Imaging crustal structure variation across southeastern Australia

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1. Introduction

The crustal structure (crustal thickness and nature of the basement) of tectonic units is not well constrained in southeastern Australia due to the presence of sedimentary rocks and of thick regolith across most of the surface. Although active seismology can provide the most reliable information on crustal architecture, such information is confined to limited regions. A complementary source of information comes from passive seismic techniques such as receiver functions utilised in this study.

The Gawler Craton contains less than 5% basement exposure in an area of roughly 530,800 km² (Payne et al., 2009). Unravelling the nature of the crustal structure and the lateral extent of crustal provinces is necessary to improve understanding of the evolution of southeastern Australia. The primary aim of this study is therefore to investigate the lateral variation of crustal structure and nature of the crust to mantle transition between the Precambrian and Phanerozoic belts of southeastern Australia in a corridor extending from the Gawler Craton to the east coast of Australia.

Since 2004 more than 7000 km of full-crustal reflection profiles have been collected across Australia providing information on the variation in Moho depth. However, the estimates of crustal thicknesses based on reflection data rely on a simple assumption of an rms velocity of 6 km/s at the base of the crust (e.g. Drummond et al., 2006; Kennett et al., 2011). The secondary goal of our study is therefore to calibrate Moho depth estimates from recent seismic reflection profiles carried out in southeastern Australia (Cayley et al., 2011; Drummond et al., 2006; Fraser et al., 2010; Preiss et al., 2010) using our receiver function measurements, which are particularly sensitive to the gradients in seismic velocities.

1.1. Tectonic setting

The Lachlan Orogen in southeastern Australia (Fig. 1b) was accreted to the Precambrian core of the continent in a sequence of stages (e.g., Braun and Pauselli, 2004; Cayley, 2011; Collins, 2002; Direen and Crawford, 2003; Gray and Foster, 2004). We show in Fig. 1 a...
simplified model of a geological map for southeastern Australia. The map area is separated in two parts: (i) the western part formed by Precambrian units: the Gawler Craton, the Curnamona Province and the Adelaide Geosyncline and (ii) the younger eastern part with Phanerozoic terranes. The Gawler Craton contains Archean–early Palaeoproterozoic supracrustal and magmatic lithologies, which are surrounded, overlain and intruded by Palaeoproterozoic (2000–1610 Ma) to Mesoproterozoic (1590–1490 Ma) units (e.g. Payne et al., 2009). The basement of the Curnamona Province is of late Palaeoproterozoic age. Palaeozoic deformation affected the marginal...
zones of the Province, but the central core is cratonic. The Adelaide Geosyncline is Neoproterozoic to Middle Cambrian basin complex with deep subsidence which experienced at least five major successive rift cycles (Preiss, 2000).

The large-scale structural elements of the eastern Phanerozoic part are mainly:

- sedimentary basins with Mesozoic and Cenozoic cover: Murray Basin, Otway Basin, Gippsland Basin and Sydney Basin,
- three orogens: the early Palaeozoic Delamerian Orogen (350–470 Ma), the Middle Palaeozoic Lachlan Orogen (450–340 Ma) and the Late Palaeozoic to Early Mesozoic New England Orogen,
- and the Newer Volcanic Province, a wide volcanic field site (around 15,000 km²) of over 400 monogenetic vents, which represents the product of the most recent volcanic activity in Australia, spanning from the Pliocene to the Holocene (e.g., Lesti et al., 2008). The magma reservoirs were estimated to be subcrustal and below 32 km based on the spacing and distribution of the density maxima (Lesti et al., 2008). The Mount Gambier volcanic sub-province is the youngest region of this intraplate continental plain basalt province. Mount Gambier has experienced recent volcanic activity: with significant volcanism at about 4000 years B.P. (Blackburn et al., 1982).

The Phanerzoic part of Australia was formed after the break-up of Rodinia around 780 Ma (Wingate et al., 1998), followed by the growth of orogenic belts along the eastern margin of Gondwana (e.g., Gray and Foster, 2004). Plate convergence in an oceanic setting along the eastern margin of Gondwana from ca. 520 Ma to 340 Ma produced crustal thickening (e.g., Gray and Foster, 2004). In the Mesozoic, southeastern Australia was the continental margin of the subducting Pacific plate.

The boundary between the Precambrian and Phanerozoic parts of Australia is often referred to as the Tasman Line. This was initially defined in Queensland as the boundary line between the craton and geosyncline in early Palaeozoic times (Direen and Crawford, 2003; Hill, 1951). The definition was based on the limited outcrop of the Precambrian rocks available at the surface. The recent reinterpretation by Direen and Crawford (2003) proposes that the Tasman Line is not a simple “line” associated to one single tectonic event: the Rodinia-Gondwana breakup, and they found no outcrop or geophysical evidence of rocks dating from the period 780 to 730 Ma at the locations of the Tasman Line. The main locus of Rodinia breakup is possibly to the east.

1.2. The use of receiver functions

The coda of teleseismic P waves contains considerable information about the structure of the Earth directly beneath a seismic station. The P-receiver function technique isolates P-to-S conversions generated at crustal discontinuities beneath the recording site (Langston, 1977, 1979) by source equalisation. Conversion primarily occurs at the Moho, which represents one of the most significant interfaces (taking into account the possible presence of sedimentary layers beneath some seismic stations) in terms of the variations in elastic properties. The principle underlying the receiver function technique is to concentrate on conversions beneath the station by removal of the instrument response and the effect of passage through the bulk of the mantle through deconvolution of different components of ground motion. In this study, we use the radial receiver function obtained by deconvolving the radial component (along the great-circle to the source) by the vertical component, as a means of investigating the nature of the crust–mantle transition beneath southeastern Australia and crustal thickness. Receiver functions are sensitive to shear velocity contrasts of interfaces located beneath the seismic station, however, they are only weakly sensitive to absolute velocities (e.g., Julià et al., 2000; Tkalcić et al., 2006). The S-velocity structure would be better resolved with a joint inversion of radial receiver function and surface wave dispersion observations (e.g. Julià et al., 2000; Tkalcić et al., 2006).

1.3. Seismic reflection data acquisition and processing

The set of seismic reflection profiles is fully described in Drummond et al. (2006), Fomin et al. (2006, 2010), and Cayley et al. (2011). The profiles have been recorded in a consistent way using the same equipment, with 18–20 s two-way time at a sample rate of 2 ms. Three in-line vibrators, each applying 250,000 N of peak force, were used as the energy source. Three upswipes of 12 s duration using the Varisweep technique were used at each vibration point. Vibroseis sweeps are typically in the range 6 to 96 Hz. Geophone groups of 12 phones were spaced at 40 m interval.

A typical processing sequence is (e.g. Drummond et al., 2006):

1) refraction statics corrections,
2) velocity analysis,
3) pre- or post-sort spectral equalisation,
4) sort to common mid-point gathers,
5) dip moveout corrections, stack,
6) post-stack migration,
7) band-pass filtering,
8) and in some cases, semblance filtering before display.

1.4. Nature of the seismic boundary between the crust and the mantle: the Mohorovičić discontinuity

The Moho is commonly defined in seismology as the boundary that marks an abrupt increase in P-wave velocity to values typical of the upper mantle (e.g., for the studies of the Australian region, see Clitheroe et al., 2000; Collins et al., 2003). However, different definitions have been used for the value of the upper mantle velocity: e.g. Vp > 7.6 km/s (Clitheroe et al., 2000); Vp > 7.8 km/s (Collins et al., 2003). The crust–mantle transition shows noticeable variations in character across Australia, with both sharp and gradational transitions reported (e.g., Clitheroe et al., 2000; Kennett et al., 2011; Reading et al., 2007; Shibutani et al., 1996). The seismic discontinuity between the crust and the mantle may not always correspond to a petrologic boundary (e.g., O’Reilly and Griffin, 1985; and discussion in Clitheroe et al., 2000). Furthermore, the origin of the Moho could be multi-genetic for continental crust (e.g., Eaton, 2005). Reflection and refraction profiles demonstrate that the crust–mantle boundary can show considerable variation. Sometimes, there is no clear boundary at the base of the crust suggesting that the impedance contrast between crust and mantle is weak. The nature of the transition from crust to mantle is reflected in the nature of the reverberations and conversions occurring beneath the station. Thus, Zheng et al. (2008) have shown that a thick transition between the crust and mantle results in diffused and weakened PpPns amplitude (the first Moho multiple) in receiver functions, while a sharp crust–mantle boundary generates strong PpPns amplitude.

2. Data analysis

2.1. Information on station deployments

The seismic data are derived from a network of 28 portable three component broad-band seismometers (Fig. 1) deployed for 9 months at a time for the SoCP (Southern Cratons to Palaeozoic) experiment and from 4 permanent broadband seismometers from the Geoscience Australia network: BBOO, ARPS, TOO and STKA in the same area. We also analyze data from one GEOSCOPE permanent seismic station at Mount Stromlo (CAN) near Canberra. The data from this study have been employed in the construction of the recent model of Moho depth in Australia: AusMoho (Kennett et al., 2011). Here we present a specific interpretation of the data in the light of the seismic measurements (receiver functions and reflection profiles) carried out in southeastern Australia.
2.2. Data and stacking

We used P waves from epicentral distances between 30 and 90° for events with \( m_b \geq 5.5 \), hand-selected for good signal-to-noise ratio (\( SNR \geq 2 \)) and cut the seismic traces at 5 s prior to and 30 s after the initial P-wave arrival. The seismograms were rotated onto the azimuth between receiver and source. A Gaussian low pass filter was applied to the waveforms before the deconvolution is performed to eliminate high frequencies (the Gaussian width is 2.5 rad/s) in order to eliminate influence from small-scale crustal heterogeneities. The P radial receiver functions (RFs) were obtained by deconvolving the radial component, directed along the great-circle between source and receiver, with the vertical component (Fig. 2) in the time domain (Ligorria and Ammon, 1999). The RF waveform was sampled at a frequency of 10 Hz giving a total of 351 data samples. The radial RFs at each seismic station were then stacked for a set of back-azimuths, with a narrow range of ray-parameters based on the following procedure:

1. Select the quadrant (back-azimuths between N0° and N90°, N90° and N180°, N180° and N270°, N270° and N360°) with the highest number of RFs.
2. Compute \( p_{\text{median}} \): the median of the ray parameters of all seismic events in this interval.
3. Select events with a ray parameter \( = p_{\text{median}} \pm 0.004 \) (s/km). Most data come from seismogenic belts surrounding Australia and this narrows down the range of useful ray parameters. For example, the useful ray parameter range for station CAN is between 0.067 and 0.075 s/km.
4. Stack the RFs selected in the previous step. Only the best data are used for stacking and we focused on obtaining the most basic information assuming a horizontally layered structure.

We show in Fig. 2 an example of individual RF traces we used for the stack. Before each stack we examined the coherency of individual RFs using the cross-correlation matrix approach from Tkalčič et al. (2011): we compute the cross-correlation coefficients for each pair of RFs. Then RFs are selected or rejected using empirical values for the minimum coefficient of cross-correlation \( \chi \) and the percentage of coherent RFs \( \tau \). More specifically, a single RF is deemed suitable for a stack if cross-correlation coefficient is higher or equal to \( \chi \) with at least \( \tau \) per cent of other RFs. In this study, for most stations, the small number of RFs (<50) results in \( \chi \approx 0.70 \) and \( \tau \approx 10\% \) or less. In the case of station SO07, \( \chi \approx 0.85 \) (Fig. 2) and we found insignificant difference of crustal thickness derived from the NA inversion of a single RF and the inversion of the stacked RF at the same station.

For station CAN, for which we have a larger number of RFs, we considered a higher value of \( \chi \): 0.90. The stacking process may induce a loss of some features of the crust–mantle boundary particularly associated with the detection of dipping and anisotropic structures that depend on the back-azimuthal variation of the RF amplitude. When the stacking bounds are tight in back-azimuth (≤10°) and epicentral distance then the signal should not be degraded (e.g. Cassidy, 1992). At CAN we have such tight bounds, but for all other stations we could miss information relevant to a dipping Moho or the presence of crustal anisotropy, as we are stacking over all RFs in one quadrant (90° variation of back-azimuth). Investigating the transverse RF variation with back-azimuth can help to determine complexity beneath a seismic station (e.g., Frederiksen et al., 2003). However, unfortunately with a temporary seismic deployment of 9 months at a time for the SoCP experiment it was not possible to obtain a broad back-azimuthal coverage in southeastern Australia. Due to this

![Fig. 2](image-url) Observed radial RFs and stacked RF for station SO07. Dashed lines show ±1 standard deviation bounds around the linear stack. Synthetic RF resulting from the NA inversion is shown at the bottom.
limitation we only consider isotropic and planar discontinuity to compare the results from all seismic stations across southeastern Australia.

2.3. Neighbourhood Algorithm inversion for 5 crustal layer and 1 mantle layer models

We characterise the crust and uppermost mantle structure with a 6-layer 1-D seismic velocity model: sediment layer, basement layer, upper crust, middle crust, lower crust, and uppermost mantle (Shibutani et al., 1996). A nonlinear inversion method, the Neighbourhood Algorithm (NA, Sambridge, 1999a, 1999b) was employed to match the observed radial RFs. During the inversion, the synthetic radial RF was calculated using the Thomson–Haskell matrix method (Haskell, 1953, 1962; Thomson, 1950). The full effects of free-surface reverberations and conversions were modelled. The NA is a direct-search method of inversion. It has the capacity to search efficiently by sampling simultaneously different regions of parameter space. This inversion technique samples regions of a multidimensional parameter space that have acceptable data fit. The NA makes use of simple geometrical concepts to search a parameter space. At each iteration, the entire parameter space is partitioned into a set of Voronoi cells (Voronoi, 1908) constructed about each previously sampled model. In our case Voronoi cells are nearest neighbour regions defined by an L2-norm. The initial sets of samples are uniformly random, but as iterations proceed only a subset of chosen Voronoi cells are re-sampled (using a random walk within each cell). This allows the NA to concentrate where data misfit is lowest. A further advantage of the NA over other direct-search methods is that only the rank of the misfit function is used to compare models. This is of particular significance as it avoids the problems associated with scaling of the misfit function and allows any type of user-defined misfit measure to be employed. The NA requires just two control parameters:

- \( n_s \), which is the number of models generated at each iteration
- and \( n_n \), which is the number of neighbourhoods re-sampled at each iteration.

After several trials, we specify a maximum number of iterations: 900 and \( n_s = 100 \) for the first iteration and \( n_s = 25 \) for all other iterations and \( n_n = 25 \) for all iterations. The a priori distribution of the misfit function is assumed to be Gaussian (e.g. Sambridge, 1999a, 2001). In this study, we used a chi-square metric similar to that employed by Sambridge (2001) to compute the misfit function. The misfit function, an L2-norm, is defined as the sum of the squares of the difference between the observed amplitude of the radial RF and the amplitude of the synthetic radial RF from a 6-layer model.

One of the advantages of the NA over other direct search methods is that only the rank of the misfit function is used to compare models and this enhances the exploration of parameter space to find suitable models. Fig. 3a shows an example of the misfit function obtained using the NA at station SO07. The NA inversion provides a 1-D shear wave speed model and an estimate of the \( V_p/V_s \). However, neither the \( V_p/V_s \) ratios nor the absolute velocities are strongly constrained in the NA inversion, which is most sensitive to the presence of discontinuities. During the inversion, as in the work of Shibutani et al. (1996) the model was parameterised in terms of 6 layers with internal velocity gradients and the possibility of discontinuities at the boundaries. This leads to the 24 parameters, the \( V_s \) values at the top and bottom of the gradient zone, the thickness of the gradient zone and the \( V_p/V_s \) ratio in each zone. We used similar bounds for the 24 parameters to those of Shibutani et al. (1996) and Clitheroe et al. (2000) (Table 1). The NA method combines a Monte Carlo search technique and the properties of the Voronoi geometry in parameter space to find an ensemble of the best fitting models and performs a global optimization. We present density plots of the best 1000 data fitting 5-velocity models generated by the Neighbourhood Algorithm in Figs. 3b and 4. The model with the best fit to the data is plotted in red. Sambridge (1999b) has used a Bayesian approach to estimate resolution and confidence intervals on the 24 parameters in the RF inversion. The layer thickness distribution and the S-velocity distribution are better constrained by the NA inversion than the \( V_p/V_s \) ratio, thus we place less reliance on the \( V_p/V_s \) ratio. Sambridge (1999b) showed using marginal probability density functions that the \( V_p/V_s \) ratio from the NA inversion is better resolved in the first layer and poorly resolved in the remaining layers. The inclusion of the \( V_p/V_s \) ratio serves primarily to allow for some of the effects of the sedimentary layer beneath the stations with no a priori information (Bannister et al., 2003).

3. Results

3.1. Crustal thickness and nature of the Moho

3.1.1. Crustal thickness

We assume that the base of the transition to mantle velocities defines the Moho depth, in order to be in accordance with previous RF studies (e.g., Clitheroe et al., 2000) which produced Moho depths close to estimates from seismic refraction studies (Collins, 1991; Collins et al., 2003). Here, we take the upper mantle velocity to be \( V_p \geq 7.6 \text{ km/s} \) following Giese (2005), which means that \( V_s \geq 4.3–4.4 \text{ km/s} \) for \( V_p/V_s \) ratio in the range 1.73–1.77 at the base of the gradient. Fig. 4 presents the shear wave velocity models from the NA inversion and data fits at SO01, CAN, MUR5 and MG01. The S-velocity models from the inversion at the other seismic stations are available as an electronic supplement. The results of the RF analysis from the NA inversion are summarised in Table 2 and Fig. 5a and b. The uncertainty in Moho depth is ±2 km. To assess the accuracy of our results compared to other styles of RF analysis, we also perform the same analysis at one seismic station in Hawaii: KIP and obtained a small difference of 1 km of crustal thickness compared to Leahy et al. (2010) who used an improved H–K stacking approach. We obtain a Moho depth to be around 18 km from the RF inversion whereas Leahy et al. (2010) obtained a Moho at around 19 km depth beneath this station.

3.1.1.1. Lachlan Fold Belt or Lachlan Orogen. The Moho depths show a generally thick crust ranging from 34 up to 51 km. These values are close to those obtained by Clitheroe et al. (2000) and Collins et al. (2003) at several localities. Taking into account crustal thickness estimates from previous studies (Clitheroe et al., 2000; Collins, 1991; Collins et al., 2003; Shibutani et al., 1996), the average crustal thickness is thicker beneath the Lachlan Orogen ~43 km than beneath the cratons. The lower crustal structure obtained at CAN corresponds to a broad velocity transition zone at the Moho and a crustal thickness ~48 km. This model is fully consistent with previous refraction and RF studies (Collins et al., 2003).

The Moho depth obtained from RFs at station ARPS (34 km) is close to the value (around 35 km) estimated by Korsch et al. (2002) using a seismic-reflection transect close to this seismic site.

3.1.1.2. Murray Basin. The crust is thinner than elsewhere in the studied area. The average crustal thickness is ~32 km. The crustal thickness varies between 28 km (for MUR3) and 35 km (for MUR5). The highest crustal thickness is observed at MUR5 and this may be due to the proximity of the Lachlan Orogen.

3.1.1.3. Gawler Craton and Curnamona Province. Our study suggests a crustal thickness ranging from 28 to 45 km. The crust is thicker than east of the cratons in the Delamerian Orogen and in the Murray Basin. The Gawler Craton and the Curnamona Province show a thick crust (≥40 km) at stations: SO07, BBO0, SO08, SO10, SO11, SO12, SO13 and SO17. The values ≥40 km observed at SO07, BBO0 and
SO08 are located on the Gawler Craton whereas the estimates for SO12 and maybe SO17 are associated with the Curnamona block. The crustal thicknesses at SO07, BBOO and SO08 are greater than 40 km and could be related to their location in the Gawler Craton. Interestingly, the crustal thickness decreases at SO08 (Fig. 5a) and then increases at SO12 and SO17 (Fig. 6). We also note that the shear wave velocity variations are similar beneath BBOO and SO08 (Fig. 6) and different from the structure obtained at SO09. The 1-D shear wave velocity models show similar trends beneath SO17 and SO13. The velocity structure of the crust at SO17 is more complex than at SO12. This difference may be due to the fact that the Curnamona Province does not extend to SO17. The average crustal thickness is found to

Fig. 3. a) Misfit function against number of models for which the forward problem has been solved using the NA approach. b) The 1-D shear wave velocity models obtained by the inversion of teleseismic RFs at SO07. All 22,600 models searched in the NA inversion are shown as gray shaded area. The best 1000 models are shown in yellow and green, with the colour being logarithmically proportional to model number. The colour scale shows the increase in data fit from yellow to green. A solid red line represents the best fitting model. 0 km depth corresponds to the station elevation. H is the Moho depth value.
be around 39 km beneath the Precambrian Gawler and Curnamona units. The average Moho depth obtained in this study for the Curnamona Province: 39.5 km (for SO12 and SO16) is similar to the results from seismic reflection experiments in the same region (across the Broken Hill and Olary domains) around 39 km (Gibson et al., 1998; Coleby et al., 2006). Recent reflection profiles in the Curnamona region indicate little changes in the depth to the crust–mantle boundary despite contrasts within the crust.

### 3.1.1.4. Mount Gambier region

The average crustal thickness is 44 km. Significant crustal thickness changes (from 37 to 47 km) are observed in Mount Gambier region over relatively short distances — around 70 km. Stations MG01 and MG06 show a thick crust ≥40 km. The smaller crustal thickness estimated at MG02 could be due to the proximity to the coast and thus the zone affected by the breakup of Australia and Antarctica. We note a relatively good correlation between Moho depth and the surface topography (Fig. 5a).

The complexity of the RF workflows for 6 seismic stations (MG03, MG04, MG05, MUR1, MUR4 and SO14) prevented us from using the NA inversion to obtain accurate information about velocities and Moho depths. Large-amplitude ringing can be generated by a low-velocity layer such as a sedimentary layer underlain by a strong contrast in wave speed (e.g., Clitheroe et al., 2000). The ringing may be due to near surface limestones with surface sediments. We observed a jump in Vp/Vs in the topmost layer of the best models at stations MUR5 and MG06 with ratios higher than 2. This jump is probably due to a sedimentary layer below the sensor and was already observed by Shibutani et al. (1996) in Eastern Australia for two stations located on sedimentary basins (SB01 and SA07). Similar high Vp/Vs values (higher than 2) related to sediments were already described in the literature (e.g. Morozov et al., 2002; Yang et al., 2012; Zelt and Ellis, 1999).

For some seismic stations a sedimentary layer was detected, but we were also able to detect the Moho depth (see for instance at MUR5). At MUR3 seismic station (a less constrained location), a high Poisson’s ratio in the top layer is responsible for strong reverberations, which mask phases such as multiples. The velocity model resolution is reduced in comparison to stations on bedrock. Depth to major crustal velocity discontinuities and the Moho are the best-constrained features of this kind of model.

### 3.1.2. Nature of crust–mantle transition

Except for the stations installed in the Murray Basin, most sites show a crustal thickness greater than 36 km. Using the character of the crust–mantle transition in the results of the modelling (Fig. 4e, f, g and h) we classify the Moho transition zone as sharp if its thickness is ≤2 km, intermediate for 2–10 km, and broad if it is ≥10 km, as suggested by Shibutani et al. (1996). Our Moho estimate will lie at the base of a gradient (in conformity with earlier work (e.g., Clitheroe et al., 2000)).

Except for two seismic stations SO15 and STKA, which are close to the limit between the Precambrian and Phanerozoic terranes, most seismic stations located in the Precambrian region have an intermediate crust–mantle transition zone. All seismic sites in the Eyre Peninsula (Fig. 5) in the Gawler Craton show an intermediate nature (SO07, BBO0, and SO08).

The crust–mantle transition at MUR3 (in the Murray Basin) is sharp whereas it is intermediate at MUR2 and broad at MUR5. The stations MG01 and MG06 show a thick crust ≥40 km with a broad crust–mantle transition, whereas the crust is thinner at MG02 (37 km) with a sharp Moho transition zone. The crust–mantle boundary is deep and mostly intermediate in character beneath the Lachlan Fold Belt.

### 3.2. Calibration of reflection profiles

In recent years a number of reflection profiles have been carried out in southeastern Australia that provide valuable continuity between localised crustal properties that we can extract from RFs. However, there is limited velocity control on these full-crustal reflection profiles. As in the work of Kennett et al. (2011), we estimate Moho depth from the reflection results using an average crustal velocity of 6000 m/s, which gives a depth conversion of 3 km per 1 s two-way time (TWT). Kennett et al. (2011) showed that such a simple approximation provides good agreement between seismic reflection and RF observations in western Australia (Reading et al., 2007, 2011) with small discrepancy (around 1 km) and also between seismic reflection and RF measurements in southern Queensland. The influence of different stacking velocities is described by Jones et al. (2005). They found that for full crustal reflection images the nature of the records shows little dependence on the stacking velocities. We compare the results from the reflection work with the RFs from the stations closest to the profiles.

A west–east seismic profile was carried out near the Mount Gambier region in 2009: GA-09-S1D (Figs. 7 and 8a). A significant thickening of the crust is observed towards the Mount Gambier area at the west of the line where the Moho reflection is at ~15 s (Cayley et al., 2011). This deep Moho is very similar to that found from the RF at MG01 station (47 km). There is a relatively rapid change in Moho depth to the east, shallowing to 33 km which is also consistent with the depth of 34 km obtained at ARPS.

A group of reflection profiles have been carried out to investigate the crustal structure of the Gawler Craton and the Curnamona Province and the linkages between them. In 2003, a north–south reflection profile was acquired in the Gawler Craton, near the major ore-deposit at Olympic Dam. The seismic data from the north–south seismic profile, 03GA-D01, are represented in Fig. 8b (Drummond et al., 2006). The bottom of the reflection Moho is around 14 s TWT (~42 km) in the northern zone and 13 s TWT in the southern zone suggesting a crustal thickness of ~39–40 km. In 2008, an east–west reflection transect (08GA-G1) was performed in the Eyre Peninsula, Gawler Craton in order to investigate the crustal architecture of the eastern part of the Gawler Craton (Figs. 7 and 8c). The central and western parts of the profile show a Moho at ~14 s TWT: 42 km (Fraser et al., 2010) whereas the eastern part indicates a Moho at ~41 km (Fig. 8c). In 2009, an east–west transect (09GA-CG1) was acquired across the Flinders Ranges, South Australia. The primary goal of this survey was to elucidate the relationship, at depth, between the Curnamona Province and the Gawler Craton (Preiss et al., 2010). The Moho is poorly imaged (Fig. 8d). At the eastern end of the line the lower crust is mostly nonreflective. Weak reflections suggest a Moho depth ~13 s TWT (39 km). At the western end of the line
Table 2
Station location and results of the RF modelling with the NA inversion. Inversion quality (Quality) is assessed visually and with the misfit value. SO02 seismic station provided no clear RFs. Note that SO06 and CAN station were both at Mount Stromlo. H is the crustal thickness and N is the number of RFs used for the stacking at each station. Nature represents the character of the Moho from the shear-wave velocity model obtained at each station: thin $\leq$ 2 km, intermediate 2–10 km and broad > 10 km.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude</th>
<th>Longitude</th>
<th>H (km)</th>
<th>N</th>
<th>Quality</th>
<th>Nature</th>
</tr>
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<td>ARPS</td>
<td>−36.7699</td>
<td>141.8383</td>
<td>34</td>
<td>17</td>
<td>Intermediate</td>
<td>Broad</td>
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<tr>
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<td>148.9990</td>
<td>48</td>
<td>29</td>
<td>Intermediate</td>
<td>Broad</td>
</tr>
<tr>
<td>MG01</td>
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<td>0.5</td>
<td>Broad</td>
</tr>
<tr>
<td>MG02</td>
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<td>37</td>
<td>2</td>
<td>0.5</td>
<td>Thin</td>
</tr>
<tr>
<td>MG06</td>
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<td>140.8707</td>
<td>40</td>
<td>8</td>
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The average crustal thickness obtained in the present study from the broad-band station in southeastern Australia is 39 km (and the median value is also 39 km). This value is similar to: (i) the 38 km average thickness of continental crust in the CRUST 5.1 model of Mooney et al. (1998), a model of global crustal thickness based on previous seismic refraction data, (ii) the value of 38.8 km computed by Clitheroe et al. (2000) from earlier RF analyses across the whole Australian continent, and (iii) the value of ca. 38 km for continental crust in Australia deduced by Collins et al. (2003) using a compilation of both RF studies and seismic refraction. These crustal thickness estimates have been incorporated into the compilation of Kennett et al. (2011) that provides a significant improvement over the resolution of previous Moho maps (e.g., Clitheroe et al., 2000; Collins et al., 2003; Mooney et al., 1998). The individual values add considerable detail for Southern Australia.

We assess the uncertainty of our estimates of Moho depths by using the density plots of the best 1000 data fitting shear-wave velocity models (Fig. 4e, f, g and h) and the nature of crust–mantle transition (if it is sharp, intermediate or broad). The estimated error in Moho depths at all seismic stations is less than $\pm$ 2 km. The crust–mantle boundary is the deepest beneath the Lachlan Fold Belt suggesting the presence of underplating materials at the base of the crust. Interestingly, even though the Gawler Craton contains very old material before 2560 Ma (Swain et al., 2005), we also note a thick crust $\geq$ 40 km beneath all seismic stations located in the Gawler Craton. This thick crust may be a characteristic of the Gawler Craton crust.

4.1. Nature of crust–mantle transition
We could not observe clear multiples on the RFs at most of our seismic stations. Chevrot and van der Hilst (2000) also pointed out the absence of clear multiples in our region of study. This feature suggests the absence of sharp seismic interface at the crust–mantle boundary beneath this region.

4.1.2. Gawler Craton and Curnamona Province
The intermediate nature observed beneath the Gawler Craton is consistent with previous result from Clitheroe et al. (2000) for the station BBOO. Such intermediate nature of the crust–mantle transition suggests magmatic underplating at the base of the crust. The presence of magmatic underplating is also suggested beneath other cratons.

The mafic underplating could result from two different processes:

i) a deep crust delamination with a change of the nature of crust–mantle transition due to lithospheric mantle partly or entirely added from the ambient convecting mantle.

ii) passage of the continent overtop a mantle upwelling resulting in the thickening of the crust–mantle transition.

4.1.3. Evidence of intermediate or broad transition beneath other cratons
Zandt and Ammon (1995) using measurements of crustal Poisson’s ratio at all continent except Antarctica suggested that magmatic underplating could be involved in the formation of all cratons. A lower-crustal 10–25 km thick high velocity layer was observed beneath the Archean Wyoming Craton from seismic refraction profile (Eaton, 2005; Gorman et al., 2002) at the south end of the profile. This layer was interpreted as thick magmatic underplating. An intermediate nature of crust–mantle transition was also observed beneath the Proterozoic North Australian Craton and interpreted as a result of the presence of underplating in the lower crust possibly associated with subduction (Clitheroe et al., 2000). A thick crust–mantle transition of about 5–10 km was also observed beneath the Tainghsan region in the northern North China Craton (Zheng et al., 2006, 2008) and it is interpreted as regional delamination of the deep crust after orogenic thickened and/or magmatic underplating.

4.1.4. Mount Gambier
Recent reflection profiles near MG01 (Fig. 8a) show a complex structure beneath this region with a rapid and substantial thickening of the crust towards the location of MG01 and a possible imbricate structure (with an apparent duplication) at the base of the crust. This class of structure may explain the complexity in the observed RFs for MG01 and could be the reason for only a moderate fit (Fig. 4d) with 1-D synthetics. The thick crust in the Mount Gambier region suggests the extension of the Gawler Craton beneath Mount Gambier and the imbricate structure may be linked to convergence during the Cambrian age Delamerian Orogeny.

4.1.5. Murray Basin
At MUR2 and MUR5, no clear PpPms arrivals are visible on the RFs, which is consistent with a thick Moho transition zone beneath these
two stations. The intermediate and broad nature of the crust–mantle transition beneath MUR2 and MUR5 could be due to the proximity of these two stations with the Lachlan Orogen and of underplating. The sharp nature of the crust–mantle transition observed at MUR3 suggests direct contact of the crust on top of the upper mantle with no underplating materials.

4.1.6. Lachlan Orogen

Our results are consistent with previous observations (Clitheroe et al., 2000; Collins et al., 2003; Shibutani et al., 1996). The variations in the crustal thickness and the intermediate and broad transition between crust and mantle beneath the Lachlan Orogen may be related to the presence of magmatic underplating at the base of the crust.
Interestingly, the tomographic model from Rawlinson et al. (2010) shows an increase of $P$-wavespeed at the SO01 location and the authors interpret the high velocity zone as a result of the presence of magmatic underplating. This mafic magmatic underplating may have occurred in a backarc region of a subduction zone as a result of high degrees of adiabatic decompression melting of the asthenosphere (Collins, 2002).

4.2. Moho topography and other evidence for complex structure

The profile of average 1-D shear wave velocity models across the Precambrian region shows a dipping structure (Fig. 6) with a dipping Moho towards the south of the Gawler Craton.

The trend of the mountain belt in the Lachlan Orogen (Snowy Mountains) is approximately N–S (Fig. 5a) east of longitude 148°E. The crustal thickness tends to increase as the surface topography increases east of longitude 138°E (Fig. 5). This direction is close to the elongation direction of Moho depth surface for most of the Lachlan Orogen (Fig. 9 of Clitheroe et al., 2000). This dipping Moho could be related to a crustal thickening to the west due to the slab convergence or to the crustal thinning towards the east induced by the continent–ocean transition.

4.3. Interpretation of receiver function measurements and reflection profile

The small difference in Moho depth estimates between RF and reflection surveys is likely to arise from the different way that they sample the crust. RFs exploit conversion of $P$-to-$S$ waves generated in transmission at crustal discontinuities beneath the recorder. Whereas the reflection method uses $P$ waves that are reflected from seismic discontinuity. RFs are most sensitive to sharp velocity contrast and sensitivity to smooth velocity gradients comes from the longer
wavelength components of the incident wave. Reflection surveys are carried out with much higher frequencies and pick up minor contrasts in acoustic impedance \( \rho V \), where \( \rho \) is the density and \( V \) is the \( P \)-wave velocity). The reflection studies have better vertical resolution than our RF study because the dominant frequency of the seismic signal from the RF is around 0.4 Hz, whereas it is around 40 Hz for the reflection profiles (e.g. Drummond et al., 2006). For seismic reflection studies, the vertical resolution attainable at Moho depth would be about 100 m. For our RF study, considering a dominant frequency of ca. 0.4 Hz a transition over 2 km will be indistinguishable from an abrupt reflection. In general the interpretation of the reflection Moho is made at the base of a crustal reflectivity, and this need not have direct correspondence to the features seen in transmission. Indeed the Fresnel zone, a measure of the likely resolving length, is much different from the two methods at the base of the crust. With a crustal thickness of 50 km, the Fresnel zone for RFs (–13 km) is much larger than for reflection measurements (less than 3–4 km) (Fomin et al., 2006).

4.4. Phanerozoic versus Precambrian regions

The Moho depth in the Mount Gambier volcanic area is thick –41 km suggesting that the limit between the Delamerian and western Lachlan orogens could be located east of Mount Gambier. Reflection profiles also indicate a very thick crust beneath Mount Gambier. Cayley et al. (2011) suggest the presence of Palaeoproterozoic–Cambrian ultramafic granulites at the base of the crust from interpretation of these recent reflection profiles. A southern extension of Gawler Craton material may reach as far as the Mount Gambier.

Drummond and Collins (1986) suggested from seismic refraction studies a correlation between increasing Moho depths and province age (excluding the Archean provinces). Clitheroe et al. (2000) observed a weak correlation using a RF analysis. The Lachlan Fold Belt or Lachlan Orogen was formed in the middle Palaeozoic (from 450 to 340 Ma) and the age of the western part of the Precambrian region is dominantly Proterozoic with an Archean core to the Gawler Craton. The crustal thickness of the Lachlan Fold Belt is thicker than beneath the Proterozoic province (Fig. 5). On the west of the Lachlan Fold Belt lies the Delamerian Orogen from early Palaeozoic (550 to 470 Ma). On the east side is the New England Orogen, which was formed from late Palaeozoic to Early Mesozoic (310 to 210 Ma). Even in the Lachlan Fold Belt no clear trend is observed between surface ages and Moho topography. As pointed out by Clitheroe et al. (2000) such correlations are difficult to justify because they rely on the assumption that the Earth surface age is correlated to the age of the last crustal growth event at the base of the crust.

5. Conclusion

Receiver function modelling of broad-band seismic waveforms recorded in southeastern Australia has been employed to constrain the way in which crustal structure changes from the Precambrian cratons to the Palaeozoic fold belt. Our estimates of Moho depths range between 28 and 48 km, with a mean value equal to 39 km, similar to Moho depths for global average of continental crust (38 km). We see a transition from a 40 km depth to Moho in the Precambrian, thinning beneath the Murray Basin and then thickening again in the exposed Lachlan Fold Belt. This leads to a general correlation between the surface topography and the Moho depths. The intermediate nature of crust–mantle transition beneath the Gawler Craton and the Lachlan Orogen suggests the presence of magmatic underplating at the base of the lower crust.

We have shown that our RF estimates for Moho depth are consistent with recent reflection profiles and provide a detailed tie and calibration of the methods. These detailed results provide strong support for the approach used by Kennett et al. (2011) to produce a continent wide map of the Moho.

The very thick crust in the Mount Gambier region obtained from both RF studies and reflection data suggests that the limit between the Delamerian and western Lachlan orogens is located east of Mount Gambier. Our results therefore support the interpretation of Direen and Crawford (2003) and place the Tasman Line well to the east of the location favoured by many authors (e.g., Scheibner and Veevers, 2000).

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.tecto.2012.09.031.

References


