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Testing the limits of virtual deep seismic sounding via new crustal thickness estimates of the Australian continent

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1 Summary

2 We apply virtual deep seismic sounding (VDSS) to data collected from both permanent and temporary seismic stations in
3 Australia with the goal of examining (i) the resilience of the method to the presence of complex lithospheric structure, and
4 (ii) the effectiveness of different approaches for estimating bulk crustal properties (namely thickness and V_p). Data from
5 the permanent WRAB in the Northern Territory station is ideal for benchmarking VDSS (large number and favourable
6 distribution of recorded earthquakes), with the results from several approaches agreeing on a thickness of 40-42 km. Ap-
7 plication of VDSS to data from the temporary BILBY array, a linear distribution of broadband stations that traverses
8 central Australia, shows that strong Moho reflections can be retrieved with as few as two earthquakes even at the transi-
9 tion between crustal blocks of different character and in the presence of thick sedimentary basins. Crustal thickness varies
10 between 36-54 km and compare well with the reflectivity character of nearby deep seismic reflection lines. Furthermore,
11 we find that off-line estimates of crustal thickness, calculated by binning the source regions according to back-azimuth,
12 produce estimates of crustal thickness that are consistent with the regional geology. Overall, we find that VDSS is a pow-
13 erful technique for estimating crustal thickness and velocity due to its insensitivity to complex short-wavelength structure
14 and requirement of a small number earthquakes to produce a stable result. However, not all schemes tested for extracting
15 bulk crustal properties appear to be robust and stringent data quality checking is still required during implementation.

16 **Keywords:** SEISMOLOGY, Body waves < SEISMOLOGY, Crustal imaging < SEISMOLOGY, Crustal structure <
17 TECTONOPHYSICS, Composition and structure of the continental crust < COMPOSITION and PHYSICAL PROPER-
18 TIES, Australia < GEOGRAPHIC LOCATION

19 **1 Introduction**

20 Australia has a long and convoluted tectonic history that has led to a complicated present day lithospheric structure. Major
21 orogenic episodes have resulted in continental growth and reworking of older material (e.g. Musgrave, Arunta, Petermann
22 and Alice Springs Orogens; Cawood and Korsch, 2008; Betts et al., 2002) while major sedimentary basins have formed
23 in response to extensional tectonics (e.g. Amadeus, Officer and Georgina Basins; Walter et al., 1995). These structures
24 can be problematic for typical high-frequency passive seismic imaging tools such as P-wave receiver functions because
25 of energy becoming trapped within the low-velocity sediments or because of complex signatures preserved at the crust-
26 mantle boundary (e.g. Langston, 2011; Clitheroe et al., 2000). It is therefore desirable to implement a technique that is
27 relatively insensitive to these complications, but that can still provide important first-order information on crustal structure
28 (i.e. crustal thickness, bulk velocity).

29 The Virtual Deep Seismic Sounding (VDSS) method is a relatively new approach that utilises incident S-waves from
30 earthquake-receiver distances of 30° - 50° (Tseng et al., 2009; Yu et al., 2012, 2013). A free surface conversion from S to P
31 leads to a topside Moho reflection, analogous to conventional active source deep seismic sounding, that has an extremely
32 high signal-to-noise ratio and is low enough frequency to be insensitive to complicated small-scale structure. It has been
33 used to delineate crustal thickness variations in regions that are expected to have complex Moho signatures (e.g. Tibet
34 and Ordos Plateau; Tseng et al., 2009; Yu et al., 2012), significant sedimentary cover (Eastern USA; Parker et al., 2016)
35 and where large crustal variations likely cause differences in isostatic compensation (Western USA; Yu et al., 2016).

36 In this paper, we implement the VDSS method on two geographically colocated datasets of contrasting deployment
37 times and apply a variety of migration and waveform modelling techniques to test the limits of its applicability. Aus-
38 tralia is the ideal natural laboratory for testing VDSS given that it has complex lithospheric structures that the method is
39 hypothesised to be insensitive to and is surrounded by seismogenic zones in the desired distance range. The approach
40 can provide robust results using comparatively few traces due to the signal-to-noise ratio achieved. Despite this, we also
41 highlight that certain modelling approaches can return spurious results even for voluminous high quality datasets.

42 **2 Geological Setting and Previous Geophysical Studies**

43 The geology of Central Australia is dominated both by orogeny and widespread sedimentary sequences that range in age
44 from Paleoproterozoic to Carboniferous (e.g. Cawood and Korsch, 2008). The oldest terrane to be imaged by this study
45 is the Arunta Block (Fig. 1), sometimes referred to as the Arunta Orogen. It comprises Paleoproterozoic sequences with
46 ages between ~ 1880 - 1800 Ma, but various stages of tectonic activity have been recorded as late as ~ 1580 Ma (Cawood
47 and Korsch, 2008). The 1.30-1.10 Ga collision of the South Australian Craton (containing the Gawler Craton, see Fig.

48 1) with the previously amalgamated North and West Australian Cratons is preserved in the Musgrave Block (Betts et al.,
49 2002; Howard et al., 2015), bounded in the north by the Amadeus Basin and to the south by the Officer Basin (Fig.
50 1). Subsequent dike emplacement and volcanism between 1.10-1.00 Ga was followed by a period of quiescence that
51 culminated in uplift and further magmatism at ~ 800 Ma. The origin of this volcanism is believed to be due to the
52 impingement of a mantle plume (Zhao et al., 1994) with subsequent extension and crustal sagging forming the Centralian
53 Superbasin (encompassing the vast majority of what is now Central Australia, including the Musgrave Block, Amadeus
54 Basin, Officer Basin and southern Georgina Basin; Walter et al., 1995). The current geometry of sedimentary basins was
55 achieved by segmentation of the Centralian Superbasin due to two major phases of intraplate deformation associated with
56 Neoproterozoic-Cambrian and Carboniferous orogeny. The Petermann Orogen (580-540 Ma) was focussed at the margin
57 of the Amadeus Basin and Musgrave Block (Fig. 1). The Alice Springs Orogen (~ 320 Ma) reactivated structures along
58 the northern margin of the Amadeus Basin, most notably the Redbank Shear Zone. This focussed reactivation has led to
59 large Moho offsets being preserved across major faults by as much as 20 km in places (e.g. Goleby et al., 1989).

60 On the broadest scale, surface wave analysis has mapped the long wavelength features of the Australian continent
61 to asthenospheric depths highlighting a strong east-west change between the ancient (Proterozoic and older, >1.1 Ga)
62 domains of central/western Australia and the younger (<1 Ga) regions of eastern Australia (Fishwick et al., 2005). The
63 older terranes to the west of $\sim 140^\circ\text{E}$ (broadly corresponding with a feature referred to as the Tasman Line) exhibit fast
64 seismic velocities to depths of 200 km or greater which is indicative of a thick, cold cratonic root often observed beneath
65 surface geology of this age (Fishwick et al., 2005; Kennett et al., 2004; Polet and Anderson, 1995). Despite this finding,
66 the shallowest parts of the surface wave model (~ 75 km depth) exhibit lower than expected seismic velocities which are
67 also confirmed by body-wave observations (Kaiho and Kennett, 2000; Fishwick et al., 2008). Recent Pn tomography
68 work has identified a more complex pattern of lithospheric velocity variations within our study area in central Australia
69 and conclude that regions exhibiting low mantle P-wave velocities (<8.0 km s $^{-1}$) are likely affected by velocity gradients
70 in regions of thickened crust (Sun and Kennett, 2016). Ambient noise tomography has also shown a complex pattern
71 of crustal structure throughout Central Australia, including the signatures of the aforementioned sedimentary basins and
72 Proterozoic collisional zones (Saygin and Kennett, 2012). This type of study benefits from utilising higher frequency
73 surface waves for determination of shallower structure (<40 km), but for regions of thicker crust this sensitivity remains
74 poor (Saygin and Kennett, 2012). P-to-S energy converted at the Moho via the receiver function method again shows the
75 presence of large variations in crustal thickness that correlate with the major tectonic blocks of central Australia, with
76 thicker crust (>50 km) and a diffusive Moho signature at the northern and southern margins of the Amadeus Basin (Sippl,
77 2016). The diffusive nature of the Moho has made it difficult to derive a definitive crustal thickness at several of these
78 localities (Sippl, 2016). Active source seismic surveys provide the highest frequency probe of deep crustal structure, but
79 are generally limited to 2D transects which provide limited geographical coverage. Sharp changes in crustal thickness, in
80 excess of 25 km of vertical Moho displacement across the Redbank Shear Zone at the northern margin of the Amadeus
81 Basin (Fig. 1), were identified in early crustal-scale reflection profiles (Goleby et al., 1989; Korsch et al., 1998). A
82 large number of seismic reflection transects have been performed across the Australian continent since these early studies
83 (comprehensively reviewed by Kennett and Saygin, 2015), with major findings including that the central Australian crust

84 exhibits distinct jumps (8-10 km in places in addition to the previously mentioned variations) in Moho geometry along
85 with regions where the reflectivity character makes definitive determination of crustal thickness impossible (Kennett and
86 Saygin, 2015; Kennett et al., 2016; Korsch and Doublier, 2016).

87 Based on the complexity of the surface geology/lithospheric structures observed, the availability of broadband seismic
88 data traversing these terranes, co-located active source seismic studies and the unavoidable technical difficulties faced by
89 the common suite of geophysical techniques implemented in the region, it is clear that central Australia represents the
90 ideal region in which to test the limits and sensitivities of the VDSS methodology.

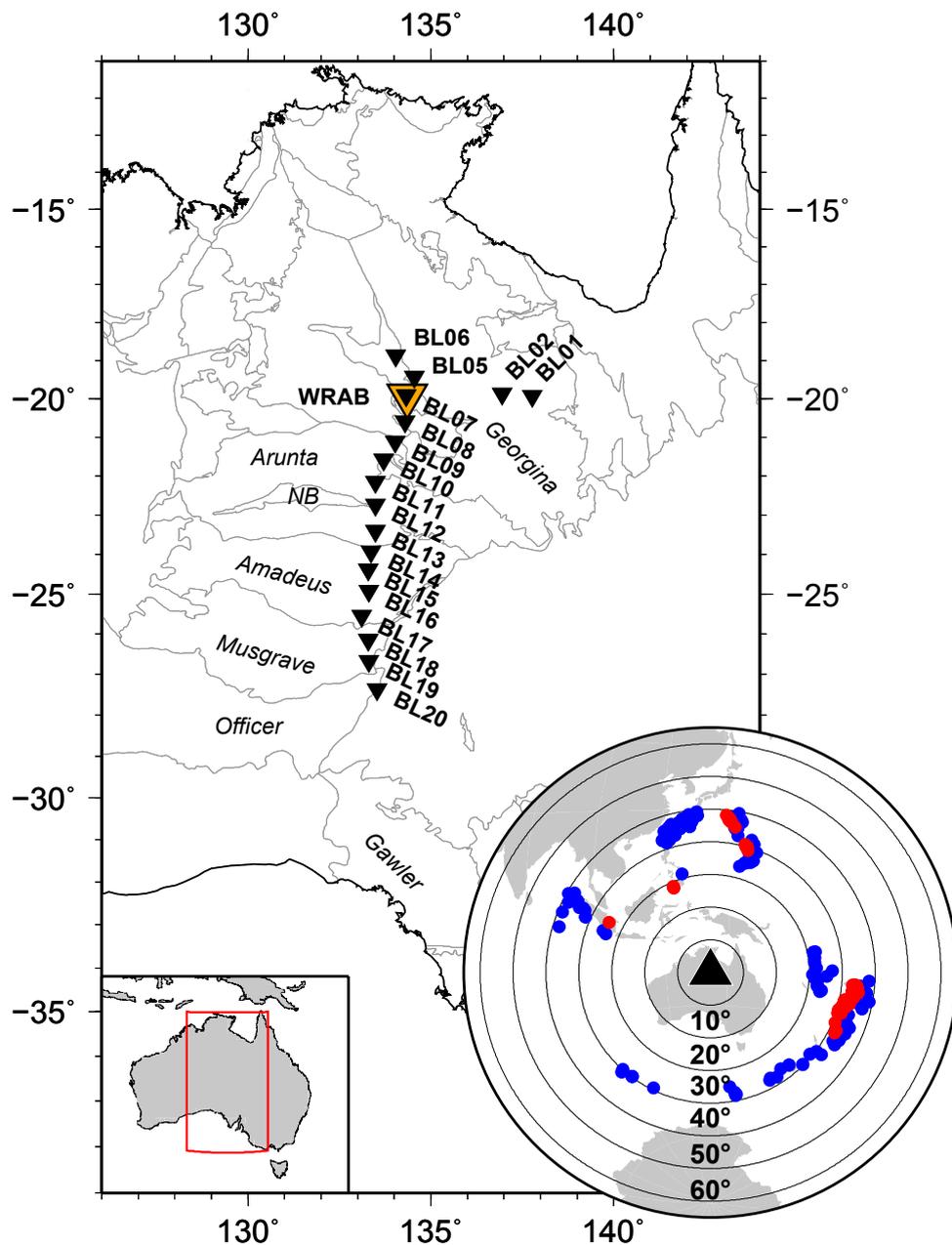


Figure 1: a) Map of the Australian continent showing the locations of broadband seismometers used in this study and the main geological terranes. NB=Ngalia Basin. Inset is the location of seismicity that contribute to the VDSS dataset both for WRAB and the BILBY seismic network. Blue dots are events with shallow hypocenters (<410 km) and red are events with deep (>410 km) hypocenters.

91 **3 Data and Methods**

92 **3.1 Seismic Stations**

93 Permanent station WRAB (Streckeisen STS-2 shallow surface vault seismometer; part of the IRIS-IDA Global Seis-
94 mograph Network) is located at the Warramunga Seismic Array in northern Australia (Fig. 1) and has been recording
95 continuously since 1994. We also incorporate data from the temporary BILBY seismic network (Rawlinson and Kennett,
96 2008), a 23 station linear broadband network that was operational between 2008 and 2011 (Fig. 1, see Sippl, 2016, for
97 details).

98 Earthquakes of M5.5+ for the distance range of 20°-60° were selected for preprocessing, with a visual inspection
99 of traces being performed to remove events that were poorly recorded. Data were initially filtered with a 2nd order zero
100 phase Butterworth band-pass filter with corner frequencies of 0.05 Hz and 0.5 Hz with the horizontal components being
101 subsequently rotated into the radial and tangential components for further processing.

102 **3.2 Virtual Deep Seismic Sounding**

103 The VDSS method relies on the fact that the topside P-to-P reflection from the Moho (following the preceding S-to-P
104 conversion at the free surface, Fig. 2) is post-critical and hence undergoes total internal reflection (Tseng et al., 2009).
105 This provides the high signal-to-noise that characterises VDSS observations, but also results in a phase shift of the $SsPmp$
106 phase that needs to be taken into account.

107 The source normalisation approach described by Yu et al. (2013) was implemented on the Australian data to remove
108 source-side scattering effects, making events suitable for VDSS analysis regardless of focal depth. We use the theoretical
109 ray parameter (p_β) and incidence angle (j) for the direct S-wave assuming the ak135 velocity model (Kennett et al., 1995;
110 Crotwell et al., 1999) to calculate the rotation required to transform the vertical and radial components into the pseudo-S
111 component (ϕ_1) using the following equation:

$$\tan(\phi_1) = (\eta_\alpha/\eta_\beta)\tan 2j, \quad (1)$$

112 where η_α and η_β are the P-wave and S-wave vertical slownesses, $(V_p'^{-2} - p_\beta^2)^{1/2}$ and $(V_s'^{-2} - p_\beta^2)^{1/2}$, respectively,
113 with V_p' and V_s' being the near surface P and S velocities (for further details, see Yu et al., 2013). Rotating into the
114 pseudo-S component provides an estimate of the shear-wave source wavelet for the VDSS analysis, with this being
115 deconvolved from the vertical component seismogram using an extended time multi-taper approach (10 s sliding window
116 length, 75% window overlap, 3 Slepian tapers; Helffrich, 2006). For WRAB, this led to 294 VDSS traces across the 22
117 year deployment history. The temporary nature of the BILBY deployment led to a significantly reduced number of usable
118 traces, varying between 2 and 19 traces per station. For each station, VDSS traces were binned and stacked by slowness
119 in 0.2 s deg^{-1} bins with 50% overlap (see Fig. 2 for WRAB and supplementary information for BILBY stations). A
120 minimum of 3 traces were required to produce a stacked trace.

121 The dataset was then interrogated using a variety of approaches with the goal of assessing their robustness and

122 repeatability both for a voluminous, high-quality dataset (WRAB) and a comparatively data-poor but geographically
 123 colocated seismic deployment (BILBY).

124 Firstly, a new migration-based approach was developed to map the seismic energy to depth assuming a single layer
 125 crust. The workflow is built upon receiver function migration (Wilson and Aster, 2003; Angus et al., 2009; Hammond
 126 et al., 2011; Thompson et al., 2011, 2015) edited to incorporate the different geometry of the $SsPmp$ phase. Given that
 127 the arrival time of this phase relative to the incident S-wave in a constant velocity medium is:

$$T_{SsPmp-Ss} = 2H(V_p^{-2} - p_\beta^2)^{1/2}, \quad (2)$$

128 where p_β is determined using the known source-receiver geometry and the ak135 velocity model (Crotwell et al., 1999),
 129 it is possible to convert the arrival time into a depth by assuming a crustal velocity. We present results for an assumed P-
 130 wave velocity of 6.5 km s^{-1} and V_p/V_s ratio of 1.73, but also provide results with V_p of 6.4 km s^{-1} and 6.6 km s^{-1} in the
 131 supplementary information. Due to its post-critical nature, the $SsPmp$ arrival undergoes a phase shift of $\sim \pi/2$, leading to
 132 the arrival time being manifested as a zero crossing as opposed to an amplitude maximum typical of other methodologies.
 133 Therefore, we migrate all traces at a given station to depth using the aforementioned approach, producing a summary
 134 depth trace through linear stacking and derive crustal thickness by determining the depth at which the zero-crossing
 135 occurs. Error estimates on this value are obtained by bootstrapping the input data 100 times (Efron and Tibshirani, 1991),
 136 randomly selecting traces from the dataset up to the original number (allowing repetition). We also attempt to remove the
 137 effect of the phase shift by calculating the envelope function (Parker et al., 2016; Bracewell, 1978; Phinney et al., 1981)
 138 for all available VDSS traces and migrating these to depth using the same approach described above. Using the envelope
 139 functions, the arrival time would be seen as a positive peak in the stacked migrated trace.

140 Previous studies have relied on waveform modelling to determine crustal thickness, and where a range of slownesses
 141 were available, to determine crustal V_p (Tseng et al., 2009; Yu et al., 2016). We use a similar approach to initially invert
 142 for H alone. Each of the slowness bins were used as input traces to model synthetic waveform data for a simple layer over
 143 a half space, allowing crustal thickness to vary between 30-55 km in 0.2 km increments. Synthetic traces were produced
 144 using the reflectivity method (Fuchs and Müller, 1971; Kennett, 1983). Each model was compared to the real data by
 145 calculating the L2 norm, with quoted values being the model with minimum misfit and associated errors calculated by
 146 repeating the process with V_p of 6.4 km s^{-1} and 6.6 km s^{-1} .

147 It is not just the $SsPmp$ phase that is present in the recorded traces, with the precursory Smp and reverberatory
 148 $SsPmsPmp$ phases also appearing as strong arrivals (see Fig. 2 and supplementary information). The time window
 149 over which the misfit was calculated is hence -10 s to +30 s relative to the direct S-wave arrival in order to incorporate
 150 these observed phases. The same approach was used to simultaneously invert for H and V_p (herein referred to as the
 151 simultaneous H- V_p inversion) with the same range of H but allowing V_p to vary between $6.1\text{-}6.9 \text{ km s}^{-1}$ in increments of
 152 0.02 km s^{-1} . In this case, quoted values are the average H and V_p from the 150 best fitting models.

153 Unlike $SsPmp$, the arrival times of Smp and $SsPmsPmp$ are dependent on the V_p/V_s ratio. In addition to this,
 154 the relative timing of many of the phases of interest (Smp , $SsPmsPmp$) and the distance at which $SsPmp$ becomes
 155 post-critical is also dependent on the crustal V_p/V_s ratio and uppermost mantle velocity. To investigate the sensitivity of

156 the dataset to these parameters, we also attempted to invert for H, Vp, Vp/Vs ratio and Pn velocity simultaneously.

157 The final method was the linear regression approach of Kang et al. (2016), where fitting a straight line through the
158 *SsPmp* travel time picks plotted in the form of p_{β}^2 vs. $T^2/4$ allows for the determination of H and Vp. Travel times were
159 derived from the envelope functions of each slowness bin to negate the phase shift of the post-critical reflection.

160 4 Results

161 4.1 Permanent Station: WRAB

162 The large dataset (294 VDSS traces) and excellent slowness coverage (13.9-15.7 s deg⁻¹) that WRAB provides is an ideal
163 dataset with which to benchmark the VDSS method (Figs. 2a-c). The main phase of interest, the post-critical *SsPmp*
164 reflection, is clearly observed with high signal-to-noise across the whole slowness range with an arrival time (relative to
165 the phase *Ss*) varying from ~6-8 s (Figs. 2b-c). The precursory *Smp* phase can be seen arriving ~6 s prior to the direct
166 S-wave in most bins, and the reverberatory *SsPmsPmp* is also clearly observed at times of 16-22 s (Figs. 2b-c).

167 The migration-based approach returned a crustal thickness of 41.2±0.5 km (mean and 2σ data uncertainty); the
168 summary migrated trace can be seen in Fig. 2a. In order to utilise the large dataset and test the stability of this result,
169 a number of data subsets were randomly created (without allowing repetition of traces) for sample sizes ranging from 5
170 traces to the maximum 294. The results of this process can be seen in Fig. 3. All data subsets (with the exception of the
171 group containing 60 traces) lie within error of the result from the entire WRAB dataset, highlighting the robustness of
172 the H estimates even with a comparatively small number of traces. Errors gradually reduce from ±3.8 km for 5 traces to
173 ±1 km or less for subsets of >70 traces (Fig. 3).

174 Despite being able to provide a single migration result for a given station, we also investigate back-azimuthal vari-
175 ations by binning data from the two dominant source regions (notably the Java subduction system to the north and the
176 Tonga-Fiji subduction system to the east). Fig. 2 shows that the migrated traces for both of these source regions are
177 extremely similar with the results being within error of each other (40.8±0.6 km for the easterly dataset and 41.6±0.9 km
178 for the northerly dataset).

179 Due to the assumption of the migration approach that the zero crossing represents the arrival time of the *SsPmp*
180 phase, there may be inherent bias in the crustal thickness estimate due to the fact that the phase shift is not equal to $\pi/2$
181 at all slownesses. This has been investigated both through tests on synthetic data for a 42 km thick crust (Vp=6.5km s⁻¹,
182 based on the migration based result of 41.2±0.5 km) and by binning the WRAB dataset into 0.2 s deg⁻¹ bins with 50%
183 overlap (Fig. 4). The maximum deviation due to the zero crossing assumption in the synthetic dataset (filled circles in
184 Fig. 4) is ~5 km below the true thickness at the high slowness end of the data coverage (15.7 s deg⁻¹). The real data from
185 WRAB does not follow the expected pattern identified in the synthetic data, with most bins below 14.6 s deg⁻¹ exhibiting
186 lower values than the synthetic values and the majority of bins above 15.0 s deg⁻¹ showing greater values than expected
187 based on the synthetic tests.

188 The envelope function migration returned a value of 39.6±0.8 km, close to the result of the VDSS trace result of

189 41.2±0.5 km. The close agreement of these two approaches is likely due to the excellent slowness coverage of the
190 WRAB dataset averaging out any potential bias due to the zero crossing assumption inherent to the VDSS trace migration
191 approach.

192 The single parameter inversion gave a similar crustal thickness of 42.4±2.4 km (Table S1) and the simultaneous H-
193 Vp inversion gave estimates of H=39.6±3.3 km and Vp=6.36±0.16 km s⁻¹ (Fig. 5 and Table S3), also within error of
194 both migration approaches. Results from the simultaneous H-Vp inversion obtained by minimising the L1 norm give
195 similar results (H=40.7±3.3 km and Vp=6.40±0.16 km s⁻¹). The four parameter inversion shows H and Vp estimates
196 consistent with the previously discussed approaches (40.1±5 km and 6.36±0.2 km s⁻¹) but only a weak dependence on
197 Vp/Vs (1.73±0.07) and Pn velocity (8.3±0.3 km s⁻¹) based on the broad spread in the histogram (Fig. S1 and table S3).

198 The envelope functions, including the *SsPmp* – *Ss* travel-time pick derived from the traces themselves and the
199 predicted travel-time curve based on the result from the migration approach, can be seen in Fig. 6. These picks were
200 then used to perform the regression analysis (Fig. 7), yielding estimates of H=33.6 km and Vp=6.1 km s⁻¹. Despite the
201 excellent slowness coverage and high quality data, and even when invoking conservative 2σ error estimates (±6.6 km and
202 ±0.32 km s⁻¹) from Kang et al. (2016), these results were quite different to other studies of the region. The potential
203 reasons for this will be discussed in subsequent sections.

204 4.2 Temporary Deployment: BILBY network

205 The temporary nature of the BILBY network and long periods in which no data were recorded (Sipl, 2016) has lead
206 to a low number of VDSS traces at most stations when compared with the WRAB dataset. The number of traces are
207 summarised in tables S1-S12 (varying between 2 and 19 per station).

208 To investigate potential back-azimuthal variations in the results, *SsPmp* Moho reflection points were calculated for
209 each VDSS trace and a mean reflection point determined for the two back-azimuthal corridors, the results of which are
210 shown in Fig. 10. Crustal thicknesses appear consistent from all directions beneath the Amadeus and Georgina Basins
211 (Fig. 8, ~24°S and ~20°S, respectively). The thinnest crust (35-40 km) is seen beneath these significant sedimentary
212 basins while the northern data, spatially coherent between several stations and coincident with the surface expression of
213 the Ngalia Basin, also exhibits crustal thicknesses of ~35 km (Figs. 8 and 10). The thickest crust is associated with the
214 margins of major crustal blocks in Central Australia. Data arriving from the east (Tonga-Fiji subduction zone) in Fig. 8
215 shows the crust thickening to ~50 km or greater at the margin of the Gawler Craton and Musgrave Block to the south of
216 the BILBY line (~27.5-26.5°S), and also at the southern extent of the Arunta Block (~23.0°S).

217 As with the WRAB dataset, the envelope functions were also migrated to depth for the BILBY network, with the
218 same pattern of crustal thickness variations as seen with the migration of the VDSS traces being evident (Fig. 8 and 9).
219 Many stations exhibit larger bootstrap error estimates using the envelope function migration, particularly where the crust
220 appears thicker (BL12 and BL19). This is likely due to the inherently broader waveforms associated with the *SsPmp*
221 arrival time that are produced when converting the data in this fashion (Parker et al., 2016).

222 Where available, the single parameter waveform inversion results for an assumed velocity of 6.5 km s⁻¹ are in good
223 agreement with the associated migration estimates with the same crustal velocity (typically within error, see Fig. 11a and

224 tables S1-S12). This is reassuring given the identical crustal velocity assumed for the two approaches.

225 The general trend of thin crust within the Amadeus Basin and thicker crust at its margins is also seen in the simul-
226 taneous H-Vp inversion results (Table S2, Figs. S26-S34). However, several stations (e.g. BL12, BL13 and BL14) have
227 Vp values that are extremely variable, e.g. 6.21 km s^{-1} and 6.81 km s^{-1} at adjacent stations (Figs. S28 and S29, respec-
228 tively) and BL12 even showing a bimodal distribution in its histograms (Fig. S27). This is highly likely to be due to the
229 large back-azimuthal variations observed and it is also these stations that tend to exhibit the poorest fits in the presented
230 slowness sections (Figs. S11-S13). Several stations again agree to within error when compared to the migration results
231 (Fig. 11b), but there are larger deviations evident than is observed when comparing the results of the single parameter
232 inversion (with the simultaneous H-Vp inversion crustal thickness estimates exhibiting significantly thinner crust than the
233 migration results).

234 5 Discussion

235 The general pattern of crustal thickness is consistent with recent P-receiver function results from the BILBY array (Sippl,
236 2016). The thinnest crust is observed beneath the Amadeus and Georgina Basins, with the thickest crust being observed at
237 the northern margin of the Amadeus Basin and the southern margin of the Musgrave Block (Figs. 8 and 10). An important
238 observation is that high frequency receiver functions are complex in nature for the regions of thicker crust, making it
239 difficult to obtain a confident estimate of crustal thickness (Sippl, 2016). In contrast, the VDSS method retrieves a clear
240 *SsPmp* phase from the Moho at all working stations including those expected to have complex Moho signatures (Fig.
241 8 and supplementary information). This corroborates conclusions from previous VDSS studies in other localities (Tseng
242 et al., 2009; Yu et al., 2012; Chen et al., 2013).

243 The AusMoho model of Kennett et al. (2011) incorporates a range of seismic data to produce crustal thickness
244 estimates across the Australian Continent. The VDSS results are compared with AusMoho in Figs. 8 and 12. Most of the
245 stations exhibit crustal thickness estimates between the two datasets that lie within error of each other, although there are
246 stations that deviate by significant amounts (10 km or more in places, BL20 and BL19 for example). These are mainly
247 where the crust is expected to be at its thickest at the margin of the Gawler Craton and the Musgrave Block, and within the
248 Arunta Block between the Amadeus and Georgina Basins (Korsch and Doublier, 2016). These are also regions where P-
249 receiver functions have a diffuse Moho signature that make it problematic to clearly define crustal thickness (Sippl, 2016),
250 likely due to structural complexities around the Moho. Despite this, a clear *SsPmp* phase is observed across the entire
251 transect (Fig. 8), even in the regions where the higher frequency methods image the Moho poorly. This could explain
252 some of the deviations from the VDSS results and those from AusMoho and Sippl (2016). Previous VDSS studies have
253 suggested that the long period and post-critical nature of the *SsPmp* phase should be relatively insensitive to complex
254 structure around the Moho and still provide a robust estimate of crustal thickness (e.g. Yu et al., 2012, 2013); our results
255 provide further evidence to support this conclusion.

256 As with this study, Sippl (2016) also identified the potential for strong lateral variations in crustal thickness be-
257 neath BILBY stations. Agreement between the receiver function dataset and our dataset is not perfect, particularly the

258 back-azimuthal variations observed at BL12 and BL13 where the VDSS results suggest thinner crust where the surface
259 expression of the Ngalia Basin lies. Different ray geometries and sampling between P-receiver functions and VDSS likely
260 mean that different structures influence the two methods. P-receiver functions would typically have Moho piercing points
261 for the Pms phase at distances of ~ 5 -15 km from the station, while the bounce point for the $SsPmp$ will vary between
262 ~ 40 -65 km from the station. Therefore, horizontal distances between sampling points for P-receiver functions and VDSS
263 are on the order of 25-60 km when assuming that the back-azimuth of the incoming rays are the same. These distances
264 could be even greater when considering different azimuthal coverage of the different datasets, highlighting the potential
265 for these approaches to be sampling very different regions even for a single station.

266 The large crustal thickness variations in this part of the study region are in agreement with previous active source
267 investigations that identify significant Moho offsets of as much as 20 km between the thinner crust of the Ngalia Basin
268 (~ 30 km) and comparatively thicker crust (~ 50 km) beneath the Central Arunta Province (Goleby et al., 1989; Korsch
269 et al., 1998). This was interpreted as being due to thick-skinned deformation along focussed, crustal-scale fault zones
270 within an intracratonic environment (Goleby et al., 1989). Our VDSS study also suggests that large lateral changes in
271 crustal thickness occur across central Australia.

272 It is also possible that the observed lateral variations may be an artefact due to differences in upper mantle velocity
273 structure between the direct S and the $SsPmp$ phases, leading to variations in differential arrival time that are not associ-
274 ated with crustal thickness or crustal velocity structure. The lateral distance beneath the station at 210 km depth between
275 the direct S and the $SsPmp$ phase is $\sim 1^\circ$ (~ 100 km). As an example, the Western USA has large variations in upper
276 mantle velocity structure (5% in V_p and 10% in V_s) over these lateral distances. This led to timing differences of as much
277 as 3 s in regions with strong lateral variations in lithospheric structure that were taken into account by Yu et al. (2016). The
278 AusREM velocity model shows maximum velocity variations across distances of 100 km of 1% or less in the vicinity of
279 the BILBY transect (Kennett et al., 2013). This would be insufficient to produce the large, sharp back-azimuthal variations
280 in crustal thickness observed at BILBY stations, making it likely that these are due to real structure as opposed to being
281 an artefact. If this is the case, the results suggest that there is heterogeneity in Central Australian crustal structure that is
282 not being fully resolved by the linear and geographically limited active source studies and the roughly vertically incident
283 nature of P-receiver functions.

284 The coincidence of active source lines provides a unique comparison between VDSS and true deep seismic sounding
285 (Korsch and Doublier, 2016), the first time this has been possible. Geoscience Australia and their associated partners
286 have acquired numerous deep seismic datasets from across the continent over the previous decades (comprehensively
287 summarised by Kennett et al., 2016), with two lines being of particular interest to this study. The 08GA-OM1, commonly
288 referred to as the GOMA seismic line, ran from the Gawler Craton through the Officer Basin, Musgrave Block and into
289 the Amadeus Basin (Figs. 10 and 12). The southern end of the BILBY VDSS data coverage (BL17-BL20) are co-located
290 with the northern extent of the GOMA line. The 09GA-GA1 line sampled the northern Amadeus Basin, the Arunta Block
291 and the southern parts of the Georgina Basin (Figs. 10 and 12). While this line is further east than our VDSS coverage, it
292 is directly along strike of the dominant structure warranting comparison between the two complimentary datasets.

293 Results from GOMA and 09GA were incorporated into the crustal component of the AusREM and AusMoho ref-

294 erence models (Kennett et al., 2011; Salmon et al., 2012), meaning the previous comparisons with this model are also
295 relevant when comparing with the active source experiments. The seismic reflection signature of the Moho can vary
296 depending on the tectonic setting, but it is common to see a reduction in reflectivity from the lower crust into the upper
297 mantle (Eaton, 2005). This reduction in reflectivity is evident in the GOMA and 09GA data, with the VDSS crustal
298 thickness estimates agreeing well with this proxy (Fig. 12). The consistency between these passive and active source
299 techniques validates the VDSS approach as a robust tool for determining bulk crustal structure.

300 The source normalisation approach of Yu et al. (2013) was implemented in order to combine both shallow and deep
301 earthquakes from the surrounding plate boundaries. With 113 events being located at depths of greater than 410 km
302 and 127 at depths of beneath 100 km, this allowed the dataset to be greatly increased. This provides strong support
303 for the applicability of this method to regions where deep seismicity with simple source-time functions are sparse. The
304 theoretical approach used here to rotate the P-SV components into the pseudo-S component in order to obtain an estimate
305 of the $SsPmp$ source wavelet also contributed to producing a high quality and easily repeatable dataset for stations that
306 are both data-rich and data-poor. As with previous studies (Tseng et al., 2009; Yu et al., 2013), our results suggest that
307 where the crustal structure has thick sediments or complex Moho structure, the VDSS method still provides a clear signal
308 with which to robustly estimate crustal thickness variations, and to a lesser extent, V_p .

309 Finally, Kang et al. (2016) used data from two stations in Australia, including a station at Warramunga, to test their
310 linear regression method. Our crustal thickness estimate of 39.6-42.4 km from WRAB are in excellent agreement with
311 their result of 42.0 ± 3.2 km. Kang et al. (2016) calculated a bulk crustal V_p of 6.51 ± 0.14 km s⁻¹, within error of our
312 estimates of 6.36-6.40 km s⁻¹ (with an associated error of 0.16 km s⁻¹) from the waveform modelling. Despite the large
313 data volume and clear move out of topside Moho reflection across the full distance range (Figs. 2 and 6), our attempt to
314 implement the linear regression method on data from WRAB led to crustal thickness and V_p estimates of 33.6 km and
315 $V_p=6.1$ km s⁻¹, respectively (Fig. 7). These are vastly different from the results of Kang et al. (2016), surprising given
316 the data volume and quality. It appears likely that the variability in the estimate for H is due to the great distance from
317 the origin of the data that the intercept is calculated from on the p^2 vs. $T^2/4$ plot (Fig. 7), with even minor changes
318 to the input data potentially creating large variations in estimates of H. Based on previous estimates of V_p (Ford et al.,
319 2010; Kang et al., 2016) and the values from the waveform modelling being in the range ~ 6.4 -6.5 km s⁻¹, the value of
320 6.1 km s⁻¹ returned from our linear regression analysis seems both inconsistent and unrealistic.

321 **5.1 Recommendations for Best Practice**

322 The analysis of contrasting VDSS datasets has shown many of the methods associated with the technique to be robust
323 even with comparatively small datasets. Despite this, some of the approaches have also been shown to return spurious
324 results even with an extensive and high quality dataset such as that from station WRAB. Based on the findings of this
325 study, we summarise below a number of key points that we consider to be best practice for the implementation of the
326 VDSS approach for future studies:

- 327 1. The source normalisation approach of Yu et al. (2013) should be used in order to maximise the usable depth range of
328 earthquakes and hence the number of usable traces.

- 329 2. For stations with limited datasets (both in terms of the number of traces and slowness coverage), analysis should
330 be limited to the determination of crustal thickness alone either by migration or through waveform inversion.
- 331 3. Despite the theoretical validity of the linear regression approach (Kang et al., 2016), application to the high qual-
332 ity data from WRAB suggest that the solutions are potentially unstable. We therefore believe that alternative
333 data analysis techniques (waveform inversion in particular) are more robust for determining estimates of crustal
334 thickness and crustal V_p .
- 335 4. Whilst both migration-based approaches (VDSS traces and envelope functions) produce largely consistent results,
336 the VDSS traces provide much more visually intuitive and interpretable results. We therefore prefer the use of this
337 approach, but potential biases due to the phase shift associated with post-critical reflection not being equal to $\pi/2$
338 across all slownesses must be taken into account during implementation. This does not preclude the production of
339 envelope functions and their subsequent migration to depth as this provides a worthwhile comparison.
- 340 5. As with most geophysical techniques, it is desirable when implementing the VDSS methodology to use several
341 approaches to investigate the stability of any recovered parameter estimates. As a minimum, we suggest that
342 determination of crustal thickness alone can be done both through migration and by waveform modelling.

343 **6 Conclusions**

344 A range of approaches for determining bulk crustal properties using the VDSS method have been tested using permanent
345 and temporary broadband seismic data from across Central Australia. Both data-rich and data-poor stations produce
346 strong $SsPmp$ phases in the 30° - 50° distance range, and hence provide robust estimates of crustal thickness using a
347 new migration-based method and a waveform modelling approach. Crustal thicknesses correlate with known geological
348 terranes and are broadly consistent with previous passive and active source studies, although local deviations of up to
349 10 km do exist. Estimates of V_p from waveform modelling range from 6.2 - 6.8 km s^{-1} where available with an average of
350 6.42 km s^{-1} , although caution should be exercised with this parameter as not all of these observations can be considered to
351 be representative due to lack of slowness coverage or large back-azimuthal variations in the signature of the $SsPmp$ phase.
352 Results from Australia provide strong evidence to corroborate previous findings that the VDSS method is resilient against
353 complex Moho signatures and sedimentary basin structure. As such, we advocate its wider use in seismic studies that seek
354 to characterise the bulk seismic properties of the continental crust, even when relatively few good quality earthquakes are
355 recorded.

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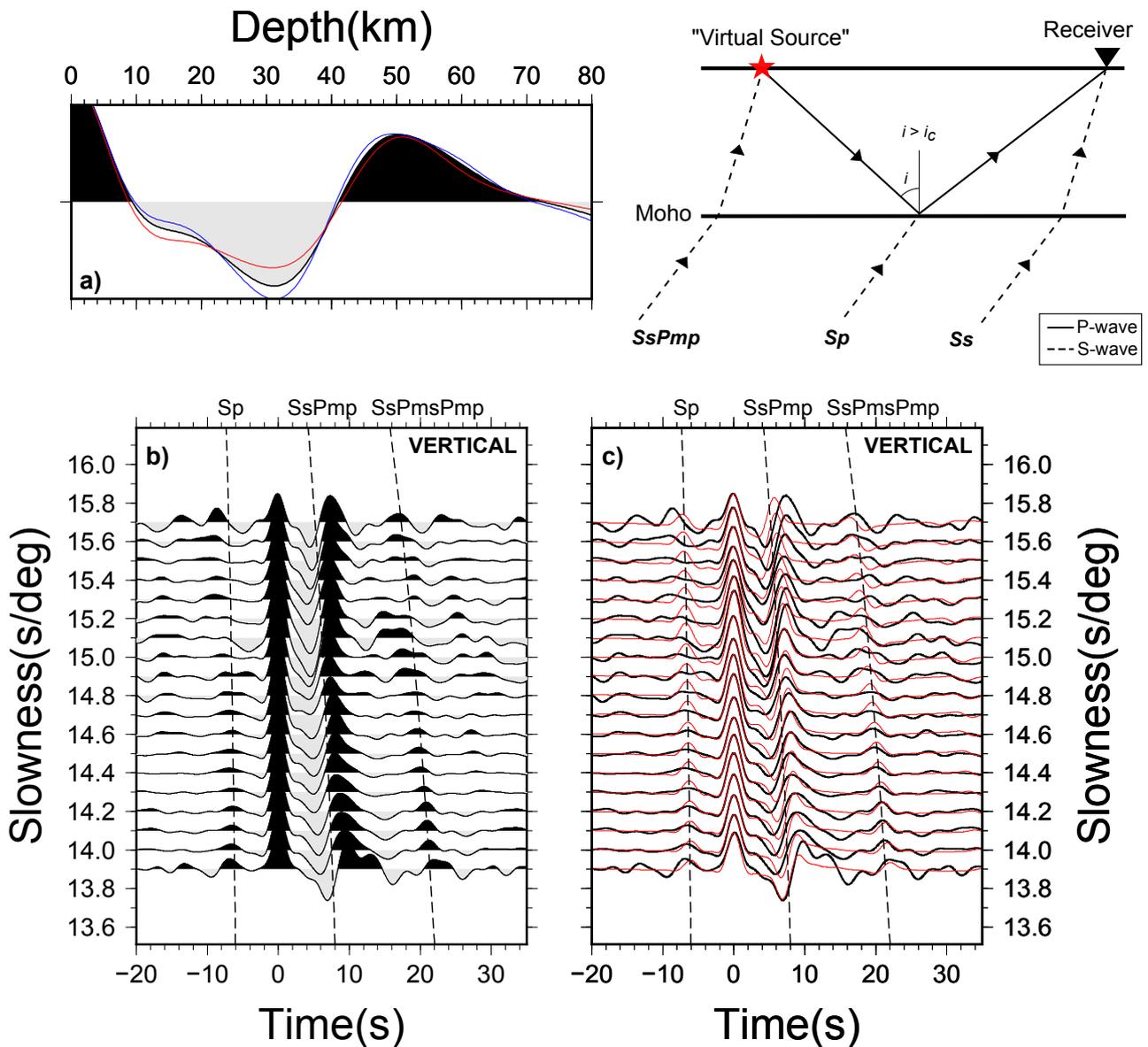


Figure 2: a) Results of the 1D depth migration for WRAB. Black line is the whole dataset, blue line is data from the easterly back-azimuthal corridor and red is data from the northerly back-azimuthal corridor. b) VDSS traces binned by slowness. Clear move-out of the $SsPmp$ phase is observed across the epicentral distance range of interest (30° - 50°), along with the presence of both the precursory Sp phase and the reverberatory $SsPmsPmp$ phase. The dashed lines are the predicted arrival times for the crustal thickness derived from the migration approach (full dataset). c) Same as b), but with the best fitting model from the single parameter (H) waveform inversion approach underlain. Amplitudes are scaled such that the direct S-waves in both the real data and the synthetic traces are equal. Inset is a ray diagram of the main $SsPmp$ waveform utilised during VDSS analysis (i =incidence angle at the Moho, and i_c =critical angle).

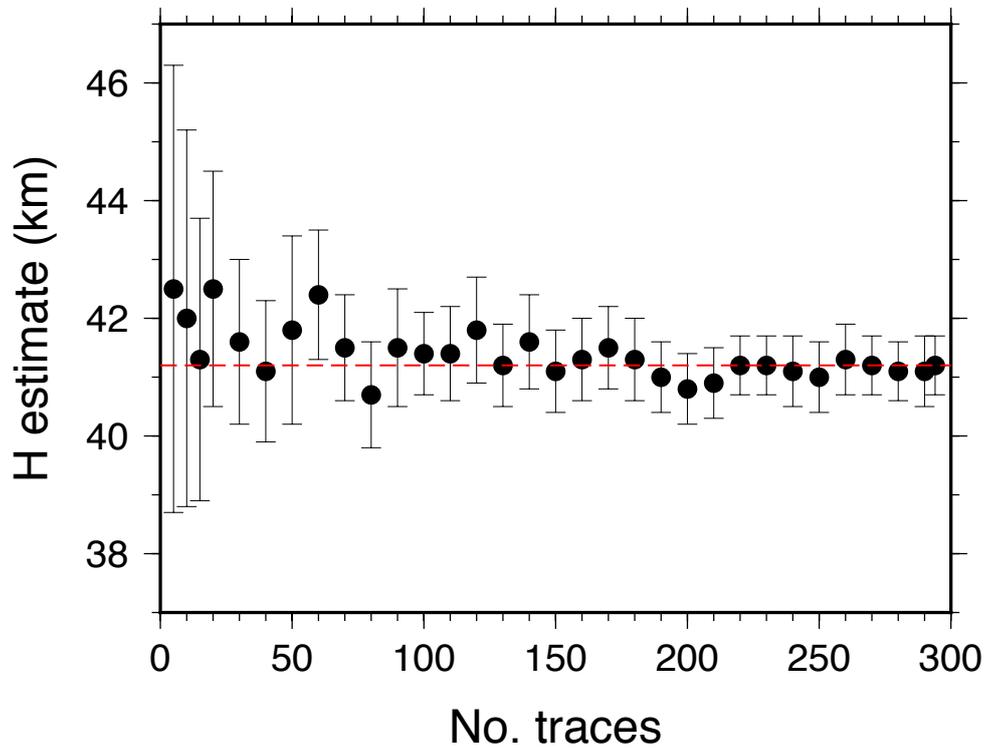


Figure 3: Migration-based results from station WRAB for a range of data subsets (traces varying from 5 to the whole dataset of 294). All subsets bar the 60-trace subset lie within error of the whole dataset result of 41.2 ± 0.5 km (dashed line).

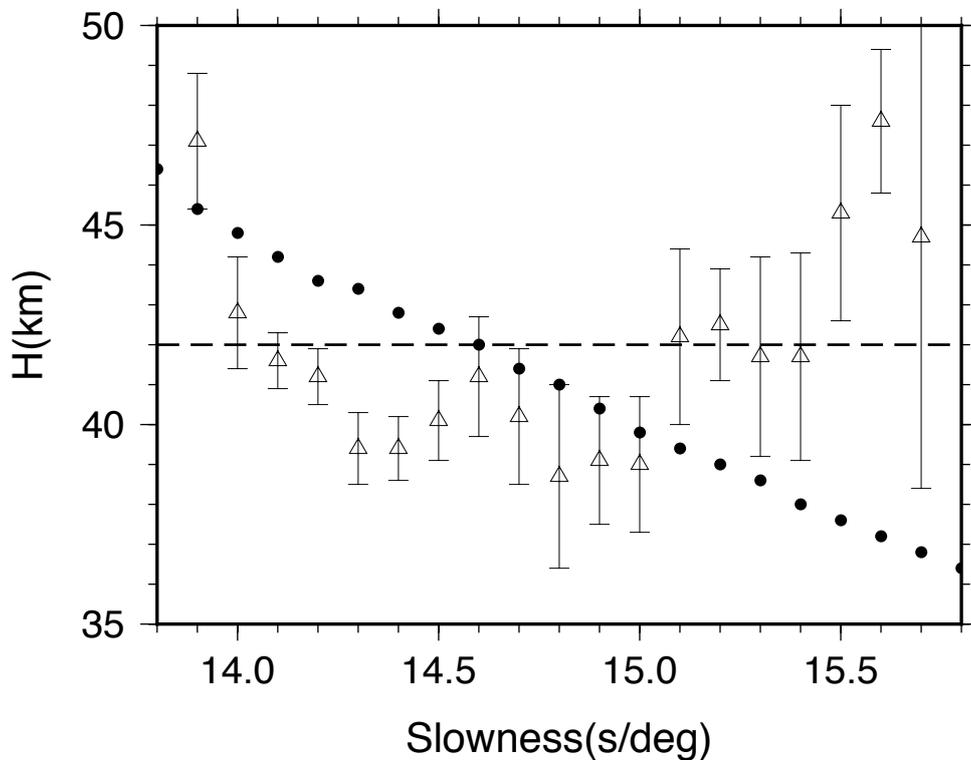


Figure 4: Comparison between the results of the VDSS trace migration approach using synthetic data for a 42 km thick crust with a uniform V_p of 6.5 km s^{-1} (filled circles) and WRAB data sorted into 0.2 s deg^{-1} bins with 50% overlap between bins (triangles). The dashed line is the true crustal thickness of the synthetic model (42 km).

WRAB

H: 39.6 \pm 3.3 km

Vp: 6.36 \pm 0.16 km/s

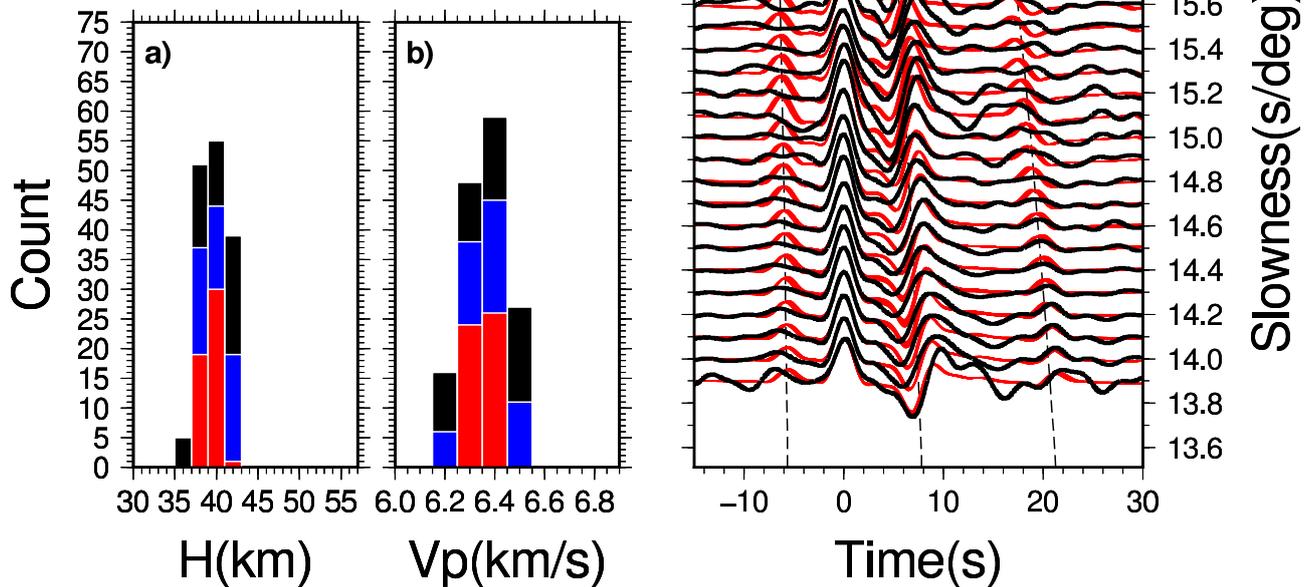


Figure 5: Results of the simultaneous H-Vp inversion for WRAB. a) Histogram of crustal thickness showing the best fitting 50 (red), 100 (blue) and 150 (black) models. b) Histogram of crustal V_p (colour convention follows a)). c) Stacked VDSS traces with slowness. The synthetic traces for the best fitting 50 models are plotted in red.

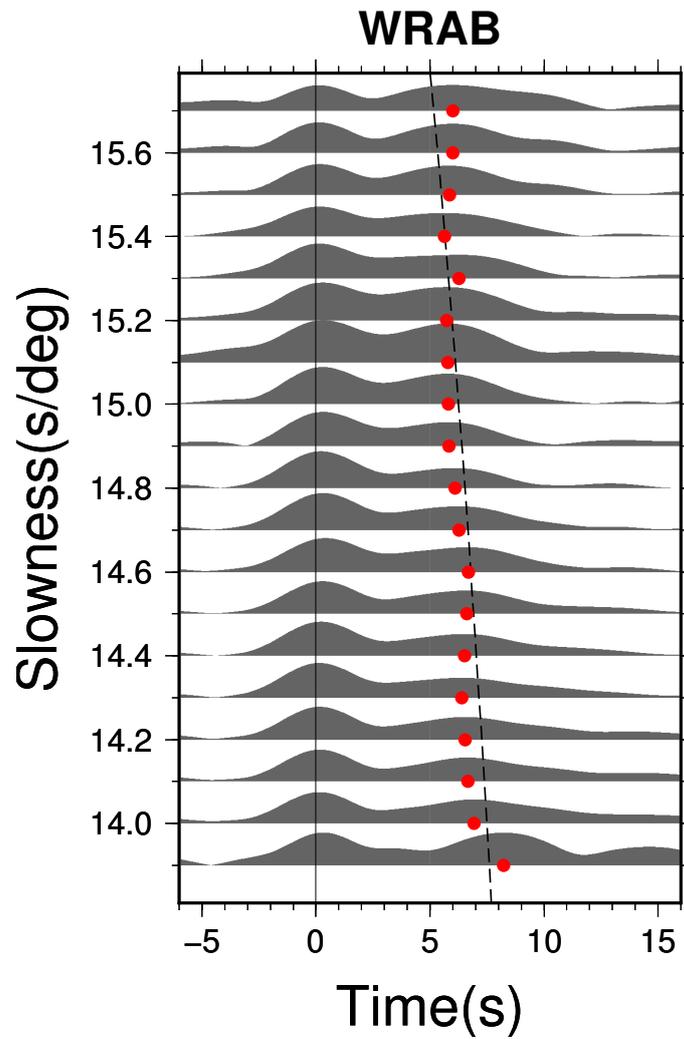


Figure 6: Envelope functions for station WRAB. Red dots are the maximum amplitude picks for the *SsPmp* phase and the dashed curve is the predicted moveout curve for the result of the migration based approach with an assumed crustal V_p of 6.5 s km^{-1} .

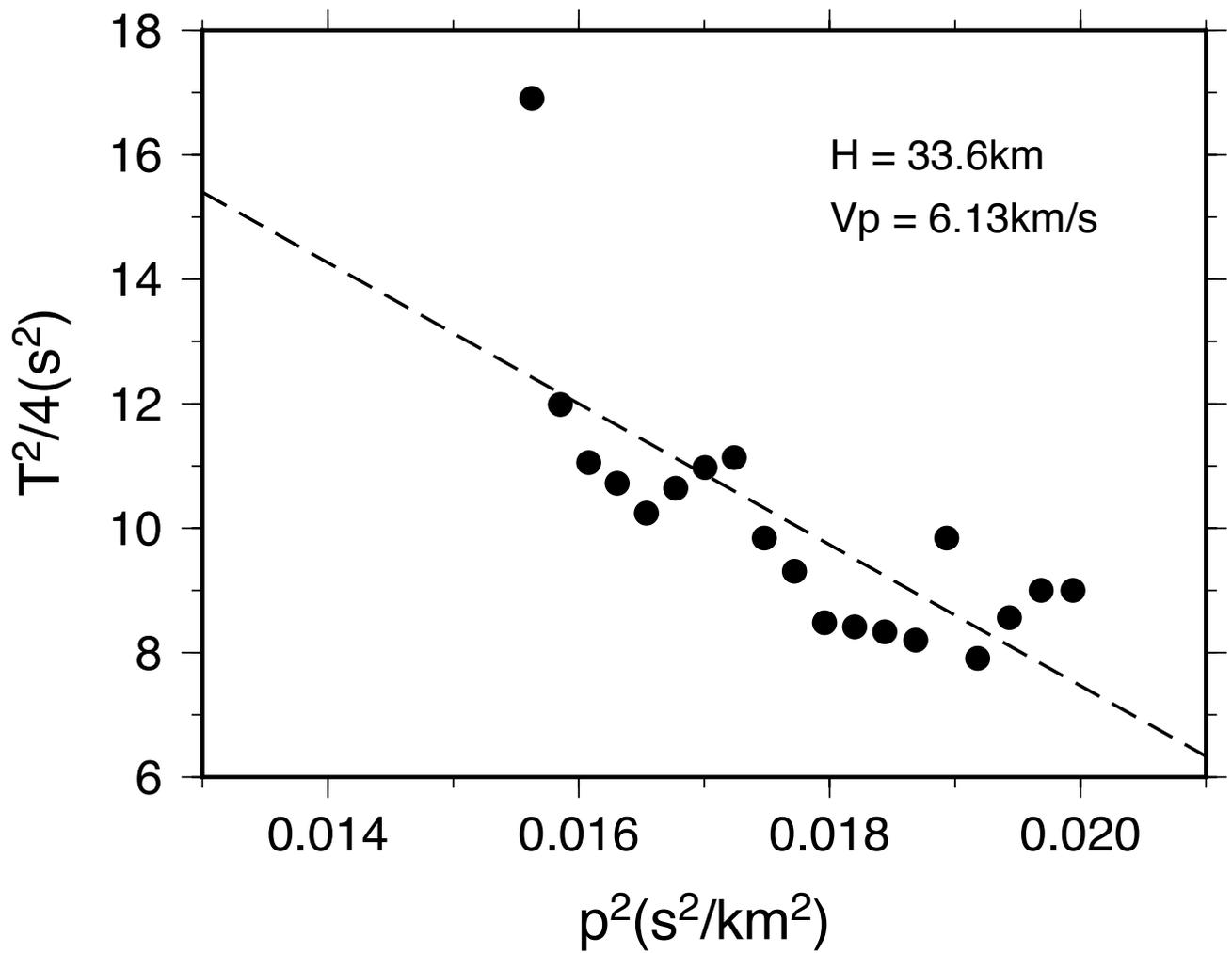


Figure 7: Result of the linear regression for WRAB. Data points are the travel time picks from Fig. 6 in p^2 vs. $T^2/4$ space and the dashed line is the best fitting straight line.

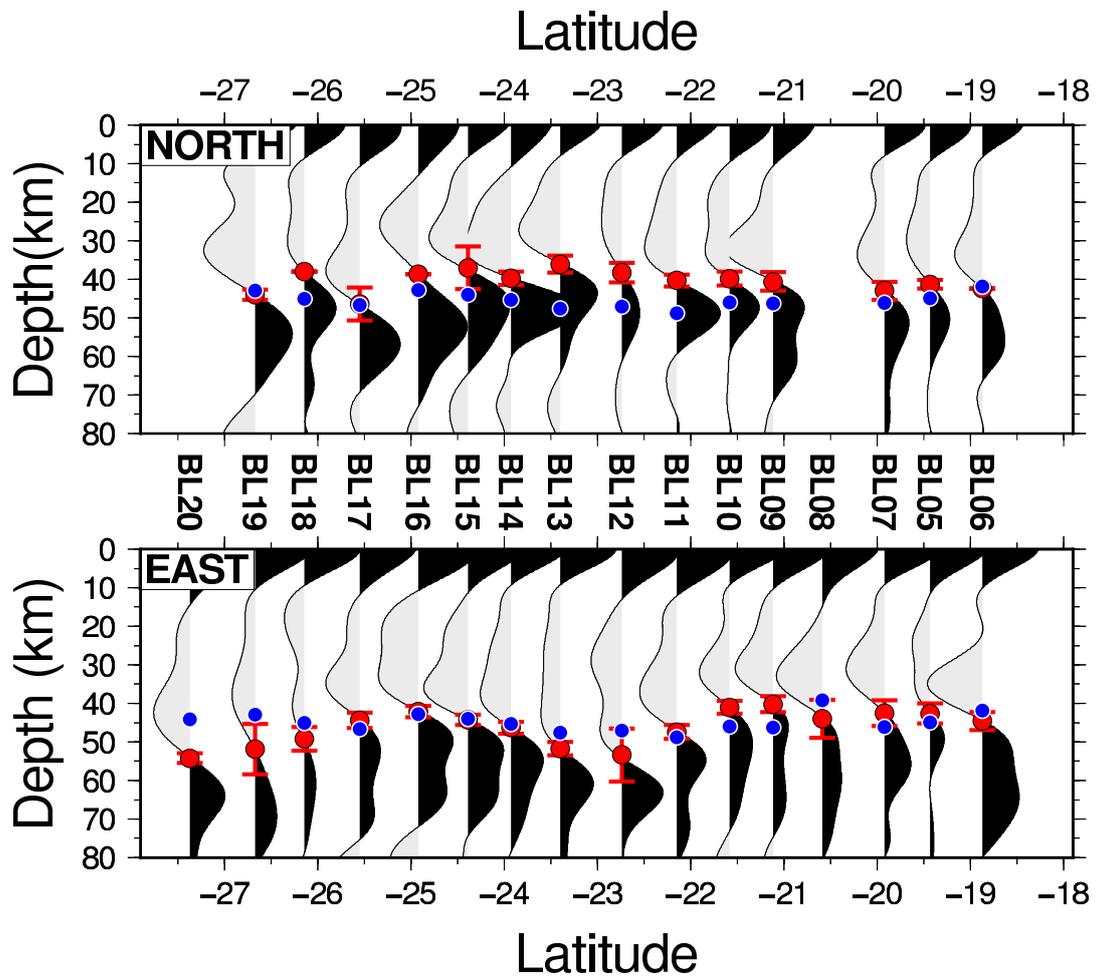


Figure 8: Cross-sections along the BILBY line both for the Java subduction source region (north) and the Tonga-Fiji subduction source region (east). Crustal thickness estimates from the VDSS trace migration-based approach (red symbols) along with the migrated VDSS traces are plotted beneath the surface location of the corresponding station. Blue symbols are the crustal thickness estimates from the nearest grid point of the AusMoho model (Kennett et al., 2011).

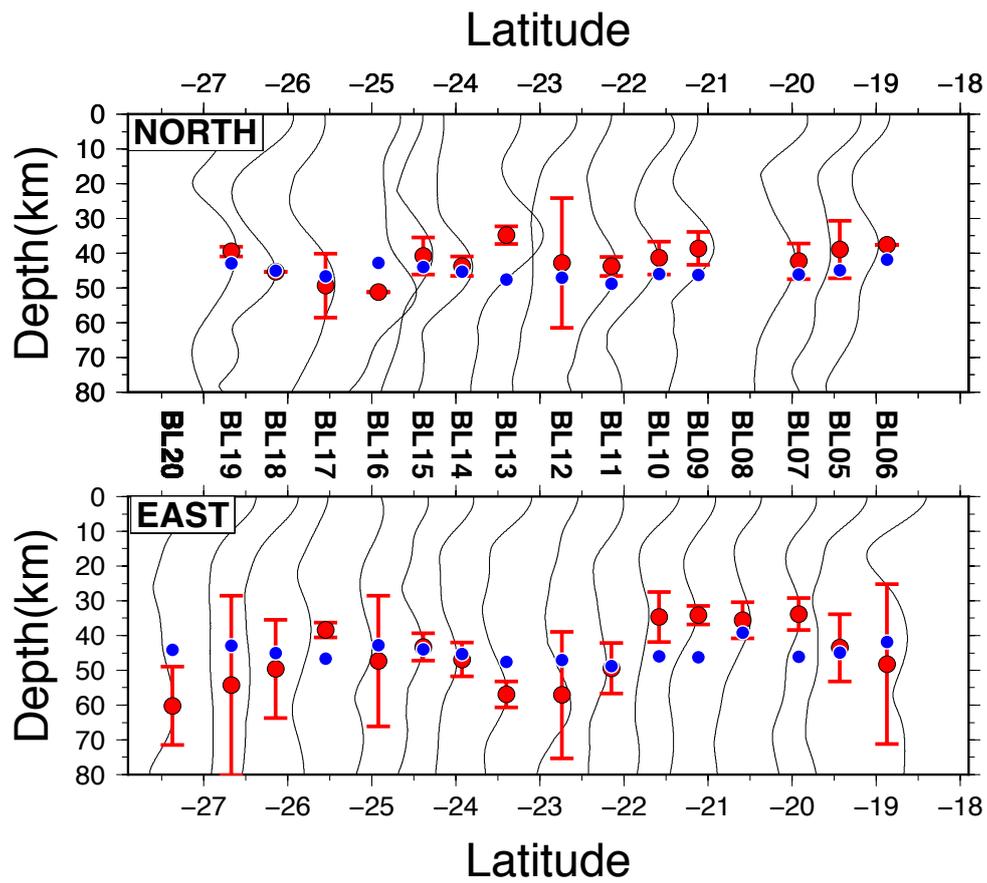


Figure 9: Cross-sections along the BILBY line both for the Java subduction source region (north) and the Tonga-Fiji subduction source region (east). Crustal thickness estimates from the envelope function migration-based approach (red symbols) along with the migrated envelope function traces are plotted beneath the surface location of the corresponding station. Blue symbols are the crustal thickness estimates from the nearest grid point of the AusMoho model (Kennett et al., 2011).

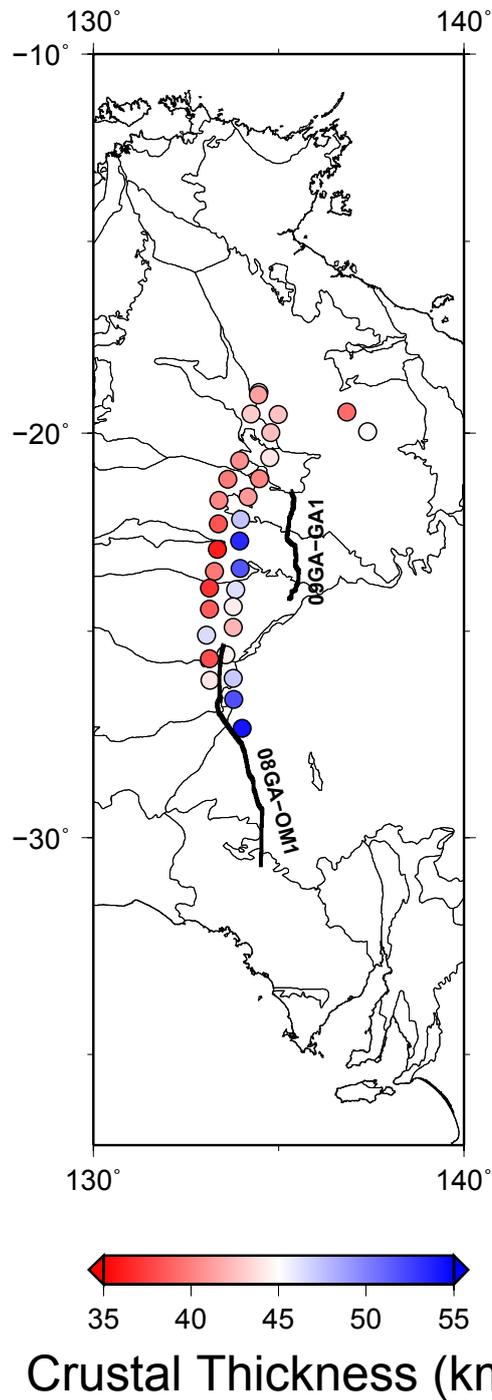


Figure 10: Map view of VDSS-derived crustal thickness variations plotted at the average location of the Moho piercing point for the northerly and easterly data subsets. The location of the two active source lines used for comparison, 08GA-OM1 (GOMA) and 09GA-GA1 (09GA), are also plotted.

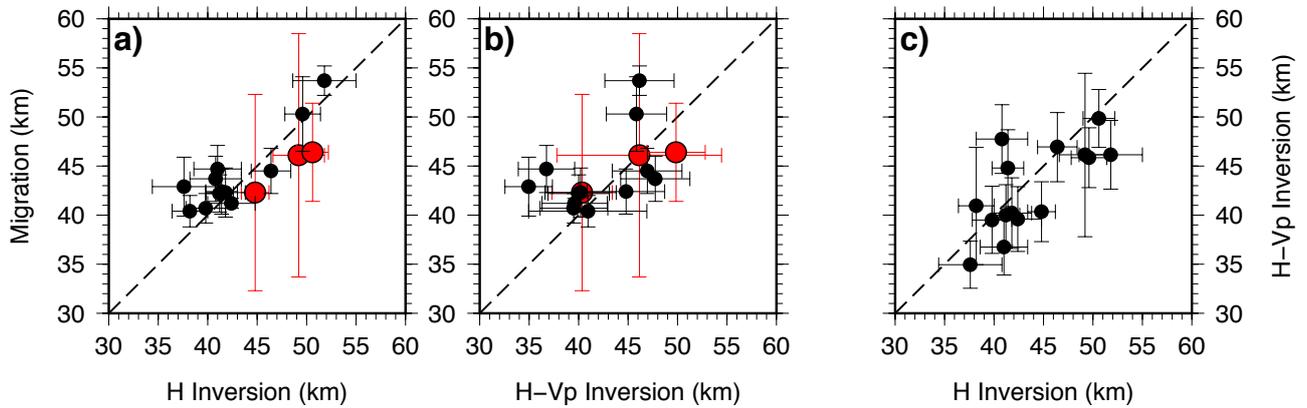


Figure 11: Comparison of single station crustal thickness estimate results from the different approaches used in the study (all available stations). Comparison of a) the migration approach with the single parameter inversion, b) the migration approach and the simultaneous H-Vp inversion, and c) the single parameter inversion and the simultaneous H-Vp inversion. Red symbols in a) and b) are stations BL12, BL13 and BL18 for which significant back-azimuthal variations are observed.

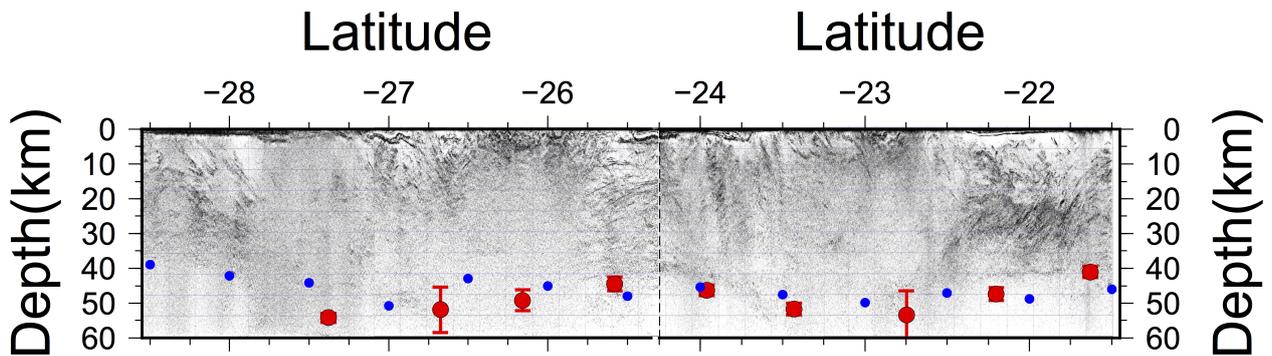


Figure 12: Comparison between VDSS-derived crustal thickness from this study (red circles, easterly back-azimuths), AusMoho crustal thickness estimates (blue circles; Kennett et al., 2011) and the migrated sections from the GOMA/09GA active-source lines acquired from the atlas of deep crustal seismic reflection profiles provided by Kennett et al. (2016). The migrated sections from 09GA and GOMA (see Fig. 10 for true location) have been translated onto a N-S cross-section.