Teleseismic travel time anomalies and crustal structure in central Australia

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Teleseismic P-wave travel time anomalies of more than 0.5s have been recorded in central Australia between sites separated by less than 100 km. Late arrivals occur at sites that are located within sedimentary basins and early arrivals occur at sites located on the exposed and eroded basement. Early arrivals correspond to relative gravity highs and late arrivals correspond to gravity lows. The differential travel times cannot be attributed to near surface crustal structure alone. The most straightforward interpretation is in terms of a Moho whose depth varies by about ±10 km from a mean value and which is deepest below the basins. This interpretation is consistent with the intracratonic basin evolution model proposed by Lambeck.

1. Introduction

The structure and evolution of the central Australian intracratonic basins and the intermediate uplifted arches, pose a number of interesting tectonic questions. Lambeck (1983a) has discussed these structures in some detail and proposed a model whose principal characteristic is that the structures formed by horizontal compressive forces acting on an inhomogeneous viscoelastic lithosphere. Uplifted areas are eroded and the downwarped areas form the sedimentary basins. When bending stresses become excessive, major faulting of the lithosphere as a whole may occur. The proposed model is based on a simple mechanical analogy, supplemented by geological evidence and gravity observations. The north-south basin structures predicted by the model are illustrated in Fig. 1 for a section extending from the Officer Basin in the south to the Ngalia Basin in the north, corresponding approximately to the profile AA' of Fig. 2. Independent tests of the model are important for understanding (1) the deep crustal structure and rheology, (2) the formation and evolution of intracratonic basins and (3) the mid-plate orogenies that have occurred, for example, in central Australia. These orogenies occurred in Early Cambrian and Early Carboniferous time (e.g., Wells et al., 1970; Forman and Shaw, 1973) and they can be considered as an integral part of the compressive basin formation model (Lambeck, 1983b).

A principal characteristic of the model is that the sedimentary sequence is believed to have been deposited on crust of average thickness of ~30-40 km and which is then depressed downwards. The intermediate arches have, on the other hand, been uplifted and eroded and the crust here is thinner than average. The base of the crust is therefore predicted to exhibit considerable undulation, possibly more than ±10 km from an average depth. A similar conclusion had been reached earlier by Forman and Shaw (1973) who noted a correlation between Bouguer anomalies and metamorphic grade and interpreted this in terms of uplift of the crust between the basins (see also Mathur, 1976).
A simple test of the model is to examine the P-wave arrival times of teleseismic events recorded at stations located across the structure: first arrivals are predicted to be early over the arches of uplifted and thinned crust and late over the deep basins lying on otherwise average thickness crust. Order of magnitude calculations predict that the differential travel time anomalies may approach 1s (Lambeck, 1983a). The initial results of such an experiment are reported here.

2. Data

Several overlapping lines of seismometers were deployed across different parts of the structure. The first line (B–B' of Fig. 2), extending across the Ngalia Basin and the Arunta Block, was part of a longer line designed to study deep mantle structure. Initial results for station anomalies have been reported by Wright (1983). The line A–A' (Fig. 2) extending from the southern margin of the Amadeus Basin to north of the Ngalia Basin was designed specifically for the study of station anomalies. This part of the experiment was carried out in three steps (lines 2–4). Line 2 consisted of sites 6–25 but this part of the experiment was marred by instrument failure. The third line (sites 1–19) extended the line to the southern margin of the Musgrave Block and overlapped considerably with, and filled gaps in, the second line. The final group of stations (20–24) and (26–34) was set up to tie together the lines A–A' and B–B' and to fill in gaps caused by the above-mentioned instrument failures. Only the data recorded on the line A–A' (station sites 1–25) of Fig. 2 will be discussed in this paper.

The choice of sites was governed partly by the known or inferred tectonic structure but mainly by the accessibility of an inhospitable terrain. In particular, it was not possible to position stations 1–4 in the south and 24–25 in the north further to the west. As a result, these stations do not pass as centrally over the quasi-linear tectonic structure as is desirable. All stations were initially located from 1:250 000 maps and aerial photography but final
Fig. 2. Location of the seismic stations across the central Australian region. The shaded regions represent the basins. Bouguer gravity anomalies are also illustrated. Note that contours are not given for all multiples of 20 mGal. A total of 34 sites were occupied but not all stations operated at the same time. The first group of stations (denoted by 1 in parenthesis following the station number) is the line BB' (stations 26–34). The second part of the experiment included stations 6–25 (2 in parenthesis) of the line AA'. The third part, sites 1–19, (3 in parenthesis) extended the line further south and filled in gaps incurred in (2) due to instrument failure. The fourth part, sites 8–24 and 26–34, (4 in parenthesis) filled in gaps in the northern part of the line AA' and ties this line to BB'.

Each site consists of a single vertical-component short-period seismometer whose output is recorded onto analogue magnetic tape together with clock records and radio time signals (e.g., Muirhead and Hales, 1980). Selected high quality earthquake records, exhibiting impulsive onsets, are digitized at 20 samples s⁻¹ and clock corrections are applied. The unfiltered seismograms for a given event are displayed on a single reduced travel-time versus distance plot. Depending on the quality of the record, time measurements are made on one or several well-defined points on the leading part of the P wavetrain.

Only earthquakes to the east of the line are considered in the present study. These events lie approximately at right angles to the north–south line AA' so that the computed relative travel times are insensitive to uncertainties in the travel-time curve. Seventeen events are from the Fiji–Tonga region at an azimuth of 92–102° and a distance range of 43–50°. Two Santa Cruz events at an azimuth of 76–79° and distances of 35–37° and one Kermadec event at an azimuth of 110° and distance of 45° have also been included. The depths of the events range from 38 to 650 km. For any one event the change in distance and azimuth across the array is only a few degrees. The angle of incidence for most P waves is from 22–25°.

Earthquake epicentre parameters are taken from the bulletin of the International Seismological Centre for line 2. For lines 3 and 4, recorded in 1982, the preliminary epicentres of the U.S. Geoparenthesis) of the line AA'.
logical Survey, National Earthquake Information Service, were used. Considerable differences between the two sets of epicentres sometimes occur and, in order to reduce location errors for earthquakes recorded on these last two lines, the only events used are those whose locations are based on 30 or more world-wide stations.

3. Method of analysis

For an event \( j(1 \ldots J) \) the travel time observed at station \( i(1 \ldots I) \) is first corrected for the focal depth of the earthquake. This corrected time is denoted by \( t_{ij}^0 \). The corresponding computed travel time \( t_{ij}^c \) is based on the Herrin et al. (1968) model. Differences \( \Delta t_{ij} \) between the two times arise from a number of sources and we can write

\[
\Delta t_{ij} = t_{ij}^0 - t_{ij}^c = \sum_{k=1}^{\gamma} \Delta t_{ij}(k) \tag{1}
\]

The contribution \( k = 1 \) is a combination of any error in the source time and the systematic difference between the arrival time and the actual peak picked on each record. For the selected events there is little change in waveform across the array and this error, \( \Delta t_{ij}(1) \), is constant for the \( I \) stations recording the event. \( k = 2 \) corresponds to an error in the earthquake coordinates. This can introduce an error that is approximately linear with the epicentral distance \( \Delta r_{ij} \) for the event. It is practically zero for a source located on the same approximate great circle as the line of stations. It reaches a maximum for an event orthogonal to the line. This error source is potentially important: location errors may not vary randomly from event to event in a given locality because of the nature of the world wide network distribution and because of near source mantle structure (see for example Bock, 1983, for errors in the location of Fiji-Tonga earthquakes). \( k = 3 \) accounts for anomalous structure in the source region. This effect can be severe unless the wavepaths traverse similar velocity structures in the vicinity of the source region and this is believed to be the case for Fiji-Tonga events recorded in central Australia (Bock, 1981).

The contribution \( \Delta t_{ij}(4) \) refers to the shortcomings of the travel-time curves. Despite much progress, the global travel-time curve remains subject to considerable uncertainty due in part to possible lateral variations in the mantle or to excessive smoothing of the travel-time data. In the distance range considered here, from 40 to 50°, slowness discrepancies of as much as 0.1 s deg\(^{-1}\) relative to the Herrin et al. (1968) curve have been noted (e.g., Muirhead and Hales, 1980; Wright, 1983). Neglect of this possible error may result is significant pseudo station anomalies which vary with distance \( \Delta r_{ij} \). \( k = 5 \) represents station location errors. These are insignificant here. \( k = 6 \) represents the errors in the time measurements from the seismograms. From a comparison of the measurements made on several points on the seismogram and from independent selection by both authors of the peaks, these errors are believed to be < 0.1 s for the selected records. The final contribution \( \Delta t_{ij}(7) \) is the sought anomalous crustal and upper mantle structure term or station anomaly. This term includes the effect of the variation in velocity along the wave path when compared to the ‘normal’ path and the change in path caused by a slope on surfaces such as the Moho. By taking events that arrive from a direction that is nearly orthogonal to the structure, this latter effect is minimized.

In view of the above remarks we write

\[
\Delta t_{ij} = a_j^0 + a_i \Delta r_{ij} + T_i + \epsilon_{ij} \tag{2}
\]

for a distance range \( \Delta r_{\text{min}} < \Delta r < \Delta r_{\text{max}} \) and an azimuth range \( \Delta \alpha_{\text{min}} < \Delta \alpha < \Delta \alpha_{\text{max}} \). For numerical reasons we use instead of \( \Delta r_{ij} \) in eq. 2 the quantity \( \Delta r_{ij} - \bar{\Delta} \), \( \bar{\Delta} \) being the mean of all \( \Delta r_{ij} \). The constant term \( a_j^0 \) for each event incorporates the above \( k = 1 \) errors. The linear term \( a_i \) represents a correction to the travel-time curve in the specified distance range. By selecting events within a relatively narrow distance and azimuth range, lateral mantle structure and systematic source location errors may also be absorbed by the \( a_i \) term. The station anomalies may be azimuth dependent and this is a further reason for selecting events within a narrow azimuth range. The \( \epsilon_{ij} \) represent unknown, ideally mainly random, corrections to the observations. With \( J \) events recorded on \( I \) stations, the maximum number of observation equations is \( I \cdot J \) (not all stations may have recorded all events). The
number of unknowns are $I + J + 1$. All observations have been given a standard deviation $\sigma_i$, of 0.1s. Since it is not possible to distinguish between a constant part of the station anomaly across the array and the $a_i^0$, the least squares solution of (2) is subjected to the condition $\sum T_i = 0$.

4. Results

The observed differential travel time anomalies, defined by

$$\delta t_{ij} = \Delta t_{ij} - \frac{1}{J} \sum_{j=1}^{J} \Delta t_{ij}$$

(3)

are illustrated in Fig. 3 for 15 of a total of 20 selected events recorded on lines 2, 3 and 4. The remaining five events produced results of equal quality. Agreement between arrival time residuals for different events recorded at the same group of stations is generally of the order of 0.1 s, as is the agreement between residuals recorded at stations common to 2 or more lines. Some events, e.g., 410, exhibit a quite marked slope across the line and this may be indicative of an event mislocation.

The least squares solution for the station anomalies is illustrated in Fig. 4. The precision estimate of each anomaly follows from the product of the solution covariance matrix with the variance of unit weight

$$\sigma^2 = \sum_i \sum_j \left( \phi_i^{*} \sigma_i^2 \phi_j \right) / n$$

where the superscript * denotes the transpose of the correction vector and $n$ is the number of degrees of freedom of the solution. The expected value of $\sigma$ is unity and the solution yields $\sigma = 1.07$. This, plus the fact that the corrections $\phi_i$ appear to

Fig. 3. Observed station residuals defined by eq. 3. The upper group of residuals are from the third part of the experiment. Only events originating in the Fiji–Tonga region are illustrated. The central group of residuals are for stations 18–24 on the line AA' from the fourth part of the experiment. Stations 18 and 19 are common to the two groups of stations. The lower group of residuals corresponds to the second part of the experiment. Stations 6, 7, 15, 16 and 19 are common with the residuals in the upper part of the Figure. Stations 19 and 24 are common with the residuals of the central part of the Figure.
follow a normal distribution, indicates that the model is consistent with the data and the a-priori standard deviation estimates.

The quality of the fit of data to the model can be further examined by computing the quantity

\[ \sigma_{\text{event}}^2 = \sum_i \epsilon_i^2 / (I^* - 1) \]

for each event, and the quantity

\[ \sigma_{\text{station}}^2 = \sum_j \epsilon_j^2 / (J^* - 1) \]

for each station. Here \( I^*(\leq I) \) is the number of stations that recorded the particular event and \( J^*(\leq J) \) is the number of events recorded by the station in question. In most instances \( \sigma_{\text{event}} < 0.1 \) s. The maximum value is 0.14 s and this may be indicative of a source location error. \( \sigma_{\text{station}} < 0.1 \) s except for stations 6 and 25 where values of 0.17 and 0.14, respectively, are found. The result for station 6 may be attributed to a known recurrent problem with the seismometer at this site during the line 2 experiment which may have resulted in the estimated arrival times to be systematically late. Compare, for example, the residuals for station 6 observed on line 2 and 3 (Fig. 3). This does not affect the solution as a whole: solutions with and without station 6 do not differ in any significant way. The result for station 25 may be a consequence of a strong azimuth dependence of the station anomaly. This is discussed further below.

The linear term \( a^1 \) is not significant for this particular data set and is, in fact, highly correlated with the \( a^0 \). This is a consequence of the small distance range represented by the data set. Solutions with and without this linear term are indistinguishable in so far as the station anomalies \( T_i \) are concerned.
5. Discussion

The station anomalies illustrated in Fig. 4 have been projected onto the section AA' of Fig. 2. Bouguer gravity anomalies have been projected onto the same section (Fig. 4). A simple relationship between gravity and travel time anomaly is not to be expected: the former, subjected to some low-pass filtering, represents volume integrals of anomalous density structure below and about the station, while the travel times are indicative of anomalous P-wave velocity along the incident ray path. Nevertheless the correlation between gravity and $T_i$ is quite satisfactory. For the Fiji–Tonga events, the ray paths are contained within relatively narrow cones whose axes lie at an angle of ~90° to the section and are inclined by ~20° to the vertical. Early station residuals (negative $T_i$) occur at stations located near the centres of the Musgrave and southern Arunta Blocks where the gravity anomalies reach their maximum values. Late arrivals coincide largely with the gravity lows located near the margins of the basins. There is also a suggestion of relatively early arrivals over the centre of the Amadeus Basin, again coincident with the relative high in the gravity field. The major difference between the two quantities occur at the complex margin of the Musgrave Block with the Amadeus Basin.

Maximum thicknesses of sediment sequences within the basins are of the order of 5 km in the Officer, southern Amadeus and Ngaila Basins and about twice that in the northern Amadeus Basin (e.g., Wells et al., 1970; Pitt et al., 1980; Wells and Moss, 1983; see also Fig. 1). Velocity and density measurements of the near surface crustal rocks are given by Frolich and Krieg (1969), Wells and Moss (1983) and on a variety of expanded-spread seismic reflection profiles shot by the Bureau of Mineral Resources, the South Australia Department of Mines and Energy, and exploration companies. The Officer and Southern Amadeus sediments are predominantly of Proterozoic age. Their seismic P-wave velocities are about (5.4–5.6) km s$^{-1}$ and their densities are mostly in the range of (2.7–2.8) g cm$^{-3}$. The northern Amadeus and Ngaila Basins contain significant thicknesses of Palaeozoic sediments over the Proterozoic sequences and here the velocities and densities are somewhat less than the above values; ~ (4.5–5.0) km s$^{-1}$ and (2.5–2.7) g cm$^{-3}$, respectively. The near surface basement rock of the Musgrave and northern Arunta Blocks along the section A–A' consist predominantly of metasediments, described as granulite–amphibolite transition rocks and granites. The southern part of the Arunta also contains mafic rocks and granulites of igneous origin. Observed seismic velocities from the above-cited sources range from about (5.1–6.0) km s$^{-1}$ while the densities range from (2.6–2.9) g cm$^{-3}$ (Mutton and Shaw, 1979). Little velocity and density contrast is therefore expected between the Proterozoic sequences and the near surface basement rocks but contrasts between the Palaeozoic sequences and this basement may be somewhat greater. In particular, these sediments may contribute as much as (0.15–0.20) s to the delays for stations over the deepest part of the Amadeus and Ngaila Basins relative to the Arunta Block stations. This compares with an observed difference of about 0.7 s between station 16 on the deepest part of the northern Amadeus Basin and station 20 on the southern Arunta. Without more detailed information on the basin sediments we do not apply these corrections.

In keeping with the tectonic model for the region, the simplest interpretation of the travel time anomalies is in terms of an undulating Moho. Let $v_c$, $v_m$ be the crustal and sub-Moho mantle velocities, respectively and $\rho_c$, $\rho_m$ the corresponding densities. Denote the departure, at station $i$, of the Moho from its mean depth by $D_i$. Then, for near vertical incidence

$$T_i = D_i (1/v_c - 1/v_m)$$

As before, $T_i$ is positive for late arrivals and $D_i$ is positive for deeper than average Moho. To a first approximation, the corresponding gravity anomaly at the station is

$$\Delta g_i = -2\pi G (\rho_m - \rho_c) D_i$$

and with (4)

$$-\frac{1}{2\pi G} \frac{\Delta g_i}{T_i} = \frac{\rho_m - \rho_c}{\rho_m - \rho_c} \frac{v_p}{v_c}$$

where $G$ is the gravitational constant. This relation
represents an upper limit to the ratio $\Delta g_i/T_i$ and is approximate for (1) stations that lie away from the suspected thrust fault across which major offsets in the Moho depth may occur, (2) where topographic gradients of the Moho are small, and (3) for near-vertical incidence of the ray path. A more general relation is

$$- \frac{1}{2\pi G} \frac{\Delta g_i}{T_i} = (1 - \beta_i) \cos \alpha_i \frac{\rho_m - \rho_s}{v_m - v_c} v_m v_c$$  \hspace{1cm} (7)

where $\alpha_i$ is the angle of incidence and $\beta_i$ is a correction to the simple Bouguer approximation upon which (5) is based. That is, it would include any effects of a dipping Moho. $\beta_i$ will generally be unknown, to be determined from the data themselves.

For stations close to a near-vertical fault, the gravity anomaly is about one half that given by (5) while the first arrivals could be early or late depending on whether the ray path is predominantly on the 'fast' or 'slow' side of the fault. In the former case $\beta_i$ in (7) equals 0.5 and in the latter case $\beta_i = 1.5$. For inclined faults $\beta_i$ will be more complex.

Figure 5 illustrates the relation between $\Delta g_i$ and $T_i$. For those stations anticipated to lie away from the faults there is, with the exception of station 25, a reasonably linear relationship given by

$$\Delta g_i = -263 T_i - 81 \text{ mGal}$$ \hspace{1cm} (8)

or

$$d(\Delta g_i)/dT_i = -263 \text{ mGal s}^{-1}$$ \hspace{1cm} (9)

With eq. 6

$$v_m v_c (\rho_m - \rho_c/v_m - v_c) = 6.3 \text{ g cm}^{-3} \text{ km s}^{-1}$$ \hspace{1cm} (10)

For those sites near the faults where the observed ratio $\Delta g_i/T_i$ falls below the regression line, the wave paths can be assumed to lie on the fast side of the fault plane. This is the case for stations 4, 5 and 6 at the Musgrave-Amadeus boundary and for stations 17 and 18 at the Arunta-Amadeus boundary. The ratio of these latter two stations agree equally well with eqs. 8 and 9. This suggests that either (1) there is no significant offset of the Moho below these sites, or (2) the wave path crosses the fault. For stations 4, 5, 6, 17, and 18 the predicted ratio, $\Delta g_i/T_i$, is (eq. 6)

$$- \frac{1}{2\pi G} \frac{\Delta g_i}{T_i} = \frac{\rho_m - \rho_c}{v_m - v_c} v_m v_c$$ \hspace{1cm} (11)

and with (10), the predicted gradient is

$$d(\Delta g_i)/dT_i = -131 \text{ mGal s}^{-1}$$ \hspace{1cm} (12)

If the ray path lies through the slow side of the fault then the ratio $\Delta g_i/T_i$ lies above the regression line (8) and the predicted ratio is

$$d(\Delta g_i)/dT_i = 131 \text{ mGal s}^{-1}$$ \hspace{1cm} (13)

This is consistent with the results for sites 1 and 24. Fitting regression lines with gradients given by (12) and (13) through the corresponding, but few, data points gives, for ray paths through the fast side

$$\Delta g_i = -131 T_i - 125 \text{ mGal}$$ \hspace{1cm} (14)

and, for ray paths through the slow side of the fault

$$\Delta g_i = 131 T_i - 125 \text{ mGal}$$ \hspace{1cm} (15)

The observed ratios are not inconsistent with these

Fig. 5. Relation between station residuals and Bouguer gravity anomalies. Stations denoted by solid circles (o) lie away from the major fault zones and line (a) is the linear regression through these points (eq. 8). Stations lying near the fault zones are indicated by the diamonds (O). Those points below the curve (a) indicate that the first arrivals pass through the 'fast' side of the fault. Line (b) is the predicted $\Delta g - T$ relation (eq. 14) for these stations. For those stations lying above the line (a) the seismic first arrivals pass through the 'slow' side of the fault and the predicted $\Delta g - T$ relation (eq. 15) is given by the line (c). The anomalous station residual (l) is for station (25).
simple models (Fig. 5) despite the likely complexity of the actual wavepaths and the small number of data points.

Of the parameters on the left hand side of (10) the best known is \( u_m \), the mantle velocity immediately below the Moho. Values of (8.1–8.3) km \( s^{-1} \) are appropriate. Lower and intermediate crustal velocities have been measured to the north of the region by Hales and Rynn (1978) who obtained 6.0 km \( s^{-1} \) down to a depth of about 12 km and 6.5 km \( s^{-1} \) for the lower crust. In so far as surfaces of density and velocity contrast can be assumed to follow the warping of the Moho, the appropriate crustal velocity in the above relation is 6.0 km \( s^{-1} \). Then, with \( u_m = 8.2 \) km \( s^{-1} \), the results (7) and (9) and an average incidence angle of \( \alpha = 23^\circ \),

\[
\frac{(v_m - v_c)}{(\rho_m - \rho_c)} = 7.2 \text{ km s}^{-1}/\text{g cm}^{-3}
\]

If \( v_m = 6.5 \) km \( s^{-1} \) is adopted, implying that the intermediate crustal boundary does not follow the Moho, then

\[
\frac{(v_m - v_c)}{(\rho_m p_c)} = 7.8 \text{ km s}^{-1}/\text{g cm}^{-3}
\]

Both ratios are consistent with Bott’s (1971) extension of the Nafe–Drake density-velocity curve to metamorphic and igneous rocks, but larger than Anderson et al. (1972) rendition of Birch’s law for the mantle for which \( \frac{dv}{d\rho} \approx 6 \text{ km s}^{-1}/\text{g cm}^{-3} \) at temperature of \( \sim 1000^\circ \text{C} \) (see their fig. 4). This latter discrepancy can be readily attributed to the assumption \( \beta = 0 \). In view of the many assumptions made, and the limited available data set, it is not particularly helpful to pursue this argument further. In particular, we cannot discriminate between the models (16) and (17).

With the relations (8), (10), (14) and (15) it becomes possible to estimate the Moho depth from either the gravity- or travel-time anomaly data. That is

\[
D_m = \frac{L_s}{\beta_1/(\rho_m - \rho_c)}
\]

or

\[
D_m = T_c \cos \alpha / \frac{\alpha \beta_1 / (v_m - v_c)}
\]

where \((\rho_m - \rho_c)\) and \((v_m - v_c)\) must satisfy the conditions (16 and 17). For \( v_c = 6.0 \text{ km s}^{-1} \) and \( v_m = 8.2 \text{ km s}^{-1} \), \( \rho_m - \rho_c = 0.31 \text{ g cm}^{-3} \) (eq. 16), and the maximum Moho undulation predicted by the model is 16 km. This increases to 22 km for \( v_c = 6.5 \text{ km s}^{-1} \) and the same \( v_m \). Figure 4 illustrates this result for the undulations relative to a nominal mean depth of 40 km. The positions of the major faults are schematic only since a unique solution is not possible from this data set alone. Further analyses of events at different epicentral distances and azimuths may help to define the orientation of these faulted zones.

6. Conclusions

The observed travel time anomalies in central Australia are consistent with the mechanical crustal deformation model of Lambeck (1983a); both in terms of variations of the Moho depth by possibly as much as 22 km and in terms of the occurrence of major discontinuities in the Moho depth at a number of places where major faulting has been predicted by the basin evolution model. Analyses of travel times at different azimuths should, to-
gether with ray tracing, give a more exact indication of the orientation of the fault planes. A preliminary examination has been made of travel times from northern and southern earthquakes but here uncertainties in the travel-time curve become an important factor in determining the station residuals because of the increased distance range. More important is that for these events the residuals are strongly influenced by any dip on the Moho. The proposed model indicates that dips of 10° may occur and these could contribute as much as 0.2 s to the station anomalies. The above estimates of the undulation of the Moho represents an upper limit due to the neglect of contributions to both gravity and travel times arising from (1) crustal structure and (2) possible variations in the mantle velocities and densities immediately below this variable depth Moho. The first neglect can be remedied only when more detailed information on the basin structure and properties of the near surface rocks become available. The second neglect can be remedied by an iterative approach to the solution, but until the azimuthal variations have been examined in detail it is premature to do so now.

The crustal structure illustrated in Fig. 4 raises a number of questions central to understanding the properties and evolution of continental crust. What are the rheological implications of this model? How has the non-hydrostatic state been maintained subsequent to its formation from Late Proterozoic to Early Carboniferous time? What is the origin of the compressive force? These questions have only been partially answered in Lambeck (1983a, c) and the present results do not contribute to this discussion other than to provide support for the general characteristics of the model. What is needed, inter alia, is deep crustal seismic reflection work across some of the principal structures. An obvious test, for example, is whether the Moho is indeed at a shallow depth below the southern Arunta and Musgrave when compared with the basins. For the two locations where deep crustal reflections have been obtained, below the northern Amadeus and Ngalla Basins, the model predicts approximately the same Moho depth and this is indeed observed (Brown, 1970; Moss, 1964). A second test is to attempt to delineate the faults beyond 2 s two-way travel time. Reflection surveys carried out by the South Australia Department of Mines across the northern margin of the Officer Basin confirm the presence of a major fault down to at least 3 s, about 1 s beyond the basin floor (see also Milton and Parker, 1973). Reflection surveys across the northern margin of the Ngalla Basin indicate similar thrust faults down to at least 2 s (Wells and Moss, 1983). No reflection information is available across the southern margin of the Amadeus. Here the model predicts significant overthrusting, possibly more so than occurs at the northern margin of this basin (see also Fig. 1). The surface expression of this fault should lie, below the thin Cainozoic cover somewhere to the north of site 6. A deep crustal reflection survey here may also help elucidate aspects of the model.

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References