

P-WAVE VELOCITY OF THE UPPERMOST MANTLE AND CRUSTAL THICKNESS IN THE CENTRAL VANUATU ISLANDS (NEW HEBRIDES ISLAND ARC)

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ABSTRACT

The *P*-wave velocity of the uppermost mantle and crustal thickness were determined for the central Vanuatu (formerly New Hebrides) Islands using data from a local seismograph network. Two different methods and data sets were used in this investigation. The first method used travel-time plots for earthquakes located well outside of the seismograph network. The second method used time-term analysis on sets of earthquakes. Both methods yielded a mantle velocity of $7.6 (\pm 0.1)$ km/sec and a crustal thickness of $27 (\pm 2)$ km for the region located just trenchward of the volcanic arc. Published refraction studies have demonstrated that the mantle velocity near the trench has a more typical value of 8.0 km/sec. This contrast in mantle velocities between the trench and the volcanic arc has also been reported for the Japanese and Kurile-Kamchatka Island arcs and, therefore, may be a general feature of subduction zones.

INTRODUCTION

The Vanuatu (formerly New Hebrides) Islands are located in the Southwest Pacific where the Australian plate is being subducted eastward beneath the Fiji plateau at the New Hebrides trench (Figure 1). The Quaternary volcanic arc is formed by the islands of Anatom, Tanna, Erromango, Ngouna, Emai, Epi, Ambrym, Aoba, and the Banks Islands. The central part of the island arc has a unique structure compared to most other island arcs because the islands of Santo, Malekula, and parts of Efate protrude west of the volcanic arc and occupy the position normally occupied by the accretionary prism complex. An aseismic ridge, the D'Entrecasteaux Fracture Zone, is being subducted along this portion of the arc. There have been few investigations of the crustal and mantle structure of this region.

Dubois (1971) and Dubois *et al.* (1973) determined the apparent velocity of the mantle refracted waves (*P_n*) recorded at stations on the islands of Santo (LUG) and Efate (PVC) from events located north and south of the stations. They deduced a model in which the velocity increases from a low of 7.4 km/sec at 20-km depth to 8.1 km/sec at 120-km depth. Another low mantle velocity of 7.4 km/sec at depths from 20 to 35 km under the island of Malekula was deduced from the inversion of *P*-arrival times at local stations (Grasso, 1981). Refraction profiles (Ibrahim *et al.*, 1980; Pontoise *et al.*, 1982) obtained in the southern Vanuatu Islands (south of Efate) yield mantle velocities between 7.7 and 7.9 km/sec at depths of 15 to 25 km beneath the island arc.

Since 1978, Cornell University and the French Office de la Recherche Scientifique et Technique Outre-Mer (ORSTOM) have developed and maintained a network of seismic stations on the Vanuatu Islands (Figure 1). New data from this network can help to constrain the velocity structure of the central part of the arc.

The methods of analysis used in this paper are standard ones based on flat-layered models. Clearly, strong lateral variations in velocities exist in the area. The crustal and lithospheric structures change drastically from the oceanic plate to the island arc, and the subducted lithosphere probably forms a major three-dimensional



velocity anomaly within the mantle. The methods used in this paper, however, focus on the velocities in the uppermost mantle beneath the crust of the island arc using profiles aligned along the strike of the island arc. The main across-the-strike lateral variations are thus largely avoided. The part of the mantle sampled is the wedge-shaped region between the island arc crust and subducted plate.

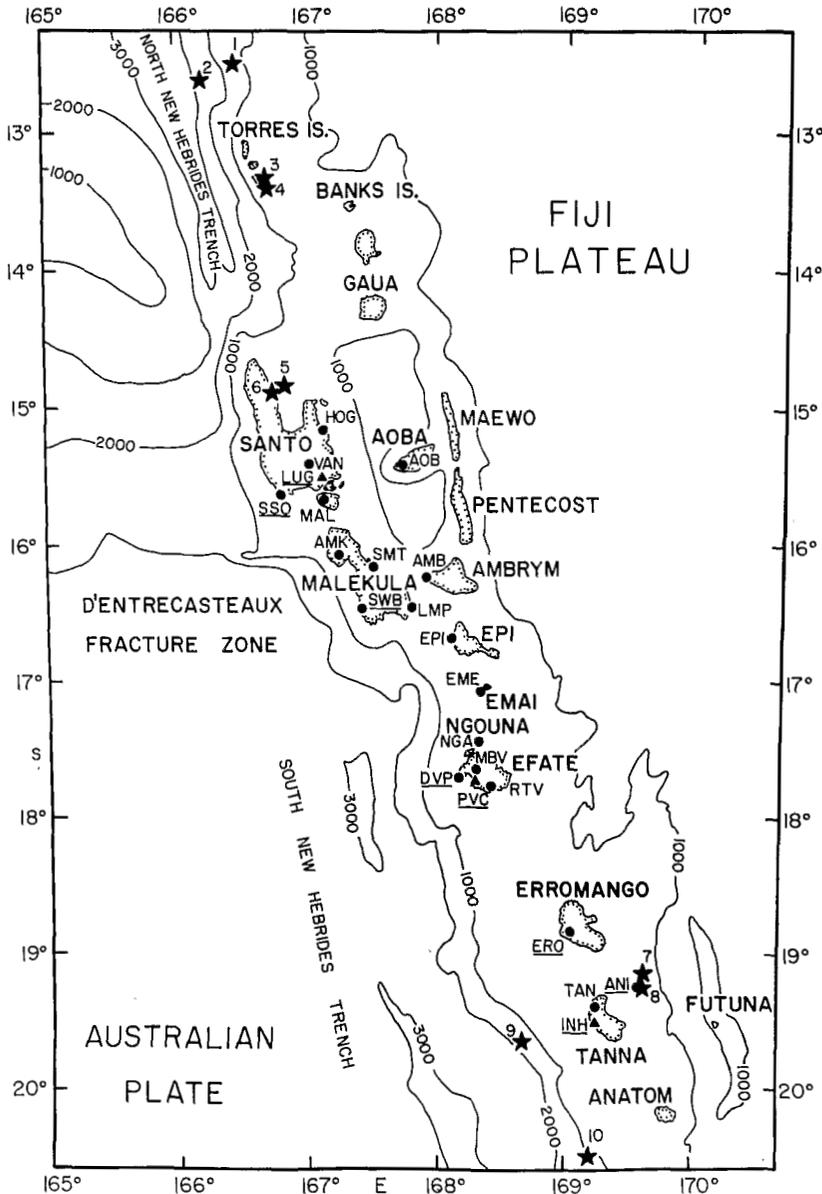


FIG. 1. Star symbols represent the locations of earthquakes used to determine the velocity of the mantle refractor beneath the seismograph network. Stations of the Cornell/ORSTOM seismograph network are shown as solid circles. Underlined stations have a vertical and one horizontal component, while other stations have only a single vertical component. Triangles are the long and short period multicomponent stations of the ORSTOM network. Bathymetric contours are in fathoms from Mamerickx *et al.* (1971). The D'Entrecasteaux "fracture zone" intersects the arc near the islands of Santo and Malekula and is outlined by the westerly bend of the 2000 fathom contour. The islands of Santo, Malekula and Efate lie trenchward of the volcanic front defined by the islands of Anatom, Tanna, Erromango, Emae, Ambrym, Aoba, and the Banks Islands.

Except for the Japanese-Kamchatka region, there are in fact very few studies of this part of a convergent plate boundary. Most of the modern refraction and reflection profiling is concentrated in the forearc region where the mantle is relatively shallow. The extensive work in the Japanese and Kurile-Kamchatka regions shows clearly that the mantle beneath the main arc has velocities of 7.6 to 7.7 km/sec, significantly smaller than the values near 8.1 km/sec found beneath many oceanic and continental areas. In particular, they are smaller than those found beneath the oceanic plate near and seaward of the trench of the subduction zone (e.g., Research Group for Explosion Seismology, 1977; Anosov *et al.*, 1980). The anomalous mantle region extends over a broad section of the island arc and is apparently not just concentrated beneath the magmatic arc itself. Low uppermost mantle velocities were also found beneath the Tonga island arc (Raitt *et al.*, 1955) and beneath the Aleutian island arc (Grow, 1973). The results of this study also reveal a region of low mantle velocities beneath the crust of the New Hebrides island arc.

APPARENT VELOCITIES OF EARTHQUAKES LOCATED OUTSIDE OF THE NETWORK

The first estimate of mantle velocity is obtained from a study of the apparent velocities of *P* waves recorded at the seismograph stations from shallow earthquakes occurring outside of the network. The method involves plotting epicentral distances versus arrival times for stations reporting individual earthquakes. The study is done with events that were well-located by the International Seismological Centre. The calculated apparent velocity gives an estimate of the mantle velocity under the network assuming the velocity structure can be represented by plane, horizontal layers.

The time-distance plots for 10 events are shown in Figure 2 and listed in Table 1. The mean velocity is 7.6 ± 0.2 km/sec. The apparent velocity from the six events north of the network varies between 7.3 and 7.8 km/sec and averages 7.66 ± 0.11 km/sec. Events 5 and 6 yield significantly lower velocities (7.3 km/sec) than the other events. These two events lie beneath northern Santo where the D'Entrecasteaux fracture zone is being subducted and are located closest to the network of seismograph stations. All other northern events yield an average velocity of 7.70 ± 0.14 km/sec. The apparent velocity of the four events south of the network is more consistent and equals 7.63 ± 0.05 km/sec.

The slightly higher velocities for the four northernmost events (1 to 4) compared to the four southernmost events (7 to 10) may not be due to their apparently greater depths (Table 1). For three of these events (1, 3, and 4) the ISC bulletin lists shallower depths (32 to 48 km) based on *pP* - *P* readings, and in all four cases the depths are poorly constrained by the *P* data alone. A second factor is that a model with plane, flat, horizontal layers may be inadequate for rays crossing the D'Entrecasteaux fracture zone, an aseismic ridge located between 15° and 16°S. In spite of these uncertainties, a rather low mantle velocity is indicated for the region from Efate to Malekula. In an effort to improve this estimate, the time-term method was used.

TIME-TERM ANALYSIS

The method of time-term analysis was applied to the *P*-wave data for three sets of events recorded at the local stations (Figure 3). A large amount of work has been done with the time-term method (see, e.g., Reiter, 1970; McCollom and Crosson,

1975) and it has proven to be very useful for determining crustal structure. Several specific features of the method which have been used here are discussed below.

The equation is

$$T_{ij} = \frac{\Delta_{ij}}{v} + C + d_i + d_j \tag{1}$$

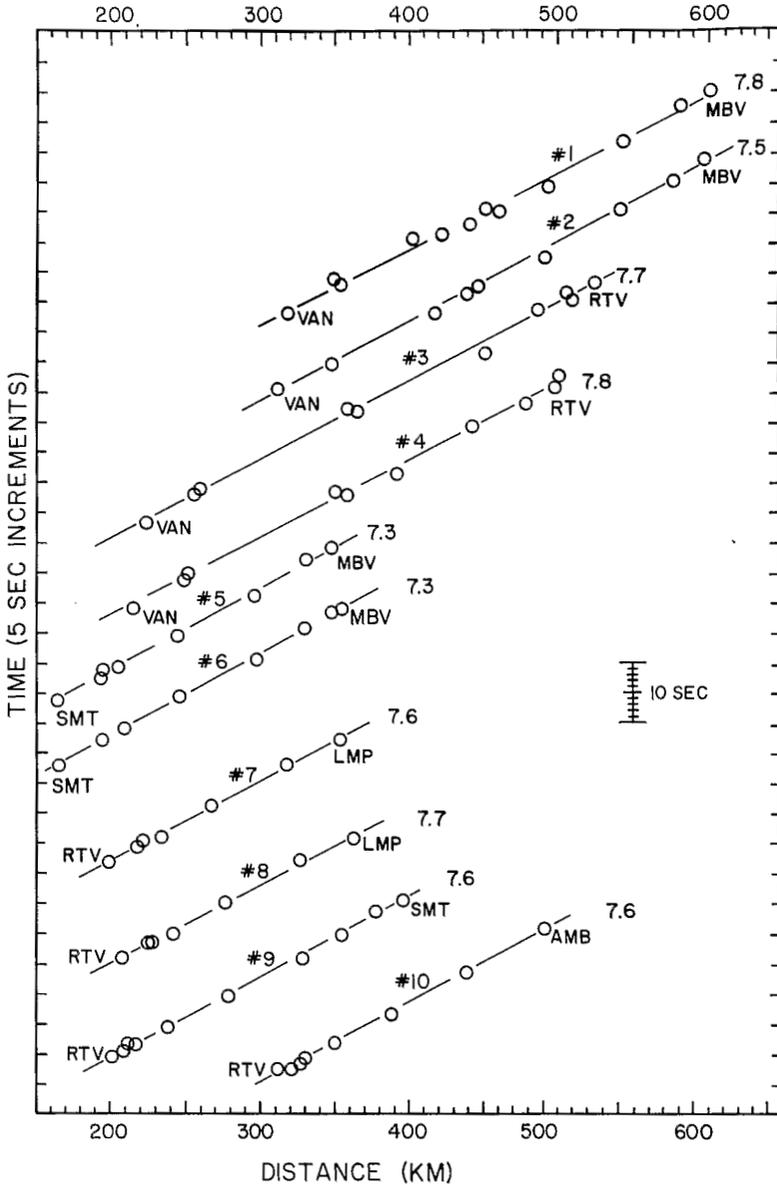


FIG. 2. Velocity of the mantle beneath the seismograph network determined from travel-time plots for the ten events in Figure 1. A fixed scale showing distance from the epicenter is used on the abscissa while a relative time scale is used on the ordinate. For each event the names of the nearest and farthest receiving stations are shown. The apparent velocity in kilometers/second is shown in the figure and in Table 1. Event numbers refer to the earthquakes shown in Figure 1 and listed in Table 1.

with

- T_{ij} = travel-time from event i to station j ,
 v = velocity of the medium (mantle) traveled by the refracted waves,
 Δ_{ij} = epicentral distance from event i to station j ,
 C = constant of the equation,
 d_j = station j time term, and
 d_i = event i time term.

TABLE 1

LEAST-SQUARES ESTIMATES OF THE APPARENT VELOCITY FROM MALEKULA THROUGH EFATE									
Event No.	Date (d/m/yr)	Origin Time	Latitude ($^{\circ}$ S)	Longitude ($^{\circ}$ E)	Depth (km)	Velocity (km/sec)	Slowness (sec/km)	Standard Deviation	
1	31/08/80	14:12:43.1	12.53	166.46	71	7.85	0.127	0.003	
2	15/11/80	03:08:53.2	12.72	166.15	73	7.54	0.133	0.002	
3	09/11/79	16:39:27.3	13.32	166.69	62	7.68	0.130	0.003	
4	02/11/79	01:32:23.2	13.40	166.70	71	7.83	0.128	0.002	
5	08/12/80	19:13:46.6	14.81	166.89	44	7.32	0.137	0.004	
6	08/12/80	21:28:32.0	14.93	166.75	27	7.34	0.136	0.002	
7	09/10/80	10:44:09.0	19.09	169.65	3	7.60	0.132	0.003	
8	09/10/80	15:29:00.4	19.31	169.62	35	7.66	0.130	0.003	
9	22/11/78	05:02:52.9	19.67	168.68	0	7.57	0.132	0.002	
10	20/12/79	13:11:42.7	20.50	169.20	38	7.55	0.132	0.003	

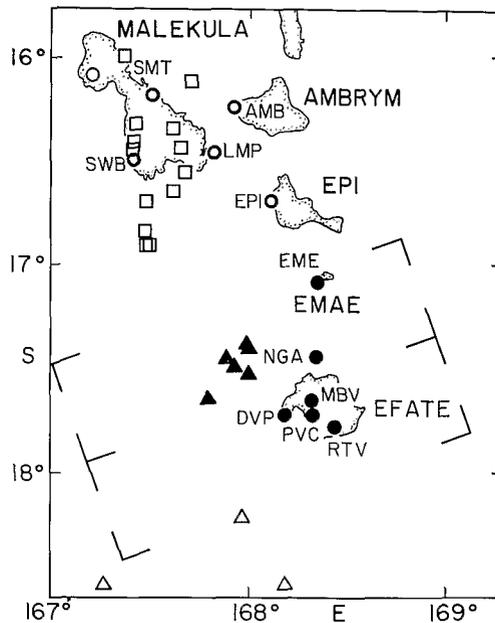


FIG. 3. Locations of the earthquakes and seismograph stations used in the time-term study (Table 2). Triangles indicate events studied using the northern stations (open circles). Filled triangles represent events located by the local network (set 1); open triangles represent events located by the U.S. Geological Survey in their *PDE Bulletin* (set 1A). The squares indicate events studied using the southern stations (filled circles) and were located by the local network (set 2). Limits of cross-section in Figure 5 are shown.

With equation (1), a set of n events recorded by a set of m stations can be studied together. The stepwise multiple regression (Draper and Smith, 1966) is used to search for $1/v$, d_i , d_j , and C . With this method only the unknown parameters which provide significant contributions to equation (1) are considered. To have a unique

solution, one event time-term and one station time-term must be fixed (see Table 2). One of the advantages of the time-term method is that the errors in the epicentral parameters and in the origin time can be absorbed in the earthquake (event) time-term. However, if the location is accurate and the medium has no lateral heterogeneity, C gives information on the thickness of the upper layer (H), d_i gives

TABLE 2
RESULTS OF TIME-TERM ANALYSIS

Set No.	No. of Events	No. of Stations	Velocity (km/sec)	H (km)	H_r (km)	C (sec)	% of Data Explained
1	6	5	7.55 ± 0.05	26	22	3.2	99.2
1A	9	5	7.55 ± 0.05	25	22	3.1	99.8
2	13	6	7.70 ± 0.10	29	30	3.2	99.4

Set No.	Date (d/m/yr)	Event Time Terms (d_i in sec)				
		Latitude ($^{\circ}$ S)	Longitude ($^{\circ}$ W)	Depth (km)	d_i (sec)	H (km)
1	03/10/78	17.40	167.98	32	-0.8 ± 0.2	33
	13/10/78	17.53	167.99	32	-0.8 ± 0.2	33
	20/10/78	17.50	167.92	22		22
	24/11/78	17.48	167.88	13	$+1.4 \pm 0.2$	3
	01/12/78	17.41	168.00	30	-1.3 ± 0.2	39
	25/03/79	17.65	167.79	22		22
1A	17/11/78	18.21	167.98		*	
	15/06/79	18.54	167.27		*	
	14/08/79	18.53	168.19		*	
2	05/11/78	16.65	167.61	15	$+0.5 \pm 0.2$	24
	03/12/78	16.44	167.65	27		30
	24/02/79	16.91	167.49	29		30
	20/03/79	16.41	167.41	13	$+1.5 \pm 0.2$	11
	25/03/79	16.85	167.47	35	-0.5 ± 0.2	36
	30/05/79	16.44	167.42	23	$+0.5 \pm 0.1$	24
	06/06/79	16.35	167.61	36		30
	12/06/79	16.91	167.48	29		30
	14/06/79	16.70	167.48	29		30
	24/06/79	16.12	167.70	32		30
	05/06/79	16.33	167.42	30		30
	25/08/79	16.00	167.37	34		30
	15/11/79	16.56	167.67	26	$+0.4 \pm 0.1$	25

Set No.	Station Time Terms (d_i in sec)										
	RTV	PVC	DVP	MBV	NGA	EME	EPI	LMP	SWB	AMB	SMT
COR†	0.0	-0.2	0.0	0.0	0.0	0.0	0.0	-0.3	0.0	+0.4	0.0
1							0‡	-0.4 (± 0.1)	X§	+0.8 (± 0.1)	X§
1A							0‡	-0.3 (± 0.1)	X§	+0.8 (± 0.1)	X§
2	+0.3 (± 0.1)	-0.1 (± 0.1)	X§	+0.1 (± 0.1)	0‡	X§					

* No event time terms (d_i) are given for the three events added to set 1 to make set 1A because their original depths are poorly constrained in the *PDE Bulletin*.

† COR are the station corrections deduced from a study of P arrival-time residuals using the output of the HYP071 location program (Lee and Lahr, 1975).

‡ 0 indicates a station with the time term fixed at zero.

§ X indicates stations that did not have any significant time-term value in the stepwise multiple regression.

information on the depth of the event (H_i), and d_j gives information on the variation of the crustal travel time under the station which may be due either to a variation of the depth of the Moho and/or the crustal velocity.

The expressions for the individual time terms are given below.

$$C = \sum_{k=1}^n H_k \cdot B_k + (H - H_r) \cdot B_n$$

$$d_i = (H_r - H_i) \cdot B_n$$

$$d_j = dH_j \cdot B_n \quad \text{or} \quad d_j = H_n \cdot dB_n \quad (2)$$

where

$$B_k = \sqrt{v^{-2}_k - v^{-2}}$$

The crust with a thickness H is assumed to be composed of n layers. Each of these layers has a thickness H_k and a velocity V_k ($k = 1, n$). H_r is the depth of a reference hypocenter whose location is assumed to be known. The event i at a depth H_i as well as the reference hypocenter at a depth H_r are assumed to be located near the Moho in the layer n . In the interpretation of d_j , it is assumed that the station time term is due to a variation dH_j of the thickness of the layer n or a variation dB_n related to a velocity change in the layer n .

The results of the time-term study and the epicenter parameters are presented in Table 2. The percentage of total variation about the mean explained by the regression (Draper and Smith, 1966) in this study varied from 99.2 to 99.9 per cent depending on the data set (Table 2). Based on refraction experiments done in the area (Ponzoise *et al.*, 1982), it has been assumed that the crust is composed of three layers with velocities equal to 3, 5, and 6.6 km/sec. The thicknesses of the two first layers are assumed to be 3 and 4 km. It is the thickness of the third layer which is deduced from the constant C .

If the depth of the reference event H_r is known, the thickness H of the crust can be deduced. The accuracy of this thickness will thus be a function of the accuracy of the depth of this reference event, of the crustal velocity and of the constant itself. For 5 km uncertainty in the depth of the reference event, 5 per cent in the crustal velocity, and 0.3 sec in C , the uncertainty on the crustal thickness is around 7 km.

If the crust has no lateral heterogeneity and if the crustal thickness under the event is H as deduced above, then the event time term d_i can be interpreted to give the depth H_i of the event. Table 2 shows that for the events used in this study, there was a good agreement between the depths given by the location with the local data and the depths deduced from the time terms.

Assuming crust without lateral heterogeneity, the station time-term d_j may be interpreted as the variation of the Moho depth dH_j , i.e., $d_j = dH_j \cdot B_3$. Alternatively, the station time terms may be interpreted as a variation of the velocity in the third layer assuming a constant crustal thickness, i.e., $d_j = H \cdot dB_3$, with dB_3 being a function of the change in velocity [equation (2)]. Table 2 shows that only the station time-terms at AMB and LMP show significant (≥ 0.4 sec) values.

A detailed study of the P -wave travel-time residuals (observed minus calculated) at the local stations for a selected set of events recorded during the 1978 land-ocean bottom seismograph experiment (Cardwell *et al.*, 1979) as well as general patterns of residuals deduced from the many events located routinely with the local network

also lead to the conclusion that there are no significant station corrections (within 0.3 sec) at most of the local stations. As with the time-term analysis, only LMP on Malekula and AMB on Ambrym have significant corrections. Station AMB is on the active volcano Ambrym which may account for the positive value of its time term. There is no obvious explanation for the negative value of the time term at LMP.

CRUSTAL THICKNESS AND MANTLE VELOCITY DETERMINED FROM THE TIME-TERM ANALYSIS

The results of the time-term analysis are shown in Table 2 for three data sets. Set 1 consists of six events shown in Figure 3 located by the Efate network and also recorded by the five stations located on Epi and Malekula Islands. The epicentral distances vary from 80 to 160 km. This data set yields a mantle velocity of 7.55 (± 0.05) km/sec beneath a 26-km-thick crust. An enlarged set of events (set 1A) containing the six events of set 1 and three more distant events listed in the *PDE Bulletin* (open triangles in Figure 3) gives the same low *P*-velocity value: 7.55 (± 0.05) km/sec beneath a 25-km-thick crust.

A "reverse profile" is obtained by another set of 13 events (set 2) located by the Malekula network (see Figure 3) and recorded by six stations near Efate Island (filled circles in Figure 3). The epicentral distances vary from 90 to 220 km. This set yields a mantle velocity of 7.70 (± 0.10) km/sec beneath a 29-km-thick crust.

The depths of all the 19 events located by the local network and studied with the time-term method vary between 13 and 36 km (Table 2) with a mean depth of 27 km. Since the foci are near the depth of the expected Moho discontinuity, the *P*-wave arrivals recorded at relatively short distances have an apparent velocity equal to or larger than the mantle velocity. The results of the time-term study agree with the study of apparent velocities and yield an average mantle velocity from Malekula to Efate of 7.6 km/sec (± 0.1).

As discussed previously, the accuracy of the crustal thickness is probably about 7 km. Thus, the calculated crustal thickness of 25 to 29 km agrees with the 21 to 27 km obtained from the study of *S*-to-*P* converted phases under Efate Island (Chiu, 1982) and with the 24 km obtained by Malahoff (1970) from Bouguer gravity anomalies. The calculated thickness also agrees with the 26 km crustal thickness beneath the region from Erromango to Anatom Islands in the southern part of the arc determined by refraction (Ibrahim *et al.*, 1980; Pontoise *et al.*, 1982).

TILT OF THE MOHO

The apparent velocities deduced from northern and southern earthquakes have the same values within the resolution of the method and thus do not indicate any resolvable tilt of the Moho caused by changes in crustal thickness along the strike of the arc. Moho tilt can also be determined from the apparent azimuths of wavefronts arriving at the subarray of five stations located on Efate Island from earthquakes near Malekula (assuming that the wave front is planar under the Efate subarray). Only the apparent azimuths for the 7 of the 13 events in set 2 with a standard error less than 10° are plotted in Figure 4. The rays coming from the events located under Malekula did not have any significant shift in azimuth. However, there is an indication of a shift in the apparent azimuth for the three most southern events. The apparent azimuth is west of the true azimuth suggesting that the rays travel in a layer dipping down toward the northeast.

Since the three southernmost earthquakes are quite near the network on Efate,

the plane wave hypothesis may not be completely valid. To check this possibility, these three events are relocated with the HYPO71 program (Lee and Lahr, 1975) using only the *P*-wave arrival times at the Efate subarray and just one *S*-wave reading to fix the epicentral distance. This is a simple way to take into account the

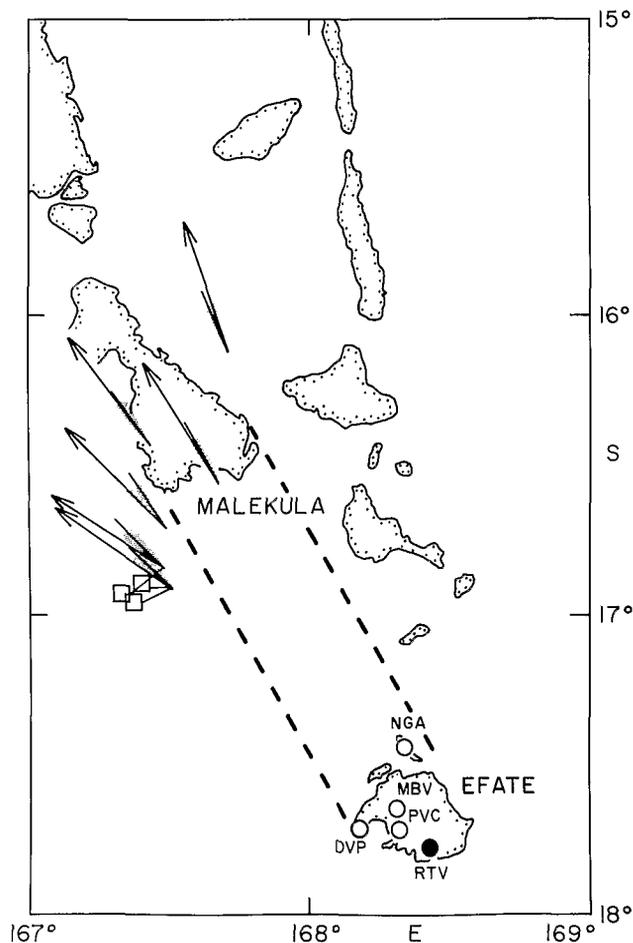


FIG. 4. The direction of the apparent azimuth of a *P* wave coming from an earthquake hypocenter of set 2 and recorded at the Efate stations (circles) is reported at the event epicenter as an arrow. This direction is obtained assuming the wave front is planar. It is deduced from the differences of the *P*-wave arrival times between the stations and a given reference station. The solid line indicates the true azimuth at the reference station (usually RTV, shown by a filled circle). The shaded area gives the range of azimuths observed at the network stations and thus indicates the limitation of the method. The open squares show the locations of three hypocenters south of Malekula obtained when the data at the Efate stations are used alone. The lines connect them to the epicenter obtained when the data of the Malekula stations are included. For these three events, both the apparent azimuths and the relocations indicate a slight tilt of the refractor down toward the north and the east directions. From the events under Malekula, no significant tilt of the refractor is observed between the islands of Malekula and Efate as schematized by the dashed lines.

curvature of the wave front. The relocation of these events (open squares in Figure 4) confirm the shift of apparent azimuth. Thus, south of Malekula, there is evidence of a slight tilt of the Moho down to the northeast. This observation is in agreement with the study of Chiu (1982) who deduced a thinning of the crust west of Efate

from the study of *S*-to-*P* converted phases. This tilt may be related to a root associated with the islands of Malekula and Efate. However, since this tilt is in the direction of the subducting plate it may be related to a discontinuity associated with the subducting plate. This relationship is illustrated in Figure 5 on a vertical section of the shallow seismicity under Efate reported by Isacks *et al.* (1981).

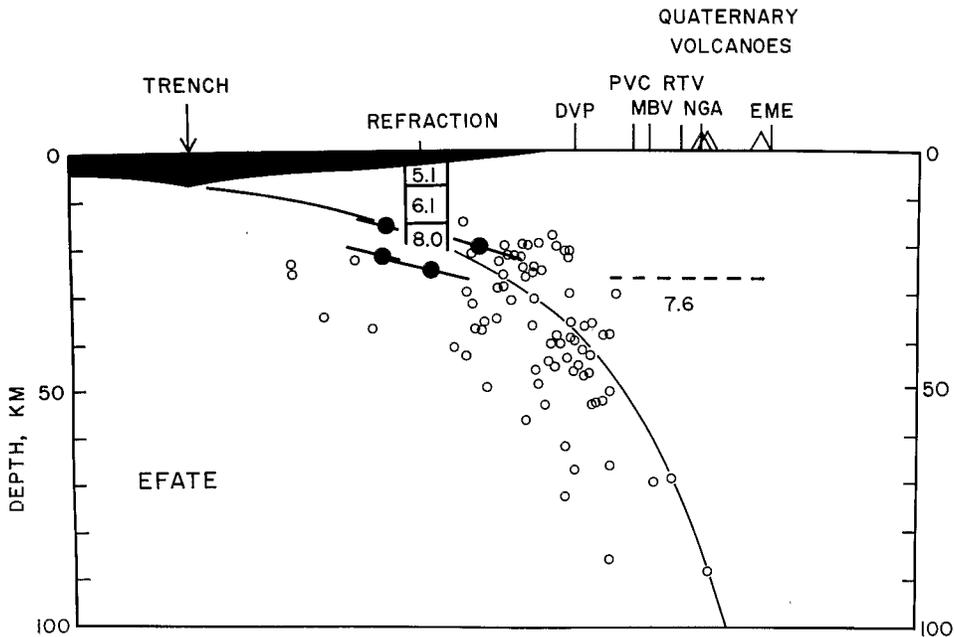


FIG. 5. Vertical cross-section oriented perpendicular to the arc near Efate (modified from Isacks *et al.*, 1981). Limits of cross-section are shown in Figure 3. The locations of some of the best earthquake hypocenter determinations are shown as small circles. Lines through the large filled circles are projections of slip vectors for nearby earthquakes ($M_S \geq 6$). The solid curve is an estimate of the top of the subducted lithosphere. The present study found a mantle velocity of 7.6 km/sec at 27 km beneath the seismograph network (dashed line). Results of a refraction survey near the trench are also shown (from Ibrahim *et al.*, 1980).

DISCUSSION AND CONCLUSIONS

The mantle velocity of 7.6 km/sec and the crustal thickness of about 27 km are obtained assuming the velocity structure is a plane, horizontally layered medium. This assumption is open to question because subduction zones are known to have quite heterogeneous structure on a regional scale. However, for the region from the middle of Malekula Island to the south of Efate Island, the data do not suggest any significant along-strike tilt of the Moho. The same velocity and crustal thickness are obtained for reverse profiles. Furthermore, the determinations of apparent azimuths do not indicate any tilt component perpendicular to strike beneath or between the two islands. Only for the region west of Malekula and near the trench is there any indication of a tilt of Moho towards the island arc. Low mantle velocities of 7.7 to 7.9 km/sec (at 20 to 26 km depth) were also determined by refraction studies for the volcanic arc near Tanna and Erromango Islands in the southern New Hebrides arc (Pontoise *et al.*, 1982).

The results of mantle velocity investigations near Efate are shown on the cross-section of Figure 5 along with a representative sample of seismicity. The sharp velocity contrast between the arc and forearc-trench regions is strikingly similar to

the velocity variations observed beneath the Japanese island arc. The Research Group for Explosion Seismology (1977) reported a very sharp transition in the mantle velocity from 7.5 km/sec beneath Honshu to 8.0 km/sec near the Japanese trench. Numerous studies are in accordance with these results (see, e.g., Kanamori, 1970; Suzuki, 1978; Okada, 1979; Asano *et al.*, 1981; Yoshii *et al.*, 1981). Similarly low velocities in the uppermost mantle beneath the main arc are reported for the Kurile-Kamchatka arc (Anosov *et al.*, 1980), the Aleutian arc (Grow, 1973), and the Tonga arc (Raitt *et al.*, 1955). These low velocities, near 7.6 to 7.7 km/sec, are characteristic of regions where mantle temperatures are elevated above those found in the mantle part of the lithosphere. Examples of such regions are actively spreading oceanic ridges and back arc basins (e.g., Ambos and Hussong, 1982) and the Great Basin of the Western United States.

The refraction results summarized on Figure 5 show an 8.0 km/sec velocity beneath the fore arc of the Efate section. The depth to this layer appears too shallow to be the mantle beneath the subducted oceanic crust, but seems to be located in the wedge of mantle between the island arc crust and the subducted plate. Studies in the other regions already referenced often show a region of more normal mantle velocities (or densities in gravity models) in this part of the wedge, and suggest that the anomalous mantle, although located seaward of the magmatic front, may not extend all the way to the descending plate.

Long profiles aligned along the strike of the Tonga-Kermadec arc (Aggarwal *et al.*, 1972) show high P_n and S_n velocities (8.4 and 4.8 km/sec, respectively) in contrast to the results of this study and the results of Raitt *et al.* (1955). These high velocities were obtained along profiles of nearly 1500 km length along which the waves apparently traveled within the subducted lithosphere. The shorter-range refraction profiles, using surface sources, probably sampled the uppermost part of the mantle wedge above the subducted plate. It is interesting that we see little indication of high velocity waves which might have refracted along the subducted plate, even though most of the shallow earthquake sources are likely to be quite close to the upper boundary of the subducted plate. It is possible that the very sharp downward flexure of the descending plate in the New Hebrides (in contrast to that in the Tonga-Kermadec region) may provide a less favorable propagation path than one through the mantle above the subducted plate. The path would have a significant lateral deviation from a great circle path and would refract into and out of the descending plate in the region of maximum downdip curvature.

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REFERENCES

- Aggarwal, Y. P., M. Barazangi, and B. Isacks (1972). P and S travel times in the Tonga-Fiji Region: a zone of low velocity in the uppermost mantle behind the Tonga island arc, *J. Geophys. Res.* **77**, 6427-6434.
- Ambos, E. L. and D. M. Hussong (1982). Crustal structure of the Mariana trough, *J. Geophys. Res.* **87**, 4003-4018.
- Anosov, G. I., S. K. Bikkenina, V. I. Fedorchenko, A. A. Popov, K. F. Sergeev, and V. K. Utnasin (1980). On the junction character of the continental and oceanic lithospheric blocks in the Kamchatka region, in *Structure of the Transition Zone*, S. Asano, Editor, Center for Academic Publication,

- Tokyo, Japan, 33-51.
- Asano, S., T. Yamada, K. Suyehiro, T. Yoshii, Y. Nisawa, and S. Iizuka (1981). Crustal structure in a profile off the Pacific coast of northeastern Japan by the refraction method with ocean bottom seismometers, *J. Phys. Earth* **29**, 267-281.
- Cardwell, R. K., B. L. Isacks, R. Louat, G. V. Latham, and A. Chen (1979). First results from a new seismograph network in the central New Hebrides island arc, *EOS, Trans. Am. Geophys. Union* **60**, 338.
- Chiu, J.-M (1982). Structural features of subduction zone determined by detailed analysis of short period waves from earthquakes recorded in the New Hebrides Island arc, *Ph.D. Thesis*, Cornell University, Ithaca, New York.
- Draper, N. R. and H. Smith (1966). *Applied Regression Analysis*, John Wiley and Sons, New York, 407 pp.
- Dubois, J. (1971). Propagation of P waves and Rayleigh waves in Melanesia: structural implications, *J. Geophys. Res.* **76**, 7217-7240.
- Dubois, J., G. Pascal, M. Barazangi, B. L. Isacks, and J. Oliver (1973). Travel times of seismic waves between the New Hebrides and Fiji Islands: a zone of low velocity beneath the Fiji Plateau, *J. Geophys. Res.* **78**, 3431-3436.
- Grasso, J. R. (1981). Inversion des anomalies de propagation des ondes sismiques: Inverse stochastique et programmation lineaire, application aux Nouvelles Hebrides et aux Pyrenees, *Thèse de Docteur-Ingenieur*, Universite de Grenoble, Grenoble, France.
- Grow, J. A. (1973). Crustal and upper mantle structure of the central Aleutian arc, *Bull. Geol. Soc. Am.* **84**, 2169-2192.
- Ibrahim, A., B. Pontoise, G. Latham, B. Larue, T. Chen, B. Isacks, J. Recy, and R. Louat (1980). Structure of New Hebrides Arc-Trench system, *J. Geophys. Res.* **85**, 253-266.
- Isacks, B. L., R. K. Cardwell, J. L. Chatelain, M. Barazangi, J. M. Marthelot, D. Chinn, and R. Louat (1981). Seismicity and tectonics of the central New Hebrides island arc, in *Earthquake Prediction, An International Review*, D. W. Simpson and P. G. Richards, Editors, Maurice Ewing Series, 4, AGU, Washington, D.C., 93-116.
- Kanamori, H. (1970). Mantle beneath the Japanese arc, *Phys. Earth Planet. Interiors* **3**, 475-483.
- Lee, W. H. K. and J. C. Lahr (1975). HYPO71 (revised): a computer program for determining hypocenter, magnitude and first motion pattern of local earthquakes, *U.S. Geol. Surv., Open-File Rept. 75-311*, Arlington, Virginia 114 pp.
- Malahoff, A. (1970). Gravity and magnetic studies of the New Hebrides Island arc, Report of the New Hebrides Geol. Surv., Port Vila, New Hebrides, 67 pp.
- McCollom, R. and R. Crosson (1975). An array study of upper mantle velocity in Washington State, *Bull. Seism. Soc. Am.* **65**, 467-482.
- Okada, H., S. Asano, T. Yoshii, A. Ikami, S. Suzuki, T. Hasegawa, K. Yamamoto, K. Ito, and K. Hamada (1979). Regionality of the upper mantle around northeastern Japan as revealed by explosions at sea, *J. Phys. Earth* **27**(Suppl.), 515-532.
- Pontoise, B., G. Latham, and A. Ibrahim (1982). Sismique refraction: Structure de la croûte aux Nouvelles Hebrides, in *Contribution a l'etude geodynamique du Sud-Ouest Pacifique*, Travaux et Documents de l'ORSTOM, no. 147, Paris, France, 79-90.
- Raitt, R. W., R. L. Fisher, and R. G. Mason (1955). Tonga Trench, *Geol. Soc. Am. Special Paper* **62**, 237-254.
- Reiter, L. (1970). An investigation into the time-term method in refraction seismology, *Bull. Seism. Soc. Am.* **60**, 1-13.
- Research Group for Explosion Seismology (1977). Regionality of the upper mantle around northeastern Japan as derived from explosion seismic observations and its seismological implications, *Tectonophysics* **37**, 117-130.
- Suzuki, S. (1978). Lateral variation of the upper mantle structure around Northern Japan and its application to hypocenter determination, *J. Fac. Sci. Hokkaido Univ.* **5**, 79-120.
- Yoshii, T., H. Okada, S. Asano, K. Ito, T. Hasegawa, A. Ikami, T. Noriya, S. Suzuki, and K. Hamada (1981). Regionality of the upper mantle around northeastern Japan as revealed by big explosions at sea, *J. Phys. Earth* **23**, 201-220.

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