In general, seismic methods provide a reliable way to image the crust–mantle interface, which is marked by a rapid increase in seismic velocity (the Moho). However, the coverage provided by seismic networks is necessarily limited due to access difficulties, and the cost and labour involved in collecting data. Gravity data provide an alternative way to model the depth to the Moho, and provide more consistent and broader coverage. We discuss the usefulness of gravity data to model Moho depth, and the advantages and disadvantages of several gravity modelling methods. As an example, a model of Australia's Moho is generated through seismically constrained gravity inversion, including an estimate of modelling uncertainty. The inversion results demonstrate that gravity inversion is generally useful, but that its usefulness is subject to the following limitations: 1 — gravity inversion cannot spontaneously generate thick, high-density crust, nor thin, low-density crust, and, unless constrained, will not generate a correct Moho where such crust exists. 2 — major errors in the definition of the a-priori density structure, in particular features that are fixed during inversion, will influence the Moho results. 3 — applying a broad range of inversion parameters is necessary to characterise uncertainty. Model variability maps for Australia show that the average error is less than 5 km. There is a general relationship with seismic coverage, but the areas of highest uncertainty are not necessarily those with the lowest seismic estimate density. Comparison with previous seismic, and seismic-gravity models of Australia's Moho indicates that low seismic data density limits usefulness due to higher uncertainty in the gravity inversion. High-seismic data density also limits usefulness because Moho depth is largely known, and there is little scope for change. The usefulness of gravity inversion is maximum under conditions where seismic coverage is moderately dense, but estimates are well distributed.
to understanding the solid earth. Furthermore, crustal thickness is a crucial constraint on many geoscientific endeavours, including the interpretation and modelling of tectonic systems (e.g. Aitken et al., 2009; Brandmayr et al., 2011; Prezzi et al., 2009; Salmon et al., 2011), geodynamic modelling of earth processes (e.g. Beaumont et al., 2000; Grobys et al., 2008; Pfeiffer et al., 2000; Valera et al., 2011), resource exploration (e.g. Begg et al., 2010; Bierlein et al., 2006) and analyses of crustal stress fields, and their influence on earthquake risk (e.g. Flesch et al., 2001; Naliboff et al., 2012).

The crust–mantle interface (hereafter considered equivalent to the Moho) is most commonly imaged using seismic methods, most notably seismic refraction studies and receiver function analyses, but also seismic reflection profiling and seismic tomography. Each of these methods is capable of generating an estimate of the local Moho depth. In refraction, the true Moho, i.e. the increase in P-wave velocity from $<7.2$ km s$^{-1}$ to $>7.6$ km s$^{-1}$, can be determined. In receiver function analysis, the discontinuity can cause the partial conversion of P-wave energy to an S-wave (or alternatively S-wave energy to a P-wave), which results in a delayed Ps-wave arrival (or accelerated Sp arrival). In this case, the crustal velocity structure can be modelled to match the waveform, so as to achieve an estimate of Moho depth (Langston, 1979). In seismic reflection, the Moho is defined not by its velocity contrast, but by its reflectivity. Moho reflectivity can be highly variable and the Moho is not necessarily reflective, in which case its depth can be estimated from the base of lower crustal reflectivity (Kennett et al., 2011). Seismic tomography, in particular ambient noise tomography (Behr et al., 2010; Stehly et al., 2009), can also yield information on Moho depth, although this is the least reliable of the seismic methods.

Although they each carry inherent uncertainties and interpretational difficulties, these seismic methods provide generally robust knowledge of the depth to the Moho, and indicate something of its character for the portion of crust directly sampled by the study. However, in the creation of extensive regional-scale or continent-scale maps, sub-optimal coverage causes several problems. Firstly, where seismic data are sparse, the model will be poorly constrained because the influence of local estimates must be extrapolated far beyond the sampled area. Secondly, irregularly spaced coverage causes a requirement for a balance between good resolution of smaller-scale features where data are closely spaced, and adequate coverage where data are more widely distributed. Thirdly, the vast majority of seismic observations are made on land, and imaging of the crust in offshore regions is often poor. It is also worth noting that the various seismic methods are most sensitive to different aspects of the seismic structures and so different Moho estimates need not correspond exactly.

Because the Moho is also a prominent density gradient, gravity data have a role in understanding the geometry of the Moho, and can help in interpreting its nature. Gravity data are often available at higher resolution and with more regular coverage than seismic studies. In particular, combined satellite-ground gravity models (e.g. EGM 2008 (Pavlis et al., 2008)) provide seamless global coverage at relatively high resolution. However, the usefulness of gravity data is limited by the lack of a reliable means to discriminate between the signal from deep crustal density contrasts, the Moho, and density variations in the uppermost mantle.

This manuscript describes the application of a gravity inversion scheme to model the geometry of Australia’s Moho, using a recent compilation of seismic Moho depth estimates (Kennett et al., 2011). Our primary aim is to generate a model of Australia’s Moho that satisfies both seismic and gravity data. However, we also seek to understand the capabilities and limitations of the method applied, and the influence of the level of seismic constraint on the resulting model.

2. Moho modelling using gravity data — a review

Gravity data have long been used to constrain models of crustal thickness (e.g. Vening- Meinesz, 1931). Throughout this time, a great variety of approaches have been used, including forward modelling, inverse modelling and process-oriented modelling. Estimation of Moho depth from the spectral content of the gravity data has also been applied (e.g. using Euler deconvolution (Tedla et al., 2011) or power spectrum analysis (Studinger et al., 1997)), however, the usefulness of such methods has been questioned (Reid et al., 2012). Due to the availability of more capable and robust methods, we do not discuss parameter estimation here. Below we summarise the philosophies behind forward, inverse and process oriented modelling, and outline several methods used within each.

2.1. Forward modelling

Forward modelling involves the generation of a model of the Earth’s crust, which can then be modified until a satisfactory fit to the gravity data is achieved. This approach has the key advantage that model geometries can be explicitly controlled, and hence, there are few problems in generating a believable Moho geometry. Forward modelling however has the problem that changes to the model are driven by user-input, and are thus subjective, and are prone to error due to misinterpretations on the part of the modeller. Several methods have been successfully used.

The least structure approach is based on the concept that the most simple model that adequately fits the data is the most reasonable, following the concept of Occam’s razor. At its simplest, this will involve a layered model with sedimentary basins, crystalline crust, perhaps with middle and/or lower crustal layers and the mantle, each with constant density (e.g. Ebbing et al., 2001; Grobys et al., 2008), although additional blocks are often required to satisfy either the gravity data, or to include other known features (e.g. Ferraccioli et al., 2011; Hackney, 2004). With least structure, there is the advantage that the number of variables is much reduced, and so the problem becomes less under-determined, and solutions become less uncertain. Further advantages include reduced labour due to the relative simplicity of the modelling process and a more obvious link between model structure and the resulting gravity anomalies. Nevertheless, the least structure approach has several drawbacks. Firstly, the Earth’s crust and mantle are not homogenous materials, and representing them as such is an oversimplification that can have serious consequences for the model. In particular, a modelled Moho geometry using constant density contrast is an end-member of the range of possibilities, and the Moho could often be better represented by including relatively small changes in the densities of the crust and/or mantle. Secondly, the level of structural complexity that can be derived from gravity data is typically low, and is often much lower than would be expected in reality. Thus least structure models are often at odds with the crustal structure derived from geological studies or seismic reflection lines. Notwithstanding these drawbacks, this method is appropriate for many studies, and provides a relatively simple and effective method where constraints are few. Continent-scale coverage is possible using this approach (e.g. Grobys et al., 2008) but only with significant effort.

The structural–tectonic approach (e.g. Aitken et al., 2009; McLean et al., 2010; Prezzi et al., 2009; Stewart and Betts, 2010; Williams et al., 2010) is based on the concept that incorporating knowledge gained from other studies (in particular near-surface detail) generates a more geologically realistic result. For example, known structures from geological cross sections may be included (McLean et al., 2010; Stewart and Betts, 2010), near-surface structure from magnetic modelling may be modelled in detail to constrain upper-crustal structure (Aitken et al., 2009), or the crustal structure imaged in seismic reflection profiles may be extrapolated to other areas (Williams et al., 2010). Where abundant data exist, these can all be combined into a detailed tectonic model, prior to modelling (e.g. Prezzi et al., 2009). So long as the densities of the surface rocks can be constrained, this approach has the advantage that, because the upper-crustal
structure is better constrained, the geometry of the deeper crust is less uncertain. However, the level of structural complexity is usually higher than that required by the gravity data, and so the level of underdeterminedness is increased. Hence, the scope for over-interpretation is increased and there is a greater risk that misinterpretations of the structure of the upper-crust will lead to erroneous Moho geometries. Due to these drawbacks, this method is best used to resolve problematic Moho geometry for relatively local areas where upper-crustal structure is well constrained by other means. 3D continent-scale coverage using this approach is possible, but a major undertaking, given the labour-intensive nature of this approach, and the requirement for in depth understanding of upper crustal structure.

2.2. Inverse modelling

Inverse modelling of the Moho surrenders the explicit control of forward modelling for an implicit approach, where a computer algorithm has control over the resulting geometry (Oldenburg, 1974; Zhdanov, 2002). Inversions have the advantage that the computer does most of the work, and large areas can be covered at high resolution within a reasonable time. In addition inversions are repeatable and do not rely as much on subjective interpreter bias, although significant decisions must still be made to drive the inversion towards a believable result. To avoid nonsensical results, the changes made during inversion must be controlled by some regularisation procedure, the exact choice of which can lead to dramatically different model results (Zhdanov, 2002).

The most straightforward approach to modelling the Moho is to invert for the geometry of a single layer-boundary with constant density contrast (Moritz, 1990; Oldenburg, 1974; Parker, 1972; Vening-Meinesz, 1931). This approach has the advantage that for a given density contrast it permits a unique solution, notwithstanding the imperfection of the input data (Parker, 1972). The application in the wavenumber domain, in either spatial or spherical coordinates (Moritz, 1990) is also exceptionally fast. Alternatively, a similar approach can be applied in the space domain (Barbosa and Silva, 1994; Barbosa et al., 1999; Bott, 1960), although more computationally intensive, space domain models are more readily constrained by known pierce-points (e.g., seismic Moho depth estimates) and geometrical changes can be constrained by weightings (Barbosa et al., 1997). These surface modelling methods are limited by the inherent assumption that the geometry of the Moho is responsible for all observed gravity anomalies. Hence, for these inversions to produce a reasonable Moho geometry, great care must be taken to filter the gravity data appropriately. Typically, the gravity effect of topography is removed by Bouguer reduction, and the gravity effect of known sedimentary basins, ice-sheets, oceans etc. are calculated and removed. In addition, wavelength filtering is applied in an attempt to remove short-wavelength gravity anomalies that represent upper-crustal features, and also very-long wavelength anomalies that represent density structure within the mantle (e.g., Braitenberg et al., 2000; Ebbing et al., 2001; Shin et al., 2009; Steffen et al., 2011). In applying such wavelength filtering, there is a risk that part of the Moho signal is inadvertently removed. There is also the risk that significant anomalies from the crust or mantle density variations are not removed in an attempt to preserve the Moho signal. Additional modifications to this method include the inclusion of gravity anomalies due to thermally induced density variations in the upper mantle (AlveY et al., 2008), and the spectral correlation of gravity anomalies with topography (Braun et al., 2007).

An alternative conventional method for gravity inversion is to apply inversion for density within a 3D grid of small cuboidal cells (Li and Oldenburg, 1998). This approach has been extensively applied to upper crustal problems, however, applications to the lower-crust or uppermost mantle are relatively few (although see Brandmayer et al., 2011; Welford and Hall, 2007; Welford et al., 2010). Despite being less popular than the surface-modelling approach, this approach is not without utility, in that it can be used to simultaneously model density variations within the crust and mantle, and by proxy, model the Moho geometry. Intelligent mesh design and sensible regularisation procedures permit reasonable resolution of the Moho surface, whilst also allowing density variations within the crust and mantle to be modelled as part of the same procedure (Welford and Hall, 2007; Welford et al., 2010). The full 3D nature of this approach, and the requirement for many cells in the vertical direction, means that this method is more computationally intensive than the other methods, and so applying it at very large scales requires significant computational resources and time.

The above approaches to the exploitation of gravity data each provide a valid, although limited, approach. Recent software developments have answered the need for flexible methods that can allow for changes to both the geometries of model features, and the density distribution within them (Fullagar et al., 2008; Guillen et al., 2008; Schmidt et al., 2011) or at least allow the density contrast at this interface to be variable (Silva et al., 2006; Sjöberg and Bagherbandi, 2011). The latter approach allows variable density contrast across the Moho, as part of a joint inversion scheme. This reduces the sensitivity of the model to extreme Moho variations, and allows some interpretation of lower crust and uppermost mantle density structure (Sjöberg and Bagherbandi, 2011). In the former category, Aitken (2010) proposed a flexible Moho gravity inversion scheme using VPmg™ software (Fullagar et al., 2008) that allows for the explicit incorporation of seismic constraints, allows for precise definition of surfaces and incorporates modelling of lateral variations in crust and/or mantle densities. Using Geomodeller™ software (Guillen et al., 2008), Schreiber et al. (2010) applied a similar concept to modelling of the Moho in the southwestern Alps. The approach of Guillen et al. (2008) involves random changes to the density of cuboidal cells or alternatively, the lithologies of cells located at a lithological boundary can be changed. Allowing the inversion to run for many iterations beyond where an asymptotic fit to the gravity data was achieved provides an estimate of the probability that an individual cell is occupied by a particular lithology, and also allows 3D mapping of the most probable lithology (Guillen et al., 2008).

The combined density/geometry approach removes some of the oversimplifications present in other inversion schemes (e.g. the assumption of constant Moho density contrast), but introduces problems of its own, in particular balancing the trade-off between changes in crust (or mantle) density against changes in Moho depth as a means to reduce the gravity misfit.

2.3. Process oriented approaches

A third approach to gravity Moho modelling involves using a Moho surface that is considered a result of a physical process, or indeed several. These modelling methods have several advantages. Firstly, the resulting Moho geometry is explicitly linked to a causal physical process, and is inherently realistic (albeit subject to the necessary simplifications). Secondly, the inputs to the modelling process are usually much better known than the Moho geometry, for example, global topography is known in impressive detail, e.g. ETOPO1 (Amante and Eakins, 2009), and the age of the oceanic crust (a primary control on crustal thickness) has also been mapped in detail (Müller et al., 2008).

For the continental crust, the process most commonly used to constrain the Moho geometry is the principle of crustal isostatic compensation either locally, or through regional flexure (e.g. Aitken et al., 2012; Jiang et al., 2004; Karner and Watts, 1983; Karner et al., 2005; Petit and Ebinger, 2000; Stern and Ten Brink, 1989; Vening-Meinesz, 1931; Watts, 1994). This can be applied as a forward problem, in which case loads are applied to the lithosphere using either a known flexural rigidity, or by applying a range of values in order to find the one that generates the closest fit to the gravity data (e.g. Aitken et al., 2012; Jiang et al., 2004; Stern and Ten Brink, 1989). Most commonly,
current topography is used as a load on an elastic plate with uniform flexural rigidity. This basic approach neglects the influence of mantle buoyancy forces, uncompensated crustal loads, palaeotopography and load evolution, and spatial and temporal variations in flexural rigidity. These can be incorporated in a forward model if required (Ferraccioli et al., 2011; Karner et al., 2005), however, they must first be known with reasonable confidence. The chief drawback with such an approach is the assumption that the crust is isostatically equilibrated (notwithstanding applied mantle buoyancy forces), and the requirement that crustal isostatic compensation is observed within the studied area.

Oceanic regions, in general, do not display crustal isostatic compensation, with the thickness and density of the lithospheric mantle and buoyancy forces within the asthenosphere being the dominant isostatic controls (e.g. Zlotnik et al., 2008). Oceanic crustal thicknesses are remarkably consistent, at approximately 7 km (Yongshun John, 1992), except in regions where flexure is observed (e.g. Watts, 1994), or the crust is otherwise anomalous, for example at oceanic plateaus, at fracture zones and at spreading ridges (Mooney et al., 1998). For these anomalous regions, gravity modelling has a role in understanding crustal thickness (Davy and Wood, 1994; Gladczenko et al., 1997; Grobys et al., 2008; Pim et al., 2008; Watts, 1994). An additional consideration is that gravity anomalies in oceanic regions, and also rifted continental margins, contain a significant gravity contribution from thermal variations in the mantle. These thermal effects can be explicitly removed from the gravity data prior to modelling (Alvey et al., 2008), however, a more sophisticated approach has been developed that simultaneously inverts for crustal thickness, stretching factor and the mantle temperature field (Chappell and Kuszniir, 2008).

Finally, many geodynamic numerical modelling algorithms permit the modelling of geodynamic scenarios which can be compared with real-world observables such as topography, basin evolution, plate-motion, the observed stress field, gravity anomalies and the geoid. Examples of such modelling includes the modelling of collisional orogens (Jiménez-Munt et al., 2005; Tang and Chemenda, 2000), slab subduction processes (Hashimoto et al., 2008; Marotta et al., 2006), continental rifting processes (Elesin et al., 2010; Lesne et al., 2000), and the thickness of oceanic lithosphere (Zlotnik et al., 2008). These methods have the capability to generate models of Moho depth that are fully consistent with a range of physical processes, including those between which complex feedbacks are observed. Currently, these methods are somewhat limited in their abilities, and have not yet been applied to broad-scale Moho mapping in 3D for “real” examples. However, such methods may be the future of process oriented modelling to understand the nature and geometry of the Earth’s Moho, and gravity data are likely to be a principal constraint on such modelling.

The chief drawback to process oriented methods is that they are unable to realise Moho geometries that are inconsistent with the process model applied. Despite this, a process-oriented approach has much utility in regions where Moho geometry is dominated by a physical process for which the main parameters are known. This requirement for the processes to be both well understood and the same throughout the area means that process-oriented modelling is most useful at the regional-scale.

In summary, each of these techniques has particular situations to which it is well suited, and other situations where it is probably unsuitable. All of these approaches suffer from the fundamental problem that a gravity anomaly of appropriate wavelength to represent the Moho could, in fact, be due to anomalous density in the crust or the mantle. Unless two of these parameters are known with reasonable certainty, resolving the other is a matter of high uncertainty. Hence, unless nature is simplified considerably, the models are highly underdetermined and many solutions may be found that will fit the data.

2.4. Methods used in this study

This study largely follows the method of Aitken (2010), but with some modifications to both the initial model and the method used. The initial model follows the same principles and uses similar datasets; however, some important differences are included. Firstly, the model covers a much larger area, so as to include more of the Australian continent and to avoid the influence of edge-effects, which can extend up to 500 km from the edge of the model (±0.5 mGal tolerance). The topographic model is now ETOP01 (Amante and Eakins, 2009), which provides consistent and seamless coverage across the whole region. The cell-size was increased to 20 km (from 15 km), however, because sensitivity to Moho geometry is on wavelengths of ~100 km or more, this affects only the imaging of crustal density variations.

Sedimentary basins are a major limiting factor on the Moho inversion process. This is because they are close to the surface, and thus possess high-amplitude gravity signals, but are often broad and so it is difficult to remove their gravity signal by wavelength filtering. The basin model used in the study of Aitken (2010) was based on the continent-scale SEEBASE compilation of sedimentary basin thicknesses (Frogtech Pty Ltd, 2005). We follow that approach here, adding sediment thicknesses from Laske and Masters (1997) in areas not covered by SEEBASE.

Although it captures sedimentary basin thickness in reasonable detail, the SEEBASE model does not include any density information. The constant density model used by Aitken (2010) was unrealistic in that it failed to capture the increasing density of sedimentary rocks with depth, leading to irresolvable misfits over the deepest basins. Ideally, density information for these basins would be derived from a combination of density measurements, and seismic velocity information, as is routinely undertaken in well-sampled basins. Such data are not available for Australia at the continent scale, and so we seek to find a model that is simple but appropriate. Basins are subdivided into three layers: an upper layer with a maximum thickness of 2 km, and an initial density of 2.4 g cm⁻³, a middle layer with initial density of 2.5 g cm⁻³ has a maximum thickness of 4 km. A lower layer with initial density of 2.6 g cm⁻³ occupies the remainder of the basin thickness. Initial, maximum and minimum densities (Table 1) are derived for each layer by converting typical seismic velocities using Gardner’s rule (Gardner et al., 1974; Telford et al., 1990). The inclusion of sedimentary basin density as a variable in inversion means that the assumptions made (excepting the broad upper and lower limits on density) are not imposed too strongly, and variations form part of the uncertainty analysis of the inversion process.

The internal structure of the crystalline crust is also important, although less so than the sediment thickness. Of course, seismic studies that resolve the Moho must also resolve the crust, and this is often achieved in impressive detail. Unlike the Moho, which is a common

<table>
<thead>
<tr>
<th>Layer</th>
<th>Initial density (g cm⁻³)</th>
<th>Minimum density (g cm⁻³)</th>
<th>Maximum density (g cm⁻³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sedimentary rocks 1</td>
<td>2.40</td>
<td>2.10</td>
<td>2.70</td>
</tr>
<tr>
<td>Sedimentary rocks 2</td>
<td>2.50</td>
<td>2.20</td>
<td>2.70</td>
</tr>
<tr>
<td>Sedimentary rocks 3</td>
<td>2.60</td>
<td>2.45</td>
<td>2.70</td>
</tr>
<tr>
<td>Upper crust</td>
<td>2.75</td>
<td>2.55</td>
<td>3.00</td>
</tr>
<tr>
<td>Lower/oceanic crust</td>
<td>2.95</td>
<td>2.75</td>
<td>3.15</td>
</tr>
<tr>
<td>Eclogite</td>
<td>3.10</td>
<td>2.95</td>
<td>3.25</td>
</tr>
</tbody>
</table>
feature of all of the Earth’s crust, intra-crustal structuring is highly variable at the regional scale, and a boundary within one tectonic block need not necessarily have a correlative in an adjacent block. This regional variability, allied with the variability in resolving power from the different methods used, mean that it is difficult to generate a continent-wide map of intra-crustal structure that is simple enough to be included in this sort of analysis. Thus, we again seek to find a model that is simple but reasonable. A three-layer crystalline crust is used, with an upper to lower crust boundary at half the crustal thickness, or at 20 km depth where Moho depth is greater than 40 km. As indicated by Aitken (2010), an extremely dense “eclogite” layer is required below ca. 40 km depth to achieve a reasonable gravity fit for Australia. The effects of deviations of this assumed crustal structure from reality are potentially significant. However, these are mitigated by the reduced sensitivity of gravity to these intra-crustal layers due to the smaller density contrasts compared to the Moho. As with the sedimentary basin density, the inclusion of the thickness of these layers and their densities as inversion parameters allows a certain degree of freedom, and the assessment of variability under different inversion conditions to be assessed.

The Moho surface is of course defined by much more seismic information than in Aitken (2010), however the model extends well beyond the limits of these data. Two areas were defined — a central region of “constrained” crust bounded by the outermost seismic constraints, and a peripheral region of “unconstrained” crust. Moho geometry in the latter was defined by a local Airy isostatic model (Eq. (1)) based on the ETOP01 surface, represented as a grid with 50 km cell size. This cell size captures the major topographic trends without being over-sensitive to local topography, and captures the wavelengths to which gravity inversion is sensitive (>100 km). Inside the constrained region, control nodes were manually added to the seismic Moho estimates to provide more consistent resolution. The same isostatic model (Eq. (1)) was used to define Moho depth at these points. The resulting Moho surface is shown in Fig. 1a.

$$M = M_0 - \left[ \frac{\rho_c - 1}{\rho_m - \rho_c} \times (T_1 - T_0) + \frac{\rho_m}{\rho_m - \rho_c} \times T_2 \right]$$

$$M$$ is Moho depth, $$M_0$$ is the reference Moho elevation, $$\rho_c$$ and $$\rho_m$$ are crust and mantle densities respectively, $$T_1$$ is bathymetric relief and $$T_2$$ is topographic relief $$T_0$$ is the reference topographic elevation. In this case, $$M_0 = -14000$$, $$T_0 = -5000$$, $$\rho_c = 2.8$$ g cm$$^{-3}$$ and $$\rho_m = 3.3$$ g cm$$^{-3}$$. The crustal thickness returned at some ocean trenches by this method was less than 6 km. In these cases, crustal thickness was set at 6 km. Topography in the area is generally low and flat, and it was not deemed necessary to limit maximum crustal thicknesses in a similar way.

The “hard” Moho constraints used by Aitken (2010) are inappropriate given the existence of seismic uncertainties and the often transitional nature of the crust–mantle boundary. This resulted in several artefacts around seismic estimates (Aitken, 2010). In this work, we apply a different approach, using the uncertainties in the seismic data as estimated by Kennett et al. (2011) to generate upper and lower bounds for the Moho surface in the vicinity of seismic estimates. In each case, the weighting was translated to an uncertainty using the formula $$U = 12 - 10$$, where U is uncertainty, and w is the weighting. Weightings were assigned on the basis of the type of data (i.e. refraction, receiver function, reflection), and in the case of receiver functions and reflection lines, inversion quality and pick quality respectively (Kennett et al., 2011). Control nodes were not constrained. This error was symmetrically applied to the Moho surface ($$M \pm U$$), thus permitting both upward and downward movements, and so takes account of random seismic uncertainties, but does not account for the transitional nature of the crust–mantle boundary, as the seismic picks of Kennett et al. (2011) were placed at the base of the transition zone, and may be systematically deeper than the Moho surface that is most consistent with the gravity data.

The mantle occupies the space from the base of the crust to the base of the model at 75 km depth, and we treat the mantle as a layer with laterally variable density, but no vertical density variations. The mantle-density grid was updated from that of Aitken (2010) (Fig 1b). The same temperature calculation and model (Goes et al., 2005) was used, but the definition of lithospheric age was modified to better represent the age of the lithosphere at depth, rather than the surface, following the interpretation of Kennett et al. (2011). Mantle densities prior to the temperature perturbation were as follows: Archaean lithosphere (type 1) and Archaean to Palaeoproterozoic lithosphere (type 2) were assigned a density of 3.31 g cm$$^{-3}$$. Palaeo-Mesoproterozoic lithosphere (type 3) and Neoproterozoic lithosphere (type 4) were assigned a density of 3.35 g cm$$^{-3}$$, Palaeozoic lithosphere (type 5) and transitional lithosphere (type 6) were assigned a density of 3.36 g cm$$^{-3}$$. Oceanic lithosphere (type 7) was assigned a density of 3.39 g cm$$^{-3}$$.

These changes to the initial model were accompanied by changes to the method. The combination of terrestrial/satellite derived free-air gravity data used in MoGGIE is replaced by the gravity disturbance derived from the EGM 2008 model. These data provide wider coverage and a consistent model across the whole modelled region. The model used was calculated at the Earth surface on a 0.1° by 0.1° grid. Very-long wavelength gravity trends relating to deep-mantle processes were removed by subtracting a separately calculated version of EGM 2008 that is complete only to degree 10. This process is considered equivalent to a high-pass filter with an approximate wavelength of 2300 km. This high-pass filtered data was then re-gridded at 20 km spacing and upward continued by 10 km, so as to match the resolution of the model whilst avoiding short-wavelength content that results from the steep topographic gradients at the edges of cells (Fig 1c).

The most important change to the method is the management of the density/geometry trade-off. We recall that the method employed alternating iterations in which the density within prisms is modified, then the geometry of layer boundaries, then density again and so on (Aitken, 2010). As well as the hard constraints listed above, the degree to which changes to density and geometry are permitted are controlled by per-iteration constraints, which must be imposed by the user. Previously a single combination was used (Aitken, 2010); however for this work, we utilise a range of values for each parameter, and so sample the range of solutions possible given the initial conditions of the model.

From these results, the mean and standard deviation of the Moho geometries can be calculated, giving an indication of the most likely model result and the variability of the results. It is important to note that, although the variability is highly sensitive to the level of

Fig. 1. a) The initial Moho surface showing the location of seismic estimates (black dots) control points (white dots) and the boundary of the “constrained” region (black line). b) Mantle density. Black lines indicate the boundaries of different mantle “age-types” labelled as follows: 1 — Archaean, 2 — Archaean to Palaeoproterozoic, 3 — Palaeo-Mesoproterozoic, 4 — Neoproterozoic, 5 — Palaeozoic, 6 — Transitional, 7 — Oceanic. c) Gravity disturbance at 10 km above the Earth’s surface. The gravity disturbance grid is smaller than the model area to mitigate edge effects.
In total, 23 inversions were run with a range of density and geometry constraints. Permitted per iteration (PPI) density changes ranged from 0 g cm\(^{-3}\) to 0.15 g cm\(^{-3}\) and PPI geometry changes ranged from 0\% to 50\%. Inversions were run for 20 iterations, however, analysis of the misfit curves indicates initial reduction in the misfit, before the onset of instability, typically occurring around iteration 8-10 (Fig. 2). The onset of instability was identified by large increases in the misfit (>0.25 mGal RMS), and also the onset of high misfit variability in the subsequent iterations. A single poor iteration was tolerated, provided subsequent iterations were stable. The iteration prior to instability was considered the final result of the inversion. Of the 23 inversion results, 12 were rejected. With a PPI density change of 0.025 g cm\(^{-3}\) models with PPI geometry change of less than 5\% were rejected on the basis that they did not generate an adequate fit to the data (RMS misfit > 10 mGal). The model with PPI geometry change of 50\% was also rejected because it generated an identical result to the model with 40\% PPI geometry change. With 25\% PPI geometry change, inversions with PPI density change of less than 0.020 g cm\(^{-3}\) were rejected because they did not generate a believable Moho geometry. Models with PPI density change of greater than 0.06 g cm\(^{-3}\) were also rejected because the results were identical to the result with a PPI density change of 0.06 g cm\(^{-3}\). The remaining 11 models were used to calculate the mean Moho surface and its standard deviation.

The mean result (Fig 3) is considered to represent the most likely Moho geometry given the initial conditions. This result shows that the initial model is largely validated, that dramatic changes are few, and that changes typically are fairly short-wavelength. Changes from the initial model are, naturally, focused within the less-well-constrained parts of the model, namely offshore regions and within central-western Australia.

The standard deviation of these results (Fig 4) shows that model Moho variability is generally low offshore, and in Tasmania. These results indicate that the result in these regions is robust under different model conditions, and indicates that seismic Moho depth estimates are compatible with the crustal density model. With some exceptions, seismic estimates merge smoothly with the gravity-modelling results in these low-variability areas. Model variability is highly changeable onshore. The largest variability is observed in an ENE trending band in central Australia, roughly corresponding with the Albany-Fraser, Musgrave and Arunta provinces. These regions possess thick to very thick crust (>45 km) but gravity anomalies are moderately low to very high, indicating high density crust (or perhaps mantle) that may not be well described by the initial model. Variability is much lower throughout most of eastern and western Australia, although areas of high variability exist. This indicates that either only minor changes to the initial model were required to satisfy the gravity field, or that these solutions were well constrained by either the seismic constraints, or by the gravity modelling process. It is interesting to note that, despite low seismic estimate density, the area between longitude 135°E to 140°E and latitude —22°S to 30°S shows very low variability in comparison with similarly constrained crust immediately to the west (i.e. longitude 125°E to 135°E and latitude —20°S to 30°S). This indicates that the robustness of modelling results is not solely related to the density of seismic data, but also reflects the constraints inherent to the gravity modelling process.

Intra-crustal boundaries were analysed in a similar fashion. The upper-crust/lower crust boundary shows generally quite low variations, with a standard deviation of less than a kilometre even in the least well constrained areas, and significantly less throughout much of the model (Supplementary Fig. 1). This variability is most similar to the crustal density anomaly map (Fig 6), indicating that in this model the upper crust behaves differently to the lower crust. The variability of the top of the eclogite layer quite closely matches the Moho variability in both magnitude and pattern, with the exception that central Australia shows less variability (Supplementary Fig. 2).

Of all these models, a single model was selected with which to represent crustal density variations. The model with permitted geometry changes of 25\% and density changes of 0.025 g cm\(^{-3}\) was selected because it achieved the lowest misfit before instability (Fig. 2), and its geometry was close to the mean Moho, with an average discrepancy of 275 m. After 40 further iterations, density-only inversion achieved a final RMS misfit of 3.31 mGal (Fig 5). Residual misfit is concentrated along the continent’s southern margin and in the Bonaparte Gulf and Woodlark basins. The existence of serpentinised peridotite has been inferred beneath the southern margin of Australia in the Great Australian Bight (e.g. Direen et al., 2011) but is not included in the model. The Bonaparte Gulf has a combination of thin crust and thick sediments that does not permit a fit to the gravity data. The residual anomaly in the Woodlark Basin likely reflects un-modelled gravity anomalies from small-scale thermal anomalies in the mantle.

The crustal density anomaly is a vertically averaged record of the percentage difference between observed density, and that of a model with the same geometry, but populated by initial density values (Table 1). This image shows that much of the continent is occupied by relatively dense crust, much of it focused in areas of thick crust (Fig 6). In some cases, these crustal density anomalies mirror the margins of tectonic boundaries (e.g. in eastern Australia) but in others, they transgress major boundaries (e.g. in central-western Australia). Oceanic crust is generally of reduced density relative to the initial model.

4. Discussion

4.1. Australia’s Moho

This model generated few fundamental changes to the Moho geometries generated in the work of Kennett et al. (2011), and...
Salmon et al. (2013), and readers are referred to those studies, and also Aitken (2010) for more in-depth discussion of the main features. Briefly, these Moho maps show moderately thin (~35 km) crust beneath the Archean Yilgarn and Pilbara cratons, except the Southwest Zone of the Yilgarn Craton, where the crust is thicker (40–45 km). The intervening Capricorn Orogen has thickened crust. The Palaeoproterozoic North Australian Craton typically has thick to very-thick crust (>40 km), with the exception of the Kimberley Craton, where the crust is thinner (35–40 km). Central Australia is dominated by thick crust (>45 km) likely reflecting crustal thickening during intraplate orogenesis, which also caused localised Moho uplifts of up to 15 km (Aitken et al., 2009; Goleby et al., 1989; Korsch and Kositcin, 2010; Korsch et al., 1998). Magmatic underplating due to the Warakurna Large Igneous Province is a further possibility to explain the thick crust here (Aitken, 2010). The Gawler and Curnamona Cratons are characterised by moderately thick crust (40–45 km). There is a rapid transition to the much thinner crust of eastern Australia, which is generally between 25 and 40 km thick. Structuring in Eastern Australia follows the N-S tectonic fabric, with several discrete blocks (e.g. New England Orogen) defined. Tasmania and the Bass Strait have the thinnest continental crust (20–25 km), reflecting crustal thinning during Gondwana breakup.

Changes that have been made to the initial model generally involve the “sharpening” of Moho steps associated with major crustal boundaries — for example the boundary between the New England Orogen and Lachlan Orogen is better defined, and resolution of the Willowra Suture is enhanced. In other areas, short wavelength features are introduced, for example the Mundrabilla Shear Zone in southern Australia. In a few areas (for example the Cape York Peninsula, and beneath the Great Sandy Desert), the gravity inversion process has generated significant changes across large areas with sparse seismic controls.

4.2. The capabilities and limitations of gravity inversion for Moho modelling

As expected, the use of gravity inversion has increased the resolution of short-wavelength Moho features, and has provided good coverage in offshore regions, and regions lacking seismic data. With this model, the changes are not so dramatic as with Aitken (2010), primarily due to the much denser network of seismic estimates, but also the averaging process applied, which smooths out any extreme values from individual results.

The results of this work indicate that gravity inversion can provide highly consistent and reasonably accurate models of the Moho geometry, even where seismic constraints are relatively sparse. The majority of the model shows a standard deviation of less than 1.5 km, equivalent to an error (3σ) of less than 5 km. High-variability areas however, have standard deviations upwards of 2.5–3 km, equivalent to an error of 7.5–9 km. These results are consistent with other models, where gravity estimates are usually within ~5 km of seismic estimates (e.g. Braun et al., 2007; Steffen et al., 2011) but are occasionally more than 10 km in error. Thus, gravity inversion methods can have comparable accuracy to the less-reliable seismic methods, but with a greater risk of highly erroneous results.

In many instances, these high-error regions occur where the density structure of the crust is “anomalous” in that it is either thick and high density (e.g. due to a mafic underplate that thickens the crust) or thin and low density (e.g. due to deep sedimentary basins over thinned crust). Because the gravity responses from the Moho and the crust cancel-out to some degree, such crustal structures cannot be spontaneously generated by gravity inversion. For a reliable model in such areas, a firm knowledge of either Moho geometry or crustal density is required. Mantle densities of course are also influential, but less so due to the narrower range of permissible densities. Forward modelling or

Fig. 3. Mean Moho elevation, taken to represent the most likely geometry of the Moho. KC — Kimberley Craton, NAC — North Australian Craton, CYP — Cape York Peninsula, WS — Willowra Suture, GSD — Great Sandy Desert, PC — Pilbara Craton, CO — Capricorn Orogen, YC — Yilgarn Craton, WMP — west Musgrave Province, MSZ — Mundrabilla Shear Zone, CC — Gawler Craton, CC — Curnamona Craton, LO — Lachlan Orogen, NEO — New England Orogen.
process-oriented modelling may prove more successful in these regions due to the greater degree of control on Moho geometry.

This leads to the twin requirements that uncertainty is reduced as far as possible, and that the remaining uncertainty is characterised, so as to guide future efforts. Uncertainty can be directly reduced by introducing new constraints from seismic Moho estimates. Uncertainty can also be reduced by better resolving the density structure of the crust and mantle. Densities in the upper crust can be derived from sampling programs or from higher resolution modelling, with appropriately filtered gravity data. Seismic velocities of the lower crust and uppermost mantle are harder to constrain, but can be obtained from models of seismic-velocity or, less directly, from more robust models of their age, composition and temperature.

Uncertainty will never be eliminated, and should be investigated by analysing the sensitivity of the modelling to at least the assumptions made in the modelling process, but perhaps also to the uncertainties in a-priori constraints. The approach used here describes the sensitivity of results to a key modelling parameter – the crustal density/crustal geometry trade-off – and provides a means to identify regions where the model requires either more data, or a more appropriate modelling approach. The approach defined above does not include tests of the robustness of assumptions made in the initial model, for example, assuming that mantle density and sedimentary basin thicknesses are known could introduce erroneous (but low-variability) results where these models are in error. In an extension to the above method, sensitivity to such parameters could be included within the sensitivity analysis with little difficulty, so long as the range of possibilities is known.

### 4.3. The influence of seismic constraint

This model has used a much greater amount of seismic information (>1200 individual estimates compared to 230) than the model of Aitken (2010), and provides a compelling opportunity to test the influence of this increase of seismic information on the model results.

We also can compare these models to a recent global gravity-only Moho model derived from GOCE data (Sampietro, 2011).

Both seismic-gravity models have imaged the overall geometry of Australia’s Moho, including moderate crustal thickness beneath western Australia's cratons, thick crust beneath the Proterozoic terranes of central and northern Australia, and predominantly thin crust beneath eastern Australia. Each also defines several short-wavelength Moho features, including major fault zones (e.g. Mundrabilla Shear Zone), Moho uplifts (e.g. central Australia), the boundaries between crustal blocks (e.g. New England Orogen) and the continental margins. The amplitudes of these features are typically smaller in the new model, reflecting the tighter per-iteration constraints on the Moho surface. The average discrepancy is 2.4 km, with the greatest discrepancies (up to 25 km) in regions of thickened crust that were unconstrained in the Aitken (2010) model (Supplementary Fig. 3). Large discrepancies occur primarily where the crust is dense (e.g. the northern Gawler Craton, the west Musgrave province).

Although the aims, resolution and scope to include local information are different for the global gravity-only model, this model captures only some of the features of Australia’s Moho. These include thicker crust in the western two-thirds of the continent, the structure of central Australia, and the difference between the oceans and the continents. However, it contains many features that are known to be erroneous, such as thin crust (28 km) beneath the Mt Isa region, which is known to have thick crust (50 km) from seismic refraction studies (Maccready et al., 1998).

Overall the discrepancy between these models averages 3.5 km, and this is generally low (<2 km) in the oceanic regions and high (>6 km) on the continent (Supplementary Fig. 4).

The level of detail in each of the seismic-gravity models, and also Kennett et al. (2011) and Salmon et al. (2013) is similar, although the new model has a more irregular surface due to the very high density of seismic estimates in some areas (e.g. southeast Australia). All of these models show greatly improved definition of Australia's Moho over the model of Collins et al. (2003). This prompts a
**Fig. 5.** The residual gravity disturbance of this final model. Positive anomalies indicate insufficient mass in the subsurface, while lows indicate excess mass. BG — Bonaparte Gulf, GAB — Great Australian Bight, WB — Woodlark Basin.

**Fig. 6.** Crustal density variations across the Australian region. Crustal density represents the vertically averaged percentage difference from the initial densities (Table 1) for the same model geometry.
discussion on the influence of the level of seismic constraint on the usefulness of gravity inversion to define Moho geometry. Gravity inversion with low-levels of seismic constraint (e.g. central Australia) is useful, and may provide an understanding of likely Moho geometry where otherwise there would be none. Although the results of gravity inversion generally compare quite well with seismic constraints there is the potential for large errors, especially where the crust or mantle density is anomalous. Unless the potential for error is quantified, this inhibits the usefulness of the method. In such situations, it may be more effective to utilise a process-oriented method, or perhaps constrained forward modelling, to limit the range of potential Moho geometries. Gravity inversion with high levels of seismic constraint e.g. south-east Australia, post Kennett et al. (2011) have also limited usefulness as the Moho surface is largely known, and there is little scope for new structure to be imaged. In these situations, gravity inversion can still provide revealing insights into the nature of the lower-crust and uppermost mantle through modelling their densities.

The most useful areas in which to apply gravity inversion are those with a moderate level of well-spaced seismic estimates e.g. eastern Australia as in Collins et al. (2003), or areas where crustal structure in a low-data density region cannot be reasonably predicted from seismic data in an adjacent well constrained region e.g. at continental margins. In each of these cases, gravity inversion can produce robust models of the Moho that genuinely add to our knowledge.

5. Conclusion

The studies discussed in this manuscript demonstrate the value of gravity modelling to generate models of the Moho. In particular, they show that gravity inversion is capable of reasonable accuracy as a means of extending coverage of Moho models beyond the limits of seismic constraints and to interpolate between gaps in coverage. They also highlight the inability of gravity inversion to model crustal structure reliably in areas to which the method is ill-suited (e.g. thick, dense crust or thin, low-density crust), unless the structure is defined by seismic methods. In such areas, more detailed gravity modelling using strong a-priori constraints from non-seismic data may be more effective.

Applying a broad range of inversion parameters to generate a variability estimate is effective in identifying areas where the result is not well constrained by either seismic or gravity methods. Although there is a general relationship with seismic coverage, the least constrained areas are not necessarily those with the lowest seismic estimate density. Therefore, the variability map contributes to the optimisation of future seismic efforts to map Moho geometry.

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.tecto.2012.06.049.

References


