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Crustal structure of the Precambrian terrains of northwest Australia from seismic refraction data

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The crust of the Pilbara Craton is 28-33 km thick, and its boundary with the upper mantle dips south at slightly less than one degree. The southern edge of the craton is marked by a sharp increase in crustal thickness about 40 km south of the Sylvania Dome in the east and along the northern boundary of the Ashburton Trough in the west. Seismically, the crust has two layers, an upper one with P-wave velocity 6.0-6.1 km s^{-1}, and a lower of 6.4-6.6 km s^{-1}. The boundary between them is depressed beneath the deepest part of the Hamersley Basin, and rises along the southern part of the basin and craton, beneath a topographic ridge of higher grade metamorphic rocks. The velocities in the lower crustal layer are highest along the southern part of the Hamersley Basin, an area known to have been depressed during the formation and filling of the basin, and then uplifted. The crust of the northern Yilgarn Craton is at least 50 km thick and, seismically, has three layers. The uppermost layer has P-wave velocities of 6.1-6.2 km s^{-1}, the middle layer, at 10-16 km depth, 6.4 km s^{-1}, and the lowest, at 32 km depth, 6.7-7.0 km s^{-1}. A zone of crust, about 50 km wide, along the northern margin of the craton was extensively reworked during Proterozoic orogenesis, and is now characterised by thinner crust than the rest of the Yilgarn Craton, extensive granitoid emplacement, intense deformation and folding of the upper crustal rocks, and low gravity values. The Capricorn Orogen between the cratons is marked by high velocities in the lower crust, probably caused by dense mafic intrusions or a higher metamorphic grade. The seismic layering within the crust is most likely caused by increasing metamorphic grade with depth. The crust is probably of average acid to intermediate chemical composition, and has a low metamorphic grade at the surface. The velocities are attributed to metamorphic grade increasing to felsic granulite at 9-16 km depth, and garnet granulite at 32 km depth in the Yilgarn Craton.

Introduction

During 1979, seismic refraction measurements were made along five profiles in the Pilbara region of Western Australia (Drummond, 1979a). Seven open-cut iron-ore mines in the northern Pilbara Block and the Hamersley Basin regularly fired large quarrying blasts, and these were used as seismic sources.

The seismic survey was planned so that the refraction profiles would traverse all the Precambrian geological provinces in the region. The survey design is superimposed on the geological map in Figure 1. BMR recorded along lines ABC, FDB and ADHE (Drummond, 1979a), and the Research School of Earth Sciences at the ANU recorded along lines GF and DC. Drummond (1979b) and Drummond & others (1981) used the intercept method (Mota, 1954) to interpret the data from lines ABC and FDB, and extended their interpretation by using computer ray tracing through their models to test for lateral structural changes.

In this paper the results of the earlier interpretations are summarised, the data from the three remaining traverses (ADHE, DC and FG) are interpreted, and the results from all five profiles are used to derive a picture of the crustal structure of the region.

Geology

The regional geology is shown in Figure 1. The tectonic nomenclature follows that of Gee (1979a, c); accordingly, a 'block' is the exposed part of an old, stabilised region called a 'craton'. In the survey area, two Archean cratons, the Pilbara and Yilgarn Cratons, are overlapped and separated by younger sediments and volcanics of several sedimentary basins.

The greenstone belts of the Pilbara Block have provided Australia's oldest isotopic ages of 3400 to 3500 m.y. (Richards, 1977; Pidgeon, 1978a; Hamilton & others, 1980), and sinuously enclose large, ovoid domes of younger, intrusive granitoid, with isotopic ages grouped around 3000 m.y. (Oversby, 1975; Pidgeon, 1978b; de Laeter & Blockley, 1972; de Laeter & others 1980, 1981) and 2600-2800 m.y. (Oversby, 1975; de Laeter & Blockley, 1972; de Laeter & others, 1980). The younger granitoids are a minor proportion of the granitoids of the Pilbara Block, and may have resulted from partial remelting of the older crust (de Laeter & Blockley, 1972).

The Yilgarn Block, which extends about 1000 km south of the area shown in Figure 1, also has a granitoid/greenstone stratigraphy. In the north, the majority of the granitoids have isotopic ages of about 2600 m.y. (Compston & Arriens, 1968; Oversby, 1975; de Laeter & others, 1980).

Volcanic and sedimentary rocks and banded iron formations of the Hamersley Basin overlap the southern edge of the Pilbara Block. Trends in the regional gravity pattern (Fraser, 1973; Wellman, 1978), and inliers of Archaean basement (e.g. the Sylvania Dome) indicate that the Pilbara Craton forms the basement of the Hamersley Basin. The rocks of this basin have marked lateral continuity, and are thought to have originally covered most of the Pilbara Craton. They are now restricted to the southern part of the craton, where the basin was deepest (Trendall, 1975; Horwitz & Smith, 1978). Where the basal unconformity of the basin is exposed, the basal volcanics are restricted to the centre of greenstone belts, implying that the granitoid plutons were still rising when the basin formed (Trendall, 1975). The intensity of folding in the basin increases southwards (Trendall, 1975).

Abutting the southern edge of the Hamersley Basin is the Ashburton Trough (the Wyloo Trough of Horwitz & Smith, 1978), which contains thick clastic sediments, turbidites, and olistostromes. The Ashburton Trough is younger than the Hamersley Basin, and

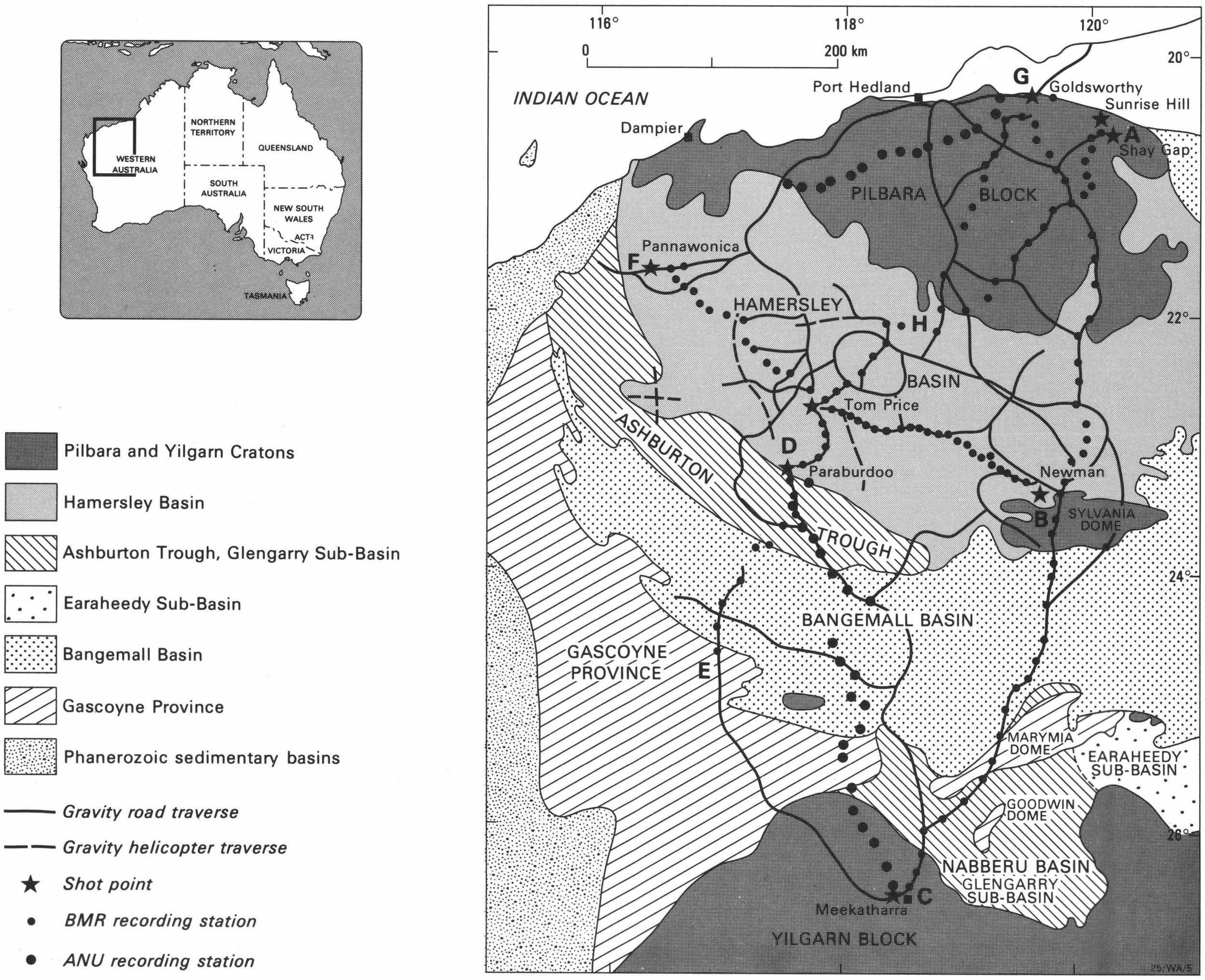

Figure 1. Geology and survey design.

is possibly contemporaneous with the Glengarry Sub-basin of the Nabberu Basin, which bounds the northern edge of the Yilgarn Craton (Gee, 1979a).

The Glengarry Sub-basin contains clastic sediments, volcanic rocks, and turbidites, which dip gently off the Yilgarn Craton. The dip and deformation of the sediments increase northwestwards across the basin (Hall & Goode, 1978; Gee, 1979b) and, although the structural relations between the Nabberu Basin and the Hamersley Basin are masked by the younger cover of the Bangemall Basin, Horwitz (1975) was able to show a mirror image symmetry of the structures in the two basins.

Clastic sediments, dolomites and cherts are the principal rock types of the Bangemall Basin; volcanic rocks are not common. Changes of facies and structural style reflect the stability of the basement on which the basin formed. Stable basement parallelled the southern boundary of the Pilbara Craton in the north, and the northern edge of the Yilgarn Craton in the south. The western and central portions of the basin had an active basement (Brakel & Muhling, 1976).

The Gascoyne Province in the west of the survey area contains reworked Archaean basement rocks, and reworked sediments from the Bangemall and Nabberu Basins and Ashburton Trough (Daniels, 1975; Williams & others, 1979). Granitoids which crop out in the province were derived from reworked Archaean basement, possibly mixed with younger magmatic or supracrustally reworked material (de Laeter, 1976; Williams & others, 1978).

The Gasycoyne Province, Bangemall Basin, and Ashburton Trough occupy the Proterozoic Capricorn Orogenic Belt (Gee, 1979c). Palaeozoic sedimentary basins bound the region on the west, north, and east.

Seismic wavegroup nomenclature

The nomenclature used in this paper is similar to that used by Drummond & others (1981), which is derived from Giese (1976).

Pg — Direct waves, which travel through the near-surface crystalline basement with velocities of 6.0-6.2 km s^{-1}.

P* — Waves critically refracted at an intracrustal boundary, and with velocities of 6.4-6.55 km s^{-1}.

P^I — Waves reflected from the intracrustal boundary and travelling with velocities intermediate between those of the Pg and P* wavegroups.

Pn — Waves critically refracted at the crust/mantle boundary, and having apparent velocities in the range 7.6-8.6 km s^{-1}.

P^M Waves reflected from the crust/mantle boundary, and with apparent velocities intermediate between those of the P* and Pn wavegroups.

P_1,P_2 — Near-surface wavegroups observed where the Hamersley Basin strata are underlain by lower velocity rocks. P_1 waves have velocities of about 5.9 km s^{-1}, and P_2 waves have velocities of 6.3-6.7 km s^{-1}.

P^L — Near-critical point reflections from a deep crustal layer in the northern Yilgarn Craton. They have apparent velocities of about 7.0 km s^{-1}.

Results

Line ABC, including blasts at G (Shay Gap, Sunrise Hill, Goldsworthy-Newman-Meekatharra)

The data from line ABC (and line GBC) were interpreted by Drummond (1979b), and a slightly revised model, shown here as Figure 2, was given by Drummond & others (1981).

In the north, along segment Shay Gap-Newman, the crust is 28 km thick at Shay Gap, 32 km thick at Newman, and is two-layered. The upper crust has a seismic velocity of 6.0 km s^{-1}, and at about 13 km depth the velocity increases to 6.4 km s^{-1}. The upper mantle velocity is 8.34 km s^{-1}. The Hamersley Basin north of Newman is represented by a thin layer with a low (5.86 km s^{-1}) velocity.

Along the segment from Newman to Meekatharra, the Pn arrivals are delayed beyond 250 km, and as no such delay is apparent in the Pn data in the reverse direction, crustal thickening to the south is inferred. Crustal thickening is consistent with intercept times from recordings of Newman and Meekatharra shots, and with delays of teleseismic arrivals from earthquakes in the Fijian region (Drummond, 1979c), and the Indonesian Arc to the north (Drummond & others, in prep.). Ray tracing has shown that, in the most likely models, the crust thickens from about 33 km south of Newman to more than 50 km at Meekatharra. The principal differences in the models occur in the lower crust along the segment between Newman and Meekatharra.

The seismic velocity in the upper crust between Newman and Meekatharra is 6.1-6.2 km s^{-1}. At 10 km depth in the north and at 16 km depth in the south, the velocity increases to 6.4 km s^{-1}. In the south of segment BC, a lower crustal layer, with a velocity of about 7 km s^{-1}, is implied by a band of large-amplitude reflections at about 7 s and 140-200 km. The crust in the south is therefore three-layered. The upper mantle velocity may decrease from 8.34 to 8.1 km s^{-1} southwards along segment BC.

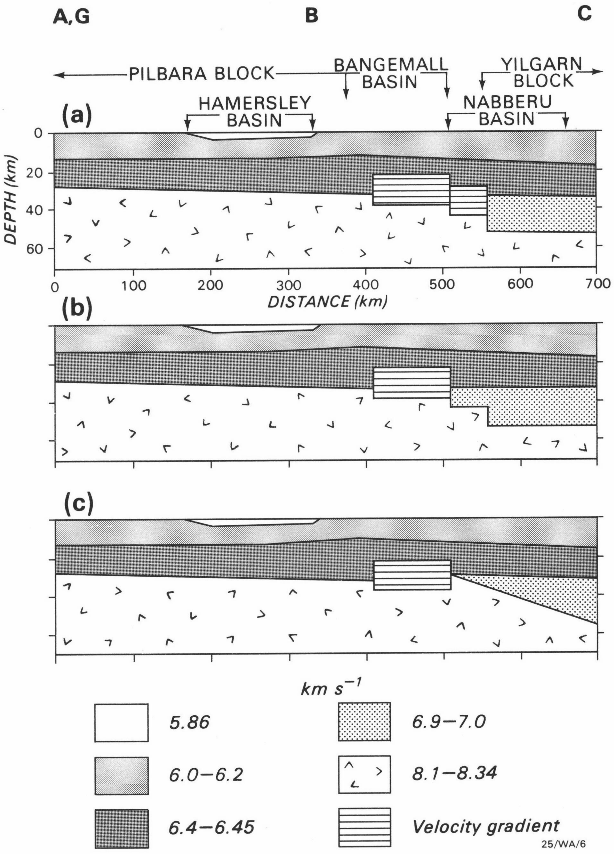

Figure 2. Three alternative seismic models for line GBC (and line ABC).
The models are those of Drummond & others (1981).

Line FDB (Pannawonica-Tom Price-Newman)

Drumond & others (1981) reported the results from line FDB, and Figure 3 shows their model. It is preliminary, because several features of their record sections could not be explained by ray-tracing methods alone.

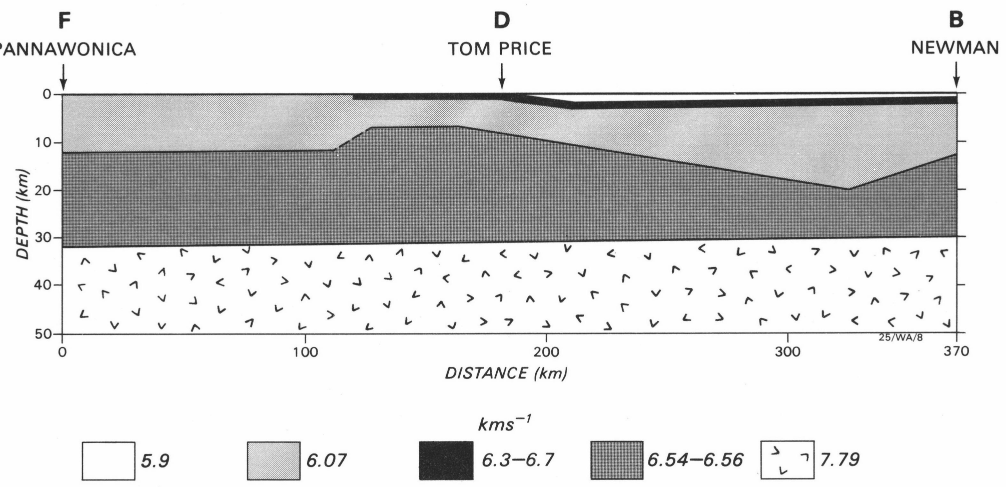

Figure 3. The crustal model derived by Drummond & others (1981) for line FDB.

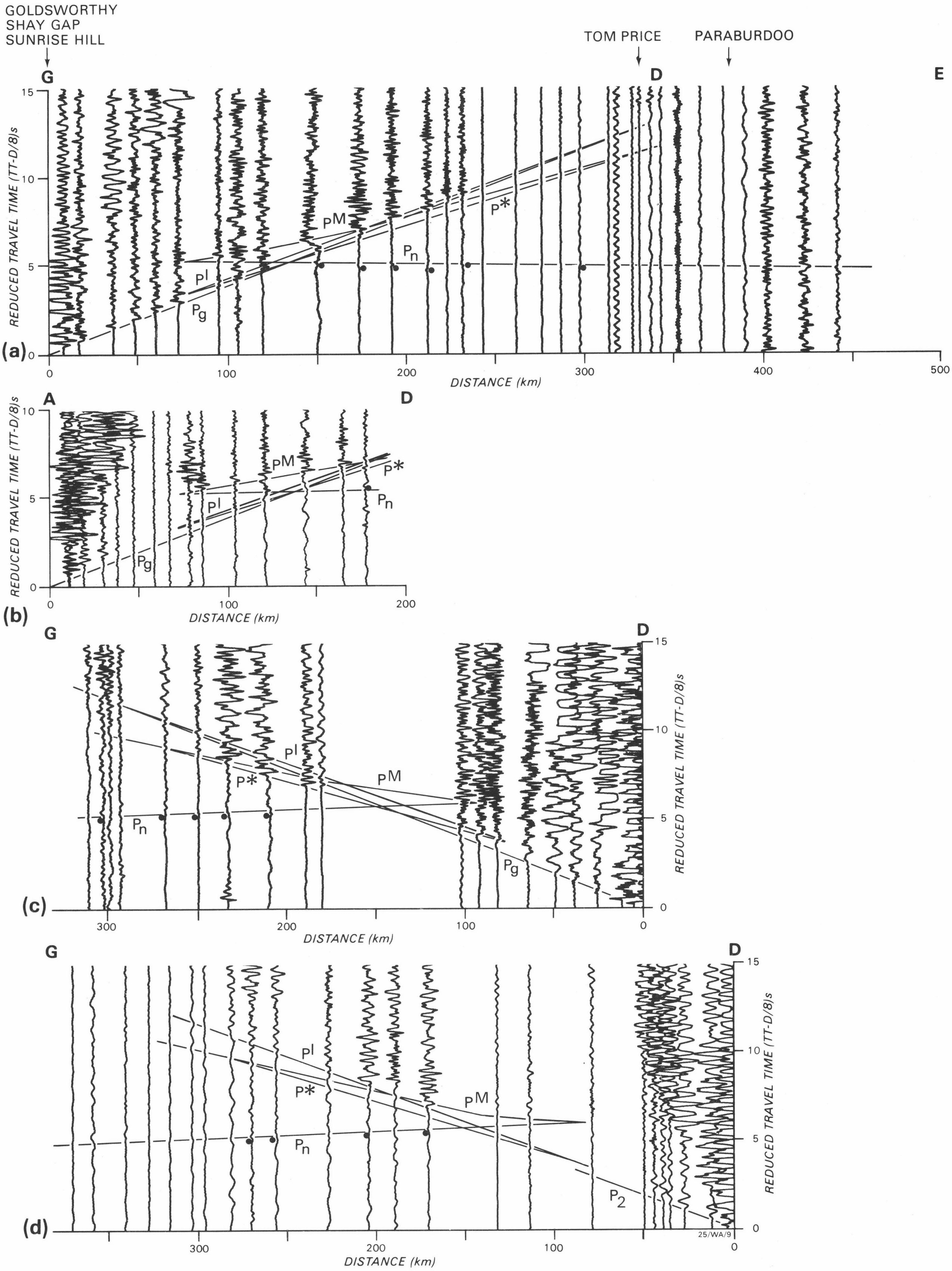

Figure 4. Record sections for line GD.
Sections a, b, c and d are for shotpoints G southwards, A southwards, Tom Price northwards and Paraburdoo northwards respectively. The superimposed travel-time curves are for the model in Figure 8a. The trace amplitudes are as recorded, and all traces were digitally filtered in the bandpass 0.5 to 8 Hz. The dots indicate the times of P_n arrivals picked from these and large amplitude records.

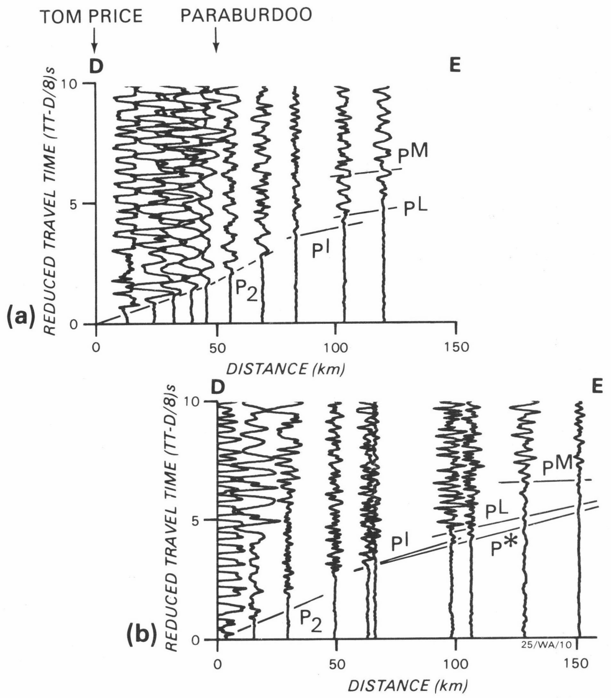

Figure 5. Record sections for line DE.
Section a is for Tom Price southwards, and section b is for Paraburdoo southwards. The superimposed travel-time curves are for the model in Figure 8b. The trace amplitudes are as recorded, and all traces were digitally filtered in the bandpass 0.5 to 8 Hz.

Between Tom Price and Newman, thin, near-surface layers with velocities of 5.9 and 6.3-6.7 km s^{-1} are interpreted as Hamersley Basin banded iron formations and volcanic and sedimentary rocks. They overlie a deeper crust in which the seismic velocities are less; the velocity of 6.07 km s^{-1} used throughout the profile was the unreversed velocity measured in the western part of the basin from a blast at Pannawonica. A lower crustal layer under the basin has a slightly higher velocity (6.54-6.56 km s^{-1}) than the corresponding layer under Newman measured along the north-south profile (Figure 2).

The boundary between the upper (6.07 km s^{-1}) and lower crustal layers has considerable topography. It is 12 km deep under Pannawonica, and shallows to 7 km depth to the west of Tom Price. Further east it deepens, perhaps reaching 20 km before shallowing to about 13 km under Newman.

Sub-Moho refracted arrivals are poorly observed along this profile, the reversed upper mantle seismic velocity was estimated at 7.8 km s^{-1}, which is considerably lower than along profile ABC. The crust/mantle boundary was modelled with very little topography; it is 31 km deep under Pannawonica, and 30 km deep under Newman. It may deepen slightly in the centre of the profile.

Lines GDC (including blasts at A) and DE (Goldsworthy-Tom Price-Meekatharra; and Tom Price towards SW)

The data from lines GD, DE, and DC are depicted as record sections in Figures 4, 5 and 6 respectively and the velocities and intercepts for the refracted phases and the P^L phase identified in, and scaled from, these record sections are listed in Table 1. In all cases, and in the following discussion, the travel times are reduced by the factor, distance/8.0. In Figures 4 and 5, the traces were digitally filtered in the bandpass 0.5 to 8.0 Hz, and the trace amplitudes are not normalised; normalised amplitudes were used in some cases in the identification and correlation of phases, but these are not shown here. In Figure 6, all record sections have had the trace amplitudes normalised so that the maximum amplitude for each trace is the same; the traces in Figure 6 were filtered in the bandpass 1.0 to 8.0 Hz. The models derived from the record sections are shown in Figure 7. The travel-time curves superimposed on the record sections in Figures 4, 5, and 6 were derived by computer ray tracing through the models in Figure 7; the following discussion serves to illustrate the wave groups used to derive the models.

Representative record sections from the line between Goldsworthy and Tom Price (GD) are shown in Figure 4. Two mines are located near D. The northern one is at Tom Price, and the southern one at Paraburdoo (Figure 1). They were treated as separate shotpoints and record sections from both are depicted.

The record section of blasts from the Goldsworthy shotpoint (G) is presented in Figure 4a. Coherent seismic energy was not observed beyond 240 km. The near-surface Pg phase is clearly seen as a first arrival to the crossover point with Pn at about 130 km. The Pn phases are very weak; dots indicate the arrivals that can be identified on larger scale records. A P^*/P^I cusp can be identified about 0.5 s after Pg, at about 70 km, and the P^M phase is clearly seen as large amplitude later arrivals in the 80-240 km distance range.

The smaller blasts at shoptpoint A (Shay Gap and Sunrise Hill), generate less seismic energy and, therefore, smaller amplitude arrivals than those at Goldsworthy. The P^M phase is, therefore, more clearly seen because the earlier arrivals do not obscure it (Figure 4b).

The Tom Price record section for line GD is shown in Figure 4c. The modelled P^M travel-times are about 0.5 s later than the recorded times. This was interpreted by Drummond (1979b) for data from shotpoint G along line GB as evidence for positive velocity gradients in the lower crust. The P^I wavegroup is clearly identified as a high-frequency phase following the low-frequency Pg arrivals.

In the record section of the Paraburdoo blasts recorded northwards (Figure 4d), the first arrivals from the shotpoint to about 50 km show a very rapid fall off in amplitude. This is indicative of a low-velocity zone in the upper crust north of the shotpoint. As will be shown in Figures 5 and 6, the near-source first arrivals probably belong to the P_2 phase which travels through the high-density, high-velocity, near-surface Hamersley Basin rocks. The P^*/P^I cusp is not clearly seen in this record section. The P^M phase is clearly seen, but the Pn phase is very weak.

The record section of Tom Price shots recorded southwards on line DE is shown in Figure 5a. The P_2 phase is recorded with strong amplitudes to about 60 km, whereafter it is delayed and its amplitudes weaken. The delay and weakening of amplitudes is also observed southwards from Paraburdoo (Figure 5b), but at only a few stations within 20 km of the blast. The delay and fall off in amplitudes corresponds, for both shotpoints, with the southern boundary of the Hamersley Basin about 10-15 km south of Paraburdoo (Gee, 1979a). The near-source upper crustal phase has, therefore, been interpreted as the P_2 phase, and not the

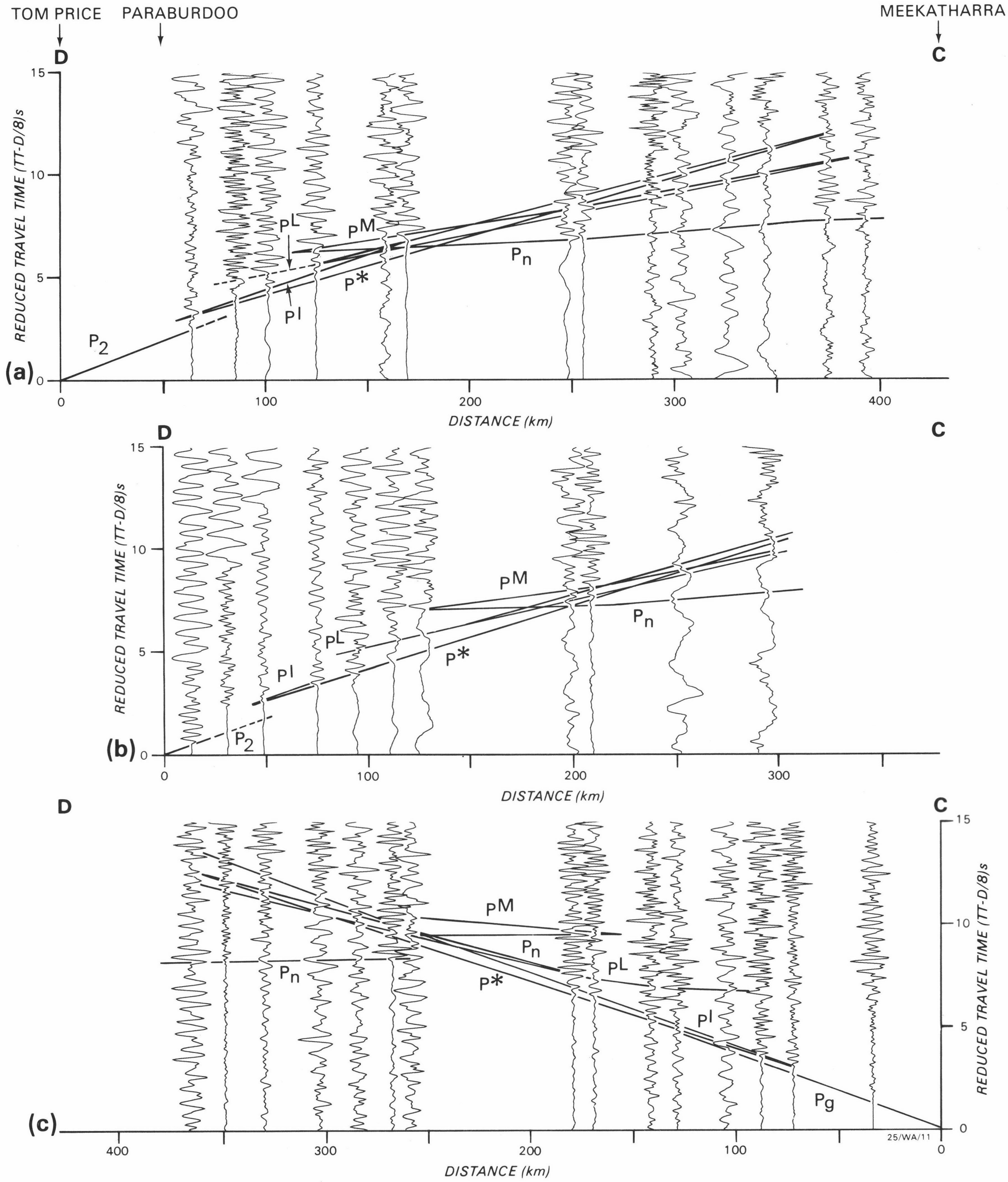

Figure 6. Record sections from line DC.
The sections a, b and c are for Tom Price southwards, Paraburdoo southwards, and shotpoint c northwards, respectively. The superimposed travel-time curves are for the model in Figure 8a. The trace amplitudes have been normalised so that the maximum amplitude is the same for all traces, and the traces in section c were digitally filtered in the bandpass 1.0 to 8.0 Hz. For some recorders, one of the horizontal components had clearer arrivals than the vertical component and has been plotted in its place. Note that the P^L cusp (the dashed curve) is closer to the shotpoint than predicted by the model.

Pg phase. This is the main reason for interpreting the near-source first arrivals in Figure 4d as P_2. Some of the lower crustal phases can also be identified in Figure 5b, in particular, the P^I and P^L phases.

Figure 6 contains the data from the ANU line between Paraburdoo and Meekatharra (DC). In Figure 6a, the P_2 phase is identified by only one early arrival, but it is identified with confidence because, at that point, the ANU line is coincident with the BMR line DE, on which the data in Figure 5a were recorded. The Pn phase is observed as strong arrivals on two traces at 240 to 260 km, but beyond that it is not clear. The traces at 375 and 395 km have large amplitude arrivals at about 8.5 s, which are delayed in time relative to Pn between 120 and 260 km, but may still be Pn arrivals. Figure 6a is similar to the record section from the line between Newman and Meekatharra (BC), where large-amplitude Pn phases are clearly evident, and can be correlated from trace to trace, and Pn is delayed beyond 250 km (Drummond, 1979b).

The P^L phase is very clearly seen in Figure 6a; the cusp on the travel-time curve for the model (Figure

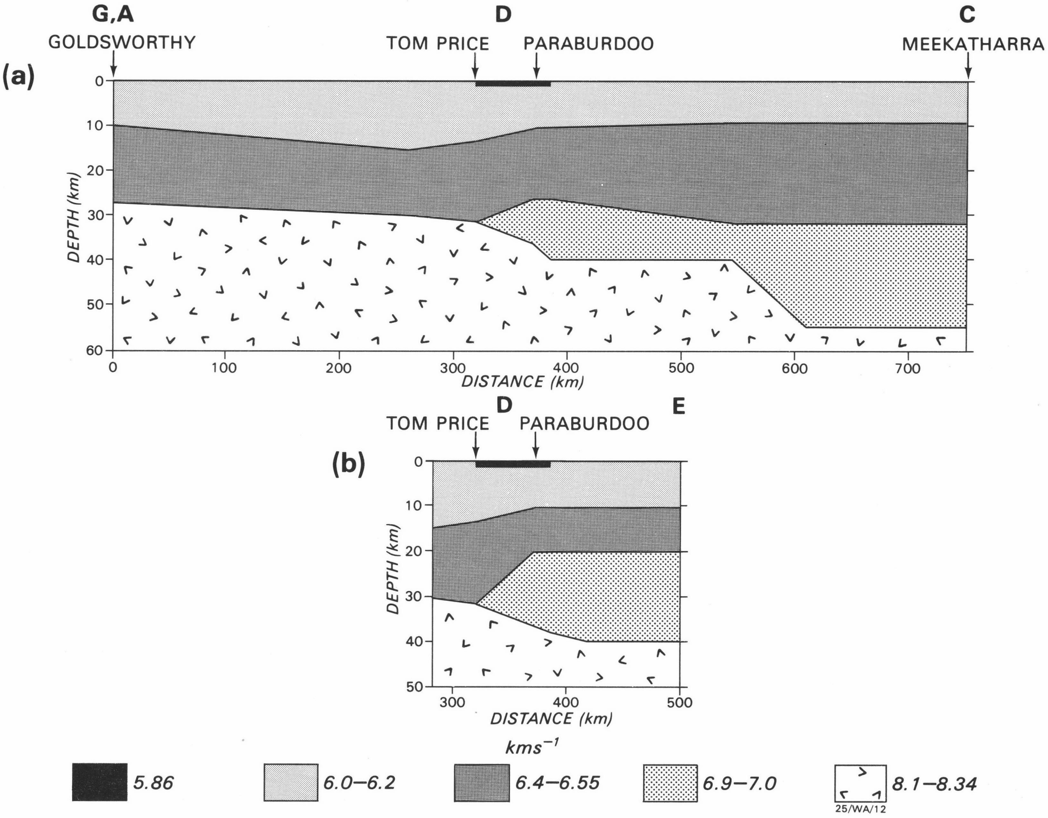

Figure 7. Seismic model for lines GDC and DE.
Figure 7a is the model for line GDC, and Figure 7b is the model for line DE.

LINE GD (Goldsworthy, Shay Gap and Sunrise Hill to Tom Price and Paraburdoo)

	Goldsworthy southwards		*Shay Gap & Sunrise Hill southwards*		*Tom Price northwards*		*Paraburdoo northwards*	
	V	*ti*	*V*	*ti*	*V*	*ti*	*V*	*ti*
P_2							6.19	0.0
Pg	6.11	0.0	6.15	0.0	6.10			
P*	6.36	1.0			6.55	1.6	6.54	1.2
Pn	8.13	5.5			8.23	6.2	8.23	6.2

LINE DE (Tom Price and Paraburdoo southwards)

	Tom Price southwards		*Paraburdoo southwards*	
	V	*ti*	*V*	*ti*
P_2	6.25	0.0		
Pg				
P*			6.25	1.2
Pn				

LINE DC (Tom Price and Paraburdoo to Meekatharra)

	Tom Price southwards		*Paraburdoo southwards*		*Meekatharra northwards*	
	V	*ti*	*V*	*ti*	*V*	*ti*
P_2						
Pg					6.13	0.0
P*	6.54	1.30	6.54	1.30	6.37	0.85
P^L	6.99	3.40	6.99	3.40	7.04	4.46
Pn	7.89	6.23	7.89	6.80	8.17	10.15[1]
					8.12	8.70[2]

[1] Later arrivals between 120-180 km.
[2] First arrivals beyond 180 km.

Table 1. Velocities ($km\ s^{-1}$) and intercepts (seconds) scaled from record sections (Figures 4, 5 & 6).

7) is at 130 km, but the clearest P^L phases are observed between 80 and 100 km. The P^L cusp is also closer to the blast than the model would predict in Figure 6b, the southwards record section of Paraburdoo blasts. This may imply that the lower crustal, 7.0 km s^{-1}, layer should be shallower in the model. Pn is not clear in Figure 6b.

The record section for shotpoint C at Meekatharra is shown in Figure 6c. The Pg phase is weak, but it is neverthless clear on large scale plots. The P^L cusp is clearer on the lower gain records, and the P^M phase is not as clear as it is for the same shotpoint along line CB towards Newman (Drummond, 1979b). Pn is observed clearly on only two traces—those at 265 and 350 km.

The data from Figure 4 were interpreted by taking reciprocal pairs of refracted phases between Goldsworthy and Tom Price (GD), and inverting them, using the intercept method, for refractor depths, velocities, and dips. The data from line DE are unreversed, and were interpreted by assuming that the layers were horizontal and that the apparent velocities were refractor velocities. The Pn delays at distances beyond 250 km in Figure 6a imply structures on the crust/mantle boundary; the boundaries between the layers and the crust/mantle boundary at either end of line from Tom Price to Meekatharra (DC) were assumed to be horizontal and the velocities to be the refractor velocities. The models from these various lines were then combined. The calculated depths were assumed to apply only beyond the offset distances to the refractors, and the composite models were then adjusted using ray tracing so that the final models (Figure 7) satisfied the travel-time data.

Figure 7a depicts the preferred model for line GDC (Goldsworthy-Tom Price-Meekatharra) and Figure 7b the preferred model for line DE (Tom Price towards the SW).

The upper crustal velocity along the profile is 6.0-6.1 km s^{-1}. In the centre of the profile, near Tom Price (D), a thin, high-velocity, near-surface layer corresponds to the Hamersley Basin rocks. It has been modelled as 1 km thick, but it may be thicker. Assessment of the thickness of such layers is difficult with ray tracing alone, and further modelling with synthetic seismograms is required. A lower crustal layer, with velocities of 6.37-6.55 km s^{-1} is present along the profile. It is 10 km deep under shotpoint G at Goldsworthy, deepens to 15 km about 260 km south of G, but then shallows again to 10 km between Tom Price and Meekatharra (DC). An even deeper crustal layer, with a velocity of 6.7-7.0 km s^{-1} is present along DC. It is 10 km thick south of Tom Price (D), and thickens rapidly in the centre of the line, to be 20-25 km thick under Meekatharra (C). The interpreted upper mantle velocity along the Goldsworthy-Tom Price-Meekatharra line varies between 8.0 and 8.2 km s^{-1}. The crust/mantle boundary is 28 km deep under Goldsworthy, deepens steadily to 31.5 km under Tom Price, and then rapidly to 40 km, and then to 55 km southwards towards Meekatharra.

The model for line DE (Figure 7b) is similar to the equivalent part of line DC to the east, except that the lowermost, 6.98 km s^{-1}, crustal layer is thicker. This may not be significant, because the seismic line is unreversed.

Line FG (Pannawonica-Goldsworthy)

Only 15 stations were installed along line FG. The data are shown in Figure 8. The recorders were set up along the northeastern half of the line, but not all operated successfully, and consequently, it is not possible to interpret the data with any degree of confidence. The model adopted for the line is shown in Figure 9 and was derived by taking the refractor depths under Pannawonica (F) from line FDG (Figure 3), and those under Goldsworthy (G) from line GDC (Figure 7), and assuming that the refractors were plane layered between F and G. The velocities used were averages of those from the northern part of line GDC (Figure 7) and the western part of line FDB (Figure 3). The travel-time curves for the model were then derived by computer ray tracing, and are superimposed on the record sections. They are generally consistent with the observed data, so the model in Figure 9 is taken to be the best available for the line between Pannawonica and Goldsworthy.

Accuracy of the models

The refracted travel-time branches for the models in Figure 7 fit the observed data to within 0.1 to 0.2 s, thereby implying an accuracy of one or, perhaps, two kilometres in the calculated refractor depths. However, greater errors may be caused by assuming that the refractor velocities are constant through the layers.

Berry (1971) showed that errors of up to 20% can arise in the refractor depths, owing to use of over-simple velocity/depth functions, such as those used here, where the velocity is assumed constant through each refractor and the boundaries between the layers are sharp. Drummond (1979b) suggested, on the basis of a study of reflected travel-time branches, that the crust/mantle boundary of the Pilbara Craton was transitional, and this may be so for the other boundaries as well. Consequently, the refractor depths from this study are probably minimum values. This should be considered when the refractor depths are quoted in the ensuing discussion. It should be noted, however, that the under-estimate of refractor depth probably applies to all profiles, so that the relative thicknesses are maintained from one area to another.

Discussion

The models from all the seismic lines are combined as a fence diagram in Figure 10. The refractor depths generally agree at the junction of the profiles within the accuracy of the modelling. Figure 11 depicts the interpretation of the seismic models in terms of the geology .

Pilbara Craton

The Pilbara Craton is 28 km thick in the north and 30-33 km thick in the south, where its boundary with the Capricorn Orogen is interpreted as the zone where the crust thickens sharply to the south. This is about 40 km to the south of the Sylvania Dome in the east, and corresponds to the northern boundary of the Ashburton Trough in the west. The crust/mantle boundary under the craton, therefore, has a dip of slightly less than one degree to the south. The dip is observed on all profiles except FDP, which lies approximately east/west and on which there is no significant dip from the west on the crust/mantle boundary.

An intracrustal boundary is indicated by the increase of seismic velocities from between 6.0-6.2 km to 6.4-6.55 km s^{-1} within the Pilbara Craton. The depth to the boundary varies throughout the region. Drummond (1979b) modelled it as 13 km deep under the eastern Pilbara Craton, but suggested that it could dip from 9 km in the north to 14 km in the south. The latter values are more in agreement with the depths from the central Pilbara Craton (Figure 7a), where the boundary dips from 10 km in the north to 15 km to the north of Tom Price. It shallows to the south of Newman in the east (Figure 2) and between Tom Price and Paraburdoo (Figure 7a). Along the southern edge of the craton, it is deepest betwen Newman and Tom Price, shallows to 7 km to the west of Tom Price, and then deepens again to 11 km in the west (Figure 3).

Southwards across the Pilbara Craton, the top of the intracrustal boundary defines a basin-like structure with a general southeast/northwest axis. The basin cuts the seismic profiles west of Newman and north of Tom Price. The basin-like structure corresponds to the

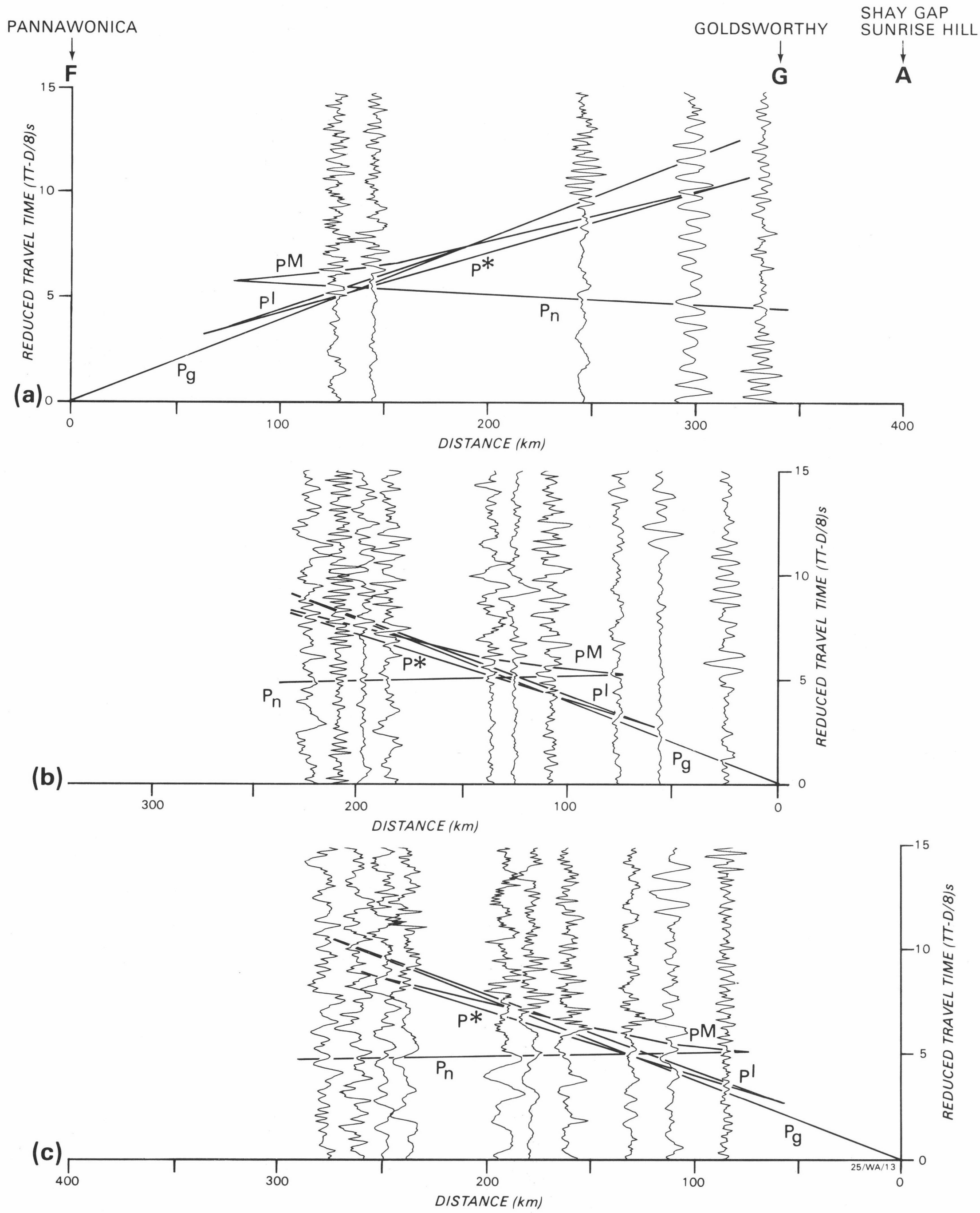

Figure 8. Record sections from line FG.
Sections a, b and c are for shotpoint F northeast, G southwest, and A southwest respectively. The superimposed travel-time curves are for the model in Figure 9. The trace amplitudes have been normalised as in Figure 6. For some recorders, one of the horizontal components had clearer arrivals than the vertical component and has been plotted in its place.

deepest part of the Hamersley Basin inferred from Trendall's (1975) isopach maps and structural diagrams. Horwitz & Smith (1978) defined an Archaean ridge along the southern edge of the Pilbara Craton which corresponds to the shallowing of the intracrustal boundary south of Newman and south and west of Tom Price.

The highest velocities (6.55 km s^{-1}) observed below the intracrustal boundary occur along the southern basement ridge and the axis of the Hamersley Basin. These velocities are due to higher metamorphic grades in the lowermost crust, caused by the depression of the crust into higher pressure and temperature regimes during the formation of the Hamersley Basin (Drummond & others, 1981). Uplift occurred after the filling of the Hamersley Basin, and can be divided into two types, regional isostatic rebound and localised diapirism of basement rocks.

Regional isostatic rebound followed the filling of the basin. This can be inferred from the exposure at the surface of some of the stratigraphically lowest members of the Hamersley Basin sequence. As well, R. E. Smith & others (personal communication) used the metamorphic grades mapped at surface exposures to estimate uplifts of 10-12 km, west of Tom Price, and 5-7 km, near Newman. The second type of uplift was localised diapirism of basement rocks to form Archean inliers within the Hamersley Basin. The largest of these is the Sylvania Dome, but several smaller ones are observed or inferred to the west and east of Tom Price. Those near Tom Price correspond to the region where the intracrustal boundary shallows to 7 km depth along profile FDB (Figures 3 and 10).

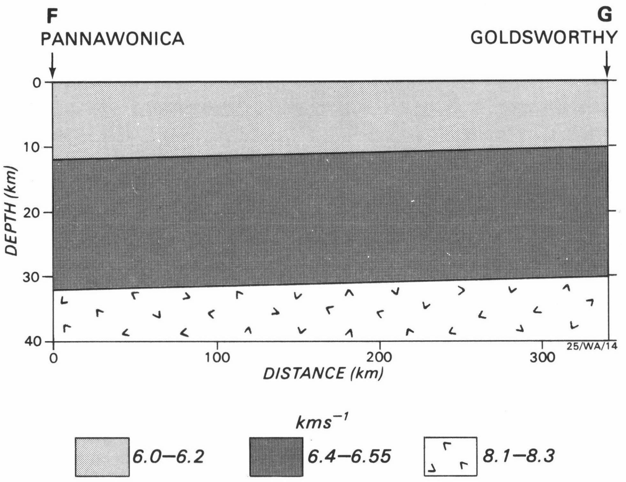

Figure 9. Seismic model for line FD.
The model was not derived from the data in Figure 8, but is an average of the models for the northern end of line GD and the western end of line FD (see the text).

The diapirism may have been caused by density inversions near the surface. Where rocks of the Hamersley Basin are seismically distinct from the underlying basement rocks, they have velocities which are higher than those of the basement. They are highest in the southeast of the basin. Further west and north, the velocities of the P_1 and P_2 wavegroups decrease as the metamorphic grade decreases (Drummond & others, 1981), and the basin rocks have not achieved sufficiently high velocities to be seen to be independent of the underlying Archaean basement. Gravity data also infer that the Hamersley Basin rocks are denser than the Archaean basement (Drummond & Shelley, 1981).

Yilgarn Craton

The crust of the Yilgarn Craton is much thicker (> 50 km) than the crust of the Pilbara Craton, and extends further north than the area mapped by Gee (1979a) as the Yilgarn Block. Intracrustal seismic boundaries occur at 10-16 km and at 32 km depth.

In the Proterozoic, the northern limit of the stable Yilgarn Craton was probably just south of the present day northern edge of the Nabberu Basin. A zone along the northern edge of the Yilgarn Craton, where the crustal thickness is intermediate between that of the Yilgarn Craton and the thinner crust of the Capricorn Orogen, probably represents part of the northern Yilgarn Craton which was reworked by tectonism in the Proterozoic. It corresponds to a large, negative gravity anomaly to the north of the Yilgarn Block. Fraser (1973) suggested that the anomaly was too intense to

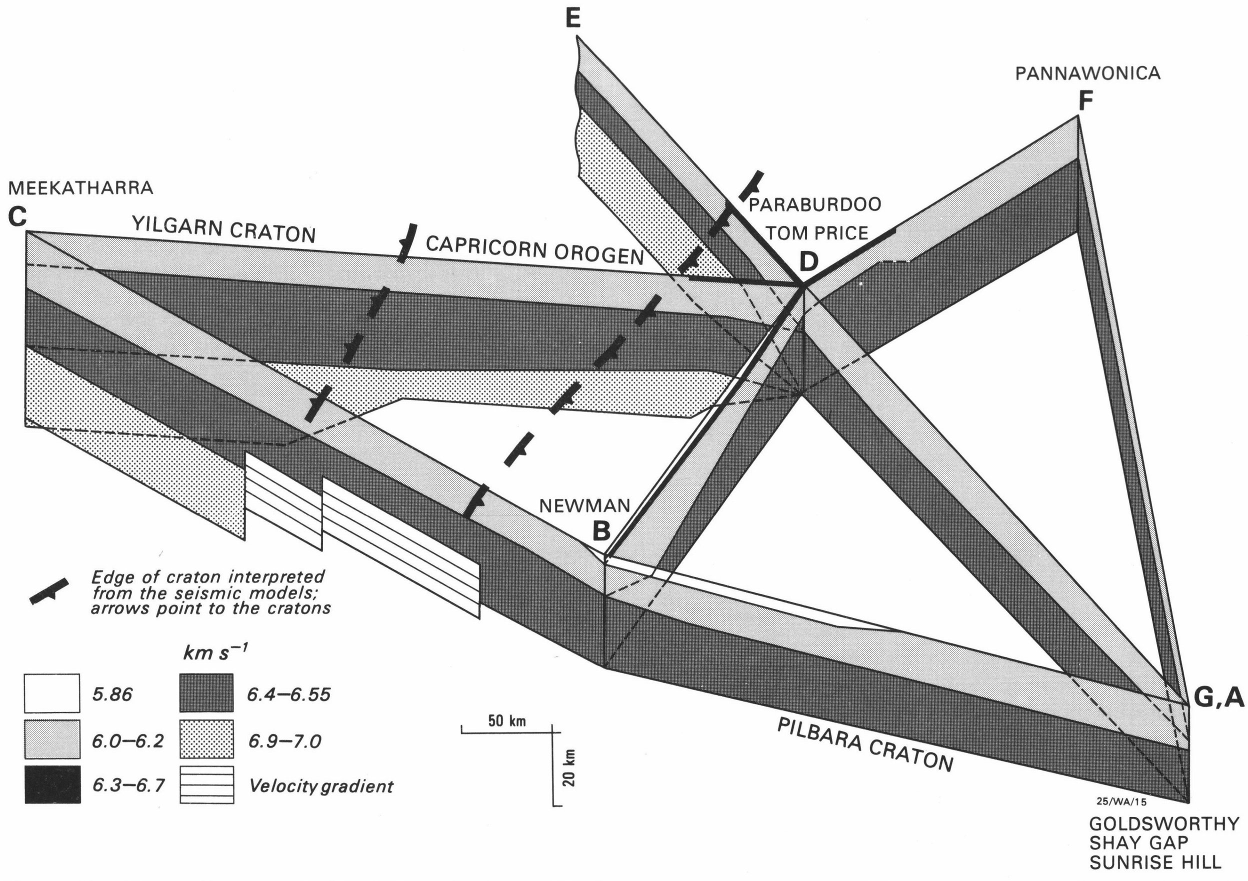

Figure 10. Fence diagram of all of the seismic models from the Intercept Method interpretation of the data from the 1977 Pilbara Crustal Survey.
The interpreted margins of the cratons are based on crustal thicknesses. The edge of the Pilbara Craton corresponds to the southern margin of thin crust, and the northern edge of the Yilgarn Craton corresponds to the northern margin of thick crust. The Capricorn Orogen between the cratons has crust of intermediate thickness.

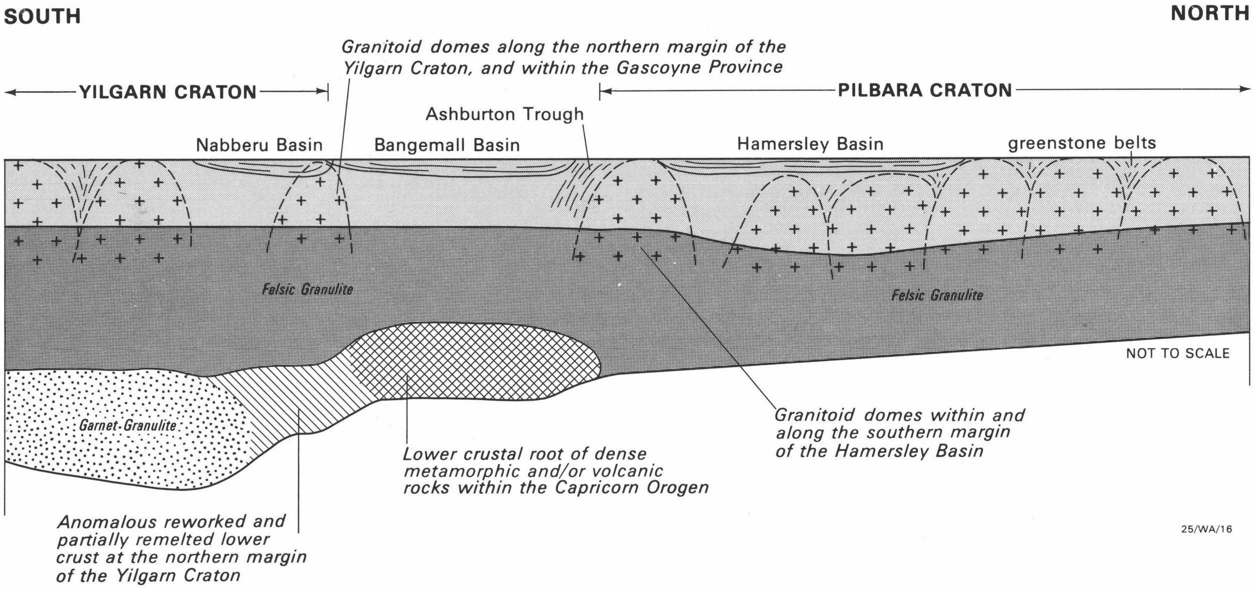

Figure 11. Cartoon sketch of a north/south section of crust through the survey area, in which the seismic models are interpreted in terms of the geology.

be purely a gravity edge effect of the Yilgarn Craton, and Horwitz & Smith (1978) correlated it with 'certain granitic and metamorphic rocks of the Gascoyne Province'.

The area contains, in the Gascoyne Province, granitoid intrusions, which have been interpreted as remelted basement and younger sediments (de Laeter, 1976; Williams & others, 1978), and, further east, the Marymia Dome at the northern edge of the Nabberu Basin. It also includes the northwestern part of the Nabberu Basin, where the sedimentary rocks and underlying basement were intensely deformed (Hall & Goode, 1978). The southern belt of stable basement in the western Bangemall Basin falls within and marks the northern limit of this zone.

Capricorn Orogen

No shotpoints were located in the Bangemall Basin or Ashburton Trough, and because the basin and trough rocks do not cause any noticebale deviation of the travel-times from those expected from normal crystalline basement, no estimate of the present day basin and trough thicknesses is possible.

The lower crust in the Capricorn Orogen is characterised by high velocities and, by inference, dense rocks. Wellman (1978) attributed the denser crust in the orogen to higher metamorphic grades. Horwitz & Smith (1978) favoured an evolutionary model for the orogen, in which the Pilbara and Yilgarn Cratons were welded into a stable basement by the end of the Archaean and then rifted to form the Capricorn Orogen. From the gravity map (BMR, 1975), they inferred dense lower crust in the orogen, and attributed it to intrusions of basic magmas into the lower crust during the rifting. However, the intrusions could also have occurred if the cratons had not been welded, but had been free to move relative to each other to relieve the stresses which must have built up during tectonism.

Chemical composition of the crust

The seismic velocities in the upper crust throughout the region, excluding those in the Hamersley Basin, range from 6.0 to 6.2 km s^{-1}. Drummond (1979b) suggested that they represent an upper crust of overall acid to intermediate chemical composition and low metamorphic grade. Below the intracrustal boundary at 9-16 km depth throughout the region, the seismic velocity increases to 6.4 km s^{-1}, and 6.55 km s^{-1} under the axis of the Hamersley Basin. Drummond (1979b) attributed the increase in velocity to the metamorphism of the acid to intermediate crust to felsic granulite. A chemical change was ruled out, because more basic rocks would have higher seismic velocities (Christensen & Fountain, 1975).

Underlying the felsic granulite layer at 32 km depth along the northern Yilgarn Craton is the thick, lower crustal layer with seismic velocities of 6.7-7.0 km s^{-1}. Drummond (1979b) suggested that this layer was caused by 'eclogite'-grade metamorphism of rocks with compositions similar to or slightly more basic than the upper crust. Drummond & others (1981) called it, perhaps less ambiguously, garnet granulite. Garnet granulites of acid to intermediate chemical composition readily give seismic velocities of about 7 km s^{-1} at depths greater than 30 km (pressures greater than 10 kbar). Basic garnet granulites often have much higher seismic velocities, and eclogite velocities at pressures of 10 kbar can exceed 8.00 km s^{-1} (Birch, 1961; Christensen, 1965; Manghnani & others, 1974; Christensen & Fountain, 1975). However, Drummond & Shelley (1981) combined the seismic and gravity data, and suggested that some chemical layering was possible in the Yilgarn Craton, with the lower crust more basic than the upper crust.

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