



Universidade de Brasília

Instituto de Geociências

Programa de Pós-graduação em Geologia

Estrutura Termal e Elétrica do Cráton Amazônico

The Thermal and Electrical Structure of the
Amazon Craton

RAPHAEL TEIXEIRA CORREA

Tese de Doutorado Nº 198

Orientador: Prof. Dra. Roberta Mary Vidotti

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“The purpose of models is not to fit the data,
but to sharpen the questions”.

“O propósito de modelos não é ajustar os dados,
mas aguçar as perguntas”.

Samuel Karlin (1924 – 2007)

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RESUMO

O Cráton Amazônico evoluiu durante o Paleo-Mesoproterozoico através de sucessivos arcos magmáticos continentais, as quais idades diminuem na direção oeste. No entanto, quando, onde e se esses arcos magmáticos colidiram com outro continente, incluindo o fechamento de bacia oceânica é um debate em andamento. Províncias Geocronológicas foram propostas para explicar a evolução geológica do cráton. Entretanto, a distribuição geográfica irregular dos dados geocronológicos associada há um certo grau de desconexão com estruturas magnetométricas e gravimétricas profundas faz surgir uma série de questionamentos em relação a continuação das províncias e o significado estrutural de suas bordas. Tendo como base a premissa que a reologia dos sólidos é controlada principalmente pela temperatura, mapeou-se a profundidade da base da litosfera magnética, assumida como uma estimativa à profundidade da temperatura Curie, de modo a endereçar a estrutura termal. Além disso, modelos de resistividade elétrica tem sido efetivos em mapear descontinuidades de escala crustal. Desta forma, adquiriu-se dados magnetotélúricos em regiões chaves, de modo que fosse possível comparar as crostas formadas durante a evolução dos arcos magmáticos. Nosso modelo termal sugere que os limites dos domínios Carajás e Bacajás devem ser revisitados, assim como o limite entre Tapajós e Iriri-Xingú. Permitiu-se concluir que o manto na porção leste do domínio Carajás e no domínio Tapajós preservaram minerais serpentinizados. O modelo de resistividade elétrica confirmou que o manto do domínio Tapajós foi hidratado e refertilizado devido ao longo processo de subducção. Essa fertilidade mantélica possui correlação com um aumento de depósitos magmático-hidrotermais. Uma estrutura de escala crustal com mais de 1000 km de extensão foi identificada. Argumenta-se que esta estrutura separa dois fragmentos tectônicos, uma vez que os mesmos possuem diferenças relevantes na direção das estruturas e no gradiente magnético. Ao longo desta feição, encontra-se i) par de longo comprimento de onda de baixo e alto Bouguer, ii) um degrau na descontinuidade de Mohorovic, iii) e um condutor na crosta inferior. Essas evidências corroboram a hipótese de uma colisão Paleoproterozoica.

Palavras-chave: Superfície Curie; Magnetotélúrico; Cráton Amazônico; Sutura Paleoproterozoica.

ABSTRACT

The Amazon Craton evolved in the Pale-Mesoproterozoic through successive continental magmatic arcs, which ages decrease westwards. However, when, where, and if these magmatic arcs collided with another continent closing an ocean-basin is a subject of ongoing debate. Geochronological Provinces have been proposed to explain the geological evolution of the Craton. However, the irregular geographic distribution of geochronological data associated with a certain degree of decorrelation with magnetic and gravity structures raised a series of questions about the continuation of these provinces and the meaning of their boundaries. Assuming that the reology of solids is primarily a function of temperature, we mapped the depth to the bottom of the magnetic's lithosphere, which is a proxy for the depth to the Curie temperature, thereby addressing the thermal structure. Furthermore, electrical resistivity models have been effective in mapping crustal scale discontinuities. Therefore, we acquired magnetotelluric data in key regions in order to compare the crusts formed during the evolution of the magmatic arcs. The thermal model suggests that the limits between Carajás and Bacajás domains should be revisited, as well as the limits between Tapajós and Irixi Xingu domains. It also suggests that the mantle of the eastern portion of the Carajás domain and Tapajós have preserved serpentized minerals. The electrical resistivity model confirmed that the mantle of the Tapajós domain was hydrated and refertilized due to long lived subduction processes. This mantelic fertility has a high correlation with the position of magmatic-hydrothermal mineral deposits. A major crustal-scale structure (more than 1000 km in extension) separates two tectonic fragments with remarkable differences in the magnetic gradient and the structures's trend. Additionally, we also find along this major boundary: i) high and low bouguer anomalies of long-wavelength paired; ii) step in Mohorovic discontinuity; and iii) a lower crust conductor. These pieces of evidence support the hypothesis of a collision in Paleoproterozoic times.

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Keywords: Curie Surface; Magnetotellurics; Amazon Craton; Paleoproterozoic Suture.

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CAPÍTULO 1

INTRODUÇÃO

1 INTRODUÇÃO

O Cráton Amazônico evoluiu durante o Paleo-Mesoproterozoico através de sucessivos arcos magmáticos continentais, as quais idades diminuem na direção oeste (Motta et al., 2022). No entanto, quando, onde e se esses arcos magmáticos colidiram com outro continente, incluindo o fechamento de bacia oceânica é um debate em andamento. Modelos de subducção de longa vida (Cassini et al., 2020, Scandolara et al., 2017), assim como modelos intracontinentais tem sido fundamentados para explicar a evolução crustal do Cráton Amazônico (Rizzotto et al., 2019a, Rizzotto et al., 2019b). Todavia, vínculos geofísicos para a crosta e o manto ainda precisam ser inseridos para aprofundar a discussão dos mesmos.

Províncias Geocronológicas foram propostas para explicar a evolução geológica do cráton (Tassinari e Macambira, 1999, Santos et al., 2000, Vasquez et al., 2008). Entretanto, a distribuição geográfica irregular dos dados geocronológicos associada há um certo grau de desconexão com estruturas magnéticas e gravimétricas profundas faz surgir uma série de questionamentos em relação a continuação das províncias e o significado estrutural de suas bordas (Juliani et al., 2013, Carneiro et al., 2019).

Ciclos de supercontinentes promovem mudanças termais e de densidade no interior da Terra e ocorrem na escala de centenas de milhões de anos (Condie et al., 2021). A área de estudo da tese (Figura 1) abrange rochas do Arqueano ao Mesoproterozóico, de modo que se especula que o Cráton tenha feito parte de dois supercontinentes: Columbia e Rodínia (Scandolara et al., 2017, Tohver et al., 2004, Motta et al., 2022).

Métodos geofísicos aplicados na escala crustal a litosférica tem contribuído para o entendimento da espessura crustal, composição crustal e mantélica, e cronologia da estruturação profunda do cráton (Albuquerque et al., 2017, Fernandes et al., 2018, Carneiro et al., 2017, Costa et al., 2020, Motta et al., 2022). No entanto, analisar a evolução crustal do cráton a partir de vínculos termais e elétricos é um campo inexplorado. Desta forma, propoe-se uma abordagem multi-disciplinar que envolve o cálculo da estrutura termal a partir de dados aeromagnetométricos. Adicionalmente, dados de magnetotelúrico e gravimetria terrestre são utilizados para gerar modelos de resistividade elétrica e densidade para a crosta em regiões chaves para o entendimento da evolução dos terrenos. Esses resultados são integrados e interpretados de forma conjunta com dados de geocronologia (idades magmáticas e metamórficas), sísmica passiva (modelos de espessura crustal e razão V_p/V_s), e recursos minerais.

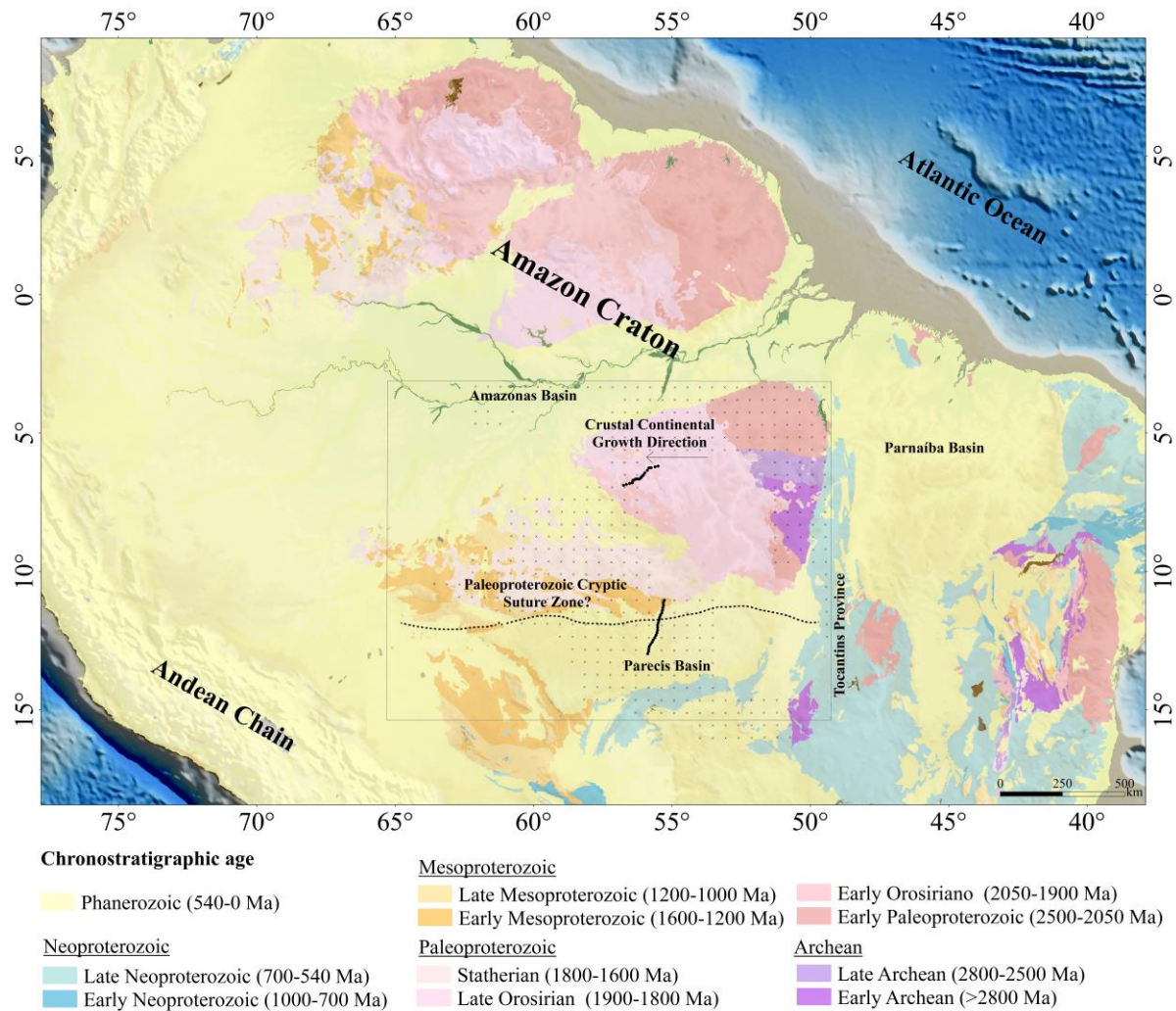


Figura 1.1 Mapa cronoestratigráfico do norte da plataforma sulamericana. O retângulo preto se refere a área de estudo. As cruzes se referem as estimativas da profundidade Curie, e as bolas se referem às estações magnetotélúricas e gravimétricas. Adaptado de Cordani et al. (2016).

1.1 Objetivos

O objetivo principal da tese é gerar modelos geofísicos que subsidiem uma análise *bottom-up*, de modo que seja possível fazer inferências sobre a composição da litosfera do Cráton Amazônico. O estudo se desenvolveu com base em algumas perguntas e hipóteses:

- Quais são as estruturas de escala crustal? Quais os seus significados?

-
- ii) O entendimento dessas estruturas permite inferir como se deu a construção do cráton no proterozoico?
 - iii) Os modelos termais e elétricos favorecem um ambiente de borda de placas ou intracontinental?
 - iv) Existe sutura no Paleoproterozoico?
 - v) Houve reativações ao longo das principais estruturas?
 - vi) É possível associar a composição mantélica às mineralizações?

1.2 Estrutura da Tese

Os resultados da presente tese foram divididos em três artigos. O primeiro intitulado “*Mapping the Thermal Structure of the Amazon Craton to Constrain the Tectonic Domains*” foi publicado no *Journal of Geophysical Research – Solid Earth* (Correa et al., 2022). Este artigo endereça a estrutura termal da área de estudo, e sua relação com dados de métodos potenciais, tomografia sísmica e espessura crustal para discutir a extensão das províncias geocronológicas.

O segundo artigo intitulado “*Tectonic and metallogenic implications of a metasomatized mantle inferred from magnetotellurics and gravity data: An example of Tapajós Mineral Province, Amazon Craton*” foi submetido à *Economic Geology*. Este artigo apresenta dados inéditos do método Magnetotelúrico para a litosfera Orosiriana, de modo que a composição do manto é discutida e o contexto geológico de estruturas de exumação profundas. O terceiro artigo intitulado “*A multi-disciplinary geophysical approach reveals a Paleoproterozoic Cryptic Suture in the Amazon Craton – Implications for Proterozoic Supercontinent Reconstructions*”. Este artigo traz uma abordagem multi-disciplinar que inclui dados de métodos potenciais, magnetotelúrico, sísmica passiva e geocronologia para investigar o significado de uma estrutura de mais de 1000 km de extensão desde a sua formação no Paleoproterozoico e os subsequentes períodos de reativação no Mesoproterozoico.

CAPÍTULO 2

ESTRUTURA TERMAL DO CENTRO-SUL DO CRÁTON AMAZÔNICO

JGR Solid Earth

RESEARCH ARTICLE

10.1029/2021JB023025

Key Points:

- We used the depth to the bottom of the magnetic layer to predict the thermal structure
- Our model suggests changes to the limits of Carajás and Bacajás domains as well as the Tapajós and Iriri Xingu domains
- Serpentinized mantle portions correlate with world-class mineral deposits in the Carajás domain

Supporting Information:

Supporting Information may be found in the online version of this article.

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Mapping the Thermal Structure of the Amazon Craton to Constrain the Tectonic Domains

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Abstract Rheology of solids is primarily a function of temperature. Mapping depth to the bottom of the magnetic layer (DBML), which is assumed as a proxy for the depth to Curie temperature, can be used to constrain the thermal structure of the lithosphere. Heat flow data are commonly used to address the thermal structure of the lithosphere. However, these data are scarce for the Amazon Craton. Therefore, as an alternative, DBML was calculated using fractal magnetic source distribution. The Amazon craton is usually described in terms of geochronological provinces models. Nevertheless, geophysical data are poorly discussed in these models. We provide a DBML model that was integrated with the potential field, seismic tomography, and crustal thickness data to address this issue. Our model varies from 10 to 80 km, and almost 90% of the estimates are in the 20–50 km range. The recurrence of long-wavelength features on independent datasets supports the assessment that the model is robust and recovered real geophysical signals. Our model suggests that the limits between Carajás and Bacajás domains should be revisited as well as the limits between Tapajós and Iriri Xingú domains. Moreover, we also correlate the model with main mineral deposits. Further, the identified DBML anomaly suggests the mantle is serpentinized in the eastern Carajás domain. This feature is highly correlated with high *p*-wave velocity indicating fertile mantle conditions, thus making this region favorable for world-class mineral deposits.

Plain Language Summary Ferromagnetic materials demagnetize when they reach the Curie temperature. On Earth, temperature and depth increases are directly proportional so that the Curie point is achieved at a certain depth. We used a specialized code to calculate Curie points/temperature for the Amazon Craton to understand the meaning of hot and colder areas. Cratons are geological bodies located distant from current tectonic activity while also showing evidence of the evolution of the past configurations of the continents. Understanding the relationship between the domains may elucidate the past configurations of the cratons and how they evolved. We compared our estimates with the depth of Mohorovic discontinuity, which is a boundary between crust and mantle and usually around 40 km for cratonic areas. We noticed that deposits of ferrous and metals base occur preferentially where the mantle is magnetic.

1. Introduction

Depth to the bottom of the magnetic layer (DBML) is a boundary limiting the extension of the lithospheric magnetic signal and has been interpreted as the depth to Curie temperature. At this point the magnetic materials lose their ferri- or ferromagnetic properties (Guimarães et al., 2014; Ravat et al., 2007). Magnetite is the most common magnetic mineral in the Earth's crust; hence 580°C was considered as a Curie point/temperature in our study.

The rheology of solids is primarily a function of temperature (Turcotte & Schubert, 2014). Therefore, assuming the DBML as a proxy for depth to Curie temperature, the thermal structure of the lithosphere can be constrained (Chopping & Kennett, 2015; Correa et al., 2016; Ross et al., 2006; Salem et al., 2014).

Although the Amazon Craton is one of the main tectonic units in South America (area of 5,6 million km²), it is probably one of the most unknown due to very large rainforest, indigenous areas, and substantial sedimentary basins. The geological evolution of the Amazon Craton is still controversial, especially regarding the number of documented magmatic arcs and the lack of evidence of suture zones. Meaning and position of the limits proposed by the geochronological model are poorly constrained by the geophysical data (Juliani et al., 2013; Santos et al., 2004; Scandolara et al., 2017; Tassinari & Macambira, 1999). The rheology of these domains is not discussed due to the lack of heat flow, crustal thickness, and high-resolution gravity data. One alternative is

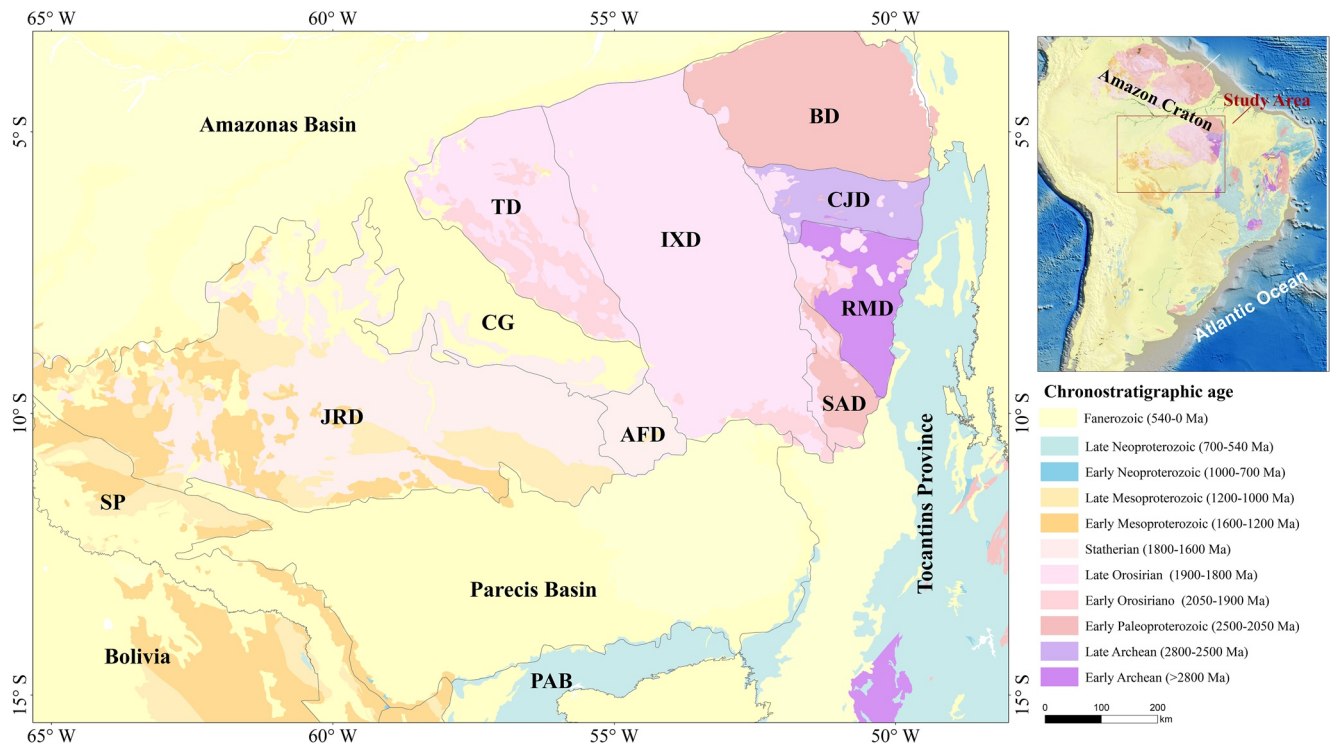


Figure 1. Chronostratigraphic map of the study area (Cordani et al., 2016). Geochronological Province limits adapted from Vasquez, Rosa-Costa et al. (2008). CJD: Carajás Domain, RMD: Rio Maria Domain, SAD: Santana do Araguaia Domain, BD: Bacajás Domain, IXD: Iriri Xingu Domain, TD: Tapajós Domain, AFD: Alta Floresta Domain, JRD: Juruena Domain, SP: Sunsás Province, CG: Cachimbo Graben, PAB: Paraguay-Araguaia Belt. Inset shows the study area in South America.

to derive DBML from the magnetic data and use it as a proxy for the depth to Curie temperature to constrain the thermal structure.

This study establishes a DBML model for the south-central portion of the Amazon Craton considering a fractal source distribution. Even though the Craton has an older stable continental lithosphere, its complex evolution since the Archean involves break up and formation of at least two supercontinents (Condie et al., 2021; Scandolara et al., 2017). Thus, we expect to find variations in the thermal structure throughout the main domains. To this end, we correlate our model with the long-wavelength potential field, seismic tomography and crustal thickness data to check whether the determined model can map features presented in independent datasets. We aim to interpret crustal-scale structures, providing insights into the geochronological province models of the Craton as well the presence of magnetic sources in the mantle. Additionally, we test if these constraints bring new information to the metallogenic systems of the Craton. The findings of this paper unlock value for future research in the Amazon Craton.

2. Geological Setting

The Amazon Craton is located in the northern portion of the South American continent. Currently, the proposed models describe it as a composition of several geochronological provinces (Santos et al., 2000; Tassinari & Macambira, 1999). Despite small differences, all models propose an Archean Nuclei in the Craton eastern portion with an arc evolution to southwestern. Juliani et al. (2013) demonstrated some incompatibilities of these geochronological models regarding magnetic and gravity data based on the E-W structures that cross the entire Craton. Thus, due to lacking consensus in the literature, we describe the geological features using the model proposed by Vasquez, Rosa-Costa et al. (2008) (Figure 1), an updated version of Santos et al. (2000), because it has more similarities with the results presented in this research.

According to Santos (2003), the Carajás Province presents the most relevant evidence of the Archean basement and divided it into two domains, the Rio Maria Granite-Greenstone (Meso-Archean 3.0–2.86 Ga) and Carajás

(2.76–2.55), which is associated with an Archean rift. Meso-archean magmatism is typical of Granite-greenstone Terrains (TTGs). The Greenstones (2.97–2.90 Ga) are composed mainly of komatiites and tholeiitic basalts. The Carajás domain includes metavolcanics, banded iron formations and granitoids. The basement is highly deformed, consisting of meso and early Archean granitoids.

The Transamazonas Province is divided into two domains, Bacajás and Santana do Araguaia. In general, the Santana do Araguaia late Archean is represented by metagranodiorite, metatonalito and monzogranite of ages of 3.06, 2.85, 2.67–2.34 Ga, respectively. Granodiorite and enderbites occur in the early Orosirian while older Archean rocks show evidence of EW and NW foliations (Corrêa & Macambira, 2014). The Bacajás Domain basement is formed by Archean orthogneiss (2.85–2.5 Ga). Mafic rocks, intermediate metavolcanics and meta-granitoids (2.36–2.31 Ga) are related to the amalgamation of an island arc to the Archean continent. Granitoids with ages of 2.21 to 2.11 Ga are associated with the onset of an active margin. Rhyacian granitoids and charnockites are evidence of a collision in 2.11 Ga. while granitoids and charnockites typical of post-collisional stages are formed between 2.08 and 2.07 Ga. The final stage marks magmatic pulses of 1.98 Ga related to the granitoids with anorogenic signature (Vasquez, Macambira, & Armstrong, 2008).

Early Orosirian also occurs in the Tapajós Domain as supracrustal retro arc basin, calc-alkaline granitoids and orthogneiss on amphibolite facies, as well as in calc-alkaline silicic volcanism and syeno and monzogranite with biotite. The latter rocks are mylonitized with porphyroclastic fabric. Also, calc-alkaline rocks with island arc signature are formed in the late Orosirian, this conclusion, however, is still controversial. In 1.88 Ga, felsic volcanic rocks and A-granites predominated in the Tapajós and Iriri-Xingu domains. The boundary between the Tapajós and Iriri Xingu Domains is controversial since the main difference between these volcanic rocks is their isotopic ϵ Nd signatures, typical of juvenile sources of 2.1 Ga in the first and related to Archean rocks in the latter (Vasquez, 2014).

According to Scandolara et al. (2017), the region covering the Jurueña Domain and Sunsás (also known as Jamari Terrane) Provinces is related to Jurueña orogeny. These authors understood it as a long live accretionary arc with a collision between 1.69–1.64 Ga. The region was reworked by intraplate orogeny between 1.42–1.37 Ga and ductile transpressive structures of 1.20–1.12 Ga.

3. Methods

3.1. De-Fractal Method

The de-fractal method (Salem et al., 2014) combines the spectral peak and centroid methods (Blakely, 1995; Connard et al., 1983; Okubo et al., 1985; Tanaka et al., 1999). These methods assume (a) a magnetic layer with infinite extension in all horizontal directions, (b) the depth to the top is small compared to the horizontal scale, and (c) magnetization is a random function of x and y .

According to Blakely (1995), the power-density spectrum of the magnetic field is given by

$$\varphi_{\Delta T}(k_x, k_y) = A(k_x, k_y) \cdot \varphi_M(k_x, k_y) \cdot (e^{-kZ_t} - e^{-kZ_b})^2 \quad (1)$$

where $A(k_x, k_y)$ is a function of the vector magnetization directions and magnetic field, $\varphi_M(k_x, k_y)$ is the power spectrum of magnetization. Z_t and Z_b are the depths to the top and bottom, respectively. The parameters k_x and k_y are the wavenumbers in x and y directions, respectively.

$$k = (k_x^2 + k_y^2)^{1/2} \quad (2)$$

According to Tanaka et al. (1999) and Ross et al. (2006), $\varphi_M(k_x, k_y)$ and the radial average of $A(k_x, k_y)$ are constant, leading to Equation 3:

$$\varphi_{\Delta T}(k) = C(e^{-kZ_t} - e^{-kZ_b})^2 \quad (3)$$

where C is a constant. For wavelengths shorter than twice the thickness of the magnetic layer, and taking the logarithm of Equation 3; depth Z_t is obtained by the slope at large wavenumbers of the natural logarithm of the azimuthally averaged power spectrum.

$$\ln(\varphi_{\Delta T}(k)) = B_1 - 2kZ_t \quad (4)$$

where B_1 is the intercept term obtained automatically in a least-square fitting. Similarly, for the long wavelengths, the azimuthally averaged frequency scaled power spectrum becomes:

$$\ln\left(\frac{\varphi_{\Delta T}(|k|)^{1/2}}{k}\right) = B_2 - kZ_0 \quad (5)$$

Where B_2 is the intercept term obtained automatically in a least-square fitting, Z_0 is the depth to the center of the layer, so that depth is obtained by adjusting a straight line to the small wavenumbers. Once Z_t and Z_0 are calculated, the depth to the bottom of the magnetic layer Z_b can be obtained by:

$$Z_b = 2Z_0 - Z_t \quad (6)$$

The peak method relies on the position (k_{peak}) of a power spectrum peak, which is a function of Z_t and Z_b . However, the peak does not always occur due to the fractal nature of source distribution or insufficient long-wavelength sampling in small window sizes (Ravat et al., 2007; Ross et al., 2006). This may yield a poorly sampled peak, thereby masking its position.

The de-fractal method (Salem et al., 2014) considers a self-similar magnetization distribution for the lithosphere. Therefore, we expect that, on average, magnetization contrasts are self-similar in both long and small scales. In fractal models, we assume the premise that magnetization is a random function with a power spectrum proportional to the wavenumber raised to power $-\beta$, where β is the fractal parameter, describing the degree of self-similarity of the magnetization (Maus et al., 1997). In a log-log scale, β is the power spectrum slope that may vary depending on the geological context. Long-range correlations increase proportionally to β . Moreover, magnetic field measured at the surface allows calculating $\alpha = \beta - 1$, where α is the fractal index (Maus & Dimri, 1994).

The de-fractal method assumes that the observed magnetic field spectrum is the product of the random magnetization model spectrum multiplied by $k^{-\alpha}$:

$$\varphi_F(k_x, k_y) = \varphi_{\Delta T}(k_x, k_y)k^{-\alpha} \quad (7)$$

Where $\varphi_F(k_x, k_y)$ is the observed magnetic field spectrum, and $\varphi_{\Delta T}$ is the spectrum of Equation 1. Therefore, isolating $\varphi_{\Delta T}$ from the equation, the fractal behavior can be removed from the observed spectrum as follows:

$$\varphi_{\Delta T}(k_x, k_y) = \varphi_F(k_x, k_y)k^{\alpha} \quad (8)$$

Once we remove the fractal effect, the centroid method is applied to obtain the depths Z_t and Z_b . We can then apply the spectrum peak method to forward model the spectrum associated with these depths (Salem et al., 2014).

In this work, we used the Fatiando a Terra python library of Uieda et al. (2013) to implement the de-fractal method while optimizing it to apply in large areas. A summary is shown in Figure 2.

3.2. Data

The data used is a compilation of several airborne magnetic surveys covering the study area. We use the compilation of Correa (2019) - Magnetic map of Brazil that resulted from the cooperation between the Geological Survey of Brazil—CPRM and several public institutions of Brazil (Figure 3). This version includes data not presented in the EMAG2 compilation (shadow area in the western portion). In Figure 3, the inset shows the airborne survey resolution. It shows robust continuity of long-wavelength anomalies between the transition from high resolution (line spacing of 500 m) to low resolution (line spacing higher than 1000 m) areas. The author corrected the database for long wavelengths following Hemant et al. (2007). Wavelengths larger than 330 km were removed using a high pass Gaussian filter and replaced by the MF7 model of Maus et al. (2008) (Figure 3). This procedure guarantees that this mesh covers wavelengths larger than the dimensions of each individual survey.

Window size (WS) is critical for obtaining reliable depths. There is a trade-off between selecting a WS large enough to sample the longest wavelength and not contaminating the spectrum with surrounding tectonic domains.

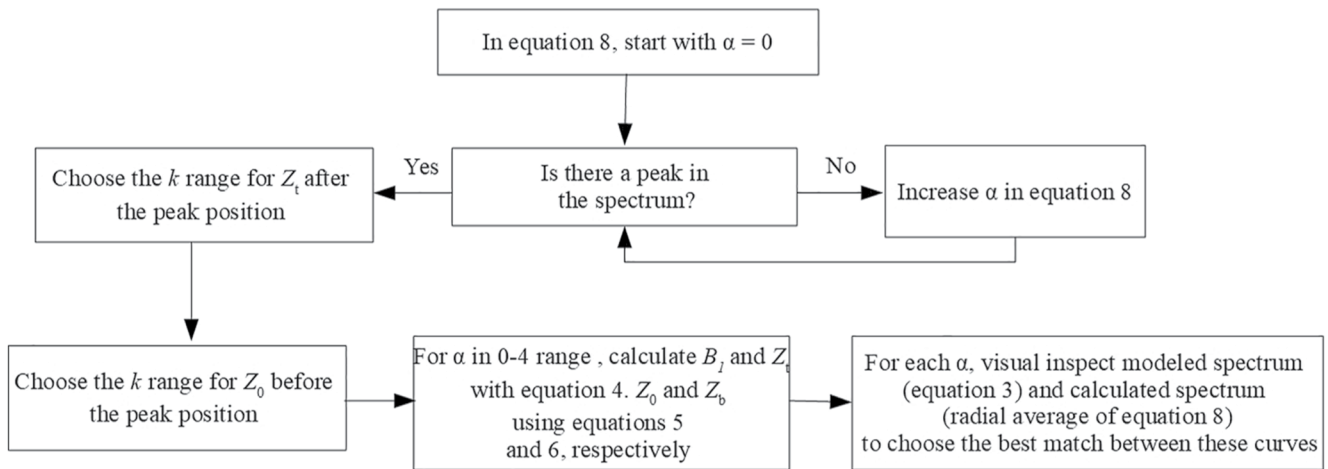


Figure 2. Flowchart of the implementation of the de-fractal method.

A WS at least six times the expected DBML is recommended to sample the long wavelengths. According to Albuquerque et al. (2017), the crustal thickness (CT) of the Amazon Craton is about 40 km; if we assume DBML close to CT, it would lead to WS 250×250 km, for example. Therefore, we have tested several WSs ranging from 200×200 to 500×500 km with 100 km increments (Figure 3). These details are described in the Supporting Information S1.

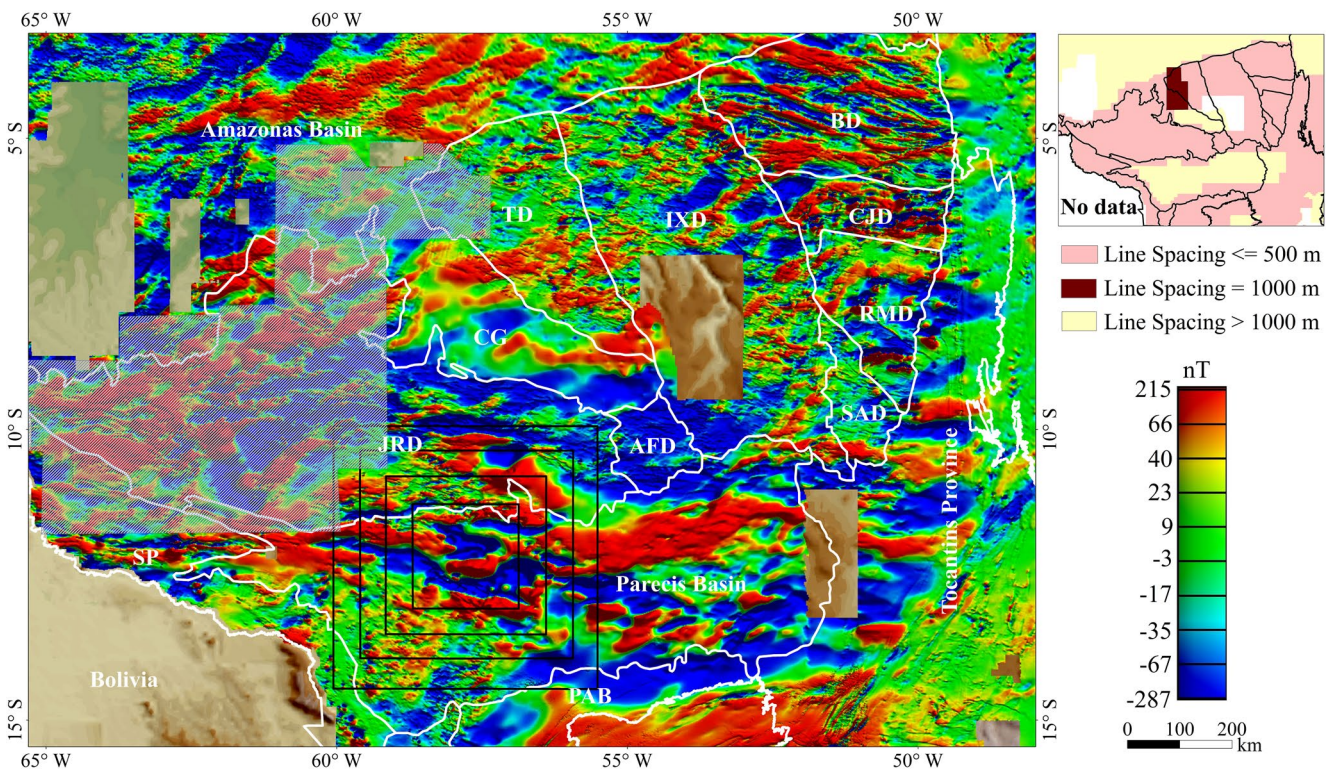


Figure 3. Magnetic anomaly in the study area. Geochronological Province limits adapted from Vasquez, Rosa-Costa et al. (2008). CJD: Carajás Domain, RMD: Rio Maria Domain, SAD: Santana do Araguaia Domain, BD: Bacajás Domain, IXD: Irriri Xingu Domain, TD: Tapajós Domain, AFD: Alta Floresta Domain, JRD: Juruna Domain, SP: Sunsás Province, CG: Cachimbo Graben, PAB: Paraguay-Araguaia Belt. The shadowed area in the western portion refers to the new data compared to EMAG2 compilation (Maus et al., 2009). Black squares are windows sizes of 200×200 , 300×300 , 400×400 and 500×500 km. We calculated Z_t for each window size to choose a trade-off between a window size large enough to sample the long wavelengths, but not too large so that adjacent tectonic domains are not included.

The depths are assumed to be in the center of the WS. We calculated DBML with 300×300 km WS where centers were moved by 50 km, totaling 543 estimates. Regions with relevant lack of data were not used. We gridded our estimates with a cell size of 12.5 km (1/4 of space between WS centers) using a minimum curvature interpolator.

4. Results

4.1. DBML Model and Its Uncertainty

In this section, we present our model and discuss its uncertainty. We compare our estimates with the centroid method, and we discuss the sensitivity of our estimates due to fractal index change. Figure 4 shows DBML and fractal index for the study area. Spatial gaps are due to sliding windows in areas where data are lacking. We opted to not interpolate these gaps to avoid artifacts.

The model average is 32 ± 11 km, almost 90% of the estimates are in the range of 20–50 km. The wavelength features bigger than the WS observed, indicate that inversion yielded real geophysical signals. The average fractal index is 1.5 ± 0.6 , while the low values concentrated mainly in the eastern portion of the Amazon Craton, increasing gradually westward, except for the western portion of Parecis Basin.

The stars in Figure 4a highlight depths calculated using the centroid method. As shown in Bansal et al. (2011) and Bouligand et al. (2009), disregarding the fractal behavior may yield to DBML overestimation whereas the correction process may lead to underestimation (Li et al., 2019). In general, the difference between the centroid and de-fractal methods is approximately 20 km, but it is more than 40 km for the JRD. Therefore, as the two methods returned depths with a difference of more than 20 km, we investigated α to check for overcorrections.

Figure 5 shows the relationship between the RMS and α for the star on JRD. Note that the RMS decreases with increasing α up to the curve inflection point at $\alpha = 2$, after which RMS increases abruptly. Ravat et al. (2016) argue that the best solutions are found neighboring the lowest RMS. For this case, visual inspection confirms the best matching is achieved in the lowest RMS ($\alpha = 2$) (Figure 5d).

Figure 6 shows the effect of α on Z_b . It is noteworthy that the method may return unrealistic solutions close to zero if α too large is used to correct the spectrum (Figure 6b). The RMS decreases up to $\alpha = 2$, moving Z_b from 60 km to ~ 20 km, however, RMS values are practically the same in the range of $\alpha = \pm 0.2$ at the lowest RMS point. Likewise, visual inspection of the matching is also subjective in this level of detail. This α range yields a difference of almost 20% in the Z_b value, in agreement with the 25% error obtained by Bouligand et al. (2009).

Li et al. (2019) pointed out that an iterative method to calculate the fractal index tends to overcorrect Z_b . Figure 7a shows the relationship between RMS and fractal index for the star in the Parecis Basin. We noticed that in the lowest RMS zone (about $\alpha = 3.8$), theoretically the region of best solutions, the de-fractal method tends to overfit the spectrum in the large wavenumbers (Figure 7b). The model yielded $Z_b = 11$ km for $\alpha = 3.8$, which is unreasonable considering a continental stable craton. Thus, choosing the solution just by looking at the lowest RMS zone may be misleading. To avoid that effect, we visually inspect all the range of solutions and the best match was found for $\alpha = 1.8$, which returned $Z_b = 28$ km (Figure 7c).

The fractal behavior has been corrected using a constant fractal parameter regardless of the changes in the geological environment (Bansal et al., 2011; Bouligand et al., 2009; Li & Wang, 2013). Nevertheless, Figure 8 shows DBML models using constant corrections. As expected, the depths decrease as α increases. For $\alpha = 0$, the average Z_b is 55 ± 16 , with some depths over 120 km. On the other hand, for $\alpha = 3$, the average Z_b is 12 ± 10 , which indicates overcorrection. For α equals 1 and 2, the models yield more reasonable results, despite low variations through the Archean crust and the proterozoic orogens. The reason for this is that the eastern portion presents α close to zero (Figure 4b) so that the Archean crust depths are underestimated for α equal to 2 and 3.

Pilkington and Todoeschuck (1993) used shallow borehole data and measured an α variation of 1.4 from sedimentary to igneous rocks. Leonardi and Kumpel (1996) found α over 3 for metamorphic rocks lifted during the Variscan Orogenesis. Bansal et al. (2010) also noticed that α decreases with depth in susceptibility logs. Therefore, an average α to correct the fractal behavior of the magnetic layer is already a simplification of the problem.

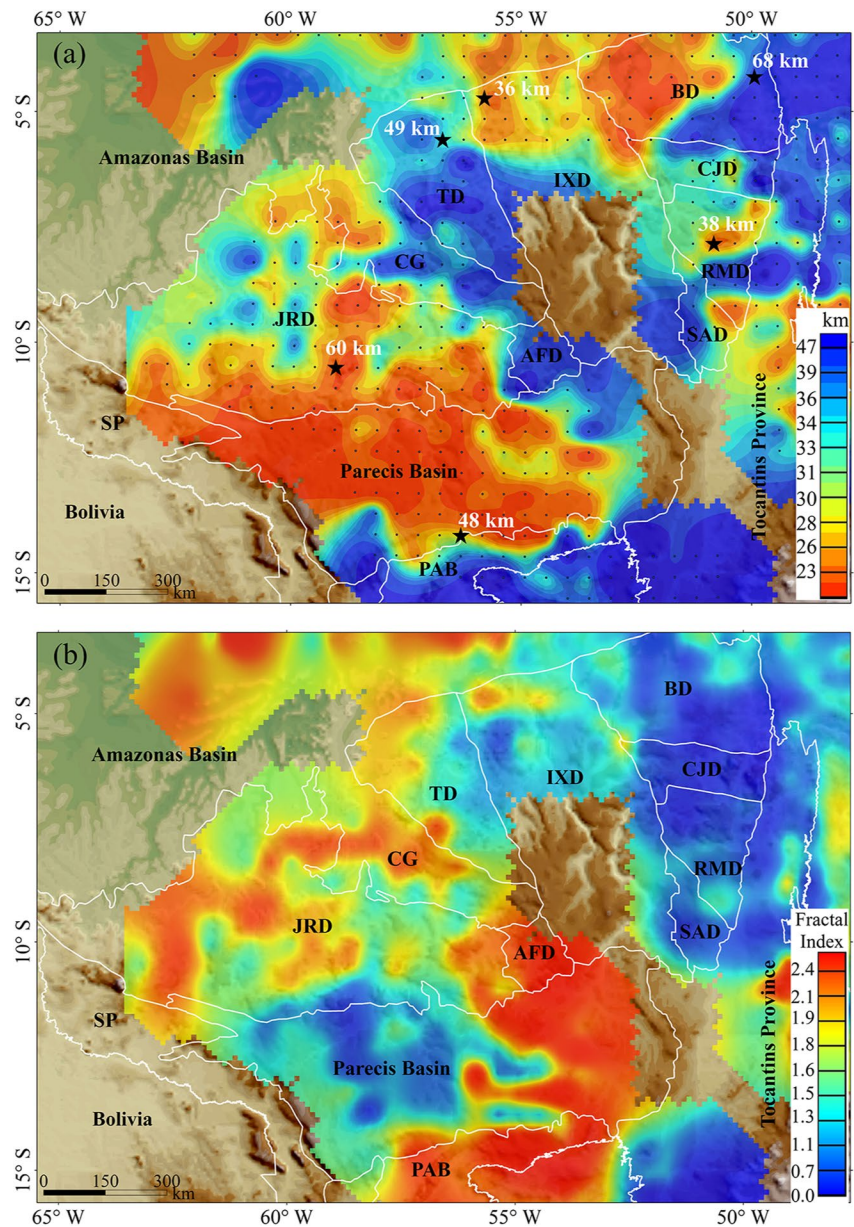


Figure 4. (a) DBML model of the study area where depths vary from 10 to 80 km for the study area. Black points are the centers of 300×300 windows used. Stars represent depths obtained by the centroid method. White lines are the limits of the Geochronological Province model adapted from Vasquez, Rosa-Costa et al. (2008). CJD: Carajás Domain, RMD: Rio Maria Domain, SAD: Santana do Araguaia Domain, BD: Bacajás Domain, IXD: Iriri Xingu Domain, TD: Tapajós Domain, AFD: Alta Floresta Domain, JRD: Juruena Domain, SP: Sunsás Province, CG: Cachimbo Graben, PAB: Paraguay-Araguaia Belt. The image shows wavelength features bigger than the WS, indicating that inversion yielded a real geophysical signal. Spatial gaps in the model denote a lack of magnetic data in these regions. (b) Fractal index of the study area. This index measures the degree of fractal behavior. Note that the Archean crust (eastern portion) present low values that increase toward the Proterozoic belts.

As our varying α model is more sensitive to the geologic domains, we consider it the preferred model for this study.

Li et al. (2017) and Idárraga-García and Vargas (2018) calculated regional DBML models based on EMAG2 compilation, even though the lack of data in this compilation is relevant for our study area (Figure 3) (Maus et al., 2009). Li et al. (2017) applied the centroid method with a constant scaling factor and obtained depths

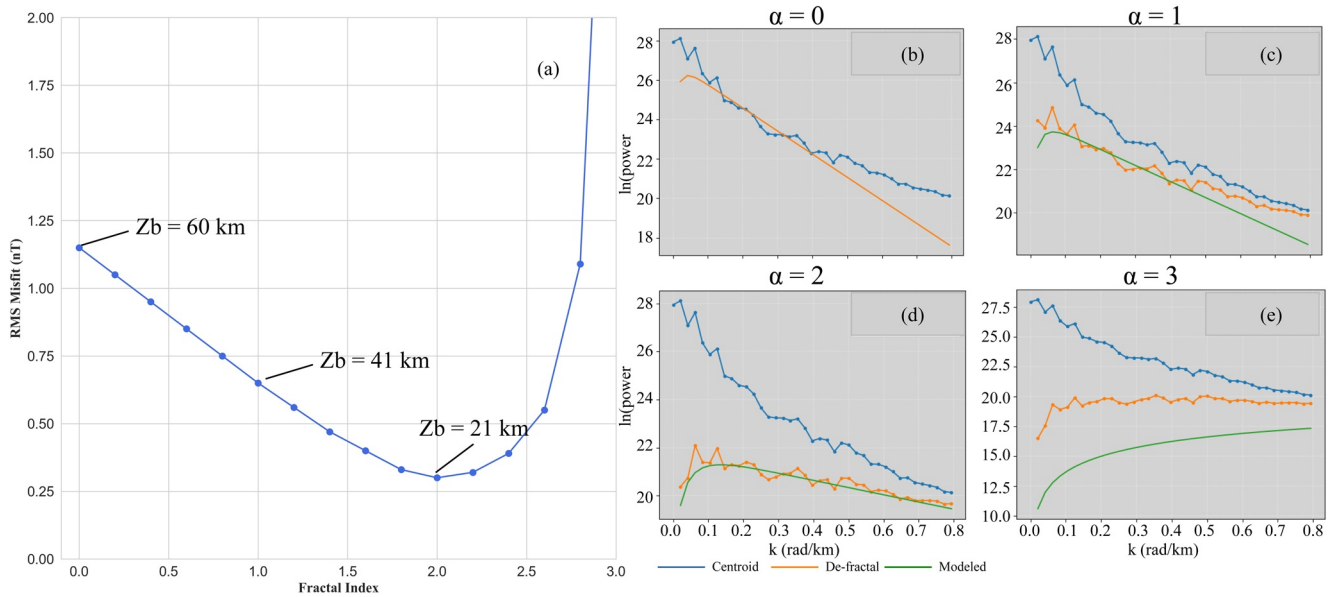


Figure 5. (a) RMS versus Fractal Index α for the star located in the RJD. (b) Visual matching between de-fractal and modeled curves for (b) $\alpha = 0$, (c) $\alpha = 1$, (d) $\alpha = 2$, (e) $\alpha = 3$. Note that the best matching is achieved at the lower RMS.

in the 30–45 km range for the South American continent. Depths are consistently larger than 40 km even for the Neoproterozoic Orogens and Phanerozoic Basins. Idárraga-García and Vargas (2018) considered an uncorrelated source model and applied the forward modeled spectral peak method. For our study area, their model also varies from 20 to 50 km, we note correlations with shallow values in BD, IXD and Parecis Basin, and deep ones in TD. However, this model is not sensitive to the heterogeneities present in the Archean domains and the RDJ. In terms of absolute values, our estimates agree with these studies while presenting additional new features.

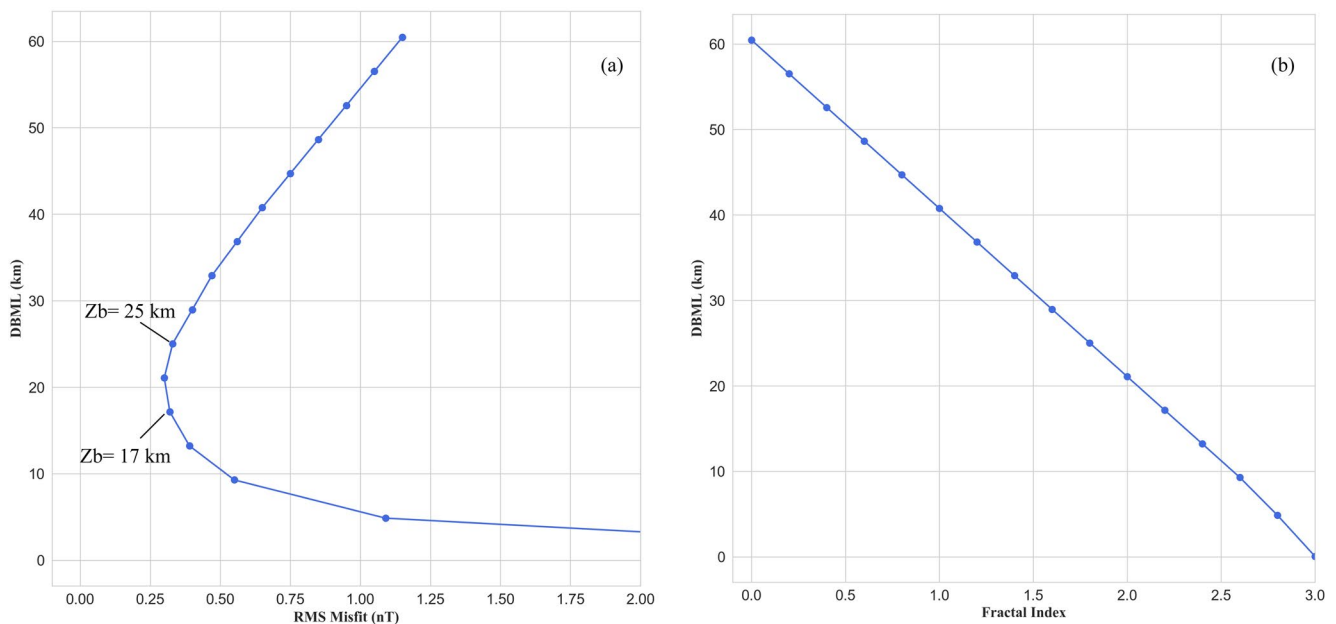


Figure 6. Sensitivity analysis of DBML for α at the lowest RMS zone. (a) Comparison between RMS and DBML. (b) Comparison between fractal index and DBML. Note that the RMS is similar in the range of $\alpha = \pm 0.2$ at the lowest RMS, so that Z_b varies nearly 20%.

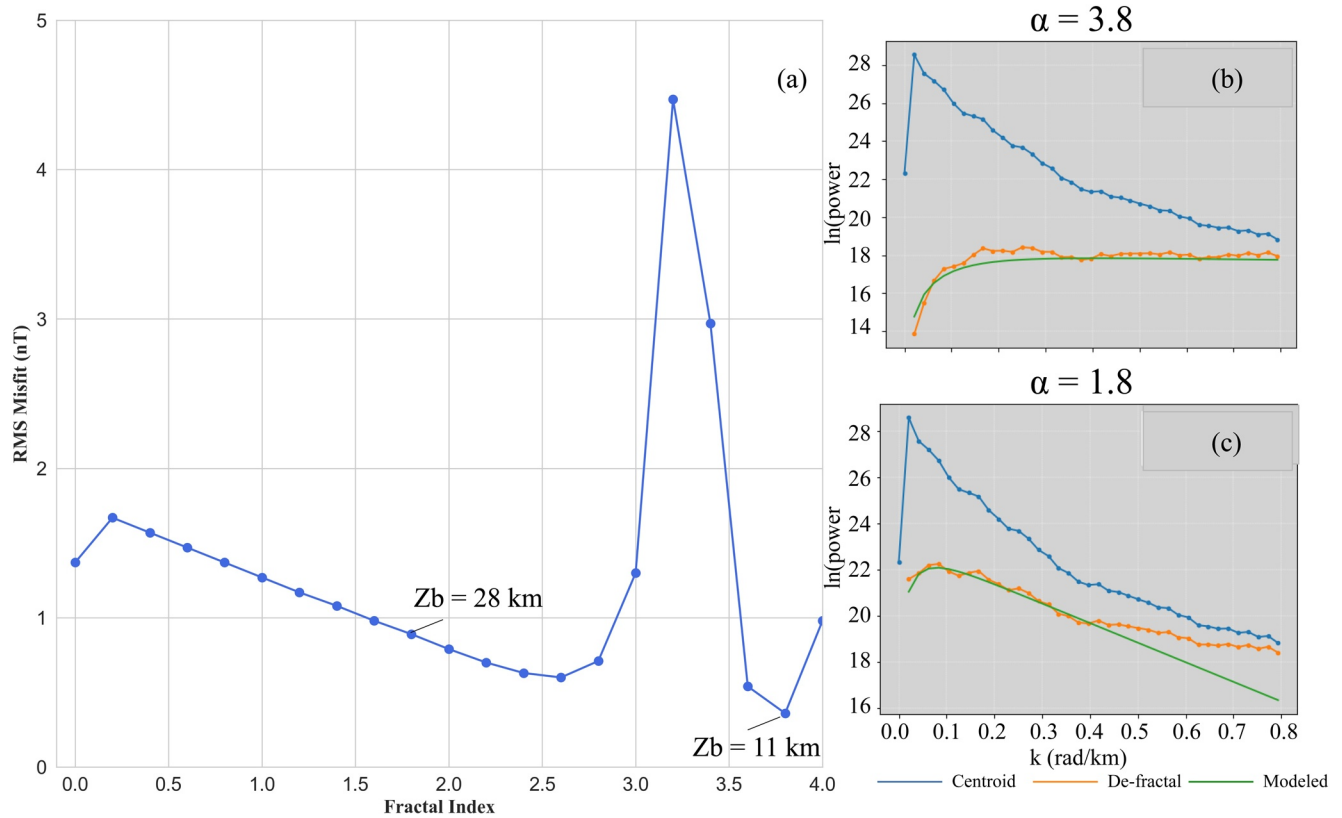


Figure 7. Effect of overcorrection due to large α . (a) Comparison between the fractal index and RMS. The lowest RMS is located at $\alpha = 3.8$. Visual inspection between de-fractal and modeled curves for (b) $\alpha = 3.8$. and (c) $\alpha = 1.8$. Because best visual matching is not always at the lowest RMS, it becomes necessary to visually inspect the entire α range.

5. Discussion

5.1. Comparison With Independent Datasets and Implications for the Tectonic Domains

In this section, we correlate the DBML model with magnetic, gravity and seismic tomography data. Figure 9a highlights the DBML model focusing on the long-wavelength changes. Assuming DBML as a proxy for the Curie temperature (more details in the Supporting Information S1), we interpret it as providing constraints for the rheology of the Amazon Craton. The elastic thickness and strength of the lithosphere are controlled primarily by temperature (Hyndman et al., 2009), therefore, a positive correlation is assumed between DBML and the rigidity of the lithosphere. We discuss if this premise holds in terms of the tectonic domains, while arguing why some limits of the geochronological model should be revised.

We enumerated main features from 1 to 9 to help the comparison with other datasets. To ensure the comparison in the long-wavelength scale, we applied an upward continuation of 15 km to the magnetic data and calculated the vertical derivative, and highlighted main crustal-scale breaks (CSB) (Figure 9b). Figure 9c displays the Bouguer anomaly of the Goco06s model (Kvas et al., 2021), and finally, Figure 9d shows the p -wave velocity for the slice of 135 km (I. S. L. Costa et al., 2020).

Although the resolution change between the DBML model and the vertical derivative is relevant, the limits between DBML features and CSB may approximately coincide or not. One reason is that DBML is calculated from a 300×300 km window, which is moved by steps of 50 km to cover the entire area. Moreover, DBML is a function of the mantle basal heat flow and also the heat flow generated due to radioactive elements within the crust, whereas the latter is mapping structural boundaries.

Vertical derivative shows that the CJD, RMD and BD present a high gradient compared to the IXD. In this sense, the Archean basement extends at least to CSB3. Likewise, we note similar Bouguer anomaly patterns

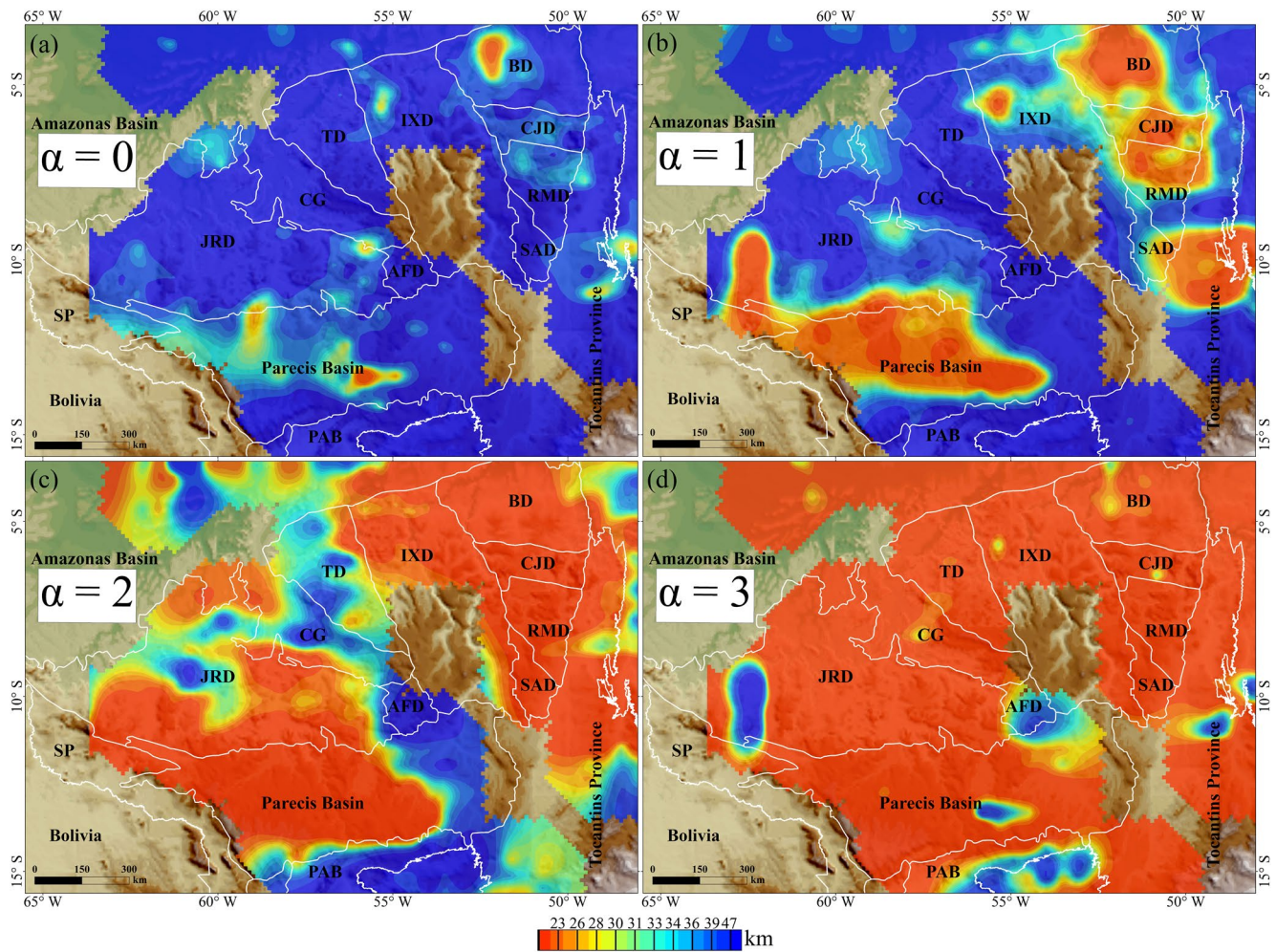


Figure 8. Maps comparing the depth to the bottom of the magnetic layer Z_b for different α values. White lines are the limits of the Geochronological Province model adapted from Vasquez, Rosa-Costa et al. (2008). CJD: Carajás Domain, RMD: Rio Maria Domain, SAD: Santana do Araguaia Domain, BD: Bacajás Domain, IXD: Irirí Xingu Domain, TD: Tapajós Domain, AFD: Alta Floresta Domain, JRD: Juruena Domain, SP: Sunsás Province, CG: Cachimbo Graben, PAB: Paraguay-Araguaia Belt. These maps are displayed with the same color bar. Note that the depths decrease for larger α . For $\alpha = 0$, depths are consistently higher than 50 km. For $\alpha = 3$, depths are consistently lower than 20 km. Therefore, we discarded these extremes. For $\alpha = 1$ and $\alpha = 2$, the models yield some direct correlations with tectonic domains.

with an NS trend, relatively high Bouguer (~ -30 mGal) surrounded by low signatures smaller than -45 mGal. This contour reflects the area of occurrence of IXD magmatism within BD, CJD, RMD domains. Additionally, the p -wave model indicates an increase in velocity from SAD-RMD to CJD-BD. In the DBML model, we noticed E-W boundaries crossing RMD and SAD (limits among features 1, 2 and 3). Figure 9b also displays E-W structures in the eastern portion of the Amazon Craton (CSB1 and CSB2). Except for feature 2, DBML tends to be shallow in RMD and deeper toward CJD, and even more eastern of CJD-BD (feature 4). F. G. Costa et al., (2020) argue that dome and keel structures in RMD and the linear belt pattern of CJD evidence a shift in tectonic style along Mesoarchean to Neoproterozoic. Deep values are observed in the center of the RMD (feature 2), whereas the DBML shallows toward the boundaries. Melting of the Archean mantle evidenced by greenstone belts influenced the rigidity of the crust, enhancing its mobility (Lourenço et al., 2016; Nebel et al., 2018).

Motta et al. (2019) suggest that CJD and BD are a single lithospheric unit with the same evolution at least up to 2.5 Ga based on Sm-Nd models and crystallization ages. Likewise, p -wave velocity corroborates this theory due to the high-velocity anomaly shared by BD and CJD. The DBML model shows that BD and CJD also share feature 4 with depths over 50 km. On the other hand, DBML shallows to 26–30 km west of BD, showing that these blocks are not a single lithospheric unit even though these shallow values may be influenced by the felsic

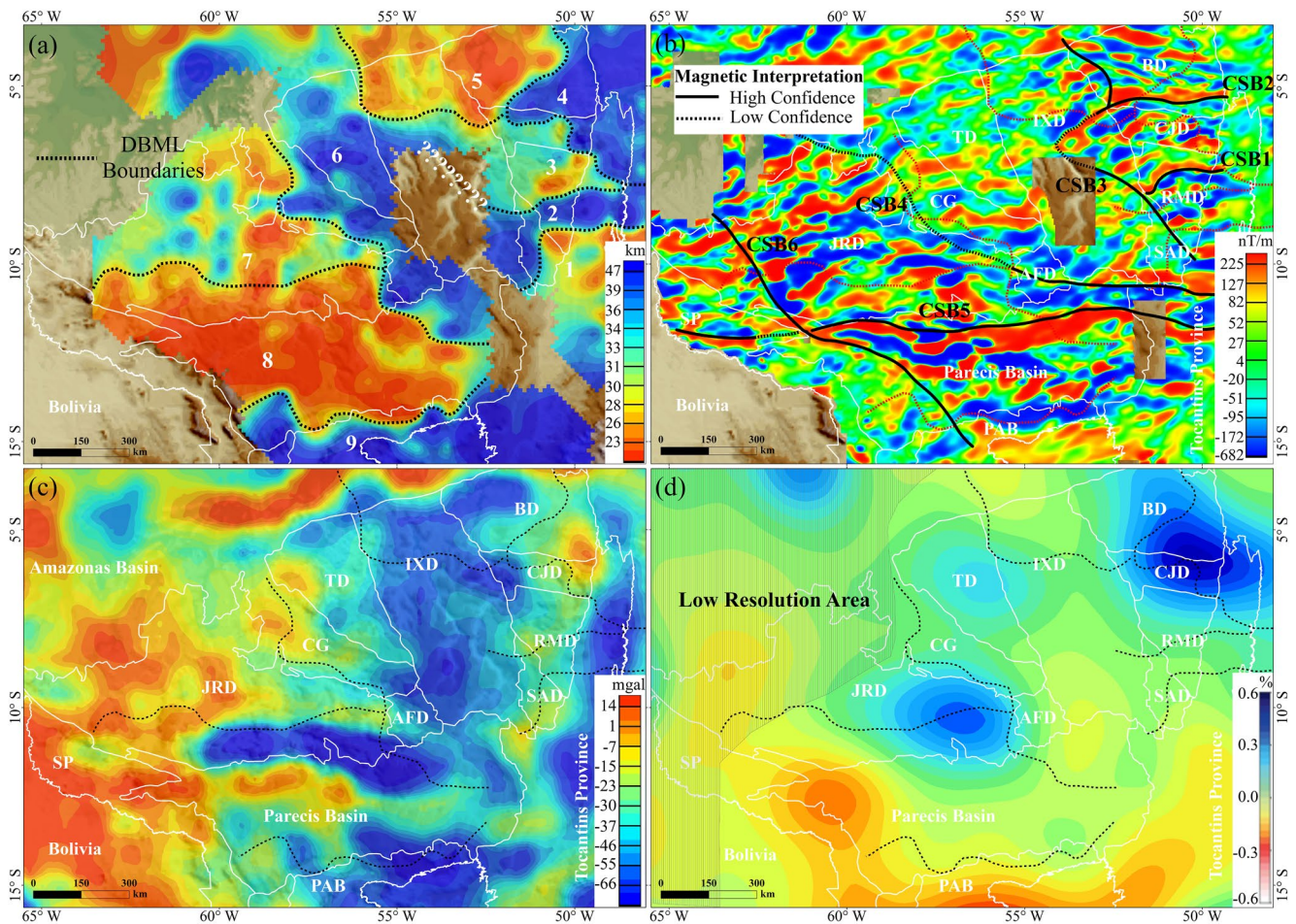


Figure 9. (a) DBML model with long-wavelength features interpreted. (b) Vertical derivative of the magnetic anomaly with upward continuation to 15 km. Crustal scale Breaks are highlighted in black. (c) Bouguer anomaly of the GOCO06s model (Kvas et al., 2021). (d) Slice of 135 km of the p -wave velocity model of I. S. L. Costa et al., 2020. White lines are the limits of the Geochronological Province model adapted from Vasquez, Rosa-Costa et al. (2008). CJD: Carajás Domain, RMD: Rio Maria Domain, SAD: Santana do Araguaia Domain, BD: Bacajás Domain, IXD: Iriri Xingu Domain, TD: Tapajós Domain, AFD: Alta Floresta Domain, JRD: Jurueña Domain, SP: Sunsás Province, CG: Cachimbo Graben, PAB: Paraguay-Araguaia Belt.

volcanic rocks of the Uatuma Silicic Large Igneous Province (Slip). In the vertical derivative data, we highlight CSB2 which distinguishes the NW trend at north from the NE and EW trend at south. We understand that this major boundary is the limit between BD and CJD.

IXD, TD and CG present similar signatures in the vertical derivative, on the other hand the Bouguer anomaly presents low values in IXD, and relatively high in TD and CG. Comparatively, DBML presents a correlation with the Bouguer anomaly with shallow values in IXD (feature 5) and deep ones in TD and CG (feature 6). Likewise, p -wave velocity presents low and high values in IXD and TD, respectively. The deep DBML, high Bouguer and high p -wave velocity in TD combined with shallow DBML, low Bouguer and low p -wave velocity in IXD corroborate the sensitivity of our model to the lithosphere rigidity.

There is no CSB separating the TD and IXD domains. Despite the lack of data in IXD, Feature 6 presents similar depths regarding Archean domains. Indeed, TD, IXD; and CJD, RMD share E-W trending structures (Figure 9b), implicating that the Archean crust extends to the western limit of feature 6. This result is corroborated by Sm-Nd T_{dm} ages of 2.5–2.9 Ga and ϵNd dominantly negative (Echeverri-Misas, 2015; Lamarão et al., 2005). Carneiro et al. (2019) and Juliani et al. (2021) argue the E-W structures are the result of the evolution of two continental magmatic arcs partially overlapped in TD. However, DBML and Bouguer anomaly do not present transitions in their signatures from south to north, neither E-W CSB is interpreted in the TD.

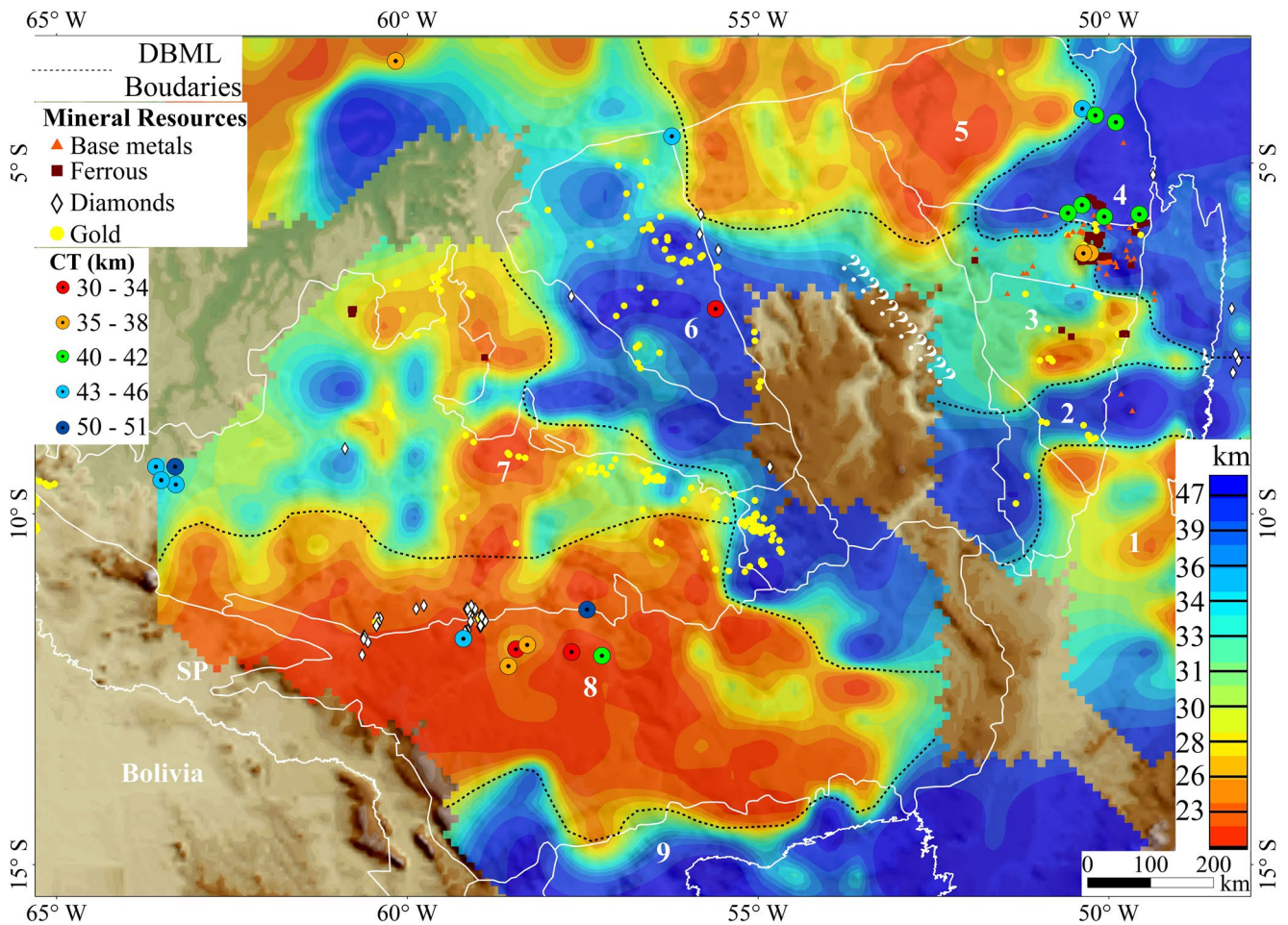


Figure 10. DBML model of the study. We also displayed the main mineral deposits and the crustal thickness estimates of Albuquerque et al. (2017). White lines are the limits of the Geochronological Province model of Vasquez, Rosa-Costa et al. (2008). The DBML features are enumerated from 1 to 9. The DBML boundaries represent changing composition and/or rheology of the Amazon Craton lithosphere.

The limit between TD and JRD is hard to define since both domains share NE-trending structures in the vertical derivative image and a similar Bouguer anomaly pattern. CSB4 marks the transition from NE to EW-trending structures. This major boundary and also the limits between features 6 and 7 support that the basement of CG is the TD.

JRD displays two main patterns (features 7 and 8) that are also present in the high contrasts of the Bouguer anomaly, especially in the eastern portion. However, this domain is probably the most heterogeneous as shown in Figures 9b and 9c. DBML deepens in the western portion of JRD as the Bouguer anomaly increases. Likewise, the vertical derivative transitions from high to low magnetic gradient from western to eastern and is marked by CSB6. We also noticed that the DBML values decrease toward the Parecis Basin (feature 8), especially in the western portion while DBML deepens toward the Paraguay-Araguaia Belt (feature 9).

The DBML model also correlates with the Bouguer anomaly in the JRD. Feature 7 coincides with a region of relatively high Bouguer anomaly, where granites and migmatites with ϵNd varying from -3.9 to 2.52 are evidenced (Rizzoto et al., 2019). In contrast, feature 8 associated with shallow DBML and low Bouguer anomaly (especially in the eastern portion) consists mainly of volcanic rocks with dominantly negative ϵNd , and sedimentary rocks as well. The p -wave model does not separate these features probably due to resolution issues. DBML shallows toward the Parecis Basin (feature 8), especially in the western portion. We attribute this signature to the thinner crust in this portion as part of the basin evolution. Scandolara et al. (2017) interpreted the CSB5 as the suture

Table 1
DBML and CT Along the Geological Domains

CT (km)	Zb (km)	Domain
46	31	BD
41.4	38	BD
42.3	41	BD
40.9	38	BD
39.4	50	BD
40.8	56	BD
41.8	49	BD
37.8	31	CJD
37.7	30	CJD
50	21	JRD
40.4	19	PB
33.4	27	PB
38.2	24	PB
43	18	PB
30	18	PB
34.9	26	PB
50.5	35	JRD
44.9	35	JRD
43	33	JRD
43.2	34	JRD
42.7	33	TD
34.4	45	TD

Note. BD, Bacajás domain; CJD, Carajás domain; JRD, Juruena domain; PB, Parecis Basin; RD, Rondônia domain; TD, Tapajós domain.

zone. Even though, we note a decrease in depth towards the Parecis Basin, our model does not map relevant contrasts through CSB5, probably due to resolution issue.

Because we are comparing magnitudes resulting from the compound influence of the upper mantle, lower and upper crusts, local crustal heterogeneities in one of these layers influence the resulting response. Recurring long-wavelength features in independent datasets support the assessment that our model is robust and recovered real geophysical signals.

5.2. Comparison With Crustal Thickness and Metallogenic Implications

The Mohorovic discontinuity (Moho) used to be understood as a boundary that limits the presence of magnetic materials (Wasilewski et al., 1979; Wasilewski & Mayhew, 1992). However, it has been shown that the mantle in the presence of fluids can become serpentinized and magnetite forms as a secondary phase (Blakely et al., 2005; Correa et al., 2016). Ferré et al. (2013) performed magnetic measurements of mantle xenoliths and concluded that magnetite occurs systematically in the mantle of cratons. We test this hypothesis for the Amazon Craton by comparing CT with DBML values. Therefore, we provide information on the presence of magnetic minerals in the mantle, which is associated with serpentinization processes. However, CT data are scarce and cover just a few portions of the Amazon Craton.

The CT of the Amazon Craton is about 40 km, with the deepest values (>45 km) observed in the JRD (Figure 10), whereas the Parecis Basin presents shallow values (<35 km). This step is separated by the CSB5, which is associated with abrupt changes in DBML and Bouguer anomaly. High and low Bouguer anomalies aligned with long wavelengths are commonly associated with suture zones.

Table 1 summarizes the data regarding the respective tectonic domains. There is no data on RMD, SAD, IXD and CG. The average CT is 38 km in CJD and approximately 42 km in BD. Assuming no relevant change in CT

for these domains, DBML tends to be shallower, except for the feature 4 area, where it is deeper than 50 km. Supergiant iron-oxide-copper-gold (IOCG) deposits with metasomatic mantellic magmas signature (Teixeira et al., 2021) corroborate these results. There are two explanations for this metasomatic mantle signature (a) Teixeira et al. (2021) argue that mantle fertilization occurred under plume-related thermal stresses, and (b) Groves et al. (2010) consider the subduction process as the main cause.

DBML is shallower than CT in the western portion (JRD and PB). Although DBML appears to be consistently shallower than the CT, features 2, 4 and 6 are potential areas for the occurrence of serpentinization.

Howell et al. (2019) highlighted the importance of the metasomatized/serpentinized lithosphere mantle on controlling metallogenic signatures in post-subduction contexts. These authors attest that ore deposits form along the mantle to the upper crust pathway of hydrothermal fluids and magmas. Therefore, we compared our DBML model with the main gold, ferrous, diamonds, and base metal deposits to seek correlations, assuming that some of the magnetite presence in the mantle are derived from highly fluid environments (Figure 10). In CJD, we note a high density of the metal base and ferrous deposits associated with feature 4. The high correlation of this feature with high *p*-wave velocity indicates fertile mantle conditions, which makes this region favorable to world-class mineral deposits. In RMD, we observe a high correlation of gold deposits with shallow DBML and in the transition of features 2 and 1. The deep values of TD and AFD also present a good correlation with the gold deposits. On the other hand, this pattern is not clear in JRD, but the gold occurs preferentially in feature 7. Diamonds seem to occur preferentially in the northern border of the Parecis Basin, which is close to the boundary between features 7 and 8 and the CSB5.

6. Conclusions

We have implemented the de-fractal method to calculate the DBML for the south-central portion of the Amazon Craton. We compared the DBML with the long-wavelength potential field, seismic tomography and crustal thickness data. The correlation between structures mapped from the independent dataset supports the assessment that our model recovered real geophysical signals.

We assumed DBML as the depth to Curie temperature to constrain the thermal structure. As the lithosphere rigidity is primarily a function of temperature, we tested if the shallow DBML correlates with the low p -wave velocity and high Bouguer anomaly and vice versa. We note this hypothesis is valid for features 1, 2, 3, 4, 5, 6 and 8, as well as for features 7 and 9 to a lesser extent. Even though these features are close in the Paleo-Mesoproterozoic domains, local crustal heterogeneities or the mantle make the interpretation more complicated. Heat flow data are necessary to close this gap, and we suggest our model as an input for future survey planning.

Our model suggests that the limits between Carajás and Bacajás domains should be revisited as well as the limits between TD and IXD. The low Bouguer anomaly and shallow DBML (feature 5) in IXD mark the area of influence of the 1.88 Ga Slip. It is also likely this event interferes with the signature of feature 3.

Deep DBML values in TD corroborate the presence of an Archean crust in the western portion of the Amazon Craton. Our model does not present an EW-trending DBML feature or EW-trending CSB in TD. The NW-trending limit between features 6 and 7 as well CSB4 are plausible structures marking the limit between TD and JRD.

Moreover, we also correlate our model with the main mineral deposits. We identified a DBML anomaly, suggesting the mantle is serpentinized in the eastern part of the Carajás domain. The high correlation of this feature with the high p -wave velocity indicates fertile mantle conditions, making this region favorable for world-class mineral deposits. Furthermore, we note a high correlation of gold deposits with shallow DBML in RM and eastern TP.

Data Availability Statement

The magnetic data used in this study were published in Correa (2019). Gravity data are available through Kvas et al. (2021). Crustal thickness data are available in Albuquerque et al. (2017). We used mineral deposits locations published in I. S. L. Costa et al., 2020.

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CAPÍTULO 3

ESTRUTURA ELÉTRICA DA LITOSFERA DE UM ORÓGENO ACRESCIONÁRIO

Tectonic and metallogenic implications of a metasomatized mantle inferred from magnetotellurics and gravity data: An example of Tapajós Mineral Province, Amazon Craton

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Abstract

The Tapajós Mineral Province (TMP) is part of a long-lived accretionary orogen and is renowned for producing at least 750 t of gold, mainly in artisanal mining. World-class deposits have been linked with lithospheric electrical conductors using magnetotelluric (MT) data. On the lithospheric scale, an assessment of the potential for larger deposits or correlated commodities is absent in the TMP. This paper presents the first electrical resistivity model for the lithosphere of the TMP. The electrical resistivity derived from a 3D MT inversion model reveals a high resistivity zone ($> 10000 \text{ ohm.m}$) interpreted as the emplacement of granitoids in the upper crust. The mid-lower crust presents resistivity between 465 and 45 ohm.m , corresponding to depths of 10-35 km. There is a low resistivity zone below 35 km (45-10 ohm.m), which represents a hydrated upper mantle due to the long-lived arc processes. Based on gravity forward modeling, we estimate that the degree of serpentinization is less than 10% due to post subduction magmatism. The electrical resistivity model allowed for the interpretation of magnetic discontinuities at

depth and revealed how part of the TMP's crust was exhumed. The high correlation between the position of the deposits and electrical conductors corroborates that MT is a valuable tool for detecting fossil fluid pathways relevant for finding new mineral deposits.

Introduction

The Tapajós Mineral Province (TMP) is a relevant gold producer in Brazil, having produced at least 750 t in the last decades (Bettencourt et al., 2016). Its evolution is related to accretionary orogen with evolution between 2.03 and 1.87 Ga. Despite well-documented arc magmatism, the lack of strong kinematic and geological evidence that indicates a collision and the formation of a major suture zone makes it tricky for geologists to join the pieces of the puzzle. Furthermore, no classical suture was observed on geophysical signatures of potential field data (Carneiro et al., 2019).

In terms of mineral resources, accretionary orogens form Au-Cu $\pm$ Mo deposits due to granitic intrusions, which exsolve fluids and volatiles responsible for transporting, disseminating, and precipitating the metals in veins and veinlets (Huston et al., 2016; Vasquez, 2018; Cassini et al., 2020). Therefore, mapping these fluid pathways is fundamental to unlocking the metallogenic potential of the TMP.

Resistivity images of the crust have been used to map crustal-scale discontinuities related to fossil fluid flows (Selway et al., 2009; Thiel et al., 2016; Heinson et al., 2018; Comeau et al., 2021; Vadoodi et al., 2021). Moreover, the resistivity can be used as a proxy for strain localization, making it possible to identify brittle-ductile transitions within the crust.

Consequently, we present an electric resistivity model for the lithosphere of the TMP in order to constrain the crustal framework and evaluate the mantle composition. A 160-kilometer MT profile was carried out on the well-known Gold dirty road

(Transgarimpeira). We test if our model can map the main airborne magnetic discontinuities. As the TMP presents gaps in the crustal level, we also check if the resistivity model can identify exhumation structures and distinguish strain levels. Then, we compared our results with the available gravity data in order to address the crustal thickness and test the hypothesis of a metasomatized mantle and its metallogenic implications.

Geological background

The Tapajós Mineral Province is located in the Amazon Craton (Fig. 3.1). The Amazon Craton is composed of the Guiana and Central Brazil Shields, which are divided by the Amazon Phanerozoic Basin. The tectonic evolution of the Amazon Craton involves an Archean nucleus in the east and the accretion of several magmatic arcs, whose ages decrease towards south-western (Santos et al., 2000; Vasquez et al., 2008). The Tapajós Mineral Province corresponds to the southern part of the Tapajós – Parima (Santos et al., 2000) Province, dated between 2.03 and 1.87 Ga. it is bounded on the east by the Iriri Xingú Domain, and on the north and west by the Amazonas and Alto Tapajós basins, respectively.

The tectonic evolution of the Tapajós-Parima Province used to be explained by two orogenesis that started with island arc magmatism, collision with a continent, continental arc, and post-orogenic to anorogenic magmatism (Santos 2001; Santos et al. 2004). However, the lack of collisional evidence such as S-type granites, fold and thrust belts, and HPLT metamorphism undermined this theory. Vasquez et al. (2017), based on peraluminous leucogranites (Vasquez et al., 2002), proposed only one collision between 2005 and 2003 Ma, followed by transcurrent post-collisional magmatic events from 2.0 to 1.96 Ga, and then by extensional events from 1.90 to 1.87 Ga. Based on magnetic data, Carneiro et al. (2019) noticed a continuity of deep E-W trending structures from the

Archean nuclei to the Tapajós-Parima Province, thereby discarding the island arc hypothesis. More recently, Cassini et al. (2020) proposed a long lived subduction that endured more than 130 Ma in four stages i) continental magmatic arc formed by anhydrous magmas (2.0 – 1.98 Ga), ii) evolved magmatic arc formed by hydrous and oxidized intrusive suites (1.98 – 1.95 Ga), iii) slab flatenning and mantle metasomatism, which eventually melts and triggered the formation of high K to shoshonitic, metaluminous to mildly peraluminous suites (1.89 – 1.88 Ga), iv) Slab break off and asthenospheric upwelling led to the collapse of the orogen (1.88 – 1.87 Ga). It is noteworthy that the quiescence period between stages two and three probably involved the switching from retreat accretionary to advanced accretionary orogenesis (Cawood et al., 2009), which is also used to explain the evolution of similar rocks in the Alta Floresta Domain (Trevisan et al., 2021).

In terms of lithostratigraphy, the Cuiú-Cuiú Complex shows calc-alkaline affinity and represents the oldest magmatism of the Tapajós Mineral Province (2033-2005 Ma, Vasquez et al., 2017). The magmatism from 2002 to 1956 Ma shows high-K calc-alkaline and shoshonitic affinities and is represented by the granitoids of the Creporizão Suite and the equivalent volcanics of the Vila Riozinho formation (Lamarão et al., 2005). After a 50 Ma gap, high-K calc-alkaline magmatism occurs between 1907 and 1874 Ma represented by the Tropas and Parauari Intrusive Suites (Vasquez et al., 2017). The Maloquinha Intrusive Suite with its acidic alkaline signature represents the last relevant magmatism of the Tapajós Mineral Province. Simultaneously, volcanics rocks of the Iriri Group are formed in the titled Uatumã magmatism, which is thought to be part of a Silicic Large Igneous Province (SLIP) by Klein et al. (2012).

The thrust faults in the Tapajós Mineral Province are locally present in the rocks of the Jacareacanga basin and the Cuiú-Cuiú complex. Their trend is evidenced by NS-

foliation in the shales and gneisses. The sedimentary and igneous rocks of this early magmatic arc have undergone high-grade metamorphism and ductile deformation, but in the eastern part, the granites and volcanic rocks of this arc were preserved, thereby indicating exhumation of the western portion (Klein et al. 2001, Coutinho 2008, Vasquez et al. 2017). This evidence is also present in airborne magnetic data, where the two crustal levels change the total gradient background (Fig. 3.2a), though this feature is not present in the Bouguer anomaly (Fig. 3.2b) making one wonder how deep this structure is. These features were transposed by transcurrent tectonics marked by NW-SE brittle-ductile transcurrent faults, which controlled the positioning of Creporizão, Parauari, and Maloquinha intrusive suites (Klein et al. 2002, Santos and Coutinho 2008).

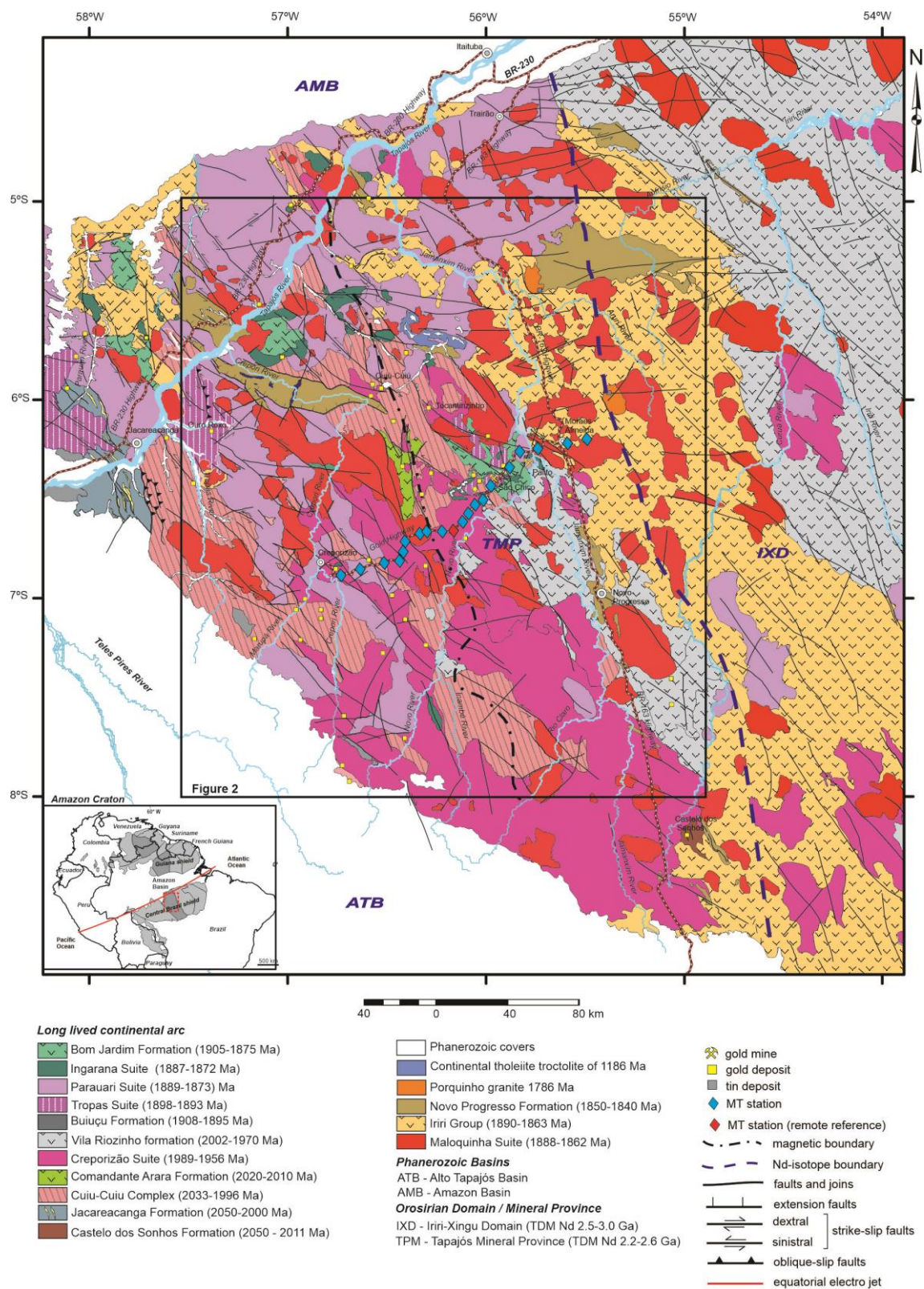


Fig. 3.1. Geological map of Tapajós Mineral Province (modified after Vasquez et al. 2008, 2017). Inset highlights the Tapajós-Parima geochronological province in the Amazon Craton (modified after Santos et al. 2000, 2004). Gold deposits of Vasquez et al. (2017).

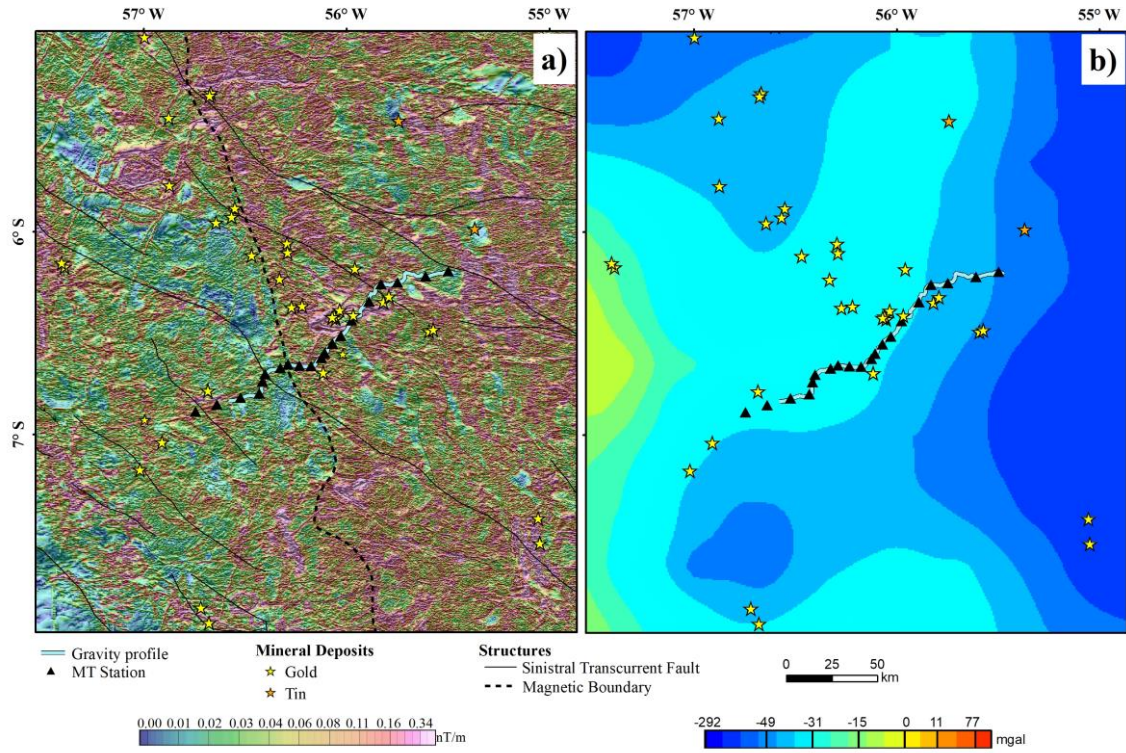


Fig. 3.2. A) Total Gradient (Li, 2006) of airborne magnetic data with 70% transparency overlaying Tilt Derivative (Verduzco et al., 2004). B) Bouguer anomaly of the model Goco06s (Kvas et al., 2021). Gold deposits of Vasquez et al. (2017).

Methods

MT data acquisition and analysis

MT is a passive electromagnetic technique that measures the time variations of both the magnetic field and the electrical field on the surface, and uses their orthogonality relation to calculate the resistivity of the subsurface (Cagniard, 1953). The depth of penetration (skin depth – δ) is proportional to the acquisition period (T) and medium resistivity (ρ). In the frequency domain, the field components are related by the equations: $\mathbf{E}(\omega) = \mathbf{Z}\mathbf{H}(\omega)$ and $H_z = -\mathbf{K}(\omega)\mathbf{H}(\omega)$

Where $\mathbf{Z}$ is impedance tensor (complex four-component), $\mathbf{K}$ is induction vector (complex two-component), $\mathbf{E}$ and $\mathbf{H}$ are the horizontal electrical and magnetic field vectors, H_z is the vertical component of the magnetic field, and $\omega = \frac{2\pi}{T}$ is the angular frequency.

We have collected 20 broad-band (BB) data sites. In addition, 10 of those had co-located long-period (LP) sites, forming a 160 km long SW - NE transect using magnetic north as a reference. Overall, the spacing is around 12 km for BB data and 24 km for LP data. Across the transition of the two magnetic blocks, we installed sites with closer spacing (Fig. 3.2). The survey was conducted with Metronix ADU07 equipment and Lemi 417M for BB and LP, respectively. Each BB site recorded data for around 40 hours, and LP recorded data between 10-20 days.

The MT data were processed, and the electrical impedances $\mathbf{Z}$ were estimated using robust techniques described by Egbert (1997). The study area may be under the influence of the equatorial electrojet - EEJ (Chave and Jones, 2012; Hamid et al., 2014). Padilha (1999) concluded that the EEJ can be used as a forcing field for MT surveys in the equatorial region, at least in the frequency range of 0.488-1000 Hz for a geomagnetic latitude of $\pm 3^\circ$. Figure 3.3 displays five channels recorded using LP equipment. One can note that E_y presents higher noise levels during the day and that the E_x is also affected in a minor order.

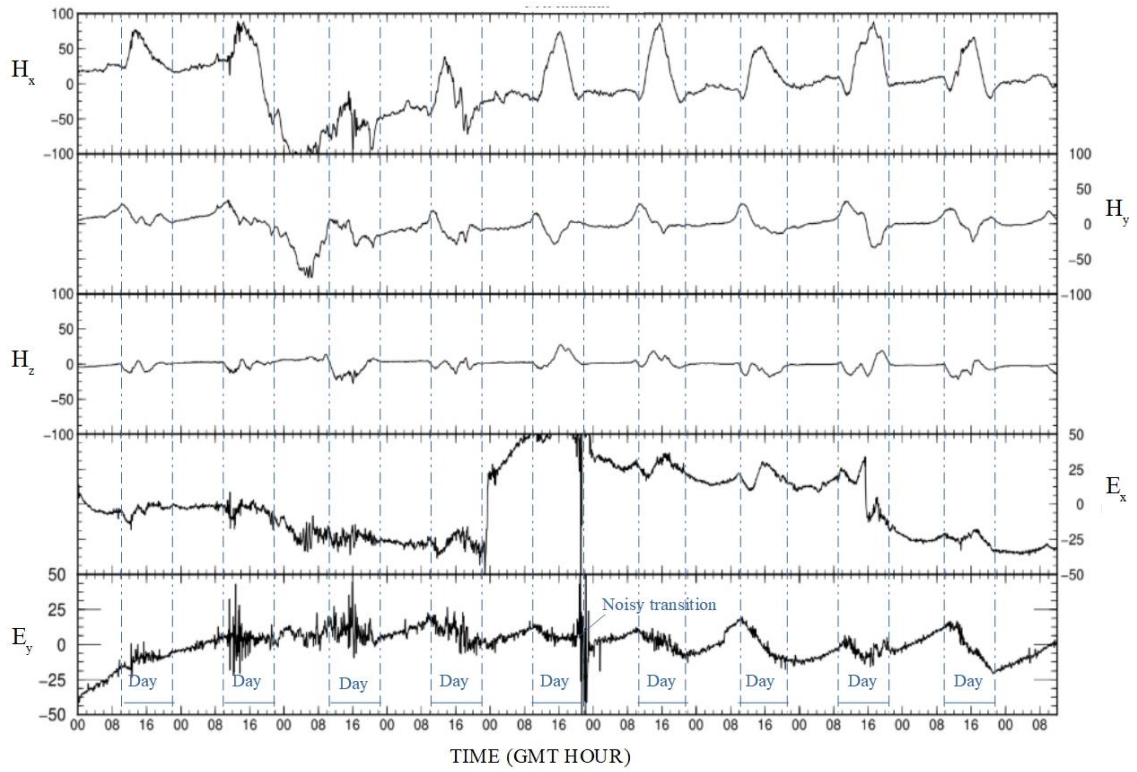


Fig. 3.3. Magnetic and electric channels measurements. Note that the EEJ reduce signal/noise ratio mainly in E_y direction, but also affects E_x channel.

Good quality BB data was achieved only up to 100 s for relevant part of the dataset. On the other hand, LP sites produced high-quality data for almost all stations up to 5000 s. This was possible due to remote station processing (Gamble et al., 1979) using only nighttime (Fig. 3.4). Otherwise, resistivity and phase curves yield noisy results. All resistivity and phase curves are available in the supplementary material.

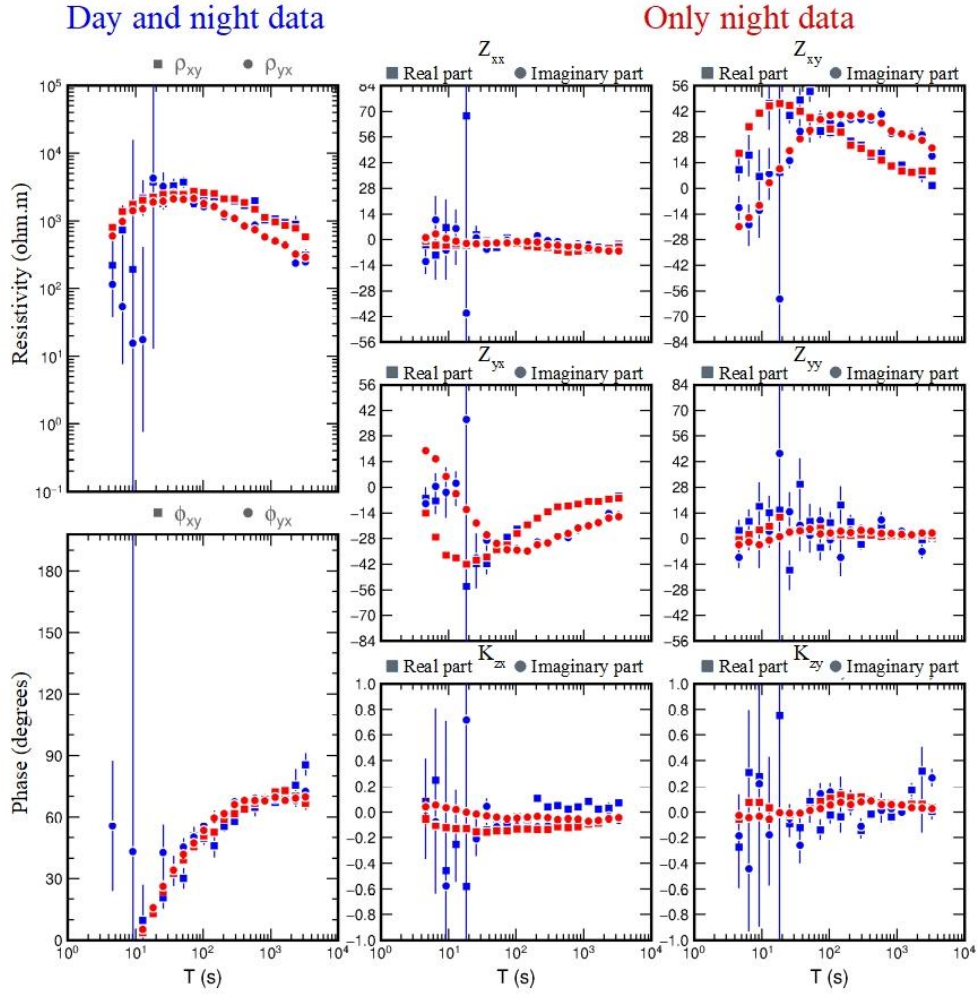


Fig. 3.4. Comparison of robust processing using both day and night data and only night data. Note that the impedance tensors which contain E_y or H_y yielded even more unstable solutions when day data are included.

Dimensionality analysis was undertaken using phase tensor analysis (Caldwell et al., 2004), which is implemented with the MTPy package (Krieger and Peacock, 2014). The phase tensor has the advantage of not requiring any assumptions about the medium's dimensionality, and it is not affected by galvanic distortion. It is formed by three rotational invariants, namely $\Phi_{\min}$, $\Phi_{\max}$ and skew (β). The latter is a measure of the asymmetry of the regional MT response being equal to zero for 1D and 2D resistivity distributions. $\Phi_{\max}$ and $\Phi_{\min}$ are the axes of the ellipse. The main axis tends to be parallel to the direction of principal induction current flow (Yin et al., 2021).

We considered ellipses with skew angle out of the range $|5|$ with 3D behavior. Phase tensor analysis shows that the data is preferentially 3D (Fig. 3.5). Moreover, we observe abrupt changes in the direction of $\Phi_{\max}$, which is also indicative of 3D behavior (Thiel et al., 2020). Figure 3.5b displays $\Phi_{\min}$, which is directly proportional to conductivity. We notice that for periods between 10^{-4} and 10^{-1} s, the crust presents low conductivity with regard to the granitoid terrain. This range period most likely reflects the upper crust. Therefore, we plot the phase tensor strike using this range as reference (Fig. 3.5c). For periods $< 10^{-1}$, the most relevant structures have N50E trending directions and rotates 10 degrees anticlockwise for longer periods between 10^{-1} and 10^3 . The wide range of possibilities for induction current flow confirms that 3D inversion is required.

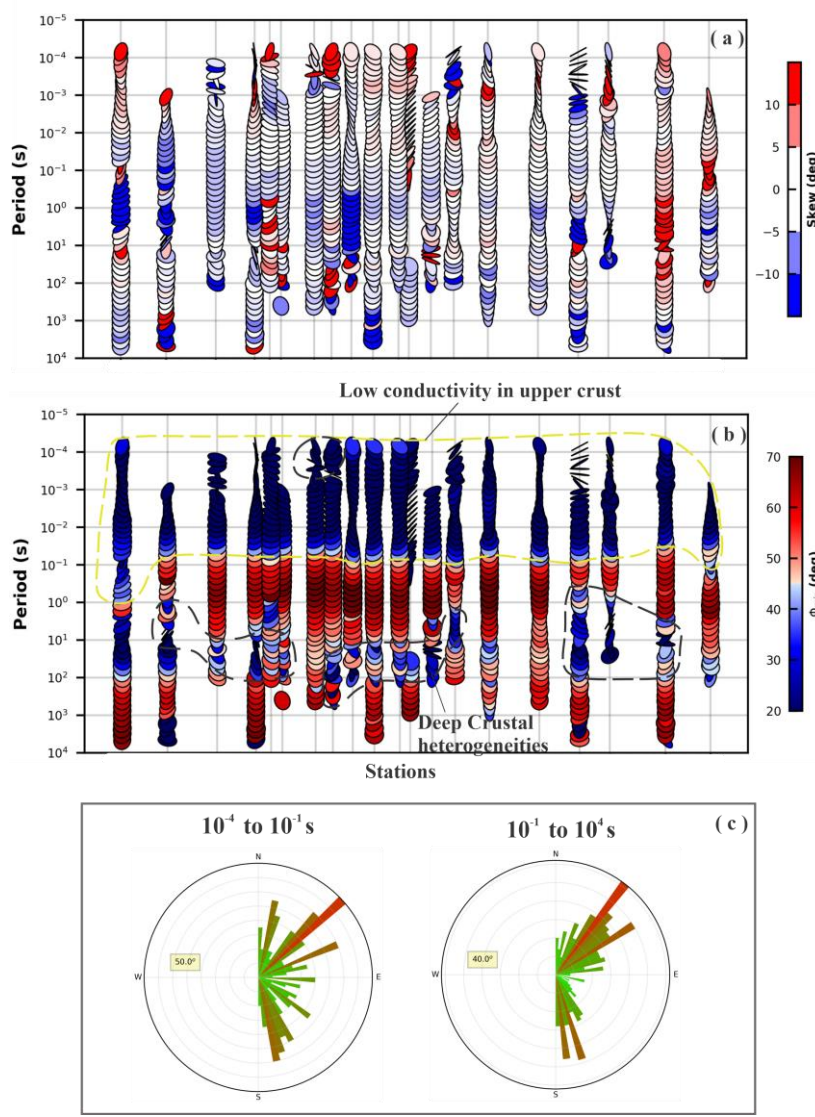


Fig. 3.5. Dimensionality Analysis. A) skew (β). B) Φ_{min} . C) Distribution of ellipses azimuth angles. Most of skew angles are out of the $|\beta| \leq 5$ range, thereby indicating that 3D inversion is required. Likewise, abrupt changes in Φ_{min} and in the direction of the ellipses also corroborate it. On the other hand, there is a stable regional geostrike in the NE direction.

Three - Dimensional Modelling of MT

We have applied the non-linear conjugate gradient approach for inversions using the ModEM code of Kelbert et al. (2014). This method minimizes the penalty function described as:

$$\Psi(m) = (d - F(m))^T C_d^{-1} (d - F(m)) + \lambda (m - m_0)^T C_m^{-1} (m - m_0)$$

where m is the model resistivity, m_0 is a starting model, C_m is the model covariance, d is data with N dimensions, $F[m]$ is the forward response of the model, C_d

is the data covariance, and λ is the damping factor, which is a trade-off between data misfit and model roughness.

We assessed the residuals deviation using the normalized root mean square (nRMS), which is given by the expression:

$$\text{nRMS} = \sqrt{\sum_{i=1}^N \frac{1}{N} \frac{d_i - F(m)_i^2}{e_i^2}}$$

Where N is the number of data points, d_i is the observed data at the i point, $F(m)_i$ and e_i are the respective forward response and data error.

We designed the mesh according to the recommendations of Miensoput (2017, and references therein) and Robertson et al. (2020). For the area of the profile, we have a chosen horizontal cell size of 2x2 km, which is around 1/6 of the mean space between stations. Fifteen padding cells were added with 1.5 increments to maintain 3-5 skin depths of distance from the stations. Vertical cell size started at 10 meters and was increased by a factor of 1.2, totaling 57 cells. As a result, cell sizes are less than 1 km from the surface to 5.7 km, less than 5 km from 5.7 to 30 km, and less than 10 km from 30 to 60 km.

We have rotated data to geographic north and removed outliers. We inverted the full impedance tensor in the period range of 0.0001 to 4672 s. We have started inversion with error floors as a percentage of $\sqrt{|Z_{xy} \cdot Z_{yx}|}$. 10% for the off diagonals, and 20% for the diagonals. Initially, we opt to downweight diagonal elements due to their typically higher noise level. The damping factor parameter was initiated at 10, and it was divided by a factor of 10 if the difference between the current and previous nRMS model was less than 0.002 %, until it reaches 10^{-4} .

The inversion approach adopted in this paper consists of testing several parameters and evaluating their influences on the resulting models. Furthermore, because

we have a reasonable geological control on the surface in addition to airborne magnetic data, we expect that the electrical structure should corroborate some of this known information.

We have started the inversion with a uniform half-space of 7141 ohm.m, which is the average resistivity of the dataset. The inversion ended after 79 iterations and a nRMS of 1.92. We tested different weights for the off-diagonal/diagonal impedance tensors such as 10/10, 5/15 %. They yielded a nRMS of 2.4 with 90 iterations, and 3.32 with 96 iterations, respectively. We have not noticed any relevant difference that should change the interpretation, and there was also an increase in nRMS as we decrease the error floor percentage of the off-diagonal tensors. Therefore, we kept the initial weights.

Then, we have run additional inversions with uniform half-spaces of 1000, 100, and 50 ohm.m. For the Alta Floresta domain (also considered part of Tapajós Province), Fernandes et al. (2018) tested prior models of 50, 100, 200, 500 and 800 ohm. They showed that the less resistive models present more structural variability. All of our models recovered a highly resistive zone in the first ten kilometers of the crust, mid-crust conductors and a gradual decrease in the resistivity with depth (supplementary material). However, prior models 1000 and 7141 ohm.m could not recover known mafic conducts mapped both on the surface as well as in the magnetic data. Therefore, we discarded them and conclude that the half-spaces of 100 (nRMS = 1.94) and 50 ohm.m (nRMS = 1.98) are more reliable because they corroborate independent information and both could be used for interpretation. We opt for the 50 ohm.m uniform half space since it yielded a higher contrast between the upper and lower crust.

Our uncertainty analysis approach focuses on comparing nRMS between stations, impedance components, and nRMS per period decade (Fig. 3.6). In general, we note that nRMS does not change abruptly among stations. Except for two stations where nRMS is

greater than 4, we found that nRMS for off-diagonal components tends to be about 2. The nRMS is relatively small for diagonal components due to lower weights (i.e., higher error floors). In Figure 3.6b, we took the average of the nRMS per decade range. We observe that nRMS tends to decrease over longer periods. Higher misfits in the first-decade range are due to the small-scale resistivity distribution that the model cannot recover with 2 km horizontal cell size. There is an upward step between 0.01 and 0.1 s, and the nRMS reaches a minimum below 1.6 in the range 1-10s. Then it rises up to 1.8 because there are fewer stations with available transfer functions for averaging data.

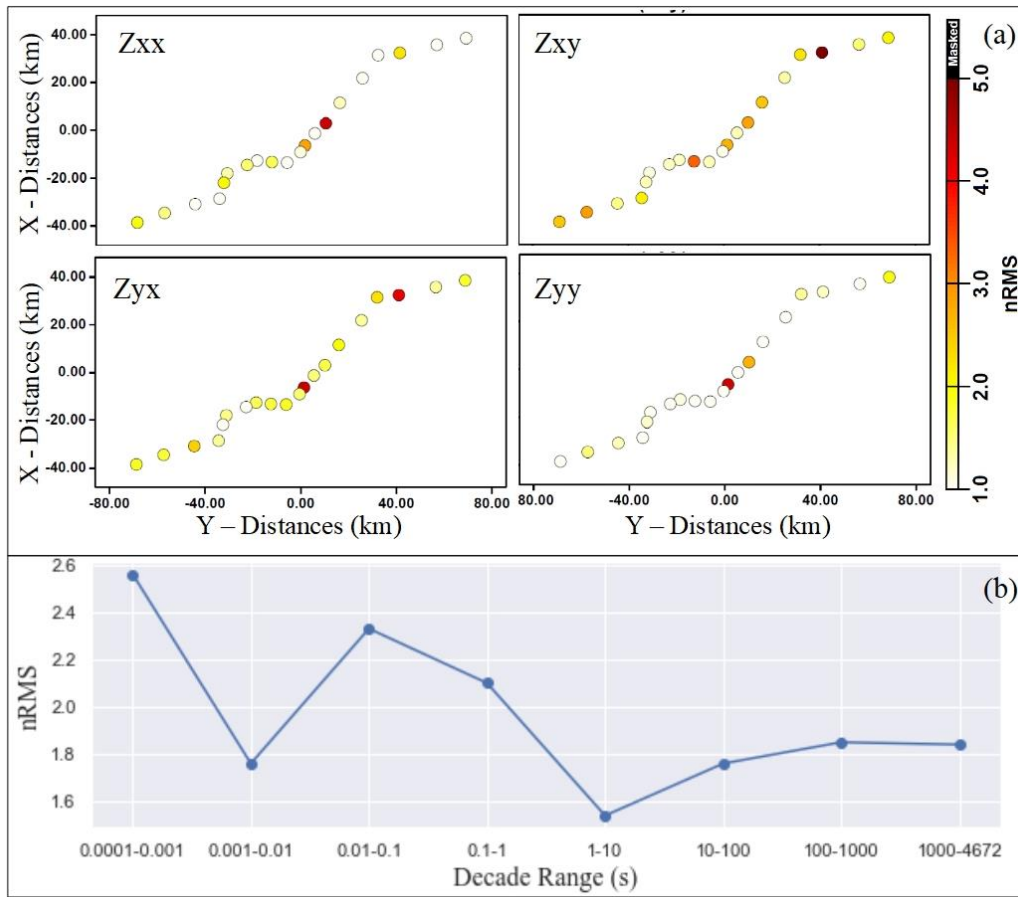


Fig. 3.6. a) Comparison of nRMS among stations and tensors impedance. B) Relation between nRMS of full impedance tensor and period. Overall, the nRMS is stable about 2 for almost for the off diagonals tensors, and it is about 1 for the diagonal tensors. As a function of period, the nRMS decreases until 10 s period, then it stabilizes about 1.8 for periods > 10 s.

In order to ensure the robustness of our model, we conducted a sensitivity analysis on individual conductors that are fundamental for interpretation. We followed the

procedure of Becken et al. (2008). We have replaced their resistivity to 10000 ohm.m and restarted the inversions to check if the model would return the conductors (Fig. 3.7). Figure 3.7b is the result of the shallow conductor. The model yielded the conductor with relatively low changes for the first kilometers, however, it decreases resistivity below 40 kilometers. We note the same behavior on Figure 3.7c referred to as the deeper conductor. Although it does not change the horizontal position, this model tends to deepen the narrow conductors. Both Figures 3.7b and 3.7c returned the conductors with equivalent nRMS, and besides, we notice an increase in nRMS on stations above those conductors. Therefore, our model is considered robust since the main features were preserved.

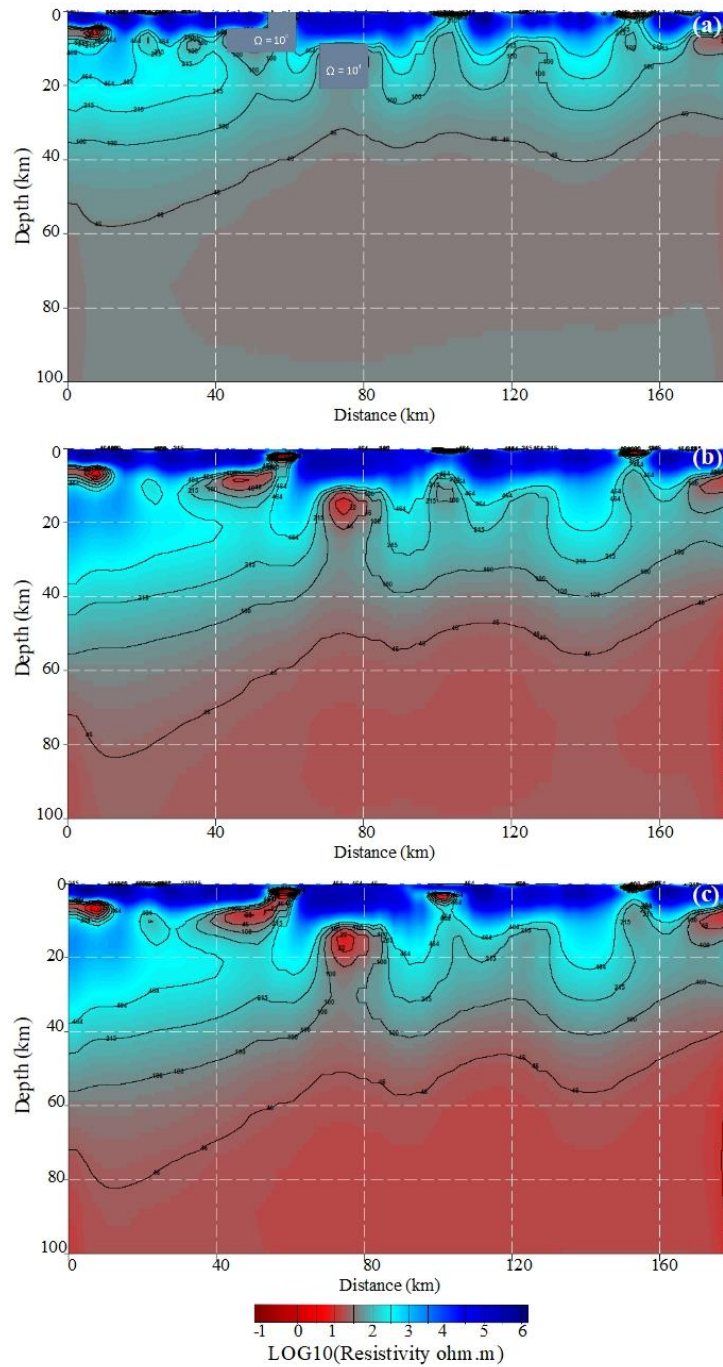


Fig. 3.7. Sensitivity Analysis. A) Preferred model. B) Result of changing the resistivity of the shallow conductor to 10000 ohm.m. C) Result of changing the resistivity of the deeper conductor to 10000 ohm.m. Although the resistivity of long periods slightly decreases, both models yielded the same conductors.

Considering a maximum period of 4672 s, if we change our average resistivity between 50 and 1000 ohm.m, skin depth equation suggests a depth resolution between 75 and 342 km. We illustrate our model until 100 km and suggest cautions for interpreting deep features (Fig. 3.8).

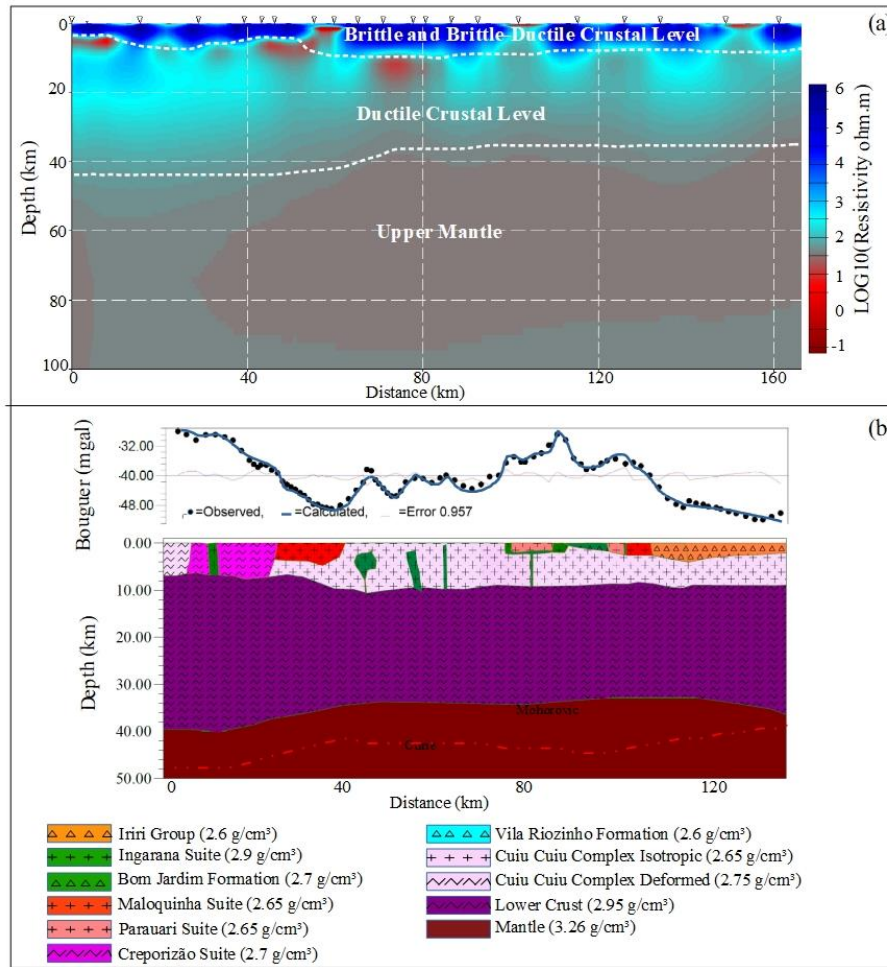


Fig. 3.8. a) Preferred model of electrical structure of lithosphere. B) Gravity 2.5D forward model. The electrical resistivity model presents three main layers: upper crust, lower crust and mantle. Their thickness is used as an input for the gravity model as well as rocks mapped. The density was chosen based on Dentith and Mudge (2014) petrophysical library.

Gravity data and 2.5 D forward modeling

The Geological Survey of Brazil - CPRM provided terrestrial gravity data, which covers almost the whole MT profile (Fig. 3.8b). The profile comprises 204 sites spaced in the range of 500-1000 m. We have applied the forward modeling described in Talwani et al. (1959), which is implemented in the GM-SYS® package of Oasis Montaj® 9.7 version. The ordinary 2D forward modeling allows changes in the profile and depth directions, whereas the layers in the perpendicular direction are presumed to extend to infinity. As a result, we propose a 2.5-D forward modeling that has an advantage over common 2D because it allows us to constrain the perpendicular extension of bodies.

In the forward gravity modeling, we incorporate as constraints the geological units mapped on the surface and consider the high resistivity zone (>10000 ohm.m) in the crust obtained from MT data as the upper crust and the range of resistivity 465-45 ohm.m as the mid-lower crust. Our approach is to set a reasonable upper crust density model and test what crustal thickness variations better fit the dataset (Fig. 3.8b). We have set densities according to the petrophysical library of Dentith and Mudge (2014).

Results and Discussion

The Crust

The electrical resistivity model reveals a high resistivity zone (> 10000 ohm.m) for the first few kilometers of the crust (Fig. 3.8a). Exceptions occur where major mafic bodies intrude the upper crust, lowering resistivity to the 1-10 ohm.m range. We interpret that this zone represents resistive granitoids intruded in the upper crust. We note that the thickness of this zone varies from 6 km in the western portion to almost 10 km in the eastern portion. This transition is marked by a low angle conductor dipping westward, and corresponds to the magnetic boundary shown in Figure 3.2. The electrical resistivity model shows that the crust in the western portion of TMP was thrust into the eastern block, which resulted in rapid tectonic denudation and exhumation of higher metamorphic facies of the Cuiú Cuiú Complex as noted by Klein et al. (2001), whereas in the eastern portion these rocks are preserved. In the mid-lower crust, resistivity decreases between 465 and 45 ohm.m (Fig. 3.8a). Evidence from xenoliths of volcanic arcs present chiefly mafic composition (Condie et al., 2022), which may explain these values. The absence of a lower crust conductor (< 10 ohm.m) reinforces that marine sediments (carbon source) were not buried to mid to lower crust depths, as a result of a collisional setting (Heinson

et al., 2021). Therefore, our findings support the hypothesis of a long lived subduction that did not evolve to collision in the Paleoproterozoic times.

Comparing to other Paleoproterozoic accretionary orogens, Cheravatova et al. (2015) show that the upper crust of the Svecoffnian orogen is also resistive, whereas the mid-lower crust tends to be less resistive. They attributed this decrease in resistivity to graphite or carbon in grain-boundaries, sulfide or oxides. Jones et al. (2005) show a similar pattern to the Trans-Hudson Orogen, whereas in the Archean domains the crust tends to be entirely resistive. Indeed, James et al. (2003) interpreted an intermediate to felsic lower crust in the Archean Kaapvaal Craton based on low Vp/Vs ratios.

The electrical model suggests the crust is thinner toward the Irixi Xingu Domain. However, the crustal thickness difference would be almost 10 km, which is unlikely since it would imply paired high and low anomalies of long wavelength in the Bouguer anomaly, which are absent (Fig. 3.2b). We addressed the crustal thickness of TMP in the 2.5D gravity forward modeling.

The gravity 2.5D forward modeling yielded a maximum crustal thickness variation of ~ 5 km. The crust tends to be thinner toward the Irixi Xingu Domain. It was expected since this domain is overprinted by a silicic large igneous province developed in 1.88 Ga, which is a process of intense crustal stretching (Klein et al., 2012), and it is in agreement with the resistivity model.

We have undertaken a sensitivity analysis in order to address limitations in our density model. Firstly, we have made changes in crustal thickness to address how amplitude and long wavelengths would be affected. We noticed that a 1 km decrease in crustal thickness yielded a positive change of 4 mGal on the Bouguer anomaly. Although Bouguer anomaly does not show high amplitudes for the area, Albuquerque et al. (2017) used the receiver function technique and derived a crustal thickness model with a 4 km

variation for the TMP and Irixi Xingú Domain. Additionally, Costa et al. (2020) show a decrease in P-wave velocity towards Irixi Xingú, which is in agreement with a thinner crust.

The Metasomatized Mantle

The upper mantle presents resistivities in the range of 45-10 ohm.m. This is uncommon since more typically the upper mantle, usually composed of peridotite in cratonic areas, is resistive (Chave and Jones, 2012; Curtis and Thiel, 2019;). Özaydın and Selway (2020) investigated how conductivity changed under various mantle composition and temperature scenarios. These authors calculated the resistivity for typical rocks of the mantle: dunites, peridotites, and piroxenites. The resistivities are higher than 10^3 ohm.m, and increase proportionally with the content of olivine. On the other hand, models simulating scenarios of hydration of anhydrous minerals or addition of graphite in the mantle show resistivities of less than 10 ohm.m. Likewise, Pommier and Le Trong (2011) conclude that serpentinization processes of about 75 km under 720 °C may decrease the mantle resistivity to 10 ohm.m. This occurs due to the oxidation of fayalite to magnetite, releasing SiO₂, which is then consumed by forsterite to form serpentine (Bucher and Grapes, 2011). Indeed, Correa et al. (2022) calculated the depth to the Curie temperature, which is interpreted as the depth of demagnetization of magnetite. They found that magnetite is present in the mantle of TMP (Fig. 3.8b).

We also tried to quantify the degree of hydration in the mantle. We decrease the density from 3.26 (10 %) to 3.1 g/cm³, which is related to a 20% degree of serpentinization (Dentith and Mudge, 2014). It yielded a maximum crustal thickness of 45 km for the west side of the profile and 32 km for the east Irixi Xingu Domain, which is unreasonable considering the prior densities set for the upper and lower crust. This difference also increases with higher degrees of serpentinization. In the TMP, the long-lived magmatic

arc hypothesis comprises several stages of mantle wedge hydration, such as the Creporizao Suite (1.98-1.95 Ga) formed by hydrous and oxidized magnetite and amphibole bearing melts, as well as the high K metaluminous to mildly peraluminous Parauari Suite (Cassini et al., 2020). According to Holwell et al. 2019, when the slab breaks off and the asthenosphere upwells, the metasomatized mantle undergoes localized partial melting, resulting in magmas with compositions ranging from high-K alkaline to silica-saturated to silica-undersaturated alkaline. These alkaline magmas are recorded in the rocks of the Maloquinha Suite. Therefore, the low resistivity values are likely to be the result of a metasomatized mantle that went locally through partial melting, and the density model shows that the post subduction magmatism reduced the degree of serpentinization to less than 10 %.

Metallogenic implications for gold deposits and correlates

The discovery rate of gold has decreased in the last decade, whereas the cost per relevant discovery has increased (Groves and Santosh, 2015). These authors argue that using a crustal to lithospheric-scale instead of the conventional deposit-scale approach may solve this issue. For instance, although orogenic gold, intrusion-related gold, Carlin-type gold, and iron-oxide-copper-gold deposits present quite different parameters at the deposit-scale, on the other hand, in terms of crustal to lithospheric-scale, these deposits show similar source-region parameters, such as the proximity to craton margins or lithosphere boundaries, though each one is formed in a specific time of the supercontinent cycle (Groves and Santosh, 2015; Huston et al., 2016,). Therefore, understanding the crustal architecture adds valuable information about these mineral systems.

In the TMP, the timing of magmatism correlates with Pb-Pb sulfide ages and Ar-Ar data for ore-related hydrothermal systems (Juliani et al., 2021 and references therein). Therefore, most of the gold deposits (and polymetallic) are classified as magmatic-

hydrothermal. Furthermore, few deposits of orogenic gold and paleoplacer are described (Veloso et al., 2013; Klein et al., 2016).

Magmatic-hydrothermal systems vary from porphyry-Au-Cu to low, intermediate, and high-sulfidation (Juliani et al., 2005; Aguja-Bocanegra, 2013; Sutherland et al., 2019; Guimarães et al., 2021). Kirkby et al. (2022) show a robust correlation between porphyry copper deposits and upper mantle conductors world-wide, which reflects the deep magmatic plumbing systems beneath these deposits. Therefore, the electrical resistivity model also validates the fertility of the mantle of the TMP for large amounts of copper.

We interpret in detail our resistivity model for the first 15 km so as to give constraints for gold mineralization controls in the study area. Figure 3.9 displays our understanding of the main geological discontinuities. According to Santos and Coutinho (2008), crustal discontinuities formed in a stress field with σ_1 aligned roughly in a E-W trending direction. We interpret sinistral NW trending structures, thrust faults, normal faults, and opening structures using this predicted cinematic. We explain the low angle C1 conductor as a fault that juxtaposes two different crustal levels.

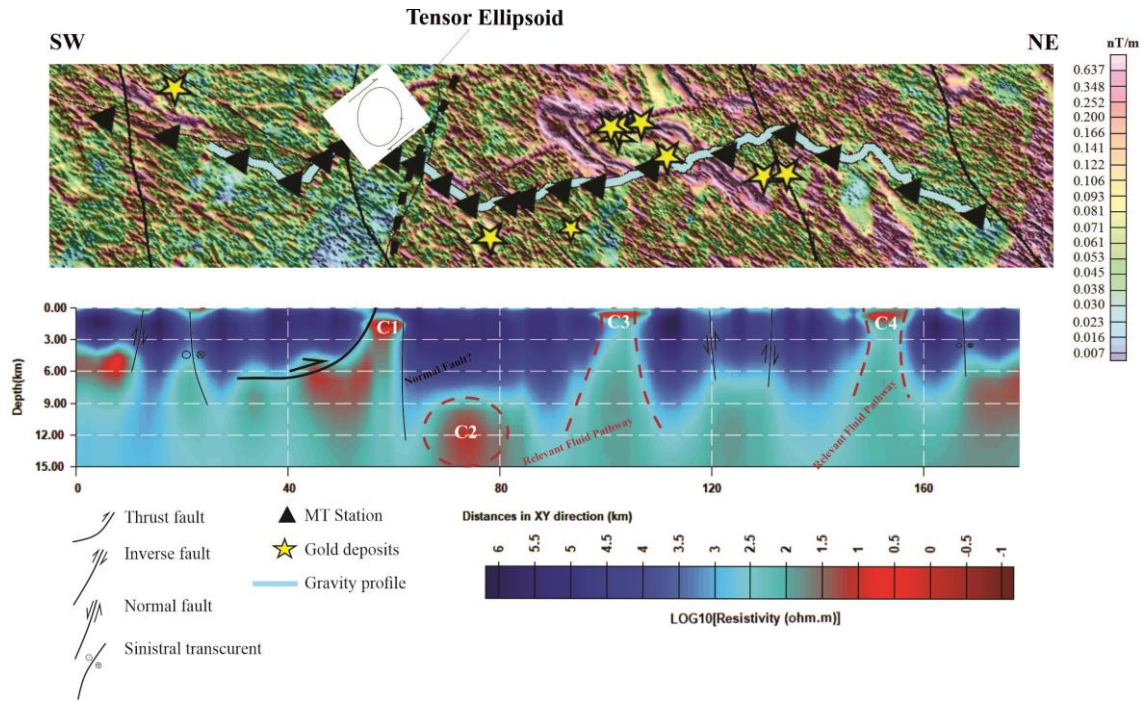


Fig. 3.9. Electrical structure of the upper crust with correlated magnetic data (Fig. 3.2a) on the surface. The structures interpreted are based on the predicted cinematic of Coutinho (2008). C1 thrust fault is the representation of the magnetic boundary in depth. Note the concentration of gold deposits at C3, and the potential for more discoveries at C4.

The vertical conductors are relevant fluid pathways as we see a high correlation with the most density of gold deposits in C3, and a high potential for more occurrences in C4. However, these conducts do not extend to the bottom of the crust. Huston et al. (2016) attest that the magmas responsible for magmatic-hydrothermal deposits ascent into the upper crust driven by buoyancy. Thus, we may infer that the ascent of the magma from the upper mantle to the upper crust has not altered the lower crust. The lack of crustal-scale shear zones implies that the deposits are controlled by the intrusions and structures developed in the upper crust.

Mid-crust conductors (such as C2) are usually explained by melt, aqueous fluids (high chlorinity), or alteration zones rich in carbon, sulfides or iron oxides (Jones, 1992, Cherevatova et al., 2015, Yin et al., 2021). We discarded the two initial hypotheses once the study area had been stable and cold since the Paleoproterozoic. The geochemical signature of hydrous and oxidized between 1.95 and 1.89 Ga may be the evidence to

explain this conductor (Cassini et al., 2022). The Fe plays an essential role as a charge carrier (increasing conductivity) so that oxygen fugacity constrained in TMP magmas collaborate to associate this mid-crust conductor as a significant footprint of the magmatic-hydrothermal systems. Fernandes et al. (2018) have also found mid-crust conductors, and they interpreted them as high concentrations of iron oxides and sulfides due to correlations with the position of gold mineral deposits. Based on geochemistry and negative $\epsilon\text{Nd}_{(1.95)}$ signature, Cassini et al. (2020) argue that the Creporizão (1.98-1.95 Ga) and Parauari Suites (1.89 Ga) are the main fertile rocks to host mineral deposits. In Figure 3.10, we summarize the main calc-alkaline magmatic hydrothermal systems of TMP regarding the crustal architecture inferred from the electrical resistivity model.

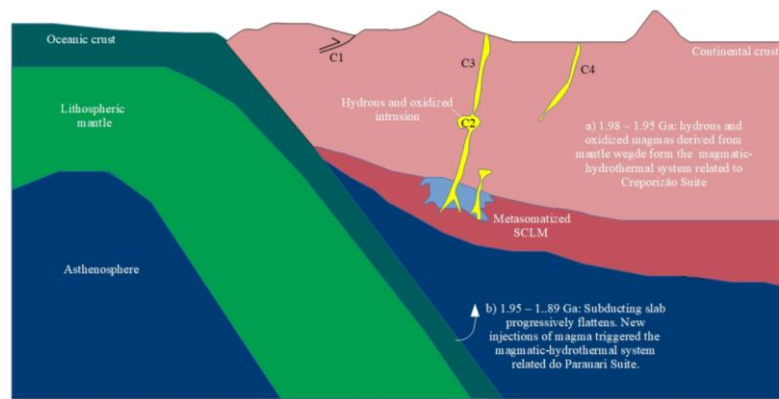


Fig. 3.10. Conceptual model for the calc-alkaline magmatic hydrothermal systems and crustal architecture inferred from electrical resistivity model (Figs. 8 and 9). SCLM: Sub continental lithospheric mantle. Adapted from Cassini et al. (2020) and Juliani et al. (2021).

The conductivity of these fertile magmas tends to increase due to the presence of magnetite and amphibole. Between 1.98-1.95 Ga, hydrous and oxidized magmas derived from the mantle wedge form the magmatic-hydrothermal system related to Creporizão Suite. Subducting slab progressively flattens metasomatizing the mantle wedge between 1.95 – 1.89 Ga, which eventually melts to form the magmatic-hydrothermal system related to Parauari Suite (Cassini et al., 2020). Notwithstanding, alkalic magmatic-hydrothermal systems form in minor orders in 1.88-1.87 Ga as a result of slab breaks off

and orogenic collapse, forming extensions structures and remobilizing fluids (Holwell et al., 2019).

Conclusions

Our work presents the first electrical resistivity model for depths up to 100 km for the Tapajós Mineral Province. We show that data processing using only nighttime data for periods > 100 s is necessary to provide high-quality data.

As the electrical resistivity model does not display a suture signature (crustal-scale conductor ou lower crust conductor), our model corroborates the long-live arc geological evolution hypothesis. Therefore, suture evidence is most likely to be westwards of the study area (Scandolara et al., 2017).

The model reveals a low resistivity upper mantle for the Tapajós Mineral Province, which is attributed to mantle wedge metasomatization and hydration, which endured around 130 Ma. Although MT is not a standard method to identify Mohorovic discontinuity, our model suggests a thinning towards Irixi Xingu domain. This behavior is also revealed in both gravity forward modeling and seismological data. A Robust correlation world-wide between mantle conductors and porphyry copper deposits reinforces the potential of TMP.

We interpret the magnetic boundary as a thrust fault juxtaposing two different crustal levels. A deposit scale comparison between the occurrences on each side should be considered to evaluate the potential for higher tons. We also identified fluid pathways on the MT model with a noticeable correlation with gold mineral deposits. There is a deep conductor around 12 km in the crust, which is the evidence of past bulky melt volume. The MT has provided relevant constraints to misunderstood structures presented on

surface and magnetic data. It has also proved valuable for detecting fluid pathways relevant for unraveling new mineral deposits.

Acknowledgments and Data Availability

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CAPÍTULO 4

EVIDÊNCIAS DE COLISÃO NO PALEOPROTEROZOICO

A multi-disciplinary geophysical approach reveals a Paleoproterozoic Cryptic Suture in the Amazon Craton – Implications for Proterozoic Supercontinent Reconstructions

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Abstract

The Amazon Craton evolved in the Paleo-Mesoproterozoic through successive continental magmatic arcs, which ages decrease westwards. However, when and if these magmatic arcs collided with another continent, closing an ocean basin is a subject of ongoing debate. It is speculated that the Baltica and Amazon cratons formed a purely long-lived accretionary margin for at least 900 Ma and only collided during the formation of the Rodinia Supercontinent at 1.2 – 1.0 Ga. Identifying relicts of suture zones is challenging since Proterozoic orogens tend to preserve the continental arc and the reworked components so that less than 10% of the orogens are composed of oceanic evidence. Moreover, the Phanerozoic cover overlays more than 50% of the proterozoic rocks in the Amazon Craton. Therefore, we propose a robust geophysical assessment to investigate the upper and lower crusts. We acquired new magnetotelluric data and integrated it to the available potential field (satellite, airborne, and ground), geochronological data (magmatic and metamorphic ages), crustal thickness and depth to Curie temperature estimates. A major crustal-scale structure (more than 1000 km in extension) separates two tectonic fragments with remarkable differences in the magnetic gradient and structure

trend. We also find along this major boundary high and low Bouguer anomalies of long-wavelength paired, a step in the Mohorovic discontinuity, and a lower crust conductor. This evidence supports the hypothesis of a collision in Paleoproterozoic times. Our findings suggest that the orogen did not remain purely accretionary and evolved to a collisional type between 1.63 and 1.69 Ga. This contrasts with the purely accretionary orogens described in the Baltica Craton, thereby undermining the hypothesis that the Amazon and Baltica cratons shared the same peri-Nuna subduction system from 2.0 to 1.3 Ga.

4.1 Introduction

Opposing tectonic settings have been postulated for the Paleoproterozoic southwestern portion of the Amazon Craton. For instance, crustal thickening models, such as accretionary continental magmatic arc, or crustal thinning models, such as intraplate rifting, have been proposed (Bettencourt et al., 2010; Scandolara et al., 2017; Rizzotto et al., 2019). Motta et al. 2022 argue that successive continental arcs were superimposed on pre-existing continental crust. However, when, where, and if these magmatic arcs collided with another continent, closing an ocean basin is a subject of ongoing debate.

Condie (2013) argues that the Amazon Craton (as well as Baltica, Makkovikian-Labradorian, and Penokean-Yavapai-Mazatzal) is part of the Great Proterozoic Orogen in the Nuna supercontinent, which remained purely accretionary for at least 900 Ma and only terminated in collision in the formation of the Rodinia Supercontinent at 1.2 – 1.0 Ga (Johansson et al., 2009; Johansson et al., 2022). On the other hand, paleomagnetic data suggest that the Amazon Craton evolved apart from the Nuna Supercontinent (Pisarevsky et al., 2014). Therefore, finding an ancient suture zone has implications for the Proterozoic supercontinent's reconstructions.

Identifying relicts of suture zones is challenging, especially in the southwestern portion of the craton, where the crust results from polycyclic tectonothermal events (Scandolara et al., 2017). Proterozoic orogens tend to preserve the continental arc and the reworked components (mean 92 %), so less than 10% of the orogens are composed of oceanic crusts, islands and plateaus, serpentinites,

and pelagic sediments (Condie, 2013). Furthermore, the Phanerozoic sedimentary cover comprises more than 50 % of the study area (Fig. 4.1). Therefore, we propose a robust geophysical assessment to investigate the upper and lower crusts. We acquired new magnetotelluric data and integrated it with available potential field (satellite, airborne, and ground), geochronological data (magmatic and metamorphic ages), crustal thickness, and depth to Curie temperature estimates.

The crustal framework resembles fault-bounded tectonostratigraphic terranes. A major crustal boundary (more than 1000 km of extension) is investigated in detail. Electrical resistivity and density models provide insights into the crust composition, and the remains of a Paleoproterozoic cryptic suture are found in the lower crust. The collisional scenario is tested using geochronological data, and the following tectonothermal events are reviewed to explain how the crust signature was overprinted.

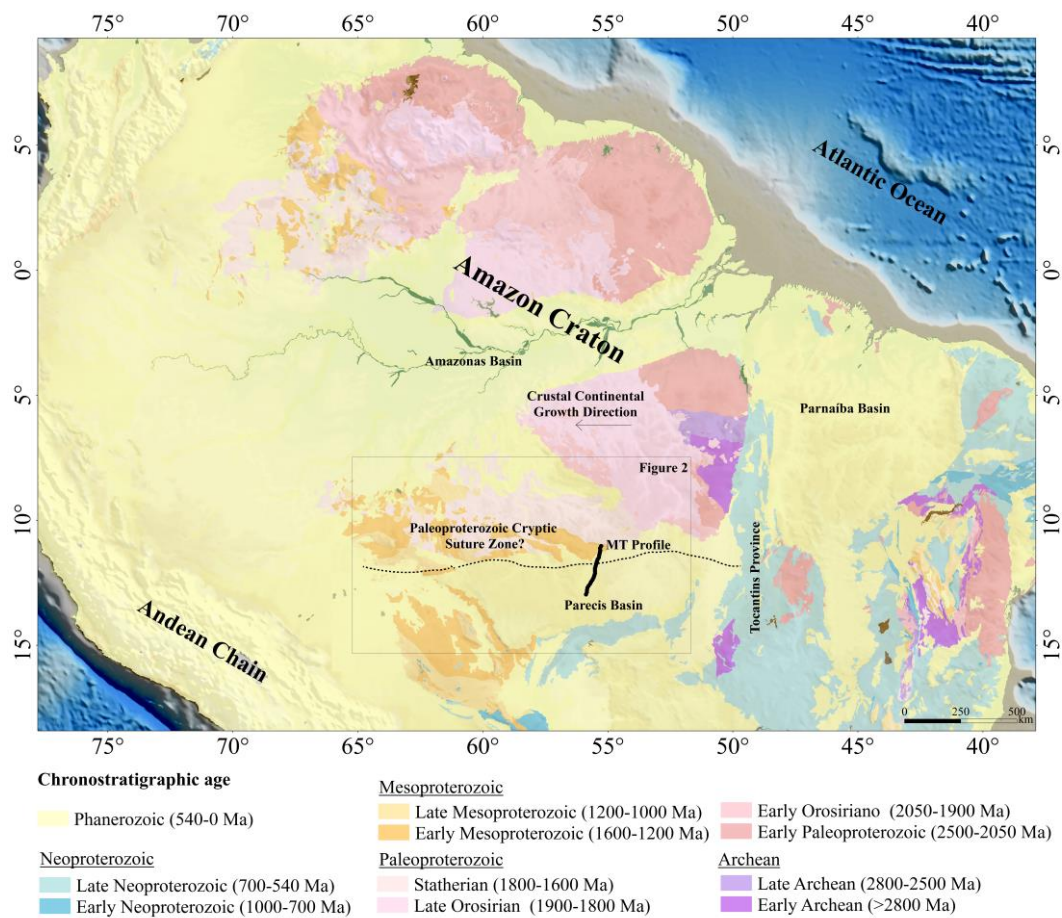


Fig. 4.1 - Chronostratigraphic map of the Amazon Craton in the context of the South American Platform. The Amazon Craton is bounded by Neoproterozoic belts (Tocantins Province) at the south and east and by the Andean Chain at the west. The rectangle comprises the study area where a crustal-scale structure lies underneath the sedimentary cover. Adapted from Cordani et al. (2016).

4.2 Geological Context

4.2.1 Regional Geology

The Amazon Craton is an Archean-Proterozoic block bounded by Neoproterozoic belts at the south and east and the Andean belt at the west (Fig. 4.1). Initially, the Amazon Craton was described as a platform of reactivation processes with ensialic orogenies and, in minor order, sea-floor spreading and subduction processes (Hasui et al., 1984; Costa and Hasui, 1997). Based mainly on Rb-Sr data, Tassinari and Macambira (1999) suggested that the Archean protocraton was formerly made of independent microcontinents (Central Amazonian, > 2.3 Ga), which were amalgamated mainly in the Rhyacian (Maroni-Itacaiúcas, age 2.20 – 1.95 Ga). Since the Orosirian successive magmatic arcs evolved with continental growth towards the west: Ventuari-Tapajós (1.95-1.80 Ga), Rio Negro-Juruena (1.8-1.55 Ga), Rondonian-San Ignacio (1.55-1.30) and Sunsás (1.3 – 1.0). Based on U-Pb and Sm-Nd isotopic data, Santos et al. (2000) constrained the Archean area (Carajás Imataca, 3.10 – 2.53 Ga) and divided the provinces in terms of provinces dominated by i) juvenile crust addition (Transamazonian, 2.25 – 2.00 Ga; Tapajós – Parima, 2.10 – 1.87; Rondônia-Juruena – 1.76 – 1.47 Ga), ii) magmatic-tectonic recycling of previous orogenic belts (Rio Negro, 1.86 – 1.52 and Sunsás, 1.33 – 0.99 Ga) and iii) underplating (Amazon Central, 1.87 – 1.50 Ga). Most of the following models are derived from these two contributions, along with local geological mapping and additional isotopic constraints (Vasquez et al., 2008, Fraga et al., 2017).

4.2.2 Local geology

The focus of this study is the area composed mainly of rocks younger than 1.83 Ga in the southwestern portion of the Amazon Craton. Most of the complexity found is due to the superposition of three tectonometamorphic events dated in 1.69-1.63 Ga (Rio-Negro Juruena Orogeny), 1.42-1.37 Ga (Rondonian – San Ignacio Orogeny) and 1.2-1.10 Ga (Sunsás Orogeny). Therefore, Scandolara et al. (2017) proposed a division into three main tectonostratigraphic terranes: Juruena, Jamari, and Jauru

(Fig. 4.2). On the northeastern portion, the Tapajós Province (1.86 > Ga) is considered the basement and outcrops only in the Juruena terrane. It consists of gneisses, volcanoclastic, metamorphosed and highly deformed plutonic-volcanic rocks and granitoids deformed locally. The geochemistry signature varies from moderately to strongly evolved, oxidized, calc-alkalic to alkali-calcic I-type magmas formed within a compressional environment (Scandolara, 2006, Trevisan, 2021). On the western portion, the Paragua Craton is an allochthonous block composed of Paleoproterozoic gneiss and granulitic complexes amalgamated to the proto-Amazon Craton in Paleo-Mesoproterozoic times (Bettencourt et al., 2010, Scandolara et al., 2017).

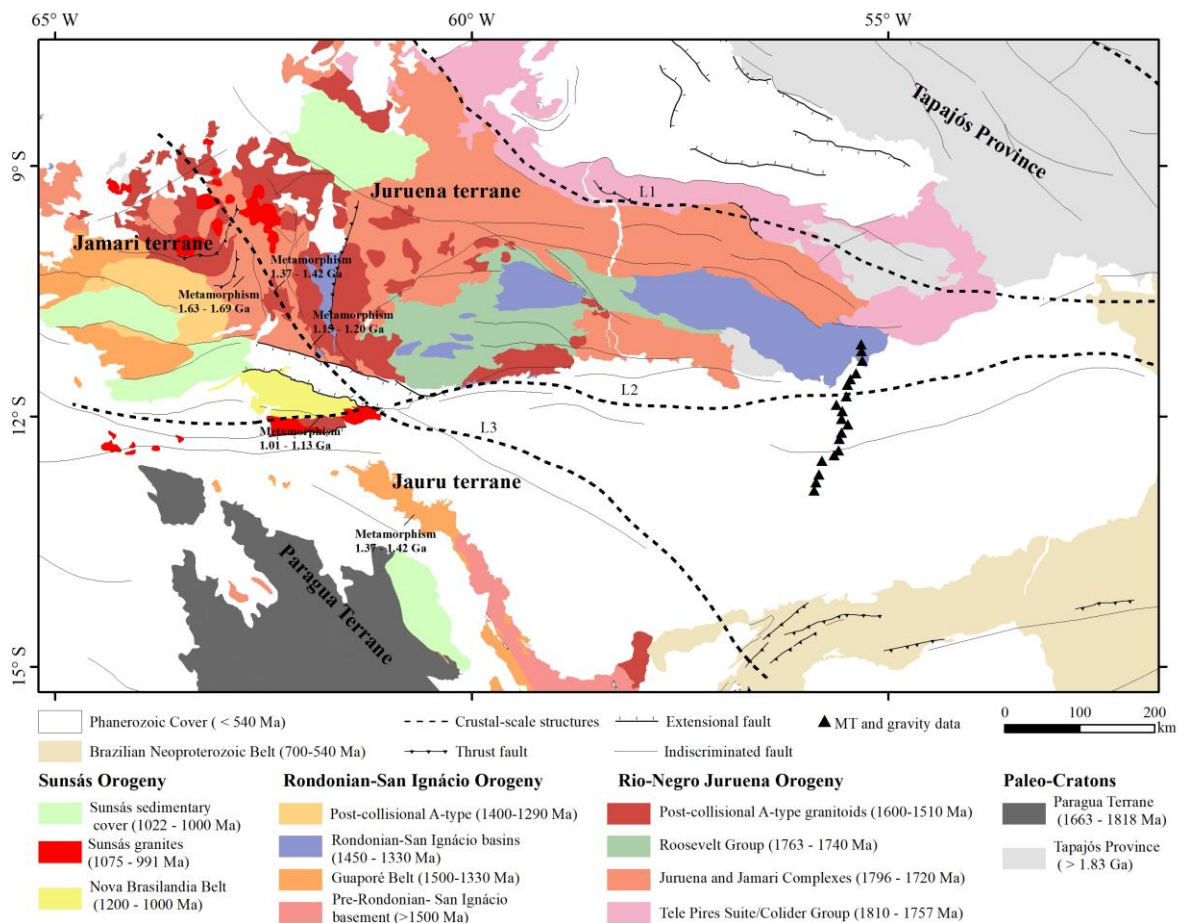


Fig. 4.2 - Geological map of the study area. Poliphasic deformation is characterized by three tectonothermal events: 1.69-1.63 Ga (Rio-Negro Juruena Orogeny), 1.42-1.35 Ga (Rondonian – San Ignacio Orogeny) and 1.2-1.10 Ga (Sunsás Orogeny). Note that Phanerozoic sedimentary rocks cover more than 50% of the study area. Crustal-scale structures (L1, L2, and L3) are interpreted from potential field data. Adapted from Cordani et al. (2016) and Scandolara et al. (2017).

4.2.2.1 Rio Negro Juruena Orogeny

The older rocks of the Rio Negro Orogeny are the Colider Group and the Teles Pires Suite, which are associated with time, space, and genetics (Rizzotto et al., 2019). The first is composed of plutonic-volcanic rocks dated between 1.81 Ga and 1.76 Ga with calc-alkalic to alkali signature and I to A-type magmas. The second consists of plutonic rocks and acid subvolcanic intrusion dated between 1.79 and 1.75 Ga with calc-alkalic to alkali-calcic and A-type magmas. Their occurrence is limited to the Juruena Terrane. Extensional tectonic settings have been proposed to explain these rocks, such as back arc, intraplate, or post-collisional (Scandolaro et al., 2017; Rizzotto et al., 2019; Trevisan et al., 2021).

Scandolaro et al. (2017) suggested that the Jamari and Juruena complexes formed contemporaneous continental magmatic arcs due to the similarities between time and geochemical signatures. According to Scandolaro et al. (2013a), the Jamari complex comprises tonalitic/enderbitic, granodioritic/charnoenderbitic, and mafic orthogneisses dated between 1.76 and 1.74 Ga. Regarding geochemistry, the orthogneisses are predominantly intermediate to acid, high-K, and metaluminous to weakly peraluminous calc-alkaline. Likewise, the Juruena Complex (1.78 to 1.74 Ga) is formed of I-type metatonalites, metagranodiorites, metagranites with high-K calc-alkaline signature; S-type granites and gabbro/gabbro-norite with subordinate norite; and diorite with tholeiitic/calc-alkaline rocks signature (Duarte et al., 2012, Scandolaro et al., 2014). Based on U-Pb ages, Scandolaro et al. (2013a) argue that these rocks were first deformed in a tectonothermal collisional event between 1.69 and 1.63 Ga. On the other hand, Rizzotto et al. (2019) divide the tectonic evolution of the Jamari and Juruena complexes and argue that most of the rocks of Juruena complex present geochemistry signature of alkali-calcic, metaluminous to peraluminous developed in a continental rift setting. The Roosevelt Group comprises low-grade metavolcanic-sedimentary rocks that have characteristics of a subaqueous depositional environment. These rocks are folded into wide antiforms and synforms with EW to WSW axes, and the age of the volcanic rocks is in the range of 1.76 to 1.74 Ga. After a

quiescence period, A-type rapakivi granite and charnockite/mangerite rocks intruded the crust between 1.60 to 1.51 Ga (Scandolara et al., 2013b). Rocks in the time interval of the Juruena and Jamari Complex are also found in the Jauru Terrane, however, with remarkable differences. For instance, Pre Rondonian-San Ignácio rocks comprise quartzites, pelitic schist, metavolcanic rocks with ages of 1.76 to 1.72 Ga., and tonalitic to monzogranites orthogneiss peraluminous to metaluminous sub-alkaline and calc-alkaline developed in an arc island environment (Geraldes et al., 2000, Bettencourt et al., 2010).

4.2.2.2 Rondonian San Ignacio orogeny

The RSI orogeny, which took place during the time interval from ca. 1.50 to 1.30 Ga, is characterized by the accretion of early Mesoproterozoic arc terranes overprinted by high-grade metamorphism during the mid-Mesoproterozoic by the inferred collision of a microcontinent – the Paragua block – against the Rio Negro Juruena continental margin (Bettencourt et al., 2010; Scandolara et al., 2017). The collision of the Paragua block against Amazonia resulted in high-grade metamorphism of the Chiquitania gneisses and the Lomas Manechis granulites at about 1.33 Ga (Bettencourt et al., 2010; Santos et al., 2008), also coinciding with cessation of arc magmatism in the Pensamiento granitic complex (Matos et al., 2009).

The Rondonian San Ignacio event affected the terranes in many different ways. There are rocks ranging from peraluminous to metaluminous sub-alkaline, calc-alkaline to A-type (1.56-1.42 Ga) associated with continental magmatic arc processes in the Jauru Terrane, as well as metavolcano-sedimentary, mafic-ultramafic, and metaluminous to weakly peraluminous sub-alkaline, calc-alkaline granites related to island arc (1.51-1.38 Ga), which culminated in several collisions of microcontinents/island arcs to the Proto Amazon Craton (Bettencourt et al., 2010, Rizzotto et al., 2013). On the other hand, there is no evidence of those rocks of the accretionary stage in the Jamari and Juruena Terranes. High-grade metamorphic rocks with a peak of metamorphism between 1.42 to

1.37 Ga occur as well as late to post-collisional A-type granites in the Jamari Terrane (1.35-1.31 Ga) and only formation of sedimentary basin in the Juruena Terrane.

4.2.2.3 Sunsás Orogeny

The Sunsás event is dominated by the extent to which metavolcanic-sedimentary belts metamorphosed in greenschist facies to amphibolite and granulite. These belts outcrops in the Jamari and Jauru Terranes and are likely to continue beneath the Parecis Basin (Tohver et al., 2004). They comprise quartzitic, conglomeratic, and carbonate units with maximum depositional age of 1.24 Ga and deformed at 1.0 Ga. Additionally, cogenetic metagranites, metagabbro dykes, and metadiabase stocks are dated in 1.1 Ga. It also occurs in synkinematic to late kinematic granites dated in 1.0 Ga. At last, tin-bearing granites intruded the crust in 0.97 Ga and marked the final stages of cratonization in the area (Tohver et al., 2004, Quadros et al., 2021).

4.3 Material and Methods

4.3.1 Potential Field Methods

Potential field methods rely on the property that any conservative field is equal to the gradient of a scalar potential. Gravity and magnetic fields suit this characteristic, and the disturbances in those fields generated due to variations in the density and magnetism of the rocks can be measured (Blakely, 1995).

We used airborne magnetic data from the Magnetic Map of Brazil, which is a compilation of public domain datasets provided by the Geological Survey of Brazil and several other public institutions of Brazil (Correa, 2019; Correa and Pinto, 2020). In the Oasis Montaj® software version 9, we applied the upward continuation filter to 15 km to remove shallow sources and a vertical derivative filter to highlight the crustal-scale structures (Fig. 4.3a). We interpreted the main magnetic discontinuities that divided the area into domains with specific gradient backgrounds and internal structural patterns.

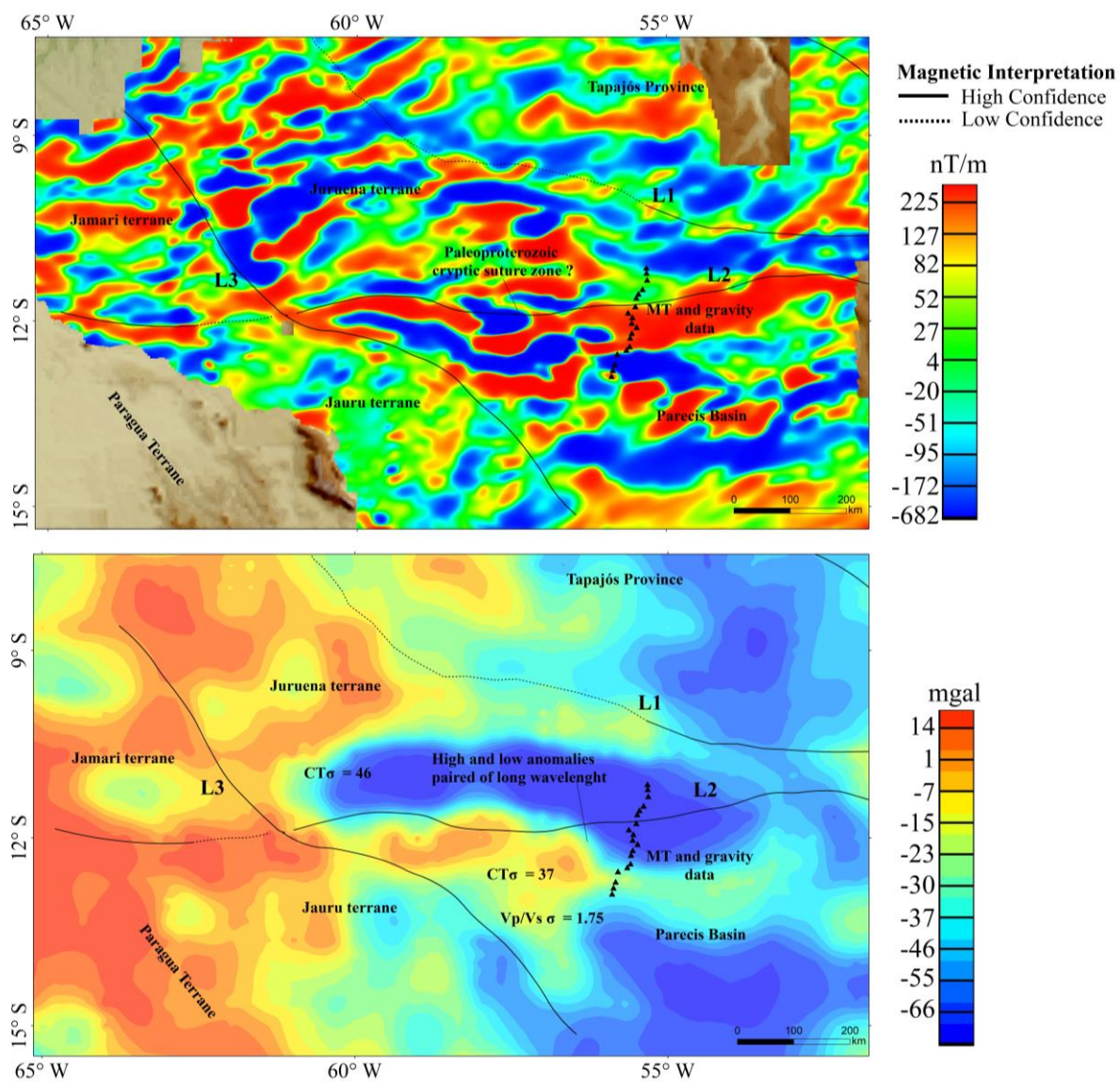


Fig. 4.3 - Potential field data. Upper panel: Vertical derivative of the magnetic anomaly (upward continued to 15 km) clipped from the Magnetic map of Brazil (Correa, 2019). Lower panel: Bouguer anomaly of the Goco06s gravity model (Kvas et al., 2021). Average crustal thickness (CT) and Vp/Vs ratio were obtained from Albuquerque et al. (2017).

We used the Bouguer anomaly of the Goco06s gravity model of Kvas et al. (2021) with a cell size of 0.25° (Fig. 4.3B). This model comprises over one billion measurements acquired over the last 15 years by 19 satellites (Fig. 4.3B). We opted not to interpret gravity discontinuities because long-wavelength Bouguer anomalies have a reasonable correlation with magnetic discontinuities, even though differences in resolution between datasets could lead to slightly different interpretations. More details on those differences will be given in the results section. Additionally, we conducted a forward modeling analysis using ground data from the Instituto Brasileiro de Geografia e Estatística (IBGE).

4.3.2 2D Gravity Forward Modeling

The forward modeling consists of setting an initial model based on known geology, then calculating the respective anomaly and comparing it with the observed anomaly. The model parameters are adjusted until the difference between the observed and calculated anomalies converges to zero. We used the GM-SYS package of Oasis Montaj®, which has implemented the Talwani et al. (1959) method. The software also minimizes edge effects by extending the layers from -30,000 km to 30,000 km in the profile direction.

A Priori geological information along the profile is scarce since well logs are restricted to the sedimentary basin or the basement top (Loureiro, 2016). On the other hand, Faria (2015) constrained the Conrad discontinuity with the radial power spectrum and the depth to the basement with 3D Euler solutions. Loureiro (2016) used seismic reflection data along some profiles to estimate the depth to the basement, while Albuquerque et al. (2017) derived a crustal thickness model from the receiver functions technique. Finally, Fontes et al. (2019) also constrained the sedimentary basin using broad-band MT data. Therefore, we used the results of these works to set a starting model with the parameters listed in Table 1.

Table 1 - Parameters of the starting model. Densities values of the Dentith and Mudge (2014) library.

Layers	Thickness (km)	Rocks	Density (g/cm³)
Sedimentary basin	1.0 – 3.0 km	Sedimentary/metasedimentary	2.42
Upper crust	13 – 11	Gneissic/Granitic	2.67
Lower crust	25 – 31	Mafic/Ultramafic	2.90
Mantle	-	Peridotitic	3.30

The calculated anomaly from the starting model yielded a Bouguer anomaly with almost 60 mGal amplitude (Fig. 4.4a). This is partly due to the 6 km increase in crustal thickness northwards.

However, the wavelengths forming a high and low were not captured, thereby requiring changes to match the observed and calculated curves.

Achieving a precise density model is unlikely in the absence of seismic reflection data. Infinite models can fit the data; however, only a few will have geological meaning. Therefore, we addressed this issue by testing some geological scenarios and comparing them to available data. In the first scenario, we fixed the sedimentary layer and the Conrad discontinuity at 14 km (Faria, 2015), and we moved the Mohorovic discontinuity to fit the data. It yielded a crustal variation of more than 30 km, thereby not in agreement with the receiver function model (Albuquerque et al., 2017). In the second scenario, we maintained the Conrad discontinuity at 14 km, fixed the Mohorovic discontinuity, and changed the thickness of the sedimentary basin. The model required a thicker sedimentary layer (~ 7 km) to explain the low gravity in the northern portion of the profile. Loureiro (2016) and Fontes et al. (2019) show that the sedimentary layer at the north boundary is considerably thinner. Finally, we fixed the sedimentary layer and the Mohorovic discontinuity and moved the Conrad discontinuity (Fig. 4.4b). It caused a variation of about 10 km, which is reasonable for a polydeformed terrane with several stages of crustal exhumation. This variation is also found on orogens in the North China Craton and Georgetown Inlier in Australia (Korsh et al., 2012; Vasanthi and Santosh, 2021). Therefore, we considered this last model reliable to give insights regarding the crustal composition and tectonics of the study area.

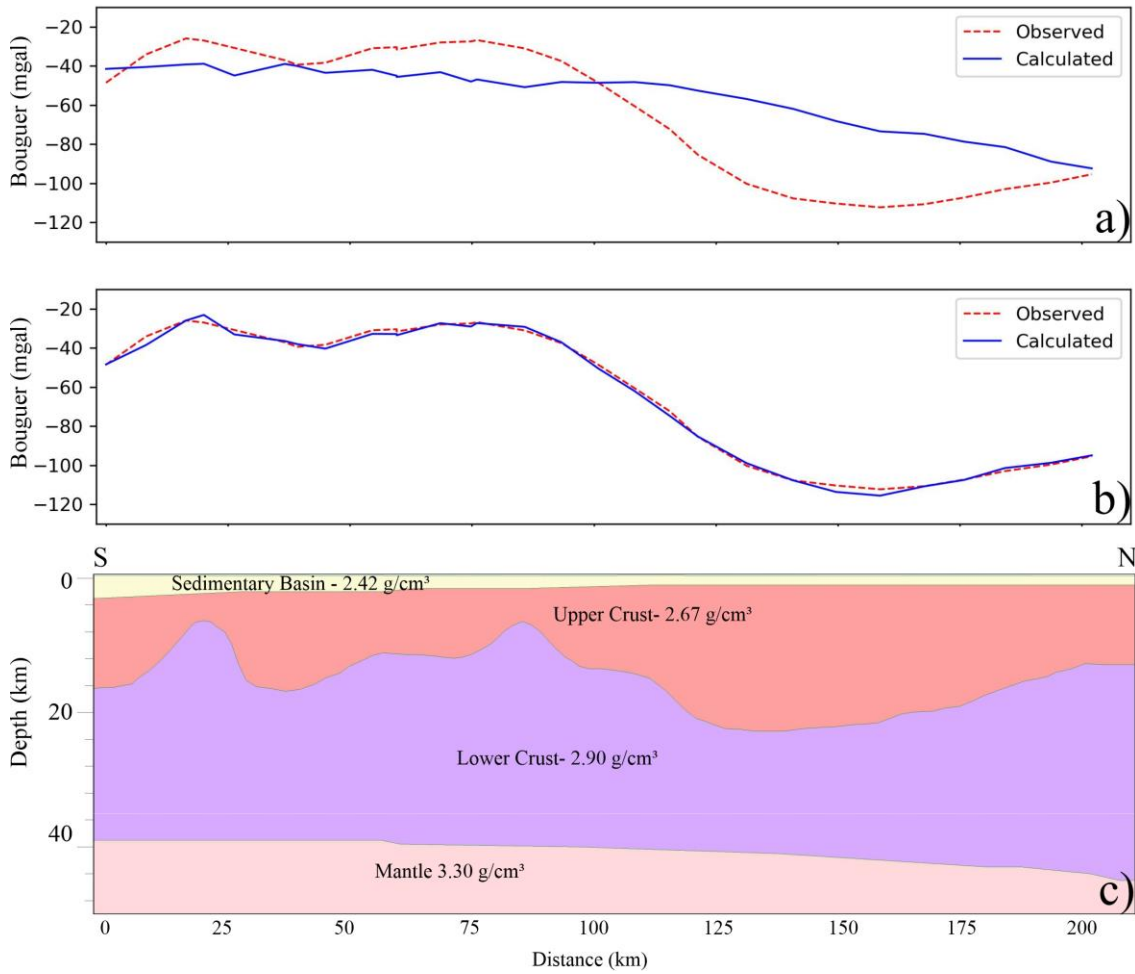


Fig. 4.4 - Gravity forward modeling. a) Comparison of observed and calculated curves obtained from the starting model (table 1); b) Comparison of observed and calculated curves from the preferred density model; c) Preferred density model shows that mafic composition is more relevant in the southern portion of the profile. Moreover, the ascending of the lower crust may also indicate structures of exhumation.

4.3.3 Magnetotelluric Method

Magnetotellurics is a passive method that measures in orthogonal directions the inducing magnetic fields (H) and resulting horizontal electric fields (E) as a function of time to calculate the Earth impedance tensor (Z) as a function of frequency (Chave and Jones, 2012). In the frequency domain, the field components are related by $E(\omega) = ZH(\omega)$, where $\omega = \frac{2\pi}{T}$ is the angular frequency and T is the period. The impedance tensor is a complex function, and it is used to determine the apparent resistivity ($\rho = \frac{1}{\mu_0 \omega} |Z|^2$), where Z is an impedance element and μ_0 is the magnetic permeability of

free space. The skin-depth δ (a proxy for depth of investigation in kilometers) is calculated by $\delta \approx 0.5\sqrt{T\rho}$.

4.3.3.1 Data acquisition and dimensionality analysis

We have acquired a 200 km transect with 21 broad-band sites at 10 km spacing, and 11 long-period instruments at 20 km intervals. Broadband data was recorded for 2 days using Metronix-ADU 07 while long-period data was measured in the 1-3 weeks range using Lemi 417M. We processed data using the robust regression code of Egbert (1997) in single-site mode. Overall, the data is high quality, with broad-band achieving 1000 s for almost all stations, and 10000 s where long-period data is available (supplementary material). However, two sites had to be removed due to interferences in the equipment during acquisition.

We addressed the dimensionality analysis and geostrike calculation with the phase tensor technique of Caldwell et al. (2004) implemented in the MT-py package (Krieger and Peacock, 2014). Z can be written in terms of the real and imaginary part as $Z = X + iY$, where the phase tensor can be calculated by $\Phi = X^{-1}Y$. The main implications are that there is no need to assume a dimensionality to the medium and that Φ is not affected by galvanic distortion. Φ is interpreted mainly by three rotational invariants, namely $\Phi_{\min}$, $\Phi_{\max}$, and skew (β). The latest is a measure of the asymmetry of the regional MT response being equal to zero for 1D and 2D resistivity distributions. $\Phi_{\max}$ and $\Phi_{\min}$ are the axes of the ellipse. The main axis tends to parallel the direction of the principal induction current flow (Yin et al, 2021).

We considered ellipses with skew angle out of the range $|\beta| > 5^\circ$ with 3D behavior. Fig. 4.5A shows the skew angle as a function of period and sites. We note that most of the periods are inside the range adopted, thereby indicating 1D (up to 1 s) or 2D behavior (periods > 1 s). However, some ellipses indicate 3D behavior for periods > 100 s.

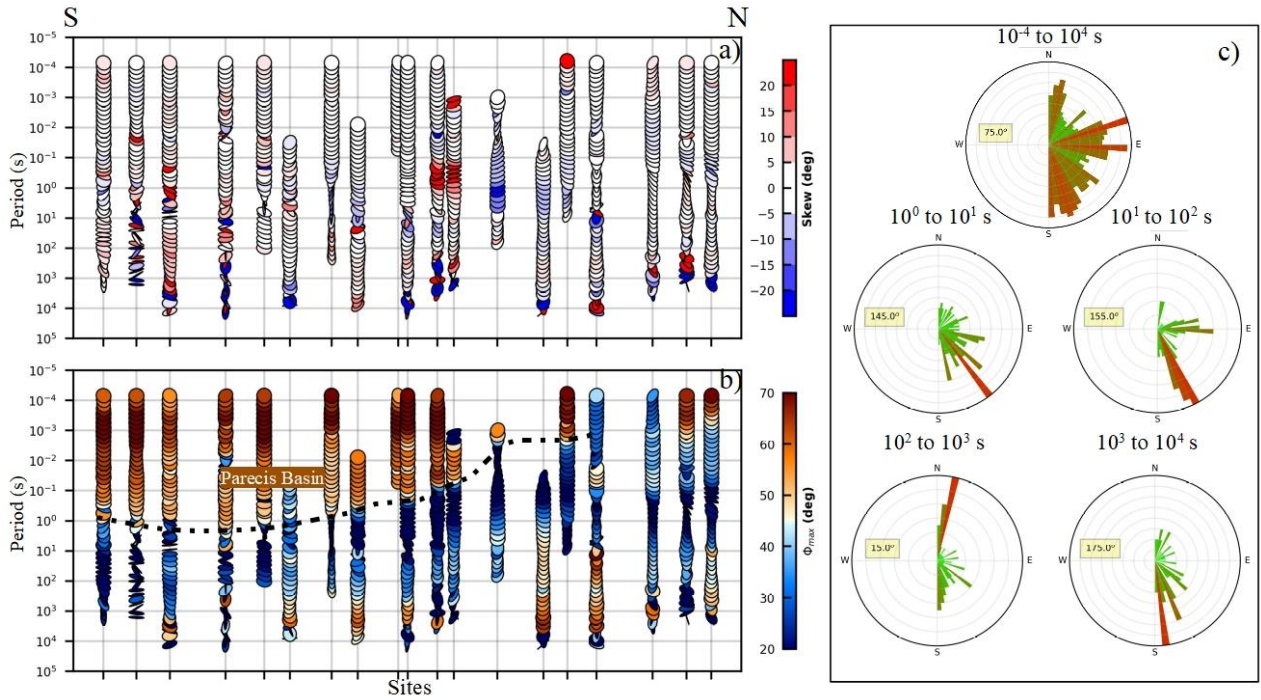


Fig. 4.5 - Dimensionality and geostrike analysis. a) Skew angle indicates 1D or 2D behavior up to 100 s, 3D behavior dominates for periods between 100 and 10000 s. b) $\Phi_{\min}$ is a proxy for conductivity. It highlights that the Parecis Basin may comprise periods < 1 s and gradually thins northwards. c) Geostrike plotted in specific period ranges. The geostrike spreads over several directions in the range of 10^{-4} - 10^4 s mainly because half of the period decades are dominated by the Parecis Basin. On the other hand, a stable geostrike is present in the basement for periods 1 – 100 s, and for the following decades, the geostrike rotates significantly due to 3D behavior

In Fig. 4.5b we plotted $\Phi_{\min}$ because it can be used as a proxy for conductivity. We note two main patterns i) $\Phi_{\min}$ larger than 50 degrees are interpreted as the Parecis Basin whose thickness decreases northwards ii) $\Phi_{\min}$ smaller than 50 degrees represents the basement, though $\Phi_{\min}$ increases for periods > 1000 s. We note that the ellipses consistently rotate in the transition from the Parecis Basin to the basement. Simpsons and Bahr (2005) argue that the dimensionality analysis should be undertaken as a hypothesis test: is there a stable geostrike? Thus, we investigate whether our dataset's strike is frequency-independent. Fig. 4.5C shows the strikes per decade of periods. Considering the entire period range, there is a high spread of directions due to the ellipses in the Parecis Basin (0.0001 – 1 s). In the basement (period > 1 s), we note a consistent geostrike direction for the decades 1-10 and 10-100 s, however, relevant rotation occurs from 100 -10000 s. Therefore, we

conclude that the stable geostrike hypothesis holds only for periods up to 100 s. Thus, we opted to invert data in 3D.

4.3.3.2 Resistivity inversion model

We inverted data in the period range of 10^{-3} to 10^4 s with 5 points per decade. We applied the nonlinear conjugate gradient for the 3D case using the ModEM code of Kelbert et al. (2014).

The model space has a total length of 969, 856, and 948 in the NS, EW, and vertical directions, respectively. The finer area of the mesh covers an area of 200 x 90 km with a cell size of 2 x 2 km, which is about 1/5 of the site spacing. Seventeen padding cells were added in 1.5 increments to minimize edge effects. Vertical cell size started at 25 m and was incremented by a factor of 1.2.

Robertson et al. (2020) show that the 3D inversion is strongly sensitive to the resistivity of the prior model. Thus, we varied our prior half-space model to cover resistivities similar to the sedimentary basin to the resistivities of the basement: 50, 100, 500, and 1000 ohm. We run inversions with error floors of 20% and 10% of the $\sqrt{|Z_{xy} \cdot Z_{yx}|}$ for the diagonal and off-diagonal impedance tensors, respectively. The smoothing factor was initiated as 10, and it was divided by a factor of 10 if the difference between the current and previously RMS model was less than 0.002% until it reached 10^{-8} .

In Fig. 4.6, we show the RMS for each half-space as a function of the sites. The RMS is higher for half-spaces of 500 and 1000 ohm, and yielded a similar result for 50 and 100 ohm.m, however, the last achieved an equivalent RMS nine iterations before. Thus, we chose the model of the prior half-space of 100 ohm.m (Fig. 4.7).

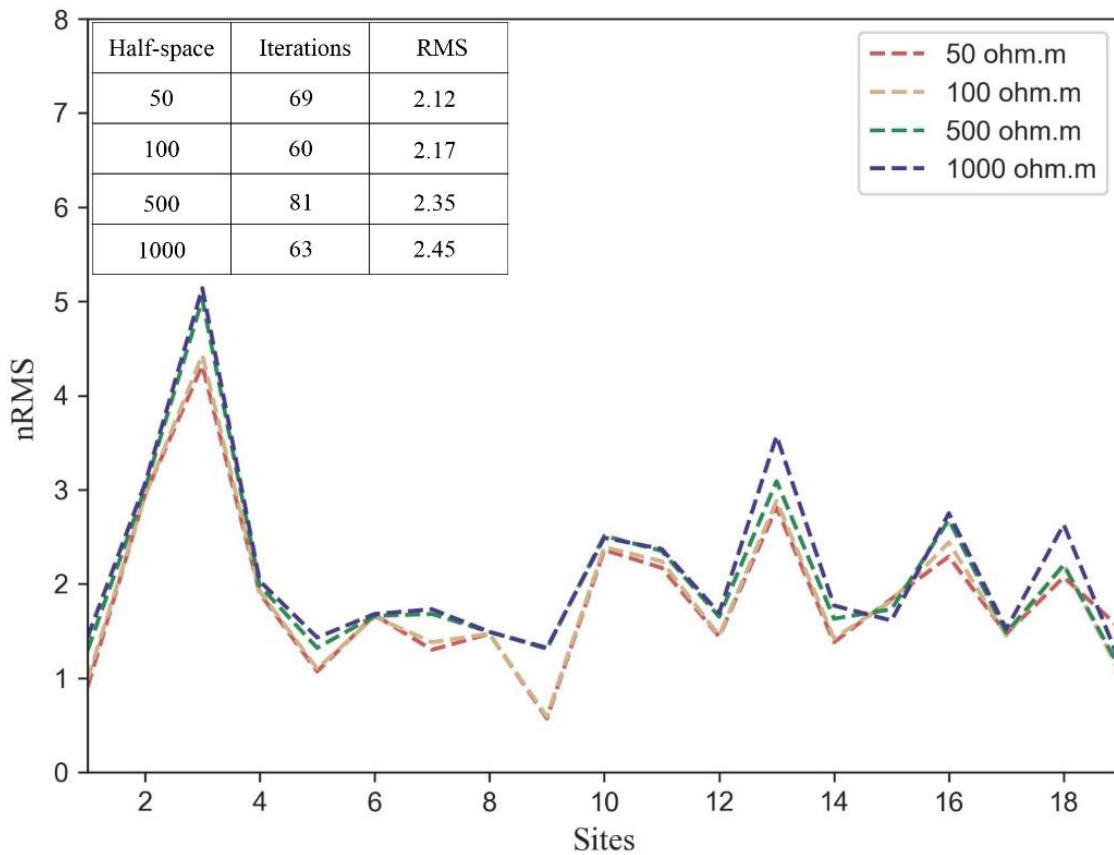


Fig. 4.6 - Normalized RMS plotted per station and homogeneous half- space. Normalized RMS is lower for the 50 and 100 ohm.m half-spaces. The inversion was more efficient for 100 ohm.m because the equivalent RMS achieved 9 iterations before.

We also carried out sensitivity tests on key anomalies to prove their robustness. The anomalies circulated in dashed lines were replaced by resistors of 500 ohm.m (Fig. 4.7). We recalculated the tensor impedances and recomputed the RMS for each site. We note an increase in the RMS for sites above these conductors, which indicates that data require these features.

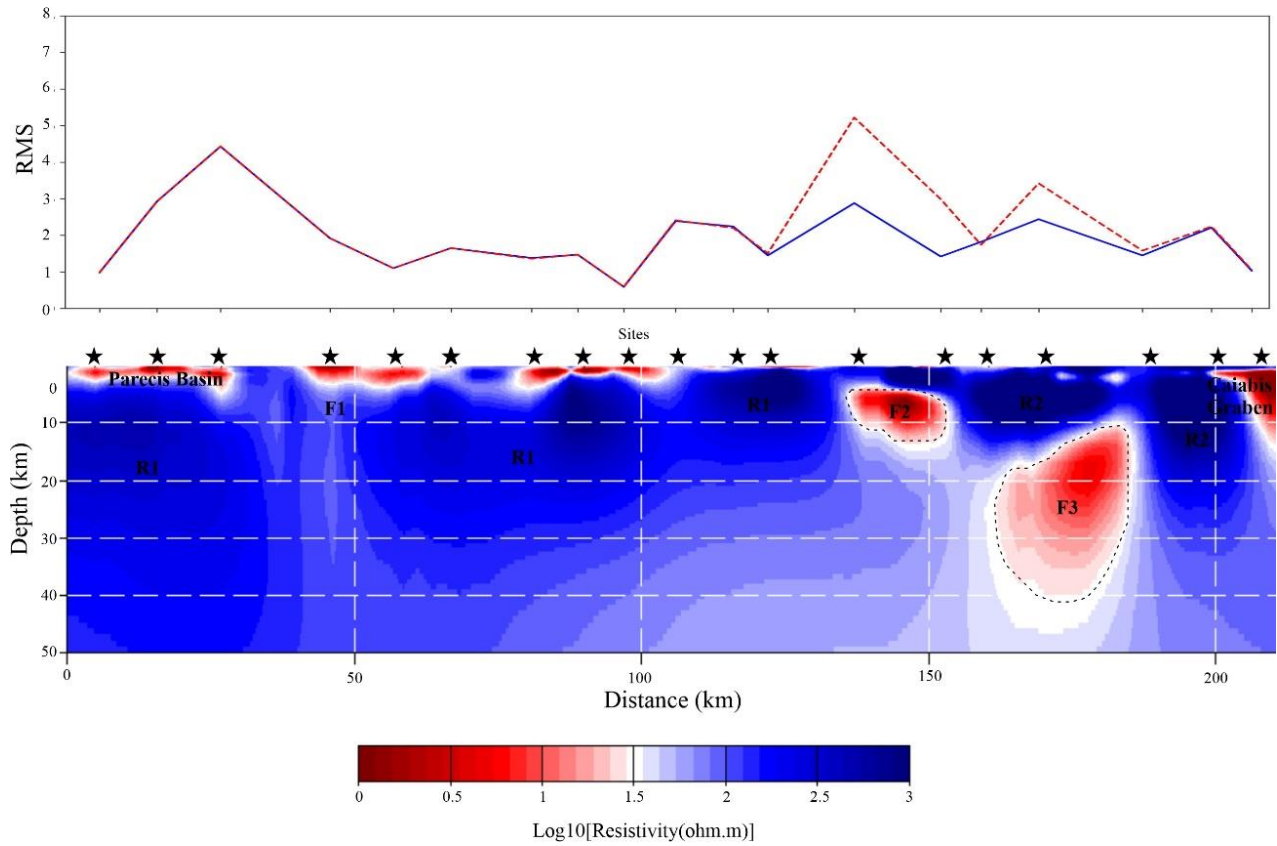


Fig. 4.7 - 3-D Inversion. Upper panel: blue line shows the RMS per site for the preferred model. The dashed red line shows the RMS of the forward response when the conductors F2 and F3 are replaced by resistivities equivalent to the background. The increase in RMS indicates that the conductors are robust and required for the model. Lower panel: Preferred resistivity model. R2 is more resistive than R1, and F1, F2, and F3 crosscut the high resistivity pattern.

4.4 Results

4.4.1 Potential Field Methods

In Fig. 4.3, we interpret three main crustal-scale structures. L1 and L3 (E-W to NW trending directions) marking transitions in the magnetic gradient background and or changes in the inside pattern of second-order structures. On the other hand, L2 with E-W direction has more than 1000 km and is marked along the wavelength of the anomaly. L1 separates a domain of low magnetic gradient and mid to low Bouguer anomaly composed of Tapajós Province as well as the rocks of Teles Pires Suite and Colider Group. We note an increase in the magnetic gradient and Bouguer anomaly towards Juruena Terrane, except in the southern portion, due to the signature of the metavolcanic-sedimentary rocks of the Roosevelt Group. The northern portion of the Juruena Terrane comprises NE trending structures associated with high Bouguer anomaly, which gradually rotate to E-W towards L2. In the

Parecis Basin, there is an increase in the magnetic gradient associated with an increase in the Bouguer anomaly. The Bouguer anomaly shows paired long-wavelength anomalies (~ 100 km) of high and low signatures approximately along the L2 structure. However, this pattern is overprinted by dominantly high Bouguer values towards Jamari and Paragua Craton. From the Parecis Basin towards the Juruena Terrane, receiver function data suggests crustal thickening (Albuquerque et al., 2017). The L3 structure crosscuts L2 and highlights a high-to-low magnetic gradient transition. There are NE trending structures in the Jamari terrane that continue to the Juruena Terrane, on the other hand, E-W structures break at the L3 structure. In the Jauru Terrane, we note a completely different pattern composed of short wavelengths and circular features.

Ground gravity data also shows long-wavelength anomalies (~ 100 km) of high and low. The peak-to-peak amplitude is about ~ 80 mgal. The crustal thickening of 6 km towards the north yielded ~ 45 mGal amplitude (Fig. 4.4a). The high Bouguer is explained by the elevation of the mafic/ultramafic material of the lower crust to at least 10 km. Consequently, an increase in granitic/gneissic material in the north fits the low Bouguer anomaly. Albuquerque et al. (2017) also indicate anomalous mafic composition for the crust under Parecis Basin, which presents the largest value of V_p/V_s ratio (1.86) of the Amazon craton. It is noteworthy that the same anomaly could also be explained by reducing the width of rising mafic material and extending it to shallower levels.

4.4.2 Magnetotellurics

For the 3-D model, Parecis Basin is constrained with a resistivity below 30 ohm.m, and the depths to the bottom are about 2 km in the southern portion and shallow northwards (Fig. 4.7). The model could not guarantee the continuity of the Parecis Basin where site spacing is over 15 km. The center-south portion of the profile is composed of basement R1 segmented by F1. We note an increase in resistivity in the upper crust of the northern portion (R2 resistors). F2 is also detected in the 3-D model, with the bottom constrained to ~ 10 km in the 3-D. Additionally, the 3-D model constrained the F3 feature, which extends from the mid-crust toward the base of the crust.

4.5 Discussion

4.5.1 The Paleoproterozoic Cryptic Suture

Tectono-stratigraphic terranes are fault-bounded fragments whose rocks differ totally or partly in geochemistry, geochronology, structural and metamorphic conditions (Comeau et al., 2020). The crustal-scale L2 structure, near high and low paired Bouguer anomalies, as well as a step in the Mohorovic discontinuity towards the Juruena and Jamari terranes, indicate that the rocks underneath the Parecis Basin may have evolved disconnected from the Juruena and Jamari terranes. Therefore, we suggest that L2 is a major terrane boundary.

There are two main hypotheses for the meaning of L2: Tohver et al. (2004) interpret that L2 is the continuation of the Nova Brasiliandia Metasedimentary Belt (NBMB) eastwards, and it marks the suturing of the southern margin of the Amazon craton in latest Mesoproterozoic times, whereas Scandolara et al. (2017) argue that L2 is the suture between the Tapajós Province and the Paraguá protocraton in Paleoproterozoic times. Although the NBMB outcrops only in the western portion of the area (Fig. 4.2), it comprises only metamorphic ages between 1.13 and 1.01 Ga. Metamorphic ages of the older event are still restricted to the rocks of the Jamari Complex, however, this may be biased because the metamorphic dataset is concentrated in the western portion of the study area (Motta et al., 2022).

The electrical resistivity model provides meaningful insights into understanding the relationship between these periods of deformation. Therefore, we integrated the electrical resistivity model with crustal thickness data, depth to the Curie temperature, and the interpreted crustal-scale structure from the magnetic dataset (Fig. 4.10). Faults and suture zones are regions of weakened crust that are usually composed of hydrothermally or metamorphosed minerals (Comeau et al., 2020). For instance, Padilha et al. (2019) mapped crustal-scale conductors associated with paleo-sutures in the São Francisco Craton. The suture signature usually consists of continuous conductors from the surface

to the mantle (Korsh et al., 2012, Jiang et al., 2019). We note that L2 structure correlates with north dipping high conductivity, even though the conductors are discontinuous. Overall, high resistivity in the upper crust ($> 10^3$ ohm.m) is explained by silicate minerals that behave as semiconductors (Selway, 2019). In the northern portion, the resistivity is consistently higher than 10^3 ohm.m, which we attribute to the metagranitoids of the Juruena complex. In contrast, the upper crust is less resistive in the center-southern portion of the model. Furthermore, vertical structures indicate that the Parecis Basin's mafic lower crust is exhumed (Fig. 4.8). Indeed, Yang (2011) demonstrates that increasing iron-bearing minerals, such as pyroxene, reduce the resistivity of the crust. The forward gravity modeling also supports indications of mafic compositions (Fig. 4.4) and the anomalously high V_p/V_s ratio (Albuquerque et al., 2017). Therefore, the electrical resistivity model corroborates that L2 divides two different crusts; however, its contact signature is not characterized by a continuous conductor from the upper crust to the mantle.

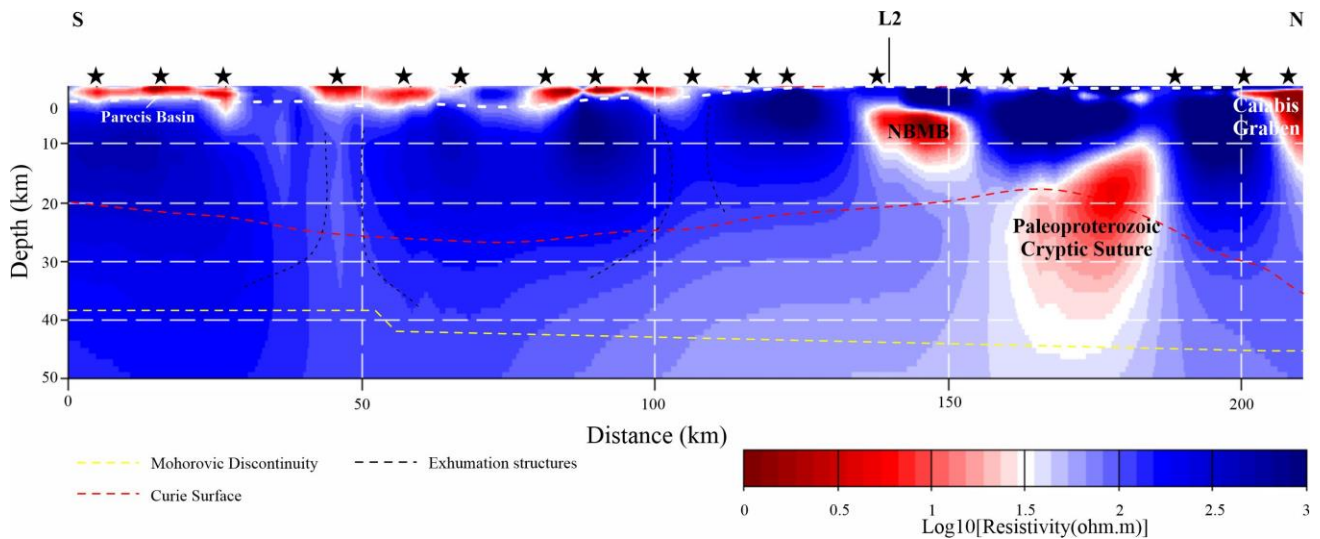


Fig. 4.8 - Preferred electrical resistivity model (from south to north). The white dashed line marks the bottom of the Parecis Basin. Note that two discontinuous conductors correlate spatially along the L2 crustal-scale structure. We argue that the lower crust conductor is evidence of the older metamorphism (1.63 – 1.69 Ga). Its upper crust signature was removed and overprinted in the formation of the NBMB between 1.2 – 1.0 Ga.

The discontinuity between the two conductors corroborates that the deformation occurred in stages. In stable continents, the upper crust presents high conductivity due to interconnected graphite or sulfide minerals, which can occur as grain boundary films or in volumetrically significant phases

(Selway, 2019). Quadros et al. (2021) attest that Mesoproterozoic turbidites of siliciclastic carbonate composition were metamorphosed in medium to high grade in the NBMB. These rocks comprise relevant amounts of sulfide minerals such as pyrrhotite, sphalerite, galena, and chalcopyrite (Bergami and Prado, 2019). Therefore, we interpret the upper crust conductor as an extension of the NBMB. Regarding the lower crust conductor, it is an unusual feature since numerical models predict that the lower crust is slightly less resistive than the upper crust, so resistivities $<10 \text{ ohm.m}$ are rare and due to specific compositions or tectonic processes (Selway, 2019). Kay et al. (2022) located a lower crust conductor in the Curnamona Province and proposed that it is caused by flake graphite from highly metamorphosed carbon-rich sediments in a collisional environment. Another vector of low resistivity in the lower crust is the amount of amphibole. Wang et al. (2012) observed an increase in several orders of magnitude of the conductivity of amphibole for temperatures $> 523 \text{ }^{\circ}\text{C}$. The Curie surface maps the depth to the temperature of $580 \text{ }^{\circ}\text{C}$ (Correa et al., 2022). Thus, we note that most of the lower crust is above this temperature. Moreover, there are some variations in the Curie surface where the shallowest points correlate with the position of the lower crust conductor, thereby indicating a transition from pyroxene-bearing to amphibolite-bearing lower crust in the northern portion. Amphibolite is less dense than pyroxene (Dentith and Mudge, 2014), so it supports the low Bouguer anomaly matching the position of the lower crust conductor.

Amphibole is either crystallized from mantle-derived melts or through the metamorphism of mafic rock in amphibolite-facies conditions (Davidson et al., 2007). Vadoodi et al. (2021) also found isolated lower-crust conductors in the Fennoscandian shield. They argued that precipitation of C-O-H fluids associated with mantle-derived melts is more plausible than transporting organic material from the surface. These fluids would be released in the mantle by the slab dehydration, providing the carbon source. We tend to favor the collisional explanation due to three main reasons: i) the 1.63-1.69 Ga metamorphism is of high pressure (7-8 Kbar), ii) the magmatism ceases during this period, which indicates that the metamorphism is driven by crustal thickening rather than magmatism iii) there are

crustal thickening towards the Jamari/Juruena terranes, even though the average is 46 km, values higher than 50 km are found (Albuquerque et al., 2017). Thus, this lower crust conductor is evidence of the Rio Negro Juruena Orogeny, and it is characterized as a cryptic suture because the evolution of the NBMB overprinted its upper crust signature. One may ask how this conductor was preserved during the formation of NBMB. Selway (2013) attests that grain-boundary graphite films are stable to ~ 900 °C, whereas the temperature peak of the Sunsás tectonothermal event reached 800 °C (Tohver et al., 2004), preserving the buried metamorphosed sediments.

To verify the collision scenario between 1.69 – 1.63 Ga, we compiled geochronological data of the Motta et al. (2022), Johansson et al. (2022), and geological survey of Brazil datasets. We plotted magmatic and metamorphic ages for the rocks at the north and south of the L2 structure (Fig. 4.9). Despite downsampling due to Phanerozoic cover in the south block, we find no correlation between the crystallization peaks of Tapajós basement and Juruena/Jamari complexes when compared to Pre Rondonian-San Ignacio rocks. Indeed, the Jamari Complex only outcrops to the north of the L2 boundary. The first correlated peak occurs at about 1550-1570 Ma, which is interpreted as post-collisional A-type magmatism (Scandolara, 2006, Scandolara et al., 2013b, Cordani et al., 2016, Bergami and Prado, 2019 Cordani et al., 2016). This magmatism could represent the collapse of the orogen, and the fact that it occurs on both sides of L2 corroborates that the two blocks were the same tectonic entity after 1.63 Ga.

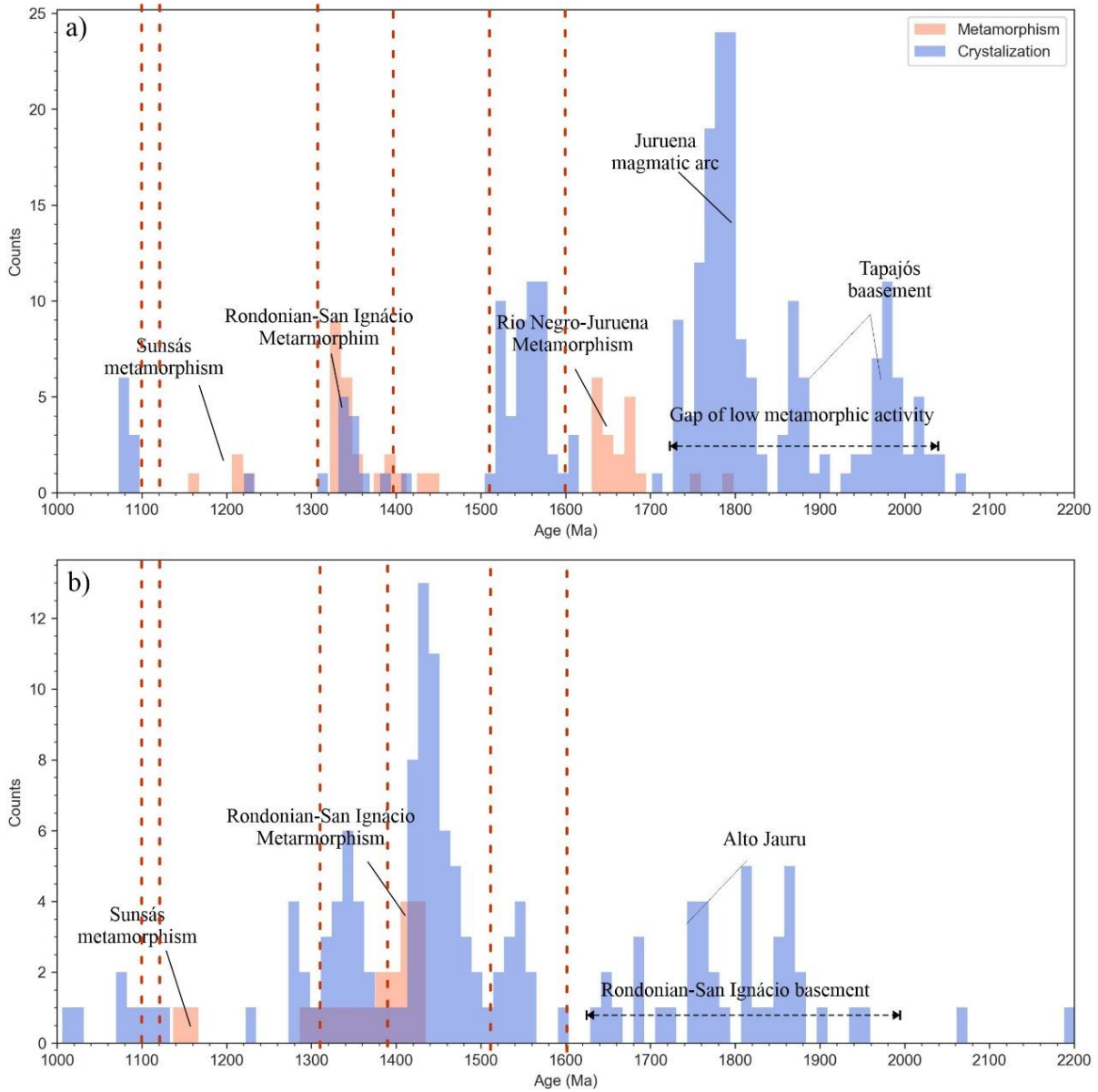


Fig. 4.9 - Comparison of magmatism and metamorphism at north and south of the L2 crustal-scale structure a) Juruena and Jamari terranes. b) Jauru terrane and Paragua Craton. The red dashed lines represent periods of A-type magmatism. The two tectonic fragments present remarkable differences in magmatism before Rio Negro-Juruena metamorphism. The magmatism peak of 1.55 Ga is present in both locations, thereby indicating that the two tectonic fragments were already a unique tectonic entity.

4.5.2 Recurrence of tectonothermal events and crustal overprint

Once formed, a fault produces permanent weakness in the crust and may localize significant strain increments regardless of the following stress field. This is particularly evident in cratons where faults may have been reactivated several times (Torgensen et al., 2022). Meert and Santosh (2022) attest that collisional orogens are good places for new oceanic crust formation during subsequent continental breakups. This topic links how the following tectonothermal events overprinted the Paleoproterozoic crust.

The high concentration of A-type granitoid (1600- 1510 Ma) in the neighborhood of L3 structure may indicate the corridor where the Paleoproterozoic crust started to break up. In Mesoproterozoic, accretion of island arc and juvenile crust continued in the margin of what today is the Jauru terrane (Bettencourt et al., 2010; Rizzotto et al., 2013), where the peak of crust formation occurs about 1440 Ma (Fig. 4.9B). Contemporaneously, the Jamari and Juruena terranes accommodated strain due to the far-field. This deformation is dated along the L3 structure in the 1.37-1.42 Ga range, marking the basin inversion. Several orogens and suture zones are speculated between Jauru Terrane and the Paraguá Craton, however, potential field data do not point out another major boundary besides L3. Because there is evidence of a collision in the Bolivia, further research should be conducted. The magmatism between 1.40 and 1.29 Ga marks the late to post-collisional phase, followed by cratonization.

In the study area, the Sunsás Orogeny is characterized by crustal reworking rather than juvenile crustal accretion (Santos et al., 2000), so magmatism after 1.2 Ga is scarce. Based on the U-PB ages of zircon, Quadros et al. (2021) show that deep marine sediments deposited between 1.24 and 1.13 Ga were followed by a period of bi-modal magmatism between 1.12 and 1.10 Ga. As aforementioned, this rifting process developed along the weakened crust close to the Paleoproterozoic cryptic suture

overprinting its upper crust signature. In a second stage, the far-field inverted and basin inversion started; lately, the L2 and L3 reactivated into crustal-scale shear zones.

Condie et al. (2021) indicate two main periods of crustal assembly in the Paleoproterozoic. These periods are related to the supercontinents Nuna (1.9 – 1.5 Ga) and Rodinia (1.1 – 0.8 Ga). Johannson (2009) and Johannsson et al. (2022) argue that the Baltica and Amazon cratons formed a purely long-lived accretionary margin between 2.0 and 1.3 Ga. Magmatism of 1790 Ma (Bispo-Santos et al., 2008) and 1535 Ma (Bispo-Santos et al., 2020) favor this hypothesis. However, Meert and Santosh (2022) argue that the link between Amazon and Baltica is fragile after 1500 Ma. Likewise, D'Agrella-Filho et al. (2021) point out that Amazonia and Baltica took a different polar wander path after 1.44 Ga. However, in terms of magmatism and metamorphism, the Hallandian and Sveconorwegian orogenies match the Rondonian-San Ignácio and Sunsás orogenies in time, respectively (Stephens and Wahlgren, 2020). The older metamorphic event, with which Rio Negro-Juruena is more than 100 Ma younger, has the greater discrepancy. Assuming that Amazon and Baltica shared the same peri-Nuna subduction system, the lack of high P metamorphism in Baltica indicates that the collision of Amazon Craton was against a microcontinent between 1.63-1.69 Ga. Indeed, the last pulse of Transscandinavia magmatism is about 1.65 Ga, followed by a quiescence period of almost 200 Ma. Therefore, even though Baltica and Amazonia could be close in the Paleoproterozoic, the same peri-Nuna subduction system could be shared only until the collision of the Amazon Craton and a microcontinent. The crustal accretion related to the Rondonian-San Ignácio orogeny occurred in an orthogonal continental margin compared to the previous one. Indeed, Scandolara et al. (2017) show evidence that the convergent vector of the Rio Negro Juruena and Rondonian San-Ignácio trend in the NS and EW directions, respectively.

4.6 Conclusions

This research has investigated polyphase deformed terranes. New broad-band and long-period magnetotelluric data were integrated into the potential field, crustal thickness, and depths to Curie

temperature to constrain the upper and lower crusts in detail. We elaborate on the meaning of fault-bounded tectonostratigraphic terranes. A major crustal-scale structure separates two tectonic fragments with remarkable differences in the magnetic gradient and structural trend. We also find along this major boundary high and low Bouguer anomalies of long-wavelength paired, a step in the Mohorovic discontinuity, and a lower crust conductor. This evidence supports the hypothesis of a collision in Paleoproterozoic times. Magmatic and metamorphism ages were analyzed to verify this scenario and we found patterns before and after the collision.

Our findings suggest that the study area did not remain purely accretionary in Paleoproterozoic times and evolved to a collisional type between 1.63 and 1.69 Ga. This contrasts with the purely accretionary orogens described in the Baltica Craton, thereby undermining the hypothesis that Amazon and Baltica cratons shared the same peri-Nuna subduction system from 2.0 to 1.3 Ga. Further research should be done to i) date the older metamorphic event in the rocks of the Jurueña Complex ii) magnetotelluric data should be acquired along the transition between Jamari and Jurueña terranes, and iii) paleomagnetic data should be reinforced to better constrain the position of the Amazon Craton in Mesoproterozoic times.

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CAPÍTULO 5

CONCLUSÕES E CONSIDERAÇÕES FINAIS

A presente pesquisa identificou estruturas de escala crustal e discutiu seus significados tectônicos com base em modelos termais e elétricos inéditos para a crosta do Cráton Amazônico. Além disso, esses resultados foram integrados com dados de sísmica passiva e geocronologia.

Os domínios Bacajás e Carajás são separados por um contato tectônico de escala crustal e apresentam diferenças significativas em termos de espessura crustal. Por outro lado, não há uma estrutura semelhante que separe os domínios Tapajós e Iriri Xingu, o que corrobora a hipótese de um embasamento arqueano comum. Analogamente, nota-se uma mudança na estrutura termal entre os domínios Tapajós e Rondônia Juruena, todavia sem a presença de uma estrutura tectônica profunda. Isso confirma a evolução de um arco acrescionário de vida longa de 2,05 -1,74 Ga.

A partir da superfície Curie, permitiu-se concluir que o manto na porção leste do domínio Carajás e no domínio Tapajós preservaram minerais serpentinizados. O modelo de resistividade elétrica confirmou que o manto do domínio Tapajós foi hidratado e refertilizado devido ao processo de subducção. Correlações de anomalias de alta condutividade no manto e depósitos de classe mundial de cobre pórfiro reforçam a fertilidade mantélica da Província Mineral do Tapajós.

Interpretou-se o evento metamórfico datado entre 1,69 – 1.63 Ga como a evolução de um orógeno acrescionário para um colisional. Na transição de uma descontinuidade crustal de mais de 1000 km de extensão, identificou-se i) um condutor na crosta inferior, ii) espessamento crustal, iii) e um par de alto e baixo Bouguer de grande comprimento de onda. É improvável que as rochas de crosta oceânica aflorem devido aos processos de reativação no Mesoproterozóico. Recomenda-se i) datar os eventos metamórficos no Complexo Juruena, ii) adquirir dados magnetotélúricos ao longo da transição entre os terrenos Jamari e Juruena e iii) avançar os estudos paleomagnéticos nas rochas Mesoproterozoicas, iv) calcular a espessura crustal nos domínios Rio Maria e Iriri-Xingú.

As evidências encontradas nesta tese corroboram a evolução crustal simplificada apresentada na figura 5.1. Um orógeno acrescionário de longa vida se desenvolveu formando os arcos continentais referentes as rochas intermediárias dos arcos Tapajós e Juruena (Figura 5.1a e b). A margem continental ativa evolui para um orógeno colisional apenas entre 1,69 - 1,63 Ga (Figura 5.1c). O consequente espessamento crustal resulta no colapso do orógeno, que é registrado pelo magmatismo intraplaca tipo A da Suite Serra da Providência. A orogênese Rondoniano San-Ignácio retrabalha os terrenos Jamari e Juruena com pico de metamorfismo entre 1,42-1,37. Neste período, a atividade magmática é escassa. Entre 1,24 – 1,10 Ga, um processo extensional é iniciado aproveitando a zona de fraqueza crustal da sutura Paleoproterozóica, onde sedimentos de mar profundo e magmatismo bimodal dão início a Faixa Nova Brasilândia, que é deformada entre 1.10 e 1.01 Ga.

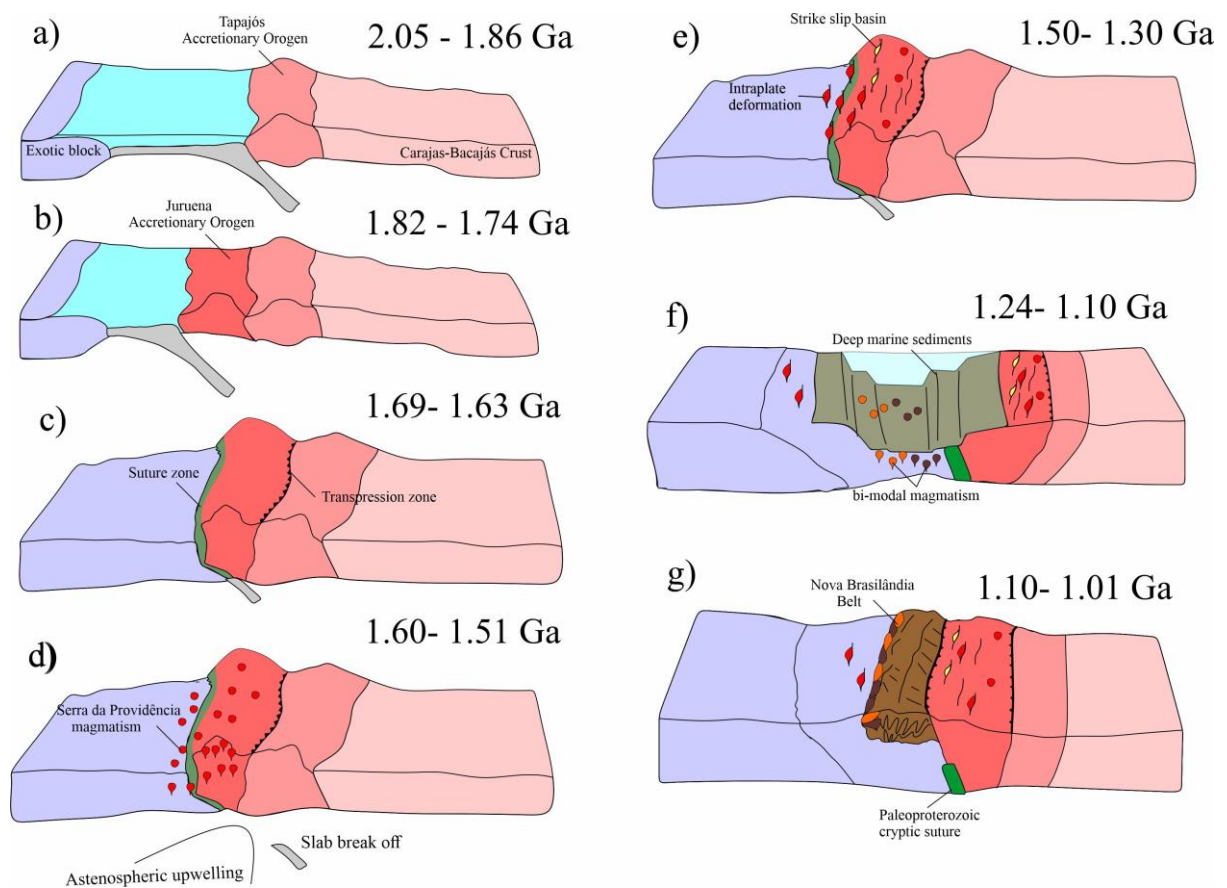


Figura 5.1. Evolução geológica simplificada do Orosiriano ao fim do Mesoproterozoico. A) Evidência: ocorrência de rochas cálcio-alcálinas típicas de arco continental sem metamorfismo de alto grau (Cassini *et al.*, 2020). B) Evidência: ocorrência de rochas cálcio-alcálinas típicas de arco continental (Scandolara *et al.*, 2017). C) Evidência: Idades de metamorfismo associadas à diminuição da atividade magmática, além de vínculos geofísicos apresentados nesta tese (Scandolara *et al.*, 2013a). D) Evidência: granitos tipo A (Scandolara *et al.*, 2013b). E) Evidência: idades de metamorfismo e idades máximas de sedimentação das bacias *strike-slip* (Scandolara *et al.*, 2017; Leite e Sales, 2003). F) Evidência: turbiditos e intrusivas félsicas e máficas (Quadros *et al.*, 2021). G) Evidência: idades de metamorfismo (Quadros *et al.*, 2021). Adaptado de Condie (2013) e Bergami e Prado (2019).

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Apêndice A – Supporting Information for Mapping the thermal structure of the Amazon Craton to constrain the tectonic domains

Journal of Geophysical Research – Solid Earth

Supporting Information for

Mapping the thermal structure of the Amazon Craton to constrain the tectonic domains

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Introduction

In Text S1, we present several tests to choose the trade-off between a window size large enough to sample long-wavelengths with minimum contamination by surrounded geological environments. In Text S2, we discuss possible meanings for the DBML model (thermal or compositional boundary) and show why it can be used to constrain the thermal structure.

Text S1. Window Size Selection

Power spectrums for each WS located in Figure 3 are shown in Figure S1. We noticed that only the 200x200 km WS does not present a peak so we discarded it. The peak is approximately at the same position for the remaining WS. This amplitude changes proportionally to the WS. The exponential decay rate decreases as the WS increases.

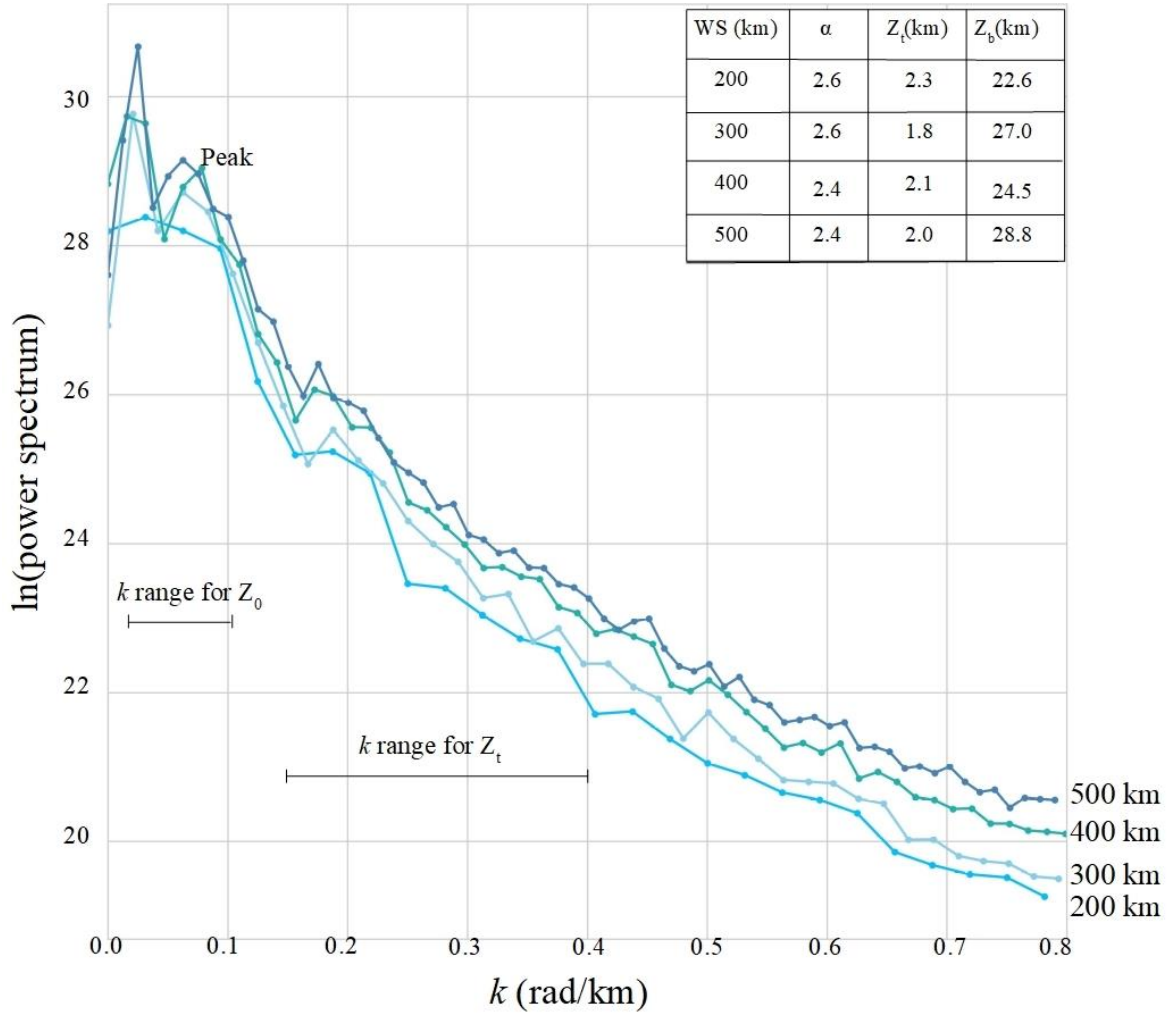


Figure S1. Power spectrum for the WS of 200x200, 300x300, 400x400, and 500x500 km. The peak position is used as a reference for choosing the wavenumber ranges. The Z_t depth is calculated fitting a straight line in the wavenumbers after the peak position. The Z_0 depth is calculated fitting a straight line in the neighborhoods of the peak in the Azimuthally averaged frequency scaled power spectrum (equation 5, not shown in this figure). There is a small variation in Z_b depths in WS larger than 300x300 km.

Choosing k wavenumber ranges to fit Z_0 and Z_t is challenging because it depends strongly on the user. In this sense, Li et al. (2013) suggested a fixed wavenumbers range, however, it can be misleading since peak position changes over different geological environments. Ravat et al. (2016) suggest fitting the centroid depth in the steepest slope using 2 or 3 consecutive points.

Our goal is to avoid choosing different wavenumber ranges in separate runs. Centroid slope controls peak position, and the depth to the top slope controls exponential decay rate. We propose the following approach i) identify the peak position; if the peak does not appear, we vary α from 0 to 4 until it appears ii) Choose the k range for Z_t after the peak position to constrain the main decay rate of the curve. iii) Choose the k range for Z_0 in the neighborhood of the peak in the azimuthally averaged frequency scaled power spectrum (equation 5, not shown in Figure S1). With these wavenumber ranges settled, the program calculates depths Z_t (equation 4), Z_0 (equation 5) and consequently Z_b (equation 6) for all α , then we compare the de-fractal curve (equation 8) and the modeled curve (equation 3) as shown in Figure S2.

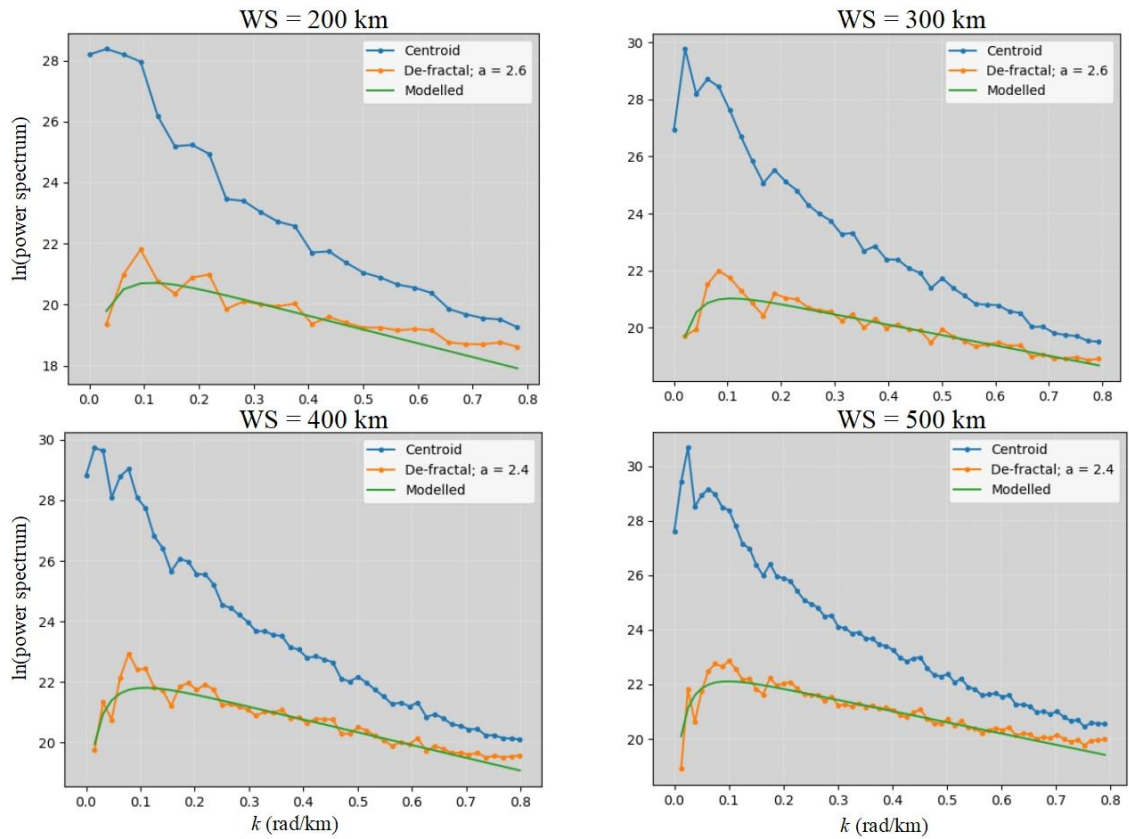


Figure S2. Spectrums of the centroid method (original), calculated (de-fractal spectrum) and modeled for the α associated with the lowest RMS. Visual inspection of

the best matching between calculated (de-fractal spectrum) and modeled spectrum for WS of 200x200, 300x300, 400x400, and 500x500 km.

Typically, the best results are around the lowest root mean square (RMS). However, we make a visual inspection of the matching between de-fractal and modeled curves for the entire α range. Salem et al. (2014) argue visual inspection is rather effective than an automatic method. The main reasons are i) the scientific problem is non-unique; ii) The solutions may be constrained by external information; iii) The different types of misfit norms, and consequently, its minimum misfits return different solutions.

We consider this approach robust since it considers different peak positions making other users choose similar k ranges, even though it may underestimate some depths if the steepest slope does not coincide. The steepest slope is usually found with 2 consecutive points so that we prefer not to fit our data with such a small number of points.

WS test results show WS of 200 and 300 km returned α of 2.6, which decreases to 2.4 for WS of 400 and 500 km. RMS misfit between the de-fractal and modeled curves are 0.42, 0.30, 0.31 and 0.33 for WS of 200, 300, 400 and 500 km, respectively. No relevant change is observed between 300 to 500 km WS. Z_t depth has a maximum change of 0.5 km, and excluding the 200 km WS, there is no relevant change in the depth of Z_b . In Figure S2, a visual inspection of the matching between curves shows 500x500 km WS presents the best result for a Z_b of 28.8 km. However, we opt for 300x300 WS since it allows computation in a more restricted area, thereby avoiding contamination of the spectrum, and it returned a Z_b with less than 2 km difference from the best WS.

Text S2. DBML meaning and its caveats

We derived a heat flow model based on Z_b estimates. Then, we argue whether it can be used to constrain the thermal structure. Assuming that heat transport is by conduction in the magnetic layer and radioactive heat production decays with depth. The surface heat flow is strongly controlled by concentrations of the radioactive, heat-producing isotopes and on the basal heating of the continental lithosphere due to mantle convection (Turcotte & Schubert, 2014). For an older stable continental craton, we may assume a low basal heat flow and constant. Therefore, the crust composition and thickness control the heat flow on the surface.

According to Li et al. (2017), heat flow on the surface Q_s can be predicted using the expression:

$$Q_s = K \frac{T_c - T_0}{Z_b - Z_s} + h_r^2 H_0 \frac{e^{\frac{-Z_b}{h_r}} - e^{\frac{-Z_s}{h_r}}}{Z_b - Z_s} + h_r H_0 e^{\frac{-Z_s}{h_r}}$$

Where K is the average thermal conductivity of the magnetic layer, T_c is the Curie temperature, T_0 is the temperature at the surface elevation Z_s , H_0 is the heat

production rate at the surface and h_r is the drop-off of heat production. Therefore, we applied this equation using our Z_b model, $K = 2.5 \text{ W(m}^\circ\text{C)}^{-1}$, $Z_s = -1 \text{ km}$, $H_0 = 2.0 \mu\text{W/m}^3$, $h_r = 10 \text{ km}$, $T_c = 580^\circ\text{C}$, $T_0 = 30^\circ\text{C}$ (Figure S3).

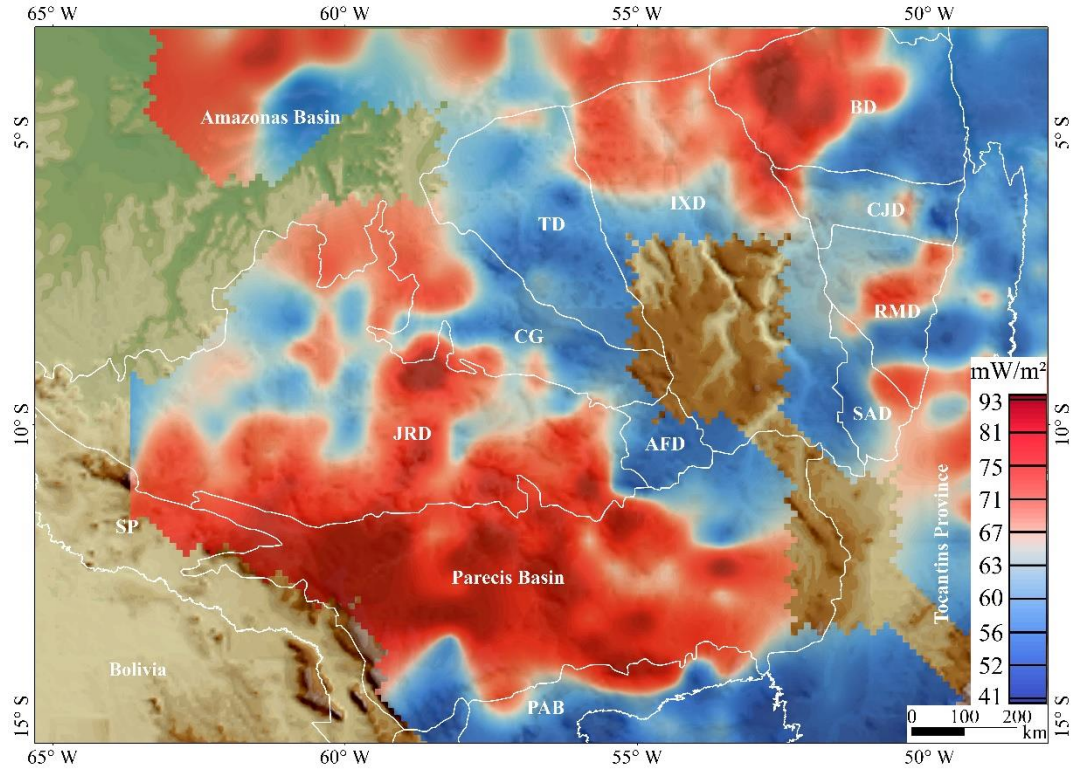


Figure S3. Heat flow on the surface Q_s . The models present an average of $67 \pm 13 \text{ mW/m}^2$, which is in agreement with heat flow derived from the continental crust. The model is strongly biased on Z_b estimates. White lines are the limits of the Geochronological Province model adapted from Vasquez et al. (2008b). CJD: Carajás Domain, RMD: Rio Maria Domain, SAD: Santana do Araguaia Domain, BD: Bacajás Domain, IXD: Iriti Xingu Domain, TD: Tapajós Domain, AFD: Alta Floresta Domain, JRD: Juruena Domain, SP: Sunsás Province, CG: Cachimbo Graben, PAB: Paraguay-Araguaia Belt.

Davies and Davies (2010) argue the world continental mean heat flux is 70.9 mW/m^2 . Goutorbe et al. (2008) found an average of $41 \pm 8 \text{ mW/m}^2$ for the Pilbara, Yilgarn and Gawler cratons in Australia, whereas, for the Proterozoic orogens, the average is $80 \pm 24 \text{ mW/m}^2$, much larger and variable. The authors attributed these changes to the depleted Archean crust in comparison with Proterozoic Regions. Our model determines an average of $67 \pm 13 \text{ mW/m}^2$ for our study area. 70% of our data is in the $40\text{--}60 \text{ mW/m}^2$ range. The areas with heat flow higher than 80 mW/m^2 are concentrated mainly in IXD, western of BD and Parecis Basin.

This model is strongly biased by the DBML model and relies on the premise that the Z_b is the depth to the Curie Temperature. However, Z_b may instead map a compositional boundary. For instance, the Mohorovic discontinuity (Moho) used to be understood as a boundary that limits the presence of magnetic materials (Wasilewski et al., 1979; Wasilewski & Mayhew, 1992). Therefore, we analyzed extreme scenarios to check for compositional or thermal boundary explanations.

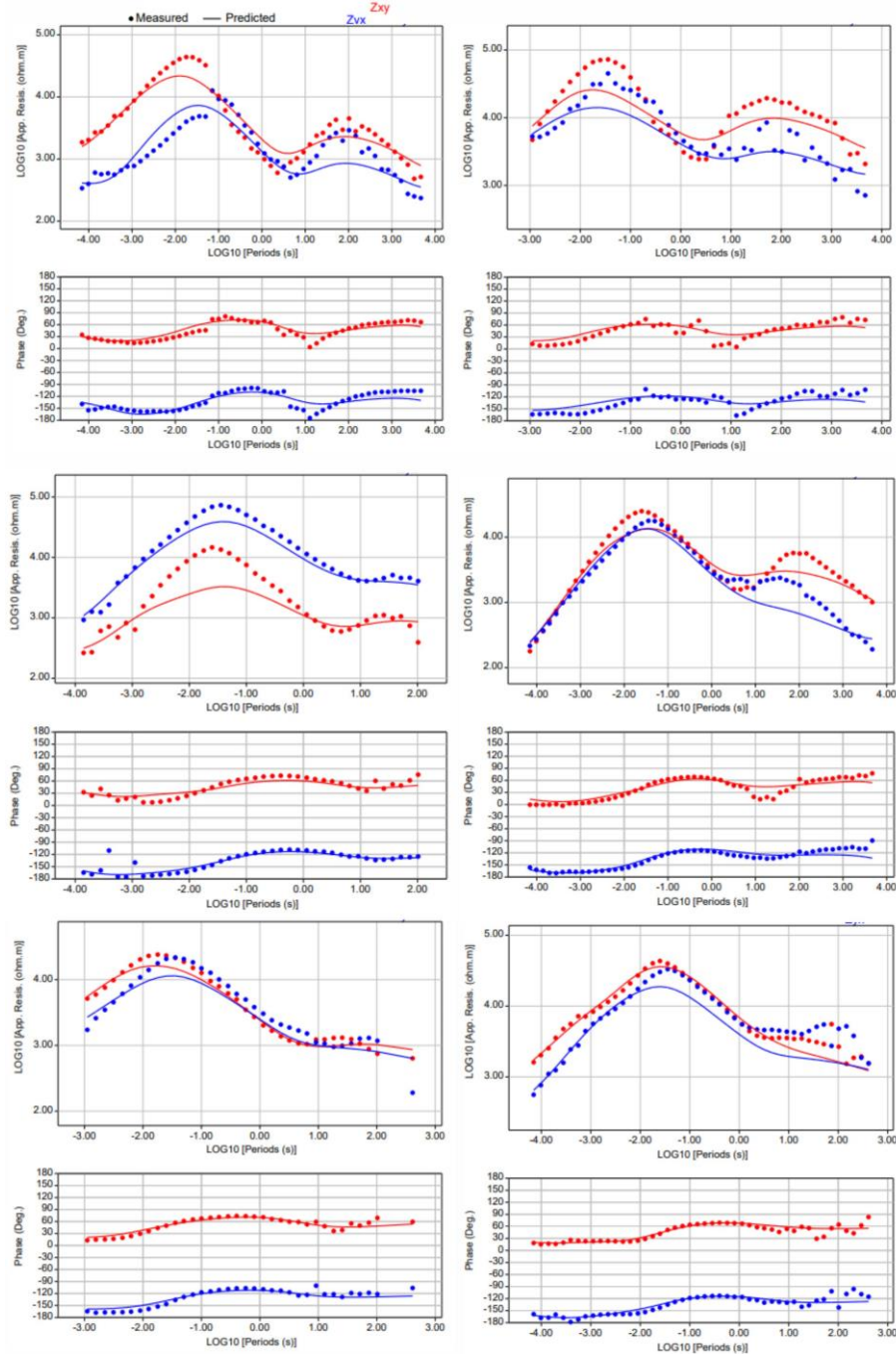
If Z_b represent the depth to the 580 °C temperature, shallow values of our model around 20 km yield a heat flow over 80 mW/m² and a thermal gradient of ~ 30 °C/km. Thus, there are no magnetic sources below this depth, and the longest wavelength is related to the 20 km deep sources. One can have insights regarding the degree level of depletion of the lower crust. On the other hand, depths of 50 km represent heat flow of 40mW/m² and a thermal gradient around 11,6 C/km. Alexandrino and Hamza (2008) have obtained geothermal gradients for the Salvador Craton in Brazil with heat flow in the range of 36-80 mW/m² range. Therefore, our estimates are within the range of expected results.

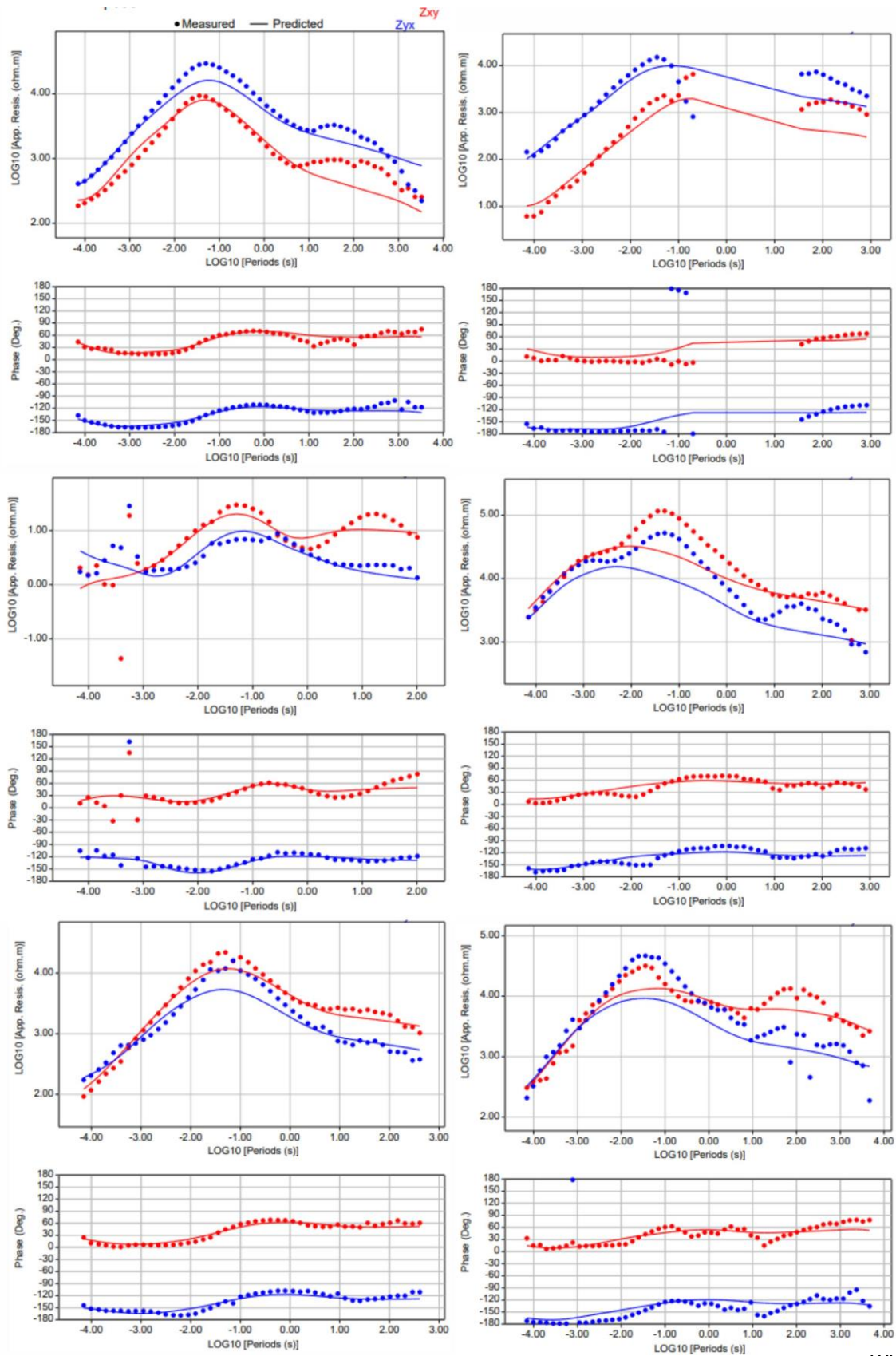
If depths represent a compositional boundary, we may have shallow depths marking some rock strain boundary or volcanic rock layer and thereby underestimating the DBML. Shallow depths concentrate in the i) Parecis Basin and ii) IXD and western of BD. In the first, depths of 20 km are too deep to represent the depth to the basement, and the Moho is around 35 km (Albuquerque et al., 2017). In the latter, the IXD represents the main area of occurrence of the Uatumã Silicic large igneous province (Slip) dated at 1.88 Ga. In North America, the Sierra Madre Occidental is the largest Slip, whose crustal thickness reaches 55 km, while its surrounding is 30 km thinner (Bryan & Ferrari, 2013). Thus, our estimates are likely to be related to the bottom of the Slip. If depths over 50 km represent a compositional boundary, consequently a low geothermal gradient is necessary for preserving the magnetic proprieties. In that scenario, the thermal and compositional boundary may coincide.

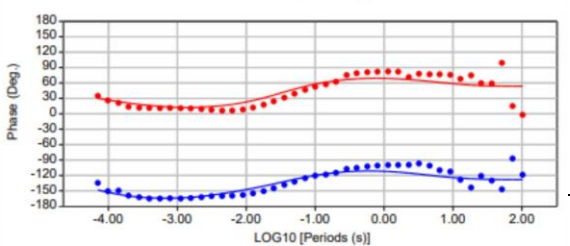
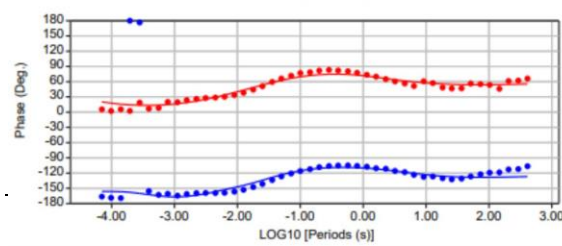
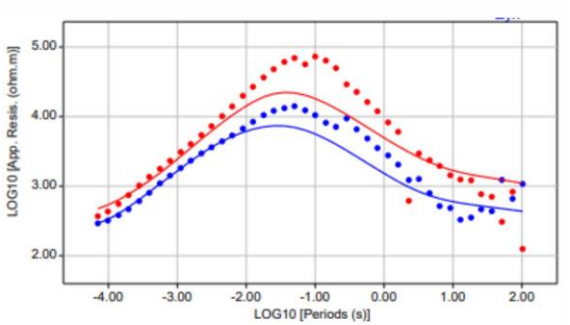
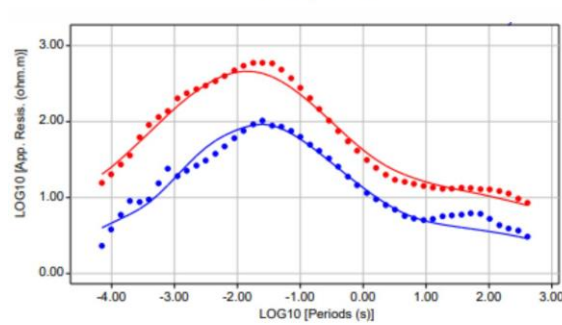
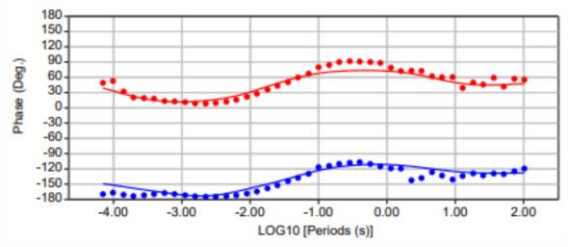
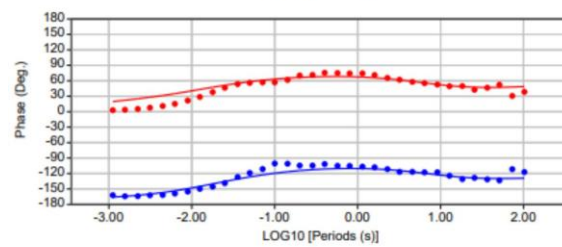
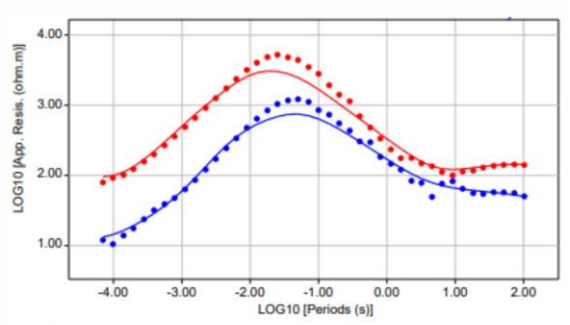
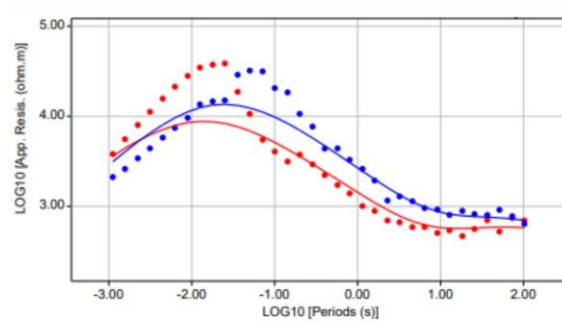
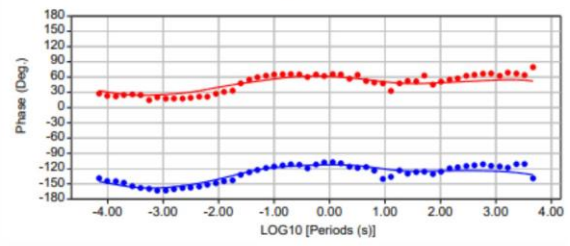
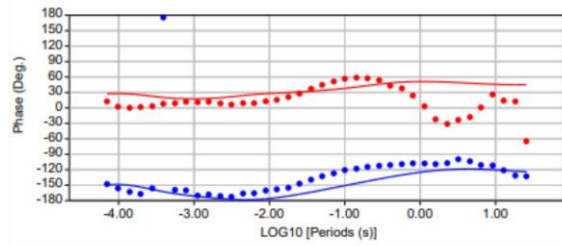
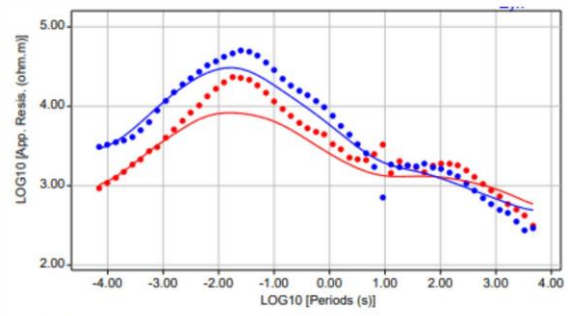
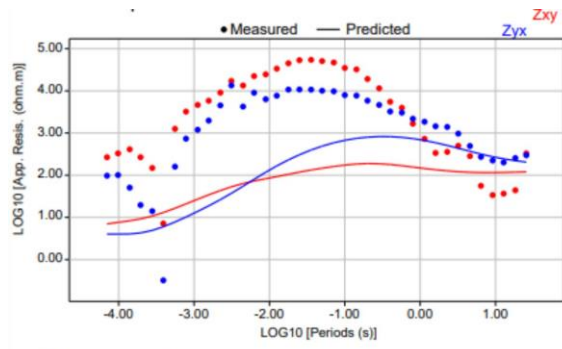
Of all these scenarios, only shallow values considering a compositional boundary underestimate the heat flow. As we do not have depths of the order of basements underneath the sedimentary basin (~ 6 km), we may assume a low probability for this scenario. In the case of IXD, Slip rocks are relatively enriched in potassium, thorium and uranium (Klein et al., 2012), thereby increasing the radiogenic heat. This fact may also explain the shallow values. Our model yielded reasonable estimates for the heat flow on the surface, and the geologic constraints support our depths. Therefore, we understand that our DBML model can be used as a proxy for the depth to the Curie temperature. However, in the absence of heat flow data, we can only guarantee that we mapped Curie temperature surface variation rather than its absolute values.

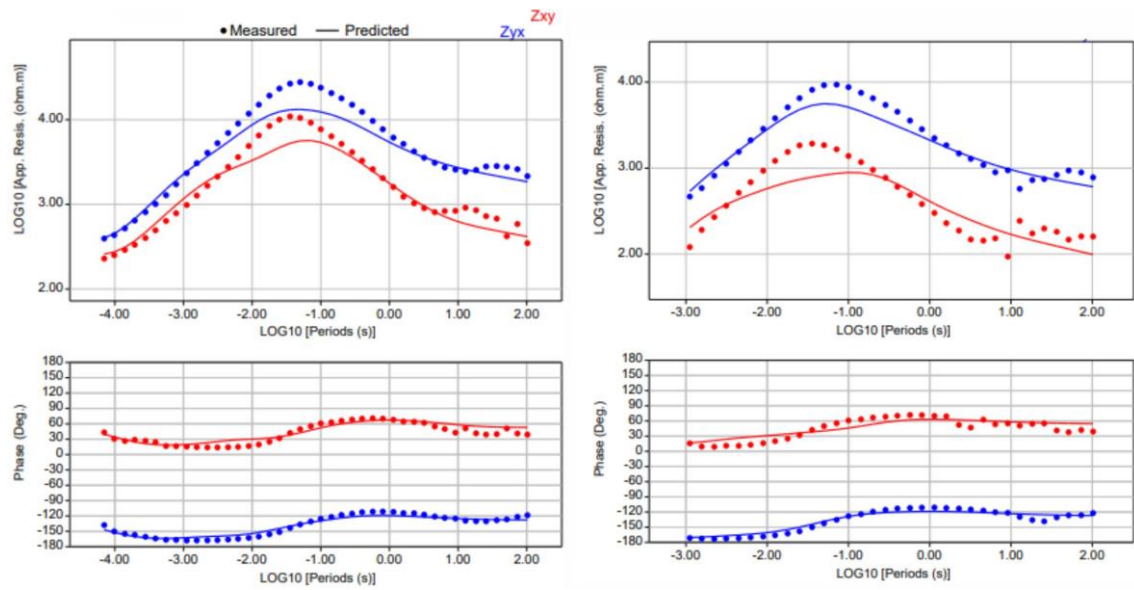
Apêndice B – Supporting Information for Tectonic and metallogenic implications of a metasomatized mantle inferred from magnetotellurics and gravity data: An example of Tapajós Mineral Province, Amazon Craton

Comparison of measured and predicted resistivity and phase curves for each station of the 3D preferred model

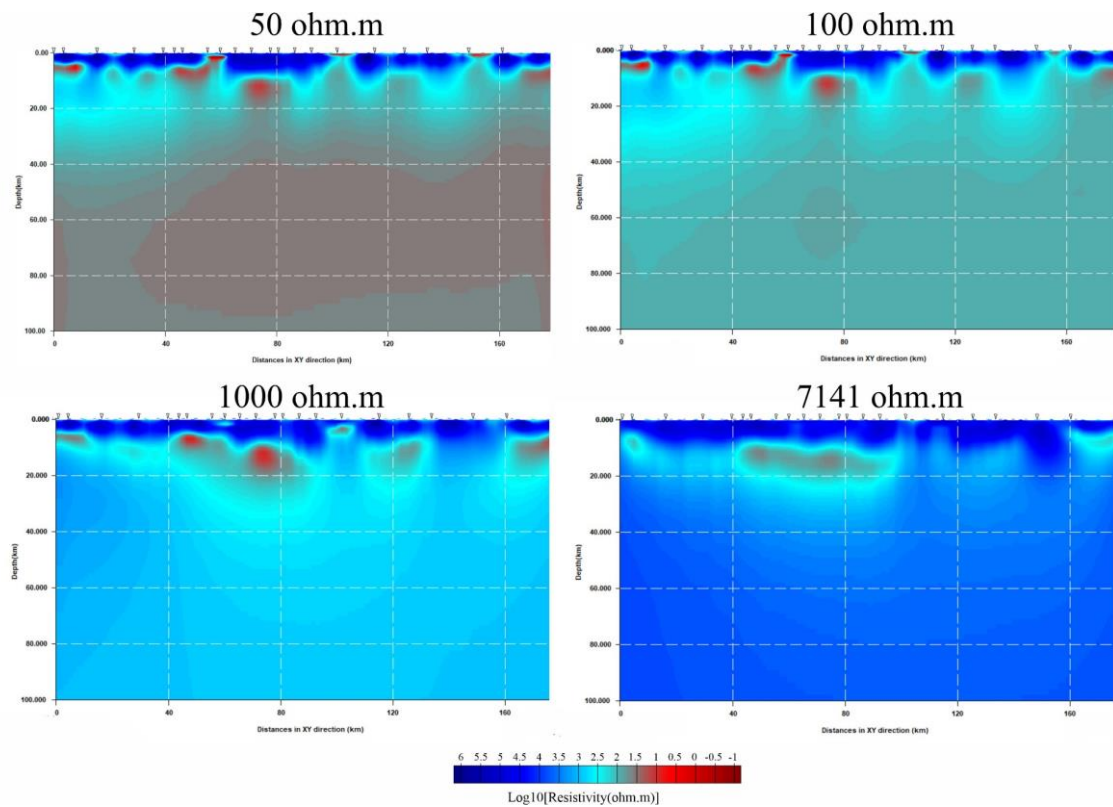








Inversion results for different prior models (homogeneous half-spaces).



Apêndice C – Supporting information for A multi-disciplinary geophysical approach reveals a Paleoproterozoic Cryptic Suture in the Amazon Craton – Implications for Proterozoic Supercontinent Reconstructions

Comparison of measured and predicted resistivity and phase curves for each station of the 3D preferred model.

