A crustal thickness model of Antarctica calculated in spherical approximation from satellite gravimetric data

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SUMMARY
The ice cap covering Antarctica has long limited our understanding of the continental-scale crustal model due to its inaccessibility and the resulting logistical difficulties when executing geophysical field work, such as seismograph deployment. Resolving a high spatial resolution crustal model for Antarctica where seismographs are sparsely distributed stimulates scientific interest in this relatively less studied continent. In this study, we utilize satellite gravity observations from the global gravity model EIGEN-6C4 to create an alternative crustal thickness model of Antarctica. The gravity data were corrected for sediments, topography and ice cover. Furthermore, we considered the gravity effect due to vertical deformation of the lithosphere caused by ice load besides the earth’s curvature in the modelling. We inverted the corrected gravity data using the regularized Bott’s inversion method in spherical approximation and constrained the results by seismic observations. This crustal thickness model shows a thicker average crust in East Antarctica and a thinner one in West Antarctica. The thickest crust is in the Gamburtsev Subglacial Mountains with a Moho depth of over 40 km. The thicker crust is particularly evident along the Transantarctic Mountains and the Dronning Maud lands. Comparisons with existing models show a good correlation in gravity-constrained areas. Differences appear in the sedimentary basins and crust with thickness closer to seismic point observations. Overall, our crustal model is relatively improved than the existing gravity derived models.

Key words: Gravity anomalies and Earth structure; Antarctica; Gravity inversion; Crustal structure.

1 INTRODUCTION
Resolving a high-resolution crustal model of Antarctica stimulates scientific interest. The inaccessibility of this continent limits the effectiveness of standard data collection methods making Antarctica a geophysical challenge. The development of the crustal model of Antarctica, however, is essential for understanding the processes responsible for the formation and splitting of Gondwana and the dynamics of plate tectonic motion. The early crustal models in Antarctica were derived from both seismic and gravity observations (Evison et al. 1962). Crustal models from seismic observations nevertheless still suffer from low resolution over Continental Antarctica, which result in greater trade-offs and poorly resolved crustal models due to inaccessibility, cost constraints and unevenly distributed seismic measurements. Advent of satellite gravity has greatly improved crustal thickness mapping in areas not adequately covered by seismic data (e.g. Tedla et al. 2011; Reguzzoni et al. 2013; Tugume et al. 2013; Van der Meijde et al. 2013; van der Meijde et al. 2015a; Llubes et al. 2017; Steffen et al. 2017). For nearly two decades, satellite gravity missions have collected better observational data over areas with sparsely seismic coverage that were previously not well-studied. Satellite gravity observations combined with terrestrial gravity data are available up to degree and order 2109 in most regions on the Earth (e.g. Förste et al. 2014). The lack of terrestrial data, however, limited the possibility to produce a model with degree and order 2109 in Antarctica. Availability of satellite gravity observations has fortunately led to improved crustal thickness estimates at continental scale of Antarctica.

Earlier crustal studies in Antarctica started in the 1960s (Evison et al. 1962) with the first continental scale model obtained from a compilation of seismic observations in the 1990s (e.g. Groushinsky et al. 1992). Over the years, crustal models of Antarctica have been produced from seismic observations (Baranov & Morelli 2013;
Laske et al. 2013; An et al. 2015; Baranov et al. 2017) and gravity observations (O’Donnell & Nyblade 2014; Baranov et al. 2017; Llubes et al. 2017). For instance, the CRUST1.0 model (Laske et al. 2013) is based on the active and passive seismic observations. Nevertheless, this model suffers from extensive crustal thickness estimation based on gravity constraints or geological terrane age estimates due to the sparse distribution of seismic observations. ANTMoho (Baranov & Morelli 2013) is based on the compilation of seismic studies obtained from 1960 to 2011 with very few points in West Antarctica (WA). AN1-Moho (An et al. 2015) also inferred a Moho model of Antarctica based on the analysis of seismic data from 122 broad-band seismic stations and over 10 000 Rayleigh wave fundamental-mode dispersion curves. They included recent additional seismic observations in WA (Chaput et al. 2014) and East Antarctica (EA; Feng et al. 2014), besides the cleaned seismic observations of ANTMoho in which poorly estimated points were removed. Despite the improvement, this model still suffers from uneven data distribution. Conversely, continental models from gravity observations have evenly data distribution. In recent years, Block et al. (2009) inverted satellite data from the Gravity Recovery and Climate Experiment (GRACE) Gravity Model 3C (GGM03C; Tapley et al. 2007) using the Parker–Oldenburg algorithm (Parker 1973; Oldenburg 1974) to obtain the Moho model of Antarctica. Using the same methodology O’Donnell & Nyblade (2014) produced a continental-scale crustal thickness model but separately inverted GOCC03S satellite gravity data (Mayer-Gürr et al. 2012) for WA and EA. They constrained their model using seismic crustal thickness estimates from ANTMoho. These models used relatively older and lower resolution global gravity models and an inversion method assuming a flat surface, which is not ideal for continental Antarctica. Baranov et al. (2017) developed three models, gravity only, seismic only, and combined seismic–gravity model based on the GOCO03S satellite gravity data (Mayer-Gürr et al. 2015) using the Vening Meinesz-Moritz’s inverse problem of isostasy in spherical approximation (Sjöberg 2009) and seismic points from the ANTMoho and the Chaput et al. (2014) model. They constrained their model using the ANTMoho data set and seismic data from Chaput et al. (2014). However, the ANTMoho seismic constraints contained some approximation which was thought to have large uncertainties and errors (An et al. 2015). Generally speaking, there is a high lateral variation of the crustal structure between seismic models and gravity models, despite clear spatial similarities of tectonic provinces in both models.

In this study, we present a crustal model calculated in spherical approximation for Antarctica from EIGEN-6C4 gravimetric data ( Förste et al. 2014). Furthermore, we also account for the gravity effect due to the vertical deformation caused by ice load on Antarctica, thus producing a Moho model on a fully relaxed lithosphere. Previous crustal models did not account for the effect of the ice load on the gravity signal, thus modelling the Moho on a vertically deformed lithosphere. We instead corrected the gravity effect which is caused by sediments, topography, ice cover and the vertical deformation on the gravity data. The corrected gravity data were inverted using regularized inversion based on Gauss–Newton’s formulation of Bott’s method (Bott 1960; Silva et al. 2014; Uieda & Barbosa 2017). The inversion approach combines Bott’s methods and two techniques of regularization and tesseraloids. Tesseraloids, which are spherical prisms (Uieda et al. 2015), were used instead of rectangular prisms for inversion and forward modelling. For a large landmass of approximately 14 million km$^2$ with the ice cover that reaches 4 km in thickness, the approach that considers the effect of ice load and the earth’s curvature in modelling the Moho provides a better approximation of the crustal model.

The paper is arranged in the following form: In Section 2, we discuss the tectonic setting of Antarctica. Data sets and methods used are summarized in Section 3. In this section, we present an overview of gravity reduction process and the regularizing inversion. We also show how the input parameters were estimated. The results of crustal thickness modelling and the uncertainties associated with the model are presented in Section 4. In Section 5, we compare our model to existing information on crustal thickness estimates and discuss them. We summarize our work in Section 6.

2 BRIEF OVERVIEW OF ANTARCTICA

Antarctica is the southernmost continent located south of the 60° latitude with a total area of ~14 million km$^2$ (Boger 2011). Antarctica is covered by a thick layer of ice and the bedrock is only exposed in few coastal regions (Fig. 1). The ice shield has a maximum thickness of ~4897 m and a mean ice thickness in Antarctica of ~1900 m (Fig. 1b; Lythe & Vaughan 2001; Fretwell et al. 2013). Antarctica has a stable cratonic block in EA and assemblages of crustal blocks in WA, divided by the Transantarctic Mountains (TAM). The identification of different tectonic blocks followed a correlation of exposed areas to coastal regions of Africa, India and Australia (Kroner 1993) and their rock affinities (Boger 2011). Furthermore, velocity and traveltime of seismic waves in lower crust and upper mantle identified EA as an Archean cratonic block with higher seismic velocities than WA, with the change in seismic velocities along the TAM (Eittreim 1994; Reading 2004; Winberry & Anandakrishnan 2004; Agostinetti et al. 2005; Lawrence et al. 2006).

EA contains ancient cratonic blocks of Precambrian origin (4600–541 Ma), formed from the amalgamation of Achaean nuclei, subglacial orogens and rifts. The geological structure of EA was formed through the three major tectonic events (Jacobs et al. 2003). These events are (1) the Grenvillian Orogeny (1.1 Ga) that resulted in the formation of supercontinent Rodinia (Rino et al. 2008), (2) the formation of Gondwana supercontinent between 610–510 Ma during the Ross/Pan-African event (Kriegsmann 1995) and (3) the breakup of Gondwana around 160 Ma (Boger 2011). Two main geographical features in EA are the orogenic mountains of the Dronning Maud Land and the Gamburtsev Subglacial Mountains (GSMs; Fig. 1a). Dronning Maud Land encompasses several crustal blocks ranging in age from Archean to Early Paleozoic. The GSMs are covered by ice with high bedrock elevations (~2500 m) and young Alpine-age topography (Lythe & Vaughan 2001). A set of subglacial basins, with a maximum depth to ~5 km, were also formed in EA including Wilkes Basin (WB, Fig. 1a) and Aurora Subglacial Basin (ASB, Fig. 1a).

WA consists of several crustal blocks including Marie Byrd Land (MBL, Fig. 1a), the Filchner–Ronne Ice Shelf (FRIS, Fig. 1a), the Ross Sea and Ross Ice Shelf (RIS, Fig. 1a), Ellsworth–Whitmore Mountains (EWM, Fig. 1a) and the Antarctic Peninsula. These crustal blocks were assembled into a single geologic unit (Fig. 1a; Dalziel & Elliot 1982). WA underwent Cenozoic rifting and Cenozoic episodes of extensive volcanism and tectonism (~60 Ma), which created a depression with lowering relative to EA (Behrendt et al. 1991). WA contains the West Antarctic
Figure 1. (a) Subglacial bedrock topography of Antarctica from Bedmap2 (Fretwell et al. 2013). Also shown are tectonic terranes of Antarctica: AB—Aurora Subglacial Basin; EL—Enderby Land; EWB—Ellsworth-Whitmore Mountains; FRIS—Filchner–Ronne Ice Shelf; GSM—Gamburtsev Subglacial Mountains; LR—Lambert Rift; MBL—Marie Byrd Land; nVL—northern Victoria Land; RIS—Ross Ice Shelf; TI—Thurston Island; VH—Vostok Highlands; WB—Wilkes Basin; WARS—West Antarctic Rift System. (b) Ice thickness of Antarctica. Also shown are the identified boundaries of West Antarctica and East Antarctica, and the location of seismic crustal thickness estimates from An et al. (2015). Subglacial bedrock topography and ice thickness data were downloaded from the Curtin University data repository (http://ddfe.curtin.edu.au/models/Earth2014/).

Rift System (WARS, Fig. 1a), which is one of the largest, enigmatic and complex continental extension in the world as assemblage of accreted terrains (Wörner 1996). WARS separates the elevated TAM and the uplifted area of MBL, a large intraplate volcanic province in Antarctica (Hole & LeMasurier 1994). The northernmost part of WA is the Antarctic Peninsula, with bedrock elevation of ∼2 km (Fretwell et al. 2013) which is believed to have formed by uplift and metamorphism of marine sediments during the late Palaeozoic and the early Mesozoic eras, ∼358–65 Ma (Bentley 1991).

The TAM is a mountain range between WA and EA which are ∼3500 km long and 200 km wide, with an elevation of ∼4 km above sea level (Fig. 1a). The only well-mapped boundary in Antarctica that separates EA and WA is also located along the mountain ranges (Finn et al. 2006). Despite this, the TAM are the largest mountains in the world to have formed in non-collisional way (ten Brink et al. 1997) with no evidence of a compressional origin (Studinger et al. 2004).

3 MATERIALS AND METHODS

3.1 Gravity data

This study used the global gravity model EIGEN-6C4, which is available up to degree and order 2109 ( Förste et al. 2014). EIGEN-6C4 is based on the Gravity Field and Ocean Circulation Explorer (GOCE) model that combines the GRACE data, the GOCE data, the DTU10 global model (Andersen et al. 2010) and the EGM2008 global model (Pavlis et al. 2012). The GOCE mission was launched in 2009 to determine the gravity field and geoid with high precision and spatial resolution of about 80 km (Floberghagen et al. 2011). It complements the GRACE mission, which is available up to the degree/order of 360 (e.g. Förste et al. 2008). EIGEN-6C4 incorporates data from the GRACE mission collected from February 2003 to December 2012 (Förste et al. 2014).

The EIGEN-6C4 raw data, defined as the magnitude of gradient of the potential calculated on or above the ellipsoid including the centrifugal potential (eqs 7 and 121–124 of Barthelmes 2014), were computed at 50 km height over ellipsoid as shown in Fig. 2(a). The height of 50 km over ellipsoid is a compromise between resolution and computational efficiency of the processing algorithm while maintaining tectonic features in the data as also discussed in Uieda & Barbosa (2017). This is mainly because the forward modelling algorithm becomes slower at lower altitude (Leonard Uieda, 2017, personal communication). The data were downloaded from the ICGEM website (http://icgem.gfz-potsdam.de/home) at a resolution of 0.5°. GOCE models in Antarctica are devoid of terrestrial gravity data, thus, having a lower resolution as in other continental areas (Hirt et al. 2016). Furthermore, the inclination of the GOCE satellite made it difficult to collect data south of 83.3° latitude, a region known as ‘Polar gap’, thereby, making only GRACE data as available satellite data in this region (van der Meijde et al. 2015b). GRACE and GOCE gravity data have a spatial resolution of about 115–125 and 70–80 km, respectively (van der Meijde et al. 2015b).
The downloaded gravity data were corrected for the effect of topography, ice and sediments as illustrated in Fig. 2 before they were used in the inversion process. Descriptive statistics for the values obtained during the gravity reduction are summarized in Table 1.
gravity effect due to the ice load and a correction for the gravity effect due to sedimentary basins. The latter is obtained using the algorithm after Uieda & Barbosa (2017). However, we rearranged the algorithm and introduced correction for the gravity effect of ice, and vertical deformation at the Moho boundary due to ice load and glacial isostatic adjustment (GIA). Sediment thickness data (1° resolution; Laske & Masters 1997; Laske et al. 2013) and topographic data (0.1° resolution; Hirt & Rexer 2015) were re-gridded to a spatial resolution of 0.5° to make them comparable with gravity data. We also applied a 77 km low pass filter to Earth2014 data (bedrock topography and ice thickness) before calculating the Bouguer effect from topographic map, 77 km being a maximum spatial resolution of GOCE models in Antarctica (Lübbers et al. 2017). Earth2014 (Hirt & Rexer 2015) uses the BEDMAP2 database (Fretwell et al. 2013) for bedrock and ice topography in Antarctica.

The scalar gravity of ellipsoidal reference earth (normal earth) was removed from the raw gravity at every point. Fig. 2(b) shows the resulting gravity disturbance calculated using the equation by Li & Götte (2001). This process removed all other signal, with only anomalous signals to the normal earth remaining. This anomalous signal, which is due to the gravitational attraction of topography, ice load and sedimentary basins, was removed from the gravity disturbance. In the process of calculating the Bouguer effect, the densities of the bedrock, the ice layer and the oceanic region surrounding the Antarctica landmass were set to 2670, 910 and 1040 kg m$^{-3}$, respectively. The Bouguer effect of topography (Fig. 2c) and glacial effect (Fig. 2d) were removed from the gravity disturbance. The complete Bouguer disturbance anomaly (Fig. 2e) results from the application of both the Bouger plate and the terrain correction for subglacial bedrock topography and ice thickness. Antarctica has a number of sedimentary basins with three layers as shown in CRUST1.0 (Laske et al. 2013). Gravity effect of the three sedimentary layers was separately calculated based on the density contrast between the crust and the upper mantle. $g^e(x_i)$ is the gravity anomaly, $g(x_i, \Delta p, p^{k-1})$ is the computed gravity anomaly at a point calculated from the Moho depth model, $A^e$ is a Jacobian matrix, $\mu$ is the regularization parameter, $\Delta p$ is an L $\times$ M finite-difference matrix representing L first-order differences between adjacent tesseraids and $p^k$ is the M-dimensional parameter vector containing M-Moho depths.

Table 1. Statistic summary of the gravity reduction process, values are in mGal and deformation values are in metres.

<table>
<thead>
<tr>
<th>Gravity disturbance</th>
<th>Minimum</th>
<th>Maximum</th>
<th>Mean</th>
<th>SD</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ice effect</td>
<td>149.41</td>
<td>3.51</td>
<td>54.57</td>
<td>46.01</td>
</tr>
<tr>
<td>Terrain effect</td>
<td>123.85</td>
<td>-360.34</td>
<td>-97.53</td>
<td>126.9</td>
</tr>
<tr>
<td>Bouguer disturbance</td>
<td>375.75</td>
<td>-193.73</td>
<td>30.51</td>
<td>164.06</td>
</tr>
<tr>
<td>Deformation</td>
<td>0</td>
<td>1156</td>
<td>-350</td>
<td>349</td>
</tr>
<tr>
<td>GIA effect</td>
<td>0</td>
<td>-30</td>
<td>-9</td>
<td>9</td>
</tr>
<tr>
<td>Effect of sediments</td>
<td>-101</td>
<td>-3.91</td>
<td>-22.81</td>
<td>16.87</td>
</tr>
<tr>
<td>Sediment free Bouguer disturbance</td>
<td>407.75</td>
<td>-182.33</td>
<td>53.32</td>
<td>172.53</td>
</tr>
</tbody>
</table>

The 3-D Moho topography of Antarctica was estimated using a regularized inversion based on Gauss–Newton’s formulation of Bott’s method (Bott 1960; Silva et al. 2014; Uieda & Barbosa 2017). The inversion has undergone three phases of methodological evolution. Bott (1960) introduced an iterative inversion approach for estimating a relief between basement and sedimentary layers from gravity values. Silva et al. (2014) extended the method into a Gauss–Newton formulation by introducing an identity matrix in the Jacobian matrix. Uieda & Barbosa (2017) transformed the ill-posed inverse problem into a well-posed matrix. They achieved this in two ways: (1) by replacing the rectangular prism with spherical prism or tesseraids, and (2) by introducing a regularization parameter in the inversion which smoothen the unstable solutions due to instability of the Bott (1960) method while maintaining the fit between solutions and data. The use of tesseraids in forward modelling is a milestone in continental scale studies as it considers the earth curvature, unlike the Parker-Oldenburg method that assumes a flat earth. The solution of the inversion at kth iteration is presented by the parameter perturbation vector (eq. 1), $\Delta p^k$:

$$\Delta p^k = A^{T} R^{-1} R A^{T} \left[ g^e(x_i) - g(x_i, \Delta p, p^{k-1}) \right] - \mu R^{-1} R p, i = 1, 2, ..., N, \tag{1}$$

where $\Delta p$ is the density contrast between the crust and the upper mantle, $g^e(x_i)$ is the gravity anomaly, $g(x_i, \Delta p, p^{k-1})$ is the computed gravity anomaly at a point calculated from the Moho depth model, $A^e$ is a Jacobian matrix, $\mu$ is the regularization parameter, $R$ is an L $\times$ M finite-difference matrix representing L first-order differences between adjacent tesseraids and $p^k$ is the M-dimensional parameter vector containing M-Moho depths.

### 3.3 Seismic crustal thickness constraints

Similar to other gravity inversion methods, the regularized Bott’s method suffers from inherent non-uniqueness, which requires prior information of the physical characteristics of the crust and upper mantle. To overcome the non-uniqueness problem, an approach of Van der Meijde et al. (2013) which validates gravity derived crustal thickness model with seismic crustal thickness estimates was adopted. There are two sets of independently compiled seismic points in Antarctica which can be used as constraints (Baranov & Morelli 2013; An et al. 2015). In this study, we used AN1-Moho as...
shown in Fig. 1(b) to estimate reference Moho depth and density contrast, besides validating the optimal crustal thickness model. AN1-Moho data are based on the compilation of three seismic studies (Baranov & Morelli 2013; Chaput et al. 2014; Feng et al. 2014). The compilation is robust as it adds more seismic point in WA from POLENET-ANET receiver functions (Chaput et al. 2014) and EA (Feng et al. 2014). Furthermore, the compilation removed poorly estimated points of the ANTMoho data set.

3.4 Estimating the inversion parameters

Required input parameters for the regularized Bott’s method are a reference Moho depth and a density contrast between the upper mantle and the lower crust, with an additional regularization parameter. These three parameters were estimated using the hold-out method (Kim 2009). We utilized the available AN1-Moho data and refined gravity data to estimate the optimum parameter values. Gravity data were first split into training and testing data to be used in this stage. The training data are a regular grid with data at every grid point, twice the original grid spacing, and the remainder constitutes the testing data. The training data were used to estimate the optimal inversion parameters. Conversely, the testing data were used to check the quality of the chosen regularization parameter while AN1-Moho estimates were used to select the optimal reference Moho depth and density contrast.

We estimated the optimal regularization parameter that smooths and stabilizes the inversion results while maintaining the fit with the observed data. We assume a randomly chosen Moho depth of 36 km and density contrast of 400 kg m$^{-3}$ during the inversion. The choice of these random input parameters do not affect the final choice of the regularization parameter (Uieda & Barbosa 2017). The optimal value of the regularization parameter is the one that best predicts the testing data, that is to say minimizes the mean square error (MSE). We set 32 values for regularization parameter on a logarithm scale between $10^{-12}$ and $10^{-6}$. Each of the 32 regularization parameters, together with a randomly chosen reference depth, was used to invert the training data. The inversion was followed by a forward modelling using tesseroid to check the fit between the predicted testing data due to the resolved Moho topography and the testing data. The optimum value of a regularization parameter was $10^{-3}$, with the lowest mean square error (MSE), calculated using eq. (2), between the two data sets:

$$\text{MSE}_0 = \frac{n^2}{N_{\text{test}}} \sum (d_{\text{test}} - d_{\text{test}}^0)^2,$$

where $d_{\text{test}}$ are the testing data, $d_{\text{test}}^0$ are the predicted testing data from forward modelling and $N_{\text{test}}$ is the number of points in the testing data, when used for estimating the regularizing parameter. When the reference Moho depth and density contrast are determined, $d_{\text{test}}^0$ is a vector of seismic crustal thickness data, $d_{\text{test}}$ is a vector of predicted crustal thickness value on the same location as obtained from forward modelling and $N_{\text{test}}$ is the number of crustal thickness data sets.

The estimated regularization parameter and AN1-Moho estimates were used to estimate the reference Moho depth and density contrast. We set 11 values of density contrast between 200 and 700 kg m$^{-3}$ after Tenzer et al. (2012) and Llubes et al. (2017) with a step of 50 kg m$^{-3}$. Previous studies have shown that the global density contrast between upper mantle and lower crust ranges from 280 to 480 kg m$^{-3}$ (Tenzer et al. 2012). However, Llubes et al. (2017) used a higher value of 630 kg m$^{-3}$ in the gravity inversion in Antarctica, which produced a ‘good’ model relatively to existing crustal thickness models. Thus, we set the upper bound of the density contrast range to 700 kg m$^{-3}$ to include all possible density contrasts from previous studies. Then, we set 15 values of reference Moho depths between 25 and 40 km with a step of 1 km, based on the average crustal thickness of EA and WA (Llubes et al. 2003, 2017; Block et al. 2009; Baranov & Morelli 2013; Baranov et al. 2017). This created 165 possible combinations of reference Moho depth and density contrast. Each of the 165 combinations together with the estimated optimum regularization parameter were used in the inversion of the training data. This produced 165 Moho topography models in which subglacial bedrock topography of Antarctica from Earth2014 (Hirt & Rexer 2015) was added to create 165 training data based crustal thickness models. The ice-load induced deformation (Fig. 2g) was added to each of the 165 Moho models before adding bedrock topography. This was done to make the gravity derived crustal thickness comparable to seismically derived crustal thickness, since seismic constraints were not corrected for the GIA deformation. A set of reference Moho and density contrast whose crustal thickness model minimizes the mean square error (MSE) with seismic crustal thickness estimates of AN1-Moho corresponds to optimal values, since they best fit the seismically derived crustal thickness. Then, we calculated the MSE for each of the 165 combinations. The lowest MSE calculated using eq. (1) had a reference Moho depth of 33 km and density contrast of 400 kg m$^{-3}$. We reduced the reference Moho depth values from 33 to 32 km to compensate for the GIA deformation, which can be up to 1 km (Fig. 2g; Steffen et al. 2017). The estimated three parameters were subsequently used to invert the refined gravity data using the regularized Bott’s method. The subglacial bedrock topography was added to the resolved Moho depth to produce a crustal thickness model of Antarctica.

4 RESULTS

4.1 Uncertainties of the input data and parameters

EIGEN-6C4 data in Antarctica have a spatial resolution of ~77 km due to the lack of terrestrial gravity data (Förste et al. 2014). This eliminates most high-frequency signal caused by intracrustal bodies, minor sedimentary basins and other near-surface bodies. Major sedimentary basins, topography and thick glacial bodies, however, have an effect on the gravity data. We eliminated their gravity effect in the data reduction process described in Section 3. Nevertheless, the input data and input parameters can introduce uncertainties in the recovered Moho depth and subsequently on the crustal model.

The use of either the BEDMAP (Lythe & Vaughan 2001) or BEDMAP2 (Fretwell et al. 2013) databases introduce uncertainties on the Moho of less than ±0.7 km (Baranov et al. 2017). Ice load in Antarctica changes the Moho depth due to vertical deformation of the lithosphere. The estimated lithospheric deformation is greater than 150 m but less than 1200 m. The relationship between lithosphere deformation and current GIA effect on the gravity data shows the same correlation as for Greenland (Steffen et al. 2017), though in Antarctica the deformation is greater. Since we corrected for this value, an estimated error of ~1.5 km was avoided which resulted in the recovery of Moho depth on a fully relaxed lithosphere. Effect of sedimentary basins on crustal modelling cannot be underestimated. Previous studies have removed sediments (Baranov et al. 2017) while others have not (O’Donnell & Nyblade 2014; Llubes et al. 2017). Block et al. (2009) discussed the negligible
effect of sediments in Antarctica. Baranov et al. (2017), however, showed that sediment thickness in Antarctica is considerable with a maximum thickness of 15 km, contributing an estimated uncertainty of 4 km on the Moho depth. Despite that thick sediments are mainly offshore, the uncertainty from onshore sediments can reach ~1.5 km, if the sediments are 3 km thick sediments (Baranov et al. 2017). This uncertainty is introduced due to the low (1°) resolution of the sediment data (Laske & Masters 1997) that derived using sparse seismograph collection and from the interpolation of the edges of the basins. The effect of sediment thickness, especially in marine sedimentary basins associated with thick sediments, can be as significant as ~80 mGal; with uncertainty of up to 4 km. Thus, we removed the effect of sediments in the Bouguer disturbance anomaly. The total contribution of the input data on Moho uncertainty is less than 2 km.

Gravity residuals between observed and calculated gravity data due to the Moho topography were less than ±5 mGal as shown in Fig 3(a) with a mean value of 0.12 mGal and a standard deviation of 2.56 mGal. The optimum value of a regularization parameter was 10^{-12} as shown in Fig. 3(b). Possible values between 10^{-12} and 10^{-5} produced good fit but 10^{-5} had the smallest MSE, at the same time being the largest regularizing parameter that fits the data with the predicted model. Values greater than 10^{-5} produce values that did not fit the data. Adjusting the 165 input parameter values against seismic constraints produced a reference Moho depth of 33 km and a density contrast of 400 kg m^{-3} as shown in Fig. 3(c). The seismic constraints have, on average, an error of ±3 km (Baranov & Morelli 2013; An et al. 2015), which is introduced in the resolved Moho during the parameter estimation. If we assume a mantle density of ~3300 kg m^{-3} and a density contrast of 400 kg m^{-3}, then our inversion assumed a mafic lower crust with a density of ~2900 kg m^{-3}. A change in the reference Moho by 1 km against the density contrast of 400 kg m^{-3} results in a baseline shift of the Moho depth by ~1 km (Fig. 4a) while varying the density contrast by 50 kg m^{-3} results in change of the spatial variations of the Moho topography (Fig. 4b). Changing both reference Moho and density contrast by 1 km and 50 kg m^{-3} results in an uncertainty of ~±1 km in the Moho model. Variance is highest and mostly positive in EA, and lowest and mostly negative in WA (Fig. 4c). In total, a Moho uncertainty of ±4 km due to seismic constraints and input parameters is obtained.

4.2 Crustal structure

The crustal thickness model mirrors the already existing information of a thinner WA and thicker EA separated by the TAM (Fig. 5). The crustal model has a spatial resolution of 0.5° with an uncertainty of ±6 km.

The WA is relatively thinner in comparison to the average continental crust of ~39 km. The thinnest crust in WA is located in the RIS of ~23 km west of the TAM. This thin crust extends into the marine Ross Sea in the north. It then connects to the ~25 km thin crust of WARSs through a relatively thicker crust (~29 km) between MBL and Ellsworth Mountains. Some regions, however, show a thicker crust in WA like MBL (~33–39 km), Ellsworth Mountains (~39–42 km) and the Antarctic Peninsula (~35 km). The thin crust extends southwest into FRIS (~25–30 km) and the Ronne Ice Shelf (~23 km). The thickness of the Antarctic Peninsula ranges from 37 km along the central part but thins out towards the coastal areas to ~30 km. The Ellsworth Mountains bridges thicker Antarctic Peninsula in the north and the TAM in the South.

EA is relatively thicker than WA. The East Antarctic average crustal thickness is similar to the global average of ~41 ± 6.2 km (Christensen & Mooney 1995). The crustal thickness of EA shows a depression in highly elevated areas especially in the Dronning Mud Land and the GSM. The thickest crust is located in the GMSs of ~48 km. The thick region of GMSs is connected to the coastal area by the thinnest crust, ~22 km, in EA located in Lambert Rift. Northeast of GSM is a relatively thick crust of ~46 km belonging to Vostok Highlands. Dronning Maud Land has a thick crust of ~44 km, which thins out towards the coast to around 33 km and then drops to 28 km indicating a possible sharp Moho. The coastal areas are relatively thinner, ~24–30 km, as show in Fig. 5, for the north Victoria Land, Wilkes Land and Enders Land.

The thickness along the TAM ranges from ~37 to ~46 km. In the northern part between 160° and 170° east, near northern Victoria Land, the crustal thickness is ~42 km. The thickest crust along the TAM is located near the South Pole, of about 48 km thickness. The southwestern part of the TAM between 40° and 70° west, in the FRIS, a crustal thickness inhomogeneity of ~21–34 km was obtained. This region extends into the Weddell Sea, where a marine sedimentary basin is found and it thins out to ~26 km.

5 DISCUSSION

5.1 Spatial comparison

A spatial comparison checks the reliability of spatial variations (Fig. 6a) of the results with existing models (Table 2) and the point observations (Fig. 6b). In such studies, it is known that ±6 km difference represents statistically comparable crust estimates (Van der Meijde et al. 2013; Uieda & Barbosa 2017), which is at the same level of uncertainty estimated in this study.

Seismic-only models show bigger differences than the seismic-gravity model. This was also shown by Baranov et al. (2017) though they noted a small difference in seismic constrained areas. The noted difference is mostly due to the improved data coverage in the combined model based on satellite gravity observations. A comparison to An et al. (2014) shows a clear boundary along the TAM in which EA is underestimated while WA is overestimated (Fig. 5b). Spatial difference within the confidence interval is ~70 per cent of the landmass with some region showing a better fit of ±3 km. The regions that are outside ±3 km are either in coastal areas or in sedimentary environments. They show a scattered pattern across the continent with a significant deviation of ~8 km noted on the northern edge of the FRIS in the Weddell Sea, the Lambert Rift and along the TAM.

We obtained similar discrepancies as reported in previous gravity studies of Antarctica against seismic observations (Block et al. 2009; Baranov et al. 2017; Llubes et al. 2017). The major difference is seen in the GMSs, where seismic points show a ~60 km thick crust while gravity data show a crust ranging from ~45 km to less than 50 km. Generally, EA shows a good correlation between the two data sets while WA shows discrepancies, especially in the Ross Sea, possibly due to the uncertainty introduced by the use of a single reference Moho depth of 29 km. Previous gravity modelling studies that used a single reference Moho exhibit a similar pattern (Llubes et al. 2003, 2017; Block et al. 2009), except the O’Donnell & Nyblade (2014) model who separately resolved crustal model for EA and WA. Inverting the data separately improves the fit with seismic observations but introduces a big discrepancy along the EA.
Figure 3. Parameter estimation results. Red rectangle representing the optimal values; (a) gravity residual; (b) estimated regularization parameter; (c) estimated reference Moho depth and density contrast based on AN1-Moho seismic constraints.

Figure 4. Variation in crustal thickness due to variation in parameters. (a) Change in crustal thickness due to contribution of a varying 1 km reference Moho depth; (b) change in crustal thickness due to contribution of a varying 50 kg m\(^{-3}\) density contrast; (c) change in crustal thickness due to the variation of both parameters.
and WA boundary; with more than 5 km compared to the single inversion of the whole continent.

5.2 Sedimentary basins

Sedimentary basins introduced uncertainties in the gravity data as discussed in Section 4.1. There are two sediment thickness models for Antarctica (Laske et al. 2013; Baranov et al. 2017). The Laske et al. (2013) model is based on a global sediment thickness map (Laske & Masters 1997) while Baranov et al. (2017) developed their model based on the seismic data and the BEDMAP2 subglacial bedrock relief. At best, onshore sediments from these models were estimated from sparsely distributed observations. Baranov et al. (2017) showed a notable sediment thickness of 14 km in the FRIS, where the sediments thickness of Laske & Masters (1997) indicates a 1.5 km sediment layer in the same region. The two models show a considerable difference of above 3 km in the Lambert Rift, Vostok Basin and on the RIS (Baranov et al. 2017). In our study, we stripped the sediments effect from the gravity data using the Laske et al. (2013) model. Baranov et al. (2017), however, used their sediment model to correct for the effect of sediments on the Bouguer disturbance. Nevertheless, we suggest the use of CRUST1.0 model for onshore sediment stripping in the gravity data than the Baranov et al. (2017). The reason is that Huebscher et al. (1996) and Huebscher et al. (1996) found the sediment thickness to be between 2 and 15 km along Weddell Sea shoreline boundary. Interpolating these values inland, as in Baranov et al. (2017), created a 15 km thick, inland subglacial basin on the FRIS. This could introduce an effect on the gravity data, adding uncertainty on the resolved Moho model. A comparison of our model with the Baranov et al. (2017) model shows a significant difference in sedimentary environments. Specifically, the FRIS with a 10 km thick crust and an elevation of 1.2 km below sea level has a similar thickness to oceanic crusts at an elevation of 5 km below sea level (Baranov et al. 2017). The crustal thickness estimation in Baranov et al. (2017) was possibly underestimated due to the thick sediment layer they used in their research. This is the reason why we applied the sediment stripping on the gravity data using a global sediment map.

Figure 5. Crustal thickness model of Antarctica obtained using a reference Moho depth of 32 km and density contrast of 400 kg m\(^{-3}\) with a resolution of 0.5\(^{\circ}\). Values less than 15 km are not included as these are mostly related to oceanic crust. Also shown are tectonic terranes of Antarctica (see Fig. 1 for the list of abbreviations).
A crustal thickness model of Antarctica

Figure 6. (a) Comparison of our model with CRUST1.0. (b) Differences between our crustal thicknesses and compilation of receiver functions selected by AN1-Moho.

Table 2. Descriptive statistics of model difference.

<table>
<thead>
<tr>
<th>Source</th>
<th>Model difference</th>
<th>Min (km)</th>
<th>Max (km)</th>
<th>Mean (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Our model</td>
<td>C1-CRUST1.0</td>
<td>-5.5</td>
<td>18.2</td>
<td>2.2</td>
</tr>
<tr>
<td></td>
<td>C1-Combined model</td>
<td>-16</td>
<td>22</td>
<td>2.7</td>
</tr>
<tr>
<td>(Llubes et al. 2017)</td>
<td>CRUST1.0–AN1</td>
<td>-26.0</td>
<td>19.0</td>
<td>-0.3</td>
</tr>
<tr>
<td>(Baranov et al. 2017)</td>
<td>GOCE–CRUST1.0</td>
<td>-4.5</td>
<td>29.0</td>
<td>1.2</td>
</tr>
<tr>
<td></td>
<td>Seismic-CRUST1.0</td>
<td>-14</td>
<td>18</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Seismic-CRUST2.0</td>
<td>-14</td>
<td>22</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Seismic-ANTMoho</td>
<td>-22</td>
<td>20</td>
<td></td>
</tr>
</tbody>
</table>

5.3 West Antarctica

The crustal structure beneath WA is well studied using seismic data (Chaput et al. 2014; Shen et al. 2018). The thin regions of WA correlate well with the troughs and lowly elevated topography, which indicate extensional thinning of the crust. Winberry & Anandakrishnan (2004) suggested that these thin crusts were formed during a focused crustal extension on a weak pre-existing lithospheric zone. Crustal structure results show a correlation of the Antarctica shoreline with other continents. For instance, the coastal areas of the Antarctic Peninsula have a thick crust of ~30 km, which is similar to the ~33 km thick crust in the Patagonia Province in South America (Van der Meijde et al. 2013; Uieda & Barbosa 2017), suggesting a once connected crustal body of the Gondwana Supercontinent. We argue that there is no strong isostatic compensation in Marie Baud Land as compared to EA in which high elevation areas with ~km above sea level have crust thicknesses greater than 45 km, for example, the GSMS. The thick crust in Marie Baud Land was also apparent in other seismic models (Chaput et al. 2014; Shen et al. 2018) and gravity derived crustal structures (e.g. Block et al. 2009) while some seismic models found a thin crust (e.g. Baranov & Morelli 2013).

5.4 East Antarctica

EA is relatively thicker than WA with average crustal thickness similar to global average values of ~41 ± 6.2 km (Christensen & Mooney 1995). The crustal thickness pattern resembles the amalgamation of different crustal blocks with complex tectonic processes. EA has had significant tectonic activities, thus the remnants of Gondwana and Rodinia evolution are preserved and can be seen in the crustal thickness. Moreover, the crustal model in EA shows a depression in highly elevated areas especially in the Dronning Maud Land and the GSM; indicating an isostatically compensated crust based on an Airy isostasy model (Airy 1855). Previous studies have indeed shown that the elevated areas in EA have sufficient crustal roots to support the isostatic theory (O’Donnell & Nyblade 2014). The thickest crust beneath the GSMS, ~48 km, is less than some previous estimations which found a thickness of more than 50 km (O’Donnell & Nyblade 2014), ~55–58 km (Hansen et al. 2010).
and ~55 km (Baranov & Morelli 2013) but similar to other gravity studies (Block et al. 2009; Baranov et al. 2017). The relatively thinner crust beneath the GSMs in our study exhibits the effect of uncertainties in the input data and parameters resulted in overcompensation in WA and underestimation in EA. The thick crust of Dronning Maud Land has well-compensated crustal roots relative to the surrounding crust, which supports the isostatic balanced theory. Before the breakup of the Gondwana Supercontinent, shoreline crusts surrounding Dronning Maud Land were connected to the southern tip of Africa, as noted in the similarities of crustal structure to the resolved crust of Southern Africa of ~33 km (Nguuri et al. 2001; Kgawane et al. 2009).

6 CONCLUSION

We have produced a continental-scale crustal model for Antarctica using the regularized inversion based on Gauss–Newton’s formulation of Bott’s method. The methodology combines the improved Bott’s method with the regularization and the tesseroid techniques. Thus, allows construction of a crustal model that is smooth but fits the data calculated in spherical approximation that accounts for the earth’s curvature. Our model produced a Moho on a fully relaxed lithosphere, after considering the gravity effect of the deformation of the lithosphere due to ice load. We also show that an average thicker crust in EA and a thinner one in WA are similar to spatial patterns of previous models. The thickest crust is located beneath the GSMs with a Moho depth of over 40 km. Moreover, other thicker crusts are also noted beneath the TAM and Dronning Maud Land. Comparison of our model with point observation data, CRUST1.0, and the combined seismic-gravity model show a good correlation in well-constrained areas, with significant differences noted in sedimentary basins. We also note a trade-off of prior information between the reference Moho and density contrast in gravity inversion which introduced uncertainties in the Moho depth estimates.

Despite the improvement on the existing model of Antarctica, a lot of work is needed to produce a more robust crustal model. For instance, our study introduced steps on gravity reduction process in Antarctica where the loading of the massive ice cover on the vertical deformation of the lithosphere could not be neglected. But uncertainties still exist due to the limitation of the input data. First, when gravity data with degree and order 2019 for Antarctica become available, the spatial resolution of the crustal model will be greatly improved. Besides, some important effects on gravity data from sediments need further consideration as the currently adopted sediment models contain uncertainties and interpolation along the basin edges. As depicted in this study, the approach that exploits the joint inversion of seismic and gravity data can improve the crustal structure estimates in Antarctica particularly in seismologically constrained areas. More measurements of seismic observations can therefore improve the modelling, though it will be decades before the Antarctic continent is more evenly covered.

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