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Lithospheric thinning under the Araripe Basin (NE Brazil) from a long-period magnetotelluric survey: Constraints for Tectonic Inversion

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Abstract

The lithospheric architecture of the western Borborema Province and northern São Francisco craton of NE Brazil has been investigated through analysis of long-period magnetotelluric data acquired along a 700 km long survey, using 12 instruments. The survey samples several tectonic terrains in the Province and penetrates into the adjacent São Francisco craton after crossing the Araripe Basin, an aborted rift basin filled with Mesozoic sediments that peak at ~ 1000 m above mean sea level. High conductivities are observed at shallow depths under the main Precambrian shear zones that pervade the Province – consistent with tectonic reactivation – and as a

small patch embedded within the high resistivities that characterize the São Francisco craton. High conductivities ($\sim 25 \Omega m$) are also observed below 120 km depth between the Patos and Pernambuco Lineaments – right under the Araripe Basin – flanked by resistive ($> 120 \Omega m$) material immediately to the north and south. This deep, highly conductive body is found consistent with the presence of melt and aqueous fluids, and is interpreted as shallow asthenospheric mantle bounded by thicker lithosphere. We propose that extensional stresses in the Mesozoic stretched and thinned the lithosphere under the Araripe Basin, causing passive upwelling of asthenospheric material and lateral flow of the overlying lithosphere, and resulting in thickening of the lithosphere under the flanks and uplift of the Araripe Basin. We also hypothesize that thermal weakening of the lithospheric mantle – perhaps sustained by channeling of asthenospheric flows under the basin – would have caused regional stresses to concentrate in the brittle upper crust and contribute to basin inversion. We thus propose that a combination of localized horizontal stresses and vertical buoyancy from underlying asthenospheric material are ultimately responsible for the actual topography of the Araripe Basin.

Keywords: Araripe Basin, Borborema Province, Magnetotellurics, Conductivity, Lithosphere, Asthenosphere, Basin inversion, Partial melt

1. Introduction

The Borborema Province is the remnant of a larger Neoproterozoic mobile belt that is presently located between the São Francisco craton, the Parnaíba Basin and the coastal sedimentary basins of NE Brazil. It is generally regarded as composed of a series of Archean to Proterozoic tectonic blocks sep-

arated by shear zones that amalgamated during the Brasiliano/Pan-African orogeny (about 0.6 *Ga*) towards the end of the Neoproterozoic (Brito Neves, 1975). The end of the Pangea supercontinent, with the separation between Africa and South America, triggered the formation of a number of syn-rift basins in the Province, eventually resulting in the development of a number of coastal sedimentary basins along the continental margin, as well as several aborted rift basins within the continental interior (de Matos, 1992, 1999).

After continental break-up, the evolution of the Province was marked by episodes of intraplate volcanism and uplift. Volcanism consists of small volumes of alkaline rocks forming two orthogonal magmatic alignments: the Macau-Queimadas Alignment (MQA), on-shore and trending in an approximately NS direction; and the Fernando de Noronha-Mecejana Alignment (FNMA), mostly off-shore and trending in the EW direction. Radiometric $\text{Ar}^{40}/\text{Ar}^{39}$ dates range between 33 and 2 *Ma* for the FNMA and show a clear age progression towards the West; $\text{Ar}^{40}/\text{Ar}^{39}$ dates for the MQA, on the other hand, range between 93 and 7 *Ma*, with no clear age progression (Souza et al., 2003; Knesel et al., 2011). Both sets of volcanism had been attributed to mantle plume sources (Jardim de Sá et al., 1999; Mizusaki et al., 2002). However, the time overlap between the lines, their orthogonal arrangement, and the long-lived character support other origins such as a small-scale convection cell at the edge of the continent (Knesel et al., 2011) or lateral asthenospheric flows from a distant plume tail (Sleep, 2003; Simões Neto et al., 2017).

Intraplate uplift is best expressed through the high topography of the Borborema Plateau to the east and the Chapada do Araripe (Araripe Basin)

to the West. The Borborema Plateau is a NE-SW elongated topographic feature with maximum elevations of $\sim 1200\text{ m}$. Its elevation has traditionally been attributed to tectonic uplift resulting from either thermal doming (Ussami et al., 1999; Jardim de Sá et al., 1999) or isostatic uplift (Morais Neto et al., 2009; Oliveira and de Medeiros, 2012), although recent studies suggest it might be a high-standing, rheologically strong feature (Almeida et al., 2015; Luz et al., 2015; Simões Neto et al., 2017). The Chapada do Araripe, on the other hand, is elongated in the EW direction and can be described as an aborted rift basin that lies at $\sim 1000\text{ m}$ altitude. Its main development has been attributed to the breakup of Pangea (Assine, 2007; de Matos, 1999), and the proposed mechanisms for its actual elevation revolve around topographic inversion – in which regional uplift of the Province would have been followed by differential erosion (Peulvast and Bétard, 2015) – and basin inversion driven by EW compression from the Mid-Atlantic Ridge and Andean pushes (Marques et al., 2014).

The magnetotelluric (MT) method is a geophysical technique used for imaging the electrical conductivity of the Earth’s crust and upper mantle. It is a passive electromagnetic method that uses the propagation of the natural electric and magnetic fields through the conductive subsurface. Simultaneous measurement of electric and magnetic fields in orthogonal directions at the Earth’s surface allow inference of subsurface electrical conductivity variations from tens of meters to hundreds of kilometers depth (Chave and Jones, 2012). The electrical resistivity – or its reciprocal, the electrical conductivity – is the most variable of all geophysical rock properties; variations of eight orders of magnitude can be encountered in the Earth, and of several orders

of magnitude within the same rock type. The electrical resistivity of subsurface rocks typically varies between $10^{-2} \Omega m$ and $10^6 \Omega m$. The MT method is widely utilized to study the subsurface at crustal and lithospheric scales. Conductivity maps obtained by MT studies are used to detect interconnected fluids, discriminate between different geological formations, map-faulted areas, and detect lithospheric conductivity anomalies related to tectonic processes. These results, associated to geological observations and geophysical models, allow to image the structure of the lithosphere and to interpret them geodynamically.

Here, we present results from a long-period MT survey spanning for $\sim 700 km$ in the NS direction across the western Borborema Province and northern São Francisco craton of NE Brazil, the BODES² experiment. The MT installation sites were selected to be in close proximity to a deep, passive-source seismic deployment representing the seismic component of a multidisciplinary study led by the *Laboratório Sismológico* of the *Universidade Federal do Rio Grande do Norte* (LabSis/UFRN). The goal of the BODES experiment is to build integrated geophysical images of the Province's crust and lithospheric mantle in order to improve our understanding of its geodynamic and tectonic evolution.

²BODES: BOrborema Deep Electromagnetic and Seismic

2. Data acquisition and galvanic distortion analysis

2.1. Data acquisition and processing

In May and October of 2015, a total of 11 instruments were installed along a profile ranging from Pedra Branca in Ceará state to Ponto Novo in Bahia state (Figure 1). In 2017 the MT profile was extended with two additional stations, resulting in a total length of about 700 km and a final station spacing of roughly 60 km. Given the decay of lateral sensitivity, we expect a poor constraint of any of the structures of the upper crust other than the ones located directly underneath the instruments. We used long-period MT instruments (LEMI), borrowed from the *Pool de Equipamentos Geofísicos do Brasil* (PEGBR), together with non-polarizable Cu-CuSO₄ electrodes. The instruments recorded between 21 and 34 days at a sampling rate of 1 Hz. The registered electromagnetic time series were in general of very good quality, only displaying noisy sections at a few sites. Northeast Brazil is extremely dry and although precautions were taken to prevent electrodes from drying out, a few days of data were still lost at some locations. Time series sections affected by electrode problems were disregarded from further analysis.

In MT, the processing stage consists of the calculation of the electromagnetic transfer functions or responses by means of robust regression algorithms. The electric E and magnetic H fields are related through the impedance tensor Z in the frequency domain, according to:

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}, \quad (1)$$

and the tipper vector is the transfer function between the vertical and hori-

zontal magnetic field, expressed in the frequency domain as:

$$H_z = \begin{pmatrix} A & B \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}. \quad (2)$$

The tipper and impedances are the MT transfer functions or responses. In particular the impedance tensor is expressed as amplitudes (apparent resistivities, ρ_{ij}) and phase (ϕ_{ij}):

$$\rho_{ij} = \frac{1}{\omega\mu} |Z_{ij}|^2, \quad \phi_{ij} = \tan^{-1} \left(\frac{\Im Z_{ij}}{\Re Z_{ij}} \right), \quad (3)$$

where ω is angular frequency, μ is magnetic permeability, and the subindexes i and j indicate the impedance tensor elements.

To process the MT time series we used the multi-remote reference *EMT* code (Neukirch and Garcia, 2014). This code is particularly useful when non-stationary noise is present, and a few of the sites were particularly noisy. The remote references were chosen to be stations that recorded simultaneously. In general, the responses were found to be of good quality between 8 s and 20,000 s. Some sites were affected by low solar signal, which impacted mainly the longer periods. In addition, the electrodes used to estimate the electric field are known to decay performance for long-term deployments and, for this reason, estimates beyond 20,000 s were deemed unreliable. Responses were rotated during this initial processing stage to the geographic north reference system from the magnetic north acquisition convention. Site B11 was occupied 22 km northwest of site B01 because the latter was severely affected by strong distortion effects. After processing, site B11 responses were nonetheless similar to B01 and noisy; therefore site B11 was discarded and B01 was finally used in the modeling stage in order to keep a more regular distance between sites during the inversion process.

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136 In the MT method, the skin depth estimation has traditionally been used
137 to assess the depth of penetration of a survey. Using the skin depth esti-
138 mation and the responses from the BODES experiment, the period range
139 covered in this survey is sufficient to image the lithosphere. Given the good
140 lateral resolution of the MT method (Jones, 1987), our station spacing allows
141 to cover lateral changes in resistivity on a regional scale.

142 *2.2. Galvanic Distortion Analysis*

143 Subsurface heterogeneities below the scale of our experiment can cause
144 galvanic distortion (Jones, 2012), which can hinder the interpretation of the
145 modeled responses. Several algorithms are available to correct for galvanic
146 distortion and at the same time estimate the electric strike angle of the
147 structures. To analyze the BODES dataset we utilized the gadget algorithm
148 (Neukirch et al., 2017), which does not assume any *a priori* dimensionality
149 about the data. The gadget algorithm decomposes the impedance tensor into
150 the well-known phase tensor (Caldwell et al., 2004) and the recently intro-
151 duced Amplitude tensor (Neukirch et al., 2017). The later is used to recover
152 the galvanic distortion parameters as originally parameterized by Groom and
153 Bailey (1989). Table (1) contains the estimated distortion parameters of the
154 BODES dataset; in general, high distortion parameters (large twist and shear
155 angles) can be appreciated, which are indicative of strong distortion. The
156 strongest galvanic anisotropic distortion is found to the north of the Araripe
157 Basin, and corresponds to the location of the Patos and Jaguaribe-Tatajuba
158 shear zones, together with an array of local-scale faults (Peulvast and Bétard,
159 2015). The elliptic shape of the amplitude tensor parameters is an indication
160 of electrical anisotropy, where the strongest the anisotropy the more elliptical

the shape becomes (Neukirch et al., 2017).

The amplitude, phase tensor electrical strike angles and induction arrows (real part) are displayed in Figure (3) for different increasing periods (increasing depth of penetration). This figure shows that even at the longest periods shown here (corresponding to lithospheric depths) there is no common electric strike angle for all periods and all sites. The plot also shows the main shear zones in the study area. Considering the 90 degrees ambiguity in MT when dealing with the strike angle, the correlation of the phase tensor strike angle with the shear zones (red lines) for shorter periods becomes apparent. The galvanic distortion and strike angle analysis reflect the complex geological and tectonic history of the Borborema Province, which is crisscrossed by a network of deep shear zones that likely extend down into the lithospheric mantle (Tommasi et al., 1995). Large distortion effects are also observed south of the profile, mostly at site B08, and towards the end of the profile, coincident with the surface expression of a NS trending shear that marks the western edge of the Chapada Diamantina (within the São Francisco craton). The induction arrows point towards areas of enhanced conductivity. For shorter periods, these arrows focus on local structures, while for longer periods, they point out towards the East. Induction arrows pointing outside the profile, confirms that the geo-electrical structure under the western Borborema Province is 3D.

3. MT data modeling

3.1. Inversion approach

To develop the model that best explains the observations, we employed the *ModEM* 3D inversion code (Kelbert et al., 2014). This code uses a parallelized forward iterative solver on a finite difference grid and a non-linear conjugate gradient algorithm for the inversion.

We considered the full impedance tensor and tipper during the inversion process for our 12 sites, and the responses were decimated to 5 frequencies per decade – from 9 s to 20,000 s. We assumed an error threshold of 3% for impedances, and a 0.03 absolute error for the tipper. We also masked obvious spurious responses before inversion, so they were not taken into account. The mesh size was $155 \times 47 \times 49$ blocks, with 5 km size blocks around the center of the model. Any of the inversion calculations took between 3 to 4 days on a 16 core cluster node, involving some 150 iterations.

During the inversion process, we experimented with a number of homogeneous (1D) starting models, each with different values of resistivity. The results showed a very strong dependence on the starting models, which we attribute to the data being acquired along a linear profile. Our preferred results were obtained with a starting model consisting of a $200 \Omega m$ half-space overlaying a $50 \Omega m$ bottom layer at 130 km depth. Most importantly, we had to account for bathymetry in our inversions due to the close proximity of the Atlantic Ocean on the east and north sides of the model (see plot in Figure 2). For this, we used the GEBCO database³. Attempts to invert the

³The GEBCO_2014 Grid, version 20150318, <http://www.gebco.net>

data without including the ocean failed to reach a reasonable fit, yielding RMS misfits larger than 4.5. Additionally, to avoid introducing artifacts, we started by inverting the shorter period data and, after reaching a misfit RMS of 2 or lower, we proceeded to invert the full period-range of the MT response. We tried different approaches and starting models to invert this dataset (see the Supplementary Material section for a summary of our efforts). Our preferred model was chosen because it contained more realistic features, which could be interpreted in a geological manner.

The misfit of our preferred inverted model is 2.23 (see individual site fits in the Supplementary Material section), which fits the apparent resistivities, phases, and tipper vector reasonably well. We checked the robustness of the final model by removing specific features (changing resistivities) and numerically assessing whether they are required to fit the data. Our sensitivity tests indicate that all structures present in our preferred model are required, and that the distribution of electrical resistivity in the final model is robust.

3.2. Modeling results

Figure (4) displays the best model obtained after inversion of the MT observations. The 3D block has been cut along the surface profile to show the most relevant structures sensed by our data. Blue, cold colors indicate resistive structures, while red, warm colors signal conductive structures. The most striking feature is the deep conductor starting at about 120 km depth under the Araripe Basin, between the Patos and Pernambuco lineaments (C1 in Figure 4). Its resistivity is less than $30 \Omega m$, and both the location and resistivity are very robust features in the inverted model.

To the north and south of conductor C1, and also at lithospheric depths,

we find a number of resistive structures (R1, R2 and R3 in Figure 4). Structure R1 is located in the northern tip of the profile and reaches depths of 200 *km*. To verify that this structure is not a border effect, we recorded an additional site in Spring 2017 (site B10) more to the north. This additional observation confirmed that the thick, resistive structure is robust and not an artifact of the inversion process. Interestingly, the location of the resistive structure R1 correlates with the presence of the exposed Archean rocks (Figure 2) of the Troiá massif (Fetter et al., 2000). Structure R2, immediately south of the Araripe Basin, is also imaged as a thick, resistive structure. And, to the south, coinciding with the location of the São Francisco craton, the model also shows a thick, resistive structure (R3), reaching more than 180 *km* depth.

At shallower, crustal levels a number of conductive structures can be identified. The conductive structure C2 (Figure 4) starting at 10 *km* depth under site B08 is located on the limit between the Riacho do Pontal fold belt and the Chapada Diamantina (northwest of the São Francisco craton). Between sites B01 and B02, there is a strong conductive anomaly, although we do not have the coverage to interpret it. This conductor correlates with the surface expression of the Jaguaribe-Tatajuba lineament, and was also reported in an independent MT profile located east of the BODES profile (Padilha et al., 2014, profile A-A'). Conductor C3, under site B03, correlates with the Patos lineament and, as appreciated in Figure (4), is marked as a nearly horizontal upper crustal structure. Finally, conductor C4 is located under the Riacho do Pontal fold belt.

4. Discussion

4.1. Shear Zone reactivation, fluids and mineralization

A number of conductive bodies characterize the shallow (crustal) structure along the MT section displayed in Figure 4. Conductor anomalies C2 and C4 are located on the northern boundary of the São Francisco craton, south and north, respectively, of the contact between the Cariris Velhos belt and the Riacho do Pontal fold belt. The Riacho do Pontal fold belt has been described as a continental-scale (1000 km) accretion belt that assembled during the Brazilian orogeny ($\sim 600 Ma$), together with the Rio Preto and the Sergipano fold belts. Moreover, the accretion process would have been marked by south-bound subduction north of the Cariris Velhos (Caxito et al., 2014), and north-bound subduction to the south (Caxito et al., 2016). Indeed, the Riacho do Pontal fold belt presently contains the remains of Neoproterozoic oceanic crust. Conductors associated to subducting zones have been previously reported in the Borborema Province (Padilha et al., 2017), and interpreted as caused by the presence of metals in folds or interconnected graphite. Given the presence of sulphide mineralization and graphite in the Riacho do Pontal area (Caxito et al., 2014), we thus interpret both conductors C2 and C4 as caused by either conductive minerals or graphite deposited by subducting oceanic lithosphere along this former passive margin.

Conductor C3 (Figure 4), which coincides with the location of two major shear zones in the Borborema Province, is interpreted as the extension of those shear zones deep into the crust. This implies that shear zones in the Borborema Province have been reactivated through time and have played a role in the accommodation of tectonic stresses. Indeed, the network of

Precambrian shear zones has been proposed to have exerted key controls in the development of Mesozoic rift basins (de Matos, 1999) and during the episodes of tectonic uplift in the Araripe Basin (Marques et al., 2014) and the Borborema Plateau (Oliveira and de Medeiros, 2012). The origin of the conductive anomaly could then be the result of the percolation of meteoritic fluids at shallower depths, and the presence of conductive minerals and graphite precipitated by fluid flow.

4.2. Shallow asthenosphere under the Araripe Basin

Conductor C1 is the most prominent feature in the resistivity cross-section displayed in Figure 4. This conductive anomaly is coincident with a low velocity anomaly reported in a recent tomographic study (Simões Neto et al., 2017) that runs in the EW direction under the Araripe Basin at sub-lithospheric depths. Taken at face value, it suggests the presence of asthenospheric material at depths as shallow as 120 km, and the presence of thin lithosphere under the Araripe Basin. Moreover, this conductive structure is located close to a marked heat flow anomaly (Davies, 2013) of 124 mW/m² (median value) on the eastern edge of the Araripe Basin. Although the heat flow map resolution for South America is on the order of 0.5°, the overall increase in heat flow would be consistent with asthenospheric material placed at 120 km depth. Alternatively, the alternative heat flow map of South America of (Hamza et al., 2005) does not display such a marked anomaly; however, it shows a positive anomaly that coincides with the northern extension of our profile. In any case, a thermal anomaly could also be caused by fluid release or radiogenic material within the crust.

The top of the asthenosphere at 120 km underneath the Araripe Basin,

implies that the adiabatic temperature at that depth must be around 1350°C (e.g., Till et al., 2010). Assuming a pressure at 4 GPa , and considering the modeled resistivity of the top of the anomaly is $25\Omega\text{m}$, thermobarometric models can be used to obtain resistivities for different mineral phases. For dry olivine – and therefore only assuming thermal effects – the SO3 model of Constable (2006) would yield a temperature of 1820°C for the resistivity observed. This is obviously too hot for an asthenosphere under a continent at 120 km depth for which no obvious rifting process is presently at play.

The addition of water and/or partial melt, however, could help reconcile the observed resistivity with a shallow asthenosphere, as both fluid and melt will increase conductivity. It has been suggested that the upper mantle and in particular the asthenosphere, contains between 50 to 200 ppm of water (Green et al., 2010), which can cause low-temperature partial melting of the lithospheric mantle (Mierdel et al., 2007). In addition, some studies have suggested that the asthenosphere contains a small amount of partial melt, and recent work by Mancinelli et al. (2017) suggests that either rheological or melt gradients must be present at the lithosphere-asthenosphere boundary in order to explain the gradient of seismic velocities observed. Moreover, it has been recently suggested that mantle resistivity can reach $30\Omega\text{m}$ at asthenospheric depths (Dai and Karato, 2009) by adding water, but it cannot be lowered below this value without adding so much water that melting is initiated (Ni et al., 2011).

Without further constraints, we consider that melt and water are responsible for the anomaly, and provide upper bounds for the amount of these fluids necessary to explain the observed anomaly (C1). Considering the rela-

329 tionship of (Ni et al., 2011, eq. 3) for the presence of water and melt phases
 330 for basaltic rocks under asthenospheric conditions, the upper bound Hashin-
 331 Shtrikman (Hashin and Shtrikman, 1962) relationship and the constraints
 332 that we have for anomaly C1, we investigate the possible relationship be-
 333 tween water amount and melt fraction for two host mineral resistivities (Fig.
 334 5). A similar technique was used by Sarafian et al. (2015) to quantify the
 335 amount of melt and water in the central Pacific asthenosphere as part of the
 336 NoMelt project. The partition coefficient between mantle and melt is set to
 337 $D_{mantle-melt} = 0.006$ Hirschmann et al. (2009); Ni et al. (2011). For the bulk
 338 value of the mantle resistivity (σ_1 in eq. 3 in Ni et al. (2011)), we used a peri-
 339 dotite model (Sarafian et al., 2015) consisting of 60% *Ol* (Wang et al., 2006),
 340 35% *Opx* (Dai and Karato, 2014) and 15% *Cpx* (Zhao and Yoshino, 2016),
 341 hydrated by water left after partitioning with melt. By using a geometric
 342 average (Shankland and Duba, 1990) we estimate the overall conductivity
 343 of this peridotitic mantle. As mentioned earlier, for the final melt-hydrated
 344 periodotite, we use a HS+ model using the conductivities of peridotite and
 345 melt at each value of melt and water percentage. For the suggested values of
 346 water content in the upper mantle (50 to 200 ppm), we obtain that the melt
 347 fraction at 120 km depth under the Araripe Basin should be around 0.3 %,
 348 and 0.2% melt for an enriched mantle with 800 ppm of water. These are
 349 reasonable amounts of water and melt amount in an asthenospheric mantle
 350 (Sifré et al., 2014).

351 What is then the origin of the C1 anomaly in the shallow asthenosphere
 352 under the Araripe Basin? Mesozoic extensional events were certainly re-
 353 sponsible for initial thinning of the lithosphere and passive upwelling of the

underlying asthenosphere. Why the shallow asthenosphere would have been preserved until today, however, requires further explanation. In general, the continental lithosphere is viewed as a dry environment, so hydration from melting of the overlying lithosphere seems unlikely. It has been suggested (Mareschal et al., 1995; Chen et al., 2009) that relict metasomatism can be preserved over billions of years in the cratonic lithosphere; however, that these metasomatic materials can provide enough fluids as the ones required by our model to keep the lithosphere thin since the Mesozoic seems more unlikely. Because of their capacity to bring fluids and other incompatible elements, plumes are very effective metasomatizing agents (e.g., Baker et al., 1998), in addition to bringing carbonitic melts that enhance conductivity, but there is no evidence of a plume recently passing under the Borborema province (e.g. Storey, 1995; Schimmel et al., 2003; Knesel et al., 2011; Pinheiro and Julià, 2014) since the Mesozoic. Alternatively, olivine, with a typical value of 50 *ppm* of water storage capability (Dixon et al., 2002), could match our inferred water content; but somehow the asthenosphere would need to be preserved from freezing and thickening the lithosphere under Araripe.

At this point, we can only conjecture that lateral flow of hot, molten material could be responsible for preserving the shallow asthenosphere under the Araripe Basin. Rifting events during the Mesozoic in the Borborema Province might have created the path necessary for the melt to travel under the lithosphere. As mentioned earlier, seismic tomographic imaging of the Borborema Province's upper mantle has revealed a NE-SW trending low velocity channel – with an EW trending offshoot right under the Araripe Basin

– that is consistent with lateral asthenospheric flow (Simões Neto et al., 2017). Due to proximity, a mantle plume under the Fernando de Noronha hotspot would be the most obvious candidate for feeding the postulated lateral flow under the Province (FNMA in Figure 2). Alternatively, it had been proposed that all intraplate volcanism in the Borborema Province younger than 85 Ma is the result of lateral flow from a distant plume tail presently located under the Paraná basin (Sleep, 2003). Although lateral asthenospheric flow around the São Francisco craton has been invoked to explain seismic anisotropy patterns from SKS splitting observations (Assumpção et al., 2006), a clear connection between an active Paraná plume and the Borborema low-velocity asthenospheric channel would need to be established.

4.3. Basin inversion and ductile flow

A number of deep resistive anomalies to the north and south of the profile suggest the presence of dry, crystalline rocks in a thick lithosphere, typical of cratons. To the south (R3 in Figure 4) the anomaly is expected, as it would correspond to the São Francisco craton, and resistivities larger than $500 \Omega m$ are consistent with depleted, crystalline mantle rocks. However, resistive anomalies R1, at the northern tip of the BODES profile and R2, to the south of the Riacho do Pontal fold belt, are harder to explain.

Previous MT studies, nonetheless, had already shown a thick, resistive structure (Padilha et al., 2014, profile B-B') down to $100 km$ depth in the northwestern Borborema region. Although the authors focused on interpreting anomalies at shallow crustal levels, they did interpreted this enhanced electrical resistivity in the upper mantle to be the relic of dehydrated, sediment-depleted oceanic lithosphere that subducted during the Neoproterozo-

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404 zoic. How this structure could have survived the vigorous thermal history of
405 the NW Borborema Province is unclear; fortunately, the deeper constraints
406 presented here allow for a more plausible interpretation of this enhanced
407 resistivity.

408 Numerical modeling has suggested (Liu and Shen, 1998) that passive up-
409 welling of the asthenosphere under the Colorado Plateau may have thinned
410 the overlying lithosphere and thickened the flanking lithosphere under the
411 Sierra Nevada through a roll down process. Similarly, mantle upwelling un-
412 der the Araripe Basin might have developed the necessary lateral pressure
413 gradients in the lithospheric mantle to push the overlying lithosphere side-
414 ways and towards the flanks, explaining the thickened roots immediately
415 north and south of conductor C1 (features R1 and R2 in Figure 4).

416 Additionally, the lithospheric erosion under the basin implied in the roll-
417 down model could provide a plausible mechanism to explain the tectonic
418 inversion of the Araripe Basin. It has been proposed that tectonic inversion
419 of the Araripe Basin was due solely to compressional stresses developed from
420 the pushes of the Mid-Atlantic Ridge and the Andean front (Marques et al.,
421 2014). The problem with the model proposed by Marques et al. (2014) is
422 that the high-angle faults bounding the basin make it difficult to think of
423 pure horizontal compression as a viable uplift mechanism. Interestingly, it
424 has been observed (Assumpção et al., 2004) that intraplate seismicity oc-
425 curs preferentially on crust above thin lithosphere, suggesting that stresses
426 concentrate above lithospheric "thin spots". If our interpretation is correct,
427 the lithosphere is thin under Araripe Basin, vertical buoyancy provided by
428 the hot, underlying asthenosphere coupled with compressional stress concen-

tration in the overlying thin lithosphere could provide additional sources of uplift to make the tectonic inversion of the Araripe Basin viable.

The geodynamic framework proposed in this work to explain the uplift of the Araripe Basin is summarized in Figure (6).

5. Conclusions

We have interpreted the high conductivities under the Araripe Basin as an indication of shallow, enriched asthenospheric material, involving small amounts of fluids and melt. We have proposed that the emplacement of a shallow asthenosphere would have originated through extension and passive upwelling at the time of basin formation, and explained its preservation through geologic time by invoking lateral asthenospheric flow from a distant source.

We have proposed that the hot, buoyant asthenospheric material under the basin provided a vertical push under the Araripe Basin that helped the tectonic inversion of the basin. Additionally, the presence of this asthenospheric material under the Araripe Basin - and the correspondingly raised geotherm - might have weakened the lithospheric mantle, caused regional compressional stresses to concentrate on the brittle crust, and contribute to basin uplift.

The high resistivities flanking the conductive anomaly north and south of the Araripe Basin have been interpreted as thickened lithosphere representing lithospheric material that was eroded from under the basin and attached to the flanks through a roll-down flow process.

Shear zones have been mapped as conductive anomalies crossing the crust

in the Borborema region, suggesting the presence of conductive minerals or graphite, precipitated by fluids flow and shear zone reactivation.

To the south of the profile, the Riacho do Pontal fold belt is observed to be flanked by conductive structures, which have been interpreted as relict subduction material.

6. Data Availability

The electromagnetic time series recorded during the BODES experiment, including calibration files and field notes for processing, will be available on the data repository of the PEGBR (<http://www.pegbr.on.br>) or upon request to the main author.

7. Acknowledgments

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8. Figures

Figure 1. Topographic map of NE Brazil showing the location of the long period MT stations of the BODES experiment. Black line indicates the Borborema Province. The Araripe Basin is indicated as well with a black contour line. Red lines indicate shear zones. Site B11 was not used because of its proximity to B01. Inset plot (upper right corner) shows the location of the topographic map (red) and the extension of the 3D MT inversion model (green).

Figure 2. Geologic map of NE Brazil showing basement rock ages, main tectonic features and the location of the long period MT stations of the BODES experiment. The gray line indicates the contour of the Borborema Province. BP: Borborema Province. AB: Araripe Basin; CSF: Craton São Francisco; RdP: Riacho do Pontal fold belt; TM: Troià massif. Shear zones are labeled in italics. Blue arrows indicate the motion of the volcanic lineaments. Lithological map from the Geological Survey of Brazil – CPRM.

Figure 3. Phase tensor and induction arrows (real part) for the BODES MT responses using the Gadget algorithm at different nominal periods. The phase decomposition results with black line indicating the best strike angle, and the colors indicate the phase values. Red lines indicate mapped shear zones.

Figure 4. Slice through the MT sites of the 3D inversion model. Note that the warm colors indicate the presence of conductors, while cold colors indicate the presence of resistors. Yellow labels inside the model are referenced in text. The projection use for the modeling is a UTM, ellipsoid WGS84, zone 24M. BP: Borborema Province; SFC: São Francisco Craton; RTJ: Recôncavo-

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694 Tucano-Jatobá Basin; A: Araripe Basin; RdP: Riacho do Pontal fold belt;
695 TM: Troiá massif; PSZ: Pernambuco shear zone; PAZ: Patos shear zone;
696 SPFZ: Senador Pompeu shear zone; TF: Tatajuba fault.

697 Figure 5. Plot of the electrical conductivity as a function of water and
698 melt (Ni et al., 2011), for a mantle initially composed of 60% Ol, 25% Opx and
699 15% Cpx as suggested by (Sarafian et al., 2015). The partitioning coefficient
700 between mantle and melt has been fixed at $D_{mantle-melt} = 0.009$ (Hirschmann
701 et al., 2009) and temperature at $1350^\circ C$. Conductivity contours at specific
702 values are shown with a white, dashed line. The conductivity contour for the
703 anomaly considered in this work, $25 \Omega m$ ($0.4 S/m$), is shown as a black line.

704 Figure 6. Cartoon representing a geological interpretation of the MT
705 profile across the Borborema Basin. Dashed lines indicate the thickening of
706 the lithosphere north and south of Araripe by material attached from the
707 roll-down process.

Table 1: This table contains the recovered distortion parameters, twist, shear and anisotropy angles, for the BODES data set.

Site	Twist [deg]	Shear [deg]	Anisotropy [deg]
B10	-10.69 ± 12.67	0.52 ± 13.62	-15.65 ± 3.61
NBPB	12.32 ± 0.03	-30.70 ± 0.02	-2.65 ± 0.11
B01	-63.37 ± 0.06	-38.48 ± 0.05	-29.00 ± 0.24
B02	-44.03 ± 2.51	4.53 ± 2.83	35.19 ± 0.26
B03	18.82 ± 0.04	-0.57 ± 0.08	34.71 ± 0.02
B04	-4.96 ± 0.07	-11.76 ± 0.04	-6.69 ± 0.08
B05	7.69 ± 0.04	-26.39 ± 0.03	15.82 ± 0.31
B06	-18.23 ± 0.03	7.64 ± 0.04	19.34 ± 0.02
B08	-87.30 ± 0.03	7.80 ± 0.01	-3.84 ± 0.02
B07	16.03 ± 0.20	-35.26 ± 0.45	21.94 ± 0.17
B09	-36.42 ± 0.05	-30.72 ± 0.03	-21.96 ± 0.06
NBPN	-24.03 ± 0.06	-38.90 ± 0.02	-8.39 ± 0.18

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