Crustal Structure of the Eastern Arunta Region, Central Australia, From Magnetotelluric, Seismic, and Magnetic Data

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Abstract We present a 3-D inversion of magnetotelluric data acquired along a 340-km transect in Central Australia. The results are interpreted with a coincident deep crustal seismic reflection survey and magnetic inversion. The profile crosses three Paleoproterozoic to Mesoproterozoic basement provinces, the Davenport, Aileron, and Warumpi Provinces, which are overlain by remnants of the Neoproterozoic to Cambrian Centralian Superbasin, the Georgina and Amadeus Basins, and the Irindina Province. The inversion shows conductors near the base of the Irindina Province that connect to moderately conductive pathways from 50-km depth and to off-profile conductors at shallower depths. The shallow conductors may reflect anisotropic resistivity and are interpreted as sulfide minerals in fractures and faults near the base of the Irindina Province. Beneath the Amadeus Basin, and in the Aileron Province, there are two conductors associated strong magnetic susceptibilities from inversions, suggesting they are caused by magnetic, conductive minerals such as magnetite or pyrrhotite. Beneath the Davenport Province, the inversion images a conductive layer from ∼15- to 40-km depth that is associated with elevated magnetic susceptibility and high seismic reflectivity. The margins between the different basement provinces from previous seismic interpretations are evident in the resistivity model. The positioning and geometry of the southern margin of the crustal conductor beneath the Davenport Province supports the positioning of the south dipping Atuckera Fault as interpreted on the seismic data. Likewise, the interpreted north dipping margin between the Warumpi and Aileron Province is imaged as a transition from resistive to conductive crust, with a steeply north dipping geometry.

1. Introduction

The Arunta region in Central Australia provides a long record of intraplate deformation and, potentially, plate margin processes, back to the Palaeoproterozoic (Collins & Shaw, 1995; Betts et al., 2002; Betts & Giles, 2006; Giles et al., 2004; Scrimgeour & Raith, 2001b). Basement provinces include the 1,690- to 1,640-Ma Warumpi Province, the 1,860- to ∼1,740-Ma Aileron Province, the 1,850- to ∼1,780-Ma Davenport Province, and the 1,865- to 1,640-Ma Casey Inlier (Ahmad & Munson, 2013; Close et al., 2007; Claué-Longet et al., 2008; Scrimgeour et al., 2005). These provinces are separated by crustal-scale faults and show marked changes in seismic reflectivity across them, suggesting that they were once separate crustal blocks (Korsch et al., 2011). The Neoproterozoic to Cambrian Centralian Superbasin was deposited on these basement provinces and then uplifted and eroded into its present-day remnants, the Georgina Basin, Irindina Province, and Amadeus Basin, during the Paleozoic (Haines et al., 2001; Scrimgeour & Close, 1999; Walter et al., 1995).

The 09GA-GA1 deep magnetotelluric (MT) and seismic reflection survey was collected in 2009 under the Geoscience Australia’s Onshore Energy Security Program in collaboration with the Northern Territory Geological Survey. It extends approximately north-south across these basement provinces and overlying basins (Figures 1 and 2). The seismic reflection data were interpreted by Korsch et al. (2011) and revised by Korsch and Doublier (2016; Figure 2). In addition to defining the geometry of the bounding structures to the basement provinces, the seismic reflection data have helped to delineate the internal structure of some of these regions and the overlying basins (Korsch et al., 2011). The 09GA-GA1 MT and seismic survey follows a similar but extended traverse to the 150-km long-period MT survey, primarily across the Aileron Province, which were inverted in 2-D (Selway, 2006).
Figure 1. Location map showing geology after Ahmad et al. (2006) overlain by phase tensors at a period of 0.26 s (left) and total magnetic intensity overlain by phase tensors and induction vectors (Parkinson convention) at a period of 100 s (right). Phase tensors are colored by the maximum phase, $\phi_{max}$. Faults, including the Delny Shear Zone, shown as black lines. Station labels with prefix GB are broadband stations; those with prefix GL are long-period stations. Inset map on the bottom right shows towns as red stars and roads as yellow lines.

MT data (Duan & Milligan, 2010) follow a 340-km traverse over the Davenport and Aileron Provinces, the inferred Warumpi Province, the Casey Inlier, and overlying basins and include both broadband and long-period data. They therefore provide an opportunity to add to our understanding of the structural and geodynamic evolution of the region. The broadband data were originally processed to a period range of 0.003 to 100 s (Nakamura et al., 2011); however, the recording time of >30 hr at each site should allow processing to longer periods, providing more information on the deeper structure. A preliminary 2-D inversion of the MT data was presented and compared with the seismic reflection interpretation (Korsch et al., 2011). However, no details of the inversion process were published, and no further models from this data set have been published since.

Inverting MT profile data using a three dimensional (3-D) inversion code is a practice that has become more common, as 3-D codes have become more efficient and widely available (e.g., Milligan, 2013; Meqbel et al., 2016; Robertson et al., 2017; Patro & Egbert, 2011). The key input for inversion, the impedance tensor, is
the complex ratio of the electric to magnetic fields in the frequency domain and is related to the resistivity of the subsurface. Three-dimensional inversion allows all elements of the impedance tensor to be used. In two-dimensional (2-D) inversion, the diagonal elements are ignored following rotation of the data to geoelectric strike (Chave & Jones, 2012). This is a good approximation if the data are 2-D and a consistent strike can be defined. However, this is rarely the case. Often, the strike varies with both distance and frequency, and the diagonal components of the impedance tensor are significant (e.g., Patro & Egbert, 2011). Furthermore, 2-D inversion of 3-D data located along a profile can result in artifacts due to off-profile features being forced onto the profile (e.g., Meebel et al., 2016).

We present the broadband 09GA-GA1 MT data reprocessed to an extended bandwidth of 0.003 to 1,000 s, and inversion of these data with the long-period data using the ModEM 3-D inversion code (Egbert & Kelbert, 2012; Kelbert et al., 2014). The results are evaluated together with gravity data, magnetic inversions, and the seismic reflection data, providing additional constraints on deep structures and the evolution of this region.

2. Geological and Geophysical Characteristics of the Arunta Region
The Palaeoproterozoic to Mesoproterozoic basement provinces crossed by the 09GA-GA1 MT and seismic line are (from north to south) the Davenport Province, the Aileron Province, the Casey Inlier, and the Warumpi Province (Scrimgeour & Close, 2011). These provinces are overlain by the Neoproterozoic and Palaeozoic Georgina Basin, Irindina Province, and Amadeus Basin (Scrimgeour & Close, 2011).
2.1. Basement Provinces
The Davenport Province and Aileron Province contain formations that are time equivalents and in some cases direct correlatives, suggesting that they were joined during at least part of the Archean to Paleoproterozoic (Blake, 1978; Payne et al., 2009; Scrimgeour & Close, 2011). However, significant tectonism has affected the Aileron Province subsequently but not the Davenport Province, possibly due to its proximity to the southern margin of the North Australian Craton (NAC; Ahmad & Scrimgeour, 2013; Scrimgeour, 2006). The Warumpi Province contains more juvenile crust (1,690–1,600 Ma; Ahmad & Scrimgeour, 2013).

2.1.1. Davenport Province
The Davenport Province comprises 1,850- to ∼1,780-Ma metasedimentary and metavolcanic rocks, with minor intrusive rocks, overlying ∼1,865- to 1,860-Ma basement, which is exposed further north in the Warra-munga Province (Blake & Page, 1988; Claoué-Long et al., 2008; Donnellan, 2013). The rocks of the Davenport Province are weakly reflective (Korsch et al., 2011).

The crustal layer beneath the Davenport Province was interpreted by Korsch et al. (2011) to be located at about 3.0 to 4.7 s two-way traveltime (TWT) or about 9- to 14-km depth. Korsch et al. (2011) named this the Ooratippra Seismic Province, and it is associated with higher seismic reflectivity than the overlying rocks. Magnetic inversions suggest this layer is associated with higher magnetic susceptibility values (>0.03 SI) than the overlying Davenport Province (Chopping et al., 2011).

2.1.2. Aileron Province
The Aileron Province contains metasedimentary successions, almost all of which were deposited between 1,860 and 1,740 Ma (Ahmad & Munson, 2013; Maidment et al., 2005; Scrimgeour, 2003). The Aileron Province is exposed immediately north and south of, and extends beneath, the Irindina Province (Figure 2, Korsch et al., 2011).

The Aileron Province shows more variation in geophysical characteristics than the adjacent Davenport Province. Although the TMI image shows some high-frequency ridges where it is exposed, the inversions of Chopping et al. (2011) reveal mostly low magnetic susceptibility within the Aileron Province. The exceptions are beneath the Amadeus Basin and north of the Delny Shear Zone (Figure 2). Beneath the Amadeus Basin, magnetic inversions reveal high magnetic susceptibility values of ∼0.09 SI units. North of the Delny shear zone, there are moderate magnetic susceptibility values of ∼0.04 to 0.08 SI units.

2.1.3. Casey Inlier
The Casey Inlier comprises 1,865- to 1,640-Ma basement, which forms an inlier in the northeastern Amadeus Basin, and has been divided into three domains with different protoliths and tectonic histories (Close et al., 2007; Scrimgeour & Close, 2011). The Casey Inlier was interpreted by Korsch and Doublier (2016) to extend to ∼10-km depth on the southern margin of the Aileron Province (Figure 2).

2.1.4. Warumpi Province
The Warumpi Province consists of 1,690- to 1,600-Ma greenschist to granulite facies metasedimentary and metaigneous rocks (Ahmad & Scrimgeour, 2013; Scrimgeour et al., 2005). It is associated with moderate reflectivity (Figures 2a and 2b) and low magnetic susceptibility (Chopping et al., 2011). The Warumpi Province is not exposed in this region, however Korsch and Doublier (2016) interpreted it to underlie the Amadeus Basin in the southern end of the line (Figure 2c).

2.2. Neoproterozoic and Palaeozoic Basins
The Amadeus Basin, Irindina Province, and Georgina Basin once formed part of the Centralian Superbasin, a large intraplate basin that covered a significant part of Australia in the Neoproterozoic to early Palaeozoic (Maidment et al., 2013; Walter et al., 1995). This basin was broken up into its current constituents during the Petermann Orogeny between ∼560 and 520 Ma (Scrimgeour & Close, 1999) and the 450- to 300-Ma Alice Springs Orogeny (Collins & Teysier, 1989; Dunlap et al., 1995; Mawby et al., 1999).

2.2.1. Amadeus Basin
The northeastern Amadeus Basin overlies Warumpi and Aileron Province basement and is dominated by Neoproterozoic stratigraphy, due to uplift and erosion of the overlying units (Carr & Korsch, 2011). The 09GA-GA1 seismic line images the Amadeus Basin as being up to 3 km thick where it overlies the Aileron Province.

2.2.2. Irindina Province
The Irindina Province is a thick, highly faulted and metamorphosed Neoproterozoic to Cambrian sedimentary succession, which extends for ∼70 km in north-south extent in the center of the 09GA-GA1 MT and seismic line (Ahmad et al., 2006; Buick et al., 2001; Buick et al., 2005; Maidment et al., 2013).
Figure 3. Tectonic model of Betts et al. (2008) and Betts et al. (2015) for the development of north and south dipping convergent margins to the North Australian Craton (NAC) and present-day configuration. In this model, the Aileron Province was accreted to the NAC on a south dipping margin around 1,860 Ma. This was followed by the development of a north dipping convergent margin on the southern margin of the Aileron province, on which the Warumpi Province was subsequently accreted. Adapted from Betts et al. (2015) and (Korsch & Doublier, 2016) for the present-day configuration.

(2011) interpreted the Irindina Province to extend to ~5-s TWT or ~15 km in depth. The geometry was interpreted as a pop-up structure, bounded to the south by a northward dipping fault that connects to the Moho (Korsch et al., 2011). Pressure-temperature calculations give peak metamorphic pressures of 0.9–1.0 GPa, and ~800 °C, implying burial of Irindina Province rocks to ~30-km depth (Miller et al., 1997; Mawby et al., 1999). These pressure-temperature conditions were reached within an extensional, intraplate setting, implying that the Irindina Province rocks were part of a deep, intraplate sedimentary basin system (Maidment et al., 2013).

2.2.3. Georgina Basin
Rocks within the Georgina Basin are relatively undeformed compared to the Irindina Province, even though they are located only 50 km further north (Carr & Korsch, 2011) and are the same age (Maidment et al., 2013).

Korsch et al. (2011) interpreted a crustal boundary beneath the Georgina Basin, which separates the underlying Davenport province from the Aileron Province. The presence of significant deformation within the Irindina Province but not within the adjacent Georgina Basin likely reflects a stronger crust within the Davenport Province, such that majority of the crustal shortening occurred within the Aileron Province (Korsch et al., 2011).

3. Tectonic Evolution of the Arunta Region
3.0.1. The Arunta Region in Proterozoic Australia
Early models of the Proterozoic evolution of Australia proposed that Australian Proterozoic terranes have been in their current configuration since at least 2.5 Ga and that intraplate processes are responsible for shaping the orogenic belts in Central Australia (e.g., Etheridge et al., 1987).

More recent models describe the western two thirds of Australia in terms of the Archean to Paleoproterozoic South Australian Craton, NAC, and Western Australian Craton in various configurations over time, separated by Paleoproterozoic to Mesoproterozoic Orogens (e.g., Betts et al., 2015; Betts & Giles, 2006; Cawood & Korsch, 2008; Giles et al., 2004; Myers et al., 1996; Payne et al., 2009; Wade et al., 2006). These models agree that the Arunta region is located on or near the southern margin of the NAC; however, they differ in the extent to which this boundary was active during the Proterozoic. For example, Payne et al. (2009) argue that the North and South Australian Cratons were joined during the Paleoproterozoic.
Other tectonic models have the Arunta region as the site of a convergent tectonic boundary on the southern margin of the NAC from 1,800 to 1,600 Ma (Figure 3; Giles et al., 2004; Betts & Giles, 2006; Wade et al., 2006; Betts et al., 2015). The most recent of these (Betts et al., 2015) incorporated the seismic reflection interpretation of Korsch et al. (2011). The Atuckera Fault was interpreted to be an extension of the Willowra Suture, a boundary between the Aileron and Davenport Provinces interpreted on the 05GA-T1 seismic line (Goleby et al., 2009; Korsch et al., 2011). Betts et al. (2015) proposed that this boundary was the result of south dipping subduction between 1,910 and 1,860 Ma, culminating in the accretion of the Aileron Province against the NAC. This was followed by the development of a north dipping subduction zone on the southern margin of the Aileron Province between 1,850 and 1,840 Ma (Figure 3; Betts et al., 2015).

The Warumpi Province was accreted against the NAC around 1,640–1,635 Ma (Scrimgeour et al., 2005). The northern Warumpi Province margin was interpreted to be north dipping on the 09GA-GA1 seismic transect (Korsch & Doublier, 2016).

### 3.0.2. Orogenesis and Uplift During the Alice Springs Orogeny

During the 450-to-300-Ma Alice Springs Orogeny, crustal shortening of at least 80 km occurred in the Arunta region (Collins & Teyssier, 1989; Hand & Sandiford, 1999; Korsch et al., 1998). The Alice Springs Orogeny was associated with episodic pegmatite emplacement and fluid influx in the Aileron and Irindina Provinces, resulting in metasomatism of the host rock (Buick et al., 2008; Raimondo et al., 2011). In the eastern Arunta region, this shortening largely took place on the Illogwa Shear Zone and Bruna Detachment in the south, and the Delny and Entire Point Shear Zones in the north (Mawby et al., 1999; Scrimgeour & Raith, 2001b; 2001a; Scrimgeour, 2013a).

### 4. Seismic Data

The 09GA-GA1 deep crustal seismic reflection survey was completed along a 373-km profile in an approximately north-south direction and extends to 20-s TWT (Figure 1a; Maher, 2010; Korsch et al., 2011; Nakamura et al., 2011). Detailed acquisition parameters are listed in Nakamura et al. (2011).

In order to emphasize regions of enhanced reflectivity, we have calculated the standard deviation of the seismic amplitude over a vertical sliding window with a length of 30 points (120 ms). The section has been depth converted using velocities determined from stacking velocities. The resulting section is shown together with the original seismic data and interpretation of Korsch and Doublier (2016) in Figure 2.

This enhancement highlights many of the features interpreted by Korsch et al. (2011) and Korsch and Doublier (2016), including faults in the Irindina Province, and the three seismic packages identified by Korsch et al. (2011) within the Davenport Province and underlying Ooratippra Seismic Province (section 2.1.1). These are an upper, poorly reflective package from ~1 to 10 km; an intermediate package of moderate reflectivity (10 to 20 km); and a lower, highly reflective package (>20 km). The Moho was interpreted to be located at 40- to 50-km depth at the northern and southern end of the profile and to extend beneath the extent of the seismic data in the center of the profile (i.e., >60 km).

### 5. MT Data

#### 5.1. Acquisition and Processing

The MT data were acquired from 18 stations with both broadband and long-period instrumentation and 21 stations with broadband instrumentation only. The resulting station spacing for the 39 broadband instruments is 10 km, and for the 18-long-period instruments is 20 km. Full acquisition and processing details are provided in Duan and Milligan (2010) and are summarized here. The deployments used Australian National Research Facility for Earth Sounding instruments, which include Earth Data Loggers and copper/copper sulfate electrodes. The long-period deployments used Bartington Mag-03MS fluxgate magnetometers, while the broadband deployments used LEMI-120 induction coils. At each broadband station, the east-west and north-south magnetic and electric fields were recorded for 30–60 hr at a sampling rate of 0.001 s. Each long-period station recorded the east-west and north-south magnetic and electric fields, and vertical magnetic field, for 5 to 7 days at a sampling rate of 0.1 s.

In 2010, the broadband and long-period data were processed to period ranges of 0.003 to 300 s and 2 to 13,107 s, respectively. We have reprocessed these data using the Bounded Influence Remote Reference Processing software (Chave et al., 1987; Chave & Thomson, 2004) using remote reference data where available.
The quality of the resulting impedance tensor data was visually inspected and noisy data points masked. Good quality data were achieved over a bandwidth of 0.003 to 1,300 s at most stations. The reprocessed broadband data were merged with the processed long-period data from Duan and Milligan (2010) for analysis and modeling. At most of the long-period stations, the data quality is good from about 20 to ∼10,000 s. This provides an overlapping band of 20 to 1,300 s where both long-period and broadband measurements are available. We took advantage of this overlapping band to merge the long-period and broadband data.

At most stations, the broadband and long-period measurements were consistent in the overlapping band, making merging straightforward. However, at some sites, there is a DC shift (or static shift) between the two measurements. Static shift is a well-recognized phenomenon in MT data and results from inhomogeneities in the conductivity structure in the near surface (e.g., Árnason, 2015; Jones, 1988; Sternberg et al., 1988). In this survey, while the broadband and long-period measurements at a given location in this survey were collected close to each other, there were differences in the precise location and the time period of measurement. Therefore, the near-surface resistivity may have been different, resulting in the observed DC shifts between the broadband and long-period measurements.

At all stations except GB11 (with long-period site GL05) we shifted the long-period data to the level of the broadband. This is because the broadband data contain higher frequency signal, are generally more consistent between sites, and are considered more reliable than the long-period data. Comparison of adjacent broadband sites shows that in general, the resistivity is reasonably consistent in adjacent sites, suggesting that the broadband stations are not badly affected by static shifts. The exception is at GB11, where the Z_yx mode shows very low apparent resistivities (~1 Ωm) at periods <20 s, while at periods >20 s the data appear to be strongly affected by noise. The low apparent resistivity in this mode contrasts with the Z_xy mode at GB11, and both the Z_xy and Z_yx modes at the adjacent sites, GB10 and GB12. Visual inspection of the EX (north-south) electric field time series data shows poor signal at this site, which may explain the spurious apparent resistivities. Therefore, at this station, only the long-period data were used.

Figure 4 shows all elements of the impedance tensor as resistivity pseudosections. Figure 5 shows the apparent resistivity and phase, as well as the vertical magnetic field (tipper) for the long-period stations, for some representative stations.

5.2. Phase Tensor Analysis
The MT phase tensor is defined as

$$\Phi = \text{Re}(Z)^{-1} \text{Im}(Z)$$

and can be depicted as an ellipse (Caldwell et al., 2004). If the geoelectric structure is 1-D, implying that resistivity only varies with depth, the major and minor axes of the ellipse are the same, and the ellipse is a circle. If it is 2-D, the major or minor axis aligns with geoelectric strike, which is consistent both laterally and with period (Caldwell et al., 2004). Two-dimensional geoelectric structure means that resistivity varies only with depth and distance perpendicular to strike. Three-dimensional geoelectric structure, in which resistivity varies in all three dimensions, is indicated by either variable geoelectric strike and/or high skew angle magnitude. The skew angle $\beta$ can be calculated with the following equation:

$$\tan 2\beta = \frac{(\Phi_{12} - \Phi_{21})}{(\Phi_{11} + \Phi_{22})},$$

where $\Phi_{xy}$ are elements of the phase tensor $\Phi$. Caldwell et al. (2004) suggest that MT data are 3-D in the case where the magnitude of $\beta$ is large, which they define as $|\beta| > 3^\circ$.

The 09GA-GA1 MT data are displayed as phase tensor maps overlain on the geology and total magnetic intensity, in Figure 1 and as a phase tensor pseudosection in Figure 6.

In order to get an indication of the depth penetration associated with the phase tensors at different periods, the Niblett-Bostick transformation (Bostick, 1977; Jones, 1983; Niblett & Sayn-Wittgenstein, 1960) can be used. It is given by

$$Z_{mb} = \sqrt{\frac{\rho_{app} T}{2\pi \mu_0}},$$

where $T$ is the period of the measurement, and $\rho_{app}$ is the apparent resistivity.
Figure 4. Apparent resistivity and phase as a function of distance along the profile and period for the 09G1-GA1 MT transect. (a–d) Apparent resistivity for all four elements of the impedance tensor; (e and f) phase for the off-diagonal elements, with the $Z_{XY}$ phase shifted by 180°. Niblett-Bostick penetration depths of 1 km (solid lines), 10 km (dashed lines), and 100 km (dotted lines) plotted on the profile as a function of period for the off-diagonal elements.
Figure 5. Apparent resistivity and phase from six representative stations along the 09G1-GA1 MT transect (red and blue) with model responses from the final inversion model shown as black lines.

Niblett-Bostick depths of 1, 10, and 100 km are shown as a function of distance and a period along the profile, in Figure 4.

The phase tensors at a period of 0.26 s correlate well with the geology map (Figure 1), showing distinct character over the different geological provinces. The Georgina Basin is associated with $\phi_{\text{min}} < 45^\circ$ (indicating increasing resistivity with depth) and circular to elliptical ellipses with an east-west orientation. Over the Amadeus Basin, $\phi_{\text{min}}$ is similar, but the ellipses show a more variable orientation. In the Aileron Province, the phase tensors are more elliptical with the major ellipse axes generally perpendicular to the mapped fault orientations. This stronger ellipticity and orientation of the ellipses are consistent with the presence and orientation of the surface trace of the major mapped faults. In the Irindina Province, $\phi_{\text{min}}$ is generally $>45^\circ$ (indicating decreasing resistivity with depth), and the phase tensors show a west to northwest orientation from periods of 0.1 to 1.0 s, which aligns with the bounding faults to the Irindina Province near its margins. Finally, the two phase tensors in the Casey Inlier are both $>45^\circ$ with the $\phi_{\text{min}}$ axis approximately parallel to the two mapped faults here.

The phase tensors indicate that there are several areas in which $|\beta|$ is significantly greater than 3°, implying 3-D geoelectric structure. In the southern end of the transect, where the Amadeus Basin overlies the Casey Inlier and southern Aileron Province, the phase tensors are strongly elliptical with high skew angles ($>9^\circ$) from periods of ~1 to 100 s. Immediately south of the Irindina Province (near stations GB28–GB29),
Figure 6. Phase tensor pseudosection from south to north along the 09GA-GA transect, colored by skew angle $\beta$. Geological provinces colored as in Figure 1 shown for reference above the section. The phase tensors are oriented such that vertical on the figure is north.

The ellipses are strongly elliptical with high skew angle magnitudes from $\sim 0.03$ – 0.3 s ($\sim 1.5$- to 12-km penetration depth).

In summary, the phase tensors from this data set correlate well with the mapped geology. The phase tensors also indicate that the geoelectric structure in this region is 3-D. Therefore, 3-D inversion is required for this data. This will allow all elements of the impedance tensor to be inverted and will also allow the model to include off-profile features.

5.3. Modeling

We used the ModEM 3-D inversion code (Egbert & Kelbert, 2012; Kelbert et al., 2014) to generate a 3-D resistivity model from the 09GA-GA1 MT data. The model has a horizontal cell size of 1,500 m. The vertical cell size increases from 10 m at the surface to 20- at 200-km depth. Six padding cells were added at the base of the model, and seven at each of the sides, so the vertical extent is 1,030 km, the east-west extent is 470 km, and the north-south extent is 740 km. Topography was incorporated in the model from the ETOPO1 data set (Amante & Eakins B., 2009).

The full impedance tensor data plus tipper were used to generate the model. Error floors were set at 5% of $\sqrt{Z_{xy}Z_{yx}}$ for the impedance tensor and 0.05 for the tipper. A half-space of 100 $\Omega$ m was used as both the reference and the starting model. The model covariance (smoothing parameter) was set to 0.5 in each direction, throughout the model space.

The model was generated using a two-step process. First, the model was run using the impedance tensor data only, from a starting root mean square (RMS) misfit of 11.8 to a minimum achievable RMS misfit of 1.84, where the misfit is the ratio of the difference between the data and the model response, and the data error. The model was then restarted using this final model as a starting model, with the vertical magnetic field (tipper) data added. This second run had a starting misfit of 2.27, which reduced to a minimum of 1.97 after 100 further iterations.
Figure 7. Observed phase tensors (a and d), resistivity model with phase tensors from the model response (b and e), and residual phase tensors calculated using the method of Heise et al. (2007) (c and f). Resistivity model slices are at depth slices of −5.2-km (top panel) and −35.7-km (bottom panel) elevation relative to sea level (5.5- to 5.7-km and 36.0- to 36.2-km depth below surface depending on station elevation). Phase tensors are shown at periods of 0.5 s (top panel) and 32 s (bottom panel). Periods were selected according to the Niblett-Bostick transformation for the depth shown, using the geometric mean of the $Z_{XY}$ and $Z_{XZ}$ apparent resistivity for that period. Observed and predicted phase tensors are colored by the maximum phase; residual phase tensors are colored by the arithmetic mean residual. All plots are shown with the geological unit outlines as shown in Figure 1 (gray lines).
Many model runs were carried out to test the sensitivity of the model to different inputs. Parameters tested include the reference model, the covariance, the horizontal and vertical grid resolution, and the model grid extent. The additional model runs also included inversions with various subsets of the data (e.g., tipper only inversion, impedance only, long period only, and every second site), and with different error floors, to test the robustness of features in the model. While the topography varies by only a few hundred meters over the 380-km north-south extent of the data, models were run with and without the topography with only minor differences between the models.

Most of these inputs made little difference to the model results. The inputs that made the most noticeable impact on the results were the prior model, and the models run with subsets of the data.

As would be expected, the runs carried out on subsets of the data (e.g., tipper only, every second site, and using periods >10 s only) made noticeable differences in the model. However, in most of these models, all the features discussed were present but with differences in their geometry. The exception to this is the tipper only inversion, which showed larger differences from the inversions using impedance and tipper data. This might be expected given that the tipper contains no information on absolute resistivity and there is no tipper data available at the broadband stations.

Changing the reference model to a 10-Ωm half-space resulted in a slightly better starting misfit (10.9) but worse final misfit (2.9) than with a 100-Ωm reference model and made the output model noticeably more conductive. Conversely, changing the reference model to a 1,000-Ωm half-space resulted in a worse starting misfit (35.7) but a similar final misfit (2.03) and made the output model more resistive. The key features discussed are present in all of these models but have a different geometry when starting from 10 or 1,000 Ωm.

Intermediate reference models of 200 and 300 Ωm were also run. The model run with a reference model of 200 Ωm had slightly better final RMS misfit to the model run with a 100-Ωm reference model. However, once tipper data were added to the inversion, a reference model of 100 Ωm reached the lowest RMS misfit and was therefore chosen as the final model. The geometry of features delineated with each of these reference models was very similar and would not have a significant influence on the interpretation of the model.

5.4. Resistivity Model

Our preferred resistivity model is shown in Figure 7 as a series of depth slices, and in Figure 8 as a vertical profile together with the seismic reflection data. The fit between the data and model response is illustrated in Figure 5 for selected stations (see the supporting information for all stations) and Figure 9. The fit to the data is also illustrated in Figure 7 as observed, predicted and residual phase tensors (Heise et al., 2007) although we note that the inversion was run to fit the impedance tensor, not the phase tensor.

On the broad scale, a marked difference can be observed in the resistivity model from north to south. This difference is clear both on the vertical section and in the depth slices, which allow us to see any off-profile features the inversion has produced to fit the data. At the northern end of the profile, the Davenport Province and overlying Georgina Basin are associated with flat-lying structure in the resistivity model (Figure 8). In the southern end of the profile (>140 km in Figure 8), the resistivity model contains more vertical structure, with conductors within the Aileron Province extending upward beneath the Amadeus Basin and Irindina Province.

The most prominent anomalies in the resistivity model are in the center of the profile beneath the Irindina Province (C1; Figure 8). They are strongest from 10- to 20-km depth, with resistivities as low as 1 Ωm, but appear to connect to a moderately conductive pathway down to ~50-km depth. When viewed on the profile, the anomalies appear to terminate close to the base of the Irindina Province (Figure 8); however, when viewed as depth slices, they extend up to several conductive anomalies that can be seen either side of the profile from depths of 2–10 km below the surface (shown at 5 km in Figure 7). These are expressed as alternating resistive and conductive features, either side of the profile with a spacing (peak to trough) of about 10 km, similar to the station spacing.

We cannot have a high degree of confidence in the positioning and geometry of these features due to their spacing and location beside, and not directly below, the stations. However, these structures were persistent in the inversions. They are present in models with a higher covariance (smoothing parameter) and those with subsets of the data. The robustness of these features was also tested by removing...
Figure 8. (a) Seismic reflection amplitude data, depth converted using a velocity model derived from stacking velocities, overlain by our interpretation (modified after Korsch & Doublier, 2016). Extent of key provinces also shown. (b) Magnetic inversion result from Chopping et al. (2011). (c) Bouguer gravity anomaly data along the 09GA-GA1 profile, from the Geophysical Archive Data Delivery System database (Geoscience Australia, 2018). (d) Resistivity model extracted onto the profile and overlain, semitransparent, on the depth-converted seismic reflection data. (e) Seismic reflection interpretation and resistivity model for the top 5 km with five times vertical exaggeration. (f) Seismic reflection interpretation and resistivity model down to 60 km. (g) Interpretation of resistivity model, shown with same colors as Figure 2d. All models and data (with the exception of gravity data and shallow view of resistivity model) shown on an equal horizontal and vertical scale.
them from the model, and running an inversion. The resulting forward model had an increased global RMS misfit of 3.02. Furthermore, when an inversion was run from this forward model, these features returned.

One explanation for these is the presence of resistivity anisotropy. The ModEM code is unable to model anisotropy but may fit the data in the presence of anisotropy by placing alternating conductive and resistive structures beside the profile. A similar phenomenon was observed by Patro and Egbert (2011) in their inversion of profile data over the Deccan Volcanic Province, Western India, and by Meqbel et al. (2014) in their inversion of long-period MT data over the northwestern United States. Furthermore, Heise and Pous (2001) showed in a synthetic study that when 2-D isotropic inversion is carried out over anisotropic structure, the anisotropy appears as a sequence of conductive and resistive dykes in the inversion.

Anisotropy is the dependence of a property on orientation and can occur on a variety of scales (Wannamaker, 2005). Often anisotropy in the crust occurs on a macroscopic scale; that is, the rocks and minerals are not intrinsically anisotropic, but the presence of fractures containing a conductive phase (e.g., saline fluids, sulfide minerals, and graphite) causes them to appear anisotropic on the scale observed by MT surveys (Wannamaker, 2005).

In some cases, there are indications of anisotropy in the data. For example, if the subsurface is 1-D and anisotropic, then uniform phase splits, and small magnitude tippers are expected (Heise et al., 2006; Martí, 2014). In 2-D and 3-D models it is more difficult to uniquely identify anisotropy (Martí, 2014). Characteristics such as phases outside of the normal ranges of [0°, 90°] and [180°, 270°] and inconsistencies between the impedance tensor and induction vectors may be present; however, these do not always occur in the presence of anisotropy, and furthermore, they may result from other sources such as galvanic distortion (Pek & Verner, 1997; Heise & Pous, 2003). In this data set, we see phase splits in the periods with penetration depths of ~2- to 10-km depth (0.2 to 1.0 s). The phase tensors are moderately uniform across the Irindina Province at periods >0.1 s, although there is some variability in skew angle and ellipse orientation (Figures 1, 6, and 7). We cannot use the tipper information for this anomaly, as it is only available with the long-period data (periods >25 s). We see phases out of the [0°, 90°] and [180°, 270°] ranges at stations 27, 34, 35, and 36 (see Figure 5 and the supporting information); however, in all cases these occur at periods >10 s, which would be associated with deeper sources than the potential anisotropy, and of these only Station 27 is located in the Irindina Province (near the boundary with the Aileron Province).

The Georgina Basin is associated with a thin conductive layer (up to 2 km thick). Likewise, the seismic reflection data image the Georgina Basin as being up to 2 km thick. Below this is a resistive layer (R1; 1,000–3,000 Ωm), which extends to 9- to 15-km depth. This layer corresponds approximately to a weakly reflective seismic package (Korsch et al., 2011) although there are slight differences in the depth extent of the layer.

Beneath this layer is a conductive layer C2 with resistivities of ~5–20 Ωm. This coincides with two seismic packages described by Korsch et al. (2011), an upper layer of moderate reflectivity that overlies a highly reflective package, which was interpreted to be the basement to the Davenport Province. The southern limit of this layer agrees well with the positioning of the Atuckera Fault/Willowra Suture in the seismic reflection data, which is thought to form the southern limit of the Davenport Province (section 3; Korsch et al., 2011).

The Casey Inlier is associated with high resistivities. The Amadeus Basin over the inferred Warumpi Province is associated with low resistivities (~1–50 Ωm), but further north, where it overlies rocks of the Aileron Province, it is associated with higher resistivities (~200 Ωm). A possible explanation for this is that where it overlies the Aileron Province, it is highly deformed and thicker (and therefore may be more compacted) but within the Warumpi Province it is shallower, with significantly less deformation mapped (Carr & Korsch, 2011).
Immediately south of the Irindina Province, there is a zone approximately 30 km wide, in which a series of faults and shear zones are mapped at the surface and interpreted from the seismic reflection data to extend to depths of at least 20–30 km. This zone is resistive (R2; $\sim$3,000–5,000 $\Omega$m) and coincides with a low gravity anomaly (Figure 8). South of this resistive zone are two conductors (C3; 20 $\Omega$m) at $\sim$20- to 30-km depth beneath the Amadeus Basin, which appear to connect to a moderately conductive (100 $\Omega$m) pathway originating at $\sim$50-km depth.

6. Discussion

Our inversion model is largely consistent with the 2-D inversion of long-period data across the Aileron and Irindina Provinces, and southernmost Davenport Province (Selway, 2006). Both inversions resolve conductors beneath the Davenport and Irindina Provinces with a resistive zone associated with the Delny Shear Zone.

6.1. Conductivity Anomalies in the Irindina Province

In section 5.4 we described strong ($\sim$1 $\Omega$m) conductivity anomalies (C1) that appear to connect up to several off-profile, potentially anisotropic conductors within the Irindina Province. Both the deep and shallow parts of these anomalies are associated with high reflectivity (Figures 2 and 8).

Most of the rock-forming minerals are poor conductors (e.g., Pommier, 2014; Telford et al., 1990; Yoshino, 2010), and sources for conductivity anomalies often result from addition of fluids, through the fluids themselves or through fluid-rock interaction, which can result in the deposition of conductive minerals such as sulfides or graphite.

Saline fluids can reduce the resistivity in porous sedimentary rocks or in permeable fractured rocks (e.g., Caldwell et al., 1986; Hyndman & Shearer, 1989; Heise et al., 2008; Kirkby et al., 2015; Kirkby & Heinson, 2017; Ussher et al., 2000). Hyndman and Shearer (1989) determined that 10-km-thick conductive layers at 20- to 30-km depth with conductivities of 20–30 $\Omega$m can be explained with 0.5–3% pore fluid water assuming a fluid with a resistivity of 0.02 $\Omega$m (based on the salinity of sea water, 3%, at lower crustal conditions, i.e., 500 °C and 0.6–1.0 GPa).

At a depth of 10 km ($\sim$0.3 GPa) the resistivity of a fluid with 3% salinity would be $\sim$0.04 $\Omega$m or greater (Bannard, 1975). More saline fluids up to 25% may have a resistivity as low as $\sim$0.01 $\Omega$m, half that considered by Hyndman and Shearer (1989); however, the anomalies within and beneath the Irindina Province are an order of magnitude more conductive than those considered by Hyndman and Shearer (1989).

Moreover, the most recent known fluid influx event in the Irindina Province was during the 450- to 300-Ma Alice Springs Orogeny. The episodic fluid release during the Alice Springs Orogeny would likely have expelled much of this water. Any remaining water is likely to have since been reabsorbed by the crust (Yardley & Valley, 1997). Thus, it is unlikely that saline fluids are the cause of the anomalies seen in the Irindina Province.

Other possible sources for the Irindina Province anomalies include sulfide mineralization and/or graphite. Both graphite and many sulfide minerals are extremely conductive ($\sim$$10^{-4}$ $\Omega$m; Telford et al., 1990; Pearce et al., 2006), so only small volumes (of the order 1–2%; Watson et al., 2010) are needed to enhance the conductivity if the conductive phases are sufficiently connected. Several metal occurrences and deposits, including massive sulfides, have been identified above C1a, particularly its southern limb (Huston et al., 2011; Scrimgeour, 2013b). Furthermore, several authors (e.g., Buick et al., 2008; Raimondo et al., 2011) have noted that rock samples collected from this region are highly sheared, which means that any conductive phases are more likely to be connected. The presence of shearing is also consistent with anisotropic electrical properties as discussed in section 5.4. Due to its presence in the near surface, we therefore suggest sulfide mineralization in fractures to be the most likely cause of the conductivity anomalies near the base of the Irindina Province.

6.2. Conductivity Anomalies in the Aileron and Davenport Provinces

In the Aileron Province, there are two $\sim$20 $\Omega$m conductivity anomalies at $\sim$20- to 40-km depth beneath the Amadeus Basin (C3). Beneath the Davenport Province, there is a conductive layer at $\sim$10- to 40-km depth (C2).
In the Aileron Province, the conductivity anomalies correlate with a strong magnetic susceptibility anomaly and a positive gravity anomaly (Figure 8; Chopping et al., 2011). Saline fluids and graphite would have a weak negative effect on both the magnetic susceptibility and density, and therefore, we suggest that a magnetic, dense source is more likely.

Most unaltered rock types have low magnetic susceptibilities Telford et al. (1990). On the high end are metasedimentary rocks (0.035 SI units) and granites (up to 0.05). Resistivities for these rock types are normally >1,000 Ωm (Telford et al., 1990). Thus, alteration minerals are likely required to explain the high magnetic susceptibilities and conductivities.

Magnetic and conductive alteration minerals include pyrrhotite, magnetite, and ilmenite. All of these minerals have densities >4,500 kg/m³ (Telford et al., 1990) and have been reported at varying concentrations in massive sulfide deposits hosted within the Strangways Metamorphic Complex in the Aileron Province (Huston et al., 2011; Hussey et al., 2005).

The formulation of Werner (1945) relates the volumetric proportion and magnetic susceptibilities of the constituents of a rock to its bulk magnetic susceptibility. Using this formula, the volumetric proportion of magnetic minerals required to explain the anomaly can be calculated. To explain a magnetic susceptibility of 0.1 SI units requires volumes of at least 4% magnetite or at least 10% pyrrhotite (Telford et al., 1990; Werner, 1945). These proportions would increase the density of the rock, leading to a positive gravity anomaly as observed, and, if sufficiently well connected, would reduce the resistivity to the values seen in the resistivity model.

We consider one or both of the above sources as the most likely interpretation due to their presence in near-surface rocks and the fact that they would be expected to increase both density and magnetic susceptibility; however, we note that it is possible that there are additional nonmagnetic conductive phases (e.g., nonmagnetic sulfide minerals). While graphite could be present, it has not been noted in analyses or descriptions of surface rocks in this area (e.g., Hussey et al., 2005; Raimondo et al., 2011; Scrimgeour, 2013b).

In the Davenport province, there is a resistive layer that extends to ~10- to 20-km depth associated with generally low magnetic susceptibility and low reflectivity. Below this, there is a 7- to 50-Ωm layer (C3), associated with higher reflectivity and moderately high susceptibilities (0.02 to 0.06 SI). The southernmost extent of the conductor correlates remarkably well with the positioning of the Atuckera Fault crustal boundary (Korsch et al., 2011). The depth to the top of the conductor increases with distance north, correlating with the subhorizontal detachment surface identified in the seismic reflection data (Korsch et al., 2011) near the Atuckera fault, but sitting up to 5–10 km deeper than this detachment surface further north.

The Warramunga Formation underlies rocks of the Davenport Province (Scrimgeour & Close, 2011) and is exposed north of the 09GA-GA1 line in the Tennant Creek region. It contains banded ironstone and is intruded by Tennant Creek Suite granites that contain magnetite (Donnellan, 2013). This composition would explain the high magnetic susceptibilities seen from 10- to 20-km depth on the profile. The electrical conductivity of banded ironstone would depend on the connectivity of magnetite grains within it, but some authors have found that it is conductive (as low as 0.87 to 4.5 Ωm; Morris et al., 1980). If the banded ironstone in the Warramunga Formation is conductive, then this might explain the 7- to 50-Ωm layer seen at >10- to 20-km depth in the resistivity model and would suggest that the layer represents Warramunga Formation equivalents at depth beneath the Davenport Province.

### 6.3. Crustal Structures in the Eastern Aileron Province

Both the resistivity model and the seismic reflection data suggest the presence of deep, subvertical structure in the southern end of the profile. In the resistivity model, there are two conductors within the Aileron Province that dip steeply northward beneath the Amadeus Basin. In the seismic data, there appear to be several discrete zones of high reflectivity that coincide with these conductors (Figures 2b and 8).

Korsch and Doublier (2016) interpreted a crustal boundary between the Warumpi and Aileron provinces, at the southern margin of this anomaly (Figure 2). We interpret the features described above in the resistivity model and seismic reflection data as fluid pathways along steeply dipping structures with a similar trend as the Warumpi Province margin.
6.4. Implications for the Structural Evolution of the Arunta Region

Our resistivity model supports the presence of both a north dipping and south dipping convergent margin in the Arunta region, and the notion that the Aileron Province accreted against the NAC, as proposed by Betts et al. (2015; section 5). The positioning and geometry of the south dipping Atuckera Fault/Willowra Suture (Betts et al., 2015; Korsch et al., 2011) agrees with the southern boundary of the conductive layer in the Davenport Province (section 6.2). There is a clear difference in the resistivity and reflectivity structure across this boundary.

The resistivity model shows a vertical to steeply north dipping margin between the Aileron and Warumpi Provinces (Figure 8; section 6.3). This is in agreement with the seismic interpretation of Korsch and Doublier (2016) and the tectonic model of Betts et al. (2008) in which accretion of the Warumpi Province occurred on a northward dipping subduction zone on the southern margin of the NAC.

7. Conclusions

A 3-D inversion of MT data from the eastern Arunta region has been reprocessed and modeled to provide new insights on the geology in this region. The key findings are as follows:

- The model contains off-profile conductors in the Irindina Province, which may result from anisotropy on the scale of these MT data. The features are interpreted as sulfide mineralization in east-west striking fractures and faults.
- The resistivity model images conductive anomalies that are coincident with a magnetic susceptibility anomaly beneath the Amadeus Basin in the Aileron province. These are explained in terms of magnetic and conductive alteration minerals such as magnetite or iron sulfides.
- The resistivity model indicates the presence of flat-lying structure in the Davenport Province. In contrast, in the Aileron province, the resistivity model is more complex and includes more vertical structure, which may reflect deposition of conductive minerals along steeply dipping faults. The margin between these provinces is imaged as south dipping, with the positioning agreeing well with previous interpretations of the seismic reflection data along the same transect. This is consistent with accretion of the the Aileron Province along a south dipping subduction zone (Betts et al., 2015).
- The margin between the Aileron and Warumpi Provinces appears to be steeply north dipping, consistent with the seismic interpretation of Korsch and Doublier (2016) and the tectonic model of Betts et al. (2008).

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**Erratum**

In the originally published version of this manuscript, Equation (1) was incorrectly given as

\[ \Phi = \frac{\text{Re}(Z)}{\text{Im}(Z)} \]

The results and conclusions of the article are unaffected, because the correct formula was used in the calculations. The typographical error has been corrected, and this may be considered the official version of record.