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CRUSTAL STRUCTURE OF THE SOUTHERN McARTHUR BASIN, NORTHERN AUSTRALIA, FROM DEEP SEISMIC SOUNDING

C.D.N. Collins

Deep seismic refraction and vertical reflection recordings have been made in the southern McArthur Basin, over the Bauhinia Shelf and Batten Trough between Daly Waters and the H.Y.C. mineral deposit, and over the Wearyan Shelf between Borroloola and Westmoreland. In the Batten Trough, McArthur Group rocks have a velocity of 5.81 km/s; no velocity contrast was detected between them and basement. Over the Bauhinia Shelf, 100–200 m of Cainozoic, Mesozoic, and perhaps Cambrian sediments, overlie ?Roper Group sediments (P-wave velocity, V = 4.6 km/s). Below these are probable Tawallah Group rocks (V = 5.8–5.9 km/s). Magnetotelluric measurements define a resistivity contrast, possibly basement, at 6–9 km depth.
On the Wearyan Shelf at Borroloola, 370 m of Roper Group (V = 3.58 km/s) overlies 2.9 km of Tawallah Group (V = 5.55 km/s). At Robinson River, 650 m of Cainozoic, thin McArthur Group, and, perhaps, upper units of Tawallah Group (V = 4.81 km/s) were detected. Tawallah Group rocks (V = 5.44 km/s) crop out northwest of Robinson River and are about 2.8 km thick. Basement velocity is 6.04 km/s. Between Robinson River and Westmoreland, basement is 3.5–2.7 km deep. At Westmoreland the McArthur Basin sequence thins against the Murphy Ridge. A layer 260 m thick (V = 3.50 km/s) lies on top of a 2.4 km thick layer (V = 5.44 km/s), and the basement velocity of 5.99 km/s increases to 6.06 km/s towards Robinson River. At mid-crustal depths velocities are 5.9–6.9 km/s, and in the lower crust, to depths of 43 km in the west and 40 km in the east, 6.8–7.5 km/s. Below this, a velocity gradient is interpreted until upper mantle velocities are reached at 43–53 km depth in the west (V = 7.5–8.4 km/s) and 44 km in the east (V = 7.9 km/s). Generally, the crustal structure of the North Australian Craton is characterised by high lower-crustal velocities, broad velocity gradients, and thick crust, which probably evolved from an Archaean continental crust during Proterozoic tectonism.

Introduction

The Proterozoic McArthur Basin covers 170 000 km^2 in the Northern Territory (Fig. 1). It contains a number of important mineral deposits and has potential for further discoveries. Its geology has been described by Plumb & Derrick (1975), Plumb (1977), and Plumb & others (1980).

The structure of the southern McArthur Basin has been investigated by several geophysical methods, namely, detailed gravity, magnetotellurics, aeromagnetics, and deep seismic reflection and refraction techniques (Plumb, 1977). This paper reports, principally, the structure interpreted from the seismic methods.

The main features of the southern McArthur Basin in the survey area (Fig. 1) are the Batten Trough, about 60 km wide, and its flanking Bauhinia Shelf on the west and Wearyan Shelf on the east. The Emu Fault is considered to define the eastern boundary of the Batten Trough. One objective of the seismic investigation was to define any major velocity differences between the Batten Trough and the Bauhinia and Wearyan

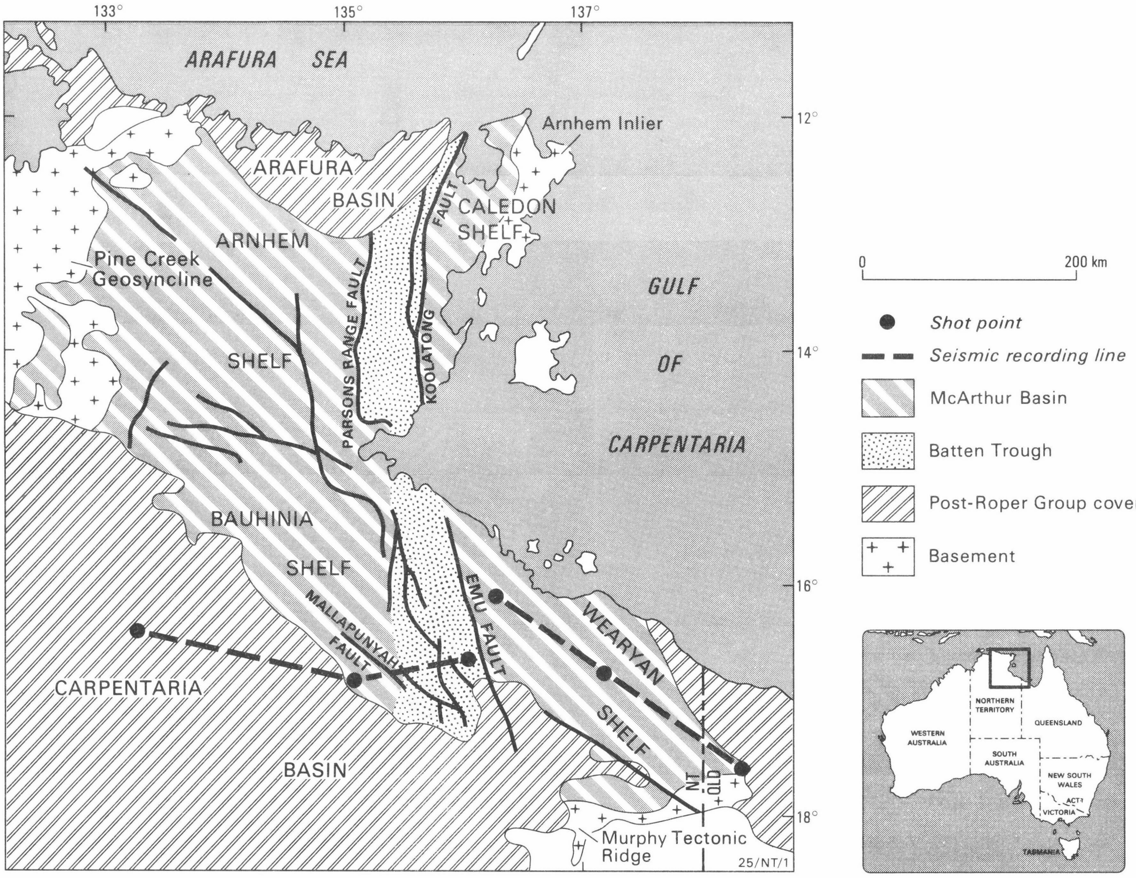

Figure 1. Major tectonic elements of the McArthur Basin (after Plumb & Derrick, 1975).

Shelves. A possible difference in crustal structure is suggested by the distribution of flood basalts, which are abundant on the shelves and very restricted in the trough.

The major faults that bound the Batten Trough may have acted as channels for ore solutions, and important geological questions that need to be answered in relation to these faults are whether they predate the McArthur Basin succession and whether the thickness of rock units changes abruptly at the faults or gradually over tens of kilometres. In particular, the Emu Fault was investigated because of its probable relation to the H.Y.C. lead-zinc deposit (Fig. 2). Other objectives of the seismic investigations were to delineate the basement of the McArthur Basin, to derive velocities and structures beneath extensive Mesozoic–Cainozoic cover above the Bauhinia Shelf, and to identify any marker horizons.

The McArthur Basin succession was deposited about 1700–1400 m.y. ago, is up to 12 km thick, and comprises the Carpentarian Tawallah, McArthur, and Roper Groups. The Tawallah Group is up to 6 km thick, and consists of sandstones and carbonates; the McArthur Group, up to 5.5 km thick, comprises predominantly carbonate rocks; and the Roper Group comprises mainly sandstones and lutites up to 5 km thick. The sediments are predominantly shallow water and often remarkably uniform over large areas.

The north-trending Batten Fault Zone, 50–60 km wide, runs down the axis of the basin and contains a very much thicker succession than those of the mildly deformed Arnhem, Caledon, Bauhinia, and Wearyan Shelves on either side (Fig. 1). It has been postulated that the fault zone corresponds to a syndepositional graben, the Batten Trough, within which the thick sequences were confined. In the survey area, the Emu Fault marks the boundary between intense faulting to the west and flat-lying sequences to the east. Since the fault was demonstrably active during at least part of McArthur Group deposition, it has been defined as the eastern boundary of the Batten Trough. The western boundary is gradational (Plumb & Derrick, 1975; Plumb & others, 1980). Later, during the deposition of the Roper Group, the site of maximum sedimentation rate shifted westwards onto the Bauhinia Shelf.

During post-depositional deformation the sense of the faulting has been reversed and the Batten Trough has become a horst. Deformation was more intense in the Batten Fault Zone than on the surrounding shelves, and was related to block faulting along pre-existing basement faults. Although vertical stratigraphic displacements of up to 7.5 km may be seen along the major faults, it has been interpreted that the overall structural development of the basin was controlled by right-lateral horizontal displacements along the north-trending Batten Fault Zone and left-lateral displacements along the northwest-trending faults (Plumb & others, 1980). Deformation of the McArthur Basin ceased before the Cambrian.

Previous geophysical work

Most geophysical surveys carried out in the area have been in the vicinity of individual mineral deposits, particularly the H.Y.C. (Plumb, 1977). Previous deep seismic surveys have been conducted west and south of the area (Hales & Rynn, 1978; Hales & others, 1980; Underwood, 1967; Finlayson, 1981, in press; Cleary, 1973; Denham & others, 1972). Regional gravity coverage extends across the McArthur Basin, and has been supplemented by detailed traverses in the survey area. Heat-flow measurements in the region have been summarised by Cull (1982b).

Field operations

The field procedures used in this study have been reported by Collins (1981), and only a brief description is given here. The crustal investigations involved recording at 71 sites along two 300-km seismic refraction traverses, one west and the other

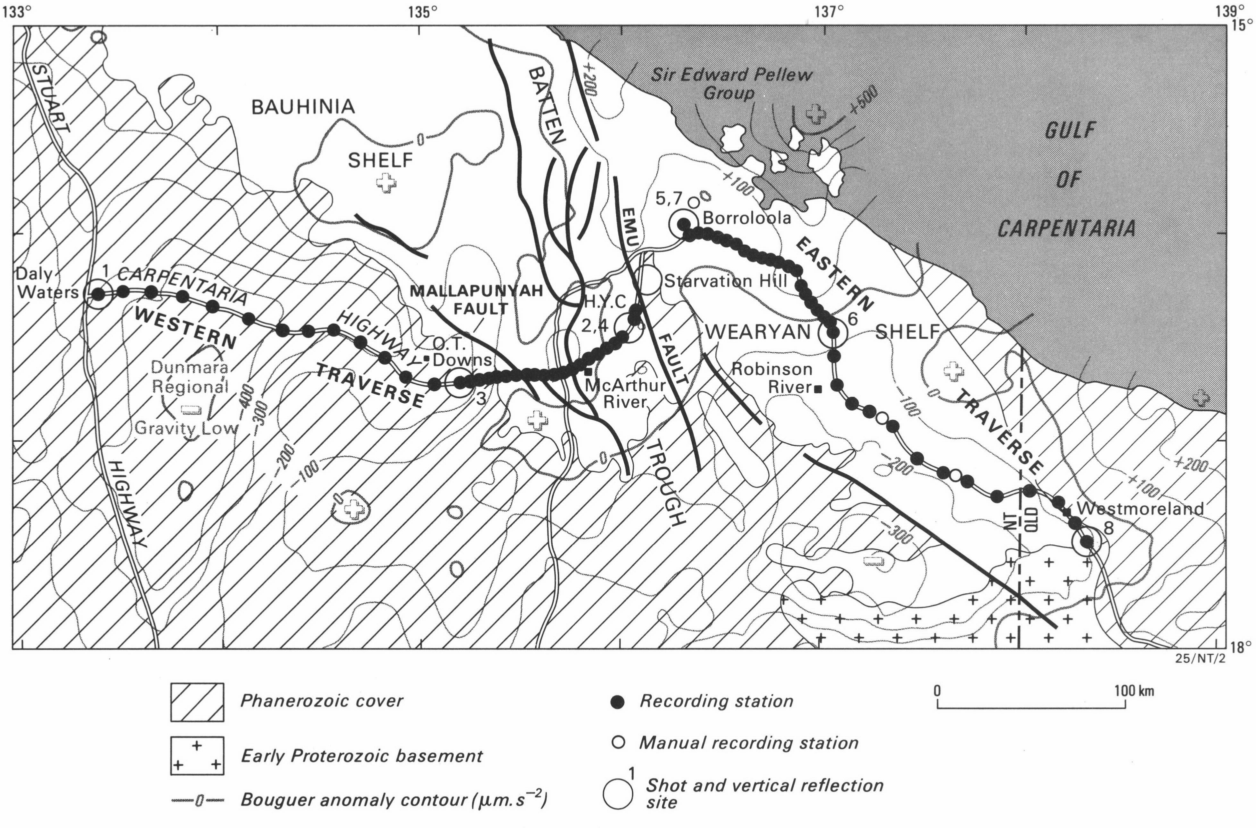

Figure 2. Location of shots and recording stations, simplified geology, and Bouguer gravity.

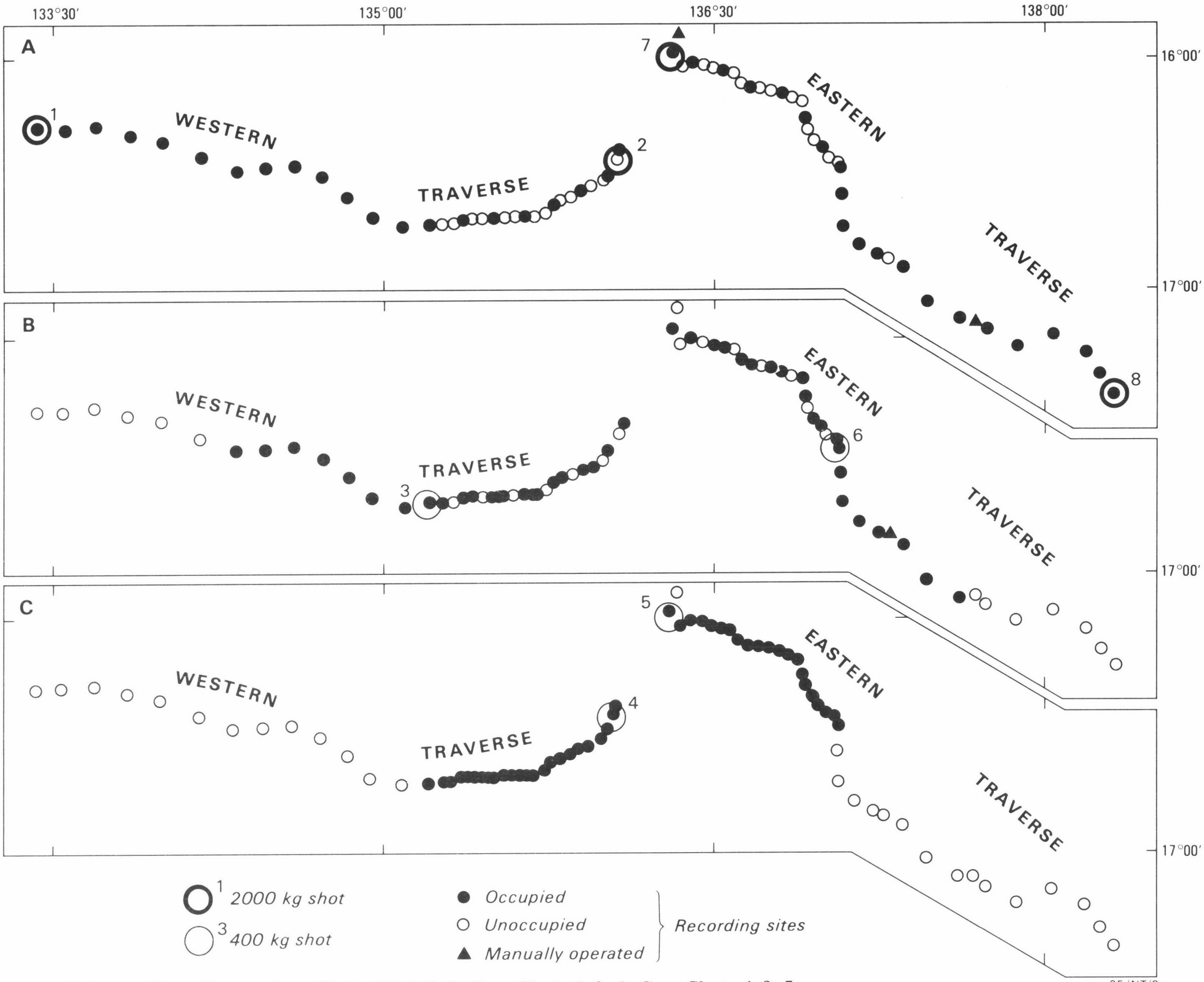

Figure 3. Recording scheme, A — Shots 1,2,7 & 8, B — Shots 3 & 6, C — Shots 4 & 5.

east of the Emu Fault, during June–July, 1979 (Fig. 2). Two large shots of 2000 kg were detonated at Daly Waters and the H.Y.C. deposit, and two smaller shots of 400 kg were detonated at H.Y.C. and 100 km west of H.Y.C., near O.T. Downs. For the larger shots, the station spacing was about 15 km, while for the small shots it was 5–10 km (Fig. 3). A similar pattern was recorded along the second traverse, east of the Emu Fault, with large shots at Borroloola and Westmoreland, and small shots at Borroloola and Robinson River. Deep reflection recordings were made at these six shot points. Figure 3 shows the recording scheme for each shot.

Recordings were made on 21 BMR automatic FM tape recording systems (Finlayson & Collins, 1980) and two manually operated recorders. Each system used a single vertical Willmore seismometer, and the amplifier gain levels, filter settings, and polarity were the same at all sites. The shots consisted of a pattern of drill holes 27 m deep, with 100kg of explosive in each hole. Small (25 kg) weathering shots were fired at each shot site, but these were recorded only to distances of about 14 km from the shot point. All shots were timed on site relative to VNG radio time signals to an accuracy of 0.02 s.

The latitudes and longitudes of the shots and stations were scaled off 1:100 000 topographic maps to within 0.05 minute of arc (approximately 100 m). Elevations ranged between 40 and 310 m on the western traverse, and 20 and 220 m on the eastern traverse. Wherever possible, the seismometer was placed on rock outcrop, but at the majority of sites this was not possible and the seismometer was buried.

Data processing and interpretation methods

The field tapes were played back and digitised at BMR, Canberra. The first-arrival times were read from the original analogue records. In most cases, the record was analogue filtered during playback, with several different passbands to enhance the arrivals against the background noise. Arrivals with impulsive onsets were read to 0.02 s. Emergent arrivals or records with low signal-to-noise ratio were read to 0.1 s or, in some cases, a greater uncertainty. The travel-times (reduced by 6.0 km/s) of all first arrivals recorded from the ends of both long traverses are plotted against distance in Figure 4.

The digitised recordings at each site were compiled into record sections. These have been digitally filtered and reduced by 6.0 km/s before plotting, and each trace has been normalised to a maximum amplitude. The long sections (0–300 km) have been filtered using a passband of 2–12 Hz, and the short sections (0–100 km), 2–15 Hz. These filter limits were chosen because they provide the best improvement in signal-to-noise ratio on most records. Individual records can often be improved further by applying different filter limits. In some cases, the first arrivals cannot be clearly seen unless a narrow passband filter is applied, and they may not be obvious on the illustrated record-sections.

The apparent velocities of arrivals from each shot were derived by linear regression, with the assumption that the travel-time curves are made up of linear segments. This may not be true for all parts of the curves, but it gives a first approximation to the velocity structure.

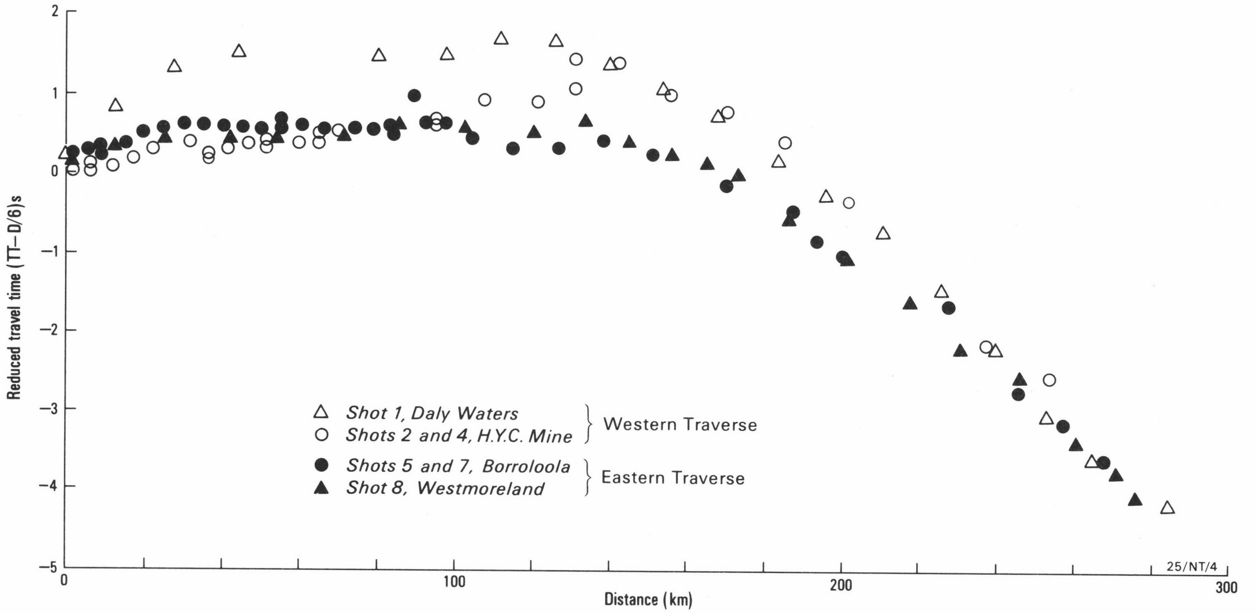

Figure 4. Reduced first-arrival travel-times.

Initial interpretation was made assuming horizontal, homogeneous, plane layering at each shot point. The models derived in this way were then tested by ray-tracing methods, and modified so that the calculated and observed travel-times agreed. They were further modified after comparison of amplitude characteristics of the record sections with synthetic seismograms calculated by the reflectivity method (Fuchs,1968).

Where the shallow layers showed a marked variation in structure, a two-dimensional gravity modelling program was used to calculate the gravity field. This was compared with the observed Bouguer gravity data, and minor variations consistent with both the seismic and gravity were made, mainly for those parts of the section for which no seismic data exist. Detailed gravity observations were made along the traverses during 1978–1980 and these have been interpreted by Anfiloff (BMR, personal communication, 1982). A ray-tracing program, capable of handling irregular structures, was used to test travel-times calculated from the model against observed arrivals where there were marked structural changes along the traverse. The synthetic seismogram program used in this interpretation can compute seismograms only for planar horizontally layered models, and therefore could not be properly used in these cases. Where possible, the model was averaged to fulfil the requirements of the synthetic seismogram program, and thus provide some information on expected amplitudes. Velocities have been corrected for Earth curvature (Mereu, 1967), but this is only significant for the lower crust and upper mantle (about 0.8 per cent at 50 km depth).

Results

Some of the major features of the seismic data can be seen in the plot of first-arrival travel-times (Fig. 4). Considering, firstly, only the data from the western traverse between Daly Waters and the H.Y.C. deposit, it can be seen that there are marked differences in the travel-times at distances less than 100 km. Shot 1 arrivals are about 1 s later than arrivals from Shots 2 and 4. Beyond 100 km, the arrivals from Shot 2 become increasingly delayed, and, from 140 km to the ends of the traverse, the travel-times from Shots 1 and 2 are substantially the same, with apparent velocity increasing with distance.

In the range 40–110 km, the arrivals from Shots 1 and 2 have similar apparent velocities. The true velocities may differ slightly at either end, but the difference between the travel-times is largely due to the presence of a near-surface low-velocity layer, indicated by the arrivals in the range 0–30 km. The travel-times from Shot 2 between 0 and 80 km exhibit local irregularities, possibly owing to structure, but, otherwise, only one phase velocity is recorded. Thus, there is, at most, only a very thin low-velocity surface layer near Shot 2.

On the eastern traverse, between Borroloola and Westmoreland, very little difference can be seen between the travel-times from Shots 5 and 7 and Shot 8 (Fig. 4). The only clear differences are slightly earlier arrivals from Shot 8 in the range 20–70 km; this trend is reversed between 100 and 140 km. The differences amount to, at most, 0.2 s. Data from both shots exhibit abrupt changes in apparent velocity at about 25 km and 140 km from the shots. From 0 to 140 km, the travel-time plots consist of approximate linear segments; beyond 140 km, the arrivals show a progressive increase in apparent velocity.

Shallow structure

In this paper, the shallow structure is considered to comprise the formations above the basement refractor, which has a velocity of about 5.8–6.0 km/s. This basement refractor is at less than 5 km depth and is evident beneath both traverses. A thin surface layer, of variable thickness and with a velocity of about 1.7 km/s, was found beneath all shot points and, presumably, occurs along the whole length of the traverses. Shallow refracted arrivals, recorded by the 3-km reflection spreads at each shot point, were used to derive an average thickness for this layer, which varied between 35 m at Westmoreland to 120 m at Daly Waters.

Western traverse

Geological constraints. Shots 2 and 4 (H.Y.C.) were situated in a thick faulted sequence of Carpentarian McArthur Group rocks, which crop out for about 80 km westwards along the traverse and are up to 5 km thick (Plumb, 1977; BMR, 1966). These are underlain by sandstone and volcanics of the Tawallah Group, which in turn overlie basement.

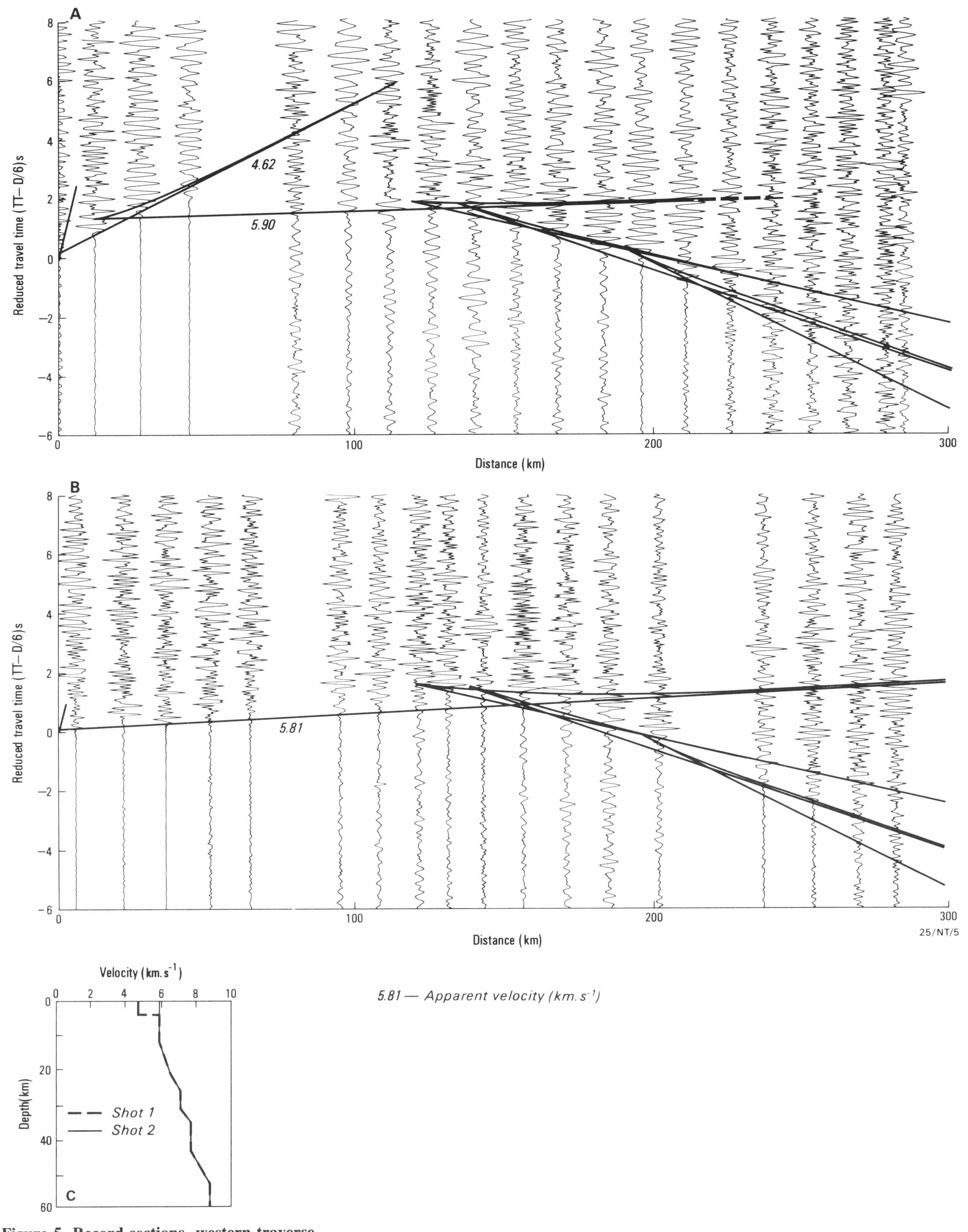

Figure 5. Record sections, western traverse.
Traces normalised and filtered 2–12 Hz. A — Shot 1, Daly Waters, B — Shot 2, H.Y.C. mine, C — velocity/depth functions.

Shots 1 (Daly Waters) and 3 (O.T. Downs) were on deeply weathered Mesozoic sediments, probably less than 100 m thick. At O.T. Downs, the sediments overlie Carpentarian sandstones and siltstones of the Roper Group, which are probably somewhat less than 2 km thick and stop about 23 km east, where their contact with the underlying McArthur Group is exposed. Westwards from Shot 3, the Mesozoic sediments completely cover the underlying rocks, but an inlier about 50 km north suggests that the Roper Group thickens westwards, perhaps to as much as 5 km, and directly overlies the Tawallah Group, owing to wedging out of the McArthur Group.

Cambrian sediments of the Daly River Basin are interpreted to underlie Daly Waters, where the total thickness of Cainozoic, Mesozoic, and Cambrian sediments has been estimated at 300 m (Brown, 1969). Elsewhere, the Cambrian Tindall Limestone alone has been shown to reach 910 m (Randal, 1973).

Seismic interpretation: The seismic record sections from all shots on the western traverse are shown in Figures 5 and 6. Below the surface layer at Daly Waters an apparent velocity of 4.62 km/s was measured. This was recorded at only two stations, so an estimate of the error cannot be made. However, it represents the average velocity of the layer, and is supported by ray tracing and gravity modelling. The estimated error limits of the other shallow velocities are shown in Figure 7. The 4.62 km/s layer is 4.1 km thick at a location 20 km east of Daly Waters; it is not directly evident at O.T. Downs (Shot 3) nor at H.Y.C. (Shots 2,4). Apart from the thin surface layer, the apparent velocity of the shallowest layer at H.Y.C. is 5.81 km/s (Fig. 6A), that of the McArthur Group carbonate rocks.

If a surface velocity of 1.70 km/s is assumed, an unusually thick surface layer of 560 m may be inferred at Shot 3.

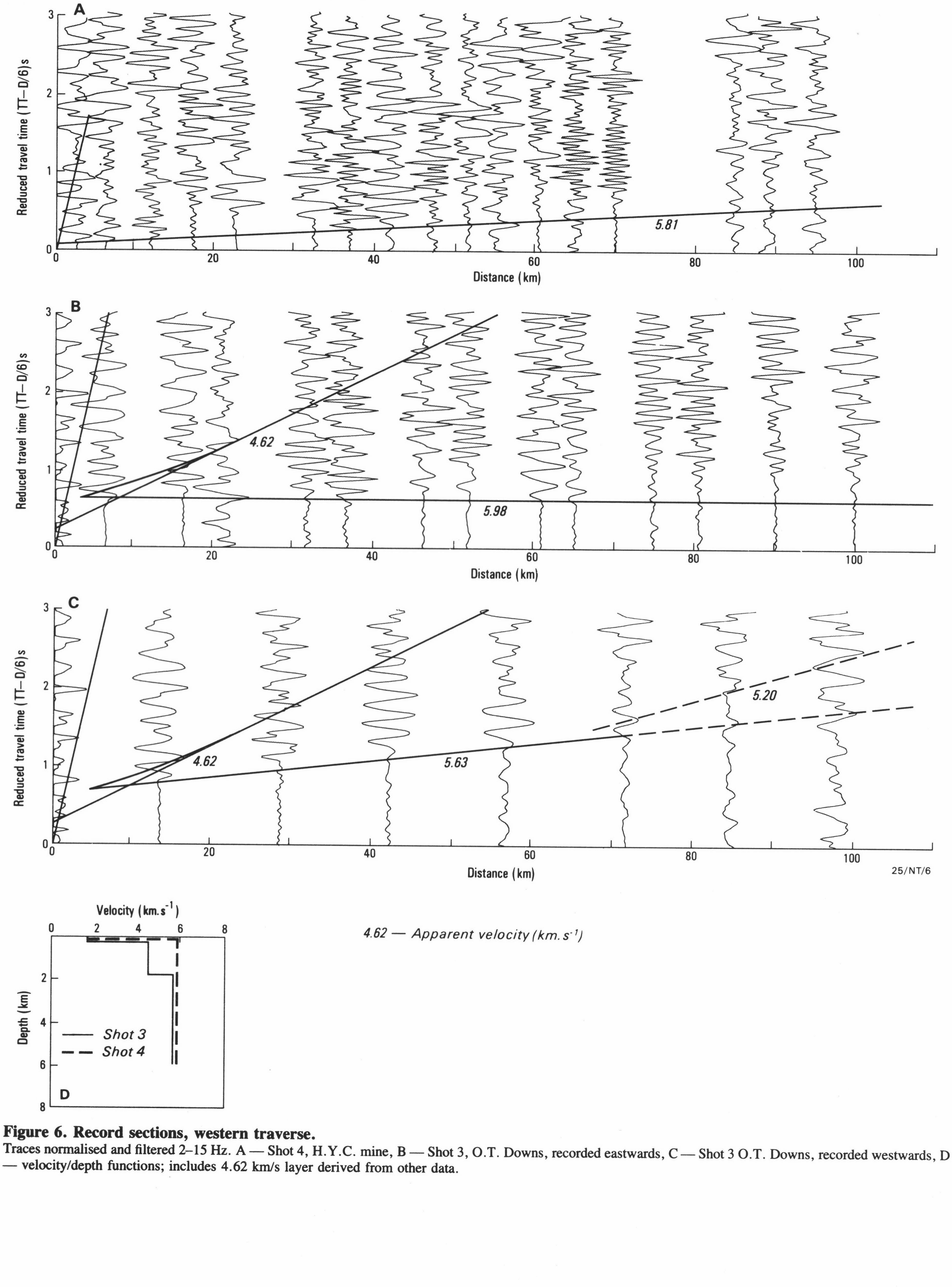

Figure 6. Record sections, western traverse.
Traces normalised and filtered 2–15 Hz. A — Shot 4, H.Y.C. mine, B — Shot 3, O.T. Downs, recorded eastwards, C — Shot 3 O.T. Downs, recorded westwards, D — velocity/depth functions; includes 4.62 km/s layer derived from other data.

However, the 4.62 km/s layer may be present under Shot 3 and may have contributed to the travel-times, because the arrivals from Shot 2 become increasingly delayed from about 23 km east of Shot 3, about the point at which outcrops of Roper Group overlying McArthur Group rocks first appear. Minimum thicknesses of the surface (1.70 km/s) layer and the 4.62 km/s layer can be calculated within the constraints of the travel-times to the nearest stations, and are 210 m and 1.4 km respectively. A vertical reflection at a depth of 1.6 km supports this interpretation. The 4.62 km/s layer is interpreted as wedging out completely 23 km east of O.T. Downs, and so correlates with the Roper Group, both in outcrop and depth below O.T. Downs.

Only the 1.70 km/s surface layer is present further east, and it thins towards H.Y.C. (Fig. 7A). Below it, the apparent refractor velocity is 5.81 km/s from Shot 4 and 5.98 km/s from Shot 3, giving a true velocity of 5.89 km/s, if a constant dip is assumed between the two shots. However, the high apparent velocity towards the east may, in part, be due to the greater thickness of low-velocity material below Shot 3, which would locally cause an increase in dip on the lower refractor; the true velocity could thus be lower. Residuals in the travel-times and the variable character of the wave-train from Shot 4 at H.Y.C. and Shot 3 at O.T. Downs (Fig. 6) are due to local structures in the McArthur Group, including a small syncline of Roper Group sediments. The data are not sufficiently detailed to map these small-scale structures, and the assumption of a homogeneous layer gives a good fit to the first-arrival travel-time data (Fig. 6).

The apparent velocity westwards from Shot 3 is 5.63 km/s, which, with the eastward apparent velocity of 5.98 km/s, gives a true velocity of 5.80 km/s. At Daly Waters (Fig. 7A) the corresponding apparent velocity is 5.90 km/s, giving a true velocity of 5.82 km/s. The data quality is poor to the west of Shot 3, probably owing to the thickening surface layers. West of Shot 3, an average dip of about 0.5° increases beyond about 60 km, where the depth to basement is 2.1 km. From extrapolation of the dips from both Shot 1 and Shot 2, the basement apparently steps downwards to the west at a point about 80 km west of Shot 3 (Fig. 7A).

The interpretation of the basement topography is supported by the gravity data along the traverse (Fig. 2). The main features in the gravity data correlate with the topography of the top of

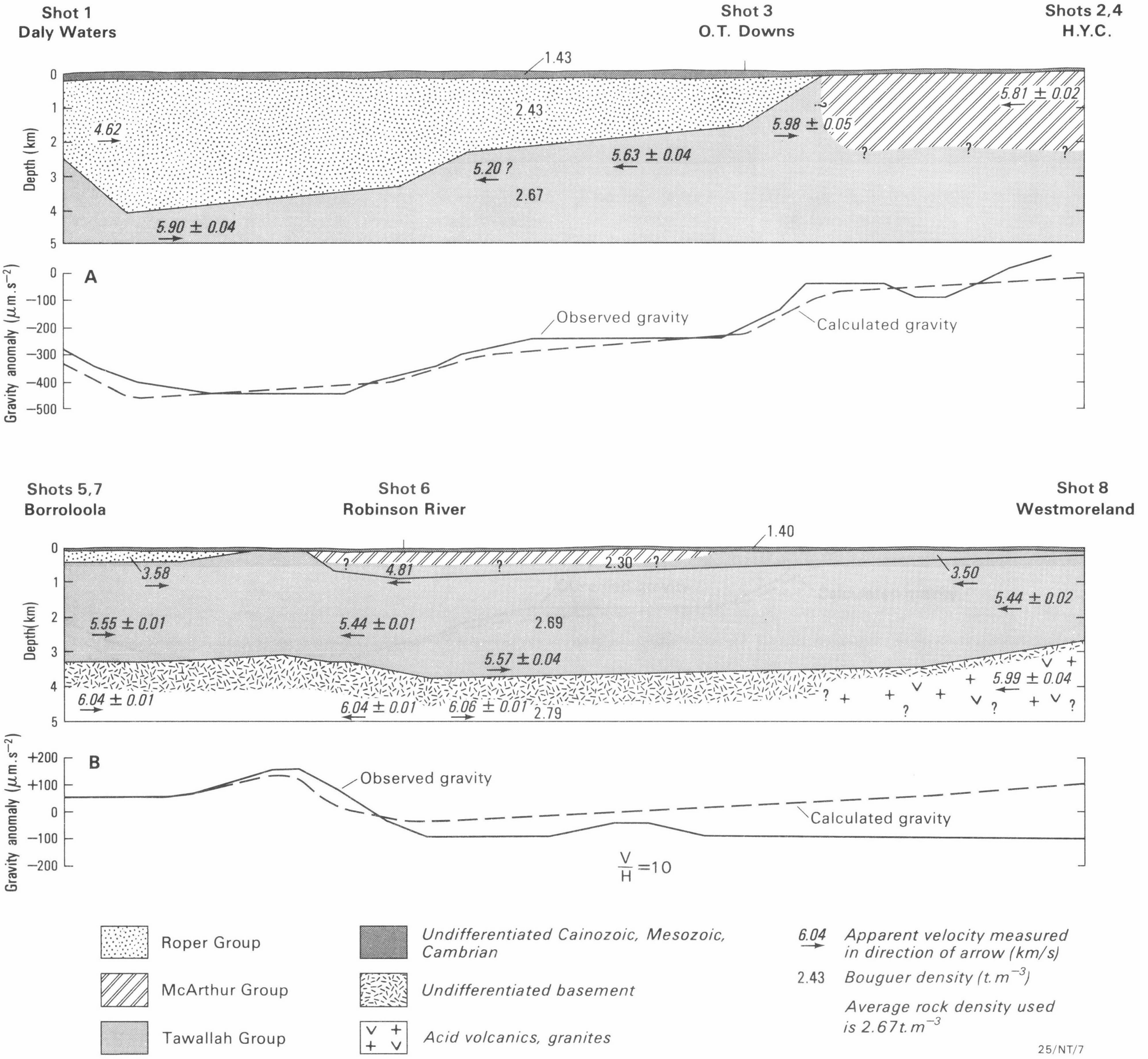

Figure 7. Seismic models and geological interpretation of the shallow structure, with observed and calculated Bouguer gravity, A — Daly Waters to H.Y.C. mine, B — Borroloola to Westmoreland.

the 5.82 km/s layer (Fig. 7A). Near Daly Waters, the gravity increases by 100 $\mu m/s^2$ over a distance of about 20 km, which could indicate basement rising to within 2.5 km of the surface at Daly Waters. This does not conflict with the seismic data, which can be interpreted only for distances greater than 20 km east of Daly Waters, owing to the offset distance. The geological data previously discussed indicate that the 4.3 km of sediments above the 5.82 km/s basement, represented by the 4.62 km/s wedge, must be a combination of about 300 m of Cainozoic, Mesozoic, and Cambrian sediments plus about 4 km of Roper Group. The Roper Group sediments, which are responsible for the Dunmara Regional Gravity Low (Fraser, 1976) (Fig. 2), must therefore extend for a considerable distance to the south.

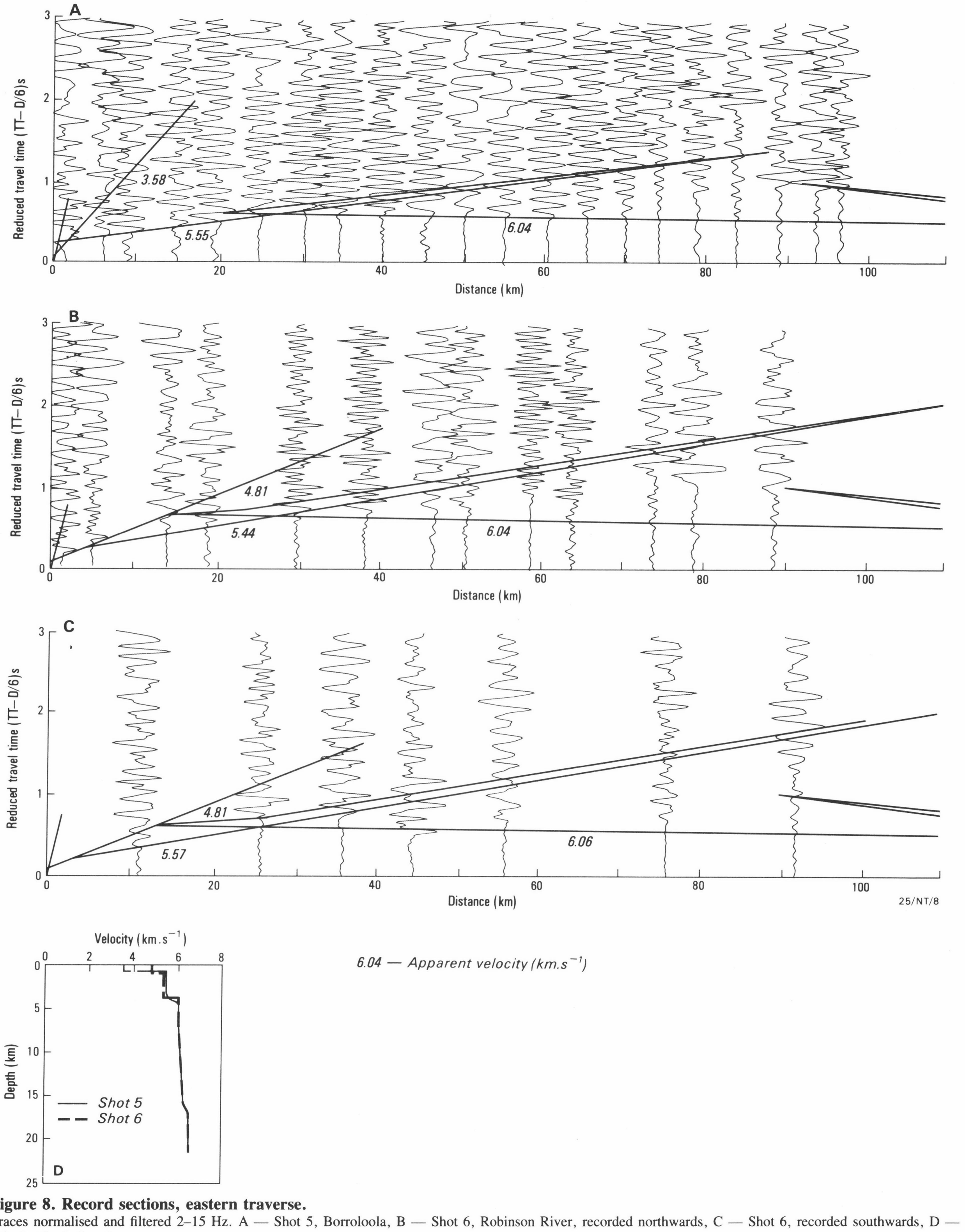

Figure 8. Record sections, eastern traverse.
Traces normalised and filtered 2–15 Hz. A — Shot 5, Borroloola, B — Shot 6, Robinson River, recorded northwards, C — Shot 6, recorded southwards, D — velocity/depth functions.

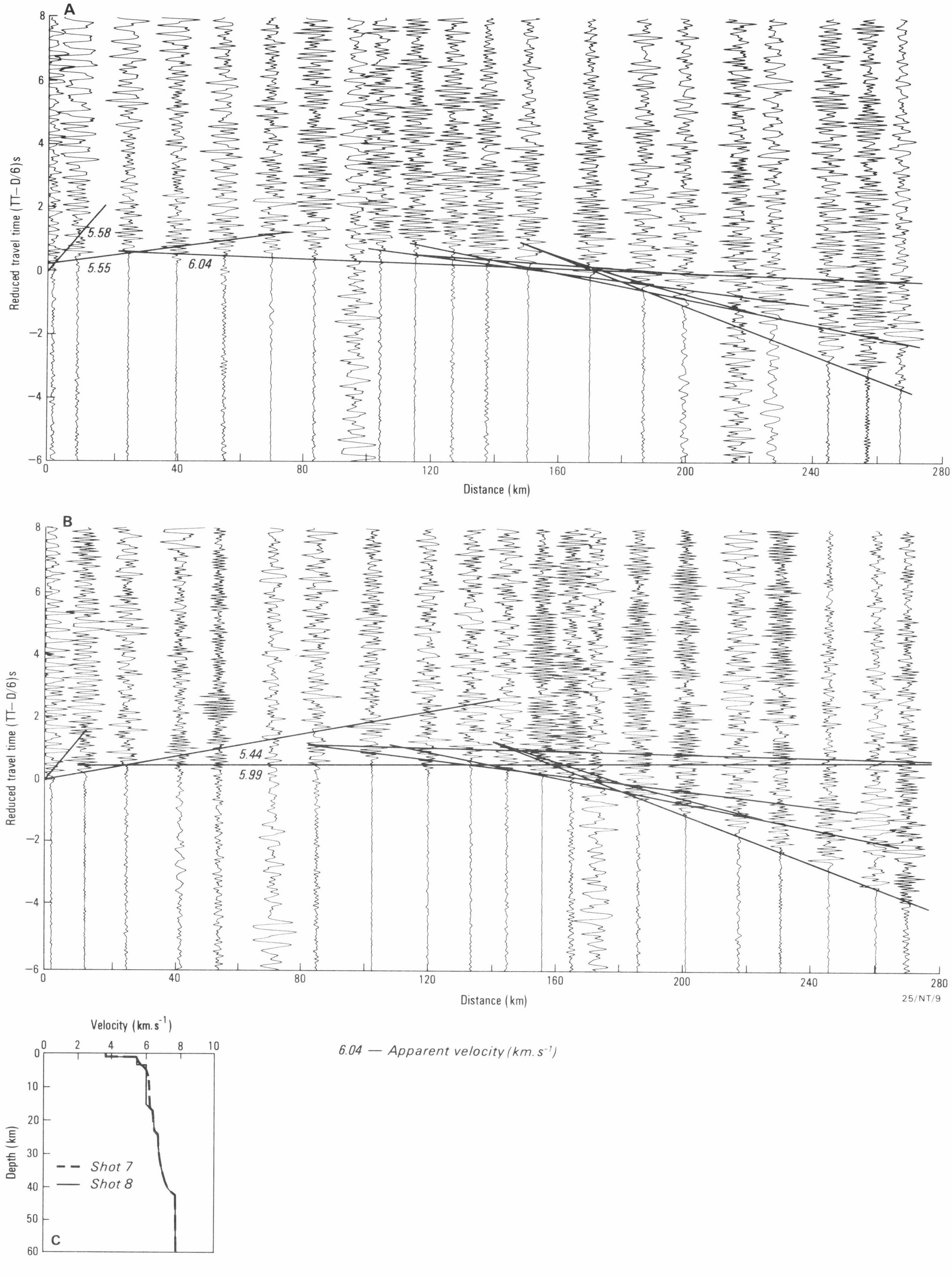

Figure 9. Record sections, eastern traverse.
Traces normalised and filtered 2–12 Hz. A — Shot 7, Borroloola, B — Shot 8, Westmoreland, C — velocity/depth functions.

Eastern traverse

Geological constraints. Shots 5 and 7 (Borroloola) were situated on Roper Group sediments east of the Emu Fault, which crop out along the traverse up to about 50 km southeast of Borroloola (Plumb, 1977; BMR, 1966). The maximum thickness of Roper Group in the area is about 500 m (Smith, 1964), and it is underlain by an unknown thickness of McArthur Group and Tawallah Group. Shot 6 (Robinson River) was situated on thin (< 100 m) McArthur Group rocks overlying Tawallah Group sediments (Yates, 1963). A basement ridge runs northwards from Robinson River homestead and probably crosses the traverse about 60 km southeast of Borroloola.

Shot 8 (Westmoreland) was situated on 200–400m of Westmoreland Conglomerate of the Tawallah Group, which overlies Cliffdale Volcanics (Grimes & Sweet, 1979). Approximately 50 km northwest of Shot 8, 3.5 km of Tawallah Group sandstones, conglomerates, and basalts of the Seigal Volcanics overlie an unknown thickness of Cliffdale Volcanics. The traverse within 50 km of Shot 8 lies over a complex area of faulting and shallowing of basement as the McArthur Basin sequence thins against the Murphy Tectonic Ridge to the south. Granites probably underlie the Cliffdale Volcanics and form part of the basement of the McArthur Basin sequence.

Seismic interpretation. Figures 8 and 9 show the seismic record sections and velocity/depth functions for the eastern traverse; the model shown in Figure 7B was interpreted. The basement arrivals from Shot 5 were slightly earlier in the range 20–50 km northwest of Shot 6, indicating a basement rise there, which coincides with the ridge indicated by the geology.

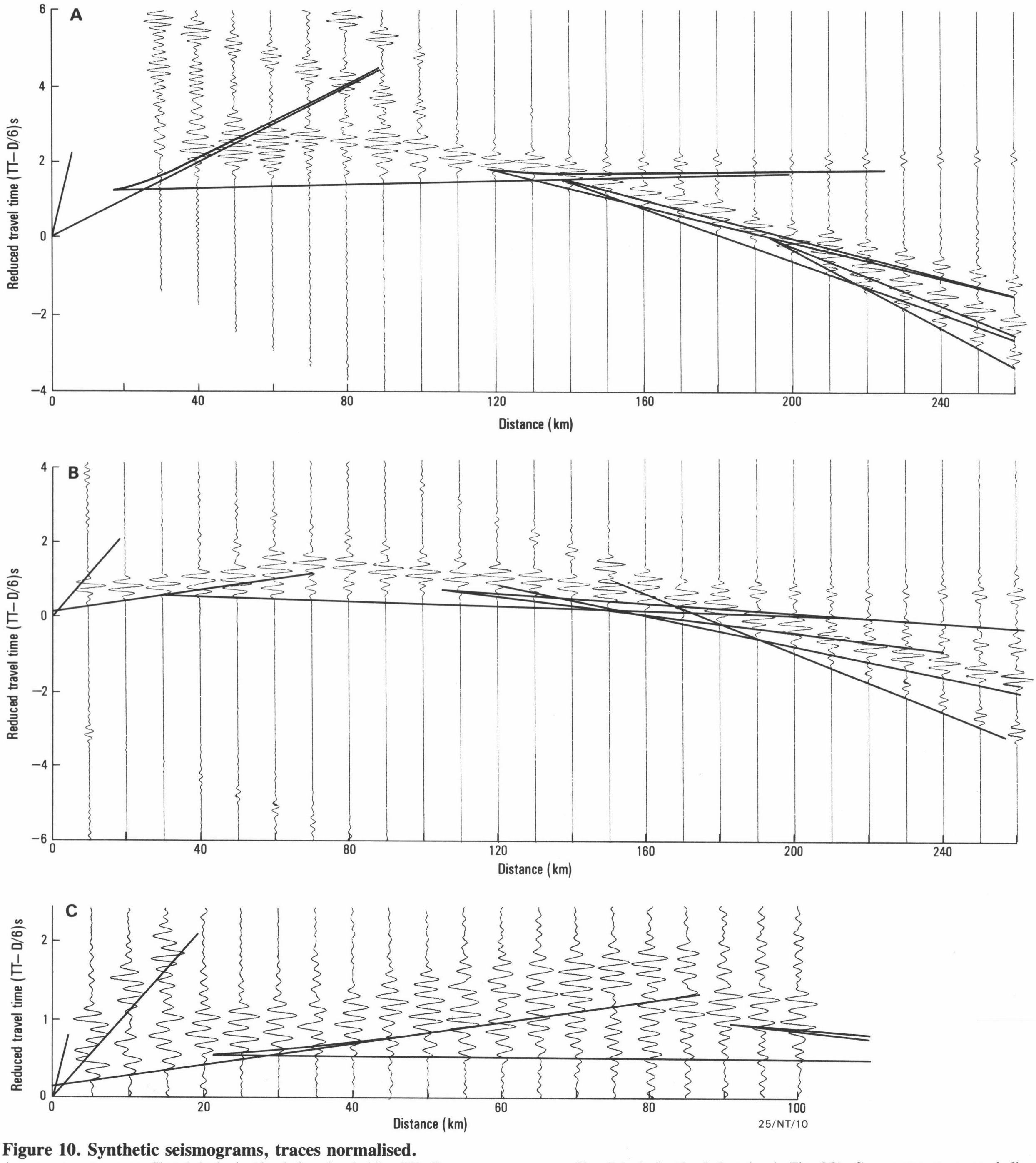

Figure 10. Synthetic seismograms, traces normalised.
A — western traverse, Shot 1 (velocity/depth function in Fig. 5C), B — eastern traverse, Shot 7 (velocity/depth function in Fig. 9C), C — eastern traverse shallow structure, Shot 5 (velocity/depth function in Fig. 8D).

Synthetic seismograms were computed for this model, using the velocity/depth function for Shot 5 (Fig. 8D); the synthetic record section is shown in Figure 10C. If the persistent wave-trains in the observed record section (Fig. 8A) are ignored, many of the details seen in the synthetics become apparent; for example the wide-angle reflection arrival at 1.4 s at 70 km. The termination of the reflection phase at about 85 km is more abrupt on the observed record section (Fig. 8A); the decrease in observed first-arrival amplitudes is also greater. This may suggest a low-velocity layer below the basement; however, modelling gave no more satisfactory results than the synthetic section shown here (Fig. 10C). The detailed shape of the velocity gradient has a marked effect on the synthetics; the shape of the gradient may even change along the traverse, but this cannot be modelled, because the structure is assumed to be laterally homogeneous. The surface layers may also vary along the traverse, and, as Khoketsu (1981) has shown, they can change the relative amplitudes of the synthetic seismograms from deeper layers.

The 3.58 km/s layer at Borroloola correlates with the thin Roper Group sediments. It was not recorded beyond the nearest station, and was interpreted from refracted arrivals recorded on the reflection spread. The 4.81 km/s layer (Fig. 7B) at Robinson River correlates with upper units of the Tawallah Group as well as a thin discontinuous veneer of McArthur Group and Cainozoic sediments. Below this, the Tawallah Group, which crops out along the basement ridge north of Robinson River homestead, correlates with the velocity of 5.44 km/s. The higher velocity of the equivalent layer at Borroloola may indicate the presence of McArthur Group carbonates, but these cannot be resolved from the data.

Pinchin (1979) measured velocities in McArthur Group rocks, using shallow seismic refraction near Mallapunyah. The velocity in the Emerugga Dolomite was 5.30–5.80 km/s, in the Tooganine Formation, 5.50 km/s, and in the Leila Sandstone Member, 6.00 km/s. The basement refractor may comprise pre-McArthur Basin metamorphics or acid volcanics and granites, as is believed to be the case towards the Westmoreland end of the traverse. Finlayson (1981) reported velocities in granites of about 5.69 km/s and a basement velocity of 6.06 km/s, within the Tennant Creek inlier, which are consistent with the present results.

The gravitational effect of the seismic model was calculated assuming two-dimensionality (Fig. 7). This is only approximately true, but the calculated and observed values agree fairly well between Borroloola and Robinson River. Between Robinson River and Westmoreland, the calculated gravity is higher than the observed gravity by up to 200 $\mu m/s^2$. This is because the traverse lies along the northeastern flank of a large gravity low (Fig. 2), which coincides with a basement granite body (Zadoroznyj & Plumb, BMR, personal communication, 1982). The basement velocity is slightly lower at the Westmoreland end of the traverse compared with Borroloola and Robinson River, and may reflect the presence of granites. The assumption of two-dimensionality of the structure used in the gravity modelling is probably not valid at this end of the traverse.

Deep structure

The record sections for Shots 1, 2, 7 and 8 are shown in Figures 5 and 9. The data quality is poor on the western traverse, particularly from Daly Waters and O.T. Downs, and this is believed to be due to near-surface effects at the shot sites attenuating the down-going energy (Pinchin, 1979). These effects have filtered out much of the high-frequency content of the signal from Shot 1 at Daly Waters (compare Figs. 5A and 5B).

A feature of all the 300-km long record sections is the lack of obvious wide-angle reflection branches, particularly from the Moho discontinuity. Later arrivals may be seen on some traces, but few can be correlated with any arrivals on adjacent traces, except at the farthest stations, beyond 250 km on the eastern traverse (Fig. 9). The reflections may be masked to some extent by the prominent wave-trains of the earlier phases, but they are not seen even where the amplitudes should increase, near the travel-time curve cusp points.

Moreover, the amplitudes of the first arrivals are large, which indicates they are diving waves rather than head waves generated at an abrupt velocity discontinuity. These two features indicate that the velocity increases gradually with depth rather than in discrete layers.

Western traverse

The velocity/depth model interpreted for the western traverse is listed in Table 1, and plotted in Figure 5C. The high velocity at the deepest part of the velocity gradient is not well defined, but the average gradient in the lower 20 km of the crust is well defined by the first-arrival data, i.e. an increase of about 0.06 km/s per km depth. The synthetic record section for this model is shown in Figure 10A.

Table 1. Velocity/depth models

The velocity/depth functions listed define the models used in the interpretation programs. Individual velocity/depth values include interpolated values which are necessary to define the gradient zones, and do not imply that the models are known to this accuracy.

Daly Waters (Shot 1)

Z (km)	V (km/s)
0.00	1.73
0.17	1.74
0.18	4.62
4.29	4.63
4.30	5.90
12.00	5.91
21.00	6.42
26.00	6.94
31.00	6.95
35.00	7.52
43.00	7.53
53.00	8.47
60.00	8.48

O.T. Downs (Shot 3)

Z (km)	V (km/s)
0.00	1.68
0.20	1.69
0.21	4.62
1.59	4.63
1.60	6.00
6.00	6.01

H.Y.C. (Shots 2, 4)

Z (km)	V (km/s)
0.00	1.68
0.07	1.69
0.07	5.81
12.00	5.82
21.00	6.42
26.00	6.94
31.00	6.95
35.00	7.52
43.00	7.53
53.00	8.47
60.00	8.48

Borroloola (Shot 7)

Z (km)	V (km/s)
0.00	1.71
0.05	1.72
0.06	3.58
0.42	3.59
0.43	5.55
1.80	5.56
2.20	5.57
2.60	5.60
2.80	5.65
2.90	5.70
2.96	5.75
3.04	5.80
3.14	5.85
3.40	5.93
4.01	6.04
8.00	6.05
15.10	6.15
15.25	6.24
15.35	6.30
15.55	6.38
15.85	6.42
16.15	6.46
16.50	6.47
21.00	6.50
22.00	6.52
22.50	6.55
23.00	6.60
23.50	6.81
26.00	6.82
28.00	6.84
30.00	6.88
32.00	6.94
34.00	7.02
38.00	7.40
40.00	7.80
60.00	7.91

Robinson River (Shot 6)

Z (km/s)	V (km/s)
0.00	1.71
0.05	1.72
0.06	4.81
0.70	4.82
0.71	5.44
3.54	5.45
3.55	6.04
8.00	6.05
12.00	6.17
15.00	6.20
16.00	6.25
16.50	6.47
21.00	6.48

Westmoreland (Shot 8)

Z (km)	V (km/s)
0.00	3.50
0.26	3.51
0.27	5.44
2.69	5.45
2.70	5.99
15.00	6.00
15.25	6.24
15.35	6.30
15.55	6.38
15.85	6.42
16.15	6.46
16.50	6.47
21.00	6.50
22.00	6.52
22.50	6.55
23.00	6.60
23.50	6.81
26.00	6.82
28.00	6.84
30.00	6.88
32.00	6.94
34.00	7.02
38.00	7.40
40.00	7.80
60.00	7.91

A model in which the velocity increases linearly with depth satisfies the travel-time data for depths below 15 km, and also the amplitude data in so far as there are no large later phases. This model does not show the prolonged, large amplitude, wave-trains visible in the observed records (Fig. 5), which are probably due to multiple reflections and scattering in the crust. However, the model in Figure 5C is preferred, because some later arrivals recorded could possibly be wide-angle reflections from the velocity gradients between 31 and 35 km, and 43 and 53 km depth.

At H.Y.C. (Fig. 5C), the 5.81 km/s layer is interpreted at 70 m below the surface; the next feature interpreted is the gradual velocity increase commencing at about 12 km depth. Any arrivals from intermediate depths are masked by these strong surface-layer arrivals (Figs. 5B & 6A). The deeper crustal section is assumed to be the same as the model derived from Daly Waters. The calculated travel-times are early by up to 0.4 s at distances beyond about 100 km from the shot. A correction has been applied to take account of the low-velocity surface layer on the western part of the traverse; however, it does not remove the total effect in the middle part of the traverse.

Eastern traverse

Some later arrivals are evident in the seismic recordings along the eastern traverse, e.g. between 90 and 125 km, Figure 9. The preferred velocity/depth model is listed in Table 1, and plotted in Figure 9C. The velocity increases gradually at depths greater than 44 km, because first-arrival amplitudes are large, but the gradient cannot be derived from the data.

The shapes of the gradient zones were constrained by the synthetic seismograms (Fig. 10B) generated from the model (Fig. 9C), and from the shallow model between Borroloola and Robinson River (Fig. 8D). Linear gradient zones were not adequate to account for the amplitudes of the observed arrivals. However, the average gradient in the lower 20 km of the crust is the same as for the western traverse, i.e., an increase of about 0.06 km/s per km depth.

At distances greater than 250 km, prominent arrivals can be seen about 1.2 s after the first arrivals from both forward and reverse shots (Shots 7 and 8) on the eastern traverse (Figs. 9A,B). Such events are not present on the western traverse record sections. The first arrivals are assumed to be Pn, i.e. arrivals refracted along the crust–mantle boundary. The amplitudes of the later arrivals are at least four times the Pn amplitudes. These arrivals may be reflections from a sub-Moho reflector. They are not simple wide-angle reflections from a sharply defined Moho, because their maximum amplitude occurs at too great a distance. They may represent a forward cusp on the Pn branch, owing to a zone of increasing velocity gradient at the base of the crust (Bullen, 1960); this has been observed in the West Australian Shield (B.J. Drummond, BMR, personal communication, 1981). Various velocity gradients were modelled, but, in all cases, the amplitude at the cusp relative to the first arrival was smaller than in the observed data.

Sub-Moho discontinuities that are shallow enough to produce strong wide-angle reflections in the observed time–distance range have been reported by Simpson (1973), Finlayson & others (1974), Hales & Rynn (1978), and Finlayson & McCracken (1981). They range in depth between 50 and 85 km. Travel-time curves were calculated for models that included sub-Moho discontinuities; none gave a satisfactory fit to the observed travel-times. The arrivals were either too early relative to the first arrivals, or were too far from the shot. To identify these arrivals properly, recordings beyond the maximum range of this survey (276 km) are required.

Vertical reflection recording

Vertical reflection recordings were made at the six refraction shot-points, and at a site near Starvation Hill between Borroloola and H.Y.C. (Fig. 2). A preliminary interpretation of these data was made by Pinchin (1980a).

At each site a 3-km recording spread was laid along the road, i.e. along the same azimuth as the long-range refraction lines, centred on the refraction shot. A short cross-spread, 1 km long, was laid at one end of the long spread, making either a 'T' or an 'L' shape with it, depending on access. The main spread had 36 recording stations at intervals of 83.33 m; the short spread had 12 stations with the same interval. Each recording station had 8 geophones per trace, in line, 5 m apart. Reflections were recorded at each site, from the large refraction shots as well as the small weathering shots. The large shots were offset laterally 15–30 m from the centres of the main spreads; the weathering shots were fired at each end of the main spreads. Digital recordings on magnetic tape were made on a DFS IV recording system; analogue records were produced in the field for monitoring purposes. Recordings were made at a sampling interval of 4 ms and the records were run for a total of 16 s.

The tapes were processed by Geophysical Services International, Sydney. This processing included true-amplitude recovery, time-variant scaling, correction for datum statics, normal move-out corrections, muting, and whitening deconvolution. At H.Y.C. and Borroloola, the two small weathering shots produced excellent reflections, so a 4-fold stack could be performed on the data. A time-varying filter was applied, with a 20–50 Hz passband at 0 s varying to an 8–30 Hz passband at 16 s. The processed section was produced as a variable area display (Fig. 11). Most records show several strong reflections and many weaker line-ups may be seen down to the bottom of the records at 16 s two-way time (about 50 km depth). Many more events were recorded on the eastern traverse than on the western traverse. At Daly Waters, and to a lesser extent at O.T. Downs, the energy was attenuated by near-surface effects (Pinchin, 1979).

Few reflectors can be traced from one shot site to the next, and these line-ups may be fortuitous, as there are numerous reflectors in each section. The most obvious arrivals that appear to be continuous are the reflection bands at about 6 and 6.5 s at Westmoreland, and 5.6 and 6.5 s at Robinson River. These reflection bands may continue to Borroloola, but they are not well defined. Many other strong reflection bands with distinctive characteristics occur both earlier and later in the sections. Some reflections have similar two-way times on two or more sections, but none can be followed from one section to the next on the basis of similar character. The lack of continuity of reflections is evidence of lateral variation of structure. Apart from record quality, there are no obvious features that distinguish the eastern traverse from the western traverse. However, none of the reflectors can be confidently traced from the eastern to the western traverse, which supports the interpretation of different velocity/depth functions from the refraction data for each traverse.

Two-way times to the major velocity changes, such as the tops of the gradient zones, were calculated from the refraction models and compared with the reflection times (Fig. 11). Velocity-gradient zones may be associated with bands of reflections, but, in general, broad gradient zones would be transparent to vertical energy. When reconciliation of refraction models and reflection data is being attempted, the differences between the methods must be considered. The vertical reflections are single probes recorded on 3-km spreads, each separated by 100–200 km. The refraction data are recorded over large distances with the result that the crustal properties are averaged along the whole ray path. Moreover, for the deeper layers, the velocity boundaries interpreted from the data apply only beyond about 50 km from each shot. If there is lateral structure, the refraction models will not apply at the shot points where the vertical reflection arrivals are recorded.

The characteristic features of the vertical reflection records, i.e. numerous reflectors, some with large dips, have been widely observed (eg. Smithson & others, 1977). Berry & Mair (1980) attempted to reconcile these results with the often different picture presented by the refraction data. They pointed out that the crustal velocity/depth distribution is determined by the

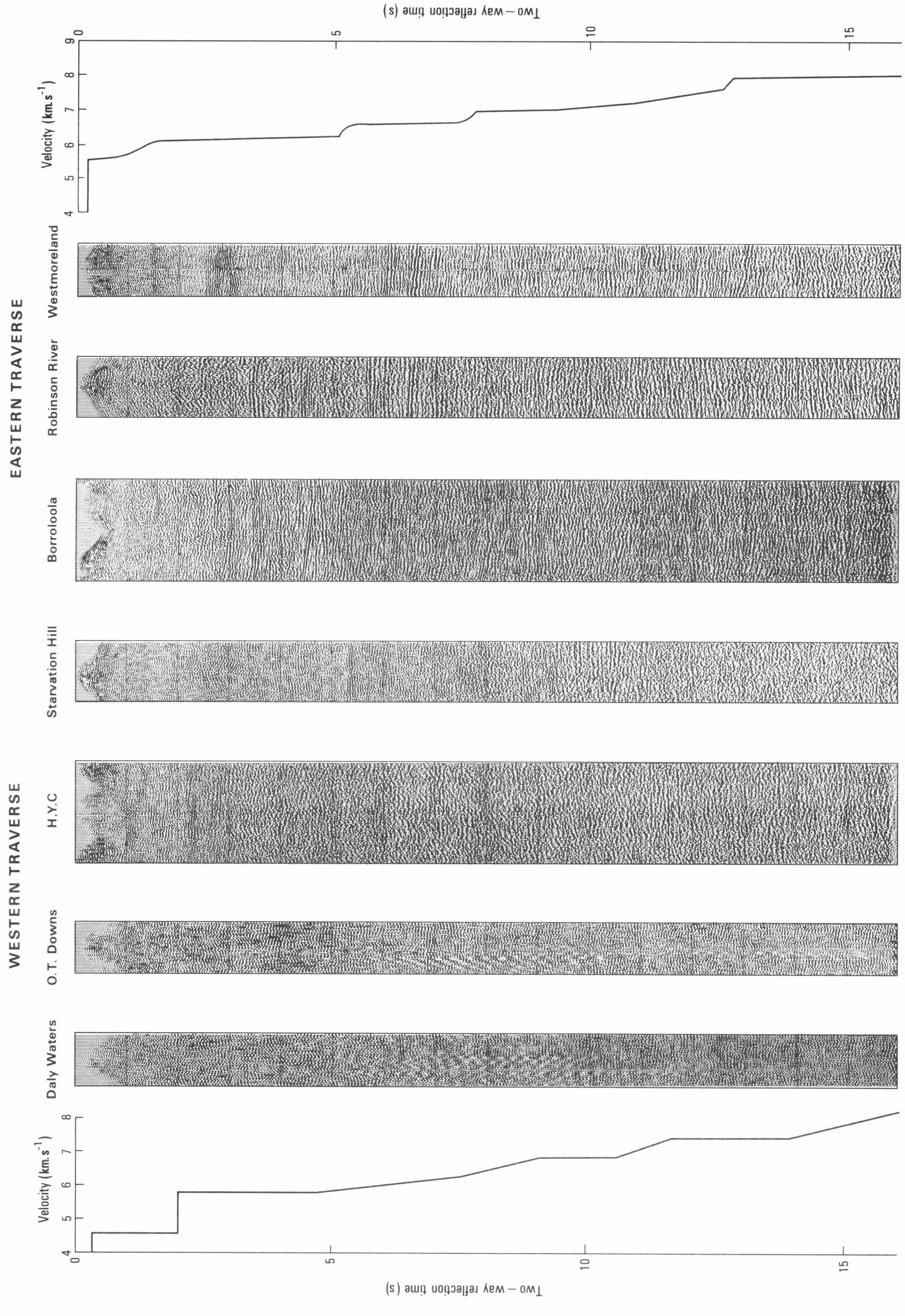

Figure 11. Vertical reflections recorded at each shot site, and at Starvation Hill.
For recording details see Pinchin (1979). The velocity/two-way-time diagrams computed from the velocity/depth function for each traverse (Figures 5(C) and 9(C)) are shown at the same two-way time scale.

following factors: the geochemistry and orogenic history, the distribution of water, and the past and present stress and temperature regimes. In general, vertical reflections are able to resolve the smaller-scale petrological features, while refraction tends to map large-scale variations, such as metamorphic grade, because velocities and depths are averaged along the traverse.

Reflection profiling over at least several tens of kilometres is essential if the true nature and extent of the reflections is to be resolved. Dipping events and diffractions cannot be interpreted from the single probes recorded here. This is obvious from the long profiles recorded by BMR in the Bowen and Eromanga basins (S.P. Mathur, BMR, personal communication), where large-scale continuity of events becomes apparent only when the profile is viewed as a whole. Randomly chosen 'probes' on the profile could give a misleading picture.

One apparent difference between recordings in the McArthur Basin and recordings in the Palaeozoic of eastern Australia is the lack of prominent wide bands of reflectors, which mark the middle to lower crust and the crust–upper mantle boundary. These can be seen even in the short (2–6 km) traverses recorded in the Lachlan Fold Belt (Pinchin, 1980b), but are not visible on the McArthur Basin records.

Discussion

The velocity/depth functions interpreted east and west of the Emu Fault have a similar overall velocity gradient with depth. However, the velocities and velocity gradients at any particular depth are different. Whether these differences occur abruptly at the fault or whether they occur gradually is not resolved except for the velocity structure within the top few kilometres. The shallow structure under the H.Y.C. deposit on the western side of the Emu Fault is markedly different from the nearby structure under Borroloola to the east, supporting geological and MT evidence for marked differences across the fault.

Both east and west of the Emu Fault, velocity gradually increases with depth. The gradients appear to be broader in the west, but this may in part be due to poorer data quality, which makes the interpretation of smaller-scale features difficult. In the lower 20 km of the crust, both traverses have the same overall increase in velocity of about 0.06 km/s per km depth.

Characteristic features of the seismic recordings are a lack of strong reflection branches, erratic large-amplitude later arrivals, strong first arrivals, long wave trains, and vertical reflections that appear to be discontinuous. Mereu & Ojo (1981) have shown that these features can be generated by small random inhomogeneities in an otherwise simple crust, in which the velocity increases linearly with depth, much as in the smooth gradient model, which satisfies the travel-time data and much of the amplitude data, for the western traverse. Moreover, they show that travel-times from this random model, plotted against distance, may form linear segments from which major horizontal velocity discontinuities may be interpreted. The size of the velocity inhomogeneities determines the depths and apparent velocities of the interpreted discontinuities. Uncertainties, such as lateral inhomogeneity in the crust, must be borne in mind when comparing the differences in detail between the eastern and western traverses. However, the major features of the velocity/depth distribution derived for each traverse are necessary to satisfy the travel-time data.

The vertical reflection records show numerous reflections, but they cannot be interpreted geologically without continuous profiling. The character of the records is different from recordings made over the Palaeozoic of eastern Australia, prominent wide bands of reflectors in the lower crust being absent. They are, however, similar to deep reflection recordings made in the southern Yilgarn Block of the Archaean Shield of Western Australia (Mathur & others, 1977); numerous reflectors throughout the records were observed, but no obvious bands of reflectors could be seen.

The basement of the McArthur Basin was identified under the Wearyan Shelf, but was not recorded in the Batten Trough, and probably not under the Bauhinia Shelf. The velocity of the layer below the ?Roper Group on the Bauhinia Shelf closely matches the velocity of the McArthur Group in the Batten Trough. Geological and MT evidence (BMR, 1966; Cull, 1982a) suggests the layer is Tawallah Group sediments. This may, therefore, explain why no velocity contrasts were detected within the Batten Trough. Apart from the relatively high surface velocities due to thick McArthur Group carbonates, which mask later arrivals, there may be little or no velocity contrast between the McArthur Group and Tawallah Group, and probably the same applies to the basement. The velocity of the Wearyan Shelf basement (6.0–6.1 km/s) is close to the velocity of the ?Tawallah Group in the Bauhinia Shelf (5.8–5.9 km/s); the Tawallah Group has a lower velocity on the Wearyan Shelf (5.4–5.5 km/s). Assuming that the basement velocity is similar on both traverses, the Tawallah Group would thus be detected, as has been observed, in the east, but not in the west.

Magnetotelluric (MT) results show a high resistivity basement beneath the Tawallah Group, or its equivalent, about 6 km deep at Daly Waters, increasing to 9.5 km below the thickest section of Roper Group further east (Cull,1982a). Seismic arrivals from any velocity contrast associated with the resistivity boundary may be masked by arrivals from the deeper velocity gradient, which begins at about 12 km depth, or else little or no velocity contrast exists, for the reasons outlined above. The latter case is supported by a lack of vertical reflections associated with the boundary.

MT results in the area are generally consistent with the seismic models (Cull,1982a). On the Bauhinia Shelf, between Daly Waters and O.T.Downs, thicknesses derived for the Roper Group from MT and seismic data are similar and show the same trends. Both MT and seismic data give a depth of about 2 km to the base of the Roper Group sediments near Daly Waters. However, from MT, the maximum depth is 5.5 km, 70 km east of Daly Waters, whereas the seismic data give a maximum depth of 4.1 km, 20 km east of Daly Waters. Both show the Roper Group gradually thinning eastward towards O.T. Downs, in agreement with surface geology.

On the Wearyan Shelf, near Borroloola, the MT results agree fairly well with the seismic model. A major resistivity boundary, believed to be basement, was interpreted at about 3.5 km depth, which correlates with the seismic basement at 3.3 km. However, south of Robinson River, MT defines the basement at 5.4 km, while the seismic basement occurs at 3.5–2.7 km. The seismic data are not sufficiently detailed towards the south to resolve the complex structures around Westmoreland, but they show that between Robinson River and Westmoreland the basement is flat-lying, in agreement with the MT data. Unknown thicknesses of Seigal Volcanics in the Tawallah Group, and the underlying Cliffdale Volcanics, may affect both the seismic and the MT data. Since there is no evidence in the seismic data for significant McArthur Group carbonates between Borroloola and Robinson River, there seems little potential for H.Y.C. type deposits east of the Emu Fault.

Small velocity inversions in the shallow layers of both traverses may exist, and indeed are likely because of the presence of high velocity carbonates and volcanics in the sediments, but there is no evidence of significant low-velocity zones in the crust. This is consistent with results from other Precambrian areas (eg. Meissner, 1976; Drummond,1979), but differs from the Tennant Creek and Mount Isa areas, where Finlayson (in press) interpreted three low-velocity zones in the upper crust. However, care is needed when generalising crustal structures to fit particular tectonic regimes. For example, the velocity/ depth function of such a vastly different area as south-western Honshu, Japan (Kohketsu,1981), bears a striking resemblance to the velocity/ depth function interpreted here. On the other hand, it is very different from the Proterozoic Cuddapah Basin, India, which otherwise has many similarities to the McArthur Basin (Plumb,1981); very strong wide-angle reflections are observed, which clearly represent abrupt velocity discontinuities (Kaila & others, 1979), particularly at the crust-mantle boundary. Similar wide-angle reflections are observed in the Archaean Pilbara Block of Western Australia (Drummond, 1979, 1981).

By comparing crustal models of the Tennant Creek Inlier, Mount Isa Block, Arafura Sea, and McArthur Basin, Finlayson (in press) interpreted a progressive increase in crustal velocities from south to north in the North Australian Craton. Also, deep crustal velocities appear to be higher under the central part of the craton than under the eastern part. Whether this is a general rule for the whole craton cannot be established until more data are available.

Inliers of Archaean basement underlie the North Australian Orogenic Province in the north (Plumb, 1979), but the extent of this Archaean basement is unknown. Plumb (1979) raised the possibility that an Archaean continental crust might underlie the whole region, because the major fracture pattern in the region is geometrically regular and has existed since at least the Early Proterozoic. The same conclusion was reached by Rutland (1981) on the grounds that the active continental margin during the Proterozoic was likely to be in the east, in the region now occupied by the younger Tasman Province; the Proterozoic crust is postulated to be the result of vertical accretion on and within an Archaean proto-continental crust. The crust in the Archaean Pilbara Block is much thinner (28–33 km) than the north Australian crust, and has relatively low velocities within the lower crust (6.4–6.55 km/s) and a well-defined crust–mantle boundary. The crust of the northern Yilgarn Block is at least 50 km thick; the form of the crust–mantle boundary is undefined, but the lowest layer interpreted has a velocity of 6.9–7.0 km/s compared with 6.4–6.55 km/s in the Pilbara (Drummond, 1981), and the boundary may be gradational. The Proterozoic Capricorn Orogenic Belt between these blocks is distinguished by intermediate crustal thickness and high velocities and velocity gradients at the base of the crust. Drummond (1981) has suggested that the Capricorn Orogenic Belt was derived from the northern Yilgarn Block, which was modified during Proterozoic tectonism; the (assumed) Archaean 'proto-continental' crust of north Australia may, therefore, have been similar in certain respects to the Yilgarn crust.

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