's (1980) relocations of some of the events listed by Gutenberg and Richter (1954). The differences between these locations are in certain cases quite substantial. For example, the largest discrepancy is nearly 200 km for the January 20, 1946 earthquake located by McCann beneath northern Malekula but located by Gutenberg and Richter near the northern end of the South New Hebrides trench (northwest of Efate Island). However, the locations of the 1944 event agree closely. Serious uncertainties in the magnitudes of the older events are also a problem. For

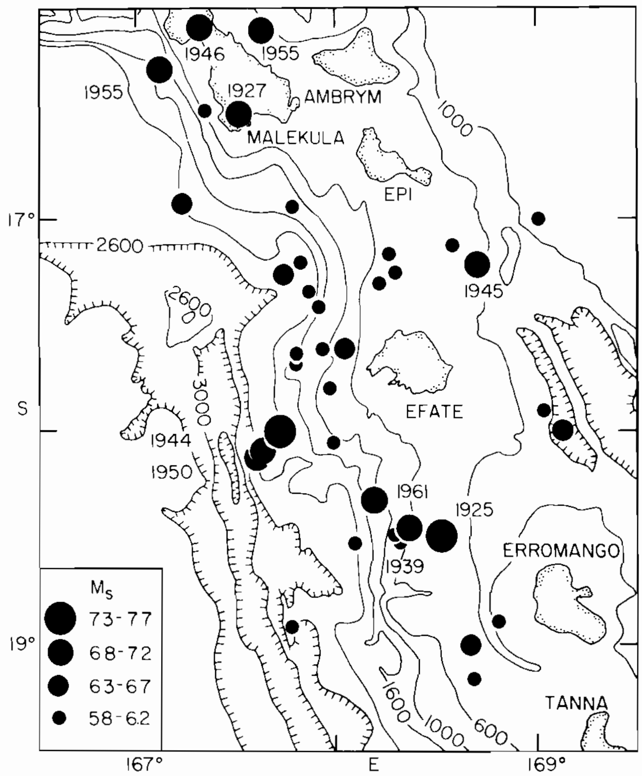


Figure 12. Large and moderate-sized earthquakes in the Malekula-Erromango region. The data for large earthquakes (Ms > 6.9) is the same as that for Figure 11 except that the December 2, 1950 earthquake is plotted with the magnitude revised by Geller and Kanamori (1977). McCann's relocation of the January 20, 1946 earthquake is beneath Santo, and located out of the figure. The data for moderate-sized earthquakes (Ms = 5.8 to 6.7) are shown for the period 1955 through 1979, with locations taken from Rothe (1965), the International Seismological Summary (ISS), the Bulletins of the International Seismological Centre (ISC), and the Preliminary Determination of Epicenters (PDE).

example, one of the largest events in the Efate region reported by Gutenberg and Richter is the December 2, 1950 shock with Ms = 7 3/4 (Richter, 1958 reports a magnitude, M, of 8.1). However, in the revisions of Geller and Kanamori (1977) the magnitude of the 1950 event is decreased to Ms = 7.2 (as plotted in Figure 12). Thus, any conclusions must be highly tentative until further analysis is made of the locations, focal mechanisms and moments of the older events.

The area from Santo through Erromango, a segment of the plate boundary of about 600 km in length, includes two groups of large, probably interplate earthquakes. Five events are located in the Santo-Malekula area and six events are located between Efate and Erromango. The 1945 earthquake is probably an intraplate event. The northernmost earthquake of the Santo-Malekula group, the January 20, 1946 earthquake, is located beneath Santo and out of the map of Figure 12. For comparative purposes we use the same moment-magnitude relationship used above in the analysis of the 1973-1974 episode. The four events beneath Santo and Malekula together contribute a net moment of about 2.6×10^{27}, while the six events located between Efate and Erromango contribute 7.1×10^{27} dyne-cm. The sum of the two groups is 9.8×10^{27} dyne-cm, a total which is close to net moment of 8.9×10^{27} dyne-cm estimated for the 1965-1974 sequence described in the last section. The pre-1965 events occur over a length of the arc nearly three times that ruptured during the 1965-1974 episode. Thus only about a third of the plate boundary from Santo through Erromango may have ruptured seismically during the 60 years before the 1965-1974 episode.

The locations in Figure 12 suggest that perhaps much of the area between Efate and Erromango may have slipped via the 1925, 1939, 1944, 1950 and 1961 events, in addition to southern Malekula via the 1927, 1946, and 1955 events. This would leave a prominent gap in the area of Efate and Epi islands and would imply that the 1965-1974 episode was the only rupture of the Santo-northern Malekula area during the past 75 years. Alternatively, the pre-1960 events beneath Malekula and Santo may have ruptured a part of the Santo-northern Malekula segment (with the 1965-1974 sequence including a repeat of such rupture) and a prominent gap remains beneath southern Malekula. In either case, the area of Efate and Epi islands appears deficient in large earthquakes during the past 75 years.

Unusual Seismicity Near Efate and Southern Malekula

Seismicity During the Past 20 Years

The largest interplate events in the southern Malekula-Erromango region during the past 20 years include the magnitude (Ms) 7.2 earthquake in 1961 and a striking spatial concentration of moderate-sized events in the Efate salient. The 1961 event has a well-determined thrust-type focal mechanism solution (Table 1 and Figure 4). The spatially concentrated activity beneath the Efate salient is developed mainly in three clusters of events which occurred in 1960-1962, 1973-1974, and 1978-1979 (Table 1 lists these events). These clusters are characterized by events with magnitudes (Ms) near 6. The largest one in 1962 had a magnitude of 6.6. In addition, an intraplate event occurred in 1966 (event 25, Figures 2 and 4) with a magnitude of 6.5.

The events reported by the PDE for the period 1961 through 1979 provide a sample of seismicity that covers magnitudes (mb) above about 4.5. The data for the New Hebrides arc are summarized in Figures 13 and 14 (Marthelot and Isacks, 1980). Although seismic quiescence preceding large earthquakes can sometimes be seen in the data of Figure 14 it is not very obvious in other cases. In particular, the activity in segment 23 in relation to the large earthquake (Ms = 8.0) of August, 1980 reveals no obvious precursory activity. The detection of such phenomena is difficult with the limited time sample and poorly defined background levels. What seems more striking in the data is the evidence for substantial variations from segment to segment in the character of the curves, variations which appear persistent at least over the time sampled.

The segments which experienced major earthquakes (e.g., segments 16-20 for the 1965-1974 episode) show very strong fluctuations in the cumulative number of events as a function of time. These fluctuations are dominated by the aftershock sequences and in one case includes a longer, exponential-like decay in the rate of occurrence during a period of about 3 years after the mainshocks.

In striking contrast to this strongly fluctuating activity, and also to all other segments, the Efate salient (mainly segment 13 but including part of 12) exhibits a very high rate of occurrence. This nest of seismic activity, a persistent feature throughout the 20 year period, coincides approximately with the concentration of magnitude 6 events described above. North of the nest the southern Malekula-Epi region (segments 14 and 15) exhibits a slight perturbation associated with the 1965 sequence, but is otherwise characterized by a moderate level of activity. South of the Efate nest the area between Efate and Erromango (including segments 10 and 11) has had a relatively low level of activity after the magnitude 7.2 event in 1961.

Small Earthquakes Located by Temporary Networks: Anomalous Cross Sections Beneath South Malekula and Efate

The data of Figure 5 (see also Figure 9) define the plate boundary only beneath Santo where the most important recent episode of interplate boundary slippage occurred 10 years prior to the

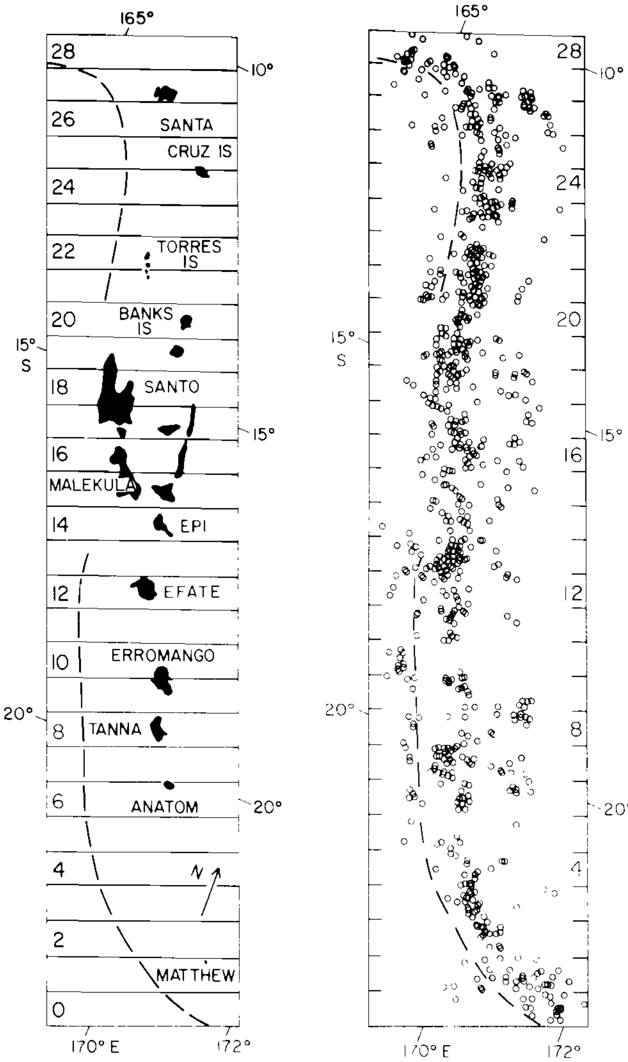


Figure 13. Left: Map of the New Hebrides island arc, divided in 29 contiguous strips each a half degree wide and perpendicular to the arc. The dashed line represents the axis of the trench. The numbers are used to identify each strip in Figure 14.
Right: Spatial distribution of shallow earthquakes (depth < 70 km) from 1961 through 1979, selected from the PDE bulletins according to the reported magnitudes and the number of stations used in the locations and includes events with magnitudes (mb) estimated to be greater than or equal to 5.

operation of the network. In contrast the Efate sections (for three different periods of sampling in Figure 5) show a very active shallow section distributed in a rather broad zone near the inferred plate boundary. The locations indicate abundant lower plate and some probable upper plate activity. The characteristically high rate of occurrence of the Efate nest is manifested in these sections.

The thrust-type events in the Efate section compared to those in the Santo and Tanna sections (see Figure 9) are located at shallower depths and closer to the surface trace of the plate boundary and have more gently dipping slip vectors. The events appear to be located at shallower depths along the interplate boundary than those in the other two sections. If this were a general feature it would indicate that the seismically tectonic part of the interplate boundary is shallower and perhaps smaller in width than that beneath Santo. Further, the central plate activity located farther east might thus be associated with a down-dip zone of creep along the interplate boundary.

The Malekula section (Figure 5) exhibits an especially diffuse distribution of shallow events near the inferred position of the interplate boundary. Many of the events are shallower than those in the Efate section and are likely to represent a concentration of activity in the upper plate. Also in contrast to the Efate segment, no thrust-type earthquakes with magnitudes greater than about 6 have occurred beneath southern Malekula during the past 20 years. Thus, the high level of intra-plate activity in the southern Malekula-Efate area might be ascribed to loading of a locked section of the plate boundary. Alternatively, the features may be associated with the complex upper plate structures which characterize the area.

Magnitude 6 Earthquakes in the Efate Nest: 1978-1979 Sequence

Time-Space Development of the Sequence

The largest interplate events (Figure 4) located near Efate since the early 1960's occurred in 1973 (event 71), 1974 (event 72), and a sequence of three events (events 85, 87, and 88) which occurred within a year during 1978 and 1979. Two weeks after the first earthquake (September 1, 1978) of the 1978-1979 sequence the new local network commenced operation and for the first two weeks of operation (the last half of September, 1978) was augmented by a network of ocean bottom seismographs as shown in Figure 6. Since then, the network on land has continued to monitor the activity at a threshold level of about magnitude (mb) 3.0. The locations were done with a flat layered model and the Hypo 71 computer program (Lee and Lahr, 1975) and by methods similar to those described by Coudert et al. (1981). The hypocenter locations with OBS readings are accurate for much of the shallow zone of activity beneath the network, but the land network alone has good accuracy only for hypocenters relatively near the islands. In the New Hebrides arc there does not appear to be a large difference in locations determined with teleseismic from those determined with local data

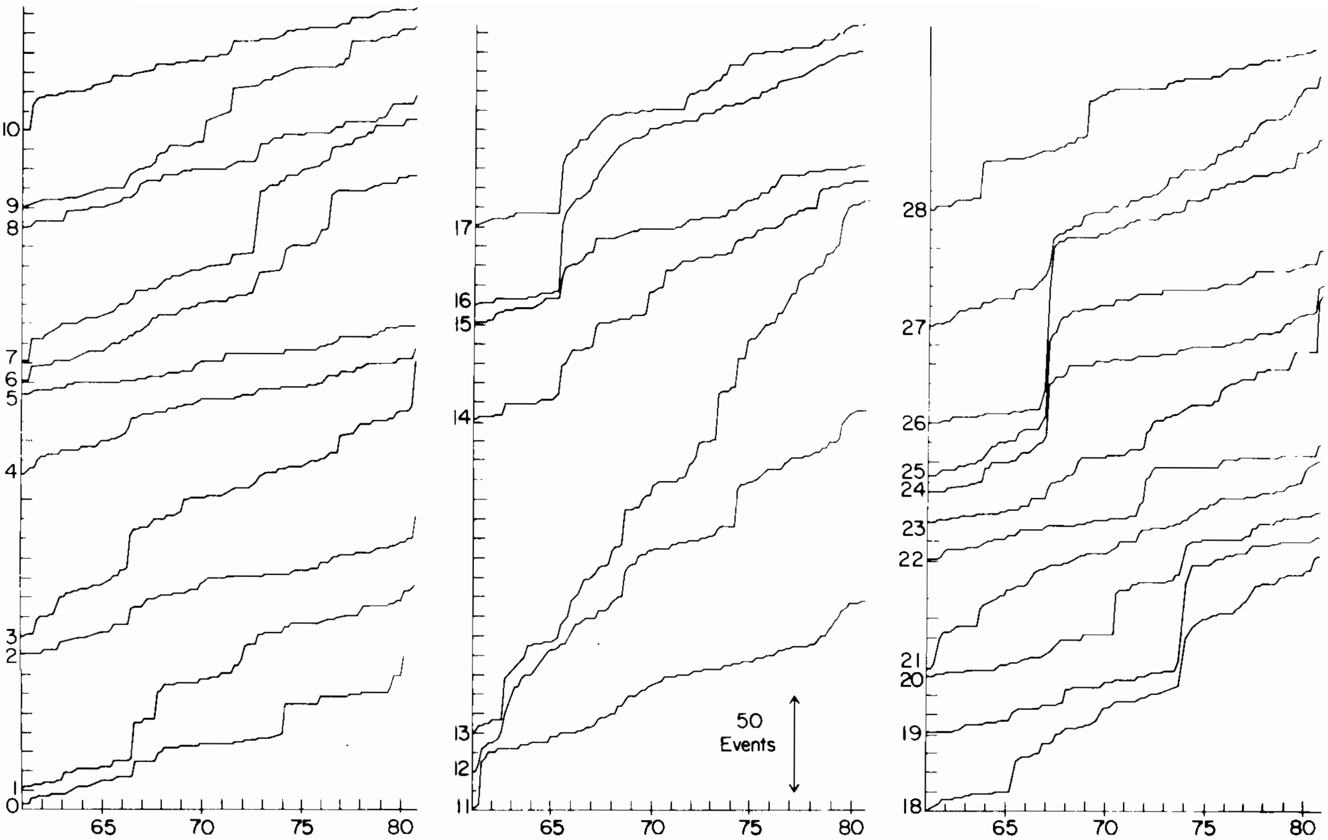


Figure 14. Curves of cumulative number versus time for all shallow earthquakes (depths < 70 km) reported in the PDE listings from 1961 through 1979 in each of the contiguous strips defined in Figure 13. The number at the origin of each curve identifies the strip along the arc from south (0) to north (28). The origin of each curve is arbitrary. The increment of time is 2 months.

(see also Coudert et al., 1981). The results are illustrated in Figure 6 and Figures 15-18. Although preliminary, these first results show several striking features outlined below which are not likely to be substantially modified by further analysis.

Foreshock and Aftershock Sequences. Although no foreshock sequence was detected by the first few stations of the network which began to operate several weeks before the September 1, 1978 event, an interesting foreshock sequence preceded the August 17, 1979 mainshock as illustrated in Figures 16 and 18. Several periods of increased activity occurred during the nine months preceding the August sequence. These include a cluster located beneath the trench (the December, 1978 cluster shown in Figure 15 but not included in the area covered by Figure 18), the striking cluster of March, 1979 (Figures 15 and 18), and increased activity during April and June, 1979. The March, 1979 cluster occurred near the boundary between the aftershock zones of the August 17 and August 26 events. During the nine days preceding the August 17 event seismic activity again increased in a series of clusters (Figure 18). The locations show a clear time-space migration in a northeast direction from the trench to the epicenter of the mainshock. The final clusters (Figures 15,16,18) occurred close to the epicenter of the mainshock. During the 3.5 hours preceding the mainshock no events were detected. The cascading or accelerating activity found by Jones and Molnar (1979) in their "stacking" of many foreshock sequences is not apparent in this particular foreshock sequence.

The zone of aftershocks of each of the August 17 and August 26 events expands in area during a period of several days as shown in Figure 17. The expanded areas overlap to some extent, but are mainly separate and contiguous. The areas of the expanded aftershock zones are, however, quite large relative to the magnitude and seismic moment of the events, as will be discussed in a following section.

Spatial Trends in Earthquake Locations and Structure of the Upper Plate. The spatial distribution of the aftershock zones and other events in the Efate nest seem closely related to predominant morphological features of the Efate salient, notably the east-west embayment of the

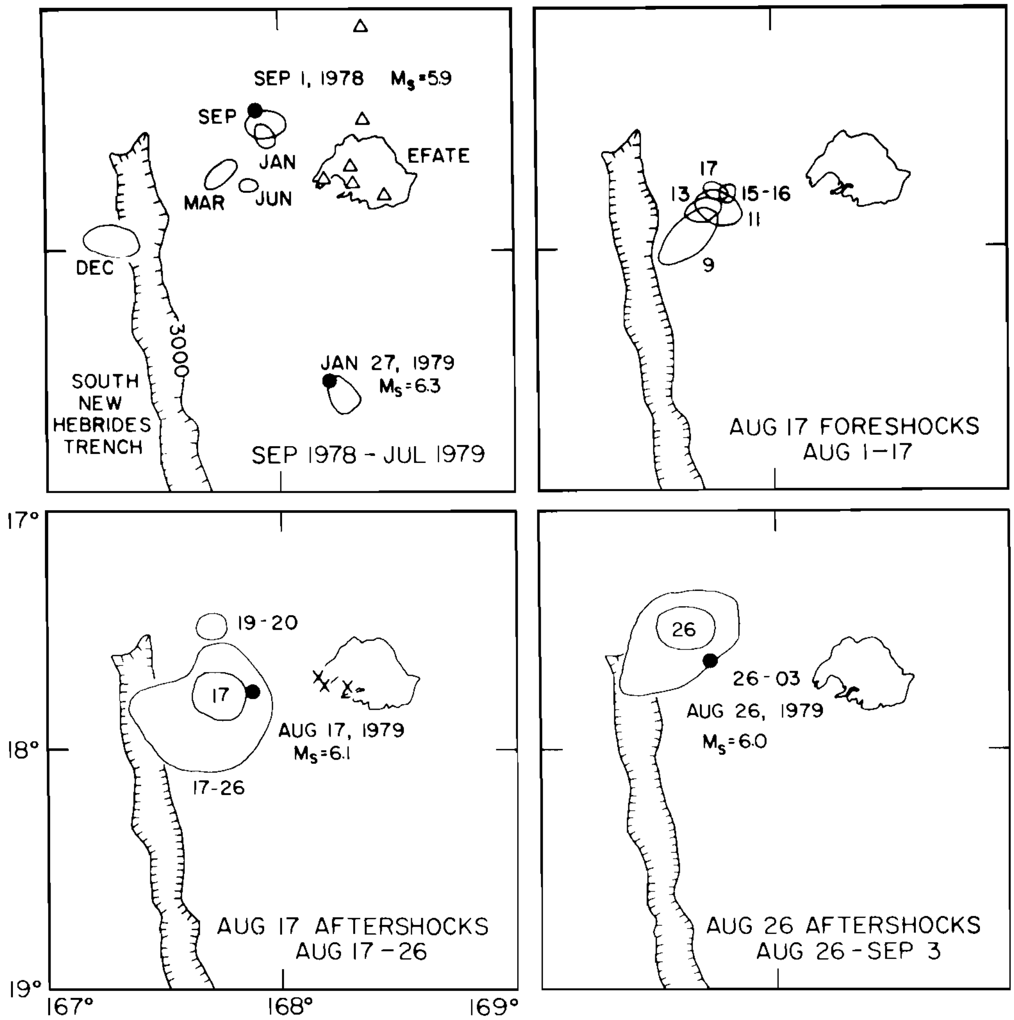


Figure 15. Maps of the Efate area summarizing the regions of concentrated earthquake activity during the period from September 1978 to September 1979. The event which initiated each cluster of earthquakes is shown by a solid circle whose diameter increases with the magnitude of the event. The extent of each cluster of earthquakes is shown by an irregular outline, and the dates are shown when the cluster was active. The X's show the location of the tiltmeter stations on Efate (lower left hand frame). The leveling array is near the westernmost two stations.

upper plate located north of Efate and the marked southwest-northeast trends defined by the bathymetry of the sea floor and the physiography and shape of northwest Efate Island. The aftershock zones of the 1978-1979 sequence are bounded to the north by the embayment feature, as shown in Figures 4,6 and 15. In Figure 6 the embayment is seen as a remarkable gap in activity. The southern limit of the aftershock zone of the August 17 event is approximately limited by the southwest-northeast trend in the bathymetry. The Pleistocene volcanoes of northern Efate (see Figure 2) are also located along that trend, as are the epicenters of the large 1944 and 1950 earthquakes (see Figure 12). Two other sets of features have the same southwest-northeast trend but are offset to the north of the morphological feature. One set (Figure 15) includes the epicenter of the September 1, 1978 event, the clusters of December, 1978 and March and June, 1979, and the boundary between the aftershock zones of the two August events. The second set includes the foreshocks of the August 17 event, the epicenters of the August 17 mainshock, the epicenters of the May 26, 1974 and February 16, 1966 (events 72 and 25, Figures 3 and 4), and the strike of the nearly vertical nodal plane of the focal mechanism of the February 16, 1966 event. The southwest-northeast trends and the east-west trends in the morphology thus seem to reflect important structures in the interacting plates,

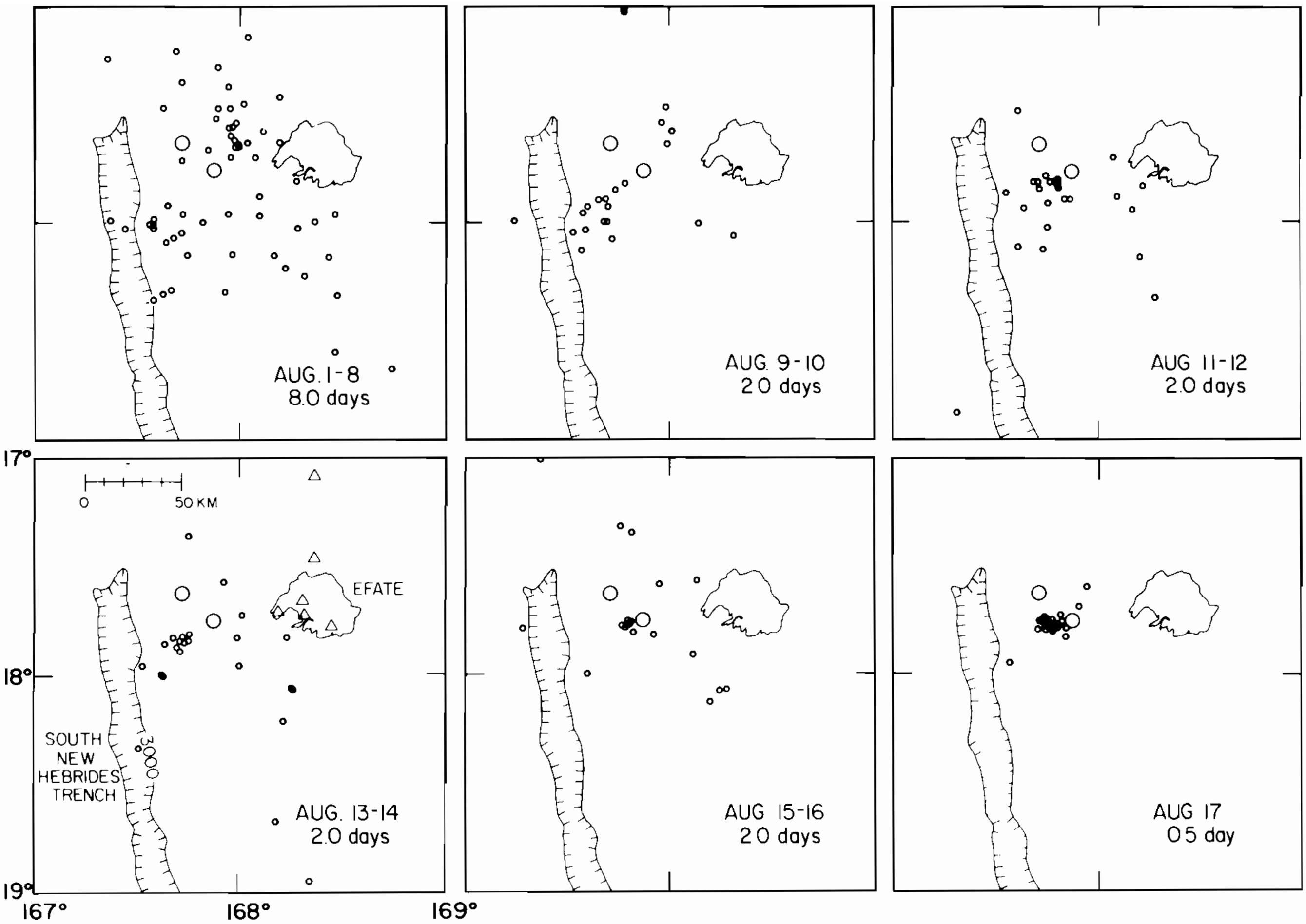


Figure 16. Maps of the Efate area showing the spatial distribution of foreshocks for six different intervals of time preceding the earthquake of August 17, 1979. Earthquake epicenters are representd by circles. The two large circles represent the locations for the mainshocks (Ms = 6) which occurred on August 17, 1979 and August 26, 1979. Triangles represent the locations of the six closest seismograph stations.

especially in the upper plate, which affect the seismicity.

The nest of seismicity near Efate is thus localized within the roughly triangular area of the Efate salient. It is likely that the salient constitutes a seismotectonic unit distinct from the southern Malekula block located to the north and possibly from the embayment of the upper plate between the Efate and Erromango salients (see Figure 3) located to the south.

Unusually Large Aftershock Areas for the August 1979 Earthquakes

Preliminary source parameters for five interplate events in the Efate nest (events 71,72,85,87 and 88) are listed in Table 2 together with those for two of the intraplate earthquakes in the area. The interplate earthquakes have body wave magnitudes (mb) between 5.5 and 5.9, surface wave magnitudes (Ms) between 5.9 and 6.1, and seismic moments (Mo) between 1×10^{27} and 3×10^{27} dyne-cm. The moments are determined by matching observed and synthetic long period P waveforms (Chinn and Isacks, 1979). The relationships between the USGS mb and Ms values, and between the Ms and Mo values are close to those found for many other interplate earthquakes (Nagamune, 1972; Geller, 1976; Kanamori and Anderson, 1975; Pucaru and Berckhemer, 1978) and indicate no obvious anomaly.

What does appear anomalous is the large size of the aftershock areas of the two events in August. Even the areas of the zones that developed only several hours after the mainshocks appear rather large. For example, if Abe's (1975) results are used to estimate an area from the P wave moment, the initial aftershock zones are too large by about a factor of 2 to 3 in area or about 10 km

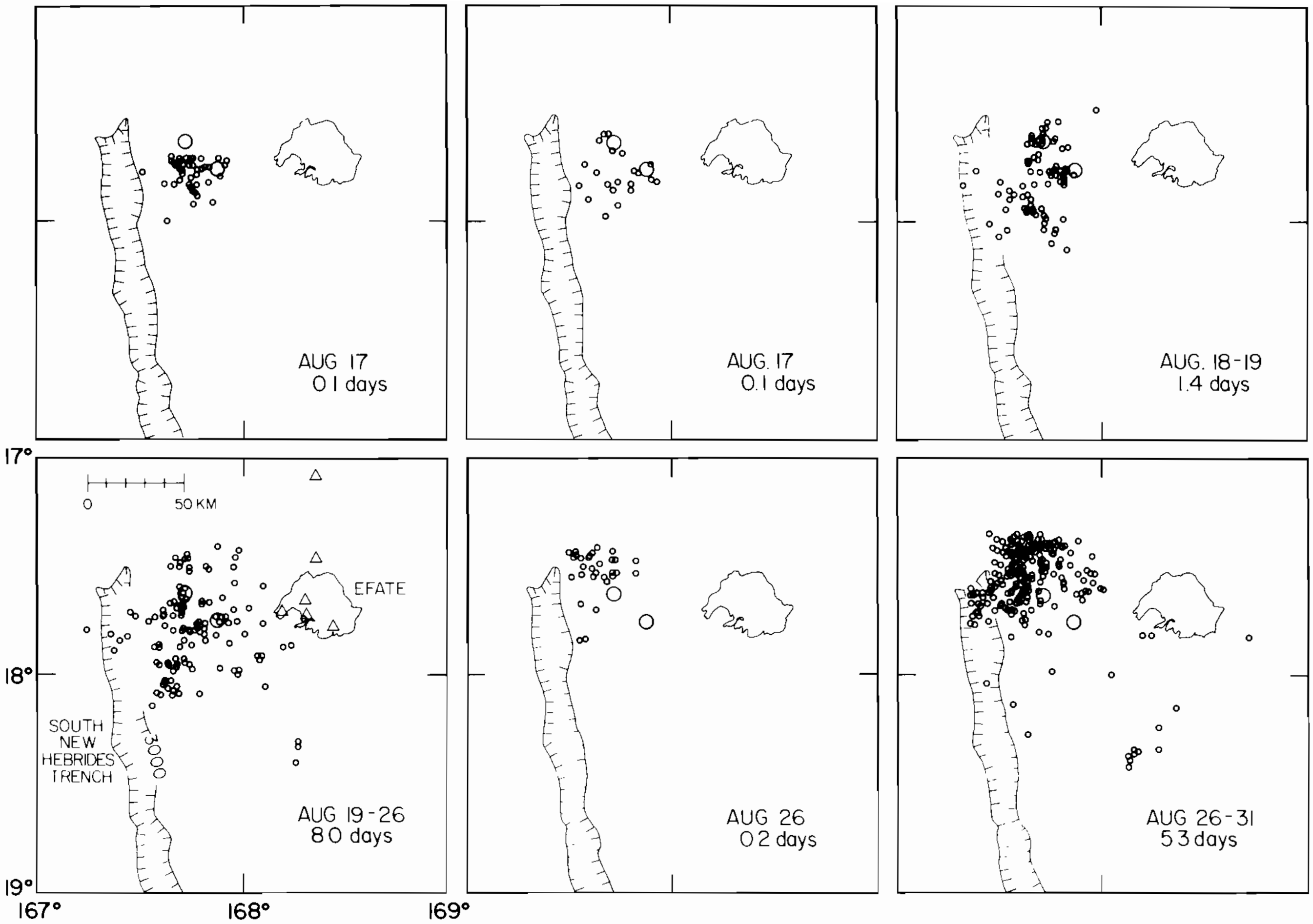


Figure 17. Maps of the Efate area showing the spatial distribution of aftershocks for six different time intervals following the earthquake of August 17, 1979. All symbols are the same as in Figure 16.

in linear dimensions. The expanded area developed in the days following the mainshock in each of the two cases is clearly anomalous. Some of this activity occurs within the sub-oceanic plate, as shown by the reliable location of hypocenters beneath and west of the axis of the trench (Figure 17), but many of the aftershocks are located near the interplate boundary.

Several limiting hypotheses can be considered to explain expanding aftershock zones. If the large area ruptured during the mainshock then, given the seismic moment, an anomalously low stress drop and small fault displacement are required. However, the 4 to 6 second durations of the source functions used in the syntheses of the long-period P waves are in better agreement with the dimensions of the small initial aftershock areas.

Alternatively, the small initial aftershock areas may coincide with the seismically ruptured areas in agreement with area-moment-magnitude relationships discussed above. If the larger aftershock area represents an area of creep between the smaller patches of stick-slip deformation responsible for the earthquakes, two possibilities can be considered: (1) creep is an ongoing phenomena that loads the isolated patches or "asperities" to seismic failure or, (2) the rupture of the patches is closely associated with major episodes of creep in time in the form of "slow" earthquakes. In particular, a post-seismic episode of creep might be reflected by the temporal expansion of the aftershock zone. The persistently high rate of occurrence of small earthquakes in the Efate nest (Figure 14) for the past 20 years could be taken to support a more continuous loading process wherein various sized isolated patches loaded by the surrounding creep fail seismically to account for the continuously high level of seismicity. The moderate-sized events of the 1978-1979 and earlier sequences may represent repetitive failure of the largest patches in the area. In this model, the expansion of the aftershock sequence would

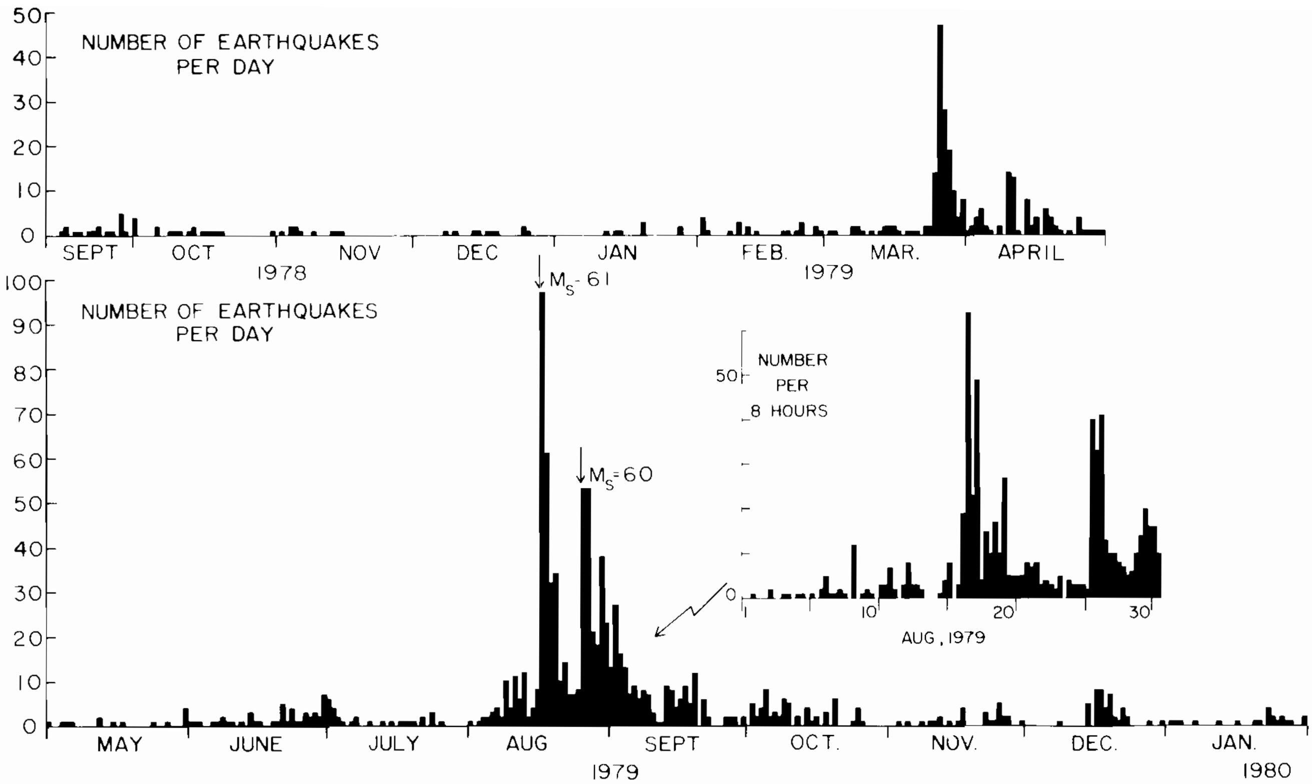


Figure 18. Histogram of earthquake activity versus time for the region defined by the aftershocks of the two August 1979 earthquakes. The insert shows clearly the small foreshock sequences which occurred on August 9, August 11, August 13, August 15, and August 17.

represent post-seismic diffusion of stress away from the patches of stick-slip behavior into the adjacent areas of creep surrounding the seismic rupture. The intervening areas of creep behavior also gives a mechanism to account for the time delays between the multiple events so characteristic of the seismicity.

Observations of surface deformations near the earthquakes might provide further evidence bearing on the question of whether a substantial component of creep occurs in episodes closely associated in time with the larger earthquakes or occurs as a longer term loading process. In the next section pertinent measurements of tilt obtained of Efate and Santo islands are described.

Tilt Measurements on Efate and Santo Islands

Tilt measurements in the New Hebrides have included the operation of a network of Kinemetrics bubble level sensors and releveling of two arrays of benchmarks located in southern Santo Island and western Efate Island (Isacks et al., 1978; Marthelot et al., 1980; Bevis and Isacks, 1981). The benchmark arrays were releveled from 2 to 4 times per year during the period since 1975 and provide measurements of tilt over baselines of nearly 1 km. The bubble level tiltmeters have proven rather noisy, with noise increasing with increasing period, but are effective as "strong-motion" instruments for signals with time scales of several hours to weeks. In the short period end of the spectrum, at periods of minutes to hours, the resolution of the recordings approaches about 0.2 microradians. The sensitivity of the detection is estimated to be about a microradian at the longer periods.

The largest events recorded near the tiltmeter network are the three events of the 1978-1979 sequence. The August 17, 1979 earthquake was the largest and closest to three tiltmeter stations on Efate Island. The straight-line distance from that hypocenter to the nearest station is about 40 km. The tiltmeter records of the three Efate stations were searched carefully for tilt changes before and after the earthquake. Except for offsets during the arrival of the seismic waves no pre- or post-seismic signals with time scales in the range of about ten minutes to a week were detected by visual inspection. Comparisons of the records at the three stations which are located at distances of 3 and 11 km from one another, provides a good means for identifying spurious drifts and other local noise sources. The same results were found for the other two more distant events of the 1978-1979 sequence. The coseismic offsets (none greater than about 1

Table 2: Source Parameters of Earthquakes Near Efate Island, 1973-1979

EQ No.	Mo/Dy/Yr	Depth[1], km	Magnitude mb	Magnitude Ms	Seismic Moment[1], X 10^{25} dyne-cm	Source Duration[1], sec
25	02/16/66	28	6.1	6.5	8.6	2
71	06/05/73	14	5.6	6.1	2.4	5
72	05/26/74	17	5.8	6.0	2.8	5
85	09/01/78	20	5.6	5.9	1.1	4
86	01/27/79	23	5.8	6.3	2.1	4
87	08/17/79	25	5.7	6.1	3.1	6
88	08/26/79	22	5.5	6.0	2.5	5

1 Determined by comparison of observed and synthetic long period P waveforms.

microradian) are not consistent in amplitude and polarity among the three stations and appear to be strongly influenced by the effects of seismic shaking of the ground at the tiltmeter site. This characteristic, also observed for smaller nearby and larger more distant events, discourages attempts to analyze co-seismic deformations.

Reliable long-term tilt measurements obtained by releveling the benchmark arrays are summarized in Figure 19. These measurements are shown by Isacks et al. (1978) and Bevis and Isacks (1981) to have a resolution of about 1 microradian. The Ratard array on southern Santo has accumulated little or no net tilt during the 5 years of measurements, but has exhibited two marginally significant excursions in 1976-1977 and 1979-1980. These excursions are possibly related to large earthquakes occurring several hundreds of kilometers north of Santo, as indicated in Figure 19.

The Efate array shows an accumulation of tilt at an overall average rate of about 1.5 microradians/year during the three year period from about 1977 to 1980. The tilt is downward toward the south, or in a direction more parallel than perpendicular to the strike of the arc. This progressive tilting is very well established by the consistency and redundancy of the array measurements as shown by Bevis and Isacks (1981). Thus, the accumulated tilt near the anomalous Efate segment is large whereas the tilt measured above the recently ruptured Santo' segment shows little accumulation of tilt.

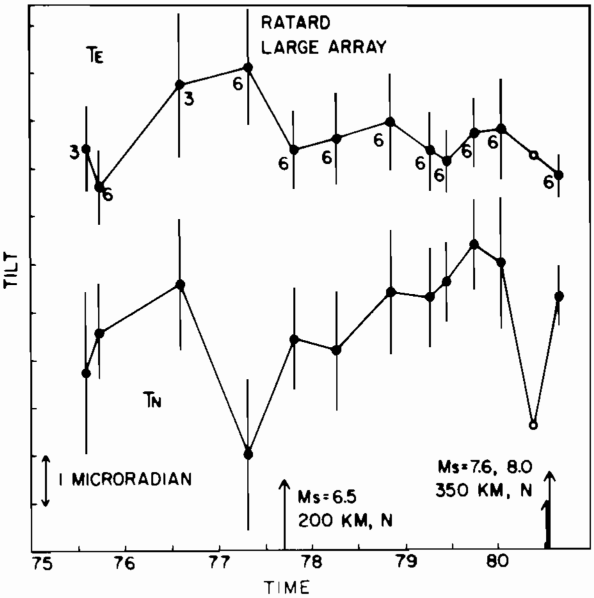
The detailed temporal development of tilt at Efate, although close to or within the noise level of the measurements, exhibits some noteworthy features that may be significant. These include a ramp-like signal in 1977 (Figure 19) which occurs during the year preceding the 1978-1979 sequence of moderately large earthquakes, a possible coseismic tilting of about 1 microradian upwards towards the west during a two month period spanning the August, 1979 earthquake sequence, and a recent reversal in the overall down-to-the-south trend of the tilt field. This last reversal is coincidental with the second large excursion seen at Ratard.

The data suggest that at a time scale of weeks surrounding the August, 1979 events (and the September, 1978 event) the tilt changes measured on Efate are not as important as the long term accumulation of tilt over the 4 years of measurements. If creep plays an important role in the interplate slippage of the Efate segment, and if the tilt signal is of tectonic origin, then the tilt data favor a long term development of creep rather than a type of slow earthquake where a large component of creep accompanies an episode of seismic slippage.

Conclusions

This study highlights spatial variations in the seismicity of the central New Hebrides that seem to be closely associated with the morphology and structure of each of the converging plates. Bathymetric irregularities of the sub-oceanic plate, notably the northern and southern bounding ridges of the DFZ, may act as asperities in the development of fault zone slippage and may also be responsible for uplift and block-like deformations of the upper plate. Irregularities in the shape and structure of the upper plate are probably indicative of variations in the width, physical state, and/or lithology of the interplate boundary. A notable example is the seaward protrusion of the upper plate in the area of Santo and Malekula islands, a protrusion that appears to increase the width of the interplate boundary. The interaction of this protrusion with the subducted topography of the DFZ may produce an area of increased coupling along the interplate boundary and increased compressional deformation in the upper plate.

In the central New Hebrides four main seismotectonic segments of the interplate boundary can be tentatively distinguished on the basis of the morphological features and characteristics of seismicity. The Santo-northern Malekula segment, including a major part of the protruding upper plate and its interaction with the DFZ, is the only segment for which seismic rupture of the interplate boundary by large earthquakes is well-established. The southern Malekula and Erromango segments have been very quiet during the past 20 years except for a magnitude 7 interplate event occurring in the Erromango segment in 1961. In contrast the segment comprising the Efate salient, located between the two quiet segments, has exhibited a persistently high rate of occurrence of small and moderate-sized earthquakes. The extent to which seismic rupture of these segments has taken place by large earthquakes occurring before 1960 is still highly uncertain, but the record suggests



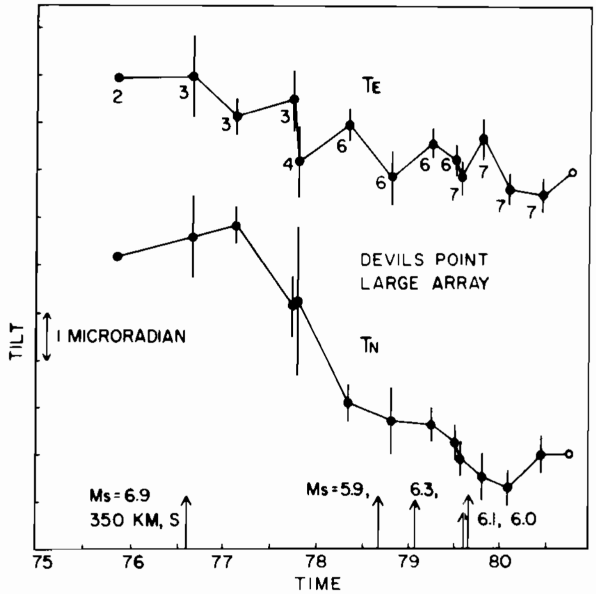


Figure 19. The north and east components of tilt observed at Devils Point, Efate Island and Ratard, Santo Island as determined by first-order relevelings of the arrays of benchmarks. Each array is clearly 1 km in dimension. The Devils Point array is located near the westernmost edge of Efate Island and the Ratard array is located near the central part of the southern coast of Santo Island (see Bevis and Isacks, 1981). The tilt components were calculated using an unweighted least squares analysis, with the September 1979 levelings as references. Also shown is the seismicity within 100 km of each array (open circles) and major events within 400 km of either array (arrows).

significant deficiencies in the number, magnitudes, and moments of events during the past 75 years in comparison with those which ruptured the Santo-northern Malekula segments in 1965-1974. The basic question is whether the variations in seismicity during the past 20 years represent different parts of a common earthquake cycle or different means by which slippage between the converging plates is accommodated.

The persistence of the anomalous seismicity of the Efate salient over 20 years suggests a long term anomaly related to the mode of plate boundary slippage. Approximate calculations show that the seismic activity, though high in respect to rate of occurrence, does not accumulate enough total moment to accommodate plate boundary slippage of 10 cm/yr in the entire Efate segment. Thus either the activity will substantially increase, or larger earthquakes will eventually occur, or a substantial fraction of the slippage occurs by creep. The last possibility is supported by the unusually large aftershock areas found for the recent sequence of moderately large earthquakes in the Efate salient in 1979. The data for this sequence suggest that relatively small, separate areas or patches of the plate boundary fail seismically and activate aftershocks in larger intervening areas. The time delays between mainshocks in the sequence and the time-space development of the aftershock sequence presumably reflect the non-elastic redistribution of stress. However, tilt measurements do not reveal evidence for large pre- or post-seismic creep movements at time scales of days to months around the times of the moderately large mainshocks of the sequence. The tilt measurements do, however, indicate significant accumulation of tilt at an average rate of 1.5 microradians per year during a 3-4 year period.

These observations can be qualitatively accounted for (somewhat speculatively) by a model in which the interplate boundary in the Efate salient includes a significant area along which slippage occurs largely by creep but within which seismic slippage occurs on relatively small separated patches. The faulting of these patches, after being loaded to failure by the surrounding creep, accounts for the high level of moderately large and moderately small earthquakes. The area of seismic slippage may thus be smaller than in other segments. There is also a suggestion that the seismically active portion of the interplate boundary may be shallower and smaller in width than in other segments. The along-strike direction of tilting observed on Efate could be accounted for in terms of deformation near a transition from the creeping Efate segment to a locked Efate-Erromango segment and/or to a locked southern Malekula segment. Thus the Efate segment may be analogous to the central California part of the San Andreas fault system (Allen, 1968) in respect to being a zone of predominantly creep behavior

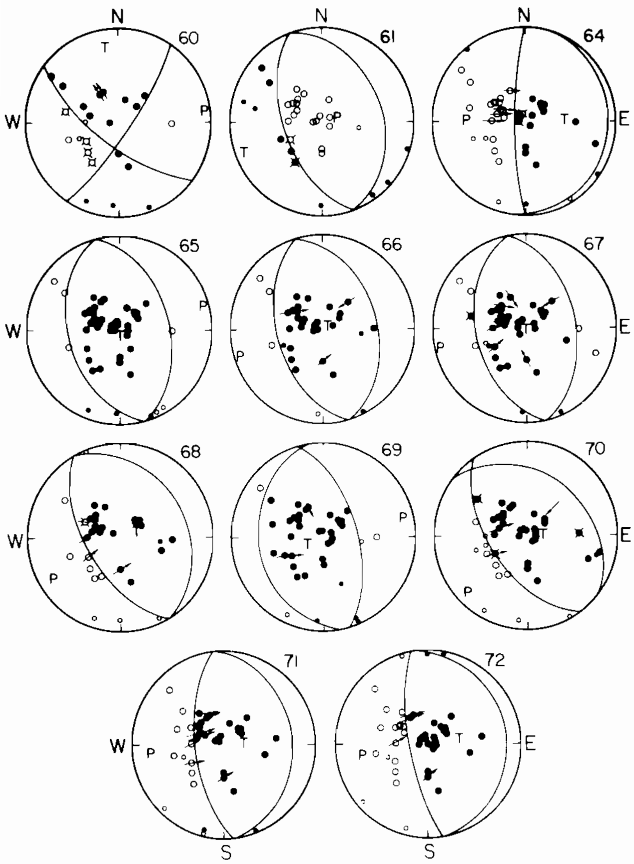


Figure 20. Focal mechanism solutions for earthquakes in the central New Hebrides, listed in Table 1. Compressions are filled circles, dilatations open circles, small symbols indicate less reliable readings, X's are readings judged to be near a nodal plane, and the arrows show first motions for S waves.

located between zones of predominantly stick-slip behavior. In particular, the moderate-sized earthquakes in the Efate nest may be comparable to the Parkfield earthquakes as recently described by Bakun and McEvilly (1979) and Lindh and Boore, 1981.

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References

Abe, K., 1975, Reliable estimation of the seismic moment of large earthquakes, J. Phys. Earth, 23, 381-390.

Allen, C.R., 1968, The tectonic environments of seismically active and inactive areas along the San Andreas fault system, in Proceedings of Conference on Geologic Profiles of San Andreas Fault System, Stanford University Publ., Geol. Sci., v. 11, 70-80.

Bakun, W.H., and T.V. McEvilly, 1979, Earthquakes near Parkfield, California: Comparing the 1934 and 1966 sequences, Science, 205, 1375-1377.

Bloom, A.L., and F.W. Taylor, 1977, Emerging Pleistocene islands in the New Hebrides island arc, Abstracts with Programs, 1977 Annual Meeting, 9, #7, Geol. Soc. Amer., p. 902.

Carney, J.N., and A. Macfarlane, 1978, Lower to Middle Miocene sediments on Maewo, New Hebrides, and their relevance to the development of the outer Melanesian arc system, Bull. Aust. Explor. Geophys., 9, 123-130.

Chase, C.G., 1971, Tectonic history of the Fiji Plateau, Geol. Soc. Am. Bull., 82, 3087-3110.

Chinn, D.S., and B.L. Isacks, 1979, Accurate depths of shallow earthquakes in the New Hebrides island arc, (Abstr.), Earthquake Notes, 49, 84.

Choudhury, M.A., G. Poupinet and G. Perrier, 1975, Shear velocity from differential travel times of short-period ScS-P in New Hebrides, Fiji-Tonga, and Banda Sea regions, Bull. Seism. Soc. Am., 65, 1787-1796.

Chung, W.-Y., and H. Kanamori, 1978a, A mechanical model for plate deformation associated with aseismic ridge subduction in the New Hebrides arc, Tectonophysics, 50, 29-40.

Chung, W.-Y., and H. Kanamori, 1978b, Subduction process of a fracture zone and aseismic ridges -- The focal mechanism and source characteristics of the New Hebrides earthquake of January 19, 1969 and some related events, Geophys. J.R. Astr. Soc., 54, 221-240.

Coudert, E., B.L. Isacks, M. Barazangi, R. Louat, R. Cardwell, A. Chen, J. Dubois, G. Latham, and B. Pontoise, 1981, Spatial distribution and mechanisms of earthquakes in the southern New Hebrides arc from a temporary land-OBS seismic network and world-wide observations, J. Geophys. Res., (in press).

Daniel, J., 1978, Morphology and structure of the southern part of the New Hebrides island arc

system, J. Phys. Earth, 26, suppl., S181-S190.
Daniel, J., C. Jouannic, B. Larue, and J. Recy, 1977, Interpretation of D'entrecasteaux zone (north of New Caledonia), in International Symposium on Geodynamics in South-West Pacific, Noumea, New Caledonia, Editions Technip, Paris, France, pp. 117-124.
Dubois, J., 1971, Propagation of P waves and Rayleigh waves in Melanesia: Structural implications, J. Geophys. Res., 76, 7217-7240.
Dubois, J., J. Launay, J. Recy, and J. Marshall, 1977, New Hebrides trench: Subduction rate from associated lithospheric bulge, Canadian J. of Earth Sci., 14, 250-255.
Dubois, J., F. Dugas, A. Lapouille, and R. Louat, 1978, The troughs at the rear of the New Hebrides island arc: Possible mechanisms of formation, Canad. Journ. of Earth Sciences, 15, 351-360.
Ebel, J.E., 1980, Source processes of the 1965 New Hebrides Islands earthquakes inferred from teleseismic wave forms, Geophys. J.R. Astr. Soc., 63, 381-403.
Falvey, D.A., 1978, Analysis of paleomagnetic data from New Hebrides, Bull. Aust. Explor. Geophys., 9, 117-123.
Frankel, A., and W. McCann, 1979, Moderate and large earthquakes in the South Sandwich arc: Indicators of tectonic variation along a subduction zone, J. Geophys. Res., 84, 5571-5577.
Geller, R.J., 1976, Scaling relations for earthquake source parameters and magnitudes, Bull. Seism. Soc. Am., 66, 1501-1523.
Geller, R.J., and H. Kanamori, 1977, Magnitudes of great shallow earthquakes from 1904 to 1952, Bull. Seism. Soc. Am., 67, 587-598.
Geological Map of the New Hebrides Condominium, scale 1:1,000,000, Ministry of Overseas Development (Directorate of Overseas Surveys), D.O.S. 1196, British Residency, Vila, New Hebrides.
Gill, J., and M. Gorton, 1973, A proposed geological and geochemical history of eastern Melanesia, in The Western Pacific, Island Arcs, Marginal Seas, Geochemistry, P.J. Coleman, ed., Crane, Russals and Co., New York, p. 543-566.
Gutenberg, B., and C.F. Richter, 1954, Seismicity of the Earth and Associated Phenomena, Princeton Univ. Press, 310 pp.
Hanks, T.C., and H. Kanamori, 1979, A moment magnitude scale, J. Geophys. Res., 84, 2348-2350.
Hasegawa, A., N. Umino, and A. Takagi, 1978, Double-planed deep seismic zone and upper mantle structure in the northeastern Japan arc, Geophys. J.R. Astr. Soc., 54, 281-296.
Ibrahim, A., B. Pontoise, G. Latham, B. Larue, T. Chen, B. Isacks, J. Recy, and R. Louat, 1980, Structure of the New Hebrides Arc - Trench system, J. Geophys. Res., 85, 253-266.
Isacks, B.L., and M. Barazangi, 1977, Geometry of Benioff zones: Lateral segmentation and downwards bending of the subducted lithosphere in M. Talwani and W.C. Pitman, III (editors), Island Arcs, Deep Sea Trenches, and Back-Arc Basins, Maurice Ewing Ser., 1, AGU, Washington, D.C., 99-114.
Isacks, B.L., G. Hade, R. Campillo, M. Bevis, D. Chinn, J. Dubois, J. Recy, and J.L. Saos, 1978, Measurements of tilt in the New Hebrides Island Arc, in Proceedings of Conference VII, Stress and Strain Measurements Related to Earthquake Prediction, U.S. Geological Survey Open-File Report 79-370, 176-221.
Isacks, B.L., and P. Molnar, 1971, Distribution of stresses in the descending lithosphere from a global survey of focal mechanism solutions of mantle earthquakes, Rev. Geophys. Space Phys., 9, 103-174.
Johnson, T., and P. Molnar, 1972, Focal mechanisms and plate tectonics of southwest Pacific, J. Geophys. Res., 77, 5000-5032.
Jones, L.M., and P. Molnar, 1979, Some characteristics of foreshocks and their possible relationship to earthquake prediction and premonitory slip on faults, J. Geophys. Res., 84, 3596-3608.
Jouannic, C., F.W. Taylor, A.L. Bloom, and M. Bernat, 1980, Late Quaternary uplift history from emerged reef terraces on Santo and Malekula islands, central New Hebrides island arc, United Nations E.S.C.A.P., CCOP/SOPAC, Tech. Bull. no. 3, 91-108.
Kanamori, H., and D.L. Anderson, 1975, Theoretical basis of some empirical relations in seismology, Bull. Seism. Soc. Am., 65, 1073-1095.
Kanamori, H., 1977, The energy release in great earthquakes, J. Geophys. Res., 82, 2981-2987.
Karig, D.E., and J. Mammerickx, 1972, Tectonic framework of the New Hebrides island arc, Marine Geology, 12, 187-205.
Karig, D.E., and G.F. Moore, 1975, Tectonic complexities in the Bonin Arc system, Tectonophysics, 27, 97-118.
Kelleher, J., J. Savino, H. Rowlett, and W. McCann, 1974, Why and where great thrust earthquakes occur along island arcs, J. Geophys. Res., 79, 4889-4898.
Kelleher, J., and W. McCann, 1976, Buoyant zones, great earthquakes and unstable boundaries of subduction, J. Geophys. Res., 81, 4885-4896.
Kim, S.G., and O.W. Nuttli, 1975, Surface-wave magnitudes of Eurasian earthquakes and explosions, Bull. Seism. Soc. Am., 65, 693-709.
Lee, W.H.K., and J.C. Lahr, 1975, HYPO-71 (Revised): A computer program for determining hypocenters, magnitudes, and first motion patterns of local earthquakes, U.S. Geol. Survey, Open File Report 75-311, Menlo Park, California, 113 p.
Lindh, A.G., and D.M. Boore, 1981, Control of rupture by fault geometry during the 1966 Parkfield Earthquakes, Bull. Seism. Soc. Am., 71, 75-116.
Louat, R., J. Dubois, and B. Isacks, 1979, Evidence for anomalous propagation of seismic

waves within the shallow zone of shearing between the converging plates of the New Hebrides subduction zone, Nature, 281, 293-295.

Luyendyck, B., W.B. Bryan, and P.A. Jezek, 1974, Shallow structure of the New Hebrides island arc, Geol. Soc. Am. Bull., 85, 1287-1300.

Mallick, D.I.J., 1973, Some petrological and structural variations in the New Hebrides, in P.J. Coleman (ed.), The Western Pacific: Island Arcs, Marginal Seas, Geochemistry, Crane, Russals, and Co., New York, p. 193-211.

Mallick, D.I.J., and G. Neef, 1974, Geology of Pentecost, New Hebrides Geological Survey, Regional Report, British Residency, Vila, New Hebrides, 103 pp.

Mallick, D.I.J., and D. Greenbaum, 1977, Geology of Southern Santo, New Hebrides Geological Survey, Regional Report, 84 p.

Mammerickx, J., T.E. Chase, S.M. Smith, and I.L. Taylor, 1971, Bathymetry of the South Pacific, Chart 12, Scripps Inst. of Oceanography, La Jolla, California.

Marthelot, J.-M., E. Coudert, and B.L. Isacks, 1980, Tidal tilting from localized ocean loading in the New Hebrides island arc, Bull. Seism. Soc. Am., 70, 283-292.

Marthelot, J.-M., and B.L. Isacks, 1980, Space-time distribution of shallow and intermediate earthquakes in the New Hebrides island arc, (Abstr.), EOS, Trans. Am. Geophys. Union, 61, 288.

McCann, W.R., 1980, Seismic potential and seismic regimes of the Southwest Pacific, preprint.

Mitchell, A.H.G., and A.J. Warden, 1971, Geological evolution of the New Hebrides island arc, J. Geol. Soc. London, 127, 501-529.

Mogi, K., 1969, Relationship between the occurrence of great earthquakes and tectonic structure, Bull. Earthquake Res. Inst., Tokyo Univ., 47, 429-451.

Nagamune, T., 1972, Magnitudes estimated from body waves of great earthquakes (in Japanese), Quarterly J. of Seismol., 47, 1-8.

Nakamura, K., 1977, Volcanoes as possible indicators of tectonic stress orientation: Principle and proposal, J. Volc. Geoth. Res., 2, 1-16.

Pascal, G., B.L. Isacks, M. Barazangi, and J. Dubois, 1978, Precise relocations of earthquakes and seismotectonics of the New Hebrides island arc, J. Geophys. Res., 83, 4957-4973.

Pucaru, G., and H. Berckhemer, 1978, A magnitude scale for very large earthquakes, Tectonophysics, 49, 189-198.

Ravenne, C., G. Pascal, J. Dubois, F. Dugas, and L. Montadert, 1977, Model of a young inter-oceanic arc: The New Hebrides island arc, in International Symposium on Geodynamics in South-West Pacific, Noumea, New Caledonia, Editions Technips, Paris, France, 63-79.

Roca, J.-L., 1978, Essai de determination du champ de contraintes dans l'archipel des Nouvelles Hebrides, Bull. Soc. Geol. France, 20, 511-519.

Rothe, J.P., 1969, The Seismicity of the Earth, UNESCO, Paris, 336 p.

Santo, T., 1970, Regional study on the characteristic seismicity of the world, Part III, New Hebrides island region, Bull. Earthq. Res. Inst., 48, 1-18.

Stephens, C.D., 1978, A short-term microearthquake survey in the central New Hebrides island arc: Closing the gap, M.S. Thesis, Cornell University, 46 pp.

Taylor, F.W., B.L. Isacks, C. Jouannic, A.L. Bloom, and J. Dubois, 1980, Coseismic and Quaternary vertical tectonic movements, Santo and Malekula Islands, New Hebrides Island Arc, J. Geophys. Res., 85, 5367-5381.

Thatcher, W., and J.B. Rundle, 1979, A model for the earthquake cycle in underthrust zones, J. Geophys. Res., 10, 5540-5556.

Utsu, T., 1974, Space-time pattern of large earthquakes occurring off the Pacific coast of the Japanese Islands, J. Phys. Earth, 22, 325-342.

Uyeda, S., and H. Kanamori, 1979, Back-arc opening and the mode of subduction, J. Geophys. Res., 84, 1049-1061.

York, J.E., 1977, Seismotectonics in intraplate and interplate regions: Eastern North America, eastern Taiwan, China, and the New Hebrides, Ph.D. Thesis, Cornell University, 152 pp.