

Structural controls on localized intraplate deformation and seismicity in Southern Australia: Insights from local earthquake tomography of the Flinders Ranges

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[1] Data from an array of 24 seismometers are used to image the crust beneath the Flinders Ranges, southeast Australia, with the goal of improving our understanding of crustal structure, rheology, and the mechanism responsible for the localized intraplate deformation that characterizes this region. A subset of P - and S -wave traveltimes is inverted to jointly recover earthquake hypocenters, P -wave velocity structure and v_p/v_s anomalies. The P -wave velocity model reveals a spatial correlation between major negative velocity perturbations and concentrations of seismicity. In particular, a cluster of seismicity is observed within a distinct low velocity region between the Archean-Mesoproterozoic Gawler Craton and the Palaeo-Mesoproterozoic Curnamona Province, from 7 to 20 km depth. We postulate that this may be associated with a pre-existing structural weakness in the crust that arises primarily from rifting between the Curnamona Province and the Gawler Craton. Another area characterized by a high level of seismicity overlies a major sequence of N-S trending Ross-Delamerian thrust faults, which correspond to a band of low v_p and particularly v_p/v_s . The lack of evidence for elevated heat flows in both of these seismogenic regions suggests that thermally induced weakness is unlikely to play a dominant role. Instead, the dynamic behavior of this intraplate region appears to be caused by a serendipitously oriented regional stress field, provided by far field forces that originate from the boundary between the Pacific and Australian plates, which acts upon preexisting structural weaknesses in the lithosphere.

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1. Introduction

[2] The majority of global seismic energy release occurs along the boundaries of tectonic plates, particularly those undergoing active convergence, where the processes of collision and subduction yield many large earthquakes. Less well understood are the factors responsible for localizing intraplate seismicity and deformation [Hillis *et al.*, 2008], a phenomenon that is responsible for the creation of numerous landforms found within continental interiors. There is much conjecture as to why concentrations of

intraplate events occur, although most hypotheses focus on the role of mechanically (e.g., faults, shear zones, failed rift, flexural deformation) and/or thermally (e.g., high heat flows, volcanic processes) induced weakness of the lithosphere [Sykes, 1978; Sandiford, 2008; Holford *et al.*, 2011; Assumpção and Sacek, 2013]. As well as helping to shed light on these matters, the study of intraplate earthquakes is also important for understanding the associated seismic hazard, which can be significant; e.g., the New Madrid seismic zone, Missouri, experienced three $M > 7$ earthquakes in 1811–1812 [Tuttle *et al.*, 2002], while more recently, a 7.6 M_w earthquake occurred in the region of Gujarat, India, in January 2001 [Negishi *et al.*, 2002; Bodin and Horton, 2004].

[3] Although there is a widespread belief that the Australian continent is a tectonically and geomorphologically quiescent intraplate region, it exhibits a significant level of earthquake activity (Figure 1), in concert with a rich record of Neogene and Quaternary faulting [Sandiford, 2003; Célérier *et al.*, 2005; Quigley *et al.*, 2006; Hillis *et al.*, 2008], symptomatic of a landscape that appears to be, in some areas, quite youthful and dynamic [Quigley *et al.*, 2010]. Indeed, one such area is the Flinders Ranges

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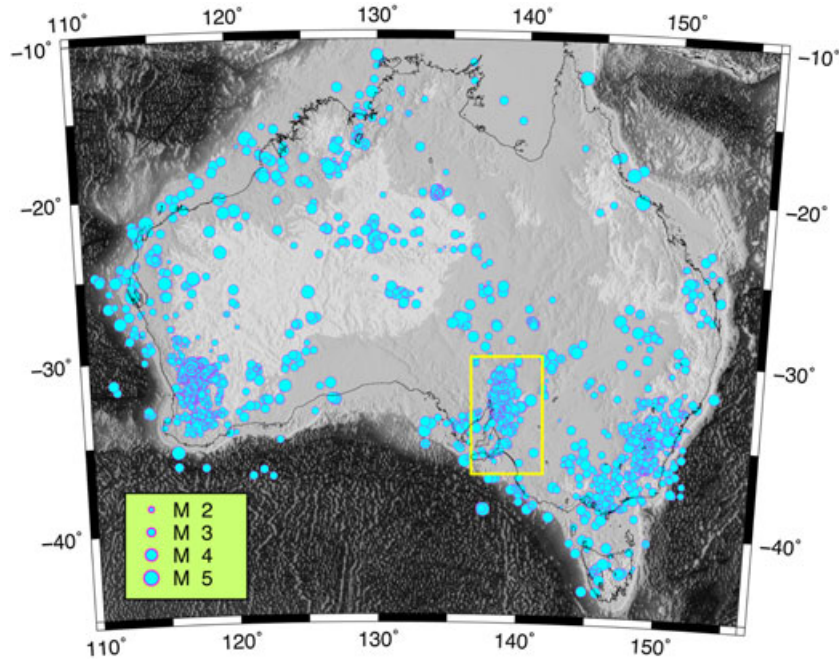


Figure 1. Distribution of seismicity in Australia over the last decade (source: Geoscience Australia earthquake database). The yellow rectangle highlights a region of relatively high seismicity, which corresponds to the Flinders Ranges.

of South Australia, which contrasts considerably with the surrounding regions (e.g., Gawler Craton and Curnamona Province) (Figure 1). The area encompassing the Flinders Ranges is dominated by earthquakes with strike-slip and reverse fault focal mechanisms [Greenhalgh *et al.*, 1994; Clark and Leonard, 2003; Hillis and Reynolds, 2003] with seismogenic strain rates up to $\sim 10^{-16} \text{ s}^{-1}$ and a maximum horizontal compressive stress direction of $\text{N}83 \pm 30^\circ \text{E}$ [Hillis and Reynolds, 2000, 2003]. The largest earthquake ever recorded in the Flinders Ranges has a magnitude of $M_s \sim 6$, although the typical value is about 3. Recently, two studies by Quigley *et al.* [2007a, b] in the Flinders Ranges have documented new and persuasive evidence from cosmogenic ^{10}Be concentrations for the formation of relatively young relief. From these new results, it has been inferred that late Neogene tectonism led, in as little as 4 My, to the formation of the entire Flinders Ranges (maximum elevation $\sim 1200 \text{ m}$); likewise, they found that the Flinders Ranges has undergone a surprisingly high erosion rate for a supposedly ancient and stable landscape. According to Sandiford *et al.* [2004], the deformation and Neogene orogenesis in southeastern Australia may have been the result of increased coupling between the Australian and Pacific Plate during the late Miocene, implying that plate boundary forces may be transmitted over thousands of kilometers into plate interiors, as recently inferred by Dyksterhuis and Müller [2008]. The long-wavelength ($\sim 150 \text{ km}$), low amplitude (200–500 m) crustal folding in the gross structure of the Flinders Ranges, manifest by a broad uplift bound by gently outward sloping basins, is consistent with a localized strain response to stresses applied in the far field [Célérier *et al.*, 2005]. Another key feature is that in this part of South Australia, the Proterozoic crust is characterized by anomalous elevated heat flows, which reflect enrichment in U, Th,

and K [Neumann *et al.*, 2000]. However, due to limited heat flow measurements, the temperature structure beneath the Flinders Ranges is still poorly known.

[4] The geodynamic framework and neotectonic activity of the Flinders Ranges suggest that the lithosphere must be mechanically weak relative to surrounding regions. However, the cause of the observed deformation remains controversial, with most explanations lying between two end-member models: (i) reactivation of zones of prior deformation caused by the increase in stress levels in the Indo-Australian plate due to changes in far-field boundary forces [Célérier *et al.*, 2005], and (ii) the presence of high heat flows, which are consistent with thermal weakening of the lithosphere [Sandiford, 2008; Braun *et al.*, 2009; Holford *et al.*, 2011].

[5] Similar examples have been found in other areas around the globe. For example, Liu and Zoback [1997] propose a model in which the localized seismicity and deformation in the New Madrid seismic zone is driven by a hotter and weaker crust, owing to a thermal anomaly (average value of 58 mW m^{-2}). However, further investigations confirmed that the surface heat flow is much smaller than previously assumed ($\sim 3 \pm 15 \text{ mW m}^{-2}$), indicating that the lithospheric strength of the New Madrid seismic zone is overall the same as its surroundings [McKenna *et al.*, 2007]. Instead, McKenna *et al.* [2007] attribute the localized intraplate seismicity to a mechanically weak area (e.g., failed rifting). Another region of pronounced intraplate seismicity is eastern mainland China (i.e., North China Craton), which has experienced several strong earthquakes ($M_s > 7$) [Wang *et al.*, 1997]. According to Wang [2001] and Wang and Cheng [2011], who have analyzed more than 2000 heat flow measurements in mainland China, there is a clear relationship between thermal weakening of

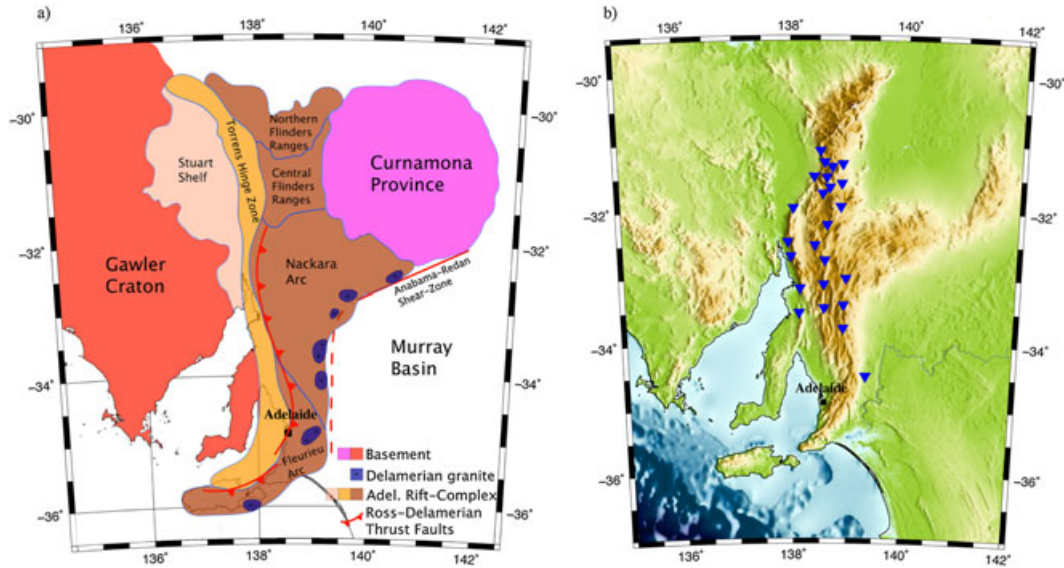


Figure 2. (a) Schematic geological map of central South Australia showing the Adelaide Rift Complex and the two basement structures (Gawler Craton and Curnamona Province). Approximate location of the Ross-Delamerian thrust faults and granites are also highlighted. (b) Location of both temporary and permanent seismometers used in this study (inverted blue triangles) along with topography. The Flinders Ranges is clearly revealed by elevated topography.

the lithosphere and localized intraplate deformation, implying that the lithospheric strength is mainly influenced by temperature variations.

[6] In this study, Local Earthquake Tomography is carried out using passive seismic data collected in the Flinders Ranges region in order to simultaneously improve hypocenter location and recover 3-D variations in v_p and v_p/v_s structure. Armed with new information on the distribution of earthquakes and variations in seismic properties throughout the crust, our goal is to better understand the origins of intraplate deformation and seismicity in Australia.

1.1. Tectonic Setting

[7] The basement rocks of southern and central South Australia predominantly constitute two Archean-Proterozoic crystalline blocks, which are overlain by a thick Neoproterozoic to Cambrian sedimentary package (average thickness of 10 km [Preiss, 2000]), the Adelaide Rift Complex (or Adelaide Fold Belt). The evolution of the Gawler Craton (Figure 2a), which dominates the western part of South Australia, is essentially expressed through two major cycles of tectonic activity in a time frame that spans ~ 2560 Ma to ~ 1450 Ma [Hand et al., 2007]. During this period, the Gawler Craton was affected by multiple tectonic events, including volcano-sedimentary basin formation, both collisional and extensional deformation, emplacement of magmatic rock suites, and regional cooling, after which the craton appears to have become a relatively stable continental block [Webb et al., 1986; Fanning et al., 1988; Daly et al., 1998; Direen et al., 2005; Hand et al., 2007; Fanning et al., 2007; Thomas et al., 2008; Payne et al., 2009]. The Curnamona Province, a nearly circular-shaped Paleo to Mesoproterozoic crustal province, lies to the east of the Adelaide Rift Complex [Robertson et al., 1998; Connor and Preiss, 2008]. The Curnamona Province consists

primarily of the Paleoproterozoic Willyama Supergroup, deposited during epi-continental rifting and accompanied by felsic and mafic magmatism between ~ 1720 – 1645 , and coeval magmatic rocks [Connor, 2000; Rutherford et al., 2007; Connor and Preiss, 2008]. Two major orogenic events have affected this basement: first, the Olarian Orogeny (~ 1620 – 1580 Ma) and subsequently the Ross-Delamerian Orogeny (~ 530 – 500 Ma) [Page et al., 2005; Foden et al., 2006; Rutherford et al., 2007]. Currently, the Curnamona Province is largely buried and bounded by younger sedimentary cover of Neoproterozoic to Holocene age. Two compelling lines of evidence suggest that from the Late Paleoproterozoic to early Mesoproterozoic, the Gawler Craton and the Curnamona Province were part of a contiguous crustal system, before rifting began at $\sim 827 \pm 6$ Ma [Wingate et al., 1998]; these can be summarized as: (i) geochronological, geochemical, and isotopic affinities between the Benagerie Volcanic Suite of north-central Curnamona Province and the upper Gawler Range Volcanics of the Gawler Craton [Wade et al., 2012], and (ii) similarities in age of early Mesoproterozoic igneous and metamorphic events [Belousova et al., 2006; Hand et al., 2008].

[8] The formation of the Adelaide Rift Complex was principally accomplished through several successive rifting events, although the present distribution of units was largely controlled by shortening during the Ross-Delamerian Orogeny [Preiss, 2010]. Failed rifting and crustal thinning (~ 800 – 516 Ma) marked an extensional period, which led to the formation of a rift- and sag-basin [Von der Borch, 1980; Preiss, 1987; Flottmann et al., 1994; Preiss, 2000]. With the onset of contractional deformation and crustal thickening due to the Ross-Delamerian Orogeny, the formerly normal faults were primarily reactivated as reverse and transpressional faults, which led to basin inversion (i.e., basement-cover décollement or some basement-involved

deformation with thrust-bound pop-up structures in the sedimentary package) and the establishment of complex regional folding [Clarke and Powell, 1989; Flottmann et al., 1994; Paul et al., 1999; Yassaghi et al., 2000; Direen et al., 2005]. Furthermore, it is important to note that the deposition of evaporitic sediments during the earliest stage of basin development played a major role in the tectonic evolution of the northern part of the basin [Paul et al., 1999]. The subdivision of the Adelaide Rift Complex has been established relative to the Delamerian tectonic overprint and is typically divided into five regions: Torrens Hinge Zone, North Flinders Ranges, Central Flinders Ranges, Nackara Arc, and Fleurieu Arc [Rutland et al., 1981; Preiss et al., 1993] (Figure 2a). Immediately to the east of the Stuart Shelf, a Neoproterozoic cover sequence overlying the Gawler Craton, lies the Torrens Hinge Zone, which is a narrow north-south belt of transition between the Gawler Craton and the Adelaide Rift Complex and is characterized by gentle folding [Preiss, 2000] and relatively thin (0–5000 m) Adelaidean sedimentary rocks [Jones et al., 2004]. The Northern Flinders Ranges are an arcuate belt of linear folds with a thick-skinned structural style caused by shortening of about 10–20% [Paul et al., 1999]. The exhumation of the basement at both of its flanks is largely controlled by inversion processes along the former extensional Norwest and Paralana Faults, which effectively represent the current boundaries of the Gawler Craton and Curnamona Province, respectively [Paul et al., 1999]. The Central Flinders Ranges are situated between the two main provinces of South Australia where the Ross-Delamerian Orogeny is least pronounced, with a shortening of generally less than 10%. Furthermore, Backé et al. [2010] have shown that here sediment thickness is less than other areas, and that distinct salt diapirs exist immediately above basement-penetrating normal faults. Despite the extensive body of knowledge surrounding the Flinders Ranges region, the basement depth, as well as the relationship between the Gawler Craton and Curnamona Province, is not well constrained due to the paucity of geophysical data and deep drill holes that intersect the basement [de Vries et al., 2008]. The Nackara Arc consists of a relatively long arcuate belt with North-South striking folds and faults formed above a regional evaporite-hosted décollement [Marshak and Flottmann, 1996]. In the Fleurieu Arc, the belt consists of a frontal imbricate fan of basement-involved thrusts, although the thicker part of the orogen is characterized by a fold thrust-belt [Flottmann and Oliver, 1994; Direen et al., 2005]. The Fleurieu Arc records both the earliest deformation and the maximum shortening (about 50%) [Marshak and Flottmann, 1996]. The internal zone of the Adelaide Rift Complex is also characterized by occurrences of equant granites that cut along the predominant fabric or, particularly the syn-orogenic granites, tend to be elongated along strike. They range in age from Mid-Cambrian to Early Ordovician [Foden et al., 2002].

2. Data and Method

[9] The data used in this study are sourced from a network of broadband seismometers, which was deployed to record earthquakes in the Flinders Ranges as a collaborative effort between Geoscience Australia, the Australian National University, and the Department of Primary Industries and

Resources, South Australia, now Department for Manufacturing, Innovation, Trade, Resources and Energy. Figure 2b shows the location of all recording stations associated with the experiment, which include nine permanent stations and 15 temporary stations that were in place between September 2003 and June 2005 [Cummins et al., 2004]. Approximately 400 earthquakes were detected by the network resulting in a data set that comprises a total of 2628 *P*-traveltimes and 2370 *S*-traveltimes. These arrival times were automatically identified from the seismic records using the multi-frequency STA/LTA detector, *dbdetect*, and a spatial grid search based locator, *dbgrassoc*, both routines of the *BRTT Antelope* software. Automatic picks were subsequently reviewed manually using *dbpick*. The RMS picking uncertainty is approximately 0.21 s for *P*-arrivals and 0.29 s for *S*-arrivals. The initial location of the earthquakes were determined by using the routine *dblocs*, based on *locs* by Bratt and Bache [1988], which exploits a nonlinear iterative least squares method. However, these are not ultimately used in the final results, except to calculate the initial traveltime misfit. The 1-D starting model used in the inversion of both *P*- and *S*-waves is a smoothed version of the layered model derived by Shackleford and Sutton [1981]. They recorded seismic data along two linear refraction profiles, approximately parallel and transverse to the N-S axis of the Adelaide Rift Complex, by using explosions at two large open cut mines as sources of seismic energy to provide one seismic velocity model for *P*-waves and one for *S*-waves. They found that the *P*-wave velocities in the upper crust (18 km), lower crust (36 km), and in the upper mantle (50 km) are 5.94, 6.46, and 7.97 km/s respectively, whereas the corresponding *S*-wave velocities are 3.43, 3.76, and 4.45 km/s. For depths greater than 50 km, the *ak135* velocity model developed by Kennett et al. [1995] has been used. Although using a data constrained 1-D starting model may be beneficial [Kissling et al., 1994], our approach of using a well constrained starting model has been widely used in seismic tomography [Eberhart-Phillips, 1990; Dzierma et al., 2012]. We also investigate the sensitivity of the 3-D solution model to the choice of initial model by running a series of tests, which involve applying sizeable perturbations to the smoothed version of the Shackleford and Sutton [1981] model prior to inversion (see Figure fs01 of Supporting Information). Results show that (i) the traveltime residuals associated with the Shackleford and Sutton [1981] model were always lower than those obtained for other initial models, and (ii) all solution models were very similar, with differences only occurring in regions of poor data coverage that are well identified by the synthetic resolution tests undertaken in the next section.

[10] A sophisticated and robust iterative nonlinear tomographic inversion package, called FMTOMO, has been modified and used to perform the inversion for hypocenter location and 3-D variations in v_p and v_p/v_s . The roots of FMTOMO began with the work of Rawlinson and Sambridge [2004a,b], in which the core innovation was the application of a fast and robust grid based eikonal solver, known as the fast marching method or FMM [Sethian, 1996; Sethian and Popovici, 1999], to solve the forward problem of traveltime prediction in laterally heterogeneous layered Earth models. For the inversion step, FMTOMO exploits a gradient-based subspace inversion scheme

[Kennett *et al.*, 1988] to solve the linearized problem of matching observed and predicted traveltimes subject to damping and smoothing regularization. Iterative application of FMM and subspace inversion aims to address the nonlinear nature of the inverse problem. A detailed description of the forward scheme implemented by FMTOMO is given by De Kool *et al.* [2006], while Rawlinson *et al.* [2006] describe how FMTOMO accomplishes the inversion step. Within each layer, the velocity field is defined by a regular 3-D grid of nodes, which constitute the control vertices of a mosaic of cubic B-spline volume elements used to create a smoothly varying, locally controlled velocity continuum.

2.1. Nonlinear Location of Events

[11] The assumption of local linearity used by FMTOMO to simultaneously adjust velocity and hypocenter parameters is employed by most Local Earthquake Tomography algorithms [Thurber, 1983; Evans *et al.*, 1994; Koulakov, 2009]. A drawback of linearization is that the source location problem is generally much more nonlinear than the velocity recovery problem, which means that a linearized approach can perform poorly in regions of significant velocity heterogeneity and/or poor initial source locations. Preliminary tests with FMTOMO applied to the Flinders Ranges data set showed that inversions for velocity structure with and without relocation turned on resulted in only small improvements in data fit; however, in the former case, final locations often significantly differed to reference locations. This implies that the assumption of local linearization used by FMTOMO is not valid for this data set and set of initial conditions.

[12] As a consequence, a new nonlinear relocation algorithm, which takes advantage of the grid-based nature of fast marching method (FMM), is developed to improve the source relocation problem. In order to compute two-point traveltimes, eikonal solvers like FMM calculate the traveltime to all points in the medium from each source. To maximize computational efficiency, the FMM source points can in fact be the receivers, if the number of earthquakes is larger than the number of receivers (invoking traveltime reciprocity), which is the case for this data set. The availability of the complete traveltime field for each receiver means that a fully nonlinear relocation scheme that is also computationally efficient is relatively simple to devise.

[13] Consider an objective function $S(\mathbf{m})$ defined by

$$S(\mathbf{m}) = (\mathbf{d}_{\text{obs}}^r - \mathbf{d}_{\text{m}}^r)^T \mathbf{C}_0^{-1} (\mathbf{d}_{\text{obs}}^r - \mathbf{d}_{\text{m}}^r) \quad (1)$$

where $\mathbf{d}_{\text{obs}}^r$ is a set of observed arrival times (from a given hypocenter) at a set of N stations with the mean subtracted, $\mathbf{C}_0^{-1}$ is the data covariance or weighting matrix, and $\mathbf{d}_{\text{m}}^r$ is a set of predicted traveltimes also with the mean subtracted. Removing the mean is useful, because it eliminates the origin time as an unknown in the inversion. Once $S(\mathbf{m})$ has been minimized, the origin time can be found retrospectively by using $\mathbf{d}_0 = \mathbf{d}_{\text{obs}} - \mathbf{d}'$, where $\mathbf{d}'$ are the observed traveltimes. Thus, the minimum of $S(\mathbf{m})$ can be readily found by a simple grid search technique, which compares the difference between each observation and each grid point in the traveltime field for all receivers. In order to achieve greater accuracy, sub-cell discretization based on interpolation of the traveltime fields is implemented. Although more sophisticated search algorithms (e.g., genetic algorithms, simulated

annealing) could be used, computation times for the nonlinear relocation of all events are of the order of 60 s, which is relatively quick. Despite the fact that the location algorithm is fully nonlinear, iterative application is still required due to the trade-off between velocity variation and source location. Ideally, it would be preferable to use a fully nonlinear inversion scheme for both velocity and source location, but the computational cost would be too great. The compromise approach taken here appears to work well in the presence of realistic crustal velocity variations, as demonstrated below.

2.2. Direct Inversion for v_p/v_s

[14] The simplest and ostensibly logical way of computing v_p/v_s is to just divide the P -wave model by the S -wave model. However, this approach has drawbacks, particularly when P - and S -wave data coverage differ, and regularization is imposed to combat solution non-uniqueness. Initial tests using this procedure, i.e., taking the ratio of separate v_p and v_s models, resulted in a pattern of perturbations that appeared to largely mimic the v_p anomalies. Standard regularization techniques (damping and smoothing) typically apply relatively arbitrary constraints on the amplitude of anomalies to the extent that interpretation generally relies on the pattern of anomalies being correct, more so than the absolute amplitude. However, the pattern of v_p/v_s anomalies has a dependence on the amplitude of v_p and v_s . Moreover, if the coverage of S -waves is much poorer than that of P -waves (which is often the case), the resultant S -wave model will tend to be smoother than the P -wave model, which can result in a v_p/v_s model that is imprinted by the smaller wavelength features inherited from the P -wave model. With the Flinders Ranges data set, a direct inversion of S - P differential traveltimes has been performed (see Walck [1988]; Thurber [1993]; Eberhart-Phillips and Reyners [2012] for more details on the method), which is principally based on three assumptions: (i) for each S -wave path between two different points, there is a corresponding P -wave path, (ii) P - and S -wave paths taken between two different points are identical, and (iii) the reference v_p/v_s model is constant. Instead of using the same formulation as Walck [1988], which is based on backprojection tomography using constant slowness cells, where the final equation requires perturbations in P -wave slowness to be computed prior to updating v_p/v_s , we use the expression

$$\delta \left(t_s - \left(\frac{v_p}{v_s} \right)^0 t_p \right) = \int_{L_s} \frac{1}{v_p^0} \delta \left(\frac{v_p}{v_s} \right) dl \quad (2)$$

where t is the traveltime for P - or S -waves along the S -wave path L_s ; v is the velocity, l is the path length, and $(v_p/v_s)^0$ is a reference model. Thus, v_p/v_s can be obtained directly without explicit knowledge of v_p or v_s , and within a purely linear framework. The synthetic tests presented in the next section clearly demonstrate that the assumptions inherent to this technique have a lesser influence on the results than *ad hoc* regularization.

[15] Another approach for constraining v_p/v_s that is commonly used requires v_p and v_s to be inverted for simultaneously with a regularization term used to control variations in v_p/v_s ratio [Michellini, 1993; Tryggvason and Linde, 2006]. This has the advantage that P - and S -wave coverage need not be identical, and that the final v_p and v_s models are fully

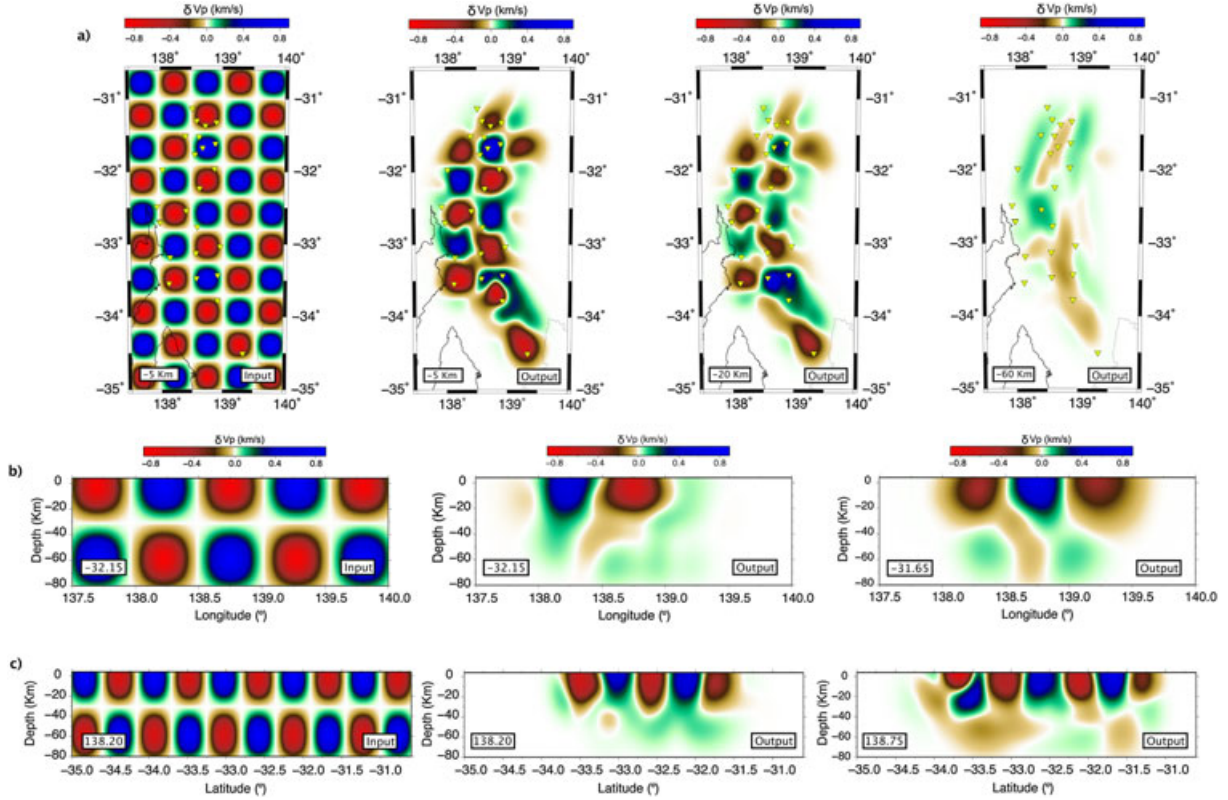


Figure 3. Checkerboard recovery test results for v_p . (a) Depth slices, (b) east-west cross sections, and (c) north-south cross sections. Gaussian noise with a standard deviation of 0.1 s was added to the synthetic traveltimes. Yellow inverted triangles represent receivers.

consistent with the v_p/v_s model. The regularization imposed on v_p/v_s usually comes in the form of damping back to a reference model. However, the reference v_p/v_s model can be quite approximate, so the regularization may be relatively *ad hoc*. Moreover, given that event locations are often determined simultaneously, and damping and smoothing regularization are applied, the number of “tuning parameters” is increased, which makes the choice of final model difficult to justify.

3. Results

[16] In order to image seismic structure in the crust beneath the Flinders Ranges, FMTOMO is used to invert for 3-D v_p and v_p/v_s variations and relocate hypocenters. We define a propagation grid, in which to perform the forward step and the nonlinear relocation, that comprises a total of more than 2 millions nodes that span 80 km in depth, 4.4° in latitude and 2.5° in longitude, with a node separation of approximately 2 km in each direction. The inversion grid spacing, which defines the smooth velocity model, is equal to ~ 12 km in latitude and longitude and ~ 11 km in depth. The complete inversion procedure is accomplished by running six iterations of a 20-dimensional subspace inversion routine with damping and smoothing parameters set according to the results of multiple trade-off curve tests (see Rawlinson *et al.* [2006] for more details), which target a model which is as smooth and close to the initial model as possible while still satisfying the data. Between each iteration, the nonlinear relocation procedure is performed using

P -wave traveltimes only (which are more accurately picked than S) and adopting a minimum relocation threshold of five observations per event, after which traveltimes are recalculated using FMM. Due to the commonality of P -wave and S -wave sources, both a P -wave and an S -wave model are recovered using the subspace technique. The recovery of v_p/v_s structure takes place using the final hypocenter locations.

3.1. Solution Quality

[17] Using the parameter values and regularization discussed above, a resolution test based on synthetic data is performed. This makes it possible to investigate solution robustness, which is dependent on path coverage and data noise. Here, this is done by applying the so-called synthetic “checkerboard test” [Hearn and Clayton, 1986; Glahn and Granet, 1993; Rawlinson and Sambridge, 2003], which involves using an identical source-receiver path configuration to the observational data set to predict traveltime residuals for a predetermined checkerboard structure defined by a pattern of alternating high and low velocity anomalies. In this case, we define both a v_p and v_s checkerboard model, which allows a synthetic v_p/v_s checkerboard model to be defined by simply dividing v_p by v_s . Gaussian noise with a standard deviation of 100 ms is added to all synthetic traveltimes in order to simulate the picking uncertainty associated with the observational data set.

[18] Figures 3 and 4 show the result of the synthetic checkerboard test (via a series of horizontal and vertical

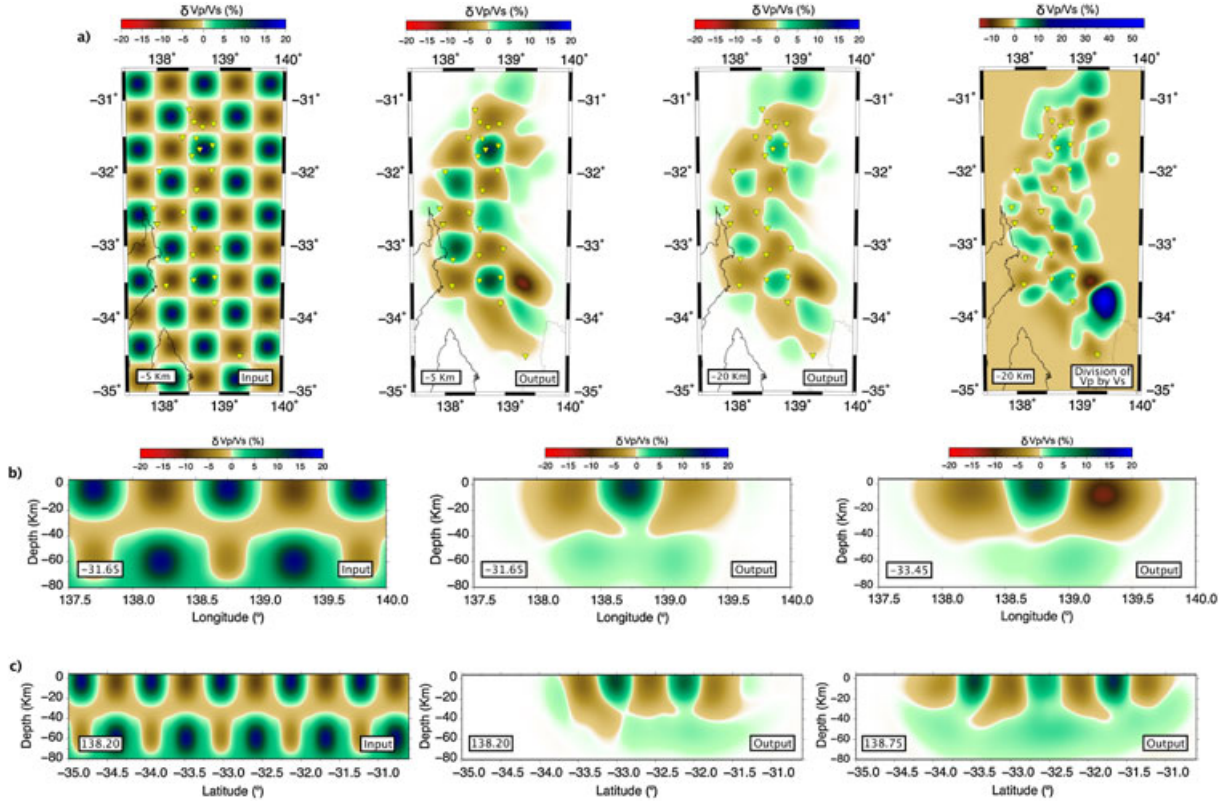


Figure 4. Checkerboard recovery test results for v_p/v_s . (a) Depth slices, (b) east-west cross sections, and (c) north-south cross sections. Gaussian noise with a standard deviation of 0.1 s was added to the synthetic traveltimes. Note how simply dividing v_p by v_s produces very poor results (top right). Yellow inverted triangles represent receivers.

slices), following six nonlinear iterations for v_p , v_p/v_s , and source location. The quality of the recovered checkerboard pattern is generally good within the crust, with relatively accurate estimates of the perturbation amplitudes, although smearing of the velocity model is evident in places, particularly in regions peripheral to the horizontal bounds of the receiver array. Resolution is progressively lost below ~ 40 km depth, where the pattern of anomalies becomes blurry and smeared out. This is caused by the predominance of crustal sources, and the tendency of waves to refract back towards the surface immediately below or at the Mohorovicic Discontinuity (Moho). The poor resolution below the southern tip of the array is due to the station configuration, with only a single isolated receiver that is somewhat removed from the rest of the array.

[19] To assess the reliability of the linear inversion method used for constraining the v_p/v_s ratio, we compare the result of the linear inversion with that produced by simply dividing the recovered v_p model by the recovered v_s model at 20 km depth (Figure 4). Clearly, the direct division approach does not produce high quality results, and demonstrates that with this class of problem, the use of conventional *ad hoc* regularization influences the result more so than approximations to the forward problem.

[20] Overall, these results indicate that our data set is sufficient to detect a realistic range of velocity anomalies in the crust beneath the Flinders Ranges.

3.2. Tomographic Solution Models: v_p and v_p/v_s

[21] The v_p solution model, obtained jointly with the fully nonlinear relocation of sources, reduces the data variance by 63%, which corresponds to an RMS reduction from 506 to 323 ms. The normalized χ^2 value shows a reduction from 13.8 to 4.9, which indicates that the final model does not fully satisfy the data. This is a common phenomenon [Rawlinson *et al.*, 2010] when dealing with real applications due to (i) estimates of data uncertainty being difficult to make, (ii) the approximate nature of the forward solver, (iii) the use of a regular parameterization, and (iv) the use of *ad hoc* smoothing and damping regularization. The last two factors severely limit the range of models that can be produced. Nevertheless, the final data fit is much better than that of the starting 1-D model, which indicates that the recovered lateral heterogeneity is largely required by the data.

[22] The crustal structure beneath the Flinders Ranges is illustrated via a series of horizontal (Figures 5a and 5b) and east-west (Figure 5c) slices taken through the v_p and v_p/v_s solution models (see Figure fs02 of Supporting Information for plots of absolute velocity, and Figures fs03 and fs04 for unprojected vertical and horizontal sections with relocated seismicity superimposed). At shallow depths (first 10 km), the pattern of anomalies is more dominated by smaller wavelength structures than at greater depths, which may partly be a function of decreasing resolution with depth. A major

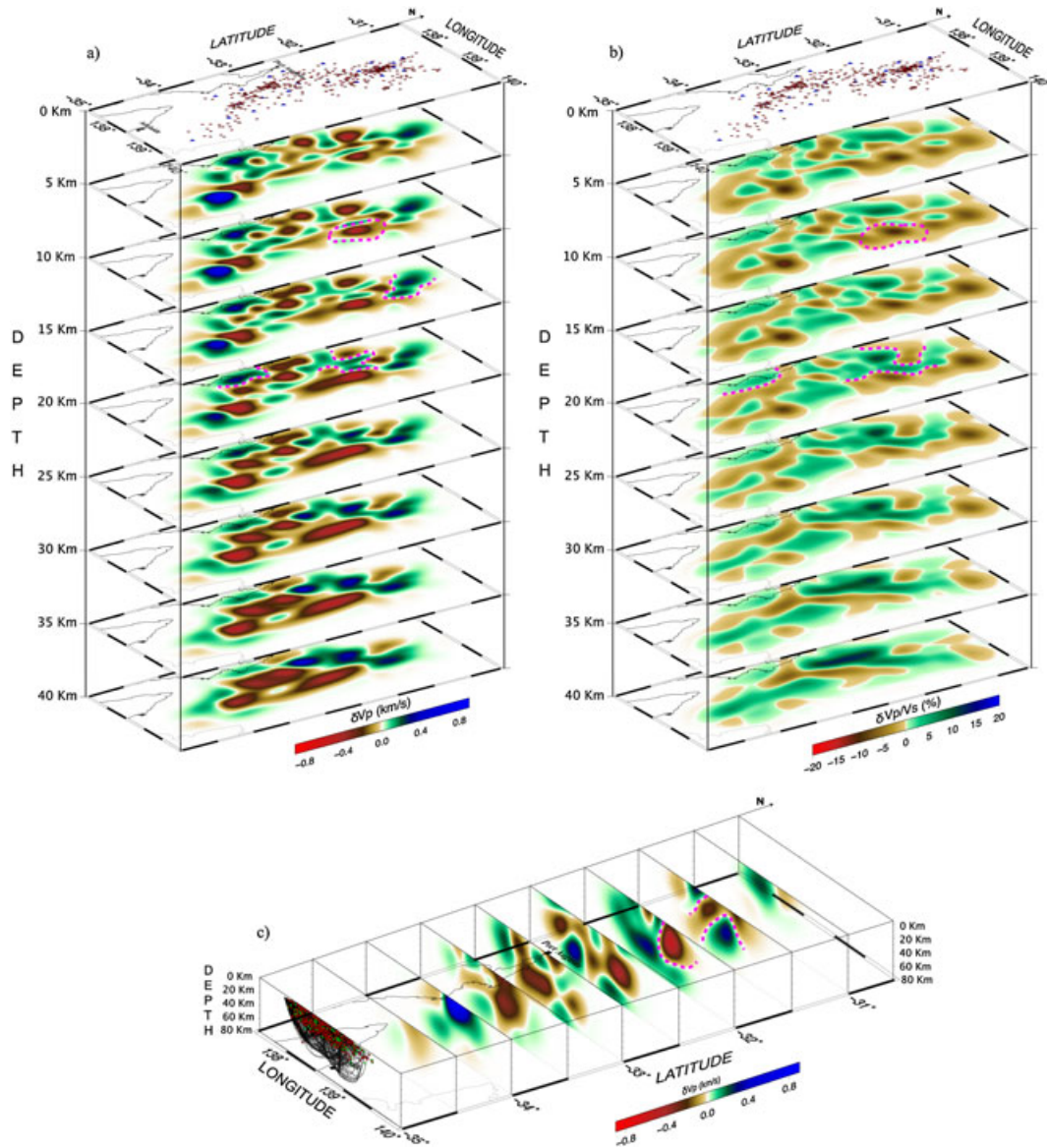


Figure 5. Horizontal and vertical slices through the final tomographic solution models. (a) δv_p , (b) $\delta(v_p/v_s)$ depth slices, and (c) E-W δv_p cross sections. Dotted magenta lines point out the main features discussed in section 3.2.

negative anomaly appears in the central-eastern region of the models at around 10 km depth. This perturbation tends to increase in amplitude and size with increasing depth, persisting through the crust in the v_p model, although it becomes significantly less pronounced in the v_p/v_s model. This is supported by the synthetic test results shown in Figures 3 and 4, which demonstrate that the level of smearing is minimal in this region, except near the margins of the model. At depths between 15 and 20 km, a positive anomaly emerges immediately west of this low anomaly in both models, resulting in a particularly prominent variation from high to low (from about 0.3 to -0.6 km/s) v_p velocity and v_p/v_s ratio (12 to -8%). There is also a positive v_p perturbation immediately north of the large low velocity anomaly at mid-lower crustal depths. It is notable that these two relatively high velocity anomalies extend at least to the base of the crust and coincide with the Central Flinders Ranges on the surface. The

southern part of the v_p solution model is dominated in the west by two adjacent anomalies that exhibit contrasting high and low velocities, following a south-east trend, approximately matching the geological contact between the Gawler Craton and the basin to the east. Elevated wave speeds to a depth of 20 km can be observed, although the resolution tests indicate that this region of the model is poorly resolved (see Figures 3 and 4). The base of the crust shows large low P -velocity anomalies of the order -0.6 km/s and tends to be concentrated in the southern region of the model; conversely, high amplitude v_p perturbations prevail in the northern region of the model (0.9 km/s). This last feature is consistent with the v_p/v_s solution model, although the positive anomalies are more prominent in the southernmost part of the v_p/v_s model. These results show that while several large scale features appear common to both v_p and v_p/v_s , significant differences are also present.

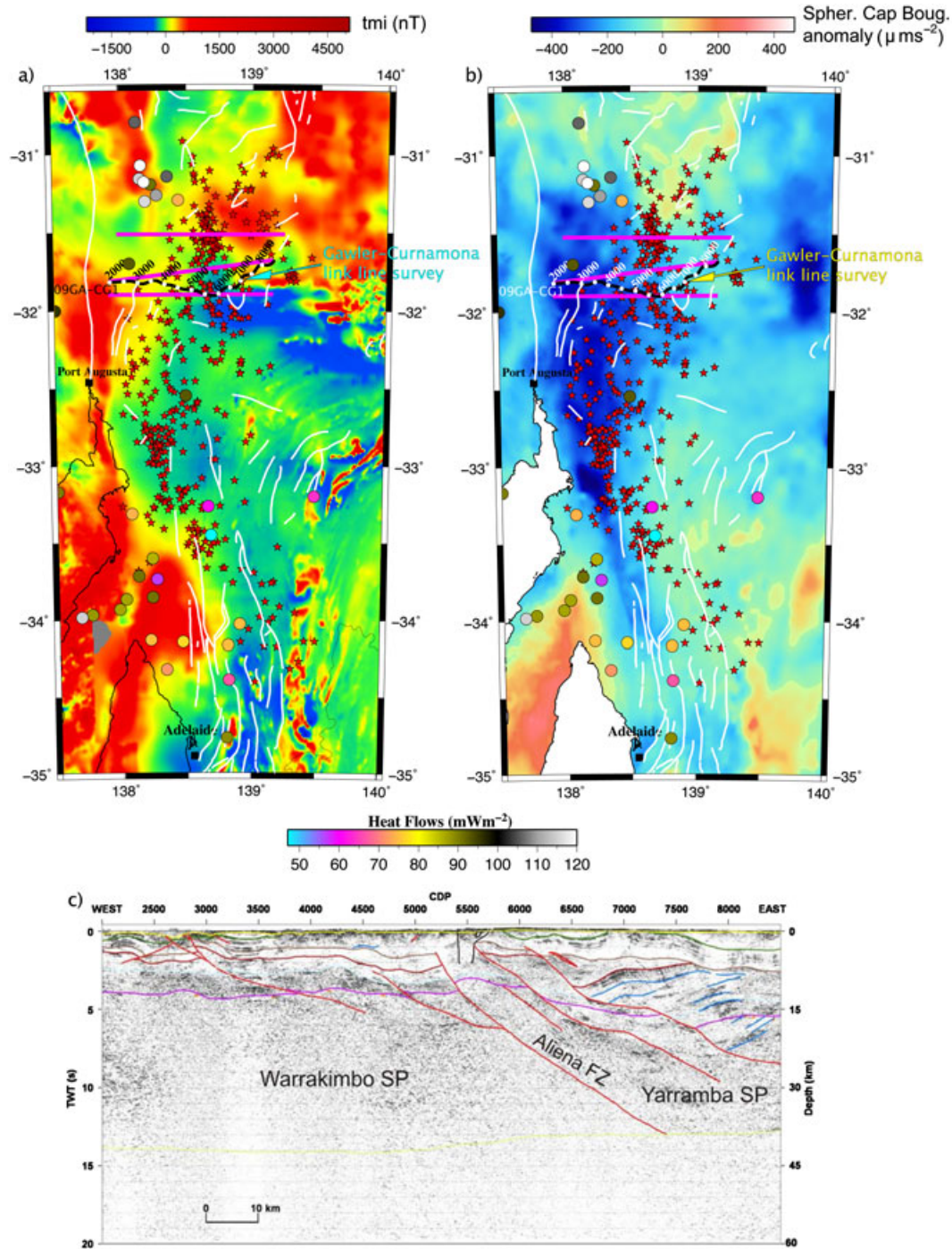


Figure 6. (a) Magnetic and (b) gravity maps of the study area. Heat flow values (colored dots), fault traces (white lines), and the seismic transect 09GA-CG1 with common depth point reflection (CDP) are also plotted, together with the location of three cross sections taken from the tomographic solution models (magenta lines) and shown in Figure 9. Red stars represent the observed seismicity. (c) Interpretation of the deep seismic reflection profile 09GA-CG1 [Preiss, 2010]. Warrakimbo and Yarramba seismic provinces correspond to Gawler Craton and Curnamona Province, respectively. tmi = residual Total Magnetic Intensity; nT = nanoTesla; SP = Seismic Province.

[23] The above features can also be readily identified in the east-west vertical sections (Figure 5c). The large low velocity anomaly is perhaps more clearly evident in the vertical sections (32°S) and appears to be ellipsoidal in shape; however, the lack of data coverage to the east makes it

difficult to evaluate its eastward extent. To the west of this feature, the positive wave speed anomaly seen in Figures 5a and 5b appears to dominate. Further north at 31.5°S, the other positive anomaly has a near circular cross section, while further south (32.5°S), it can no longer be observed.

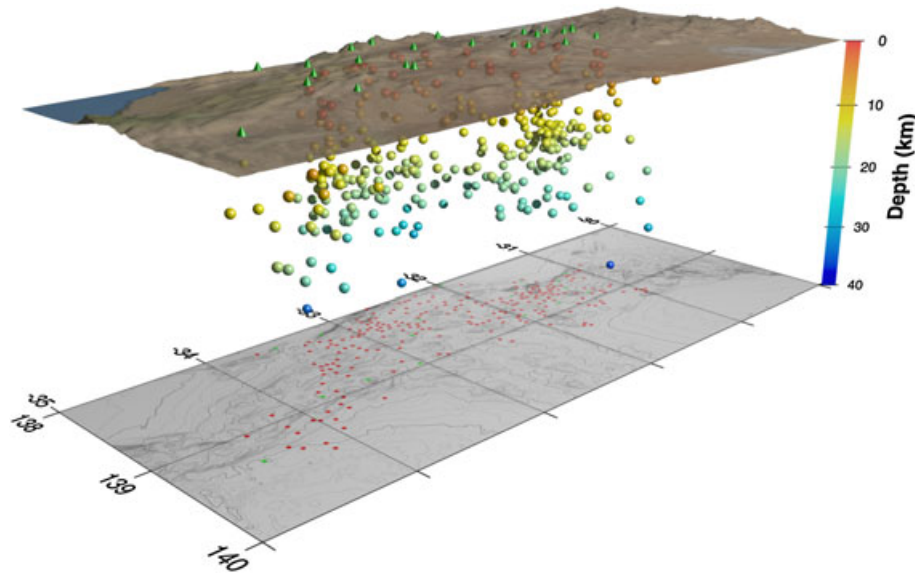


Figure 7. 3-D projection of relocated hypocenters. Green cones denote receivers, and the contour map at the base of the figure highlights topography (grid in degrees).

Of particular relevance to later discussions is the relatively narrow low velocity band that separates two high velocity bodies (31.5°S) in proximity of the Central Flinders Ranges (i.e., between the Gawler Craton and the Curnamona Province).

4. Discussion

4.1. Insights From Potential Field Data and Deep Seismic Reflection Profile

[24] To complement the local earthquake tomography results and provide major constraints on the structures involved in the deformation that characterizes the Flinders Ranges, a full synthesis of the results is made together with insights from public domain geophysical data sets from South Australia. They include magnetic and gravity data [Milligan *et al.*, 2004; Tracey *et al.*, 2007], heat flow measurements [Holford *et al.*, 2011], and a deep seismic reflection profile [Preiss, 2010].

[25] Figures 6a and 6b illustrate maps of the total magnetic intensity and the Spherical Cap Bouguer gravity field that spans the study area. Major positive magnetic and gravity anomalies are located in the north-east area of both maps, coinciding with the Curnamona Province (latitude $\sim 30.5^{\circ}\text{S}/32.0^{\circ}\text{S}$ and longitude $\sim 138.5^{\circ}\text{E}/140.0^{\circ}\text{E}$). The Adelaide Rift Complex is well delineated by negative magnetic and gravity anomalies. On its eastern flank, blocky irregular shaped Ross-Delamerian granites are characterized by both relatively high magnetic and gravity anomalies, which follow an arcuate trend from south of the Curnamona Province to the Fleurieu Arc. The Adelaide Rift Complex itself exhibits a series of subparallel magnetic anomalies roughly aligned with the topography and concentrated in the Nackara Arc. Furthermore, an elongated negative gravity anomaly occurs east of Port Augusta, which extends from around 31.5°S to 34°S and can be related to the root of the Flinders Ranges. To the west, the Adelaide Rift Complex

is confined by the Ross-Delamerian thrust faults, clearly outlined by the marked variation from negative to positive magnetic and gravity anomalies (Figures 6a and 6b). The western sector of the two potential field maps reveals the Gawler Craton, which is defined by positive gravity and magnetic values when not masked by Neoproterozoic to Cambrian platform cover. Further north-east, corresponding with the Central Flinders Ranges (latitude $\sim 31^{\circ}\text{S}/31.5^{\circ}\text{S}$ and longitude $\sim 138.5^{\circ}\text{E}/139.5^{\circ}\text{E}$), both gravity and magnetic maps display a positive anomaly, indicating that in this area the basement is shallower.

[26] In Figures 6a and 6b, heat flow values are also shown. It is well known that South Australia is transversely spanned by a heat flow anomaly, commonly referred to as the South Australian Heat Flow Anomaly (SAHFA) [Neumann *et al.*, 2000]. The SAHFA is defined by a small number of heat flow measurements that reveal high heat flow values on the south-eastern margin of the Gawler Craton (about 80 mW m^{-2}), which are approximately double that of most Proterozoic crust in Australia, and smaller values in the west (about 55 mW m^{-2}). The most prominent heat flow anomaly has been recorded in sediments from wells drilled north of the 09GA-CG1 line, where heat flow values are as high as $90\text{--}120 \text{ mW m}^{-2}$. In contrast, in the Nackara and Fleurieu Arc, heat flow values drop to an average of $\sim 65 \text{ mW m}^{-2}$.

[27] The main aim of the Gawler-Curnamona Link Line seismic survey (09GA-CG1, see Figures 6a and 6b for location and Figure 6c for interpretation) was to elucidate the relationship, at depth, between the Gawler Craton and Curnamona Province, in addition to studying the major crustal penetrating faults and shear zones that may have been potential pathways for mineralizing fluids within rocks of the Adelaide Rift Complex [Preiss, 2010]. According to Preiss [2010], the section shows a general eastward increase in the thickness in the Neoproterozoic to Cambrian cover, whose base is delineated by the magenta horizon (Figure 6c); however, this limit is hard to draw due to the absence of an obvious angular unconformity in the western region of the

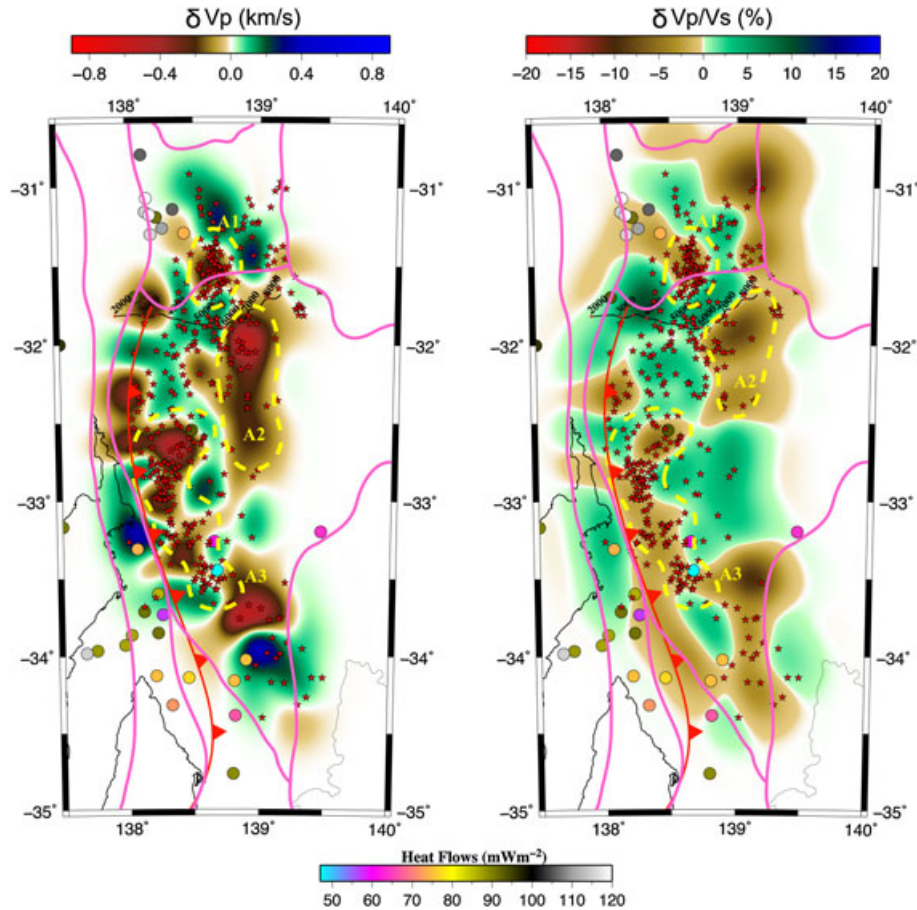


Figure 8. Two horizontal slices at 18 km depth taken from the v_p (left panel) and v_p/v_s (right panel) tomographic solution models. Heat flows values are also plotted along with observed seismicity (red stars). Yellow dotted lines highlight areas characterized by low v_p and v_p/v_s (and elevated seismicity in the case of A1 and A3); A1, anomaly 1; A2, anomaly 2; A3, anomaly 3. Magenta lines show the main geological contacts described in Figure 2a.

seismic section. A major zone of crustal-penetrating, east-dipping thrust faults (Figure 6c, red lines) appear towards the eastern end of the seismic section (termed the Aliena Fault Zone). These faults separate middle to lower crustal regions of different seismic character. From this seismic section, it has been inferred that the Gawler Craton (called Warrakimbo SP in Figure 6c) appears to be on the western part of the Aliena FZ and the Curnamona Province (called Yarramba SP in Figure 6c) on the east. The Moho is poorly imaged (Figure 6c, yellow line), although it appears to be visible at about 39 km depth at the eastern end of the profile and at about 42 km depth in the center.

4.2. Crustal Structure and Seismicity Beneath the Flinders Ranges

[28] In this study, checkerboard tests clearly demonstrate that data from the 24 stations used in this work are capable of a maximum resolution of several 10 s of km in both latitude and longitude (see section 3.1). Any interpretation of the results should account for the fact that the Earth's crust encompasses a great variety of features including faults, layers, intrusions, and so forth, which often cause large variations in wave speed over small distances. In many cases, such complexities cannot be faithfully recovered by

tomographic imaging due to data limitations and assumptions made by the inversion scheme, even though they do influence seismic wave traveltimes.

[29] The main factors affecting v_p/v_s , and the directly related Poisson's ratio, are fluid saturation, partial melt, mineralogy, porosity, and overall volume of cracks and their geometries (see *Christensen* [1996] for a complete review). For most common minerals, increases in pressure have little influence on v_p/v_s , although there is a tendency for very slight increases to occur [*Christensen*, 1996]. Similarly, v_p/v_s appears to have no significant dependency on temperature, although laboratory measurements are limited. Variations in composition produce clearer changes in v_p/v_s ; for instance, a transition from felsic to mafic igneous rocks (e.g., granites to gabbro) produces an increase in v_p/v_s , whereas a transition from mafic to ultramafic igneous rocks (e.g., gabbro to dunite) produces a decrease in v_p/v_s . Serpentinization, which often takes place at plate boundaries, tends to increase v_p/v_s dramatically; for example, hydration of dunite can increase v_p/v_s as much as 2.1. The introduction of partial melt tends to increase v_p/v_s , while the presence of water will decrease it [*Zheng and Lay*, 2006]. However, an increase in pore fluid pressure can lead to an increase in v_p/v_s . Of particular relevance to this study is how v_p/v_s might vary in zones of

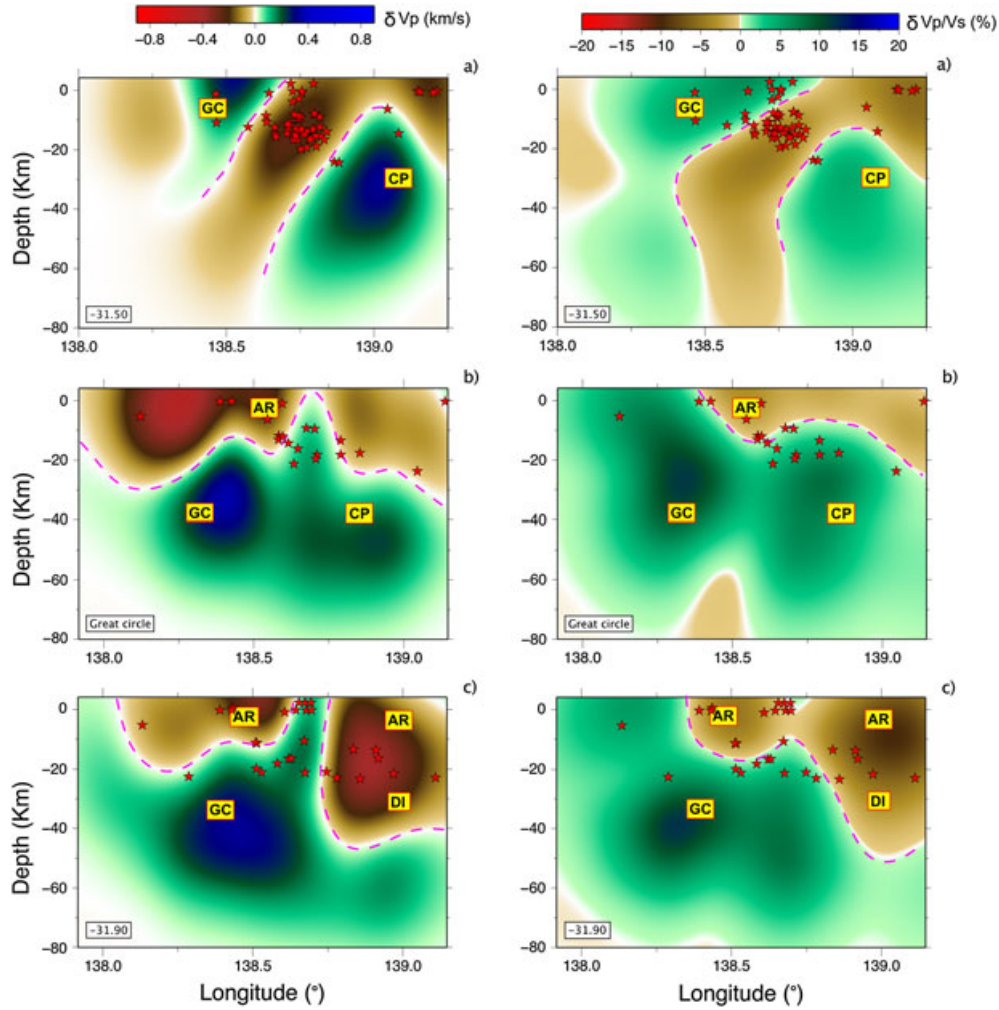


Figure 9. East-west cross sections through the (Figures 9a–9c, left) v_p and (Figures 9a–9c, right) v_p/v_s solution models with several features of interest highlighted. The exact location of the cross sections are shown by magenta lines in Figures 6a and 6b. The great circle line goes straight from the first to the last CDP of the 09GA-CG1 seismic transect. The magenta dashed line outlines the main feature of interest. GC, Gawler Craton; CP, Curnamona Province; AR, Adelaide Rift Complex; DI, Delamerian Intrusion. Red stars represent seismicity. Horizontal:Vertical aspect ratio is 1:1.

deformation. The presence of cracks, fractures, and faults may decrease v_p/v_s [Walsh, 1965], while a recent paper by Lowry and Pérez-Gussinyé [2011], in which they try to estimate continental crust thickness and seismic v_p/v_s ratio in the western United States, shows that an abundance of crustal quartz (the weakest mineral in continental rocks), which has low v_p/v_s , can give rise to a dynamic feedback mechanism, which drives deformation. In this process, ductile strain focuses in weak, quartz-rich crust which in turn triggers advective warming, hydration, and further weakening.

[30] A number of prominent and well-constrained features are revealed in the tomographic model, which allow a number of inferences to be made (Figures 5, 7, 8, and 9). First and foremost, even after source relocation, hypocenters do not appear to show clear lineations relating to any mapped faults (Figure 6; see also Figure 7 for a 3-D distribution of the relocated hypocenters). However, the distribution of seismicity in the Flinders Ranges is coincident with the current axis of the ranges. Of particular significance is the fact that

the seismicity is almost totally concentrated on the eastern side of the Ross-Delamerian thrust faults, whose trend is well resolved by the high magnetic anomaly that marks the eastern extent of the Gawler Craton (Figure 6). The majority of the seismicity occurs at depths greater than about 10 km (Figure 7; the deepest event occurs at approximately 36 km), suggesting that the brittle-ductile transition in the Flinders Ranges might not be unusually shallow as recently suggested by the presence of high heat flows [Sandiford, 2008; Holford et al., 2011]. Furthermore, the available heat flow values do not point to any correlation between regions of elevated seismicity and elevated heat flow (Figures 6 and 8), which implies that heat-induced crustal weakening is unlikely to play a major role in the deformation of the Flinders Ranges. However, further heat flow measurements are required to make a more definitive conclusion.

[31] The new model presented in this work provides solid evidence that a major change in crustal structure occurs in the central Flinders Ranges. Perhaps, the most prominent

structure of all is the presence of anomaly A1 between two high velocity anomalies (latitude $\sim 31.0^\circ\text{S}/32.0^\circ\text{S}$ and longitude $\sim 138.5^\circ\text{E}/139^\circ\text{E}$; Figure 8, but see also Figures 5 and 9). According to the interpretation of the seismic reflection transect 09GA-CG1 [Preiss, 2010], a major system of faults separates the Gawler Craton on the west from the Curnamona Province on the east. Overall, this basic setting is in agreement with our results even though direct comparison between the two classes of seismic image is difficult (reflection seismology reveals discontinuities and faults, while tomography recovers smooth variations in seismic properties). The pronounced low crustal velocity region to the south of the reflection seismic line (anomaly A2 in Figure 8) could correspond to one of the numerous Palaeo-Mesoproterozoic inliers of the Barossa Complex located in southern South Australia [Szpunar *et al.*, 2007]. They are typically poorly exposed and comprise poly-deformed metasediments, orthogneiss, migmatites, pegmatites, and post-deformational intrusives [Preiss *et al.*, 1993]. An alternative explanation is that this is a zone of felsic Delamerian intrusion, perhaps highly fractured and silica rich (v_p/v_s relatively low). The Curnamona Province appears to better match the high velocity anomaly further north (Figure 8), which also lies on a major positive magnetic and gravity anomaly (Figure 6).

[32] In the zone between the Gawler Craton to the west and the Curnamona Province to the east (anomaly A1 in Figure 8), there is by far the most pronounced cluster of seismicity concentrated between 7 and 20 km depth (see also Figure 7), which could be associated with a mechanically induced weakness in the crust arising primarily from the rifting of the Curnamona Province from the Gawler Craton [Belousova *et al.*, 2006; Hand *et al.*, 2008; Wade *et al.*, 2012] (Figure 9a), rather than elevated thermal regimes due to high heat flows for which there is little evidence. Similar to anomaly A2 further south, this region is also characterized by low v_p and low v_p/v_s (Figure 9a), which clearly points to a strong correlation between seismic properties and deformation. The underlying cause for the decreased v_p/v_s may be the presence of preexisting fractures [Anderson *et al.*, 1974] and/or mechanically weak rocks such as quartz-rich schist, although it is difficult to pinpoint an exact cause. The second main cluster of seismicity, which is also associated with a negative band of v_p and v_p/v_s perturbations (anomaly A3 in Figure 8), runs adjacent to the east dipping Ross-Delamerian thrust faults and is likely due to reactivation along this section of the fault zone. The low v_p and v_p/v_s values may be associated with both fracturing, and the presence of an intensely deformed mylonitic shear zone belt with $30\text{--}50^\circ$ east-southeast dipping foliations [Flottmann *et al.*, 1993], possibly associated with the Barossa Complex, whose rocks are typically found at the south-western edge of the Adelaide Rift Complex [Szpunar *et al.*, 2007].

[33] Given that strain localization in the Flinders Ranges likely results from the presence of preexisting structures, it remains to be explained why deformation was initiated relatively recently (less than 10 Ma). As noted earlier, there is evidence of increased coupling between the Australian and Pacific plate during the late Miocene [Sandiford, 2003; Reynolds *et al.*, 2003; Sandiford *et al.*, 2004; C  lerier *et al.*, 2005; Nelson *et al.*, 2006], which may have been sufficient to trigger deformation of mechanically weak rocks adjacent

to much more resistant cratonic material and reinitiate movement along the previously quiescent Ross-Delamerian thrust faults.

[34] South of Port Augusta, the positive wave speed anomaly along the coast indicates the presence of the Gawler Craton, although this trend is not very well defined offshore due to the smearing that can be observed in the checkerboard depth slices in Figures 3 and 4. However, the presence of the craton is confirmed by the positive magnetic anomaly and geological maps (the craton crops out on the northern part of the Yorke and Eyre peninsulas). The other positive anomaly in the south-eastern sector of the depth slice (Figure 8) ($\sim 34.0^\circ\text{S}/139.0^\circ\text{E}$) lies in a zone of poor path coverage, hence its shape and amplitude could change with the addition of more data: however, its presence may be related to the blocky Delamerian granites that extend from the Curnamona Province to the Fleurieu Arc following an arcuate trend.

5. Conclusion

[35] We have developed a hybrid inversion scheme that couples nonlinear earthquake location with iterative nonlinear recovery of v_p structure and linearized recovery of v_p/v_s to constrain the 3-D seismic structure of the Flinders Ranges using P - and S -traveltimes. The significant findings of this study are that zones of low v_p and v_p/v_s , which are consistent with compositional changes and fracturing, correlate strongly with high concentrations of seismicity, but evidence of high heat flow in these areas is lacking. We conclude that pre-existing mechanical weaknesses in the lithosphere, principally due to structure and composition, exert first-order control on the distribution of seismicity in the Flinders Ranges, which is driven by a recent (late Miocene) increase in coupling between the Australian and Pacific plates. Even if heat flows above the seismogenic zones were found to be anomalously high, it appears unlikely that this would be the root cause of strain localization.

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References

- Anderson, D., B. Minster, and D. Cole (1974), The effect of oriented cracks on seismic velocities, *J. Geophys. Res.*, **79**(401), 1–4015.
- Assump  o, M., and V. Sacek (2013), Intra-plate seismicity and flexural stresses in Central Brazil, *Geophys. Res. Lett.*, **40**, 1–5, doi: 10.1002/grl.50142.
- Back  , G., G. Baines, D. Giles, W. Preiss, and A. Alesci (2010), Basin geometry and salt diapirs in the Flinders Ranges, South Australia: Insights gained from geologically-constrained modelling of potential field data, *Mar. Pet. Geol.*, **27**(3), 650–665.
- Belousova, E., W. Preiss, M. Schwarz, and W. Griffin (2006), Tectonic affinities of the Houghton Inlier, South Australia: U–Pb and Hf-isotope data from zircons in modern stream sediments, *Aust. J. Earth Sci.*, **53**(6), 971–989.
- Bodin, P., and S. Horton (2004), Source parameters and tectonic implications of aftershocks of the Mw 7.6 Bhuj earthquake of 26 January 2001, *Bull. Seismol. Soc. Am.*, **94**(3), 818–827.
- Bratt, S., and T. Bache (1988), Locating events with a sparse network of regional arrays, *Bull. Seismol. Soc. Am.*, **78**(2), 780–798.
- Braun, J., D. Burbidge, F. Gesto, M. Sandiford, A. Gleadow, B. Kohn, and P. Cummins (2009), Constraints on the current rate of deformation and surface uplift of the Australian continent from a new seismic database and low-T thermochronological data, *Aust. J. Earth Sci.*, **56**(2), 99–110.
- C  lerier, J., M. Sandiford, D. Hansen, and M. Quigley (2005), Modes of active intraplate deformation, Flinders Ranges, Australia, *Tectonics*, **24**(6), TC6006, doi:10.1029/2004TC001679.

- Christensen, N. (1996), Poisson's ratio and crustal seismology, *J. Geophys. Res.*, **101**(B2), 3139–3156.
- Clark, D., and M. Leonard (2003), Principal stress orientations from multiple focal-plane solutions: New insight into the Australian intraplate stress field, *Geol. S. Am. S.*, **22**, 91–106.
- Clarke, G., and R. Powell (1989), Basement-cover interaction in the Adelaide Foldbelt, South Australia: The development of an arcuate foldbelt, *Tectonophysics*, **158**(1–4), 209–210.
- Conor, C. (2000), Definition of major sedimentary and igneous units of the Olary Domain, Curnamona Province, *MESA Journal*, **19**, 51–56.
- Conor, C., and W. Preiss (2008), Understanding the 1720–1640Ma Palaeoproterozoic Willyama Supergroup, Curnamona Province, Southeastern Australia: Implications for tectonics, basin evolution and ore genesis, *Precambrian Res.*, **166**(1), 297–317.
- Cummins, P., C. Collins, A. Bullock, and D. Love (2004), Monitoring of earthquakes in the Flinders Ranges, South Australia, using a temporary seismometer deployment, in *Australian Earthquake Engineering in the New Millennium*, edited by K. McCue, M. Griffith, and B. Butler, p. 22, Proceeding of the Australian Earthquake Engineering Society Conference, Mt. Gambier.
- Daly, S., G. Fanning, and M. Fairclough (1998), Tectonic evolution and exploration potential of the Gawler Craton, South Australia, *AGSO J. Aust. Geol. Geophys.*, **17**, 145–168.
- De Kool, M., N. Rawlinson, and M. Sambridge (2006), A practical grid-based method for tracking multiple refraction and reflection phases in three-dimensional heterogeneous media, *Geophys. J. Int.*, **167**(1), 253–270.
- de Vries, S., L. Pryer, and N. Fry (2008), Evolution of Neoarchean and Proterozoic basins of Australia, *Precambrian Res.*, **166**(1–4), 39–53.
- Direen, N., D. Brock, and M. Hand (2005), Geophysical testing of balanced cross-sections of fold-thrust belts with potential field data: An example from the Fleurieu Arc of the Delamerian Orogen, South Australia, *J. Struct. Geol.*, **27**(6), 964–984.
- Dyksterhuis, S., and R. Müller (2008), Cause and evolution of intraplate orogeny in Australia, *Geology*, **36**(6), 495.
- Dzierma, Y., W. Rabbel, M. Thorwart, I. Koulakov, H. Wehrmann, K. Hoernle, and D. Comte (2012), Seismic velocity structure of the slab and continental plate in the region of the 1960 Valdivia (Chile) slip maximum—Insights into fluid release and plate coupling, *Earth Planet. Sci. Lett.*, **331**, 164–176.
- Eberhart-Phillips, D. (1990), Three-dimensional P and S velocity structure in the Coalinga region, California, *J. Geophys. Res.*, **95**(B10), 15343–15363.
- Eberhart-Phillips, D., and M. Reyners (2012), Imaging the Hikurangi Plate interface region, with improved local-earthquake tomography, *Geophys. J. Int.*, **190**(2), 12211242, doi:10.1111/j.1365-246X.2012.05553.X.
- Evans, J., D. Eberhart-Phillips, and C. Thurber (1994), User's manual for SIMULPS 12 for imaging Vp and Vp/Vs: A derivative of the Thurber tomographic in version SIMUL3 for local earthquakes and explosion, *U.S. Geol. Surv. Open File Rep. OFR 94-431*, pp. 101.
- Fanning, C., R. Flint, A. Parker, K. Ludwig, and A. Blissett (1988), Refined Proterozoic evolution of the Gawler craton, South Australia, through U-Pb zircon geochronology, *Precambrian Res.*, **40**, 363–386.
- Fanning, C., A. Reid, and G. Teale (2007), A geochronological framework for the Gawler craton, South Australia, *Bull. Geol. Surv. S. Aust.*, **55**, 80.
- Flottmann, T., and R. Oliver (1994), Review of Precambrian-Palaeozoic relationships at the craton margins of southeastern Australia and adjacent Antarctica, *Precambrian Res.*, **69**(1–4), 293–306.
- Flottmann, T., G. Gibson, and G. Kleinschmidt (1993), Structural continuity of the Ross and Delamerian orogens of Antarctica and Australia along the margin of the paleo-Pacific, *Geology*, **21**(4), 319–322.
- Flottmann, T., P. James, J. Rogers, and T. Johnson (1994), Early Palaeozoic foreland thrusting and basin reactivation at the Palaeo-Pacific margin of the southeastern Australian Precambrian Craton: A reappraisal of the structural evolution of the southern Adelaide Fold-thrust Belt, *Tectonophysics*, **234**(1–2), 95–116.
- Foden, J., M. Elburg, S. Turner, M. Sandiford, J. O'Callaghan, and S. Mitchell (2002), Granite production in the Delamerian orogen, South Australia, *J. Geol. Soc.*, **159**(5), 557–575.
- Foden, J., M. Elburg, J. Dougherty-Page, and A. Burt (2006), The timing and duration of the Delamerian Orogeny: Correlation with the Ross Orogen and implications for Gondwana assembly, *J. Geol.*, **114**(2), 189–210.
- Glahn, A., and M. Granet (1993), Southern Rhine Graben: Small-wavelength tomographic study and implications for the dynamic evolution of the graben, *Geophys. J. Int.*, **113**, 399418.
- Greenhalgh, S., D. Love, K. Malpas, and R. McDougall (1994), South Australian earthquakes, 1980–92, *Aust. J. Earth Sci.*, **41**(5), 483–495.
- Hand, M., A. Reid, and L. Jagodzinski (2007), Tectonic framework and evolution of the Gawler craton, southern Australia, *Econ. Geol.*, **102**(8), 1377–1395.
- Hand, M., A. Reid, M. Szpunar, N. Direen, B. Wade, J. Payne, and K. Barovich (2008), Crustal architecture during the early Mesoproterozoic Hiltaba-related mineralisation event: Are the Gawler Range Volcanics a foreland basin fill? *MESA Journal*, **51**, 19–24.
- Hearn, T. M., and R. W. Clayton (1986), Lateral velocity variations in southern California. I. Results for the upper crust for Pg waves, *Bull. Seismol. Soc. Am.*, **76**, 495–509.
- Hillis, R., and S. Reynolds (2000), The Australian stress map, *J. Geol. Soc.*, **157**(5), 915–921.
- Hillis, R., and S. Reynolds (2003), In situ stress field of Australia, *Geol. S. Am. S.*, **372**, 49–58.
- Hillis, R. R., M. Sandiford, S. D. Reynolds, and M. C. Quigley (2008), Present-day stresses, seismicity and Neogene-to-Recent tectonics of Australia's 'passive' margins: Intraplate deformation controlled by plate boundary forces, *Geol. Soc. Spec. Publ.*, **306**(1), 71–90, doi:10.1144/SP306.3.
- Holford, S. P., R. R. Hillis, M. Hand, and M. Sandiford (2011), Thermal weakening localizes intraplate deformation along the southern Australian continental margin, *Earth Planet. Sci. Lett.*, **305**(12), 207–214, doi:10.1016/j.epsl.2011.02.056.
- Jones, L., P. Lyons, B. Goleby, A. Shearer, M. Schwarz, R. Skirrow, R. Korsch, J. Totterdell, W. Preiss, and N. Direen (2004), Gawler craton uncovered, *ASEG Ext. Abstr.*, **2004**(1), 1–4.
- Kennett, B., M. Sambridge, and P. Williamson (1988), Subspace methods for large inverse problems with multiple parameter classes, *Geophys. J.*, **94**(2), 237–247.
- Kennett, B., E. Engdahl, and R. Buland (1995), Constraints on seismic velocities in the Earth from traveltimes, *Geophys. J. Int.*, **122**(1), 108–124.
- Kissling, E., W. Ellsworth, D. Eberhart-Phillips, and U. Kradolfer (1994), Initial reference models in local earthquake tomography, *J. Geophys. Res.*, **99**, 19,635–19,646.
- Koulakov, I. (2009), Lotos code for local earthquake tomographic inversion: Benchmarks for testing tomographic algorithms, *Bull. Seismol. Soc. Am.*, **99**(1), 194–214.
- Liu, L., and M. Zoback (1997), Lithospheric strength and intraplate seismicity in the New Madrid seismic zone, *Tectonics*, **16**(4), 585–595.
- Lowry, A., and M. Pérez-Gussinyé (2011), The role of crustal quartz in controlling Cordilleran deformation, *Nature*, **471**(7338), 353–357.
- Marshak, S., and T. Flottmann (1996), Structure and origin of the Fleurieu and Nackara Arcs in the Adelaide fold-thrust belt, South Australia: Salient and recess development in the Delamerian Orogen, *J. Struct. Geol.*, **18**(7), 891–908.
- McKenna, J., S. Stein, and C. Stein (2007), Is the New Madrid seismic zone hotter and weaker than its surroundings? *Geol. S. Am. S.*, **425**, 167–175.
- Michelini, A. (1993), Testing the reliability of Vp/Vs anomalies in travel-time tomography, *Geophys. J. Int.*, **114**(2), 405–410.
- Milligan, P., R. Franklin, and D. Ravat (2004), A new generation Magnetic Anomaly Grid Database of Australia (MAGDA), *Preview*, **113**, 25–29.
- Negishi, H., J. Mori, T. Sato, R. Singh, S. Kumar, and N. Hirata (2002), Size and orientation of the fault plane for the 2001 Gujarat, India earthquake (Mw 7.7) from aftershock observations—A high stress drop event, *Geophys. Res. Lett.*, **29**(20), 10–1.
- Nelson, E., R. Hillis, M. Sandiford, S. Reynolds, P. Lyon, J. Meyer, S. Mildren, and C. Rogers (2006), Present-day state-of-stress of southeast Australia, *APPA Journal*, **46**, 283–305.
- Neumann, N., M. Sandiford, and J. Foden (2000), Regional geochemistry and continental heat flow: Implications for the origin of the South Australian heat flow anomaly, *Earth Planet. Sci. Lett.*, **183**(1), 107–120.
- Page, R., C. Conor, B. Stevens, G. Gibson, W. Preiss, and P. Southgate (2005), Correlation of Olary and Broken Hill Domains, Curnamona Province: Possible relationship to Mount Isa and other North Australian Pb-Zn-Ag-bearing successions, *Econ. Geol.*, **100**(4), 663–676.
- Paul, E., T. Flottmann, and M. Sandiford (1999), Structural geometry and controls on basement-involved deformation in the northern Flinders Ranges, Adelaide Fold Belt, South Australia, *Aust. J. Earth Sci.*, **46**(3), 343–354.
- Payne, J., M. Hand, K. Barovich, A. Reid, and D. Evans (2009), Correlations and reconstruction models for the 2500–1500 Ma evolution of the Mawson Continent, *J. Geol. Soc. London. Spec. Publ.*, **323**(1), 319–355.
- Preiss, W. (1987), *The Adelaide Geosyncline: Late Proterozoic Stratigraphy, Sedimentation, Palaeontology and Tectonics*, *Geol. Surv. South Aust. Bull.*, **53**, 438.
- Preiss, W., A. Belperio, W. Cowley, and L. Rankin (1993), Neoproterozoic, in *The Geology of South Australia: Geological Survey of South Australia Bulletin*, **53**, edited by J. F. Drexel et al., 171–203.

- Preiss, W. V. (2000), The Adelaide Geosyncline of South Australia and its significance in Neoproterozoic continental reconstruction, *Precambrian Res.*, 100(1), 21–63.
- Preiss, W. V. (2010), Geology of Neoproterozoic to Cambrian Adelaide Geosyncline and Cambrian Delamerian Orogeny, in *South Australian Seismic and MT Workshop, 2010*, edited by R. J. Korsch, and N. Kositsin, pp. 34–42, Geoscience Australia, record 2010/10.
- Quigley, M., M. Cupper, and M. Sandiford (2006), Quaternary faults of south-central Australia: Palaeoseismicity, slip rates and origin, *Aust. J. Earth Sci.*, 53(2), 285–301.
- Quigley, M., M. Sandiford, K. Fifield, and A. Alimanovic (2007a), Bedrock erosion and relief production in the northern Flinders Ranges, Australia, *Earth Surf. Processes Landforms*, 32(6), 929944, doi:10.1002/esp.1459.
- Quigley, M., M. Sandiford, L. K. Fifield, and A. Alimanovic (2007b), Landscape responses to intraplate tectonism: Quantitative constraints from ¹⁰Be nuclide abundances, *Earth Planet. Sci. Lett.*, 261(1–2), 120–133.
- Quigley, M., D. Clark, and M. Sandiford (2010), Tectonic geomorphology of Australia, *J. Geol. Soc. London. Spec. Publ.*, 346(1), 243–265.
- Rawlinson, N., and M. Sambridge (2003), Seismic traveltimes tomography of the crust and lithosphere, *Adv. Geophys.*, 46, 81–198.
- Rawlinson, N., and M. Sambridge (2004a), Multiple reflection and transmission phases in complex layered media using a multistage fast marching method, *Geophysics*, 69(5), 1338–1350.
- Rawlinson, N., and M. Sambridge (2004b), Wave front evolution in strongly heterogeneous layered media using the fast marching method, *Geophys. J. Int.*, 156(3), 631–647.
- Rawlinson, N., A. Reading, and B. Kennett (2006), Lithospheric structure of Tasmania from a novel form of teleseismic tomography, *J. Geophys. Res.*, 111(B2), B02301, doi:10.1029/2005JB003803.
- Rawlinson, N., H. Tkalcic, and A. M. Reading (2010), Structure of the Tasmanian lithosphere from 3D seismic tomography, *Aust. J. Earth Sci.*, 57, 381–394.
- Reynolds, S., D. Coblenz, and R. Hillis (2003), Influences of plate-boundary forces on the regional intraplate stress field of continental Australia, *Geol. S. Am. S.*, 59–70.
- Robertson, R. S., A. F. Crooks, P. W. Hill, and M. J. Sheard (1998), Review of the Proterozoic geology and mineral potential of the Curnamona Province in South Australia, *AGSO J. Aust. Geol. Geophys.*, 17, 169–182.
- Rutherford, L., M. Hand, and K. Barovich (2007), Timing of Proterozoic metamorphism in the southern Curnamona Province: implications for tectonic models and continental reconstructions? *Aust. J. Earth Sci.*, 54(1), 65–81.
- Rutland, R., A. Parker, G. Pitt, W. Preiss, and B. Murrell (1981), The Precambrian of South Australia, in *Precambrian of the Southern Hemisphere, Developments in Precambrian Geology Series, 2*, edited by D. R. Hunter, 309–360, Elsevier, Amsterdam, The Netherlands.
- Sandiford, M. (2003), Neotectonics of southeastern Australia: linking the Quaternary faulting record with seismicity and in situ stress, *Geol. Soc. Australia Spec. Publ.* 22, and *Geol. Soc. America Spec. Pap.* 372, 107–119.
- Sandiford, M. (2008), Enhanced intraplate seismicity along continental margins: Some causes and consequences, *Tectonophysics*, 457(3), 197.
- Sandiford, M., M. Wallace, and D. Coblenz (2004), Origin of the in situ stress field in south-eastern Australia, *Basin Res.*, 16(3), 325–338.
- Sethian, J. (1996), A fast marching level set method for monotonically advancing fronts, *Proc. Nat. Acad. Sci.*, 93(4), 1591.
- Sethian, J., and A. Popovici (1999), 3-D traveltimes computation using the fast marching method, *Geophysics*, 64(2), 516–523.
- Shackelford, P., and D. Sutton (1981), A first interpretation of crustal structure in the Adelaide Geosyncline in South Australia using quarry blasts, *J. Geol. Soc. Aust.*, 28(3–4), 491–500.
- Sykes, L. R. (1978), Intraplate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism, and other tectonism post-dating continental fragmentation, *Rev. Geophys.*, 16(4), 621–688, doi:10.1029/RG016i004p00621.
- Szpunar, M., B. Wade, M. Hand, and K. Barovich (2007), Timing of proterozoic high-grade metamorphism in the barossa complex, southern South Australia: Exploring the extent of the 1590 ma event, *MESA Journal*, 47, 21–27.
- Thomas, J., N. Direen, and M. Hand (2008), Blind orogen: Integrated appraisal of multiple episodes of Mesoproterozoic deformation and reworking in the Fowler Domain, western Gawler Craton, Australia, *Precambrian Res.*, 166(1–4), 263–282.
- Thurber, C. (1983), Earthquake locations and three-dimensional crustal structure in the Coyote Lake area, central California, *J. Geophys. Res.*, 88(B10), 8226–8236.
- Thurber, C. (1993), Local earthquake tomography: Velocities and vp/vs-theory, in *Seismic Tomography: Theory and Practice*, edited by H. Iyer, and K. Hirahara, pp. 563–583, Chapman and Hall, London.
- Tracey, R., M. Bacchin, and P. Wynne (2007), AAGD07: A new absolute gravity datum for Australian gravity and new standards for the Australian National Gravity Database, *ASEG Ext. Abstr.*, 2007(1), 1–3.
- Tryggvason, A., and N. Linde (2006), *Geophys. Res. Lett.*, 33(7), L07303.
- Tuttle, M., E. Schweig, J. Sims, R. Lafferty, L. Wolf, and M. Haynes (2002), The earthquake potential of the New Madrid seismic zone, *Bull. Seismol. Soc. Am.*, 92(6), 2080–2089.
- Von der Borch, C. (1980), Evolution of late proterozoic to early paleozoic Adelaide foldbelt, Australia: Comparisons with post-permian rifts and passive margins, *Tectonophysics*, 70(1–2), 115–134.
- Wade, C., A. Reid, M. Wingate, E. Jagodzinski, and K. Barovich (2012), Geochemistry and geochronology of the c. 1585 ma Benagerie Volcanic Suite, southern Australia: Relationship to the Gawler Range Volcanics and implications for the petrogenesis of a Mesoproterozoic silicic large igneous province, *Precambrian Res.*, 51, 206–207.
- Walck, M. (1988), Three-dimensional Vp/Vs variations for the Coso region, California, *J. Geophys. Res.*, 93(B3), 2047–2052.
- Walsh, J. (1965), The effect of cracks in rocks on Poisson's ratio, *J. Geophys. Res.*, 70(20), 5249–5257.
- Wang, C., X. Zhang, Z. Lin, Q. Wu, and Y. Zhang (1997), Crustal structure beneath the Xingtai earthquake area in North China and its tectonic implications, *Tectonophysics*, 274(4), 307–319.
- Wang, Y. (2001), Heat flow pattern and lateral variations of lithosphere strength in China mainland: Constraints on active deformation, *Phys. Earth Planet. Inter.*, 126(3), 121–146.
- Wang, Y., and S. Cheng (2011), Lithospheric thermal structure and rheology of the eastern China, *J. Asian Earth Sci.*, 46, 51–63.
- Webb, A., B. Thomson, A. Blissett, S. Daly, R. Flint, and A. Parker (1986), Geochronology of the Gawler Craton, South Australia, *Aust. J. Earth Sci.*, 33(2), 119–143.
- Wingate, M., I. Campbell, W. Compston, and G. Gibson (1998), Ion microprobe U–Pb ages for Neoproterozoic basaltic magmatism in south-central Australia and implications for the breakup of Rodinia, *Precambrian Res.*, 87(3), 135–159.
- Yassaghi, A., P. James, and T. Flottmann (2000), Geometric and kinematic evolution of asymmetric ductile shear zones in thrust sheets, southern Adelaide Fold–Thrust Belt, South Australia, *J. Struct. Geol.*, 22(7), 889–912.
- Zheng, Y., and T. Lay (2006), Low Vp/Vs ratios in the crust and upper mantle beneath the sea of Okhotsk inferred from teleseismic pMP, sMP, and sMS underside reflections from the Moho, *J. Geophys. Res.*, 111(B1), B01305.