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Seismic Investigation of the Papuan Ultramafic Belt

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Summary

Seismic survey data from a profile along the north-east Papuan peninsula coast indicate velocities which can be correlated with those of oceanic layers 1, 2 and 3, but with a crustal thickness of between 20 and 25 km. Distinct differences in the crustal layering are determined between the region where the Papuan Ultramafic Belt crops out along the coast and the region of the Trobriand Platform. The crustal thickness is more than twice the total thickness of gabbro and basalt components of the ophiolite suite exposed inland. Later seismic arrivals suggest the presence of a low-velocity zone below the Moho and a return to upper mantle velocities at depths of about 50 km.

Introduction

The ophiolite suite of rocks in eastern Papua described by Davies (1971) is widely considered to be a classic example of a fragment of oceanic crust and upper mantle obducted on to a continental mass during orogenesis in an area of continent/island arc convergence (Coleman 1971; Moores & Vine 1971; Moores 1973). Dewey & Bird (1970) have stressed the importance of the study of such rock sequences in the development of mountain belts and indicate that ophiolites are present in many of the major intra-continental orogenies such as those in the Urals and in the Appalachian–Caledonian belts (Dewey & Bird 1971). The time–space relationships of the younger ophiolite belts therefore play an important role in placing constraints on interpretations of continental interaction and accretion at plate boundaries.

The East Papua Crustal Survey conducted during October–December 1973 was a seismic investigation of the major structures in the region of the Papuan Ultramafic Belt and was designed to add more definitive geophysical constraints to interpretations of the nature and evolution of the Belt. The results presented in this paper are some initial observations which are of importance in providing a framework for more detailed interpretation.

Papuan Ultramafic Belt

Davies (1971) and Davies & Smith (1971) described the Papuan Ultramafic Belt and its geological setting in the Papuan peninsula in some detail. The belt is a peridotite-gabbro-basalt complex which crops out over a length of 400 km and width of 40 km to the north-east of the Owen Stanley Range which runs the length of the Papuan peninsula (Fig. 1). The complex comprises, from top to bottom, some

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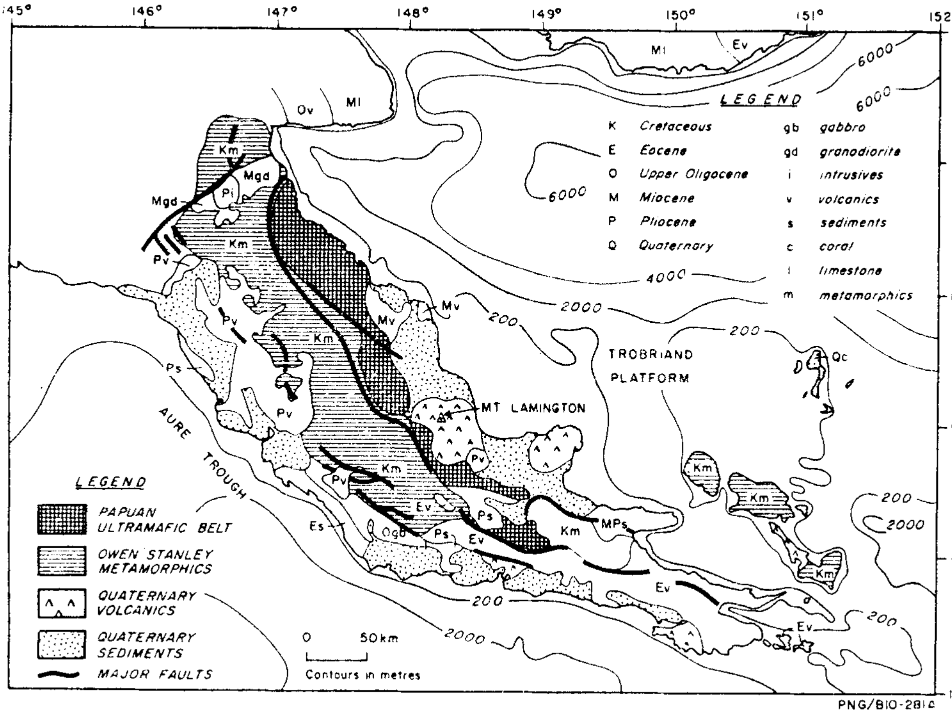

FIG. 1. Simplified geology and bathymetry of the Papuan peninsula region.

4–6 km of basalt and spilite in massive form and as pillow lavas with some dacite, 4 km of high-level gabbro, granular gabbro, and cumulates, and 4 to 8 km of ultramafic rocks of which up to 0·5 km are cumulates and the remainder are non-cumulate harzburgite, dunite, and orthopyroxenite. The basalt, gabbro and cumulate ultramafics probably crystallized in the Cretaceous and were intruded by tonalite in the Eocene. The ophiolites were emplaced in Eocene or Oligocene time.

Mesozoic schist and gneiss of varied metamorphic grade form the sialic core of eastern Papua and abut the ultramafic suite along the Owen Stanley Fault system. The metamorphics were formed during the Eocene from Cretaceous or older sediments derived from continental Australia, the most common grade being greenschist facies; lawsonite occurs in the metamorphic rocks along the entire length of the Owen Stanley Fault.

Regional gravity observations have been made throughout the Papuan peninsula (St John 1970; Milsom 1973; Finlayson, in preparation). Milsom (1973) has made a detailed interpretation of the gravity data and considered the constraints on the emplacement of the ophiolites and their possible origin. The data are compatible with the ophiolites being slabs of oceanic or frontal-arc overthrusts with dip angles of 25°–60°, but the means of emplacement is still not uniquely determined. The dips are much larger than those determined by seismic work (13°–19°) by Finlayson *et al.* (1976). Milsom (1973) has considered other modes of emplacement for the ophiolite structure in eastern Papua and regards the ' orogenic thickening ' hypothesis (Rabinowitz & Ryan 1970) and the ' pluto-volcanic ' hypothesis as unlikely. He regards it as possible that the overthrust material may be derived from a marginal basin rather than normal oceanic crust.

The structures to the south-west of Papua in the Coral Sea Basin have been investigated by Ewing, Hawkins& Ludwig (1970), Mutter (1975) and Gardner (1970). These

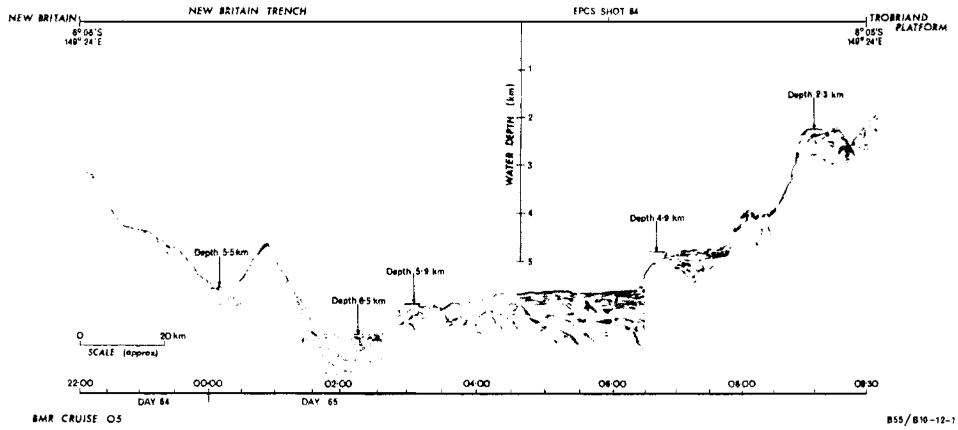

FIG. 2. Seismic sparker traverse across the Solomon Sea from the Trobriand Platform to New Britain (location shown in Fig. 3).

demonstrate that crust of oceanic thickness is present in the Basin between the Coral Sea Plateau and the Papuan Platform. Crustal extension during the early Eocene is proposed to account for formation of the Basin (Davies & Smith 1971). In the north-west Coral Sea a Mesozoic and Tertiary sedimentary section more than 10 km thick (St John 1970) overlies a predominantly continental type crust extending north-east from Australia (Finlayson 1968; Brooks 1969).

North-east of Papua the central part of the Solomon Sea Basin has a crust of oceanic thickness (11–14 km) but the upper mantle velocities ($7 \cdot 7$–$8 \cdot 0\ \mathrm{km\ s^{-1}}$) tend to be low (Furumoto *et al.* 1970). Profiling records from the Basin show it is characterized by rough topography (Fig. 2) with pockets of sediments. The free-air gravity values in the centre of the Basin are predominantly positive (0–100 mGal) with considerable variation in the areas of high structural relief at its margins (Rose, Woollard & Malahoff 1968).

The tectonic framework of the region has been studied, mainly on the basis of seismicity, by a number of authors (Denham 1969; Milsom 1970; Ripper 1970; Johnson & Molnar 1972; Krause 1973; Luyendyk, MacDonald & Bryan 1973; Curtis 1973). The Melanesian region is regarded as an area of convergence between the Pacific and Indian–Australian lithospheric plates with the smaller North Bismarck, South Bismarck and Solomon plates in the interaction zone. Minor seismicity at shallow and intermediate depth occurs along the eastern Papuan peninsula, and the formal solution to plate motions requires a transition in this area between a spreading centre in the Woodlark Basin and the Benioff zones along northern New Guinea and New Britain (Krause 1973). Vertical movements also have played, and continue to play, an important role in the tectonics of the peninsula, both upwards in its eastern part (Smith 1970; Davies & Smith 1971) and downwards as is evident along the drowned coastline where the Ultramafic Belt crops out on the coast (von der Borch 1972).

In the southern part of the Ultramafic Belt large Pliocene and Quaternary strato-volcanoes occur in the Mount Lamington–Hydrographer Range–Managalase province round seismic station LMG (Fig. 3), and in the Mount Victory–Mount Trafalgar area round the seismic station at Tufi (Fig. 3). Mount Lamington (Taylor 1958) and Mount Victory (Smith 1969) have been active in the last hundred years. The tectonic significance of the volcanology of eastern Papua has been discussed by many authors (Jakes & White 1969; Jakes & Smith 1970; Johnson, Mackenzie & Smith 1970;

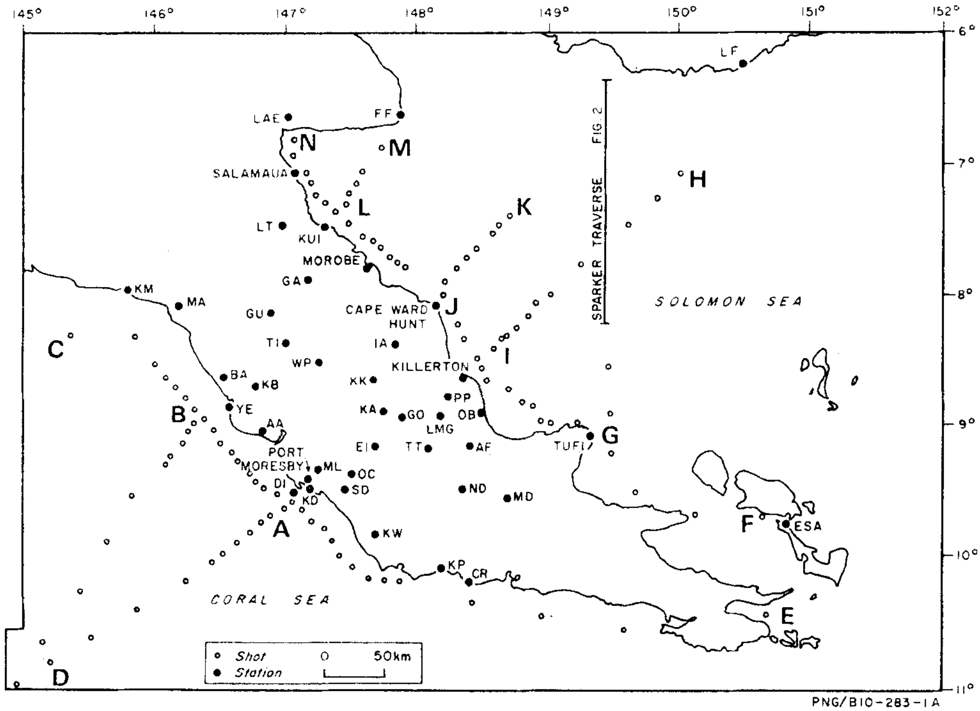

FIG. 3. East Papua Crustal Survey: shot and recording station locations, and position of sparker traverse shown in Fig. 2.

Smith 1973; Finlayson *et al.* 1976), and the structure of the Trobriand Platform (Fig. 1) has been reviewed by Bickel (1974).

Seismic survey

The Australian Bureau of Mineral Resources, Geology and Geophysics conducted the East Papuan Crustal Survey during October–December 1973 in co-operation with the Australian National University, the University of Queensland, the Geological Survey of Papua New Guinea, the University of Hawaii, Preston Institute of Technology and Warrnambool Institute of Advanced Education (Finlayson, in preparation). One-hundred and eleven shots were fired in the Coral and Solomon Seas and recordings were made at 42 seismic stations in Papua New Guinea (Fig. 3). The shots were mostly of two sizes, 180 kg and 1000 kg, detonated 100 m below the water surface where depth of water permitted. A timing accuracy of 0·01 s tied to the VNG Australian Post Office radio time signals was achieved for most shot instants and a positional accuracy of 0·1 minutes of arc was aimed at, using an extended SHORAN navigation system.

All seismic stations recorded VNG radio time signals for use as their time standard. Station positions were determined prior to seismic field work using a Decca trisponder trilateration network tied to the Australian Map Grid. Four seismic recording stations were part of the permanent PNG seismic observatory network and recorded on photographic or smoked paper recorders. Most of the remaining stations used high-gain seismic amplifiers coupled to slow-speed tape recorders, the majority operating unattended, either continuously (Muirhead & Simpson 1972) or programmed to operate for the first quarter of an hour every hour when shot firing was scheduled. Analogue

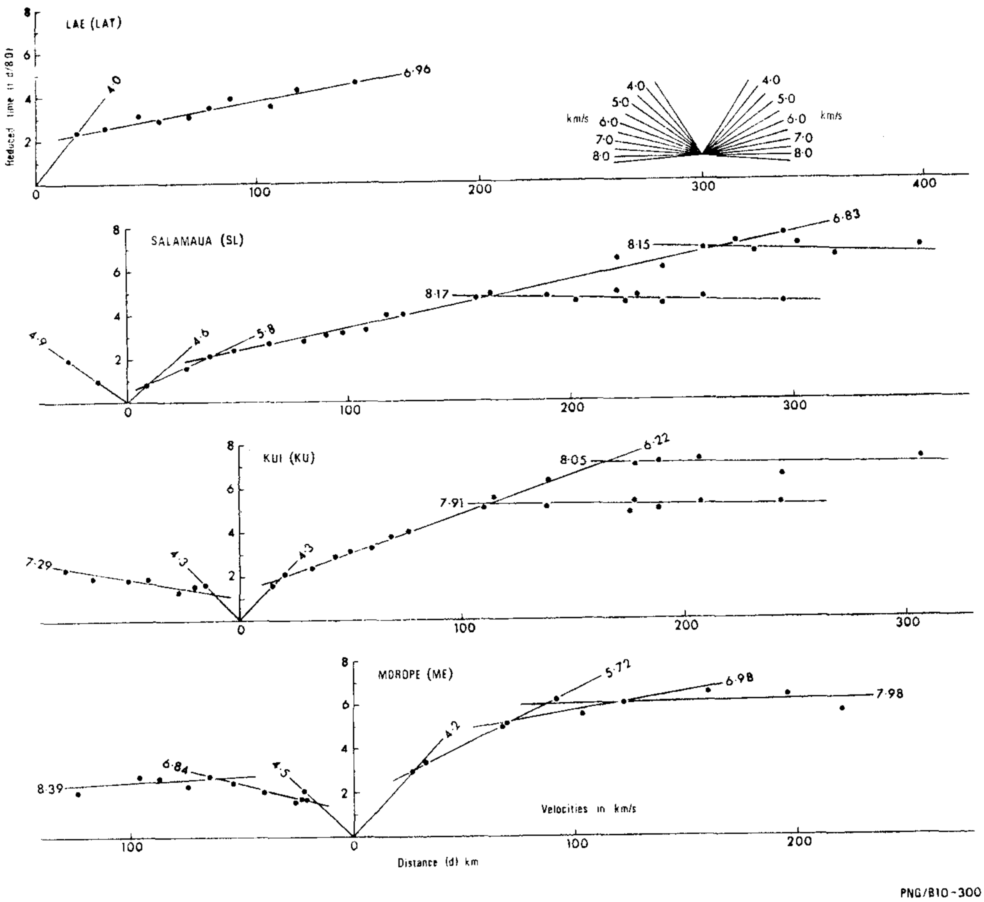

FIG. 4. Reduced travel-time plots for recording stations Lae, Salamaua, Kui and Morobe from shots along the north-east Papuan peninsula coast.

records were subsequently produced from all tapes and significant arrival times were picked by two or more seismologists. The details of the survey operations and data from the survey have been described by Finlayson (in preparation).

Seismic recording along the north-east Papuan coast

An initial interpretation has been made of data recorded along the north-east Papuan coast from Lae to Tufi (Fig. 3). Several recording stations (Salamaua, Kui, Morobe) along this line are closely associated with outcrops of Papuan Ultramafic Belt rocks. This line was chosen so that structures normal to the geological strike would have their least effect on the interpretation and water depth corrections to the seismic travel times would be small (water replacement velocity $= 4 \cdot 0\ \mathrm{km\,s^{-1}}$).

Figs 4 and 5 are reduced travel-time plots of the first-arrival data recorded at the seven recording stations along the north-east coast; Lae (LAT), Salamaua (SL), Kui (KU), Morobe (ME), Cape Ward Hunt (CW), Killerton (KL), and Tufi (TI). In a few instances, subsequent seismic arrivals are plotted where they are important for this interpretation. Examples of records from Salamaua are shown in Fig. 6. The travel-time data have been plotted according to the approximation that the shots and stations can be considered as reversed profiles between Lae and Tufi, an approximation which

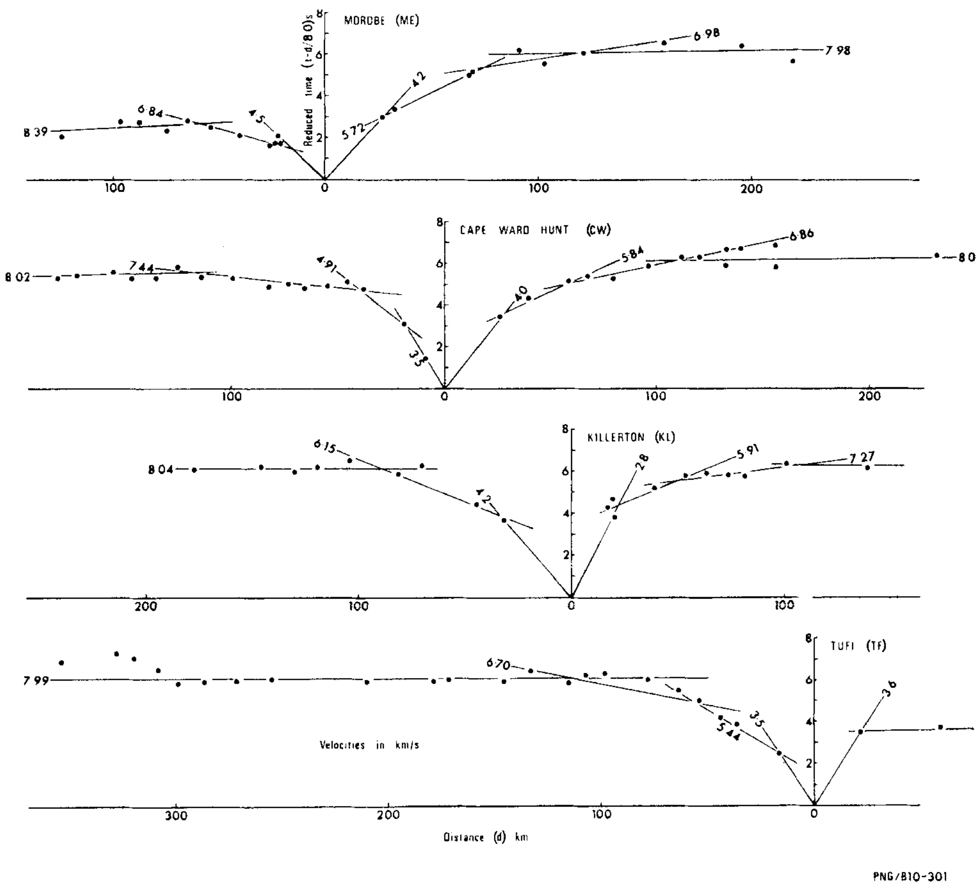

FIG. 5. Reduced travel-time plots for recording stations Morobe, Cape Ward Hunt, Killerton and Tufi from shots along the north-east Papuan peninsula coast.

is reasonable except at the shorter shot-to-station distances. Least-squares straight-line fits were applied to sets of data points where it was thought that refracting horizons had been identified. The data were weighted according to a subjective 'quality' attributed to each arrival when the records were read; the weights were 5, 3, or 1 according to whether the arrivals were judged good, fair, or poor (Finlayson, in preparation). The least-squares parameters are listed in Table 1 and the velocities so determined are indicated in Figs 4 and 5. An estimate of the velocity in the uppermost layer is obtained by constraining lines through the datum points from the nearest shots to the origin.

The travel-time plots indicate a difference in character between the data plotted north-west of Morobe towards Lae and those plotted south-east of Morobe towards Tufi. This major difference is attributed to changes in crustal structure in the Morobe region, where the shooting line crosses the boundary between the Papuan Ultramafic Belt and the Trobriand Platform.

A prominent intracrustal refractor with a velocity of approximately 7 km s^{-1} is identified between Lae and Morobe (Fig. 4); the small intercepts associated with this refractor indicate its relatively shallow depth. The cross-over distance between the 7 km s^{-1} refractor and the travel-time branch associated with arrivals directly recorded through the near-surface layers is, in most cases, between 20 and 25 km. From the data available an average velocity for these layers is 4·4 km s^{-1}.

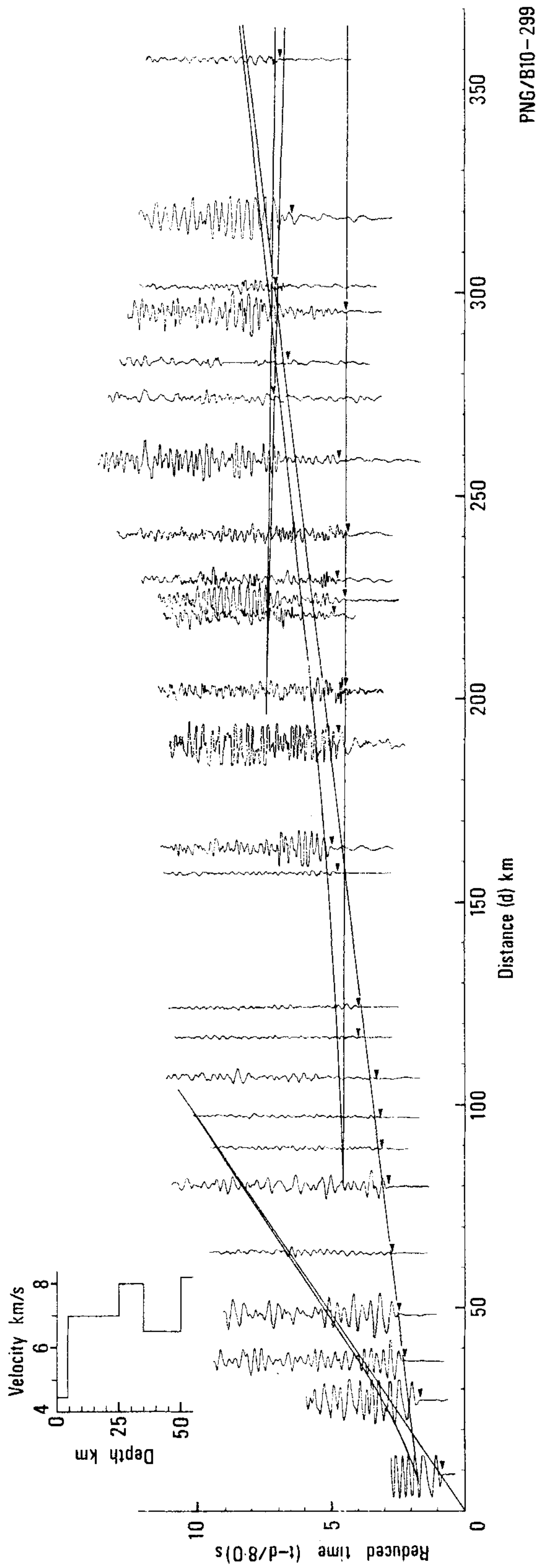

FIG. 6. Un-normalized record sections at Salamaua from shots south-east along the Papuan peninsula coast, *together* with the travel-time curve for the PUB-1 crustal model.

Table 1

Least-squares velocities and intercepts from stations along the north-east Papuan coast

Station	Direction	Velocity ($km\ s^{-1}$)	Intercept (s)	No. of data	Rms residual (s)
Lae	SE	$6·96 \pm 0·09$	$1·99 \pm 0·14$	10	0·21
Salamaua	SE	$5·80 \pm 0·02$	$0·33 \pm 0·04$	5	0·04
		$6·83 \pm 0·03$	$1·27 \pm 0·10$	16	0·21
		$8·17 \pm 0·09$	$5·21 \pm 0·30$	10	0·17
		$8·15 \pm 0·18$	$7·52 \pm 0·83$	6	0·22
Kui	NW	$7·29 \pm 0·18$	$1·24 \pm 0·16$	7	0·17
	SE	$6·22 \pm 0·04$	$1·24 \pm 0·09$	11	0·13
		$7·91 \pm 0·12$	$4·83 \pm 0·33$	7	0·15
		$8·05 \pm 0·25$	$7·22 \pm 0·88$	5	0·31
Morobe	NW	$6·84 \pm 0·22$	$1·31 \pm 0·18$	7	0·17
		$8·39 \pm 0·37$	$3·00 \pm 0·46$	6	0·24
	SE	$5·72 \pm 0·08$	$1·61 \pm 0·14$	5	0·10
		$6·98 \pm 0·37$	$3·91 \pm 0·73$	6	0·33
		$7·98 \pm 0·26$	$5·96 \pm 0·52$	6	0·38
Cape Ward Hunt	NW	$4·91 \pm 0·20$	$1·60 \pm 0·30$	3	0·09
		$7·44 \pm 0·14$	$4·38 \pm 0·21$	9	0·19
		$8·02 \pm 0·20$	$5·47 \pm 0·44$	8	0·20
	SE	$5·84 \pm 0·19$	$2·36 \pm 0·28$	4	0·12
		$6·86 \pm 0·07$	$3·75 \pm 0·16$	10	0·16
		$8·00 \pm 0·35$	$6·20 \pm 0·80$	5	0·35
Killerton	NW	$6·15 \pm 0·10$	$2·69 \pm 0·21$	4	0·13
		$8·04 \pm 0·17$	$6·24 \pm 0·31$	7	0·20
	SE	$5·91 \pm 0·41$	$3·40 \pm 0·39$	5	0·29
		$7·27 \pm 0·19$	$4·91 \pm 0·26$	7	0·24
Tufi	NW	$5·44 \pm 0·12$	$1·58 \pm 0·22$	6	0·16
		$6·70 \pm 0·26$	$3·64 \pm 0·48$	8	0·39
		$7·99 \pm 0·05$	$6·03 \pm 0·15$	14	0·22

Between Morobe and Tufi the 7 $km\ s^{-1}$ refractor is not such an obvious feature on the travel-time plots (Fig. 5) but can still be seen in the range 50–100 km with arrival times affected considerably by the sedimentary structure of the Trobriand Platform. At distances between 20 and 50 km a refractor can be identified with apparent velocities ranging between 4·9 and 6·1 $km\ s^{-1}$ and having a mean of 5·66 $km\ s^{-1}$. The shots nearest to the recording stations indicate a mean apparent velocity of 3·7 $km\ s^{-1}$ but at Killerton arrivals with an apparent velocity of 2·8 $km\ s^{-1}$ are recorded from shots over a sediment sequence greater than 3 km thick, and this velocity is taken to be the velocity of direct recorded waves. The 3·7 $km\ s^{-1}$ mean velocity probably corresponds to the minimum velocity in acoustic basement, which may be correlated with magnetic basement determined by CGG (1971) for the Trobriand Platform.

Apparent velocities of the order of 8 $km\ s^{-1}$ are not seen as first arrivals on the time–distance plots at distances less than 100 km except at Morobe, where data distribution is not good. The reliable data from all stations indicate that apparent velocities are in the range 7·91–8·17 $km\ s^{-1}$. An interesting feature of the arrivals at Salamaua and Kui is that there appears to exist a later branch of the travel-time plot with an apparent velocity slightly greater than 8 $km\ s^{-1}$, the delay being about 2·0–2·5 s (Fig. 4). The increased amplitude of the later arrivals would suggest that they may be wide-angle reflections; a decrease in the energy of the first refracted arrivals is also apparent. Thus a retrograde branch of the travel-time curve may be inferred which leads to a subsequent, higher-velocity, refracting branch.

Structural interpretation

The arrival data plotted in Figs 4 and 5 have been interpreted by using routine time-depth methods on as many pseudo-reversed profiles of shots and stations as possible. The approximations associated with the assumed coincidence of shot and stations 10–15 km apart at each end of the reversed profiles became evident when reciprocal times could not be matched closely in all cases; however, the interpretation attaches greater importance to those profiles where reciprocal times matched to better than 0·3 s. The wavefront method of Schenck (1967) and the time-term method of Willmore & Bancroft (1960) were also applied to various data sets along the profile. Identification of arrivals at some stations was assisted by the compilation of seismic record sections (Fig. 6).

The sedimentary structure has been estimated from the aeromagnetic work conducted by BMR over the Papuan peninsula (CGG 1969, 1971, 1973) and the BMR marine sparker traversing conducted along the north-east Papuan Shelf (von der Borch 1972) and Trobriand Platform (Tilbury, 1975).

The shots between Lae and Morobe were fired offshore from the shelf area, which extends seaward for between 11 and 16 km. Acoustic basement offshore was identified with Papuan Ultramafic Belt rocks, which appear to have been stable and slowly subsiding since the early Pliocene (von der Borch, 1972). The shelf break occurs at between 109 and 117 m depth and sediments are present on the 1:6 gradient slope with a two-way reflection time of at most 0·25 s, i.e. a thickness of approximately 0·2–0·4 km. Aeromagnetic interpretation of the slope seaward of the Papuan Ultramafic Belt indicates that magnetic basement follows the seabed depths closely and no great thickness of sediment exists (CGG 1969).

South-east of Morobe magnetic basement depth is estimated at 1–2 km to beyond Cape Ward Hunt, then increasing to about 3 km in the Killerton region and decreasing again towards Tufi (Fig. 7). Magnetic basement is associated with the basaltic component of the ophiolite suite of rocks at Morobe and with the possible extension of this component under the Trobriand Platform (Finlayson *et al.* 1976). The 2·8 km s^{-1} seismic velocity measured at Killerton from the nearest shots is taken to indicate the velocity of the sedimentary sequence.

The 4·4 km s^{-1} velocity indicated for the uppermost layer between Salamaua and Morobe is associated with the near-surface rocks of the Papuan Ultramafic Belt, which are shown to continue offshore at shallow depth (von der Borch 1972). The 3·7 km s^{-1} velocity indicated from shots close to most of the stations south-east of Morobe is less than that for the near-surface ophiolite suite rocks and may indicate the sequence of Pliocene to Recent volcanics associated with the Trobriand Platform (Smith 1973). It may in fact not represent a single rock type but rather a series of rocks which are undistinguished by magnetic interpretation methods (CGG 1971).

The 5·66 km s^{-1} horizon is evident at depths of 3–7 km on the Trobriand Platform. The widely ranging apparent velocities determined on the travel-time plots may be associated with a layer of non-uniform velocity resulting from the rapid decrease in compressibility caused by the closure of cracks and grain boundaries at 1–2 kb pressures. Manghnani & Woollard (1968) have demonstrated the wide range of velocities apparent in laboratory tests on basalts at these pressures, and it is probable that the 5·66 km s^{-1} layer under the Trobriand Platform shallows in the Morobe region and becomes the surface layer with velocity 4·4 km s^{-1} between Morobe and Salamaua. It is also possible that a thin 5·66 km s^{-1} layer is not detected from first arrival data in that region.

The refractor with velocity 6·8–7·0 km s^{-1} is identified at depths of 3–6 km between Salamaua and Morobe, the shallower depths being interpreted in the Salamaua region (Fig. 7). The refractor deepens markedly south-east of Morobe under the Trobriand Platform. Depths in the Cape Ward Hunt to Killerton region are

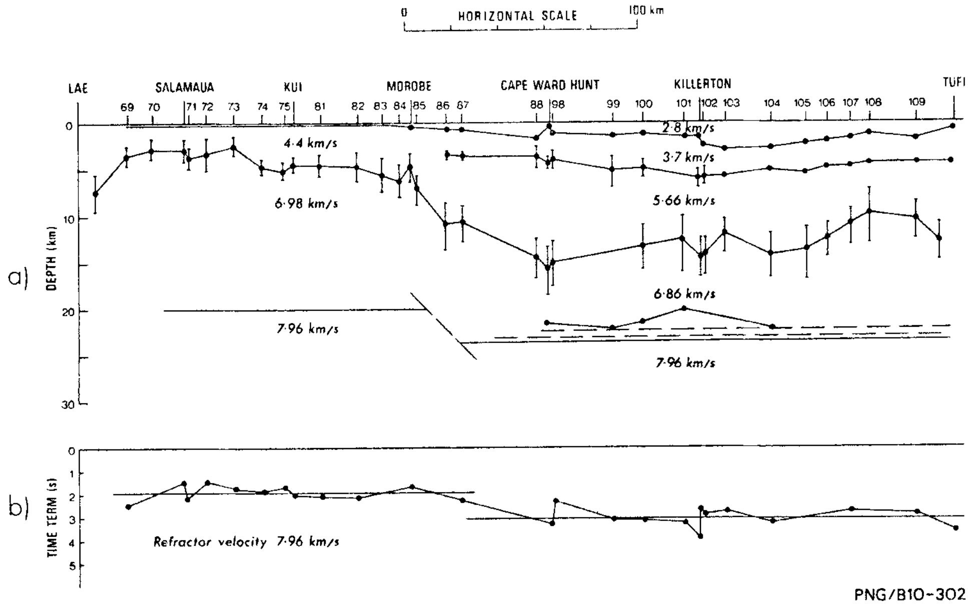

FIG. 7. (a) Crustal structure along the Lae-Tufi profile. (b) Time-terms to the $7 \cdot 96 \text{ km s}^{-1}$ refractor.

about 13–15 km, shallowing into the range 10–13 km towards Tufi. The bars on the plotted datum points in Fig. 7 indicate the extremities of depths interpreted from various data sets using different interpretation methods.

The depth to the 8 km s^{-1} refractor is more difficult to interpret because of the greater distances required to observe refractions as first arrivals and the crustal structure changes which occur half way along the profile, therefore a simple interpretation is presented at this time.

Time-depth interpretations have been made from data across the Trobriand Platform between Morobe and Tufi and the depths are indicated in Fig. 7. The good data sets from Morobe, Cape Ward Hunt, Killerton and Tufi indicate velocities in the range $7 \cdot 98$–$8 \cdot 02 \text{ km s}^{-1}$ with no significant dip. Simple layered modelling indicates an average depth of 22–23 km.

Finlayson *et al.* (1976) have shown how the depth to this refractor decreases to about 8 km along a profile between Killerton and Mount Lamington (LMG), demonstrating the gradients (approximately 19°) to be expected in this refracting horizon normal to the geological strike. Along the strike, however, it is expected that the gradients would not be as great, and a time-term interpretation was sought by constraining the refractor velocity to $7 \cdot 96 \text{ km s}^{-1}$, the mean of all apparent velocities along the profile. The resultant time-terms are plotted in Fig. 7 and it is evident that the average for the Salamaua–Morobe region (1·92 a) was 1·14 s smaller than the average for the Trobriand Platform (3·06 s). Computation of the average depths indicates that under the north-west part of the profile the refractor is at 20·0 km and under the Trobriand Platform at 23·7 km.

The later arrivals evident on some of the seismic traces can be used as additional evidence for the interpretation along the profile. Simple ray-tracing methods were applied to layered models to try to account for the prominent later arrivals at Salamaua (Fig. 6) from shots beyond 200 km mentioned earlier in this paper. The characteristics of seismic traces from various velocity-depth profiles have been described by many

Table 2

Seismic velocities (km s^{-1}) *from the Papuan Ultramafic Belt and Cyprus*

Formation	Outcrop	Cyprus in formation	Laboratory 0–2 kbar	At sea	Oceanic crust	Along profile	Papuan Ultramafic Belt	Laboratory
Upper sediments	—	—	—	2·1	Layer 1	2·8		
Pillow lavas	2·9	3·3	2·9–4·3	3·3	Layer 2	3·7, 4·4		
	3·7	—	—	4·4–4·5	—	—		
Basal group	4·8	5·1	—	4·4	—	—		
Diabase	4·9–5·1	6·4	4·9–5·7	5·5	Layer 3	5·66		
Gabbro	5·5	—	6·6–7·0	6·5	—	6·86–6·98	Gabbro	7·44–7·45 2–3 kbar
					Upper mantle	7·91–8·17	Cumulate ultramafics	7·37–8·06 3 kbar
							Non-cumulate ultramafics	8·01–8·17 3–4 kbar
	Lort & Gray 1974					East Papua Crustal Survey		Kroenke *et al.* (1974)

authors (Fuchs & Müller 1971). To achieve the characteristics of the profile recorded at Salamaua a deeper refractor was introduced with a velocity of about $8 \cdot 2 \mathrm{~km~s}^{-1}$ and a low-velocity zone above it, not only to account for the substantial delay in the arrival of reflections but also to approximate the relative amplitude features. The PUB–1 model and travel-time curve superimposed on Fig. 6 is one model which can be used.

Discussion

Berry (1971), among others, has drawn attention to the uncertainties in interpretation of first-arrival seismic data and indicated that the depths to a refractor for some velocity-depth models consistent with the travel-time plots may be 20 per cent greater than those from the simplest consistent model. No attempt therefore has been made to relate the statistical uncertainties indicated in Table 1 to structural uncertainties; the errors should be taken merely as an indication of the fit of the datum points to the least-squares straight lines.

The velocities along the Lae-Tufi profile can be compared with laboratory velocity measurements on Papuan Ultramafic Belt rocks by Kroenke *et al.* (1974), Their results, together with results from the Cyprus ophiolite complex (Lort & Gray 1974), are listed in Table 2. The Cyprus velocity data are correlated with oceanic crustal structure of Moores & Vine (1971) and the velocities found in eastern Papua are similar to those data. However, the layer thicknesses found in ocean basins are commonly 1–2 km for layer 2 and 4–6 km for layer 3, giving a total thickness of solid crust of about 8 km (Drake & Nafe 1968) compared with a thickness greater than 20 km along the Lae-Tufi profi.e. The 22–23 km depth determined off the Killerton coast is in broad agreement with Milsom's (1975) coastal value of 25 km based on gravity data. Further seismic analysis is required, however, to determine the detailed structure of the ophiolite sequence over the whole of the region. The total crustal thickness along the Lae-Tufi profile is considerably greater than 10 km, which is thickness of the basalt and gabbro components of the ophiolite suite exposed at the surface farther inland.

Finlayson *et al.* (1976) have drawn attention to the structures below the Moho demonstrated by the PUB–1 model fitted to Salamaua data (Fig. 6). Similar seismic features can be seen in the recordings beyond 150 km (Fig. 3) at Musa (MU), Cape Rodney (CR), Kupiano (KP), and Kwikila (KW) from shots along the north-east coast. The time-delay and amplitude data associated with deeper refractors seem to be satisfied only by introducing a low-velocity zone of the proportions of that in the PUB–1 model. It may be speculated that the material for such a low-velocity zone is derived from the Mesozoic north-east margin of the Australian continent, but this will clearly require further investigation.

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