Lithospheric structure and melting processes in southeast Australia: new constraints from joint probabilistic inversions of 3D magnetotelluric and seismic data

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Key Points:

• We apply a novel approach for joint probabilistic inversions of 3D magnetotelluric and seismic data.
• We use the new method to image the lithosphere-asthenosphere system beneath southeastern Australia.
• The imaged lithospheric structure provides insights on metasomatism and melt production in the region.

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Abstract

The thermochemical structure of the lithosphere controls melting mechanisms in the mantle, as well as the location of volcanism and ore deposits. Obtaining reliable images of the lithosphere structure, and its complex interactions with the asthenosphere, requires the joint inversion of multiple data sets and their associated uncertainties. In particular, the combination of seismic velocity and electrical conductivity, along with proxies for bulk composition and elusive minor phases, represents a crucial step towards fully understanding large-scale lithospheric structure and melting processes. We apply a novel probabilistic approach for joint inversions of 3D magnetotelluric and seismic data to image the lithosphere beneath southeast Australia. The results show a highly heterogeneous lithosphere with deep conductivity anomalies that correlate with the location of Cenozoic volcanism. In regions where the conductivities have been at odds with sub-lithospheric temperatures and seismic velocities, we observe that the joint inversion provides conductivity values consistent with other observations. The results reveal a strong relationship between metasomatized regions in the mantle and i) boundaries of geological provinces, elucidating the subduction-accretion process in the region; ii) distribution of leucitite and basaltic magmatism; iii) independent geochemical data, and iv) a series of lithospheric steps which constitute areas prone to generating small-scale instabilities in the asthenosphere. This scenario suggests that shear-driven upwelling and edge-driven convection are the primary mechanisms for melting in eastern Australia, contrary to the conventional notion of mantle plume activity. Our study presents an integrated lithospheric model for southeastern Australia and provides valuable insight into the mechanisms driving surface geological processes.

Plain Language Summary

The lithosphere is the Earth’s outermost rigid layer and the focus of important geological processes, such as earthquakes, volcanism, and mineralization. The location of these processes often coincides with deep discontinuities in the lithosphere. Imaging the structure of the lithosphere using geophysical techniques is crucial to fully understand the nature of these processes.

The most reliable images of the lithosphere are obtained by joint analysis of two or more geophysical data sets. In particular, the combination of magnetotellurics (an electromagnetic technique) and seismic data holds great potential because of their complementary sensitivity to the Earth’s properties. Furthermore, our understanding of the variability of the lithospheric structure is enhanced by combining this joint analysis with a probabilistic approach. This is because probabilistic methods provide a large number of models that can explain the data. Given the substantial data coverage in southeast Australia, we use a new probabilistic approach for the joint analysis of magnetotelluric and seismic data to image the lithospheric structure beneath this region. Our results show a complex lithosphere in line with the location of volcanism and the tectonic history of the region. The lithospheric composition derived from the models provides significant insights into melt production in the area.

1 Introduction

The magnetotelluric method (MT) has great potential for investigating metasomatism and tectono-magmatic processes in the lithosphere (e.g., Wannamaker et al., 2008; Comeau et al., 2015; Aivazpourporgou et al., 2015; Wannamaker et al., 2014; Bedrosian, 2016; Kirkby et al., 2020; Blatter et al., 2022; Özaydın & Selway, 2022; Cordell et al., 2022). Due to its sensitivity to fluid and/or melt content, MT is particularly useful for probing the connection between deep melt/fluid pathways and their surface expressions, such as the location of ore deposits (e.g., Griffin et al., 2013; Heinson et al., 2018; Kirkby et al., 2022) and volcanic centers (Wei et al., 2001; Comeau et al., 2015). However, MT is not free from limitations.
For instance, MT struggles to image structures beneath good conductors because the electromagnetic (EM) fields, which diffuse downward into the Earth, decay rapidly within the conductors. This results in minimal to no field diffusion occurring downwards, particularly when significant damping occurs at the top of a conductor (Jones, 1999). The MT method is also ambiguous in discerning the different factors that affect electrical conductivity, such as temperature, water/melt content and composition. Unlocking the full potential of the MT method requires the development of methodologies that can assign meaningful physical interpretations to conductivity anomalies and discriminate between their causes (Selway, 2014).

A widely adopted approach to reduce feature ambiguity is the combination of MT with other geophysical data sets by joint inversions (e.g., Khan et al., 2006; Gallardo & Meju, 2007; Jegen et al., 2009; Moorkamp et al., 2010; Bennington et al., 2015; Afonso, Rawlinson, et al., 2016; Jones et al., 2017; Blatter et al., 2019; Manassero et al., 2021; Liao et al., 2022; Wu et al., 2022). Taking advantage of the complementary sensitivities of different data sets to the properties of interest, joint inversions minimize the range of acceptable models consistent with the available data and can increase model resolution (e.g. Moorkamp et al., 2007; Afonso et al., 2013a; Afonso, Rawlinson, et al., 2016; Afonso, Moorkamp, & Fullea, 2016). For example, in the case of MT and seismic data, both data sets are sensitive (to different degrees) to the background thermal and compositional structure of the lithosphere. However, only MT is strongly sensitive to minor conductive phases (e.g., hydrous minerals and graphite), hydrogen content or small-scale melt/ﬂuid pathways (Karato, 1990, 2006; Evans, 2012; Yoshino, 2010; Khan, 2016; Selway, 2014). In this way, joint MT + seismic inversions have great potential to improve the resolution of conductivity structures (e.g., Moorkamp et al., 2007, 2010; Gallardo & Meju, 2007), detect regions of partial melting and fluid pathways in the lithosphere (cf., Selway & O’Donnell, 2019; Evans et al., 2019; García-Yeguas et al., 2017; Bennington et al., 2015), and understand their relationship with the location of ore deposits (e.g., Takam Takougang et al., 2015) and metasomatized lithologies (e.g., Snyder et al., 2014).

In addition to the benefits of joint inversions, valuable information about model uncertainties can be obtained by simulation-based probabilistic approaches (Tarantola, 2005; Rosas-Carbajal et al., 2013; Afonso, Rawlinson, et al., 2016; Manassero et al., 2021). Rather than outputting a single best-ﬁtting model, probabilistic approaches provide a distribution of models and their associated probabilities according to their performance in explaining the observations. Thus, probabilistic inversions naturally address the non-uniqueness problem in geophysics (particularly in MT) and quantify the model ambiguity (Tarantola, 2005; Gregory, 2005). Probabilistic methods, however, require the evaluation of millions of possible models, each in turn requiring the computation of a forward solution. This limits simulation-based probabilistic approaches to problems where fast forward operators are available. In the case of 3D MT inversions, fully probabilistic methods have been infeasible due to the large CPU time required by the associated forward problem (Miensopust et al., 2013). To address this limitation, Manassero et al. (2020) developed a novel strategy based on reduced order modeling (referred to as RB + MCMC) that allows fast and accurate approximations of the forward solution and performing joint probabilistic inversions of 3D MT data with other data sets (Manassero et al., 2021). Potential applications and the efficiency of the method to solve the joint inverse problem of MT and seismic data were demonstrated with whole-lithosphere synthetic examples in our previous paper (Manassero et al., 2021).

In this work, we apply the method of Manassero et al. (2020) and Manassero et al. (2021) to a dense array of collocated MT and seismic data in southeastern Australia. This region is known to have experienced multiple orogenic events that resulted in a complex crustal architecture, but its deep lithospheric roots remain poorly characterized (Rawlinson et al., 2016). It also hosts one of the most voluminous intraplate volcanic provinces on Earth, the Eastern Australian Volcanic Province (EAVP, Johnson et al., 1989). Half of the erupted volume in the EAVP is age progressive (Wellman & McDougall, 1974). This
volcanism has been linked to a hot mantle plume, initially by radiometric dating (Wellman & McDougall, 1974) and later through plate velocity modeling correlating with radiometric dates (Sutherland et al., 2012; Davies et al., 2015), while the melting mechanism responsible for the other half is less clear (Wellman & McDougall, 1974; Shea et al., 2022). Here we focus on the mapping of whole-lithosphere 3D structures and sub-lithospheric temperature anomalies in order to investigate i) the origin of the intraplate magmatism with no clear plume signatures, ii) the connection between deep melt/fluid pathways and the location of volcanic centers, and iii) how lithospheric structure may have influenced melting generation and transport. Southeastern Australia is also a region with abundant xenolith-derived datasets (cf. Shea et al., 2022, and references therein) that can be used to validate the results from our joint inversion.

2 Geological background

The Tasmanides in southeastern Australia are a complex orogenic system that developed from west to east through repetitive cycles of subduction, accretion and lithospheric deformation along the eastern margin of Gondwana (Glen, 2005, 2013; Champion et al., 2016; Rosenbaum, 2018; Moresi et al., 2014). This region is broadly divided into the Delamerian Orogen in the west (early-Palaeozoic) and the younger (mid-Paleozoic) Lachlan Orogen in the southeast (Figure 1.a). Much of the geological complexity in the area can be explained by a geodynamic model of a microcontinent collision and the subsequent development of an orocline, referred to as the Lachlan Orocline model (Cayley, 2011; Cayley & Musgrave, 2015; Moresi et al., 2014; Musgrave, 2015). The major structures described in this model are curved crustal geometries with an eastward rotation (Musgrave, 2015), which have been imaged by gravity, magnetic and potential field data (e.g., Musgrave & Rawlinson, 2010; Nakamura & Milligan, 2015; Nakamura, 2016, see Figure 1.c); ambient noise tomography in the crust (e.g., Young et al., 2013; Pilia et al., 2015, see Figure 8.d) and MT conductivity models (Aivazpourporgou et al., 2015; Kirkby et al., 2020; Heinson et al., 2021). Important first-order information about the lithospheric structure beneath southeast Australia has been obtained from conventional studies, such as ambient noise and teleseismic tomography (Rawlinson et al., 2016; Davies et al., 2015; Young et al., 2013), xenolith thermobarometry (e.g., Lu et al., 2018), thermal modeling (e.g., Tesauro et al., 2020) and a recent 3D conductivity model (Kirkby et al., 2020). This latter conductivity model showed for the first time that some of the crustal structures associated with the orocline persist below the Moho, providing new insights into the lithospheric architecture and geodynamic history of the region.

Throughout the late Mesozoic and the entire Cenozoic, eastern Australia was subject to voluminous mafic intraplate volcanism, which formed the extensive Eastern Australian Volcanic Province (EAVP, Johnson et al., 1989; Sutherland et al., 2012; Shea et al., 2022). Several regions across eastern Australia contain recent eruptions; in north-eastern Australia, the Kinrara vent contains lavas ∼7 ka ± 2 ka (Cohen et al., 2017), while Mount Gambier, Newer Volcanics (NV) in southeastern Australia contain lavas ∼4–7.5 ka (Blackburn et al., 1982; Smith & Prescott, 1987).

The EAVP comprises 67 separate volcanic centers with two dominant volcanic center compositions: basalt and potassic leucitite (Figure 1.b). Although basaltic volcanics erupted through a thinner lithosphere (< 110 km) along the eastern and south-eastern seaboard, the leucitite volcanic centers lie on thick lithosphere (> 125 km) in central New South Wales and central Victoria (Davies & Rawlinson, 2014; Rawlinson et al., 2017). This leucitite suite represents the most atypical and extraordinarily enriched melt compositions reported for mafic melts in eastern Australia (Cundari, 1973; Birch, 1978). In particular, they represent melts from pervasively metasomatized source assemblages, likely a Ti-bearing oxide phlogopite websterite ± apatite (see the recent review by Shea et al., 2022). The lack of anhydrous peridotite and the abundance of hydrous minerals in their mantle source assemblages is of
particular importance to this work, indicating widespread mantle metasomatism beneath eastern Australia.

The EAVP is also unique in the sense that about half of the volcanism is age-progressive and commonly linked to a hot mantle plume (e.g., Sutherland et al., 2012; Davies et al., 2015), while the remaining volcanic centres show no age-progression and no obvious melting mechanism (Wellman & McDougall, 1974). To further exacerbate this issue, lava compositions throughout the EAVP (including both age-progressive and non-age-progressive volcanic centers) argue for low-temperature melting of metasomatized mantle source lithologies. Contrary to the presence of a hot mantle plume, these compositions suggest mild, but localized perturbations in mantle temperatures (Shea et al., 2022).

3 Methods and data sets

3.1 Data

The input data used in our joint probabilistic inversion include magnetotelluric (MT) data from the AusLAMP array (Australian Lithospheric Architecture Magnetotelluric Project) in southeast Australia and P-wave velocities from the tomography model of Rawlinson et al. (2016). Long-period MT data were acquired at 298 AusLAMP stations (blue triangles in Figure 1f) across a ~55 km spaced array covering an area of 950 × 950 km. The MT data are the full impedance tensor for periods between 6.4 to 40,000s. Data errors were provided by the impedance estimation process (Kirkby et al., 2020). Details about data acquisition and processing are given in Kirkby et al. (2020). Error floors represent a lower bound for the data errors, which means that no estimated data errors can be less than the specified error floors. Error floors are set to 5% of max(\(|Z_{xx}|, |Z_{xy}|\)) for the components \(Z_{xx}\) and \(Z_{xy}\) and 5% of max(\(|Z_{yy}|, |Z_{yx}|\)) for the components \(Z_{yy}\) and \(Z_{yx}\). This is justified by the fact that most of the errors in the impedance estimation come from the electric fields (Simpson & Bahr, 2005), which means that the absolute error in \(Z_{xx}\) will be the same as the absolute error in \(Z_{xy}\), as they share the \(E_x\) channel. The same happens for \(Z_{yx}\) and \(Z_{yy}\) as they share the \(E_y\) channel. We assume uncorrelated data errors that follow a double exponential distribution (see next section).

The P-wave velocity model used in this study (Rawlinson et al., 2016) was constructed from teleseismic tomography using data from the mainland component of the WOMBAT transportable seismic array (Rawlinson et al., 2015). In order to account for the unresolved crustal component of the teleseismic arrival time residuals, the model of Rawlinson et al. (2016) includes a detailed crustal model from ambient noise tomography (Young et al., 2013) and the Moho from AuSREM (Figure 1d, Kennett & Salmon, 2012) as starting model of the tomographic inversion. (see details in Rawlinson et al. (2016)). Despite this additional constraint, uncertainties in absolute velocities within the crust remain relatively high, especially in the lower-crust and Moho (Young et al., 2013; Rawlinson et al., 2016). We obtained seismic velocities by interpolating the velocity model on a data-point grid of 50×50 km at the surface (shown in red dots in Figure 1f) and 24 points between the surface and 340 km depth. The velocity data errors are assumed to be uncorrelated and normally distributed with a standard deviation of 1%, according to Burdick & Lekić (2017).

3.2 Bayesian inversion

In the Bayesian or probabilistic approach to the inverse problem, inference about the model parameters \(m\) given observed data \(d\) is based on the so-called posterior probability density function (PDF):

\[
P(m|d) = \frac{P(d|m)P(m)}{P(d)} \propto \mathcal{L}(m)P(m) \propto exp(\phi)P(m), \tag{1}
\]
Figure 1. Maps showing (a) orogens in southeastern Australia, grey outlines denote geological provinces (Raymond et al., 2018). This panel includes a map of Australia and the study area in blue. (b) Mesozoic to Cenozoic sedimentary basins (Raymond et al., 2012), leucitites volcanoes (orange) and basaltic volcanics (pink) after Shea et al. (2022). The basaltic in NV are highlighted in purple. (c) Total magnetic intensity map (TMI, Nakamura & Milligan, 2015). (d) Moho depth from the AusREM model (Kennett & Salmon, 2012) where 5km contour lines are shown in dashed-grey. (e) Mean crustal RHP from Haynes et al. (2020). (f) Elevation map of southeast Australia including AusLAMP MT stations (blue triangles) and the location of the velocity data (red dots). Panel (a) and (c) show major tectonic boundaries are outlined in grey. White triangles indicate stations where data fits are shown.
where \( P(m) \) denotes the prior PDF describing all the information on the model’s parameters prior to the inversion (e.g., prior geological or petrological knowledge in the area of study). \( \mathcal{L}(m) \) is the likelihood function, which is specified by the statistical distribution of data errors, and \( \phi \) is the misfit of model \( m \). In the case of MT, we assume uncorrelated data errors that follow a double exponential or Laplace distribution (Tarantola, 2005; Menke, 2018). The data misfit is then given by (Tarantola, 2005):

\[
\phi = -\sum_{i=1}^{N} \frac{|g_i(m) - d_i(m)|}{s_i}.
\] (2)

The double exponential distribution is more robust against outliers and is often a more realistic assumption than the normal distribution (e.g., Shearer (1997); Farquharson & Oldenburg (1998); Rosas-Carbajal et al. (2013)). For the velocity data, we assume uncorrelated and normally distributed errors. In this situation, the misfit takes the following form:

\[
\phi = -\frac{1}{2} \sum_{i=1}^{N} \left( \frac{g_i(m) - d_i(m)}{s_i} \right)^2.
\] (3)

For each data set, \( g \) is the solution of a particular forward problem for model \( m \), \( N \) is the total number of data and \( s_i \) denotes the standard deviation for the \( i \)-th data error.

Posterior PDF of the data and parameters is commonly approximated using sampling-based Markov chain Monte Carlo (MCMC) algorithms (Gilks et al., 1995). In our joint inversions of independent data sets, we use the Delayed Rejection Adaptive Metropolis (DRAM) scheme of Haario et al. (2006) in combination with the Cascaded Metropolis (CM) approach (Tarantola, 2005; Hassani & Renaudin, 2013; Manassero et al., 2021). Details about the general inversion framework (RB+MCMC) are given in Manassero et al. (2021) while particular details about the sampling strategy, prior information, and the initial seismic velocity and electrical conductivity models used for the current inversion are given in Section S1 of the Supplementary Material.

In Figure 2, we show examples of both MT and seismic velocity data, together with the posterior PDF over data after the joint probabilistic inversion. The color scale within the figures indicates the likelihood of the models effectively explaining the data (increasing from black to yellow). As expected in probabilistic inversions using MT data, model variability increases with depth, i.e. longer periods (see results in, e.g., Rosas-Carbajal et al., 2013; Manassero et al., 2020). In general, we observe poorer fits for the diagonal components compared to the off-diagonal components. Further discussion on the data fitting is provided in the Supplementary Material, along with additional figures.

### 3.3 Model parameters

In order to define the model parameters and their interdependence in the joint inversion, we distinguish between the primary and secondary parameters (Khan et al., 2006; Afonso et al., 2013a; Manassero et al., 2021). The latter are directly linked to the properties used to solve forward problems in their classic forms (e.g., \( V_p \) and electrical conductivity). The former are the fundamental thermodynamic parameters, namely, temperature (T), pressure (P) and bulk composition (C). These control the magnitudes of the secondary parameters in the mantle via equations of state and thermodynamic constraints (this applies to the mantle only; for crustal parameters see below). A specific configuration of the primary parameters in the 3D space defines what we refer to as the *background state* (or background contribution). In this way, the P-wave velocity and electrical conductivity in the mantle, defined by the background state, can be written as \( V_p(T, P, C) \) and \( \sigma_b(T, P, C) \).

As shown by Afonso et al. (2013a, 2013b) and Manassero et al. (2021), an efficient way to parameterize the background state is to divide the 3D space into \( m \) rectangular columns (Figure 3.a) and use the following model parameters in each column (Figure 3.b): i) \( n \)
Figure 2. I. Posterior PDFs of MT data for station L4R at 142.51 E, 33.993 S. Field data and error bars are plotted in blue. Panels (a), (b), (c) and (d): Posterior PDFs of the real and imaginary parts of the off-diagonal components ($Z_{xy}$ and $Z_{yx}$). Panels (e), (f), (g) and (h): Posterior PDFs of the real and imaginary parts of the diagonal components ($Z_{xx}$ and $Z_{yy}$). The data has been scaled by the square-root of the period (T) in all panels. We note that most data errors provided by the impedance estimation process (Kirkby et al., 2020) for this station are smaller than the error floors, hence the expected variability present in random errors is not seen. The location of the station is shown in Figure 1f. II. Posterior PDFs of P-wave velocity data for stations (a) ST15 located at 139.71 E, 32.74 S and (b) ST182 at 144.4 E, 33.20 S. P-wave velocity data and error bars are plotted in blue. For those locations, the LAB depths corresponding to the mean, lower and upper bound of the 68% CI models are shown in solid and dashed gray lines, respectively. In all panels the likelihood of the models to effectively explain the data increases from black to yellow.
‘thermodynamic nodes’ distributed throughout the mantle, and ii) the depth to the thermal
lithosphere-asthenosphere boundary (LAB). The depth of the LAB can be defined using
different proxies depending on the type of observations, such as seismic velocities, electrical
conductivity, petrogenetic, rheologic or thermal properties (Eaton et al., 2009; Fischer et al.,
2010; Birkey et al., 2021, and references therein). The thermal definition of the LAB, as the
boundary between the conductive lithosphere and the convective asthenosphere (Turcotte
& Schubert, 2014), allows for practical modeling and estimation of thermal structures in
the lithosphere (Jaupart & Mareschal, 2010; Hasterok & Chapman, 2011). The thermal
definition of the lithosphere is also directly related to the strength of the rock and the
seismic or electrical definitions through equations of state (Afonso, Moorkamp, & Fullea,
2016, and references therein). Given these advantages, we adopt the thermal LAB definition
from Afonso, Rawlinson, et al. (2016) as the depth at which the isotherm reaches 1250°C.
The LAB depths allow us to solve for a lithospheric conductive temperature profile, while
the temperatures of the thermodynamic nodes placed in the sub-lithospheric mantle are
allowed to vary during the inversion as required by the inverted data. The computation
of the pressure (P) and definition of the bulk composition (C) are described below and in
Section 3.5, respectively.

Since electrical conductivity is also highly sensitive to factors other than T, P, and
C (e.g., hydrogen content, localized fluid/melt pathways, presence of hydrated phases or
graphite), we expand the space of secondary parameters and write \( \sigma = \sigma_0(T, P, C) + \sigma(X) \),
where X stands for any factor that cannot be captured by the background. This means
that \( \sigma(X) \) is a representation of any anomalous conductivity associated with processes su-
perimposed on the background state (Manassero et al., 2021). This anomalous conductivity
contribution and the conductivity in the crust (\( \sigma_c(X) \)) are parameterized with l conductivity
nodes (blue dots in Figure 3.c) distributed throughout the whole domain (see details in
Section S3 in Supplementary Material and in Manassero et al., 2021).

To parameterize the rest of the properties of the crust, we divide the crust into three
layers (sediments, upper crust and lower crust) from the surface to the Moho (see Figure 3.a).
Within each column, layers have fixed thicknesses and their own set of physical properties:
thermal conductivity \( k \), volumetric radiogenic heat production (RHP) and P-wave \( V_p \)
velocity. During the inversion, only \( V_p \) is allow to vary within their assigned uncertainties
(Rawlinson et al., 2016); all remaining parameters are assumed constant. The thermal
conductivity of the crustal layers are set to \( k^1=2.8, k^2=2.6 \) and \( k^3=2.3 \) W/m°C. The crustal
RHP is obtained using the mean crustal RHP from a previous 1D joint probabilistic inversion
(Figure 1.e, Haynes et al., 2020) while the Moho depths are taken from the regional AusREM
model (Figure 1.d, Kennett & Salmon, 2012). We also incorporate one conductivity cell
below each MT station (Figure 3.c) as extra parameters to account for the galvanic distortion
effect produced by near-surface inhomogeneities beyond the resolution of our model (Jones,
2011; Chave & Jones, 2012; Avdeeva et al., 2015). Similarly to the approach used in ModEM
(Kelbert et al., 2014), these cells are placed in the first (thin) layer of the numerical mesh
used to solve the MT forward problem.

3.4 Forward Problems and Model Discretization

The main forward problems solved during the probabilistic inversion are the 3D MT
problem and the conductive heat transfer in the lithosphere. The MT forward problem used
here (Zyserman & Santos, 2000; Manassero et al., 2020) follows the secondary field formul-
ation of Douglas Jr et al. (1999, 2000), where the electromagnetic fields are considered as the
superposition of primary fields induced by a layered Earth and secondary field generated
by the conductivity distribution over the 1D Earth. The absorbent boundary conditions of
our formulation (Sheen, 1997; Manassero, 2019; Manassero et al., 2021) permit the use of
a small number of padding cells at the domain boundaries, thereby reducing the mesh size
without compromising the solution. Our formulation, boundary conditions and numerical
solution of the forward problem using finite elements (FE) are described in detail in Man-
Figure 3. Discretization of the 3D volume and parameterizations used for the background properties and conductivity anomalies (modified from Afonso, Rawlinson, et al. (2016)). The coarse discretization is represented by 1D columns. Each column comprises three crustal layers, two mantle compositional layers, LAB and 8 mantle thermodynamic nodes (orange dots) where the energy minimization problem is solved. The intermediate discretization (Figure 3.c and grey lines in 3.b) comprises the finite elements (FE) used to solve the MT forward problem. The finest discretization (red points along the geotherm in 3.b) is used to solve the steady-state heat transfer equation within the lithosphere. The background seismic velocities ($V_p(T,P,C)$) and electrical conductivities ($\sigma_b(T,P,C)$) obtained at the thermodynamic nodes are linearly interpolated to the fine scale mesh to create the seismic velocity model and to the FE mesh for the computation of the conductivity model and the MT forward solution, respectively. The anomalous conductivity distribution in the mantle ($\sigma(X)$) and the conductivity in the crust ($\sigma_c(X)$) are parameterized with conductivity nodes (blue dots in Figure 3.c). The inversion parameters are: the depths of the LAB, the temperature of the sublithospheric thermodynamic nodes, $V_p$ of the crustal layers, the value of the conductivity nodes and cells below each MT station (Figure 3.c).
assero (2019) and Manassero et al. (2021). Details on the conductive heat transfer modeling can be found in Afonso et al. (2013a, 2013b) and in Afonso, Rawlinson, et al. (2016). In what follows, we focus on the model discretization and the derivation of the seismic velocity and background conductivity models, given a realization of the primary parameters.

1. Coarser discretization: The study area is subdivided into 441 columns of size $0.45^\circ \times 0.45^\circ \times 410$ km (Figure 3.a). Each column (Figure 3.b) is subdivided in crustal layers and mantle thermodynamic nodes, placed every 50 km in the vertical direction. These nodes are used to obtain stable mineral assemblages and physical properties in the mantle by Gibbs free-energy minimization (Afonso et al., 2013b).

2. The intermediate discretization (Figure 3.c) comprises the finite elements (FE) used to solve the MT forward problem. The entire conductivity model is discretized into a total of $63 \times 63 \times 36$ FE (or cells), each with a horizontal size of $17 \times 17$ km and variable vertical dimensions, reaching a minimum thickness of 100 m at the top. The overall dimensions of the domain are $1071 \times 1071 \times 611.9$ km. The model includes three padding cells at the boundaries of the domain and four layers of air (with a total air thickness of 106 km). As part of the second field formulation, we use a layered Earth with 100 $\Omega$ m in the mantle up to 410 km and 10 $\Omega$ m below that, reflecting the conductivities of the transition zone (Grayver et al., 2017). These behave as boundary conditions for the model.

3. A fine mesh (2 km, red dashed line in Figure 3.b) is used to solve the steady-state heat transfer equation within the lithosphere (via a FE algorithm), subject to Dirichlet boundary conditions at the LAB ($T_{LAB} = 1250 ^\circ C$) and at the model’s surface ($T_S = 10 ^\circ C$) (Afonso et al., 2013b).

During the probabilistic inversion, a realization of the background parameters includes a specific LAB depth and temperatures for all the sub-lithospheric thermodynamic nodes in the entire domain, both randomly sampled from their prior distributions. After solving for the conductive geotherm corresponding to the sampled LAB depth, we interpolate the temperatures to the lithospheric thermodynamic nodes (i.e., those thermodynamic nodes that reside inside the lithosphere). The pressure is computed at all thermodynamic nodes using a quadratic lithostatic-type approximation (see Section S2 of Supplementary Material). Using these T, P and a pre-defined composition, we retrieve all thermo-physical properties at the thermodynamic nodes from pre-computed tables. These tables are calculated by Gibbs free-energy minimization with components of the software PerpleX (Connolly, 2009; Afonso et al., 2013b) and the database and thermodynamic formalism of Stixrude & Lithgow-Bertelloni (2011), within the CFMAS system ($CaO, FeO, MgO, Al_2O_3, SiO_2$). All thermophysical properties computed at the thermodynamic nodes are linearly interpolated to the fine mesh to create the corresponding seismic velocity model, and to the FE mesh for the computation of the conductivity model (see details in Section S2) and the MT forward solution.

### 3.5 Mantle composition

The pre-computed tables and their equilibrium assemblages are computed using a mean bulk mantle composition (i.e., specific CFMAS compositions) of 44.3 wt% $SiO_2$, 2.8 wt% $Al_2O_3$, 8.5 wt% $Fe_2O_3$, 39.3 wt% $MgO$ and 2.7 wt% $CaO$. We estimate this mean composition by averaging eight spinel lherzolites xenoliths (see Table S2 in Supplementary material) that were entrained in EAVP lavas. We use major element compositions from Irving (1980), O’Reilly & Griffin (1987), Griffin et al. (1987) and unreported samples from Bokhara River, which cover the area of interest. Since this is the most recent volcanism in eastern Australia, these xenoliths are the most representative samples of current mantle compositions available.
The use of an average mantle composition is justified by the fact that $V_p$ and electrical conductivity have second-order sensitivity to (dry) bulk mantle composition (see Figure S1, S2 and Özaydın & Selway, 2020; Trampert et al., 2001; Goes et al., 2000). We also assume a dry mantle composition for the background properties. The reasons for this choice are: i) $V_p$ is not significantly affected by the small amounts of water commonly observed in mantle samples (Yu et al., 2011; Cline II et al., 2018, and references therein), and ii) the conductivity nodes can represent any positive anomalies (e.g., water content) over the background values (which represents the most resistive end-member at the given T-P-C conditions), this reduces the number of parameters by two (see Manassero et al., 2021).

3.6 Mantle water content as a proxy for metasomatism

Using outputs of the joint probabilistic inversion (thermal structure, conductivity models, equilibrium assemblages and mineral compositions), we can estimate the bulk water content in the mantle (i.e. hydroxyl or $OH^-$ bound to nominally anhydrous minerals) that would be required to explain the inversion results. Importantly, the water content as estimated here lumps all unmodeled chemical effects resulting in high conductivity (e.g. connected sulfides, presence of melt) and it is taken as a proxy for mantle metasomatism (see Discussion).

The water content computations are done using the software MATE (Özaydın & Selway, 2020), which includes several experimental models for electrical conductivity, water partitioning and solubility (based on petrological studies). In particular, we used the electrical conductivity models of Gardès et al. (2014), Dai & Karato (2009a), Liu et al. (2019), and Dai & Karato (2009b) for olivine, orthopyroxene, clinopyroxene and garnet, respectively.

The solutions for the water content lie between the bounds defined by the dry lithosphere (i.e., 0 ppm) and the maximum bulk water content calculated using the olivine water solubility model of Padrón-Navarta & Hermann (2017). The experimental coefficients used in the water partitioning are: $D^{OH}_{opx/ol} = 5.6$, $D^{OH}_{cpx/opx} = 1.9$ of Demouchy et al. (2017) and $D^{OH}_{gt/ol} = 0.8$ of Novella et al. (2014); which reflect the sub-solidus conditions found in the continental lithospheric mantle in southeastern Australia. Since we aim to portray variations of water content in the mantle rather than fitting the real water content seen in xenoliths, the choice of the experimental parameters is adequate for our calculations. All water calculations are done using the calibration of Withers et al. (2012) for olivine, and the calibration of Bell et al. (1995) for pyroxenes and garnet.

The electrical conductivity of each individual mineral phase is turned into bulk conductivity through the Generalised Archie’s Law (Glover, 2010) with cementation components ($m$) of $m = 2$ for orthopyroxenes, $m = 4$ for clinopyroxenes and garnet, and $m < 1$ for olivine (perfectly connected). The Generalised Archie’s Law is preferred over the conservative estimates of Hashin-Shtrikman lower-bound since it allows us to incorporate the effects of specific minerals in the conductivity values, such as highly-interconnected phlogopites. The main cementation components used here provide values similar to the Hashin-Shtrikman lower-bound for a lherzolitic matrix (Özaydın & Selway, 2020).

4 Results

4.1 Thermal structure of Southeast Australia

The depth to the LAB obtained from the joint probabilistic inversion of seismic velocities and MT data is shown in Figure 4; the complete 3D temperature structure is shown via depth slices in Figure 5 (first three columns). Figure 4 also includes a recent LAB model obtained from a low-resolution 1D joint probabilistic inversion of elevation, surface heat flow, Rayleigh wave dispersion curves, and geoid anomalies (Haynes et al., 2020; Afonso et al., 2013b) and the estimated LAB depths from two recent seismic tomography models.
in eastern Australia (Davies et al., 2015; Rawlinson et al., 2017). The maximum absolute difference in lithosphere thickness inside the region is ~200 km, with shallow LAB depths (<100 km) in the eastern and southern ends of the continental block and deep LAB depths (>250 km) beneath the eastern part of the Curnamona Province (CP) and the northern part of the Delamerian Orogen. We observe a clear correlation between lithospheric structure and the locations of recent volcanism (e.g., Figure 4.a): leucitite volcanic centers correlate with regions of intermediate lithospheric thickness (125-160 km) while the basaltic volcanoes are located in regions where the LAB is shallower than 120 km. At least two clear and step-like changes in LAB depth are observed in our model along a transect from the CP to the southeast corner of the model and across the leucitite volcanoes (see also Figures 11).

The first order features of our LAB model are in good agreement with the mean LAB obtained by Haynes et al. (2020), even though the data sets used in each inversion are different. Many features in our LAB model are also present in those derived from seismic velocity models (Davies et al., 2015; Rawlinson et al., 2017). In particular, we observe similar LAB depths beneath the basaltic volcanoes and west of 146°E, where a wedge-like structure follows the curvature of the Stawell Zone (SZ, see Figure 3.1). All models show a thickening of the lithosphere towards the northwest part of the region. However, some significant discrepancies are found in the CP and towards the center of the model. Beneath the CP, our LAB depths are considerably deeper (>70 km) than those of Davies et al. (2015) and Rawlinson et al. (2017). Although some of this difference could be attributed to the fact that the mantle composition in this area is likely more depleted (and thus ‘faster’) than the average bulk mantle composition used here (see Section 3.5), additional calculations shown in Figure S3.a of the Supplementary Material reject this possibility as the main cause. Rather, the main reason is the different definitions of LAB adopted in these works. The seismic LAB is commonly defined as the transition from a faster lithospheric lid to the lower-velocity, convecting asthenosphere (Fischer et al., 2010; Eaton et al., 2009). While in shallow lithospheric environments there is a marked minimum in \( V_p \) near the LAB (and thus easy to pick), deep lithospheric environments are characterized by smooth \( V_p \) profiles, making it harder to choose the thermal LAB unambiguously (Figures S3.b and 2. II) based only on \( V_p \) profiles. Therefore, this explains why our LAB estimates are similar to those in Davies et al. (2015) and Rawlinson et al. (2017) in thin lithospheric settings, but diverge in absolute magnitude as the LAB gets thicker. We also note that the limitations in vertical resolution within body-wave tomography often lead to significant uncertainties in the LAB depth (Tan & Helmberger, 2007; Fischer et al., 2010). A method commonly used to resolve this ambiguity is the joint interpretation of surface-wave models with S-receiver functions (Eaton et al., 2009; Fischer et al., 2010). This method offers good sensitivity to the LAB depth and mid-lithospheric discontinuities (MLDs). In the first order, our LAB estimates largely agree with the LAB depths derived using S-receiver functions in southeastern Australia (Kumar et al., 2007; Ford et al., 2010; Birkey et al., 2021). The main difference is observed beneath the CP, where we estimate a deeper LAB than those estimated by Kumar et al. (2007) and Birkey et al. (2021) (~70 km and ~100 km, respectively). In addition, Birkey et al. (2021) suggest the presence of an MLD in this area. See the discussion in 5.1.3.

The depths to the thermal LABs obtained after an RB+MCMC inversion using MT data alone (Figure S5 in Supplementary Material) are shown in Figure S6 (Supplementary Material). Compared to the results from the joint inversion, these figures show large variability and no clear trend in the thickness of the lithosphere from the CP to the southeast corner of the model (a feature observed in all other models). This comparison illustrates well the facts that i) MT alone has difficulties in discriminating thermal causes from other factors controlling the electrical conductivity in the mantle (Jones, 1999) and ii) other types of data (e.g. seismic) should be included when imaging lithospheric structure.
Figure 4. Depth of the thermal LAB. (a) Mean model after the joint probabilistic inversion; (b) mean model obtained after a 1D joint probabilistic inversion (Haynes et al., 2020; Afonso et al., 2013b); (c) and (d) lower and upper bounds of the 68% confidence interval (1 standard deviation from the mean), respectively. (e) and (f) depth of the LAB after Rawlinson et al. (2017) and Davies et al. (2015), respectively. The location of leucite-bearing volcanism are shown in blue and standard basaltic volcanoes in grey. The 140 km-contour of the LAB depth is shown in dashed-grey line and the outline of the tectonic provinces in solid grey lines. The location of the Stawell Zone (SZ) is marked in panel (a).
Figure 5. Columns (1)-(3): depth slices from the (1) mean model and those models corresponding to (2) the lower and (3) upper bound of the 68% CI of the posterior PDF for the temperature. Columns (4)-(6): depth slices from the models corresponding to (4) the lower and (5) upper bound of the 68% CI, and (6) mean of the posterior PDF for the P-wave velocity; Column (7): P-wave velocity model of (Rawlinson et al., 2016). Selected depths are shown on the left of the figure. In all cases, velocities are plotted relative to 1-D reference model AusREM at 34.4°S, 145°E shown in Figure S4 of the Supplementary Material.
4.2 Seismic velocity structure

Depth slices of the P-wave velocity structure predicted by our model are shown in Figure 5. The P-wave velocity model of Rawlinson et al. (2016) is also shown for reference. In all cases, the velocities are plotted relative to the AusREM model at 34.4°S, 145°E (Figure S4 in Supplementary Material). We observe that the inversion succeeded in reproducing the Vp structure of the input model (Rawlinson et al., 2016). In particular, the mean P-wave velocity down to 100 km is practically identical in both models. The Newer Volcanic province stands out as a low-velocity anomaly at depths between 60 and 80 km, whereas basaltic volcanoes in the middle of the Eastern Province (~149°E, 34°S) correlate well with deeper low-velocity anomalies.

Some minor discrepancies between our results and the model of Rawlinson et al. (2016) are observed at depths > 100 km. For example, we obtain slightly higher seismic velocities (0.6% higher on average) in the depth range 100-180 km at the eastern end of the model. Supplementary tests allow us to attribute these differences to the constraints imposed by the MT in the joint inversion. Similarly, we obtain slightly slower velocities throughout the whole model at 200-220 km depth (see Figure 5). At these depths, the local discrepancies are a consequence of the different physical parameterizations used in this work and by Rawlinson et al. (2016). However, neither the original tomography model of Rawlinson et al. (2016) nor our model has sufficient resolution at these depths to justify further comparisons.

4.3 Electrical conductivity structure

The conductivity models for the crust and mantle predicted by the joint inversion are shown in Figures 6 and 7, respectively. For comparison, these figures include the results obtained from a recent deterministic inversion of MT data (Kirkby et al., 2020), using the ModEM software (Kelbert et al., 2014). The main structures observed in the conductivity models are comparable (within the uncertainties of the model) to those of the model of Kirkby et al. (2020) at all depths.

4.3.1 Crust

Figure 8 illustrates the agreement between the conductivity structure in the crust and other sources of information: sedimentary basins (Raymond et al., 2012), magnetic anomalies (Nakamura & Milligan, 2015) and a shear velocity model (Pilia et al., 2015). In particular, Panel (a) shows that the extent of the Paleozoic to Cenozoic sedimentary basins in the region is well outlined by the mean conductivity model at 2 km depth. A comparison between Panels (c,e,f) and Panel (b) highlights the correlation between total magnetic anomalies and conductivity features at different depths. Examples of these are (A) a conductor in the CP; (B) a SW-NE linear structure close to the NW limit of Murray Basin; (B') and (B'') a NW-NS linear structure and a conductor within Murray Basin; conductors (C) and (D) in the Tabberabbera Zone; (E) a N-S conductor aligned with the western border of the Eastern Province; (F) two resistive structures west of the Sydney Basin; (G) the Sydney Basin; (H) a conductive region aligned with the north limit of the Northwest and Central NSW provinces; (I) a highly resistive region in the Stawell Zone near the NSW-VIC border; (J) a circular structure in the middle of the model; (K) a high-conductivity anomaly; and (L) a conductor east of the CP.

The conductor (A) correlates well with the conductor seen in the MT study of Robertson et al. (2016), using data from 74 AusLAMP stations placed in the Ikara–Flinders Ranges and CP, and in the recent study of Kay et al. (2022), using a densely spaced MT modeling scheme. We note that we have stations only on the east of this conductor and its extension cannot be well defined by our inversion. Comparing Panels (c) and (d), we observe that the main conductivity structures correlate well with the velocity anomalies imaged by the shear-wave velocity model of Pilia et al. (2015). We note that concentric geometries at 29
km depth, such as conductor (J) and structures west of the model, resemble the features of the Lachlan Orocline model revealed by potential field and passive seismic data (c.f. Kirkby et al., 2020).

### 4.3.2 Mantle

The conductivity models between 80-250 km depth (Figures 7) largely resemble the ModEM model of Kirkby et al. (2020). In particular, we observe a similar north-eastward orientation of the conductors in the middle of the model (C1, C2, C3, C4 and C5 in Figures 7 and 9). When comparing mantle conductivity with the mean LAB structure in Figures 9, our models suggest that there is some correlation between LAB topography and these mantle conductors (cf. Kirkby et al., 2020). In particular, the general NE-SW trend of the conductors tends to follow the LAB depth structure. The conductor C5 aligns well with the LAB wedge northwest of the model, whereas C3N and C4S tend to follow the 120-140 km LAB depth iso-surfaces. C1, C2 and C3S are located in regions where the LAB depths < 120 km. We also observe a high-conductivity intra-lithosphere structure beneath the CP (C6), which agrees well with the structure imaged by previous MT studies (e.g., Robertson et al., 2016; Thiel & Heinson, 2013). A high-conductivity region (C4) is observed below the central-leucitite volcanoes. In our models, the extent of this region is larger and more connected than in the ModEM model.

The main difference between our conductivity model and that of Kirkby et al. (2020) is that the sub-lithospheric conductivities along the south-east coast and in the middle of the region are higher in our model (R1 and R2 in Figure 7). The same is true when we compare our model with the results of a probabilistic inversion of MT data only (Figure S5, Supplementary Material). The high resistivity (> 10^4 Ωm) values in MT-only inversions are at odds with the mantle resistivity range obtained for sub-lithospheric temperatures and pressures (Fullea et al., 2011; Naif et al., 2021). We observe that, due to the constraint imposed by seismic data in the joint probabilistic inversion (via the thermal structure), unrealistically high resistivity values are not present in our model.

Another example of the constraint imposed by seismic data in the conductivity models is shown in Figures 9. At 140 km depth, we observe that the conductors in the east (C1, C2 and C3S in Figure 9.c) are located within a region defined by a 1250°C-contour (Figure 9.b). These mantle conductivity structures correlate well with the location of the eastern basaltic volcanics and a stripe of low P-wave seismic velocities (Figure 9.d). At the same time, the stripe of high seismic velocities beneath the east coast at a depth of 140 km is a clear example of the constraint imposed by the MT data in the velocity models. This stripe is not seen in the models of Rawlinson et al. (2016) and correlates with a relatively cold and highly resistive mantle (Figures 9.b-c).

The plausibility of small and narrow features at depths greater than 100 km should be taken with caution. Although our probabilistic inversion provides a range of possible structures and conductivity values, it is worth noting that they all fall within the same parameterization. In 3D inversions of continental-scale data sets, small-scale structures can be easily moved or removed by the sampling process without compromising the model fit.

### 4.4 Joint assessment of bulk water content and temperature maps

The bulk water content maps derived from the mantle conductivity models are shown in Figures 10. As mentioned in Section 3.6, we emphasize that “bulk water content” is a lumped proxy for general mantle metasomatism and therefore their absolute values must be taken with caution. We observe that most of the localized conductive anomalies above the background require relatively high bulk water contents. This is the case for the following structures (Figure 10): C1, C2 and C3 beneath the eastern basaltic volcanoes; C4 below the
Figure 6. Conductivity in the crust from the joint probabilistic inversion. Columns (1)-(3): depth slices from the (1) lower, (2) upper bound of the 68% percentile and (3) mean conductivity models of the posterior PDF. Column (4): conductivity model of (Kirkby et al., 2020). Selected depths are shown on the left of the figure and the boundaries of geological provinces are shown in grey lines.
Figure 7. Mantle conductivity from the joint probabilistic inversion. Columns (1)-(3): depth slices from the (1) lower, (2) upper bound of the 68% percentile and (3) mean conductivity models of the posterior PDF. Column (4): conductivity model of Kirkby et al. (2020). The location of leucitite-bearing volcanism are shown in blue and standard basaltic volcanoes in grey. Selected depths are shown on the left of the figure. Dashed-black lines highlight conductors in the mean model and resistors in the ModEM model below 123 km depth.
Figure 8. (a) Sedimentary basins overlying mean conductivity model at 2 km depth and 200 Ωm-resistivity contour in dash lines. (b) Total magnetic anomalies after Nakamura & Milligan (2015) (c) Mean conductivity (overlying the magnetic anomalies in grey scale) and (d) shear wave velocity model after Pilia et al. (2015) at 2 km depth. (e-f) Mean conductivity models at 12 and 29 km depth overlying the magnetic anomalies in grey scale. We refer the reader to the main text for a description of structures A-L. Boundaries of geological provinces are shown in grey lines.
Figure 9. (a) Mean LAB depth. Contours of the LAB depth every 20 km are shown in grey-dashed line. Mean models at 140 km of (b) temperature (c) electrical conductivity and (d) P-wave velocity relative to 1-D reference model AusREM at 34.4°S, 145°E. The 1250°C-contour (corresponding with the thermal LAB) is plotted in dashed-black in (b-d). Panel (c) shows the location of the geological provinces and conductors in dashed blue. The location of leucitite volcanoes are shown in blue triangles and the surface outcrop of basaltic volcanics are shown in grey in all panels. Panel (b) shows five transects which are discussed in section 5.1.
central leucitites; \( C_5 \) on the eastern boundary of Delamerian Orogen; \( C_6 \) beneath the CP; and the deep localized conductor \( C_7 \) at \( \sim -30.5^\circ N, 147^\circ E \), beneath the northern leucitites.

Figures 11-12.I and Figure S7 (Supplementary material) show vertical slices of the conductivity, water content, and temperature along the four transects depicted in Panel (b) of Figure 9. The transects in Figure 11.I cross most of the geological provinces on the west and demonstrate a striking correlation between known geological boundaries and the alternation between wet/dry portions of the lithosphere. The joint assessment of these transects clearly shows that the lithospheric mantle beneath the CP (\( C_6 \)) corresponds to a highly conductive, hydrated, and cold region. We observe a high-conductivity anomaly (\( C_5 \)) below the Stawell Zone that crosses the LAB. Although the high temperatures found in this region (\( T_2 \) in Figure 11.Ic) can partially explain its conductivity structure, Figure 11.Ib indicates that a large part of this anomaly is related to metasomatism (or incipient melting?). The high-conductivities observed in region \( C_{3N} \) (beneath Tabberabbera Zone) and the conductor \( C_{NV1} \) (at \( \sim 90 \) km depth beneath the NV) can be entirely explained by a relatively large water content. We observe that, while the conductivity of \( C_{3S} \) at \( \sim 200 \) km can be explained by the high anomalous temperatures found in that region (\( T_1 \)), a substantial part of its conductivity at \( \sim 150 \) km is explained by the presence of water (or melt?). Two shallow conductors \( C_{NV2} \) and \( C_{NV3} \) are found at \( \sim 20-75 \) km depth beneath the NV.

Along the transects in Figure 11.II, the LAB shows a small perturbation over a large conductive anomaly at \( \sim 50 \) km and two defined steps at \( \sim 300 \) and \( \sim 750 \) km. These features correlate with the location of the northern leucitites, central leucitites and basaltic volcanoes, respectively. They also correlate with the location of high-conductivity regions (\( C_7, C_4 \) and \( C_{3S} \)) in the sub-lithosphere. When comparing the conductivity and the bulk water content along this transect, we observe that while the water contents at \( C_7 \) and \( C_4 \) are relatively large, that of \( C_{3S} \) is considerably lower. A series of crustal conductors are also observed beneath both the leucitites and basaltic volcanoes.

Figure 12.I, which traverses the eastern basaltic volcanoes and the NV, shows the continuation of \( C_{3S}, C_{NV2} \) and \( T_1 \) beneath the NV. A high-conductivity and wet region (\( C_1 \)) is observed in the sub-lithospheric mantle below the Eastern Province. This deep and wet structure is also seen in Figure S7 and is correlated with a high temperature anomaly (\( T_3 \)). A shallow semi-hydrated structure (\( C_{EP1} \)) is observed right below the basaltic volcanoes. The relationship between these features is hard to reconcile, unless we relax the assumption that the entire conductivity anomaly over the background is due purely to water content. We discuss this further in the next section.

5 Discussion

5.1 Mantle metasomatism and volcanism in southeast Australia

Mantle metasomatism occurs when incipient melts or fluids react with mantle rocks (predominantly peridotite). These reactions can i) affect the modal proportions of peridotites, ii) introduce new volatile-bearing phases (phlogopite, amphibole, apatite, and carbonates) and, in some pervasive cases, iii) create new lithological domains, such as small portions of pyroxenite ± volatile-bearing phase lithologies (e.g. O’Reilly & Griffin, 1987). The generation of volatile-bearing phases reduces the solidus temperature (Wallace & Green, 1988; Foley et al., 2009; Pintér et al., 2021) and increases the electrical conductivity of the mantle domain (Selway, 2014).

The bulk water content we report in this study acts as a general proxy for metasomatism or mantle fertility, i.e., the inclusion of phases (metasomes) that increase the electrical conductivity of the mantle. Therefore, this proxy lumps together a number of factors not explicitly modeled in this work, for example, i) the presence of phases such as graphite or sulphides for depths above 75-120 km (Selway, 2014; Özaydın & Selway, 2020); ii) co-
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![Bulk Water Content](image)

**Figure 10.** Caption on the next page.
Figure 10. Bulk water content and mantle conductivity models from the joint probabilistic inversion. Columns (1)-(3): water content maps obtained from the (1) the lower, (2) upper bound of the 68% CI and (3) mean conductivity models. Column (4): depth slices from mean conductivity models of the posterior PDF. The location of leucitite-bearing volcanism and basaltic volcanoes are shown in orange and turquoise in (3); and blue and in grey in (4). Selected depths are shown on the left of the figure. We refer to the main text for an explanation of structures C1-C7. Selected depths are shown on the left of the figure.

existing water and phlogopite in cold mantle below 75-120 km depth; or iii) presence of melt in regions of elevated temperatures.

The results of Section 4.4 indicate widespread mantle metasomatism in southeastern Australia (Figures 10-12.1). Clear correlations are observed between the location of volcanic centers and regions of metasomatized mantle and conductive crust. In particular, these regions are i) C3S, CNV1, CNV2 and CNV3 below the Newer Volcanics; ii) C1 and CEP1 beneath the Eastern Volcanics (also C4, CEP2 and CEP3 in Figure S7); iii) C4 beneath Central Leucitites; and iv) C7 below Northern Leucitites. Xenoliths entrained in these lavas show strong evidence of metasomatism at the source (Yaxley et al., 1991, 1998; Shea et al., 2022), further validating the presence of metasomatic agents in these regions (Frey et al., 1978). Other metasomatized mantle regions show a strong link with subduction/accretion processes rather than volcanism. These are C6, C5 and C3N beneath CP, Stawell Zone and Taberraberra Zone, respectively.

One of the key benefits of our inversion is that we can dissociate the effects of the temperature from other factors controlling the conductivity structures and, ultimately, map anomalies associated with mantle metasomatism. While the current state of our methodology does not allow us to discriminate the different metasomatic factors (presence of melt, water, or phlogopite, for example), these can be inferred via the joint assessment of the conductivity, temperature, and metasomatism models in different regions. Further analysis with contributions from melt modeling may be required to understand the full scope behind metasomatism and the genesis of melts in southeast Australia. This work is left for a forthcoming publication.

5.1.1 Basaltic volcanoes

The existence of shallow mantle upwellings and a thin lithosphere provides a favorable setting for both mantle metasomatism and extensive decompression melting to take place (Aivazpourporgou et al., 2015). The LAB is very shallow in the Newer Volcanics (NV) and Eastern Volcanics (EV), allowing mantle upwellings to reach depths at which decompression melting of peridotitic rocks in addition to melting of metasomatized domains is possible resulting in the primitive basaltic melts. We observe that the metasomatized regions C1S and C1 correlate with the location of sub-lithospheric high-temperature anomalies (T1 and T3, respectively), meaning metasomatized regions also contributed to the primitive basaltic melts. The extensive decompression melting of peridotite, driven by high temperatures, will dilute melts from the metasomatized regions resulting in the basaltic lavas seen in Eastern Australia. From the low velocities observed at 60-80 km beneath the NV (Figure 5), Rawlinson et al. (2017) interpreted T1 as a mantle upwelling and the source of the NV (see also Rawlinson et al., 2015). Similarly, Rawlinson et al. (2015) interpreted the low-velocity anomaly corresponding to T3 (Figures 5) as a deep mantle source for the EV. All of the above point to ideal conditions for the metasomatism to be linked with the generation of extensive basaltic magmatism in this region.
Figures 11.I and 12.I show a clear conductive pathway from $C_{3S}$ to $C_{NV1}$-$C_{NV3}$ and from $C_{1}$ to $C_{EP1}$, while the mantle beneath the volcanics is relatively dry. To further illustrate the relationship between the location of volcanics, shallow conductors and mantle metasomatism, Figure 12.II shows the average conductance at 20-50 km depth and the average water content near the LAB beneath the EV. According to these results, basalt fields tend to associate with shallow conductors (Figure 12.IIa) and dry mantle (Figure 12.IIb). This relationship indicates that metasomatic agents percolated from deep mantle sources and traversed toward the crust. On their ascent, they precipitated conductive minerals forming shallow conductors ($C_{NV2}$,$C_{NV3}$, $C_{EP1}$, $C_{EP2}$, $C_{EP3}$ in Figures 11.I, 12.I and S7). We also note that basalts tend to be located in the surroundings of the most metasomatized regions of the lithosphere rather than on top of them; something that has been observed also in kimberlites worldwide (Özaydın & Selway, 2022). The dry mantle regions beneath the basaltic fields may represent residual mantle that has already experienced a melting event, exhausting the region of metasomes and leaving behind a dry residuum. Alternatively, the basaltic melts could have been generated by metasomatized mantle in the vicinity of the basalt fields, but not directly below, and transported via oblique and complex lithospheric pathways (Tapu et al., 2023; Özaydın et al., 2024).
Figure 11. Caption on the next page.
Figure 11. I. Vertical slices along transect A-A’ (crossing most of the geological provinces) of a) the mean conductivity model, b) bulk water content derived from the mean conductivity mode and c) temperature. II. Vertical slices along transect D-A’ (crossing the leucitite volcanics and basaltic volcanics in the south) of a) the mean conductivity model, b) bulk water content derived from the mean conductivity mode and c) temperature. d) Intermediately connected ($m=2.5$ in blue) and poorly connected ($m=5$ in green) phlogopite in a dry lherzolitic matrix that fit the observed conductivities along the transect. The Moho and LAB depths along that transects are shown in dashed lines in all panels. The elevation and location of the geological provinces is shown at the top of the figures.

5.1.2 Leucitite volcanoes

The leucitite lavas have melt compositions comparable to lamproites and were derived from an atypical mantle assemblage of phlogopite-bearing pyroxenite (Shea et al., 2022; Foley et al., 2022). Due to these lava compositions, Kirkby et al. (2020) interpreted the conductors beneath the central leucitites as regions of metasomatized mantle with hydrous minerals such as phlogopite. Given the high conductivities ($<100\ \Omega\ m$) and bulk water content ($\sim 200$ ppm) observed around $C_4$ (Figures 11.II), our results indicate a high probability of the presence of volatile-bearing minerals, supporting the above interpretation.

Using the water calculation setup described in Section 3.6 and the phlogopite conductivity model of Li et al. (2017), we calculated the electrical conductivities of lherzolite with 5 and 10 % vol. of 0.52 w.t. fluorine-bearing phlogopite (average fluorine value of mantle rocks, Özaydın et al., 2022) for both perfectly connected ($m=1.1$, modified Archie’s law) and sparsely populated phlogopites ($m=6$, modified Archie’s law). The results show that perfectly connected cases are 2.5 orders of magnitude more conductive than the observed conductivities in the region, while the conductivity for sparsely populated/disconnected cases was near the lower bound of the observed conductivities (Figure S8 in Supplementary Material). These results suggest that a lherzolite with 5-10 % vol of partially connected ($6 < m < 1.1$) phlogopite explains the conductivities in $C_4$. We therefore interpret the high conductivities in $C_4$ as a phlogopite-bearing lherzolite with small percentages of partially to sparsely connected phlogopite. These results are also illustrated in Figure 11.II.d, which shows the percentage of partially connected ($m=2.5$ and $m=5$) phlogopite that can explain the observed conductivities beneath the leucitites volcanoes.

We note that the temperatures inferred at $C_4$ are close to the solidus for phlogopite (Thibault et al., 1992), suggesting that melt of phlogopite-bearing pyroxenite can be produced at these temperatures. This finding is consistent with the work of Byrnes et al. (2023), where the temperatures and composition of magmas at the Leucite Hills in Western United States, that are compositionally similar to the East Australian Leucitites, are likely sourced from melting of phlogopite-bearing mantle lithologies. The authors found that these volcanoes are located on the “warming front” of metasomatized lithosphere.

The northern leucitites sit above a high-conductivity and metasomatized region below the LAB ($C_7$). The ultrapotassic compositions of these lavas suggest low-degree partial melting (Cundari, 1973), which is consistent with the relatively cold temperatures found in the region. Furthermore, potassium-rich magmas are produced by melting a metasomatized mantle that has been enriched in phlogopite (Xu et al., 2017; Förster et al., 2019). We calculated the effect of phlogopites in this region and found similar results to those of the central leucitites (Figures S8), favoring the scenario with partially connected phlogopites. Compared with the central leucitites, the higher conductivities and colder temperature in this region provide favorable conditions for the presence of existing phlogopites that survived previous melting events.
5.1.3 Eastern Curnamona Province

Figure 11.I.a-b shows a successive alternation of conductive/wet and resistive/dry lithospheric domains that resemble the west-to-east subduction-accretion process in eastern Australia (Glen, 2005; Shea et al., 2022). The joint assessment of the fields depicted in Figure 11.I also suggests that the region C6 experienced pervasive mantle metasomatism. Given the cold temperatures, the lack of present magmatism, and the geological history of southeastern Australia, the metasomatic events C6 are likely related to accretion processes by which successive subduction and orogenic events introduced metasomatic agents into the mantle. Overtime, this process preferentially metasomatized the old, thick lithosphere beneath the eastern corner of the Curnamona province. The presence of hydrous minerals in this metasomatized region is likely to produce significant reductions in seismic velocity, leading to mid-lithospheric discontinuities (MLD, Selway et al., 2015; Byrnes et al., 2023). Results from S-receiver functions (Station STKA in Birkey et al., 2021) reveal the presence of an MLD beneath Curnamona Province. Our results indicate that this MLD is likely caused by low-velocity hydrous minerals in the region.

The eastern side of the crustal conductor (A) in CP (described in Section 4.3.1) can be seen in Figure 11.Ia. Kay et al. (2022) interpreted these shallow conductors as the deposition of interconnected graphite. However, given that graphite films are not stable at shallow depths (Zhang & Yoshino, 2017; Yoshino et al., 2018), we observe that a more feasible explanation for these crustal conductors are carbon-rich fluids sourced from the deep metasomatized region C6 (Thibault et al., 1992).

5.2 Implications for magma generation beneath eastern Australia

Age-progressive volcanism in the EAVP, particularly along the Cosgrove track (Davies et al., 2015), has been widely attributed to long-lived mantle plume activity (Wellman & McDougall, 1974). However, Shea et al. (2022) recently argued that primitive melt compositions throughout the EAVP (MgO < 15wt%, total alkali’s > 3wt%, and high TiO2 > 2 wt%), including all or most of the age-progressive volcanism, could be produced by melting a metasomatized mantle source at temperatures lower than those expected in a deep mantle plume. The petrological and geochemical evidence summarized by Shea et al. (2022) and in the previous sections indicates that primitive melts, particularly those associated with the leucitites, originated from a hydrous pyroxenitic component rather than from a peridotitic mantle lithology. Since the solidi of hydrous pyroxenites (Foley et al., 2022) is substantially lower than that of anhydrous peridotites (Walter, 1998), partial melting of such lithologies is possible with only slight perturbations above ambient upper mantle temperatures. The lack of high temperature plume-derived melts (high MgO basalts and low total alkali) in the EAVP, but widespread alkaline basaltic volcanism can be reconciled by stepped-edge processes driving mild temperature perturbations being commonplace throughout eastern Australia.

Our results show a series of steps in the LAB that correlate well with the location of basalts and leucite volcanoes (Figures 11.II and 13). Thermomechanical models have shown that such steps in the LAB constitute areas prone to generation of sublithospheric small-scale, edge-driven convection (EDC) instabilities and partial melting (e.g., Zlotnik et al., 2008; Van Wijk et al., 2010; Davies & Rawlinson, 2014; Ballmer et al., 2011; Afonso et al., 2008; Duvernay et al., 2021a). In particular, Duvernay et al. (2021b) showed that shear-driven upwelling (SDU) and EDC processes that account for water content in the upper mantle can produce enough melt to explain the total melt volume in the EAVP.

Considering all of the above, a plausible model arises for the melt generation in southeastern Australia, which we summarize in Figure 13. The relative motion of the Australian plate and the asthenospheric mantle created a favorable asthenospheric flow (possibly toward the south-southwest) which allowed the generation of EDC/SDU cells beneath steps in the LAB (e.g. above C4 and C5S). Such a convective flow can detach metasomatized
Figure 12. I. Vertical slices along transect B'-B'-B (across the basaltic eastern volcanoes and the NV) of a) the mean conductivity model, b) bulk water content derived from the mean conductivity model and c) temperature. The Moho and LAB depths along that transect are shown in dashed lines in all panels. The LAB depth along that transect is shown in dashed-black line in all panels. The elevation and location of the volcanics and the NV are shown at the top of the figure. II. Relationship between Cenozoic eastern volcanics and the parameters derived from the electrical conductivity model: (a) Conductance of the lower-crust (∼20–50 km), (b) water content calculated around the LAB depth (∼100–120 km).
lithologies from the lowermost portions of the lithosphere (on the thick side of the step) and drag them into the upwelling limbs of the EDC/SDU convection cell, where they would preferentially melt (given their lower solidus) and create the primary metasomatizing melts. The latter would subsequently react with the lower portions of the lithosphere (in the thin side of the LAB step) and create the source of the leucitite volcanoes. This scenario is similar to the one presented by Shea et al. (2022) and supported by our results and the abundant petrological evidence summarized by these authors.

If EDC/SDU cells are the main driver for the leucitite volcanism, they need to be able to produce melts for a relatively short period of time (non regenerative) to explain the punctuated nature of this volcanism. In fact, numerical simulations clearly show that ED/SD instabilities are ephemeral in nature and can be impeded by small perturbations in the sublithospheric flow (e.g., King & Anderson, 1995; Duvernay et al., 2021b, 2022). The presence of an anomalously hot upwelling or plume can also enhance or shut down melting near lithospheric steps (Mather et al., 2020; Duvernay et al., 2022), making it difficult to separate the two mechanisms without further knowledge of the underlying mantle dynamics. Thus, while our results provide a firm ground for interpretations on the origin of the EAVP and the roles that lithospheric structure and composition played, we cannot assess the deep mantle processes that controlled the large-scale sublithospheric flow.

6 Conclusions

We performed a joint probabilistic inversion of 3D magnetotellurics (MT) and seismic velocity data to constrain the lithospheric structure, metasomatic domains and melting processes in southeast Australia. Our methodology minimizes the non-uniqueness of the MT problem and provides quantitative information on model uncertainties via full posterior distributions. This information is crucial to assigning meaningful interpretations to electrical conductivity anomalies in terms of temperature versus metasomatism/compositional anomalies.

We image a highly heterogeneous lithosphere beneath eastern Australia that we link to geodynamic and tectono-magmatic processes across multiple scales. In particular, we detect widespread, but highly irregular mantle metasomatism throughout the region, pointing to complex interactions in the asthenosphere-lithosphere system. We also image alternating conductive/wet and resistive/dry lithospheric domains that correlate with the location of major geological provinces, resembling the west-to-east subduction-accretion process that formed eastern Australia. A series of steps in the present-day thermal structure correlate with the location of intra-plate volcanic centers and moderate thermal anomalies in the sublithospheric mantle, suggesting a genetic link between these three features. Basaltic volcanism is preferentially located in regions of very thin lithosphere and dry mantle, whereas leucitite volcanoes are located in regions of highly metasomatized mantle, intermediate lithospheric thickness, and localized lithospheric steps. These results, together with recent petrological and geochemical evidence for relatively low temperatures in the melting region (Shea et al., 2022), suggest that the interaction between a complex, metasomatized lithospheric structure and localized mantle upwellings (e.g. via edge-driven convection or focusing of moderately hot mantle upwellings) is probably responsible for much of the volcanism in the EAVP, rather than a deep, hot mantle plume.

Lastly, it is generally accepted that the metasomatized lithospheric mantle plays a critical role in the generation of major ore deposits. The ability to map metasomatized mantle domains opens up new avenues for exploring mineral resources.

Acknowledgments
Figure 13. 3D rendering views of the LAB (red surface), mean conductivity and temperature models depicting interactions between mantle metasomatism and steps in the LAB. Dashed black arrows show the flow of the asthenosphere and shearing of enriched mantle material into EDC-cells (circular dashed lines). Local hotspots along the LAB where enriched mantle material crosses its solidus in the sub-lithospheric mantle are indicated by grey blobs. Grey arrows indicate incipient melts that may travel across the LAB from deep to shallow portions of the lithospheric mantle. Leucitites and basaltic volcanic centers are shown in yellow triangles and blue circles, respectively. The Moho is shown in blue over the conductivity model. The figure includes the topography in southeast Australia.
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7 Open Research

The processed edi files for the MT data (Kyi et al., 2020; Duan & Kyi, 2018) are available at http://dx.doi.org/10.11636/Record.2020.011 (NSW) and http://dx.doi.org/10.11636/Record.2018.021 (Vic). The MT impedances used for the inversion (Kirkby et al., 2020) and the Vp velocity model of Rawlinson et al. (2016) are available on Zenodo (European Organization For Nuclear Research & OpenAIRE, 2013) at https://doi.org/10.5281/zenodo.8378540. The inversion software for this research (Manassero, 2023) is available at https://doi.org/10.5281/zenodo.10139638.

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