Propagation of *P* Waves and Rayleigh Waves in Melanesia: Structural Implications

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The records of the Orstom seismological stations in New Caledonia and New Hebrides were used for a study of the propagation of *P* waves and Rayleigh waves in the South Pacific seismic belts, particularly in the New Hebrides area. After a brief study of the seismicity of the New Hebrides arc, it is shown that *P* arrivals from deep-focus earthquakes in Fiji and from the Longshot explosion are late in Melanesia, and that local differences exist between stations. A study of *P*-wave propagation from normal earthquakes of the New Hebrides implies an important difference between the stable areas of New Caledonia in the Loyalty Islands and the seismic area of New Hebrides. It is shown that the propagation has an azimuthal influence, and this influence is interpreted as a lithospheric plate effect and as a root of mountain effect. In the second part of the paper, Rayleigh-wave dispersion is studied by a systematic investigation of influence coefficients of layer parameters. The structural framework of the area between Australia and the andesite line is presented. In this area the crust is thicker than in oceanic regions, being about 20 km on the average, about 15 km in basins, and 25–30 km under the Coral Sea platform and ridges. Theoretical models to explain body-wave and Rayleigh-wave propagation give results in fair agreement with observations in the New Hebrides anomalous area. It is found that there is a crust where *P*-wave velocity is high on a low-velocity upper mantle. Speed increases with depth from 7.4 km/sec at a depth of 20 km to 8.1 km/sec at a depth of 120 km. In the third part of this paper the results and geological implications are discussed in the sea-floor spreading and continental-drift concepts. There is fair agreement with the hypothesis that spreading occurs from the Indian Antarctic and Pacific Antarctic ridges. The area to the east of New Hebrides (North Fiji basin, Fiji Islands, and Lau ridge) would be a large stable plate between two mobile plates of lithosphere converging and thrusting under the Tonga-Kermadec and New Hebrides arcs. According to the propagation of seismic waves under New Caledonia and the Loyalty Islands and to geological observations, it is suggested that New Caledonia and Norfolk ridge would be an island arc whose activity ceased in the Oligocene.

Over the last 10 years, the records of the 30 seismological stations operating in the southwest Pacific have greatly extended knowledge of the structure of this area. With data from this network, the seismicity of the island arcs of New Guinea, Solomon Islands, New Hebrides, Tonga, Kermadec, and New Zealand (areas 13 and 14 of *Gutenberg and Richter* [1965]) have been described in basic papers by *Adams* [1963] and *Eiby* [1964] (deep-focus earthquakes of New Zealand); *Sykes* [1964] (deep-focus earthquakes of New Hebrides); *Sykes* [1966] (earthquakes of the Fiji–Tonga–Kermadec seismic belt); *Sykes et al.* [1969] and *Mitronovas et al.* [1969] (spatial distribution of earthquakes of the Fiji–Tonga–Kermadec seismic belt); and *Denham* [1969] (seismicity of New Guinea and the Solomon Islands).

Study of the propagation of surface waves has revealed the main structural features of the area between Australia and the andesite line. Such studies have been made in Australia [*Bolt and Niazi*, 1964], in Fiji [*Kuo et al.*, 1962; *Hunkins and Kuo*, 1965], in New Zealand [*Thomson and Evison*, 1962], and in New Guinea [*Brooks*, 1969]. The paths cross areas of different structural type: oceanic basins, continental areas, oceanic trenches, and island arcs, and the results obtained show intermediate types of dispersion. *Santo* [1961*a*, *b*, 1963] divides the Pacific into regions, giving each region a numerical classification that depends on the Rayleigh-wave dispersion (1 is a typical oceanic structure, and 7 is a typical continental structure). The divisions of the area studied here have classification from 3 to 5.

The 7 stations of the Orstom network (see Figure 5) lie in the center of this area and are suitably placed to supply details of the structure of the New Hebrides and New Caledonia arcs. Thus the stations at Luganville and Port Vila have given evidence of anomalous propagation of P waves along the New Hebrides arc, and the long-period instruments at Port Vila, Noumea, and Koumac have given models of the structure of the New Hebrides arc, the New Caledonia ridge, and its southward extension as the Norfolk ridge [*Dubois*, 1965, 1968].

Seismic profiles in the Coral Sea basin [*Shor*, 1967] show sediments, 'second layer,' and oceanic crust all thicker than normal for an oceanic station. A normal mantle lies at a depth of 19 km. Rayleigh-wave dispersion is in good agreement with these data.

Observations of body waves and Rayleigh waves in the New Caledonia and New Hebrides area will be presented and used to investigate the structure of the region in the context of expansion and ocean-bottom spreading. The basic observations of *Oliver and Isacks* [1967] in the Tonga–Kermadec region led them to propose the hypothesis of a sinking lithosphere underthrusting an island arc. *Molnar and Oliver* [1969], in their global study of lateral variation of attenuation in the upper mantle, observed a poor and atypical propagation of S_n across the convex side of the New Hebrides arc, over two paths between the New Hebrides and Charters Towers. The stations at Noumea and Koumac lie close to the convex side of the arc and are suitably placed to enable a study of this problem.

Following *Cullen* [1967] and *Summerhayes* [1967], I have assumed that expansion in this area would occur toward the north-northeast (New Hebrides) and to the north (Solomon Islands) from the Indian Antarctic ridge. Geological features, seismic data, and the propagation of seismic waves suggest that New Caledonia, the Loyalty Islands, and the Norfolk ridge compose an island arc whose activity ceased in the Oligocene.

Karig [1970] and *Packham and Falvey* [1971] have recently used marine geophysical data (gravity especially) to argue that marginal seas are areas where oceanic crust is being formed immediately behind the andesitic arc, by a process analogous to that believed to operate on mid-oceanic ridges.

In this paper we shall first study the seismicity of the New Hebrides island arc and the propagation of P and S waves, and secondly we shall try to construct theoretical models using the parameters obtained in the first part and to see if they are in general agreement with observed Rayleigh-wave dispersion and the concept of sea-floor spreading and continental drift.

Seismicity of the New Hebrides Islands arc. A belt of great seismic activity extends from 10°S to 24°S. Important volcanic activity is associated with the activity (the active volcanoes of Gaua, Ambrym, Lopevi, Tanna, and Matthews). The belt of epicenters is 200 km in width, and the shallow and normal earthquakes lie near its western side, along and near the oceanic trench that lies parallel to the island arc. The intermediate foci (100 to 200 km in depth) lie beneath the line of active volcanoes on the eastern side of the belt. The deep-focus earthquakes (600 km) are confined to a limited area under the North Fiji basin, 300 km to the east of Banks Islands. No earthquake between 300 km and 600 km in depth has been observed during the last 10 years.

Epicenters determined by the International Seismological Center (ISC) for earthquakes from 1961 to 1966, inclusive, are shown in Figures 1 and 2.

Three vertical sections oriented perpendicular to the arc show hypocenters projected from distance within ±200 km of each section (Figure 3). Only shocks for which data from more than 20 stations were available have been shown. The accuracy claimed in ISC bulletins was always better than ±10 km in position and ±20 km in depth. The depths obtained from P-wave arrivals were compared with those based on the pP–P interval as a further check on their accuracy. Table 1 shows that for eight arbitrarily chosen earthquakes in the years 1965–1966, the values of depth obtained by the two methods agree within the stated limits of error.

Thus we shall accept shocks for which the

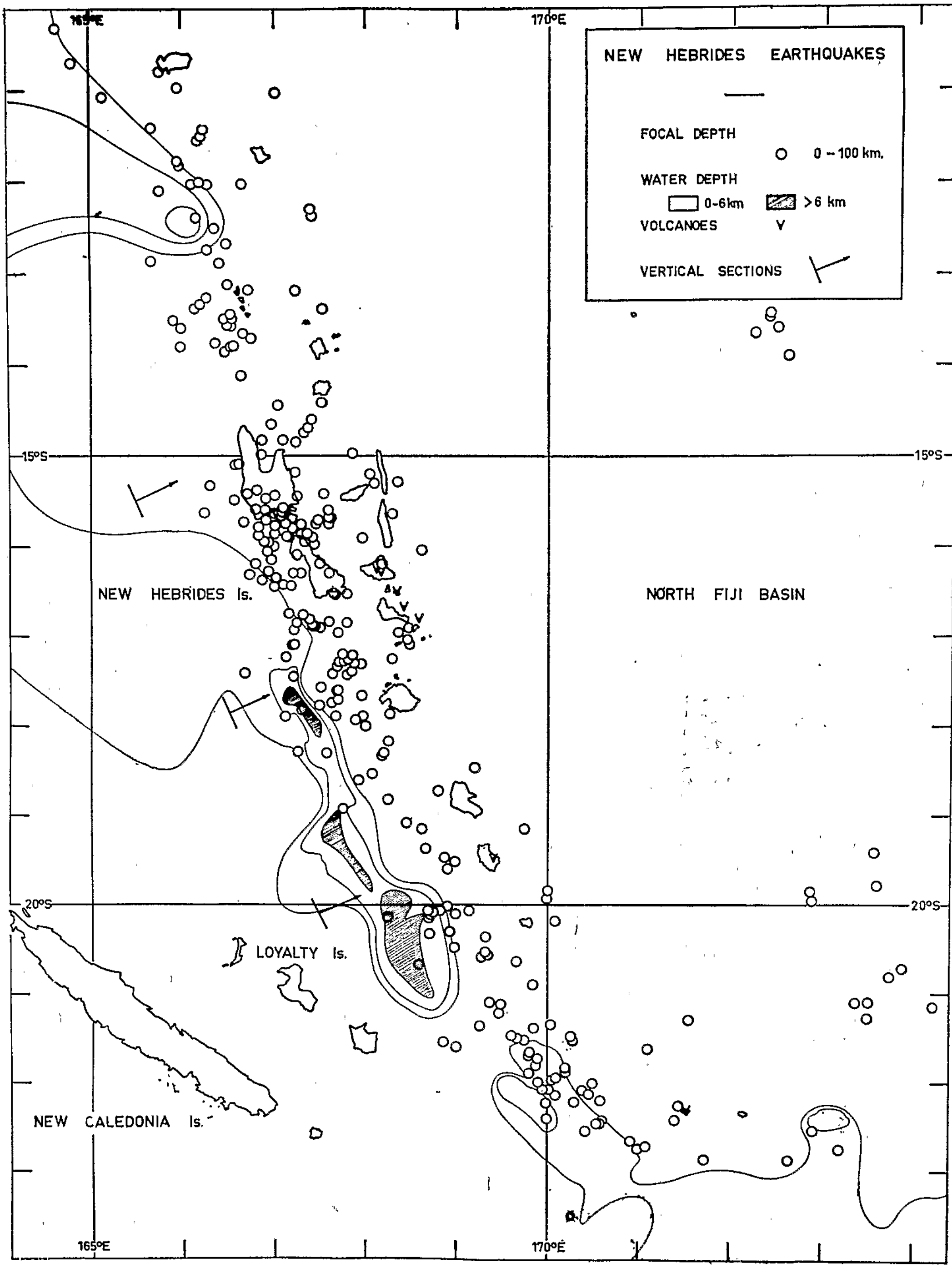

Fig. 1. Earthquakes in the New Hebrides region. Focal depth is 0–100 km.

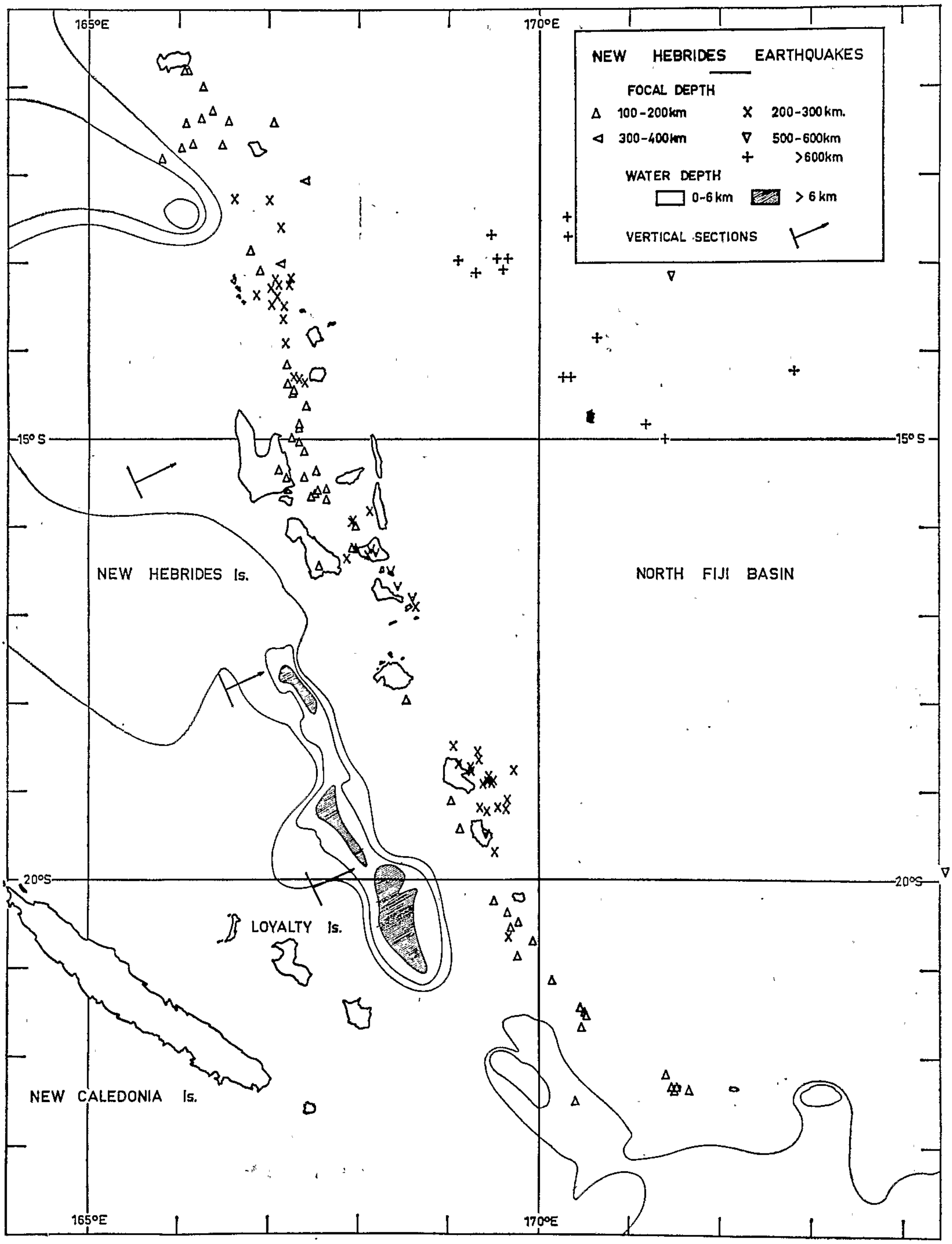

Fig. 2. New Hebrides earthquakes. Focal depth is greater than 100 km.

error in the depth obtained from *P* readings is less than ±20 km; data from more than 20 stations are used.

The map and vertical sections show the foci concentrated near a plane dipping beneath the island arc at about 55°. It is remarkable that there is a gap in focal depths below 330 km and that deep-focus earthquakes are limited to a belt running NNE to SSW in the area between latitudes 12°S and 15°S and longitudes 169°E and 174°E. On October 1, 1965, a shock with a focal depth of 546 km occurred to the south of this region at 20°S, 174°E. This is the deepest shock in this area for some ten years.

STUDY OF *P* AND *S* WAVES TRAVEL TIMES

Earthquakes recorded in Melanesia. *P* waves from the earthquakes of the southwest Pacific are clearly recorded in New Caledonia and the New Hebrides. Deep-focus earthquakes near Fiji and shocks in the Solomon Islands, New Britain, and New Guinea are common.

Histograms of travel-time residuals for deep Fijian earthquakes have been constructed for each station. This classical method of analysis has been used by many authors [*Tryggvason*, 1964; *Carder et al.*, 1966; *Cleary and Hales*, 1966]. The scatter of *P* residuals follows approximately a normal distribution, which is shown on each histogram (Figure 4). The width of each class is 0.5 sec. If x_{ij} is the residual associated with station i and source j, determined by using the *Jeffreys and Bullen* [1940] tables, the mean of residuals at station i is $\langle x_i \rangle = \Sigma x_{ij}/n$ where n is the number of shocks. The values shown by the histograms scattered about the smooth normal curve because of (1) errors in measured arrival time, caused by variations in paper speed, misidentifications of first phases, etc.; and (2) errors in origin time and epicenter location. Station means for Fiji earthquakes are listed in Table 2.

From an analysis of variance it is possible to derive results that do not appear from the histograms. Following *Carder et al.* [1966], we use an array x_{ij} and define $\langle x \rangle$ as the mean residual, $\langle x_i \rangle$ as the mean residual associated with station i, and $\langle x_j \rangle$ as the mean residual associated with earthquake j. Estimates of variance are then made for station means, for earthquake means, and for variance associated with random errors. The ratio of the variance associated with the earthquake or the station to the variance due to the random error is called the *F* ratio and is compared with the tabulated values of the *F* distribution. Application of the method to the Fiji earthquakes recorded at stations in New Caledonia and the New Hebrides reveals significant differences between stations and earthquakes ($F_1 = 13.8$ for sources and $F_2 = 66.5$ for stations). The significant differences between stations are attributed to differences in the structure beneath them. The elimination of one shock out of three gives a value of $F_1 = 2.3$, which is significant when compared with the tables. Local heterogeneity at the source may be the reason for this discrepancy in the third part of the population of shocks. It appears from this study that Melanesia is characterized by late arrival of *P* waves from Fiji earthquakes, and that local differences exist within the Melanesia area.

Explosion Longshot. Records of the 'Longshot' nuclear test at three Melanesian stations show late arrivals. The origin time is accurately known, and the U.S. Coast and Geodetic Survey (now the National Ocean Survey) lists residuals (observed and calculated) with respect to the Jeffreys-Bullen tables.

When the values of residuals at Noumea, Koumac, and Luganville are compared with the means of the residuals for European and American stations at the same epicentral distance and with Australian data, it is found that:

1. Arrivals of *P* waves at Melanesian stations are 3.5 sec later than arrivals at continental stations in Europe and America, for epicentral distances of about 70°.
2. Arrivals at Melanesian stations are 1 to 1.5 sec late with respect to those at Australian stations (Table 3).

Interpretation. The late arrivals in Melanesia may arise in three ways: at the source, as a function of azimuth; along the path in the mantle; or beneath the recording station. *Davies and McKenzie* [1969] suggest that the global distribution of residuals can be explained on the assumption that the dipping plate effectively 'pulls' the focus toward stations with large negative residuals.

This may explain part of the delay between Melanesia and Europe and Melanesia and America, but not the delay between Melanesia and Australia (1.0 to 1.5 sec), which probably arises from some local cause, such as a low-velocity upper mantle in the region to the north [*Dubois*, 1966]. We shall compare this result with the delay for Fiji earthquakes. Toward the west and northwest, the results are different. For deep-focus earthquakes in the Banda Sea, *Pascal* [1970] has found a delay of 0.7 sec under Australia for paths between Australian and Melanesian stations. The plate assumption may explain this particular anomaly [*Davies and McKenzie*, 1969].

The third possibility is to attribute the anomaly to the deep mantle [*Toksöz et al.*, 1967]. This would imply very marked lateral variations if the differences between the paths Longshot–West Australia and Longshot–Melanesia are to be explained. We shall discuss only the two first interpretations.

Study of the residuals shows that arrivals from the north, east, and southeast are usually late and probably arise from a low-velocity upper mantle under the Melanesian stations. Histograms and analyses of variance have shown the existence of local anomalies inside the area studied. The method of station pairs provides a more precise knowledge of these anomalies.

The difference between the residuals at two given stations $x = (O - C)_1 - (O - C)_2$ is also the difference between the theoretical and observed travel times. When x is positive, the arrival at station 1 is the later, and vice versa. The origin time does not appear in the expression, and errors due to uncertainty in the position of the source are very small. However, the value of x is affected by the measurement errors at stations 1 and 2.

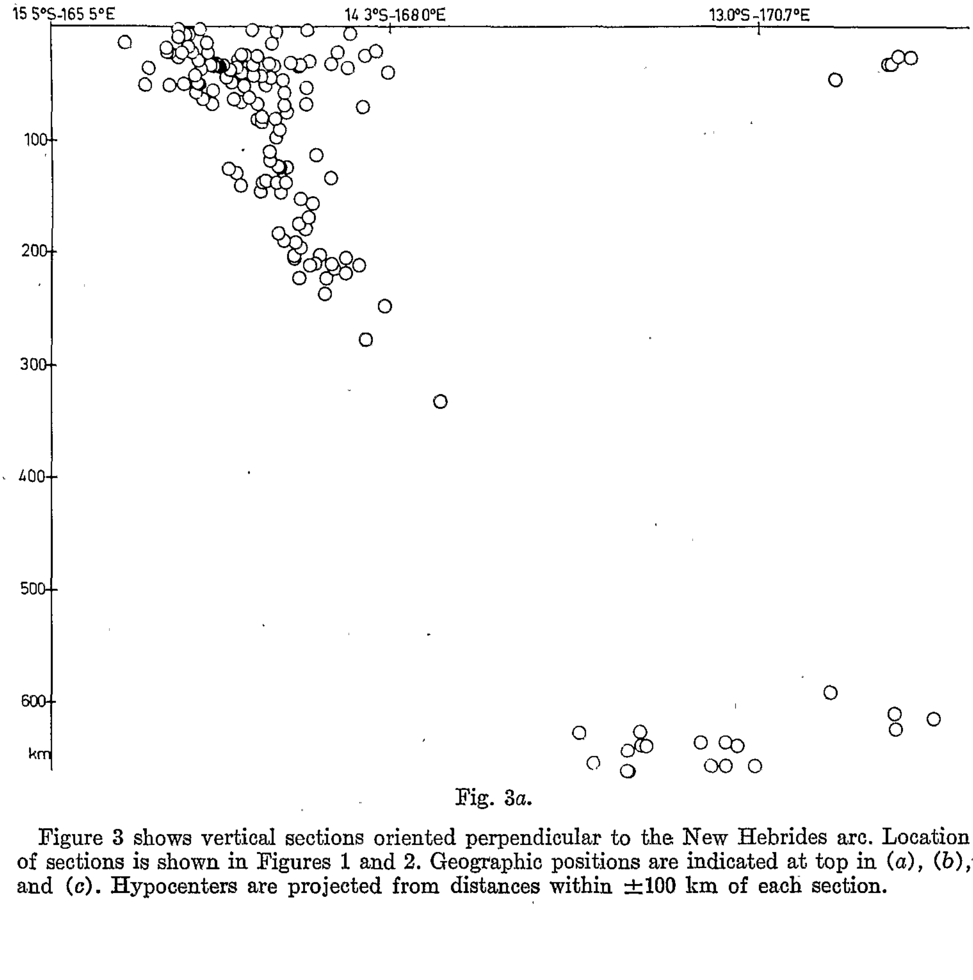

Fig. 3*a*.

Figure 3 shows vertical sections oriented perpendicular to the New Hebrides arc. Location of sections is shown in Figures 1 and 2. Geographic positions are indicated at top in (*a*), (*b*), and (*c*). Hypocenters are projected from distances within ±100 km of each section.

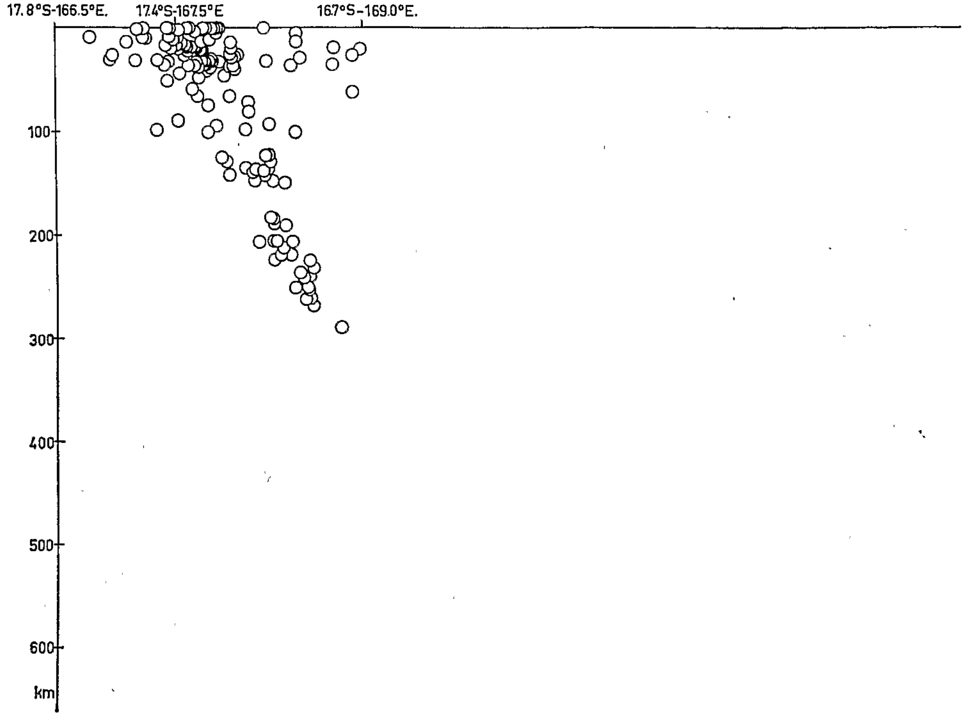
17.8°S-166.5°E.
17.4°S-167.5°E
16.7°S-169.0°E.
100
200
300
400
500
600
km

Fig. 3*b*.

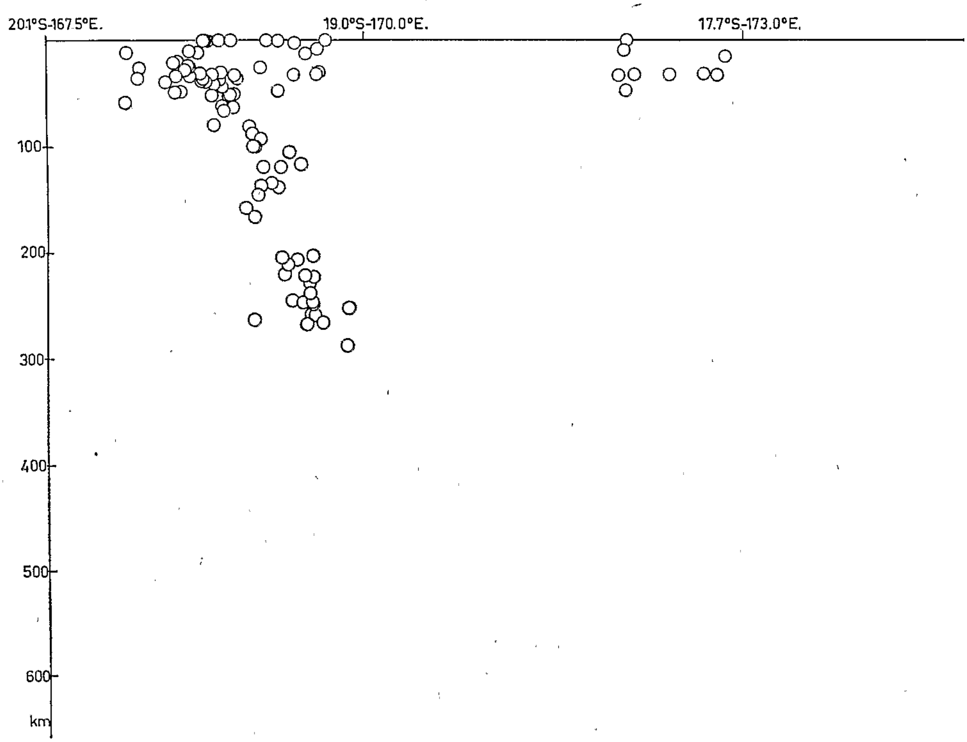
20.1°S-167.5°E.
19.0°S-170.0°E.
17.7°S-173.0°E.
100
200
300
400
500
600
km

Fig. 3*c*.

TABLE 1. Earthquake Depths Obtained by Two Methods

Date	Time	Coordinates	Standard Error, sec	Depth from P, km	Number of P Readings	Depth from $pP - P$, km
Mar. 24, 1965	07h 59m 39.1s	16°24 ± 0.028°S 167°95 ± 0.028°E	1.07	188 ± 4.2	90	194.2 ± 2.6
March 25, 1965	07h 17m 21.9s	14°32 ± 0.027°S 167°33 ± 0.029°E	0.93	209 ± 4.3	77	203.3 ± 1.1
July 15, 1965	08h 01m 17.7s	13°07 ± 0.063°S 166°87 ± 0.064°E	1.29	132 ± 8.9	25	122.8 ± 2.3
Sept. 1, 1965	06h 38m 37.0s	14°46 ± 0.021°S 167°26 ± 0.023°E	1.05	195 ± 3.6	137	181.6 ± 1.1
Sept. 7, 1965	08h 28m 52.0s	15°61 ± 0.033°S 167°08 ± 0.034°E	1.21	33 ± 4.3	68	37.6 ± 1.2
Oct. 1, 1965	13h 22m 28.4s	19°89 ± 0.023°S 174°46 ± 0.025°E	1.02	546 ± 3.7	182	541.5 ± 2.5
Feb. 4, 1966	10h 39m 11.5s	15°93 ± 0.023°S 167°91 ± 0.023°E	1.07	183 ± 3.2	202	180.6 ± 1.8
July 22, 1966	08h 25m 55.1s	15°99 ± 0.021°S 167°95 ± 0.024°E	0.99	190 ± 3.0	148	187.3 ± 2.4

The results for the New Hebrides and New Caledonia station pairs are plotted in Figure 4, together with histograms of x for Noumea–Koumac, Noumea–Port Vila, and Port Vila–Luganville for Fiji earthquakes. The means of the differences are listed in Table 4.

In the geographical area under consideration, lateness of arrival can be correlated with local topography. P waves are late when crossing a ridge. For example, for the paths Fiji–Noumea and Fiji–Koumac, the Koumac arrival is later than the arrival at Noumea. The delay probably occurs beneath the central chain, which is crossed only on the path Fiji–Koumac (Figure 5). The angle of incidence of the ray arriving at Koumac is 40°, and it enters the 35-km-thick crust 40 km from the station (see below). The difference of 0.7 sec could imply a difference of 15 km in the thickness of the crust under the southwest coast and under the central chain of New Caledonia, assuming a mean P-wave velocity of 6.5 km/sec.

Observations of New Hebrides earthquakes. Besides the sources in Melanesia that were considered above, many earthquakes occur in the New Hebrides seismic belt (see Figure 1). The records of these shocks obtained at the New Hebrides and New Caledonian seismic stations yield important information about propagation conditions in this area. Here also the method of station pairs reveals a correlation between late P arrivals and local topography; for example, Koumac arrivals are 0.79 ± 0.16 sec later than those at Noumea for paths Matthew–Koumac and Matthew–Noumea, which have the same azimuth as the paths Fiji–Koumac and Fiji–Noumea. No differences have been found between Espiritu Santo–Koumac and Espiritu Santo–Noumea.

In a further analysis, a least-squares method was used to compute the propagation equation for near earthquakes in the New Caledonia–New Hebrides area. A linear time–distance relationship was assumed to hold for epicentral distances between 300 and 800 km. Propagation equations of the form $T = \Delta/V + T_0$, where T_0 is the intercept on the T axis, are listed in Table 5. T is in seconds, Δ is in km and, V is in km/sec. For details of the computation see *Dubois* [1969].

The method was applied to the seismic paths New Hebrides–Noumea, New Hebrides–Koumac, New Hebrides–Ouanaham, New Hebrides–Port Vila, and New Hebrides–Luganville, using normal-depth earthquakes. The (T, Δ) relation was found to be linear in the area to the west of the New Hebrides but not along the seismic belt. Table 5 gives the propagation equation and the value of F: variance S_2^2, which corresponds to the dispersion to the regression line over S_{rep}^2 variance due to reproducibility error [see *Snedecor and Cochran*, 1967].

TABLE 2. Station Means of Travel-Time Residuals for Fiji Earthquakes

Stations	Station Mean, sec	Standard Deviation, sec	Number of Observations	Probable Error of Mean
Noumea	+1.18	1.35	110	±0.09
Koumac	+2.16	1.71	147	±0.10
Ouanaham	+0.33	1.33	34	±0.16
Port Vila	+1.43	1.53	114	±0.10
Luganville	+0.85	1.65	82	±0.10
Lamap	+0.67	1.52	54	±0.12
Lonorore	+0.56	1.45	45	±0.13

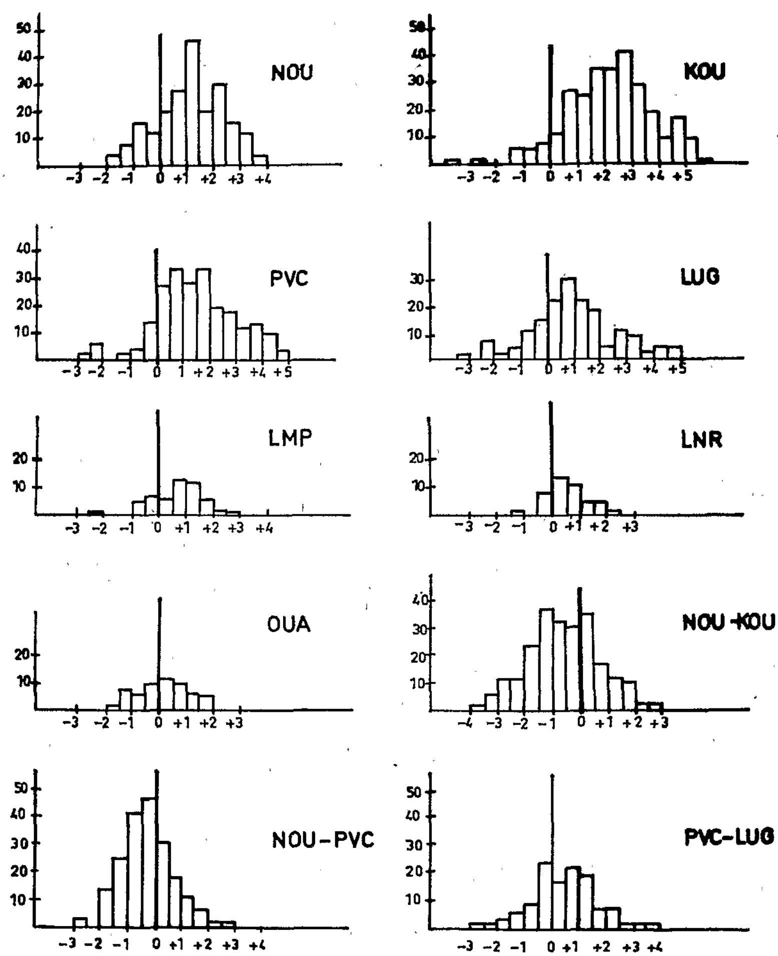

Fig. 4. Histograms of time residuals of P waves from Fiji deep-focus earthquakes at New Caledonia and New Hebrides stations: Noumea (NOU), Koumac (KOU), Ouanaham (OUA), Port Vila (PVC), Luganville (LUG), Lamap (LMP), and Lonorore (LNR). Interval of class is 0.5 sec. Histograms for stations pairs NOU-KOU, NOU-PVC, and PVC-LUG are shown.

TABLE 3. Travel-Time Residuals for Explosion Longshot at about 70° Epicentral Distance

Melanesia			Europe 4 stations		America 11 stations		Australia and New Guinea		
Stations	Epicentral Distance, D deg	$O-C$, sec	D°	$O-C$, sec	D°	$O-C$, sec	Stations	D°	$O-C$, sec
LUG	67.5	−1.9					PMG	66.6	−2.9
KOU	72.9	−1.3	70°	−5.1 ± 0.1	70°	−5.0 ± 0.2	DAR	76.1	−2.3
NOU	74.3	−1.2					CTA	76.8	−2.8

Stations: Luganville, Koumac, Noumea, Port Moresby, Darwin, and Charters Towers.

To the west of the New Hebrides, the propagation equation suggests a P-wave velocity of 7.9 km/sec in the upper mantle [*Bullen*, 1963]. Assuming a mean P-wave velocity of 6.5 km/sec in the crust, the crustal thickness under the central chain is 36 ± 4 km at the point where the New Hebrides–Noumea ray enters the crust, about 40 km to the northeast of Noumea. Taking this together with the result given by the station pairs, we may assume a crustal thickness of 20 km under the southwest coast of New Caledonia, and 36 km under the central chain.

Propagation along the New Hebrides belt. For propagation along the seismic arc, the equation is not linear. Many earthquake epicenters are collinear with Port Vila and Luganville, and it is therefore possible to compute the apparent velocity between the stations with good accuracy, since errors in the origin time and the position of the epicenter are eliminated in the computations. Table 6 shows the value of the apparent velocity with respect to epicentral distance for 13 normal-depth earthquakes (Figure 6).

The apparent velocity is low and increases from 7.2 km/sec to 7.9 km/sec (except in shocks 4 and 5). The Herglotz-Wiechert method applied to these results gives an upper-mantle structure in which the P-wave velocity increases linearly from 7.4 km/sec at a depth of 20 km, to 8.1 km/sec at a depth of 120 km.

The study of a local swarm of earthquakes in the Tonga Island area (near Luganville and Port Vila) suggests that here is a higher-velocity crust in the New Hebrides area. Thus the Hebridean anomalous area is characterized by a low P-wave velocity in the upper mantle, and a high velocity in the crust.

S waves. The S waves from normal-focus earthquakes in the New Hebrides are very well recorded at Noumea (Figure 7). As in the case of P waves, the method of least-squares can be used to obtain the propagation equation:

$$T = \Delta/(4.506 \pm 0.043) + 7.67 \pm 1.19 \qquad (n = 47)$$

This corresponds to a value of 0.256 ± 0.008

TABLE 4. Analysis of P-Wave Differences for Station Pairs

	Fiji Earthquakes				Northwest Earthquakes			
Station Pairs	Mean Differences, sec	Standard Deviation, sec	No. of Observations	Probable Error of Mean, sec	Means of Station Pairs, sec	Standard Deviation, sec	No. of Observations	Probable Error of Means, sec
Noumea-Koumac	−0.70	1.60	118	±0.10	+0.21	1.32	50	±0.13
Noumea-Port Vila	−0.31	0.91	105	±0.06	...	...	...	...
Port Vila-Luganville	+0.46	1.12	65	±0.11	−0.39	1.25	45	±0.14

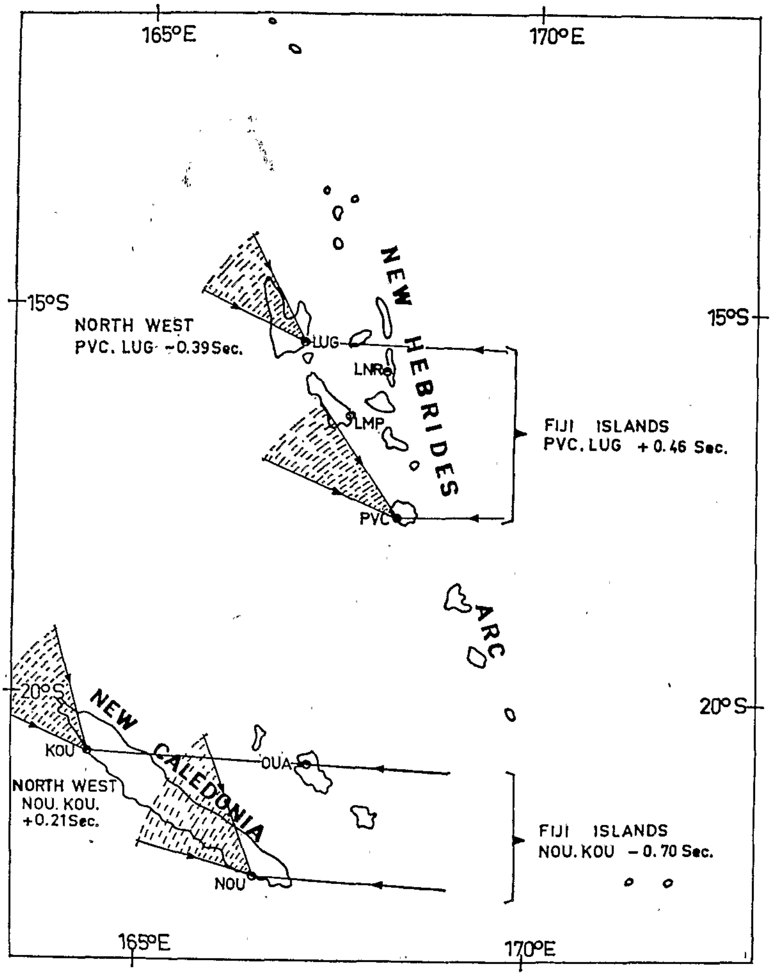

Fig. 5. Results for New Hebrides and New Caledonia stations pairs. The delay probably occurs beneath the central chain, which is crossed on the path Fiji–Koumac, and beneath Port Vila on the path Fiji–Port Vila.

for the Poisson ratio in the upper mantle in the area extending from the New Hebrides to New Caledonia. *Molnar and Oliver* [1969], in their study of lateral variation of attenuation in the upper mantle, observed poor propagation of S_n across the convex side of the New Hebrides Arc over two different paths from the northern New Hebrides to Charters Towers across the stable Coral Sea, but they were unable to explain why the S_n propagation was poor. On the other hand,

TABLE 5. P-Wave Propagation Equations: Test of Linearity

Stations	Propagation Equations	S_2^2	S_{rep}^2	F	Number of Earthquakes
Noumea	$T = \Delta/(7.878 \pm 0.022) + 3.18 \pm 0.27$	0.832	1.364	0.61	257
Koumac	$T = \Delta/(7.915 \pm 0.027) + 3.39 \pm 0.29$	···	···	···	265
Ouanaham	$T = \Delta/(7.890 \pm 0.045) + 3.00 \pm 0.37$	···	···	···	85
Port Vila	$T = \Delta/7.6 + 2.8$	5.78	1.357	4.26	76

TABLE 6. Normal-Focus Earthquakes on the Same Line as Luganville and Port Vila

Earthquakes	$\delta\Delta$, km	δT, sec	V_a, km/sec	Δ_m, km
Sept. 15, 1963	274.3	38.2	7.19	343
Sept. 29, 1963	274.3	38.0	7.22	387
Dec. 24, 1963	268.6	37.3	7.21	406
Dec. 30, 1963	273.0	35.5	7.80	421
Oct. 26, 1963	272.3	35.0	7.77	470
July 3, 1963	268.9	36.5	7.38	486
Dec. 3, 1963	273.4	37.0	7.38	542
Sept. 18, 1963	274.5	36.0	7.65	713
Sept. 15, 1963	271.9	34.4	7.90	735
Sept. 16, 1963	272.8	34.5	7.91	757
Sept. 17, 1963	274.1	35.0	7.85	773
Sept. 25, 1963	275.2	36.0	7.65	790
Sept. 25, 1963	275.6	35.5	7.76	799

$\delta\Delta$ is the difference in epicentral distance, δT is the difference in P-wave arrival, V_a is the apparent velocity, and Δ_m is the epicentral distance to a point midway between the two stations.

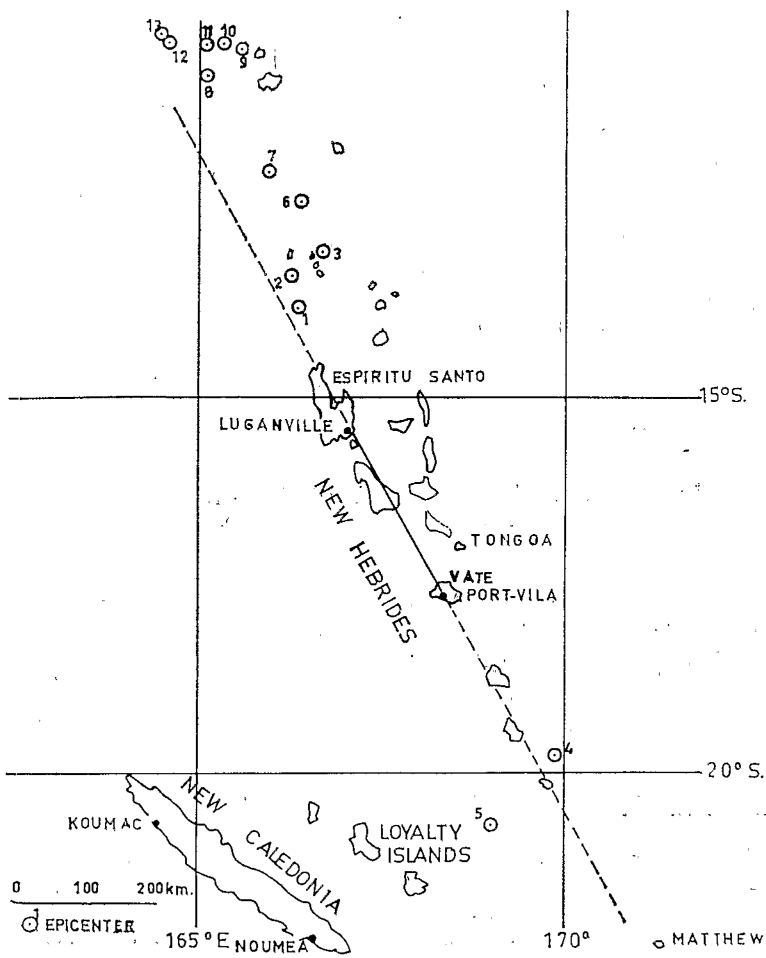

Fig. 6. Earthquake epicenters used to compute apparent velocity of P waves between Luganville and Port Vila with respect to epicentral distances.

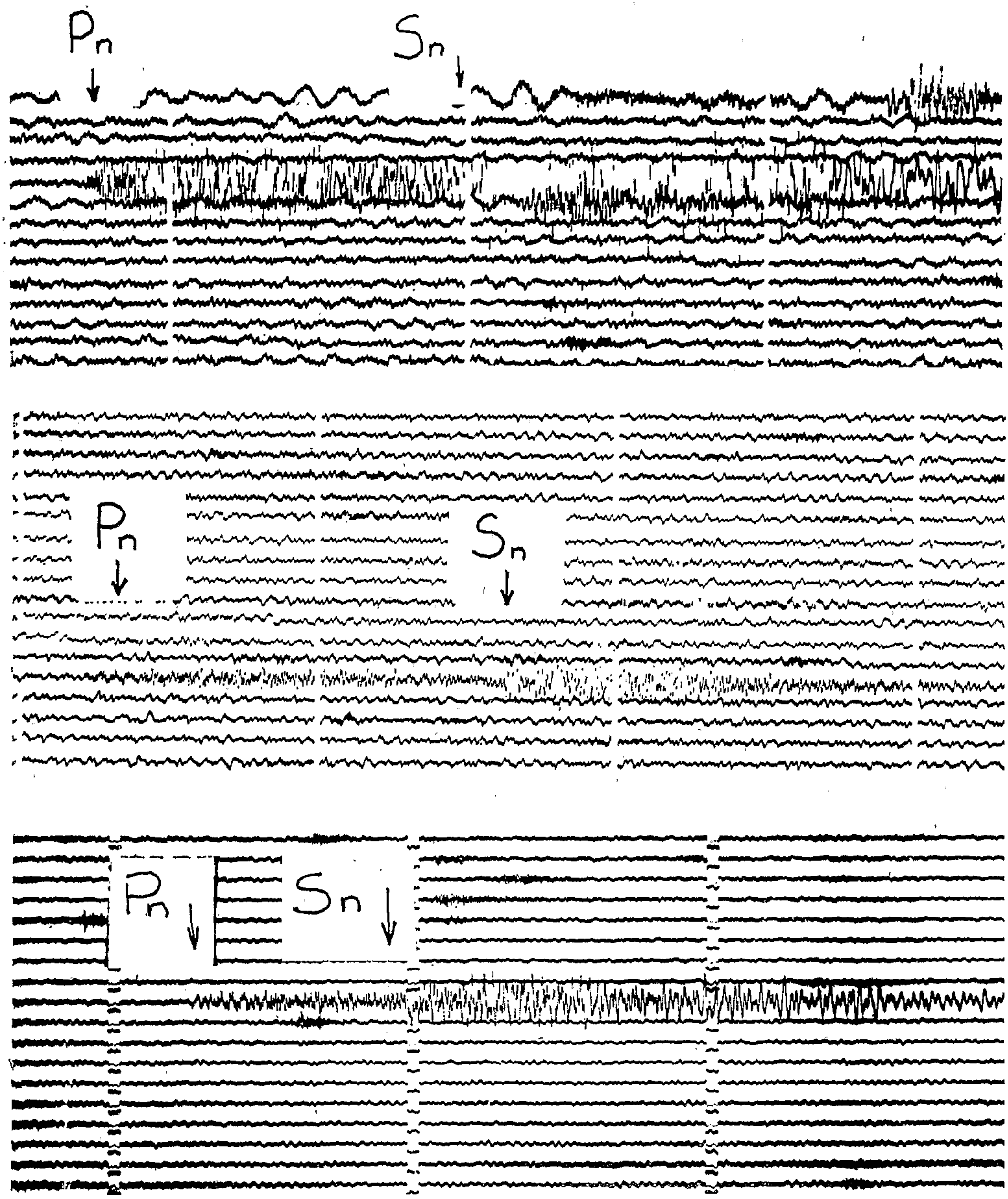

Fig. 7. Seismograms at Noumea and Port Vila. Normal depth-focus earthquakes in Espiritu Santo Island region. S_n is strong at Noumea and poor at Port Vila.

the transmission of *S* waves is good over the whole of the path from the New Hebrides to Noumea (Figure 7). This observation will be discussed below. Along the New Hebrides belt, propagation of *S* is poor to Port Vila and Luganville (from normal to 150 km depth focus).

Propagation of Rayleigh waves. Group velocities of Rayleigh waves were obtained for 100 southwest Pacific earthquakes recorded at Noumea, Koumac, and Port Vila. Appropriate allowances for the magnification curve of the instrument and for phase shift were made.

To fit the theoretical curves to the observations it is helpful to define 'influence coefficients' c such that a variation $\Delta\alpha$ in a parameter α produces a change ΔC in the phase velocity C: $\Delta C = Cc_\alpha \Delta\alpha / 2\alpha$. The subscript α refers to P-wave velocity, β to shear-wave velocity, and ρ to density. Calculations were done on the CDC 3600 computer; a program titled VPROCJ, written by N. Jobert, for various layers derived from the study of body waves was used. In Figure 8 the ratios $\Delta C/\Delta\alpha$, $\Delta C/\Delta\beta$, and $\Delta C/\Delta\rho$ are plotted against T, the period of the ground motion for the fundamental Rayleigh mode and the first shear mode. The influence on the group velocity U was computed from C variations by using the Newton-Gregory interpolation method [*Jeffreys and Jeffreys*, 1950]. A computation for the effect of layer thickness was done, mainly for oceanic and sedimentary layers [*Dubois*, 1969].

Experimental results. *Kuo et al.* [1962], *Hunkins and Kuo* [1965], and *Saito and Takeuchi* [1966] have studied the area between Australia and the Andesite-line. The Noumea network lies in the center of this area and can be used to establish structural subdivisions within

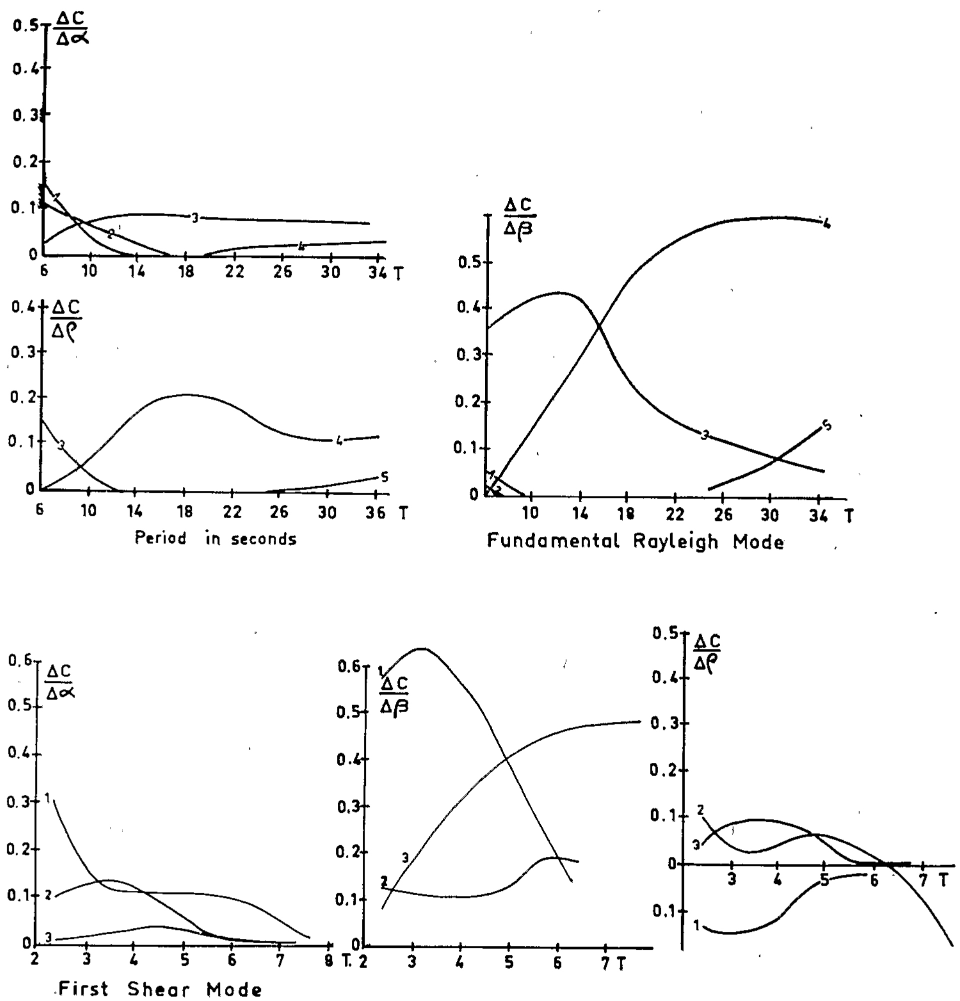

Fig. 8. Influence of the parameters α (P-wave velocity), β (shear-wave velocity), and ρ (density) on C Rayleigh-wave phase velocity against T (the period of the ground motion) in sedimentary layers 1 and 2, layer 3 in the crust, and layers 4 and 5 in the upper mantle (Models E and T, after *Dubois* [1969]).

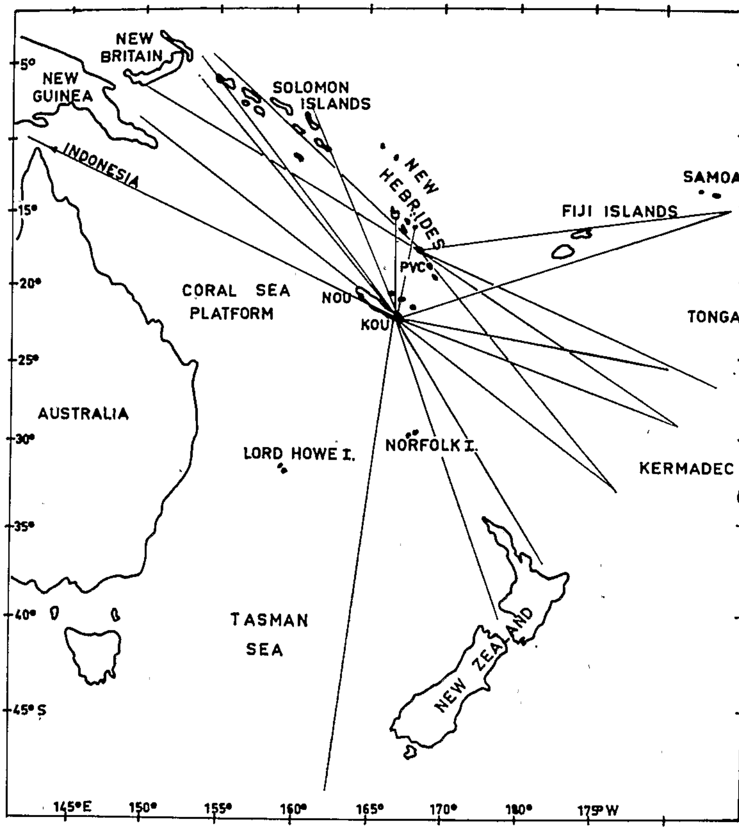

Fig. 9. Seismic paths on which the group velocity of Rayleigh waves has been computed.

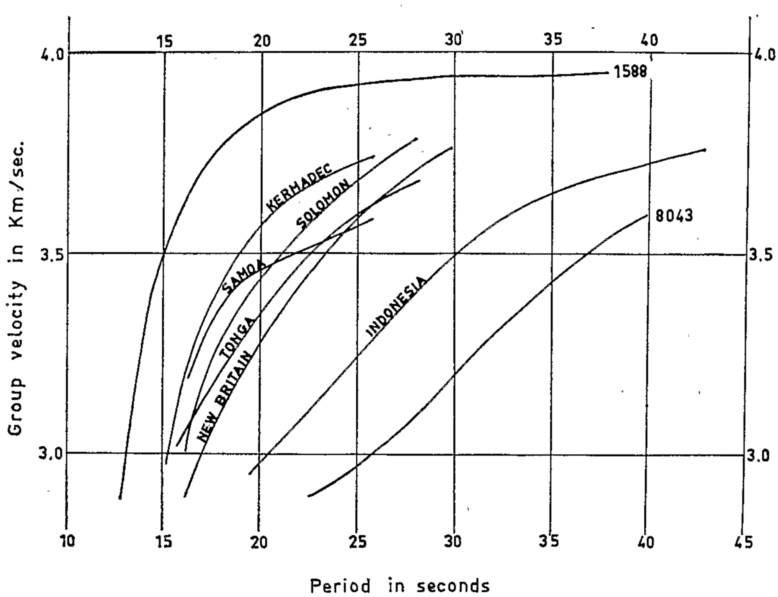

Fig. 10. Group velocity against period of the ground motion observed at Noumea for different seismic paths. Comparison with continental 8043 and oceanic 1588 theoretical models [*Dorman et al.*, 1960].

it (Figure 9). Figure 10 shows the group velocities of the Rayleigh waves characteristic of those areas. Comparison with theoretical dispersion curves [e.g. *Saito and Takeuchi,* 1966] suggests the structure given below. In a more detailed paper, the group velocity over the different paths (Figure 9) and comparisons with possible models are made [*Dubois,* 1968].

In the oceanic area surrounding New Caledonia, the crust is 20 km thick and thus is thicker than the classical oceanic crust. In the northwestern part of the Coral Sea, a thickness of 25 km is indicated. In the Fiji–Tonga–Samoa area, the indicated thickness is 22–28 km, and in the North and South Fiji basins and the Tasman Sea, it is 15 km or less. A thickness of 25 km is found along the Norfolk and Lau ridges. *P*- and S-wave velocities in the upper mantle beneath the Coral Sea and the New Caledonia–New Hebrides area are low.

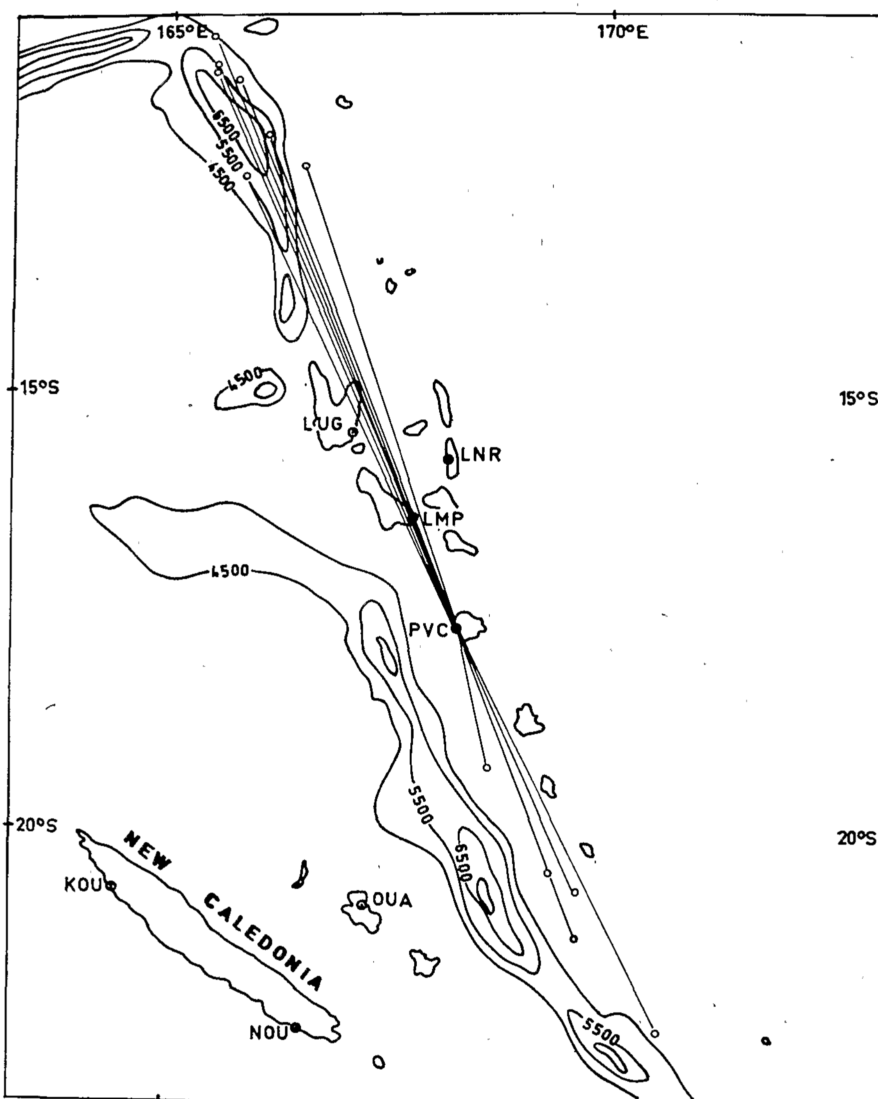

Fig. 11. Seismic paths along the New Hebrides arc.

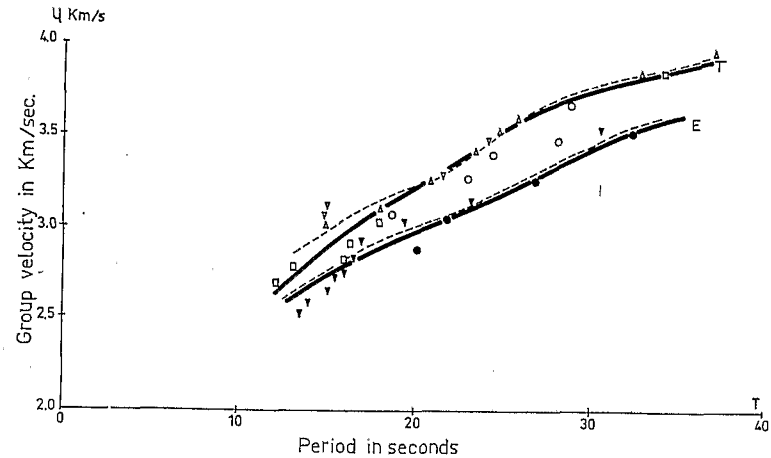

Fig. 12. Group velocity of Rayleigh waves, fundamental mode. Open symbols represent data from trench epicenters, black symbols represent data from eastern slope epicenters (dashed lines). Theoretical dispersion curves of models T and E are indicated by solid lines.

New Hebrides and New Caledonia area. The northern part of the New Hebrides anomalous belt has been studied using records from seven Santa Cruz Island earthquakes recorded at Port Vila (Figure 11). Rayleigh-wave dispersion curves show that little difference exists between a group of four earthquakes under the trench and a group of three earthquakes on the eastern slope (see Figure 12 and Table 7).

By using the method of fitting previously discussed, we constructed theoretical models E and T corresponding to two groups of observed data (Table 8). The Rayleigh-wave models are similar to the body-wave models proposed in the earlier part of this paper. In fact, the linear gradient of *P* and *S* waves in the upper mantle agrees with the dispersion curves of the fundamental and first shear modes (Figure 13) for both the north and the south groups of earthquakes. For the first shear mode, the sloping interface correction [*Tryggvason*, 1964] was applied when the bathymetric profile was rough, because first shear mode propagation is influenced mainly (Figure 13) by variations in the thickness of the oceanic layer.

The separation of the seismological stations in New Caledonia and the New Hebrides is too great to permit a complete study of the phase velocity. However, in some circumstances a phase-velocity record may be used. In one case, when an epicenter was located on the line joining the stations at Noumea and Koumac (a distance of about 300 km), the same train of Rayleigh waves was easily identified on the two seismograms (Figure 14). A comparison between the observed and the computed dispersion curves (*Haskell's* [1953] flat-layers method) is made in Figure 15 and Table 9.

Experimental data fall between the theoretical models I_1 and I_2 and G_2 and G_3. The crustal thickness under the west coast of New Caledonia ranges from 22 to 30 km, according to the adopted values of α, β, and ρ.

General Discussion and Geological Implications

The data available for an interpretation of the structure of the New Hebrides arc include epicenter and focal-depth locations, measurements of the velocity of seismic waves, geological surveys of the Melanesian region, and scattered geophysical observations at sea. According to the diagram by *Oliver and Isacks*, [1967], the lithosphere is dipping under the northern Fijian basin, with the earthquake foci lying on a surface that dips at about 55° to the east-northeast, until it is interrupted at a depth of 350 km (Figures 3 and 16). The observed seismic-wave velocities do not conflict with this picture, and they suggest some additional points in its favor. In general, *P* waves from the north and east (Fiji and Longshot) arrive late,

TABLE 7. Rayleigh-Wave Dispersion Data

Date	Origin Time, h	m	s	South Latitude	East Longitude	Average of Water Layer Thickness, meters		Dispersion data	Observations
Jan. 1, 1967	04	04	06.5	11°98	166°01	1430	Fundamental	*T* 32.2 26.7 21.7 20.0 *U* 3.54 3.27 3.06 2.90	
							1st shear	*T* 9.6 7.2 6.0 5.7 4.6 *U* 3.93 3.78 3.28 2.98 2.06 *T/H* 6.70 5.05 4.19 3.98 3.22	East slope, not filtered.
Jan. 1, 1967	14	18	51.4	12°39	165°78	2200	Fundamental	*T* 20.8 19.5 16.5 13.0 12.0 *U* 3.16 3.02 2.89 2.80 2.72	
							1st shear	*T* 7.8 9.4 9.0 6.2 5.4 4.7 3.0 *U* 3.72 3.66 3.57 3.48 3.24 3.16 3.06 *T/H* 3.52 4.28 4.10 2.82 2.46 2.14 1.35	Under the trench, not filtered.
Jan. 1, 1967	21	58	57.8	11°13	165°46	2350	Fundamental	T 37.0 32.8 25.8 24.6 23.4 20.6 18.0 15.0 13.6 *U* 4.00 3.78 3.64 3.56 3.45 3.29 3.13 3.05 2.98	
							1st shear	*T* 4.2 3.5 3.0 2.9 *U* 3.18 3.75 3.00 2.52 *T/H* 1.79 1.49 1.28 1.23	Under the trench, not filtered.
Jan. 2, 1967	19	59	58.2	12°33	166°44	1415	Fundamental	*T* 28.4 28.0 24.2 22.9 18.6 15.0 *U* 3.69 3.50 3.42 3.28 3.09 3.02	
							1st shear	*T* 7.8 6.0 3.8 3.0 2.9 2.9 *U* 3.58 3.43 3.32 3.00 2.71 2.58 *T/H* 5.50 4.24 2.69 2.12 2.05 2.05	East slope, not filtered.
Jan. 3, 1967	11	05	15.0	11°18	165°44	2350	Fundamental	*T* 34.0 30.1 18.0 16.3 14.9 *U* 3.78 3.56 3.04 2.93 2.84	
							1st shear	*T* 6.0 4.8 *U* 3.28 3.05 *T/H* 2.55 2.04	Under the trench, not filtered.
Jan. 3, 1967	12	32	09.2	10°85	165°39	2660	Fundamental	*T* 24.0 21.5 18.8 15.3 15.0 15.0 *U* 3.50 3.30 3.24 3.13 3.05 2.96	Under the trench, not filtered.
Jan. 16, 1967	04	44	27.3	11°30	165°68	1760	Fundamental	*T* 30.5 23.1 19.2 17.0 16.5 16.0 15.6 15.2 14.0 13.5 *U* 3.56 3.16 3.04 2.94 2.85 2.76 2.74 2.67 2.60 2.54	

TABLE 7. *(continued)*

Date	Origin Time, h m s	South Latitude	East Longitude	Average of Water Layer Thickness, meters		Dispersion data	Observations
					1st shear	T 8.0 7.0 6.3 5.8 5.5 5.2 4.9 4.7 4.5 4.3 U 3.98 3.84 3.70 3.60 3.52 3.44 3.32 3.26 3.20 3.14 T/H 4.55 3.98 3.58 3.30 3.12 2.95 2.78 2.67 2.56 2.44	East slope, filtered.
Jan. 16, 1967	10 31 37.8	20°60	169°62	344	1st shear	T 5.2 4.2 3.9 3.8 3.3 1.4 U 3.09 3.01 2.98 2.92 2.83 1.90 T/H 4.97 4.02 3.74 3.64 3.16 1.34	
March 20, 1967	19 07 52.2	22°14	170°55	1191	1st shear	T 5.0 3.9 3.1 2.7 2.0 1.8 1.4 U 3.07 2.71 2.33 2.10 1.95 1.67 1.53 T/H 4.19 3.27 2.60 2.26 1.68 1.51 1.17	
Apr. 5, 1967	06 55 29.0	19°19	168°64	763	1st shear	T 2.2 2.0 1.8 1.6 1.1 1.1 U 2.63 2.39 2.31 1.53 1.34 1.33 T/H 2.88 2.62 2.36 2.10 1.44 1.44	
June 13, 1967	03 11 59.0	21°23	169°64	542	1st shear	T 2.8 2.5 2.1 1.9 1.6 1.5 U 3.31 3.11 2.80 2.53 2.33 2.18 T/H 5.17 4.62 3.88 3.51 2.96 2.78	
July 13, 1967	10 04 19.0	20°40	169°27	880	1st shear	T 5.8 3.1 2.1 2.1 1.6 1.6 1.2 U 3.44 2.91 2.17 1.55 1.41 1.12 1.08 T/H 6.59 3.52 2.38 2.38 1.82 1.82 1.36	
July 21, 1967	19 28 08.0	19°17	168°60	801	1st shear	T 4.6 3.0 2.3 2.1 1.7 1.2 U 3.47 2.84 2.25 1.86 1.63 1.54 T/H 5.74 3.74 2.87 2.62 2.12 1.50	

TABLE 8. Theoretical Models E and T

Layer	Depth, km	H, km	α, km/sec	β, km/sec	ρ, g/cm³
		Model E			
Water		1.5	1.50	0	1.00
	1.5				
Sediments		3.5	2.50	1.18	2.10
	4.5				
'Oceanic' layer		15.0	6.70	3.80	3.00
	19.5				
'Oceanic' layer		15.0	7.00	4.05	3.00
	34.5				
			7.60	4.45	3.20
Upper mantle		70.5			
	105.0		8.30	4.70	3.34
		Model T			
Water		2.5	1.50	0	1.00
	2.5				
Sediments		2.0	2.50	1.18	2.10
	4.5				
'Oceanic' layer		17.5	6.70	3.80	3.00
	22.0				
			7.60	4.45	3.20
Upper mantle		83.0			
	105.0		8.30	4.70	3.34

possibly as a result of their passage through the region that lies above the seismic surface. To the northwest, these late arrivals are no longer observed. This is in agreement with the plate hypothesis of *Davies and McKenzie* [1969]. The travel-times of P waves propagated along the arc are consistent with the existence of a velocity gradient at depths between 20 and 120 km. To explain the late arrivals, it is necessary to assume that this gradient persists to a depth of 300 km. Observations of Rayleigh waves are consistent with this model.

The width of the abnormal belt, on the evidence of the similarity of P-wave patterns recorded at Lonorore and Lamap, is at least 100–150 km. Within the belt, the boundary between crust and mantle is poorly defined, and S waves are strongly attenuated at depths between 100 and 150 km [*Molnar and Oliver,* 1969]. These observations refer to paths between the New Hebrides and Solomon Islands (Figure 9).

Where does the expansion arise? *Summerhayes* [1967] and *Cullen* [1967], suggest that it occurs on the Indian-Antarctic ridge, since the North Fiji basin is a large stable plate lying between two mobile lithospheric plates that converge and thrust beneath the Tonga–Kermadec and New Hebrides arcs. The secondary expan-

TABLE 9. Theoretical Models I, H, and G

Models	First Layer (Sediments)*	Second Layer: α, km/sec	β, km/sec	ρ, g/cm³	Thickness, km: 1	2	3	4	Mantle†
I	α = 4.00 km/sec	6.20	3.58	2.80	19	22	28		α = 7.95 km/sec
H	β = 2.30 km/sec	6.50	3.76	3.00	19	22	28	33	β = 4.60 km/sec
G	ρ = 3.00 g/cm³	6.00	3.40	2.50	15	20	22		ρ = 3.34 g/cm³

* Thickness, 2 km.
† Thickness, ∞.

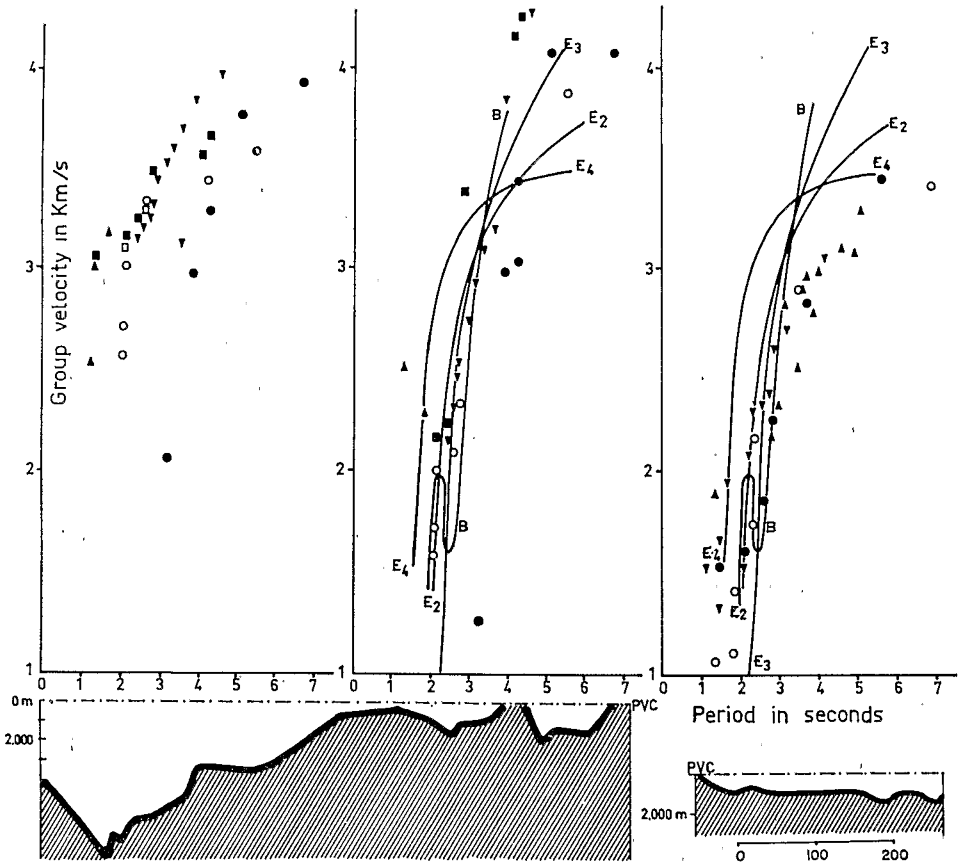

Fig. 13. Group velocity of first shear mode observed at Port Vila. *Left and center,* Santa Cruz earthquakes before and after sloping interface correction [*Tryggvason,* 1964]; *right,* southern New Hebrides earthquakes. Theoretical dispersion curves (solid lines) correspond to model E, Table 8, in which the thickness of the 'oceanic' layer varies from 15 to 25 km [*Dubois,* 1969]. Scale at lower right is in kilometers.

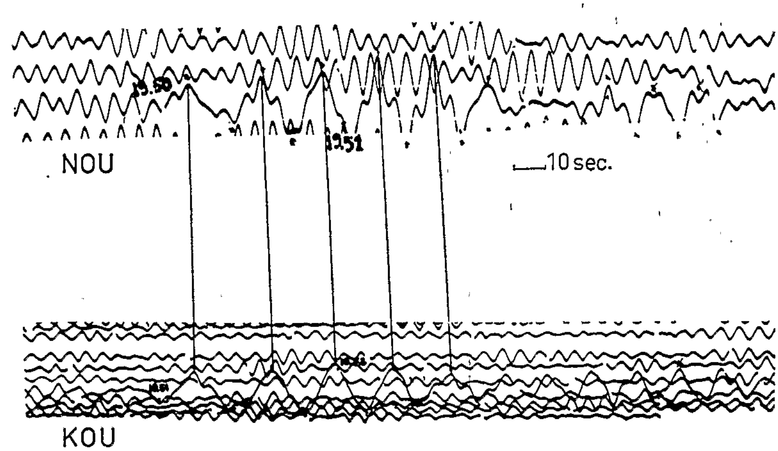

Fig. 14. Seismograms at Noumea and Koumac represent correspondence of phases of Rayleigh waves from an earthquake in Tonga (December 11, 1967; 21°6S, 174°3W; 33-km depth).

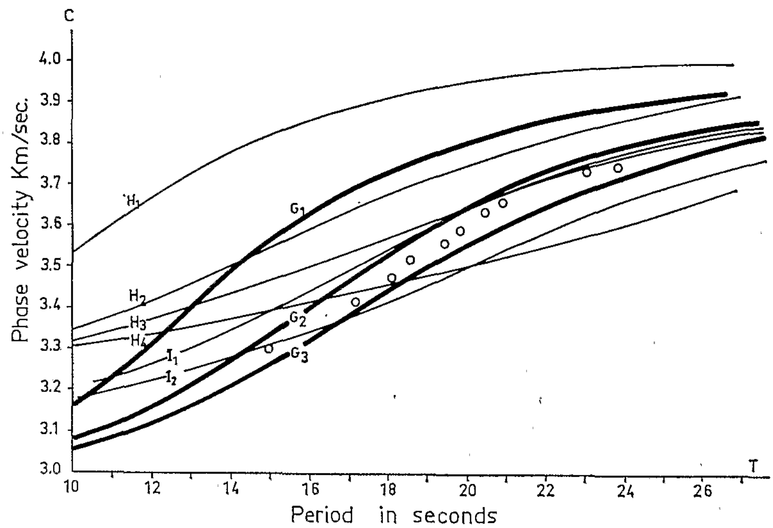

Fig. 15. Comparison between the observed and the computed dispersion curves, models I, H, and G.

sion centers on the concave side of the arc, suggested by *Packham and Falvey* [1971] and *Karig* [1970], could also be invoked. In the southern New Hebrides there is an inter-arcuate basin of this kind, east of the active line of volcanoes, and known as the Erromango rift [*Puech and Reichenfeld,* 1969].

If the patterns of P-wave arrival times in the New Hebrides and New Caledonia are considered together with the geology, it becomes possible to regard the New Caledonia–Loyalty Islands system as an island arc whose activity ceased in the Oligocene. The presence of a fossil lithosphere and a wedge of upper mantle material with a comparatively low P velocity would explain why the arrivals from the north and east are late but do not explain the early arrivals from the northwest, which traverse lithospheric paths. The interpretation is complicated by the relative differences that result from the roots of the central mountain chain. The explanation appears to be related to the

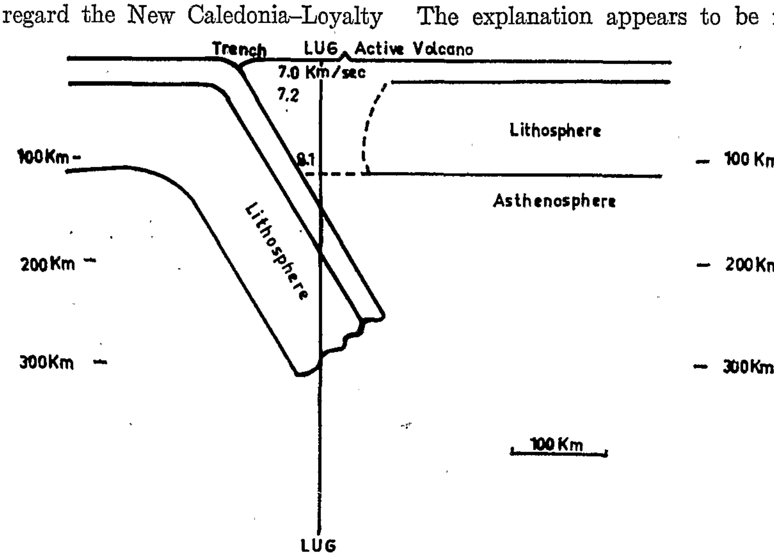

Fig. 16. Vertical section oriented perpendicular to New Hebrides arc, according to hypothesis of *Oliver and Isacks* [1967]. Interruption of the lithosphere corresponds to the lack of hypocenters at a depth greater than 300 km.

topography, but this can only explain half of the 2-sec delay.

The similarity of the anomalies in propagation is paralleled by a similarity in structure. On this view, the southwest New Caledonia basin becomes a trench partly filled with sediment [*Houtz et al.*, 1967], the Oligocene volcanoes being represented by the alignment of the Loyalty Islands. Lava beds at the Mare Island have an age of 29 ± 4 m.y. [*Chevalier*, 1968]. The peridotites would have been deposited between the paroxysmal phases, during the Oligocene. A gravity anomally of 170 mgal suggests that the peridotites form a sheet about 8 km thick in the eastern part, overlapping areas of sediment and Eocene basalt to the west [*Guillon*, 1971]. After this, from the Miocene onward, underthrusting in its present position in the New Hebrides would have taken place [*Ewing and Ewing*, 1967], whether the expansion stopped or not.

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REFERENCES

Adams, R. O., Source characteristics of some deep New Zealand earthquakes, *N.Z. J. Geol. Geophys.*, *6*, 209–220, 1963.

Bolt, B. A., and M. Niazi, Dispersion of Rayleigh waves across Australia, *Geophys. J.*, *9*, 21–35, 1964.

Brooks, J. A., Rayleigh waves in Southern New Guinea, 1, Higher mode group velocities, *Bull. Seismol. Soc. Amer.*, *59*, 945–958, 1969.

Bullen, K. E., *An Introduction to the Theory of Seismology*, 3rd ed., p. 381, Cambridge University Press, London, 1963.

Carder, D. S., D. W. Gordon, and J. N. Jordan, Analysis of surface foci travel time, *Bull. Seismol. Soc. Amer.*, *56*, 815–840, 1966.

Chevalier, J. P., Expédition Française sur les récifs coralliens de la Nouvelle Calédonie, vol. 3, Editions de la Fondation Singer Polignac, Paris, 1968.

Cleary, J., and A. L. Hales, Analysis of travel times of *P* waves, *Bull. Seismol. Soc. Amer.*, *56*, 467–489, 1966.

Cullen, D. J., Mantle convection and sea floor spreading in the South West Pacific, *Nature*, *216*, 356–357, 1967.

Davies, D., and D. P. McKenzie, Seismic travel-time residuals and plates, *Geophys. J.*, *18*, 51–63, 1969.

Denham, D., Distribution of earthquakes in the New Guinea Solomon Islands region, *J. Geophys. Res.*, *74*, 4290–4299, 1969.

Dorman, J., M. Ewing, and J. Oliver, Study of shear velocity distribution in the upper mantle by mantle Rayleigh waves, *Bull. Seismol. Soc. Amer.*, *50*, 87–115, 1960.

Dubois, J., Sur la vitesse de propagation des ondes *P* le long de l'arc seismique des Nouvelles Hebrides, *C. R. H. Acad. Sci.*, *260*, 2275–2277, 1965.

Dubois, J., Temps de propagation des ondes *P* à des distances épicentrales de 30 à 90 degrés, région du Sud-Ouest Pacifique, *Ann. Geophys.*, *22*, 642–645, 1966.

Dubois, J., Étude de la dispersion des ondes de Rayleigh dans la région du Sud Ouest Pacifique, *Ann. Geophys.*, *24*, 359–372, 1968.

Dubois, J., Contribution à l'étude structurale du Sud-Ouest Pacifique d'après les ondes sismiques observées en Nouvelle Calédonie et aux Nouvelles Hébrides, *Ann. Geophys.*, *25*, 923–972, 1969.

Eiby, G. A., The New Zealand subcrustal rift, *N.Z. J. Geol. Geophys.*, *7*, 109–133, 1964.

Ewing, J., and M. Ewing, Sediment distribution and the mid-ocean ridges with respect to spreading of the floor, *Science*, *156*, 1590, 1967.

Guillon, J. H., Contribution to the structural study of the New Caledonia—Norfolk Island arc (abstract), *Proc. 12th Pacific-Sci. Congr.*, in press, 1971.

Gutenber, B., and C. F. Richter, *Seismicity of the Earth and Associated Phenomena*, p. 310, Hafner, New York, 1965.

Haskell, N. A., The dispersion of surface waves in multilayered media, *Bull. Seismol. Soc. Amer.*, *43*, 17–34, 1953.

Houtz, H., J. Ewing, M. Ewing, and A. G. Lonardi, Seismic reflection profiles of the New Zealand Plateau, *J. Geophys. Res.*, *72*, 4713–4729, 1967.

Hunkins, K., and J. Kuo, Surface wave dispersion in the Tonga-Fiji Region, *Bull. Seismol. Soc. Amer.*, *55*, 135–145, 1965.

Isacks, B., J. Oliver, and L. R. Sykes, Seismology and the new global tectonics, *J. Geophys. Res.*, *73*, 5855–5899, 1968.

Jeffreys, H., and K. E. Bullen Seismological tables, Brit. Ass., Gray-Milne Trust, 50 pp., 1940.

Jeffreys, M., and B. S. Jeffreys, *Methods of Mathematical Physics*, 2nd ed., p. 708, Cambridge University Press, London, 1950.

Karig, D. E., Ridges and basins of the Tonga-Kermadec Island arc system, *J. Geophys. Res.*, *75*, 239–254, 1970.

Kuo, J., J. Brune, and M. Major, Rayleigh wave dispersion in the Pacific Ocean for the period range 20 to 140 seconds, *Bull. Seismol. Soc. Amer.*, *52*, 333–357, 1962.

Mitronovas, W., B. Isacks, and L. Seeber, Earthquake locations and seismic wave propagation in the upper 250 km of the Tonga Island arc, *Bull. Seismol. Soc. Amer., 59,* 1115–1135, 1969.

Molnar, P., and J. Oliver, Lateral variations of attenuation in the upper mantle and discontinuities in the lithosphere, *J. Geophys. Res., 74,* 2648–2682, 1969.

Oliver, J., and B. Isacks, Deep earthquakes zones, anomalous structures in the upper mantle, and the lithosphere, *J. Geophys. Res., 72,* 4259–4275, 1967.

Packham, G. H., and D. A. Falvey, A hypothesis for the formation of marginal seas in the Western Pacific, *Tectonophysics, 11,* 79–109, 1971.

Pascal, G., Etude des O-C des ondes *P* de 30° à 100°, Relation avec l'hypothèse des plaques, *Rapp. Orstom Centre Noumea,* New Caledonia, 1–13, 1970.

Puech, J. L., and C. Reichenfeld, Études bathymétriques dans la région des îles Erromango, Tanna et Anatom (Nouvelles Hébrides), *C. R. H. Acad. Sci., 268,* 1259–1261, 1969.

Saito, M., and H. Takeuchi, Surface waves across the Pacific, *Bull. Seismol. Soc. Amer., 56,* 1067–1091, 1966.

Santo, T. A., Division of the southwestern Pacific area into several regions in each of which Rayleigh waves have the same dispersion characteristics, *Bull. Earthquake Res. Inst. Tokyo Univ., 39,* 603–630, 1961*a*.

Santo, T. A., Dispersion of Love waves along various paths to Japan, 1, *Bull. Earthquake Res. Inst. Tokyo Univ., 39,* 631–651, 1961*b*.

Santo, T. A., Division of the Pacific area into seven regions in each of which Rayleigh waves have the same group velocities, *Bull. Earthquake Res. Inst. Tokyo Univ., 41,* 719–741, 1963.

Shor, G. G., Seismic refraction profile in Coral Sea basin, *Science, 158,* 911–913, 1967.

Snedecor, G. W., and W. G. Cochran, *Statistical Methods,* p. 593, Iowa State University Press, Ames, 1967.

Summerhayes, C. P., New Zealand region volcanism and structure, *Nature, 215,* 610–611, 1967.

Sykes, L. R., Deep-focus earthquakes in the New Hebrides region, *J. Geophys. Res., 69,* 5353–5355, 1964.

Sykes, L. R., The seismicity and deep structure of island arcs, *J. Geophys. Res., 71,* 2981–3006, 1966.

Sykes, L. R., B. L. Isacks, and J. Oliver, Spatial distribution of deep and shallow earthquakes of small magnitudes in the Fiji-Tonga region, *Bull. Seismol. Soc. Amer., 59,* 1093–1113, 1969.

Thomson, A. A., and F. F. Evison, Thickness of the earth's crust in New Zealand, *N.Z. J. Geol. Geophys., 5,* 29–45, 1962.

Toksöz, M. N., M. A. Chinnery, and D. L. Anderson, Inhomogeneities in the earth's mantle, *Geophys. J., 13,* 31–59, 1967.

Tryggvason, E., Arrival times of *P* waves and upper mantle structure, *Bull. Seismol. Soc. Amer., 54,* 727–736, 1964.

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