The 3D crustal structure of the Wilkes Subglacial Basin, East Antarctica, based on joint inversion of gravity and magnetic data using variation of information

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July 20, 2023

Abstract

Direct geological information in Antarctica is limited to ice free regions along the coast, high mountain ranges or isolated nunataks. Therefore, indirect methods are required to reveal subglacial geology and heterogeneities in crustal properties, which are critical steps towards interpreting geological history. We present a 3D crustal model of density and susceptibility distribution in the Wilkes Subglacial Basin and the Transantarctic Mountains (TAM) based on joint inversion of airborne gravity and magnetic data. The applied "variation of information" technique enforces a coupling between gravity and magnetic sources to give an enhanced inversion result. Our model reveals a large-scale body located in the interior of the Wilkes Subglacial Basin interpreted as a batholithic intrusive structure, as well as a linear dense body at the margin of the Terre Adélie Craton. Density and susceptibility relationships are used to inform the interpretation of petrophysical properties and the reconstruction of the origin of those crustal blocks. The petrophysical relationship indicates that the postulated batholitic intrusion is granitic, but independent from the Granite Harbour Igneous Complex previous described in the TAM area. Emplacement of a large volume of intrusive granites can potentially elevate local geothermal heat flow significantly. Finally, we present a tectonic evolution sketch based on the inversion results, which includes development of a passive continental margin with seaward dipping basalt horizons and magmatic underplating followed by two distinct intrusion events in the Wilkes Subglacial Basin with Pan-African ages (700 - 551 Ma) and Ross ages (550 - 450 Ma).

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Key Points:

- Crustal density and susceptibility distribution model based on joint inversion of gravity and magnetic data using variation of information
- Crustal heterogeneities related to intrusive crustal block and the craton margin are revealed using density and susceptibility relationship
- A new tectonic evolution sketch includes two intrusion events, which are separated in time and space.

Abstract
Direct geological information in Antarctica is limited to ice free regions along the coast, high mountain ranges or isolated nunataks. Therefore, indirect methods are required to reveal subglacial geology and heterogeneities in crustal properties, which are critical steps towards interpreting geological history. We present a 3D crustal model of density and susceptibility distribution in the Wilkes Subglacial Basin and the Transantarctic Mountains (TAM) based on joint inversion of airborne gravity and magnetic data. The applied "variation of information" technique enforces a coupling between gravity and magnetic sources to give an enhanced inversion result. Our model reveals a large-scale body located in the interior of the Wilkes Subglacial Basin interpreted as a batholithic intrusive structure, as well as a linear dense body at the margin of the Terre Adélie Craton. Density and susceptibility relationships

are used to inform the interpretation of petrophysical properties and the reconstruction of the origin of those crustal blocks. The petrophysical relationship indicates that the postulated batholitic intrusion is granitic, but independent from the Granite Harbour Igneous Complex previous described in the TAM area. Emplacement of a large volume of intrusive granites can potentially elevate local geothermal heat flow significantly. Finally, we present a tectonic evolution sketch based on the inversion results, which includes development of a passive continental margin with seaward dipping basalt horizons and magmatic underplating followed by two distinct intrusion events in the Wilkes Subglacial Basin with Pan-African ages (700 - 551 Ma) and Ross ages (550 - 450 Ma).

Plain Language Summary

Most rocks in Antarctica are hidden beneath a thick icesheet. Therefore, indirect techniques are required to reveal rock provinces below the ice and within Earth's crust. Rocks simultaneously influence the gravity and magnetic field through their physical properties (density and susceptibility). Here we use both the gravity and magnetic field to reveal rock provinces beneath the ice and use the relationship between density and susceptibility of the rocks to interpret the distribution of granitic rocks in the area of the Transantarctic Mountains and the Wilkes Subglacial Basin in East Antarctica. Granitic rocks can lead to elevated heat flow due to radiogenic decay of minerals within the rock and influence the overlaying icesheet. Lastly, we use our subsurface model of rock provinces to speculate on the tectonic evolution of the region.

1 Introduction

The Wilkes Subglacial Basin (WSB) is located between the Transantarctic Mountains (TAM) and the Terre Adélie Craton (Figure 1). It was first described based on radar data in the 1970s[*Drewry*, 1976] and stretches ca 1600 km from the George V Coast towards the South Pole, while its width decreases from ca 600 km close to the George V coast [*Ferraccioli et al.*, 2009b] to < 100 km towards the South Pole [*Studinger et al.*, 2004].

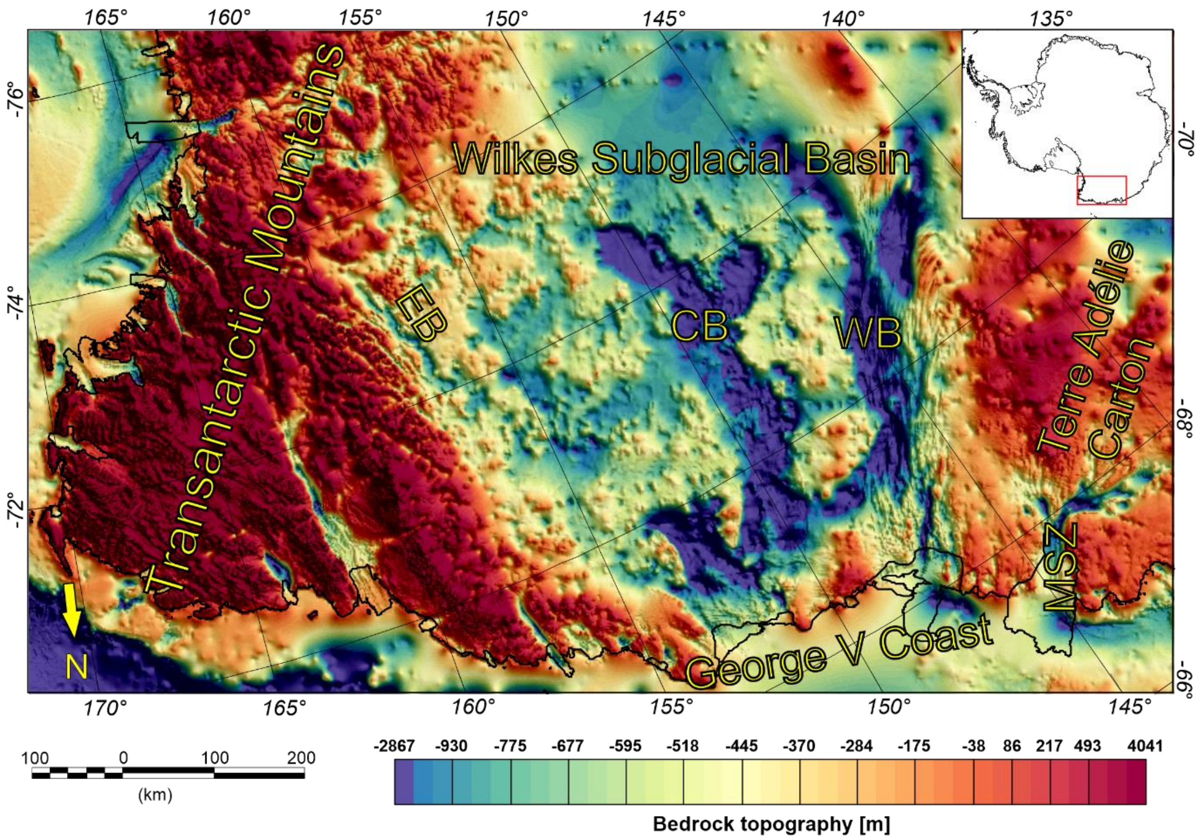

Figure 1: Bedrock topography of the Transantarctic Mountains and Wilkes Subglacial Basin (WSB) from the Bedmachine model version 2 [Morlighem et al., 2020]. EB: Eastern Basin; CB: Central Basin; WB: Western Basin; MSZ: Mertz shear zone. Black lines mark ice grounding lines and ice shelf extents from the SCAR Antarctic Digital Database.

The WSB hosts one of the largest areas of bed topography below sea level in East Antarctica, reaching depths of more than 2 km below sea level within the locally more deeply incised sub-basins [*Morlighem et al.*, 2020] (Figure 1). Such sub sea level basins pose a potentially high, but poorly understood, risk for the stability of the East Antarctic Ice Sheet (EAIS) and therefore for future sea level rise, as they are more vulnerable to melting by warming of the adjacent ocean. Such melting could potentially trigger mechanisms of unstable retreat [*Pollard et al.*, 2015; *Schoof*, 2007]. Recent studies suggest a significant long-term contribution from the WSB region to sea-level rise within the next two centuries accompanied by major retreat of the ice sheet in the WSB region by the year 2500 [*DeConto and Pollard*, 2016; *Stokes et al.*, 2022]. For the WSB itself, competing models for its evolution have been proposed since its discovery including a rift basin[*Steed*, 1983], extended terrane [*Ferraccioli et al.*, 2001] or flexural down-warp of cratonic lithosphere as a consequence of the TAM uplift [*Stern and ten Brink*, 1989; *ten Brink et al.*, 1997]. However, the modern landscape formation of the WSB is believed to result from lithospheric flexure associated with the TAM uplift combined with glacial erosion [*Ferraccioli et al.*, 2009a; *Jordan et al.*, 2013; *Paxman et al.*, 2018; *Paxman et al.*, 2019]. The geology of the WSB remains disputed as the occurrence of direct geological samples is limited to ice free regions along the coast, or isolated nunataks, while the origin of geological material transported to the coast by glaciers is often ambiguous. Adjacent to the Mertz shear zone, on the side of the Terre Adélie Craton, ≥2440 Ma old paragneiss and granitoids are exposed, while on the WSB side of the Mertz shear zone ca. 500 Ma old granites have been mapped[*Goodge and Finn*, 2010]. Aeromagnetic measurements in the WSB have been used to infer the presence of Beacon Supergroup sedimentary strata intruded by rocks of the Ferrar Large Igneous Province [*Ferraccioli et al.*, 2009a]. Geological interpretations of the interior of the WSB are mainly derived from radar, gravity and magnetic airborne measurements (e.g.[*Jordan et al.*, 2013]). A prominent positive magnetic anomaly exists in the central WSB the origin of which is hypothesised to be either an intrusive arc associated with subduction [*Ferraccioli et al.*, 2009a], or a thinned crust as a result of rifting [*Ferraccioli and Bozzo*, 2003]. Another prominent feature is a positive linear gravity anomaly associated with the craton margin, which was interpreted as up-thrusted crustal material along the craton flank [*Studinger et al.*, 2004].

The neighbouring Transantarctic Mountains are the largest non-contractional mountain range on Earth, separating the warmer lithosphere of the Cretaceous-Tertiary West Antarctic rift system and the colder and older provinces of East Antarctica [*Morelli and Danesi*, 2004; *Robinson and Splettstoesser*, 1986; *ten Brink and Stern*, 1992]. Direct geological information is richer in the TAM compared to WSB since more rock outcrops are present in the high mountain range. The five dominating Ross Orogen geological units in the TAM are from east to west the Robertson Bay Terrane, the Millen Schist, the Bowers Terrane, the Wilson Terrane and Granite Harbour Igneous Complex, which are all intruded by the Jurassic Ferrar sill complex and in places overlain by the associated extrusive Kirkpatrick basalts [Estrada et al., 2026 and references therein]. Zircon-age dating from sedimentary rocks shows a decrease in age from the Wilson Terrane eastwards to the Robertson Bay group from Pan-African age (551 – 700 Ma) to Ross age (450 -550 Ma) [*Estrada et al.*, 2016]. The age of the Granite Harbour Igneous Complex relates to the Ross Orogeny [*Estrada et al.*, 2016], while the source for the Pan African material remains speculativeThe subglacial geology in the WSB and TAM region is largely hidden beneath a 2-3 km thick ice sheet. Understanding the subglacial geology and crustal properties is crucial to constrain the influence of the solid Earth on the stability of the overlying ice sheet. Radiogenic heat production is predicted to contribute up to 40% to the surface geothermal heat flow [*Artemieva and Mooney*, 2001; *Haeger et al.*, 2022; *Hasterok and Chapman*, 2011]. However, due to the lack of information on subglacial geology and crustal properties, incorporating accurate thermal crustal parameters such as radiogenic heat production and thermal conductivity is challenging and therefore current geophysical derived geothermal heat flow models commonly use global average values instead [*Haeger et al.*, 2022; *Lösing and Ebbing*, 2021; *Lowe et al.*, 2023; *Martos et al.*, 2017; *Shen et al.*, 2020; *Stål et al.*, 2021]. For more in depth discussion of current Antarctic geothermal heat flow models the reader is referred to[*Reading et al.*, 2022] and [*Burton-Johnson et al.*, 2020].

The objectives of this study are to identify crustal structures, crustal geological provinces and intrusive bodies and constrain their dimension in 3D, and also to identify the Terre Adélie Craton boundary. For this purpose, we conduct joint inversion of gravity and magnetic data to obtain petrological parameter distribution in terms of density and susceptibility. The geophysical inversion uses the joint inversion framework JIF3D [*Moorkamp et al.*, 2011], which uses the Variation of Information (VI) to introduce a coupling between the inverted density and susceptibility sources [*Lösing et al.*, 2023; *Moorkamp*, 2021; 2022]. Subsequently, the density and susceptibility relationship of the inversion model is used to identify crustal rock provinces and infer potential rock types. The geophysical and petrophysical interpretation of the inversion results is the basis for our tectonic evolution model.

2 Method

2.1 Joint inversion of gravity and magnetic data

Joint inversion of gravity and magnetic data is carried out in JIF3D [*Moorkamp et al.*, 2011]. JIF3D is a 3D joint inversion framework for geophysical data sets including magnetotelluric, seismic, magnetic data as well as scalar and tensor gravity data. JIF3D utilizes a limited memory quasi-Newton approach [*Avdeev and Avdeeva*, 2009] for optimization. For a complete mathematical description of the JIF3D inversion framework the reader is referred to [*Moorkamp*, 2021; 2022; *Moorkamp et al.*, 2011].

Inverting simultaneously for crustal density and susceptibility distribution using gravity and magnetic data in a joint inversion framework is well established [*Bosch et al.*, 2006; *Fregoso and Gallardo*, 2009; *Frey and Ebbing*, 2020; *Gallardo-Delgado et al.*, 2003; *Guillen and Menichetti*, 1984; *Shamsipour et al.*, 2012]. However, joint inversion based on Variation of Information (VI), which allows the coupling of physical parameters, has only recently became popular in geophysical joint inversion applications [*Haber and Holtzman Gazit*, 2013; *Lösing et al.*, 2023; *Mandolesi and Jones*, 2014; *Moorkamp*, 2021; 2022]. VI is related to the concept of mutual information (MI) [*Moorkamp et al.*, 2011]. VI describes the amount of shared information contained in two variables, meaning a low VI value indicates that both variables are dependent, while a high VI value indicates that information about variable 1 does not reveal meaningful information about variable 2 [*Lösing et al.*, 2023; *Mandolesi and Jones*, 2014; *Moorkamp*, 2021; 2022]. Variation of Information is defined as:

$$VI(x,y) = 2H(x,y) - H(x) - H(y) \quad (1)$$

Here $H(\mathrm{x}) = -\sum_i p(\mathrm{x_i}) \log p(\mathrm{x_i})$ is the Shannon Entropy; $p(\mathrm{x_i})$ is the probability density approximated by kernel methods [*Mandolesi and Jones*, 2014]; $H(\mathrm{x})$ and $H(y)$ are the marginal entropies and $H(\mathrm{x,y})$ is the joint entropy[*Lösing et al.*, 2023; *Moorkamp*, 2021; 2022]. VI is incorporated into the objective function Φ_{joint} as:

$$\Phi_{joint} = \Phi_{d,grav} + \Phi_{d,mag} + \lambda_1 \Phi_{reg,\rho} + \lambda_2 \Phi_{reg,sus} + \lambda_3 \Phi_{VI} (2)$$

where $\Phi_{d,grav}$ and $\Phi_{d,mag}$ are the Root-Mean-Square (RMS) misfit between observed and inverted gravity and magnetic data; $\Phi_{reg,\rho}$ and $\Phi_{reg,sus}$ are regularisation terms for the density and susceptibility distribution, controlling the smoothness of the inverted model, Φ_{VI} is a coupling term which includes Variation of Information of recovered density and susceptibility and λ represents

the weighting factors of the individual terms.

VI inversion has been successfully applied to magnetotelluric and seismic data [*Mandolesi and Jones*, 2014], magnetotelluric and gravity data [*Moorkamp*, 2021; 2022], and gravity and magnetic data [*Lösing et al.*, 2023]. [*Lösing et al.*, 2023] demonstrated that VI inversion can recover density and susceptibility distribution, if it is assumed that the gravity and magnetic signals have an identical source, by performing VI inversion tests on synthetic data.

2.1.1 Inversion setup

The density and susceptibility inversion models are discretised into meshes with equal horizontal cell sizes of 7.5 km. The vertical cell size is 1 km at the surface and increases with depth by a factor 1.1 for each cell. Increasing vertical cell size with depth is introduced to account for decreasing resolution with increasing distance to the source in potential field applications. A horizontal cell size of 7.5 km is chosen because the input gravity and magnetic data have a resolution of 10 km (see section 3). A cell size of 7.5 km allows the inversion to adjust on average more than one cell to fit the inversion data but is also not so small that it does not represent the resolution of the input data or introduce artifacts. Additionally, a padding area of 20% around the study area is added to avoid edge effects. The resulting inversion mesh contains 244 cells east-west direction, 140 cells in north-south direction and 21 cells in the z direction (60 km below sea level). To constrain the model geometry, the bedrock topography from BedMachine version 3 [*Morlighem et al.*, 2020], Curie point depths (CPD) [*Lowe et al.*, 2023], and Moho depths [*Pappa et al.*, 2019a] are used.

The aim of the VI inversion is to invert jointly the gravity and magnetic field for the density and susceptibility of the crust. Only cells between the bedrock interface and the Moho interface are allowed to vary during the inversion. However, rocks lose their magnetic properties at the Curie temperature. Therefore, the Curie Depth Points from [*Lowe et al.*, 2023] are introduced as an additional constraint. Joint VI inversion is carried out for all cells between the bedrock interface and the CPD interface. Between the CPD interface and the Moho interface only density inversion is carried out and no susceptibility values are obtained for this section of the crust. The input gravity field for the density inversion is corrected for masses below the Moho interface by the Antarctic lithospheric model from Pappa et al., (2019b) (see section 3.1). All cells are set to a starting value of 0 kg/m^3 for the density anomaly model and 0 SI for the magnetic model and are iteratively updated during the VI inversion process. The first inversion run is performed with a high coupling weight in order to enforce a tight coupling between the inverted models. Subsequently, a second inversion run is carried out, which uses the resulting density and susceptibility values from the first inversion run as a starting model but a lower coupling weight is applied. The rationale behind using two different coupling weights is that a tighter coupling in the first inversion run favours geometrical structures in both the density and susceptibility, while a second inversion run with a lower coupling provides the inversion algorithm with more freedom to fit the observed gravity and magnetic field better [*Moorkamp*, 2022]. Coupling values were established by trial and error. The induced magnetic field strength is set to 64981 nT with an inclination of -84.4° and a declination of 147.4° based on the definitive magnetic reference field (DGRF) for a longitude of 155° and a latitude of -73° for the year 2005 (the year of the aerogeophysical survey WISE-ISODYN, which is the main survey in the study area). Additional parameter settings of the inversion are given in table 1.

Table 1 Inversion parameters

Coupling	25,000 run 1 and 15,000 run 2

Regularization Density and Susceptibility	10
Error Gravity [mGal]	2 mGal
Error Magnetics [nT]	15
Min. / Max Density [kg m^{-3}]	± 250
Min. / Max Susceptibility [SI]	± 0.1
Magnetic Field strength [nT]	64981
Inclination [°]	-84.4
Declination [°]	147.4
dens_covmod (Maximum depth of density variation).	Moho depth
sus_covmod (Maximum depth of susceptibility variation).	CPD depth
coupling_validity (Maximum depth where coupling is assumed).	CPD depth

Subsequently the relationship between the inverted relative densities and susceptibilities are used to characterize 3D crustal structures and to identify crustal units with similar relationships.

3 Data

This section describes the geophysical data and the boundary condition used during the VI inversion to obtain a 3D crustal density and susceptibility distribution model.

3.1 Gravity data

Bouguer gravity anomaly data are taken from AntGG [*Scheinert et al.*, 2016]. The AntGG gravity compilation includes airborne, terrestrial, and shipborne measurements, which are provided as a grid with a grid spacing of 10 km. Data gaps in AntGG are filled in with Bouguer anomaly data from the Ganovex VII – ItaliAntartide XV survey [*Reitmayr et al.*, 2003] and recent ground station measurements conducted within the Italian National Program for Antarctic Research activities [*Zanutta et al.*, 2018] (Figure S1). This data compilation is gridded with a grid spacing of 10 km and a blanking distance of 40 km using the minimum curvature gridding function in Oasis Montaj to produce the Bouguer anomaly grid of the WSB and TAM region. The Bouguer anomaly grid is subsequently upward continued to a constant observation height of 10 km (Figure 2a) using the Compudrape algorithm in Oasis Montaj. The Bouguer anomaly map shows a large negative signal in the TAM region, where a crustal root is present beneath the mountain range. A prominent positive linear feature exists in the Bouguer anomaly map following the western edge of the Wilkes Subglacial Basin and might indicate the transition zone at the margin of the Terre Adélie Craton [*Goodge and Finn*, 2010]. To remove the gravity contribution below the Moho interface the gz gravity component from the Antarctic continent-wide forward lithospheric model from Papa et al. (2019b) was subtracted (Figure S2). This model contains a homogeneous crust and a variable mantle. By subtracting this model, we corrected for a variable mantle and Moho depth variation and therefore the resulting residuals belong to the crust. The lithospheric model is presented at 10 km observation height and is the reason all input data are upward continued to this constant height. In the resulting residual gravity map (Figure S2) the large-scale negative signal beneath the TAM is absent, while the relative positive linear crustal structure is preserved. Furthermore, the mean value (92.5 mGal) of the residual gravity field is removed to shift the residual gravity field to a mean level of 0 mGal (Figure 2b).

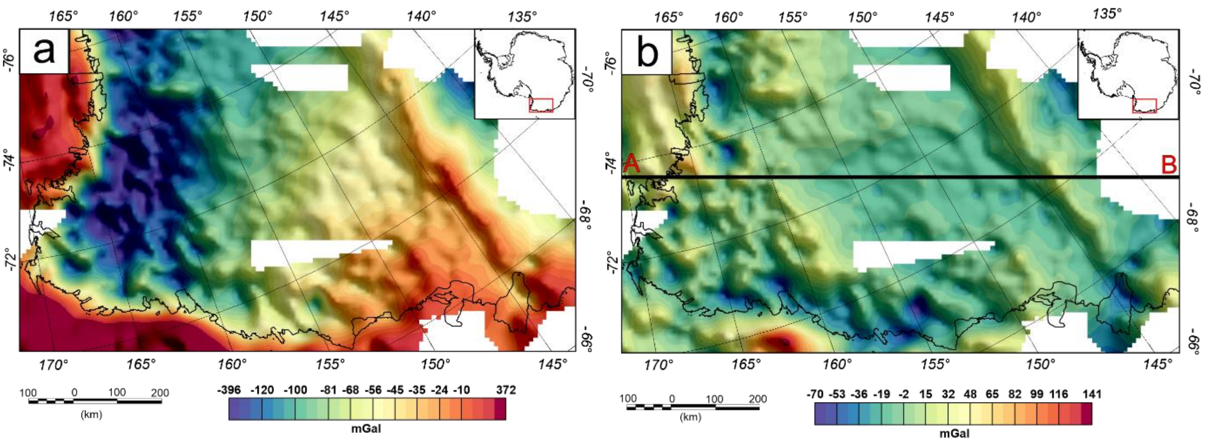

Figure 2 a) Bouguer Anomaly compilation including AntGG data [Scheinert et al., 2016] and regional gravity data [Reitmayr et al., 2003; Zanutta et al., 2018]. b) Residual gravity map obtained by subtracting gz gravity component from the lithospheric model of [*Pappa et al.*, 2019b] and then subtracting the mean value of the residual field. Both gravity grids have a grid spacing of 10 km. Black line indicates location of cross section profile in Figure 8.

3.2 Magnetic data

The ADMAP-2 compilation includes 3.5 million line-km of aeromagnetic and marine magnetic data in Antarctica and the Southern Ocean south of 60 °S [*Golynsky et al.*, 2018]. The gridded ADMAP-2 product (Figure 3a) has a grid spacing of 1.5 km and its production included subtraction of the International Geomagnetic Reference Field, diurnal effects correction, high-frequency error correction, levelling, regional gridding, and merging of regional grids into a continent-wide compilation [*Golynsky et al.*, 2018]. For the inversion, the magnetic data from the ADMAP-2 compilation is regridded with a grid spacing of 10 km to match the resolution of the gravity data (section 3.1). Subsequently the magnetic data are also continued upward to a height of 10 km (Figure 3b) to be consistent with the gravity data. The upward continuation of the magnetic data functions as a lowpass filter, removing high frequency content, while broad magnetic anomalies are preserved. The magnetic anomaly grid shows a broad linear feature orientated southeast – northwest (F1) in the central part of the WSB the origin of which has been hypothesised to be a failed rift or an arc-related intrusive body [*Ferraccioli and Bozzo*, 2003; *Ferraccioli et al.*, 2009b]. Perpendicular to this features a positive magnetic anomaly is visible (F2). A strong linear anomaly exists west to the WSB towards the craton margin (F3), where the magnetic anomaly rapidly increases to values of ~1500 nT compared to the dominating ± 300 nT range in the WSB and TAM area. Smaller magnetic anomalies are observed offshore to the east (F4) and along the Rennick Graben and Matusevich glacier region (F5).

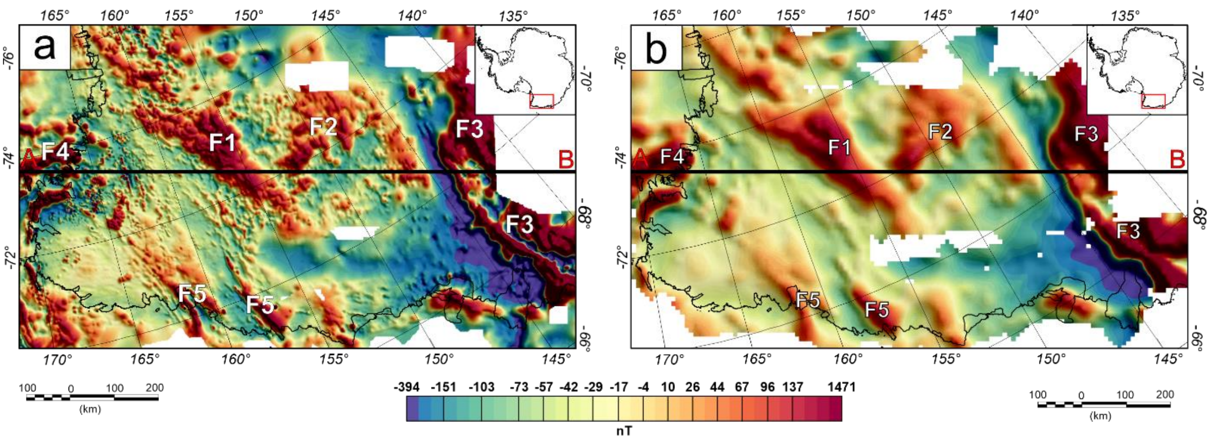

Figure 3 a) Magnetic anomaly grid from ADMAP-2 [Golynsky et al., 2018] with a grid spacing of 1.5 km. b) Regridded ADMAP-2 magnetic data with a grid spacing of 10 km and upward continued at a constant height of 10 km to match the resolution and upward-continued height of the gravity compilation. Black line indicates location of cross section profile in Figure 8. Features F1-F5 relate to magnetic anomalies discussed in the main text.

3.3 Moho depth and Curie depth.

Moho depth estimates (Figure 4a) are taken from satellite gravity inversion [*Pappa et al.*, 2019a]. The Moho interface is used as the bottom boundary condition for the density inversion.

CPD estimates are taken from [*Lowe et al.*, 2023] and used as a bottom boundary condition for the susceptibility inversion. Below the CPD the crustal rocks have lost their magnetic properties and therefore the joint inversion of magnetic and gravity data needs to be limited to crustal depth above the CPD. The CPD estimates from [*Lowe et al.*, 2023] have a 20 km resolution and are interpolated on a 7.5 km grid, matching the cell size of the inversion mesh, by applying statistical kriging using the python package PyKrig [*Murphy et al.*, 2022]. After interpolation, the CPD map shows some values deeper than the Moho depths from Pappa et al. (2019a) in the offshore area and along the coast. A CPD below the Moho would indicate that the upper mantle is magnetic and although this possibility has been suggested [*Ferré et al.*, 2014]), we discard CPD values below the Moho boundary (Figure 4b).

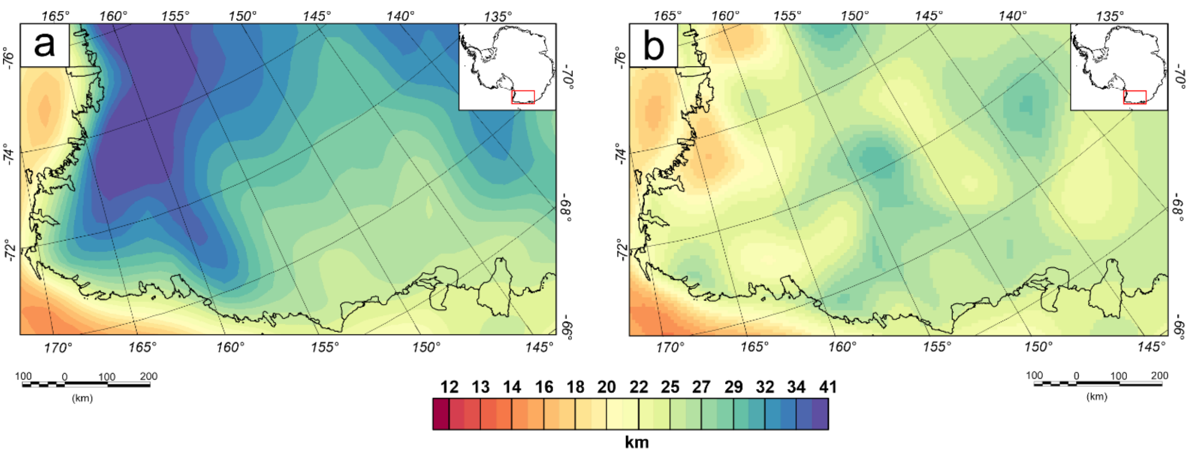

Figure 4: a) Moho depth map derived from Satellite gravity measurements [Pappa et al., 2019a]. b) Curie point depths estimated [Lowe et al., 2023] based on kriging interpolation, clipped to be shallower than the Moho.

4 Results

The VI inversion of gravity and magnetic data for the WSB and TAM region is carried out in two subsequent inversion runs with varying coupling factors of 25 000 for the first 100 iterations (inversion result shown in the supporting information file) and 15 000 for subsequent 25 iterations, while all other parameters are kept fixed as described in section 2.1.1 and table 1. The Root Mean Square error (RMSE) after the combined 125 iterations is 1.5 mGal for the gravity inversion and 0.7 nT for the magnetic inversions (Figure 5 a). The RMSE error decreases for the gravity and magnetic inversion sharply after the transition from a coupling of 25 000 to 15 000 but the amplitude of decreasing RMS seems not significant at first glance. However, the long wavelength residual in the magnetic inversion model decreases significantly in the residual maps after lowering the coupling (Figure 6f & Figure S4)

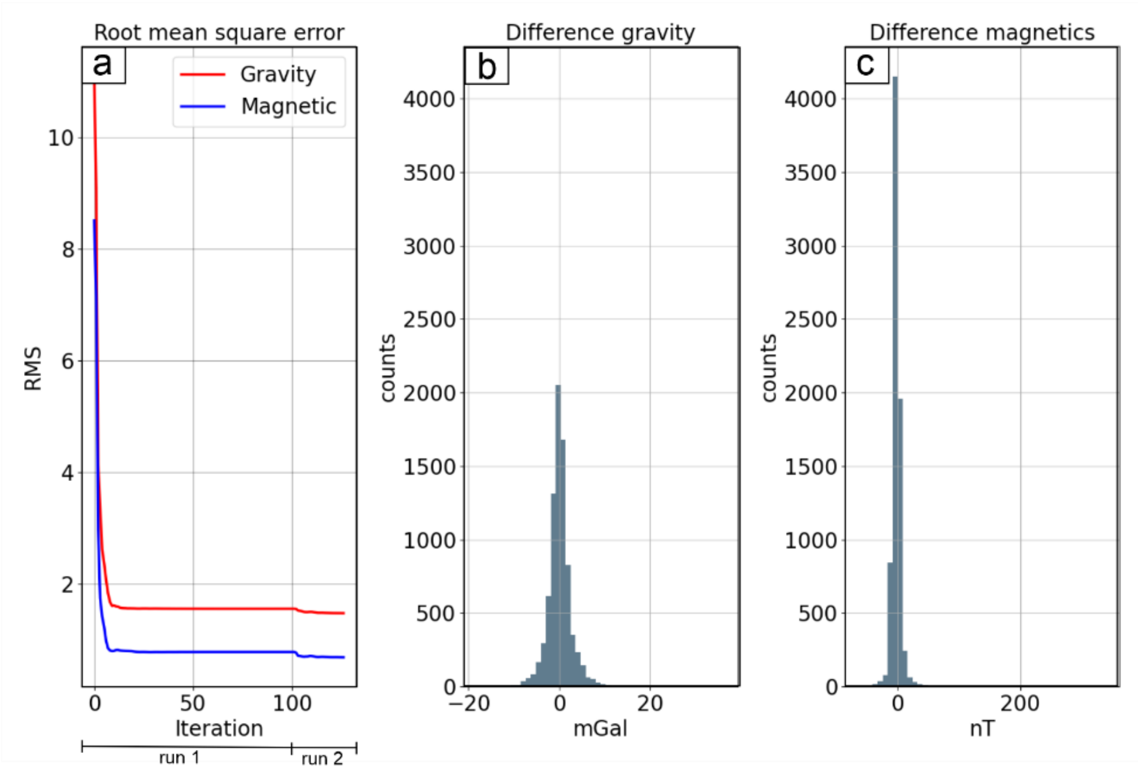

Figure 5 a) Root mean square (RMS) error between observed and inverted gravity (red) and magnetic (blue) field for each inversion iteration. b) Gravity residual histogram between observed and inverted gravity field. c) Magnetic residual between observed and inverted magnetic field.

The amplitude and distribution of anomalies within the gravity and magnetic field are reproduced well by the inversion (Figure 6a-c). The end members of the residual between observed and inverted values are + 37 and -18 mGal with a standard deviation of 3 mGal for the gravity inversion and for the magnetic inversion +341 and -65 nT with a standard deviation of 13 nT (Figure 5 c, d). The difference map between observed and inverted values are shown in figure (6 e, f). Both residual maps show a good agreement between observed and inverted gravity and magnetic field with low residuals for most of the study area. The highest misfit amplitude in the gravity field is located in the area of the TAM, in the northeastern part of the study area. Airborne surveys are sparse in this area and the AntGG [*Scheinert et al.*, 2016] compilation has significant data gaps in this region. Gravity ground stations [*Reitmayr et al.*, 2003; *Zanutta et al.*, 2018] are used to fill the data gaps (Figure S1b). A possible source of the higher amplitude in the misfit cluster in the TAM region could be contributed by the low spatial coverage of the ground stations and the offset in frequency content between the airborne and ground station data and could therefore be associated with local effects.

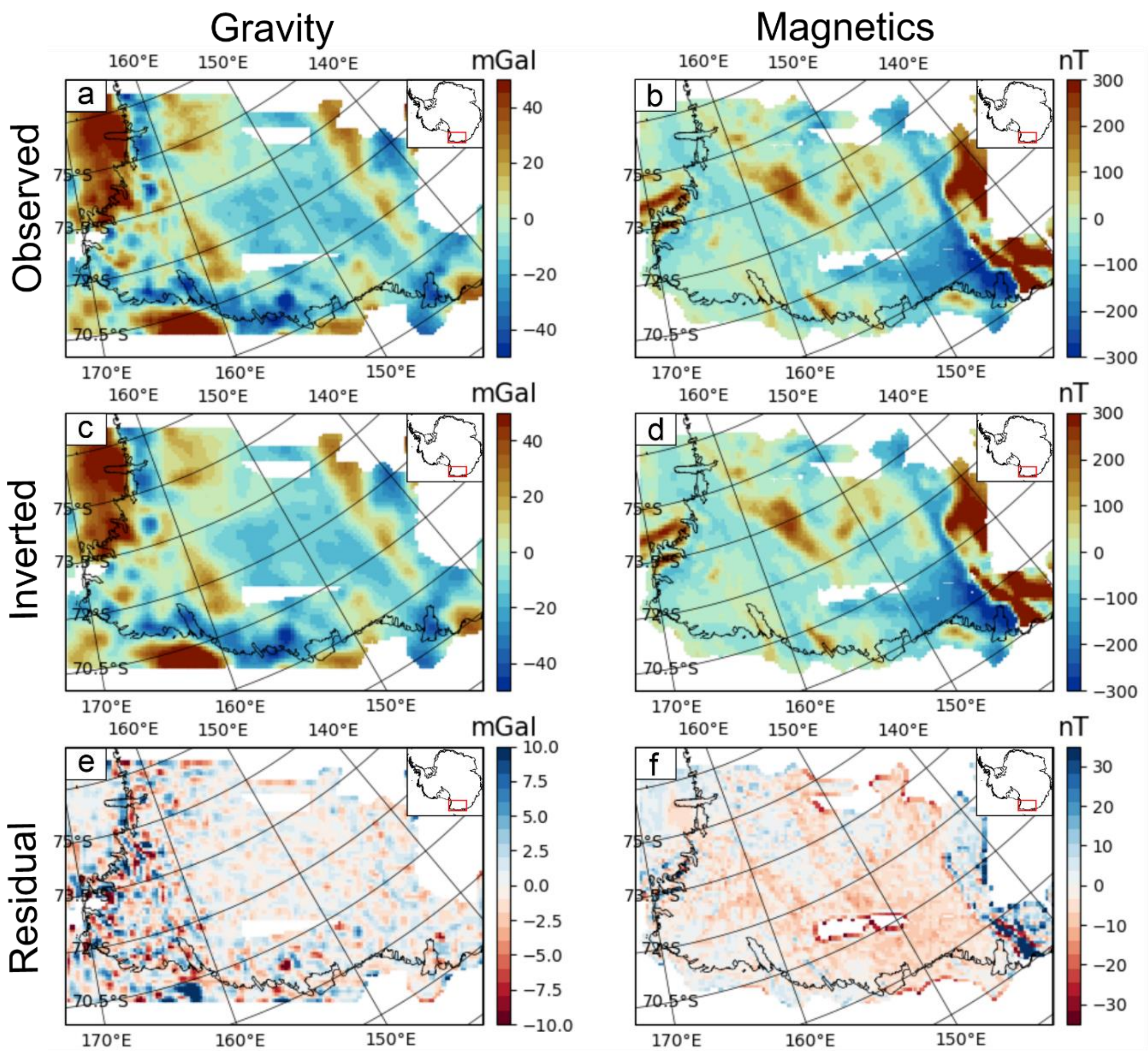

Figure 6: a) Gravity inversion input data. b) Magnetic inversion input data. c) Final inverted gravity field. d) Final inverted magnetic field. e) Difference map between observed and inverted gravity fields (6a minus 6c). f) Difference map between observed and inverted magnetic field (6b minus 6d).

The largest misfits between the observed and inverted magnetic data are located around prominent data gaps within the WSB and the large data gap in the south, as well as in the west of the study area where the amplitude of the magnetic anomaly map increases sharply from 300 nT to over 1500 nT. It is somewhat expected that the inversion algorithm struggles to reproduce a rapid variation in the magnetic field of over 1200 nT since inversion algorithms favour smooth models. On the other hand, an error of 200 nT in a region with a field strength of over 1500 nT is less dramatic than compared to the TAM and WSB region with amplitudes of 300 nT. Additionally, a long wavelength feature exists in the magnetic residual map, which we don't further address due to the low magnitude of below 5 nT.

Density and susceptibility examples at 5.5 km, 11.9 km and 20.5 km depths are presented in Figure (7). The linear southeast-northwest anomaly in the central WSB appears to be connected at depth with the neighbouring anomaly with a southeast northwest orientation, while both anomalies are separated at shallower depths.

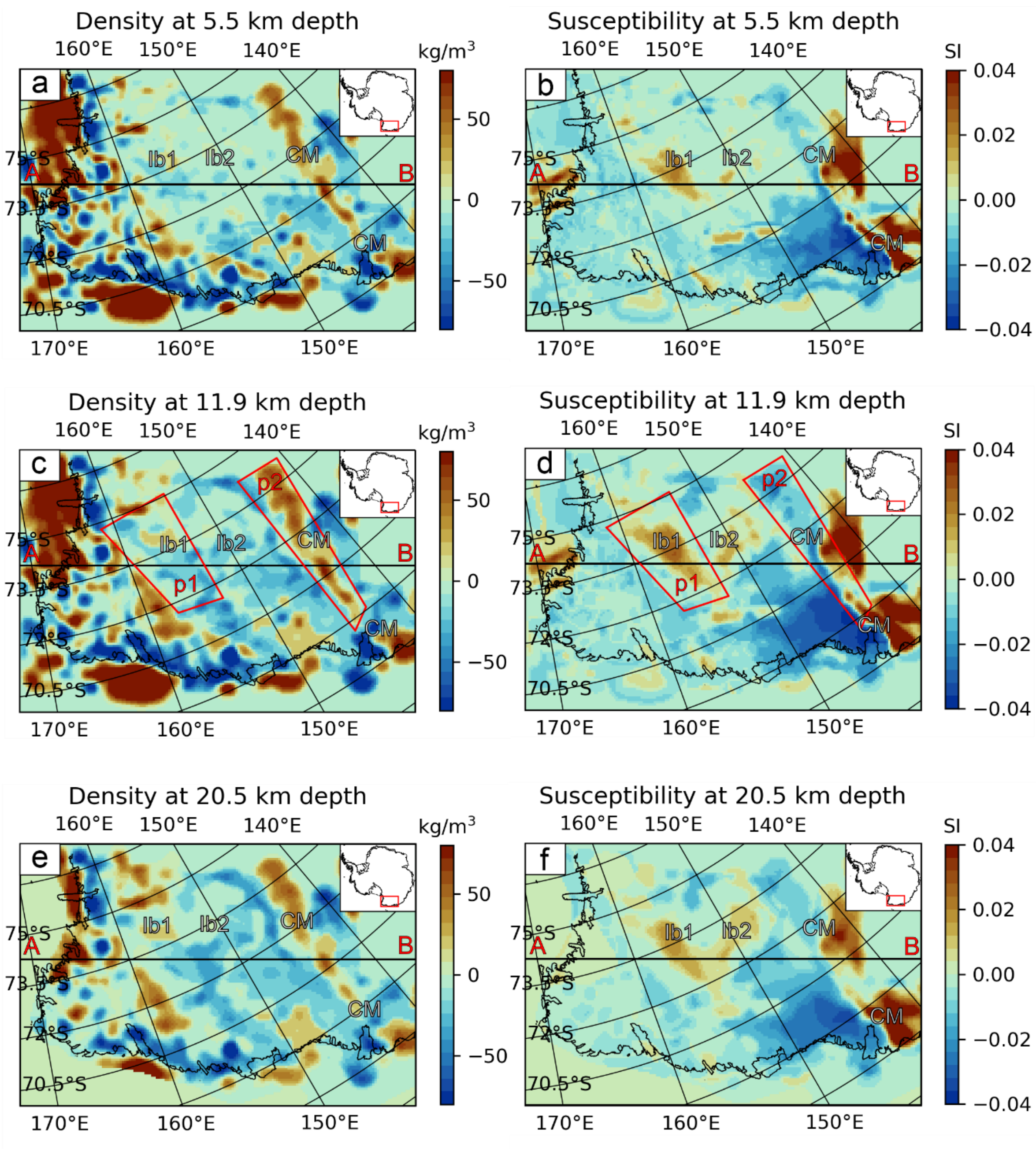

Figure 7: a) Inverted density at 5.5km depth. b) inverted susceptibility at 5.5 km depth. c) inverted density at 11.9km depth. d) inverted susceptibility at 11.9 km depth. e) inverted density at 20.5 km depth. f) inverted susceptibility at 20.5 km depth. Black line indicates location of cross section profile in Figure 8. P1 and P2 indicate the polygons for extracting 3D distribution of density and susceptibility bodies. IB: intrusive body; CM: craton margin.

The aim of the inversion is to find density and susceptibility distributions that are geometrically connected and simultaneously can explain the observed gravity and magnetic field. The underlying assumption is that a crustal rock has density and susceptibility values, which influence simultaneously the gravity and magnetic field. A cross section along profile AB (Figure 2b; Figure 3a, b; Figure 8 a-f) shows the similarities in the geometry of both petrophysical quantities. A large-scale negative density anomaly is located centrally in profile AB (black dotted circle), while the susceptibility inversion model shows a high susceptibility anomaly with identical geometry. Adjacent to this anomaly is another negative density anomaly (red dotted circle) but in this case the susceptibility values are also negative. This illustrates that a common geometry of both quantities is inverted, but the relationship between the density and susceptibility values is not linear. The cross section through the inverted density and susceptibility model indicates that the source for the linear magnetic anomaly in the central WSB is connected at depth to the neighbouring anomaly source with a perpendicular orientation (Figure 8). These results illustrate the advantage of using a joint inversion approach to obtain the density and susceptibility distribution (see supporting information Figure S6-9).

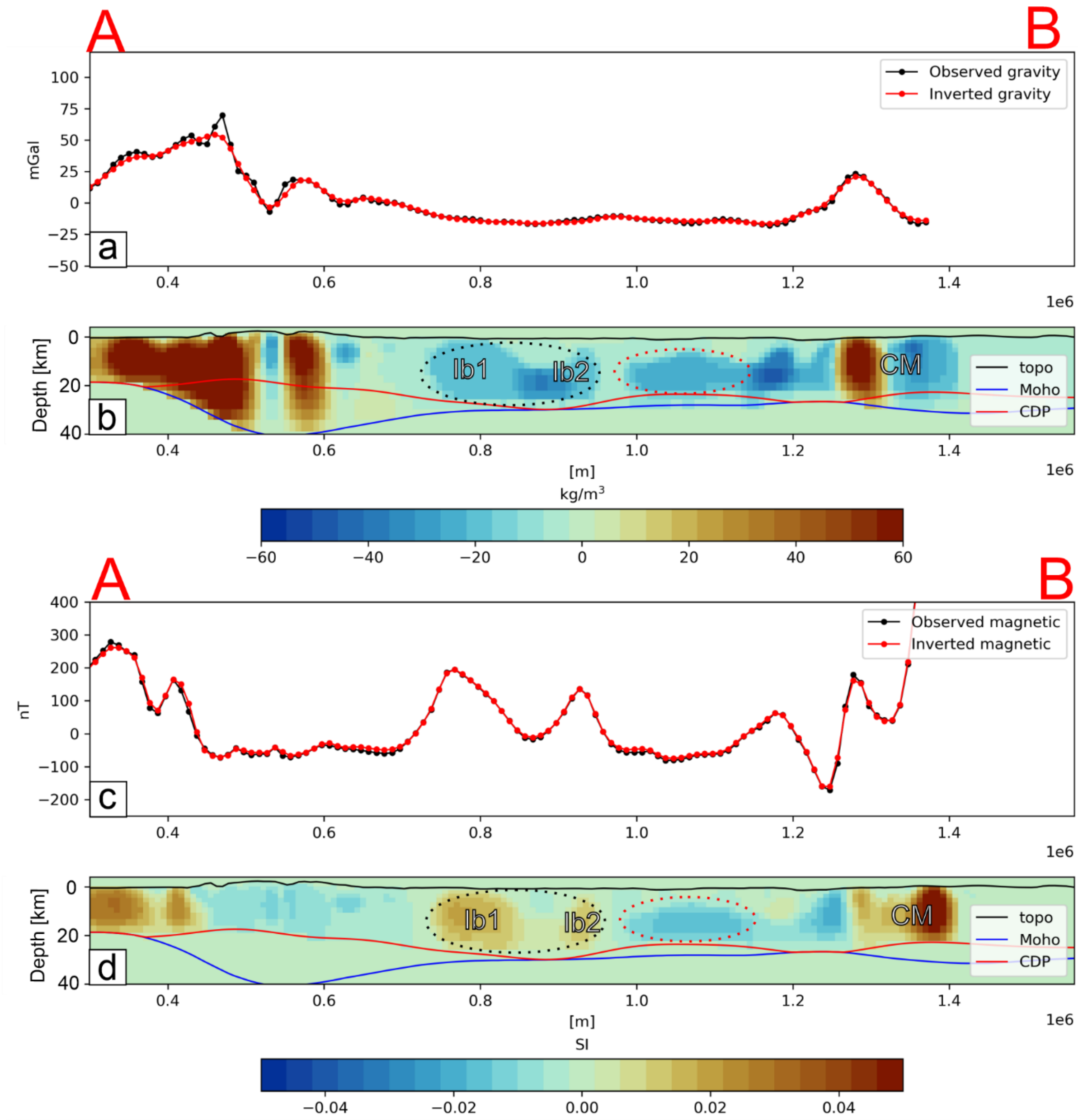

Figure 8: Observed (black line) and inverted (red line) gravity field along profile AB. Location of profile AB is given in (Figure 7). Cross-section of the inverted density model along profile AB. c) observed (black line) and inverted (red line) magnetic field along profile AB. d) cross-section of the inverted susceptibility model along profile AB. IB: intrusive body; CM: craton margin.

A density and susceptibility cross plot illustrate the parameter relationship between both quantities (Figure 9). The relative inverted densities range from -160 to 250 kg/m^3, while the inverted susceptibilities range from -0.06 to 0.9 SI. However, the density and susceptibility histograms indicate that the inverted density values are predominantly between ± 50 kg/m^3 and between ± 0.02 SI for the inverted susceptibility model.

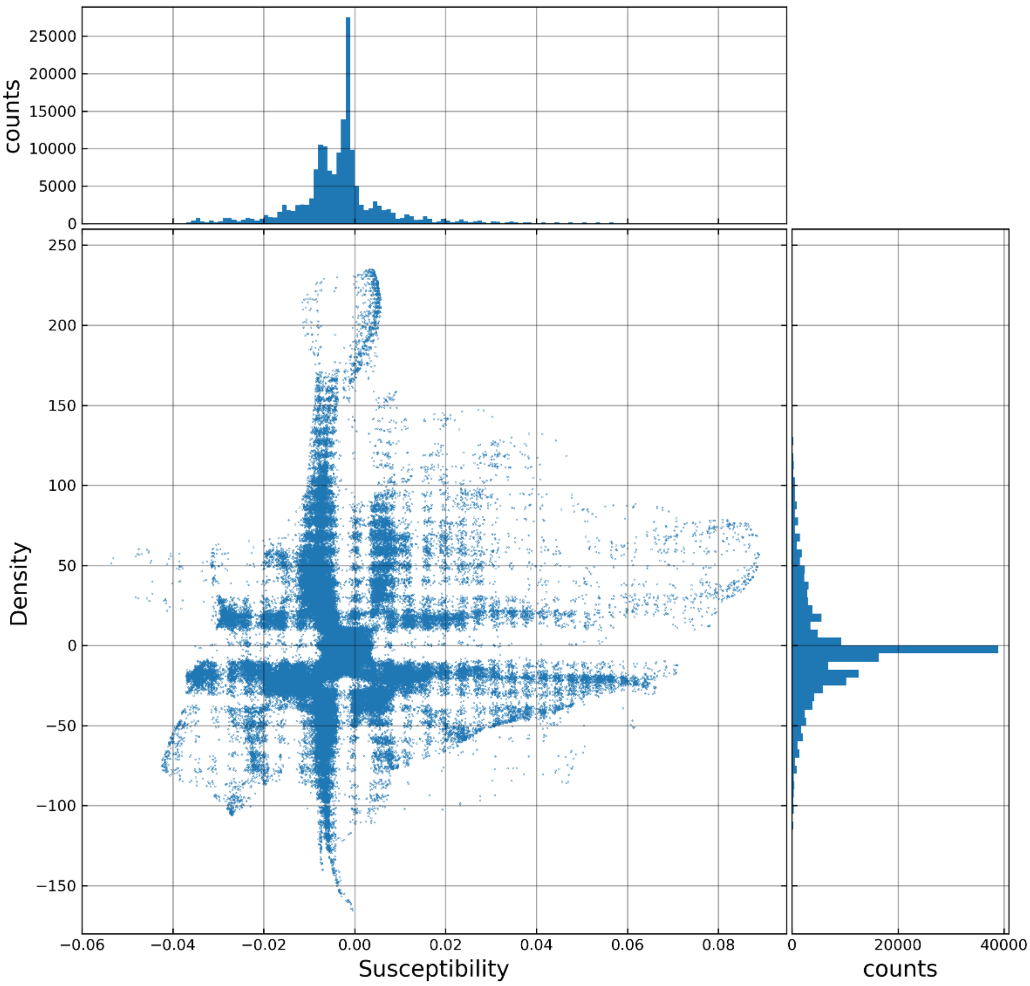

Figure 9: Inverted density and susceptibility cross plot and density and susceptibility histograms of the VI inversion model.

High relative density values above 100 kg/m^3 are located exclusively offshore while large negative values below -100 kg/m^3 are located as distinct cluster in the TAM and along the coast (Figure 10 a-b). High relative susceptibility values above 0.025 SI are located offshore and to a much larger extent at the western edge of the WSB at the inferred craton margin (Figure 10 a-b). Furthermore, strong negative relative susceptibility values below -0.03 SI are limited to the Craton margin (Figure 10 a-b).

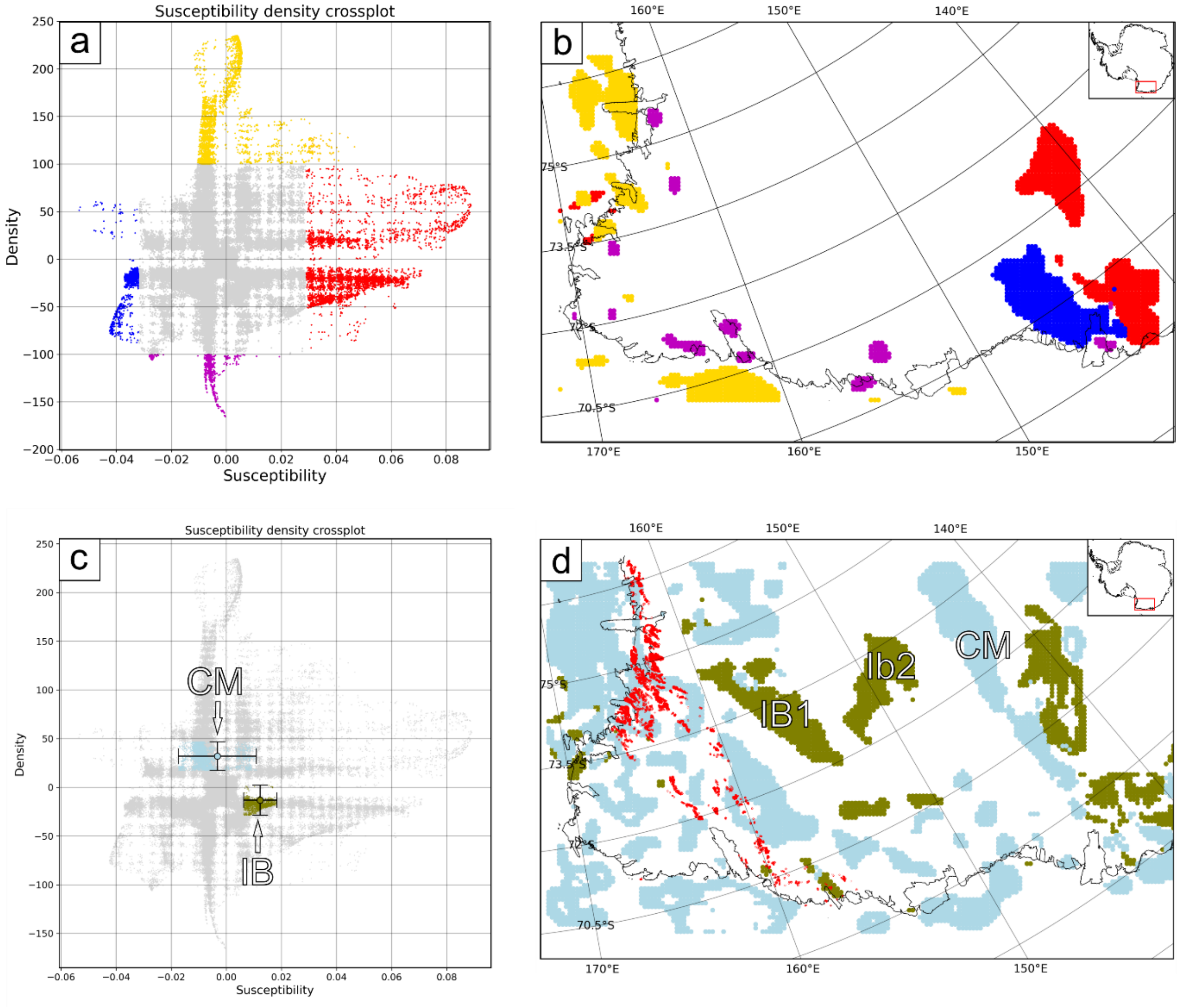

Figure 10: Density vs susceptibility cross-plots and province characterisation. a) Extreme susceptibility and density groupings. High inverted relative density values above 100 kg/m^3 (yellow dots), low relative inverted density values below – 100 kg/m^3 (purple dots), high relative inverted susceptibility values above 0.029 SI (red dots) and low relative inverted susceptibility values below -0.032 SI (blue dots) superimposed on density and susceptibility cross-plot of the entire inversion model. b) spatial distribution of extreme density and susceptibilities highlighted in a). c) density–susceptibility relationship of the interpreted intrusive bodies (olive green dots) and craton margin feature (light blue dots) superimposed on density and susceptibility cross-plot of the entire inversion model. d) location plot of density and susceptibility relationships within the standard deviation range of the extracted bodies in Figure 10 c. Granite Harbour Intrusive Complex (red) taken from GeoMAP [*Cox et al.*, 2023]. IB: intrusive body; CM: craton margin.

To constrain the geometry and properties of particular sub-surface source bodies we extracted the density and susceptibility values within the polygons p1 and p2 (polygon location shown in Figure 7 c-d) and subsequently used thresholding of the susceptibility or density values to recover the approximate geometry of sources associated with specific anomalies. The source for the prominent positive magnetic anomaly in the WSB region is extracted by windowing the joint inversion output in this area for relative susceptibility values larger than 0.003 SI. The volume of the extracted feature amounts to ~286 000 km^3. The source for the linear high gravity anomaly on the western flank of the

WSB was extracted by windowing the joint inversion output in this area for relative density values larger than 10 kg/m^3. The mean and standard deviation of the density and susceptibility relationship within those two extracted prominent features are superimposed on the density–susceptibility cross plot of the whole inversion model (Figure 10 c). Both features cluster around distinct areas in the cross plot and are easily distinguishable (Figure 10 c). The Central WSB source (IB1) has a mean relative density that is slightly negative and a mean relative susceptibility that is moderately positive, while the mean relative density is positive, and the mean relative susceptibility is negative for the feature along the craton margin (CM). The standard deviation for the susceptibility of the craton margin feature is significantly larger compared to the intrusion signal (Figure 10c). The standard deviation range of density and susceptibility for each body is subsequently used to filter the entire inversion model to find the location of rocks with matching density and susceptibility relationships across the study area (Figure 10 d). The location of rocks with matching susceptibility and density relationships reveals petrophysical similarities associated with the magnetic anomaly (F2 in Figure 3) adjacent to the central WSB magnetic anomaly (F1 in Figure 3), which potentially has the same origin (IB1, IB2 in Figure 10 d). The extent and volume of those 3D structures can be extracted from the model (Figure 11).

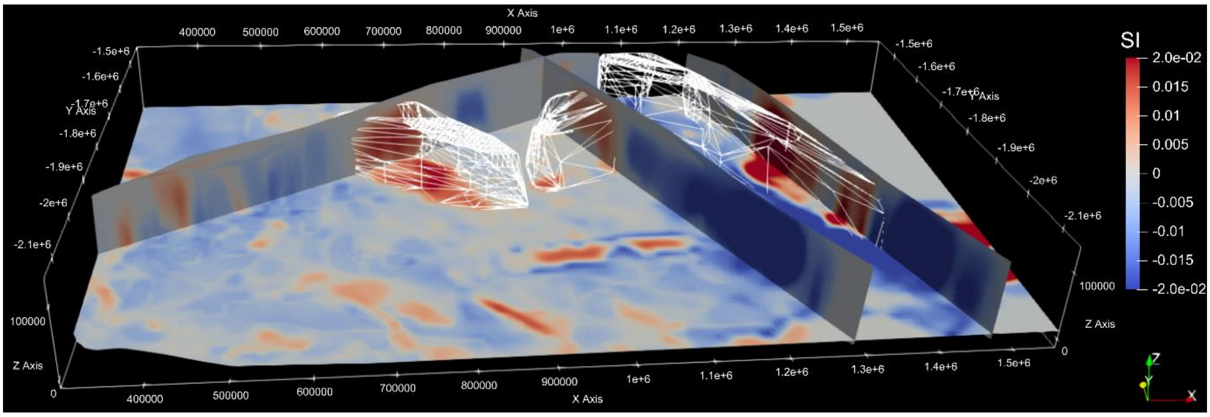

Figure 11: 3D intrusion and craton margin bodies (white wireframe) superimposed on slices through the 3D inverted susceptibility model. View looking grid north (approximately SSW geographically) along the axis of the WSB, with two interpreted intrusions in the centre and craton margin source body to the right.

5 Discussion

Shifting the relative density values to absolute values by adding the average crustal background density of 2670 kg/m^3 indicates that the density of the extracted feature in the centre of the WSB ranges from 2620 to 2690 km/m^3 combined with moderate susceptibility values. Based on its petrophysical properties and the large volume, we interpret this feature to be a granitic batholith intrusion. A previous geophysical interpretation that the origin of this feature is associated with thinned crust as a result of rifting [*Ferraccioli et al.*, 2001] can be ruled out since such a tectonic feature would have substantially higher densities than observed in the inversion model. This potentially granitic batholitic intrusion has the identical petrophysical signature in terms of density and susceptibility as the neighbouring body labelled IB2 in Figure (10d) which is associated with the magnetic anomaly labelled F2 in Figure (3 a, b). The density and susceptibility distribution along the AB profile (Figure 8 b-d) and the depth slices (Figure 7f) indicates that the sources for these anomalies (IB1 and IB2) are connected at depth. The total volume of the proposed batholith amounts to ~470 000 km^3, which is a significant addition to the volume of the upper crust at the time of emplacement. This block of intruded material very likely deviates from the surrounding rocks in terms of radiogenic heat production and thermal conductivity. Such broad crustal inhomogeneities or crustal blocks should be considered in future geothermal heat flow models of East Antarctica, even though the implementation is very challenging because the precise geochemical composition and the thermal petrophysical characteristics are unknown due to a lack of rock outcrops. The absence of the proposed granitic signature in the TAM is highly interesting, since granites from the Granite Harbour Igneous Complex are mapped and sampled by geological surveys in this area [*Cox et al.*, 2023]. This suggests that the petrological signature of the proposed batholithic intrusion body is fundamentally different from granites of the Granite Harbour Igneous Complex (Figure 10 d red dots) which was emplaced during the Ross Orogeny [*Estrada et al.*, 2016]. We speculate that the intrusion in the central WSB happened during the Pan African time (551 – 700 Ma) and is the source material of the Pan African-age zircons in the Priestley Formation [*Estrada et al.*, 2016]. Furthermore, a potential batholithic intrusion on such a scale will most certainly affect local crustal inhomogeneities and the local heat flow budget even though the lack of direct geological samples and heat flow measurements makes it challenging to quantify the extent. Even though the potential batholitic intrusion in the WSB is based on geophysical interpretation and direct petrophysical and thermal property measurements are lacking comparing the inferred batholith to direct measurements of potentially similar intrusive rocks in the Weddell Sea [*Leat et al.*, 2018], the southern Przdz Bay [*Carson et al.*, 2014] and Cornwall [*Beamish and Busby*, 2016] can inform which magnitude of heat flow elevation could be expected for such a batholith, especially for East Antarctica were surface heat flow values are rarely predicted to exceed ~60 mW/m^2 [*An et al.*, 2015; *Fox Maule et al.*, 2005; *Haeger et al.*, 2022; *Lösing and Ebbing*, 2021; *Lowe et al.*, 2023; *Martos et al.*, 2017; *Shen et al.*, 2020; *Stål et al.*, 2021]. A granitic batholithic intrusion will increase the heat flow in this region significantly, especially when the granites are rich in heat producing elements such as potassium (K), thorium (Th), and uranium (U). Evidence of high heat producing granites exist in the Weddell Sea, which are responsible for significantly increasing the local heat flow [*Leat et al.*, 2018]. Those granites, which intruded the West Antarctic crust are rich in Th and U up to 60.7 and 28.6 ppm, respectively and lead to heat production of up to 9.06 µW/m^3 [*Leat et al.*, 2018]. The local heat flow is predicted to reach 70-95 mW/m^2, which is 15-30 mW/m^2 higher than the surrounding West Antarctic crust [*Leat et al.*, 2018] and is even higher compared to the colder East Antarctic crust. Cambrian granites of southern Prydz Bay, East Antarctica, are reported to increase local surface heat flow above 120 mW/m^2 [*Carson et al.*, 2014]. The Cornubian granite batholith province in Cornwall, England, is one of the highest heat flow regions in the United Kingdom [*Beamish and Busby*, 2016]. The Cornubian granite batholith stretches roughly 200 km through southwest England and is comparable in size to the batholith in the central WSB. The heat flow within the Cornubian granites is reported to be up to 138 mW/m^2 [*Beamish and Busby*, 2016]. This comparison illustrates the magnitude of heat flow which is potentially underestimated in geophysical geothermal heat flow models if the crustal domain is treated with global average values instead of accounting for crustal heterogeneities.

The density and susceptibility relationship for the linear structure at the craton margin is more ambiguous than the proposed batholith structures, due to the larger standard deviation in the susceptibility values. Similar density and susceptibility relationships are present offshore, and onshore along the cost and in the TAM, but these areas may be lithologically distinct from the craton margin structure. We propose that the linear craton margin feature can be interpreted as the remnant signature of a former magmatic rifted continental margin where thick sequences of seaward dipping basalt horizons are often combined with mafic underplating. Both features require densities above 2700 kg/m^3, in line with the inversion result. Additionally, both features host the potential of significant remanent magnetisation, as indicated by negative relative susceptibility. An alternative model for the positive gravity anomaly associated with the margin of the East Antarctic craton inboard from the WSB proposed by [*Studinger et al.*, 2004] is that of a section of up-thrust crustal material loading the craton margin. Our inversion does not show the strongly asymmetric pattern of densities expected for a flexural loaded margin, but we can-not rule out some amount of compressional re-working of the former rifted margin during the Ross Orogeny.

Our inversion results and the density – susceptibility relationship indicate that a large-scale batholitic intrusion is present in the WSB. We speculate that this was emplaced during the Pan African time, and is not connected to the Granite Harbour Igneous Complex, which was emplaced slightly later during the Ross Orogeny [*Estrada et al.*, 2016]. Furthermore, the highly variable susceptibility values in the west towards the craton margin combined with the high-density bodies could indicate thick seaward dipping basalt horizons coupled with mafic magmatic underplating, which can both have high susceptibility and potentially hold strong remanent magnetisation, giving rise to the recovered negative relative susceptibility values. Based on those findings we propose a four-step tectonic evolution (Figure 12), which includes the initial development of a rifted continental margin with seaward dipping basalt horizons and mafic underplating, followed by passive margin development and two distinct subsequent intrusion events to emplace the proposed “Central Batholith” and the Granite Harbour Igneous Complex. We speculate that these were resulted from subduction-related magmatic arc formation. The age of the initial rifting event is unknown, but evidence of rifting and passive margin development in the Central Transantarctic Mountains suggests that this may have occurred around 670 Ma [*Goodge*, 2020; *Goodge et al.*, 2002]. We propose that the Central Batholith was emplaced after development of a subduction system against the former passive margin. This batholith may be the source for the ~650 Ma zircon ages found in sedimentary rocks in the Priestly formation [*Estrada et al.*, 2016]. Emplacement of the Central Batholith was followed by the Ross age (550 - 450 Ma) emplacement of the Granite Harbour Igneous Complex.

In summary, we propose a four-stage geodynamic evolution model for this region. Stage 1: ~670 Ma: Continental breakup and a developing magmatic rifted margin with seaward dipping basalt horizons and mafic underplating. Stage 2: ~660 Ma continued ocean spreading and sediment deposition on the shelf of the passive margin. Stage 3: ~650 Ma Arc development as a result of subduction against continental margin. Large scale emplacement of Batholith intrusion, potential re-working and back thrusting of mafic components onto the craton. Stage 4: ~500 Subduction zone retreat associated with secondary arc development and emplacement of intrusive rocks of the Granite Harbour Igneous Complex in bound to the TAM.

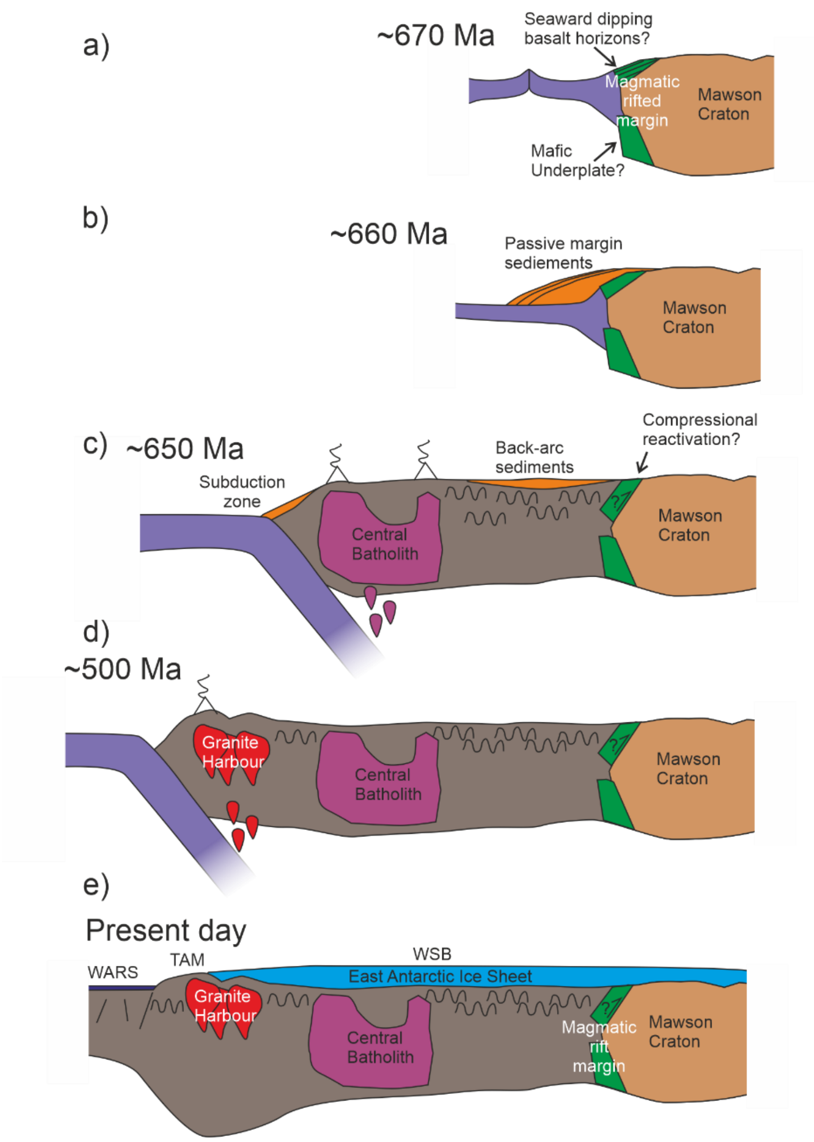

Figure 12: Tectonic evolution sketch: a) Rifting of Rodinia supercontinent and development of magmatic margin. b) Passive margin development and sediment deposition. c) Subduction zone development, emplacement of Central Batholith, re-working of passive margin and potentially deposition of back-arc sediments. d) Subduction zone migrates further out-board and Granite Harbour intrusive suite is emplaced. e) Present day geological section.

6 Conclusion and Future Work

We present a density and susceptibility distribution model for the Wilkes Subglacial Basin and the Transantarctic Mountains using joint inversion of gravity and magnetic data based on variation of information coupling. This model provides insight into the heterogeneity of the 3D crustal structure in The WSB and TAM region and allows quantification of the volume of crustal provinces which should be considered in future lithospheric scale thermal studies. The inversion model images a large low density and moderate susceptibility body in the central WSB, which we interpreted as most likely to be an intrusive granite batholith based on the inverted petrophysical properties. A failed rift scenario for the origin of this body can be ruled out based on the inverted petrophysical relationship. Furthermore, the density and susceptibility relationship and cross-section of the inversion model indicates that this structure is connected to the adjacent low density moderate susceptibility body, which potentially has the same origin but nearly perpendicular orientation. If so the volume of the total granitic intrusive body increases to ~470 000 km^3, which is a considerable amount of addition volume of the upper crust at the time of emplacement phase that has the potential to have an enduring effect on the local heat flow due to radiogenic decay of heat producing elements. Despite the lack of direct heat production and heat flow measurements, correlation to well-studied granite intrusion provinces in the Weddell Sea [*Leat et al.*, 2018], southern Prydz Bay [*Carson et al.*, 2014], and Cornwall [*Beamish and Busby*, 2016] emphasise the influence local crustal structures such as intrusive bodies can have on the local heat flow budget.

Examining the density and susceptibility relationship we found that the batholith intrusion has a different petrophysical signature to the Granite Harbour Igneous Complex. Furthermore, the edge of the craton is identified by the inversion in the form of a positive linear gravity anomaly west of the WSB. Based on our findings the tectonic evolution of the WSB requires two distinct intrusion events emplacing the batholith and subsequently emplacing the Granite Harbour Igneous Complex. The inverted crustal model supports the idea of a passive continental margin with thick seaward dipping basalt horizons and mafic underplating. However, the alternative idea of up-thrusted crustal material at the craton edge cannot be ruled out based on the geophysical inversion model. Combined scenarios between these endmember models are also possible.

This study highlights the crustal heterogeneities on a regional scale in East Antarctica and gives evidence that treating the crustal domain with a set of constant global average values in geophysical geothermal heat flow models is a simplification that might underestimate the geothermal heat flow beneath the ice sheets. Despite the many challenges, next generations of geophysical heat flow models are required to consider crustal heterogeneities to increase the understanding of the contribution from the solid earth to the cryosphere and ultimately the stability of the ice sheet. Therefore, further geophysical, and geological research on the Antarctic subglacial geology is necessary to understand the thermal state of the most remote continent on Earth.

Acknowledgements

Funding for this research was provided by NERC through a SENSE CDT studentship (NE/T00939X/1). ML acknowledges additional funding through the Gray-Milne Travel Bursary provided by the British Geophysical Association and the Small IT grant awarded by the School of GeoSciences, The University of Edinburgh. The authors thank the developers of open scientific software products and colormap products which were utilized in this study: JIF3D is available under a GNU General Public License (v3) via subversion at https://svn.code.sf.net/p/jif3d/jif3dsvn/trunk/jif3D; NumPy [*Harris et al.*, 2020], Matplotlib [*Hunter*, 2007]; Pandas [*McKinney*, 2010]; Geopandas [*Jordahl et al.*, 2020]; Cmrameri [*Crameri*, 2023]; Cartopy [*Elson et al.*, 2022]; Shapley [*Gillies et al.*, 2022]; SciPy [*Virtanen et al.*, 2020]; PyKrige [*Murphy et al.*, 2022] and Jupyter notebook [*Kluyver et al.*, 2016] as well as the developer of the commercial software Geosoft, especially the plug-in “Compudrape” and their publisher Seequent.

Open Research

Python code in form of a Jupyter notebooks to reproduce the inversion and plot the inversion results are temporarily available at GitHub (https://github.com/MaximilianLowe/VI_inversion_WSB_TAM_JGRse). The GitHub repository will be transferred to zenodo after the manuscript is accepted.

Inversion models will be available and archived at the British Antarctic Survey Polar data centre as netCDF files. The data is transferred but not jet published (usually takes a few weeks). In the meantime, for the review process the netCDF files can be access through following link: https://www.dropbox.com/scl/fo/uqlsm50cf6ykkcje1ke41/h?rlkey=8y90rjwlwefmnlj4l1qdwxoxu&dl=0

BedMachine Antarctica version 3 [*Morlighem et al.*, 2020] is freely available at https://nsidc.org/data/nsidc-0756/versions/3

ADMAP-2 magnetic data [*Golynsky et al.*, 2018] is freely available at https://doi.pangaea.de/10.1594/PANGAEA.892724

AntGG [*Scheinert et al.*, 2016] is available at: https://doi.pangaea.de/10.1594/PANGAEA.892724

Curie point depths [*Lowe et al.*, 2022] are available at: https://ramadda.data.bas.ac.uk/repository/entry/show?entryid=b8dcbaa9-3ac0-42bd-95a5-6b5961cbcb7e

Moho depth [*Pappa et al.*, 2019a] are available from: https://agupubs.onlinelibrary.wiley.com/action/downloadSupplement?doi=10.1029%2F2018GC008111&file=GGGE_21848_DataSetsS1-S6.zip.

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