Upper-mantle structure of the Borborema Province, NE Brazil, from *P*-wave tomography: implications for rheology and volcanism

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SUMMARY

We have performed a tomographic study of the upper mantle under the Borborema Province by inverting relative *P*-wave traveltimes residuals (including the *PKP_{df}* phase). The events were recorded by more than 50 broad-band and short-period stations deployed across the region, resulting in a total data set of 1912 relative residuals. A multichannel phase cross-correlation technique was utilized to develop the measurements, which were then inverted through an iterative, non-linear inversion scheme based on the subspace inversion method. The Fast Marching Method, a grid-based numerical algorithm that traces an interface evolving over a narrow band of nodes that are updated solving the eikonal equation by finite differences using upwind entropy, was considered to solve the forward problem. Traveltimes from outside of the 3-D model were calculated using a global reference model. The resulting tomographic images reveal structure down to depths of ~600 km, under an area extending approximately 800 km in the EW direction and 900 km in the NS direction. The most important features revealed by the tomographic images include: (i) a relatively fast lithospheric mantle under the southern Borborema Plateau, when compared to the lithospheric mantle north of the Patos Lineament; (ii) a marked, shallow (< 150 km) low-velocity anomaly under the northeastern most corner of the Borborema Province; and (iii) a low-velocity channel bordering the Borborema Plateau at asthenospheric (250–400 km) depths. The lithospheric velocity contrast is interpreted as arising from a colder, stronger lithosphere south of the lineament, while the asthenospheric low-velocity channel is interpreted as resulting from lateral flow from a distant mantle plume (located in SE Brazil). We argue that the rheological contrast validates a stretching model recently proposed to explain the Plateau's elevated topography, and that the postulated lateral flow represents the source of magmas feeding intraplate volcanism in NE Brazil for the past ~80 Ma.

Key words: South America; Body waves; Seismic tomography; Intra-plate processes.

1 INTRODUCTION

The study of the Earth's upper mantle is important for determining the tectonic, thermal and geodynamic evolution of the continents, as well as for improving our understanding of the asthenospheric mantle flow right beneath them. The theory of plate tectonics, established for over 30 yr as the paradigm against which tectonic processes must be benchmarked, has been successful in explaining a variety of geologic phenomena such as the occurrence of earthquakes and volcanoes, and the formation of mountain belts around plate margins. Most intraplate phenomena, such as the formation of basins and plateaus in the continental interiors or the origin of intraplate volcanism, on the other hand, are still a matter of debate.

A prime example within stable South America is the Paraná basin of SE Brazil, for which the stratigraphic record suggests up to six different episodes of sedimentation and erosion that include the eruption of extensive flood basalts (e.g. Milani & Ramos 1998). In the Brazilian Northeast, an active debate revolves around the origin of (postulated) tectonic uplift of the Borborema Plateau and the 'coeval' Cenozoic volcanism found along the Macau-Queimadas Alignment (MQA).

The MQA can be described as an approximately north–south trending alignment of alkali volcanic rocks that erupted in Mesozoic times (Fig. 1). The alignment is characterized as small volume and long lived (93–7 Ma), lacks a clear age progression, and has a time overlap with the east–west trending Fernando de

Figure 1. Topographic map of the Borborema Province with the main geological features highlighted. Red dots indicate the location of the Cenozoic volcanic bodies, and the dashed lines mark the trend of the corresponding magmatic alignments (Macau-Queimadas and Fernando de Noronha-Mecejana). The purple pentagons indicate the location of mesas covered by the Serra dos Martins formation and the white pentagons the location of AFTA samples from Morais Neto *et al.* (2009). The cooling ages are indicated in the inset. The boundaries of the Borborema Plateau are marked through a solid blue line (Adapted from Almeida *et al.* 2015).

Noronha-Mecejana alignment (FNMA) offshore (see e.g. Mizusaki *et al.* 2002). The Borborema Plateau, on the other hand, is generally introduced as an east–west oriented topographic feature with average elevations of ~1000 m a.m.s.l. (Morais Neto *et al.* 2009; Oliveira & Medeiros 2012). Its Cenozoic uplift has been postulated from the presence of sediments of the Serra dos Martins formation on top of high-altitude mesas, for which fission track analysis place a maximum deposition age in the Palaeogene (Oliveira & Medeiros 2012). Although some authors have challenged the postulated time of uplift (e.g. Morais Neto *et al.* 2009), it is generally accepted that there is a time overlap with MQA volcanism.

A commonly accepted model for the origin of volcanism along the MQA is based on the postulated existence of a small-scale convection cell under the northeastern-most corner of the continent (Kneisel *et al.* 2011). According to this model, the small-convection cell would have eroded the lithosphere immediately above, triggering melts along its downwelling edge, and feeding the surface volcanics that now form the MQA alignment. The same small-convection cell was also invoked by Oliveira & Medeiros (2012) to explain plateau uplift. According to those authors, some of the melts produced by the small-convection cell would have ponded under the plateau's crust and formed a layer of mafic cumulates that isostatically elevated the surface above. Interestingly, this model has been recently challenged

by Luz *et al.* (2015a,b), for whom the Borborema Plateau should rather be regarded as a high-standing feature unrelated to MQA volcanism. The newly proposed framework is based on a strong rheology for the lithosphere making up the elevated, thick-crust regions of the southern Borborema Plateau, in contrast to a weaker rheology for the lithosphere making up the low-lying, thin-crust regions of the surrounding Sertaneja Depression. Extensional processes in the Mesozoic would have stretched and thinned the rheologically weaker lithosphere, which would have subsequently subsided and formed the Sertaneja Depression, leaving the rheologically stronger lithosphere under the Borborema Plateau relatively undeformed and at a higher elevation. The postulated rheological contrast is largely based on an interpreted mid-crustal detachment zone—that would have developed during crustal stretching—observed in regions of thin crust but missing in regions of elevated, thick crust (Almeida *et al.* 2015). The rheological contrast is expected to hold down to the bottom of the lithosphere, so further evidence for the strong rheology of the southern Plateau on a lithospheric scale is required to confirm the newly proposed tectonic framework for the region.

One of the main tools for exploring the upper mantle is seismic tomography (e.g. VanDecar 1991; Ritsema *et al.* 1998; Schimmel *et al.* 2003; Bastow *et al.* 2008; Rocha *et al.* 2010; Hansen *et al.* 2012), in which velocity anomalies are mapped at various depths from inversion of relative seismic traveltimes measured at seismic stations. Variations in seismic velocity can result from changes in chemical composition, temperature, partial melts, and/or fluids (Deschamps *et al.* 2002; Artemieva *et al.* 2004); luckily, Artemieva *et al.* (2004) demonstrate that—at least at lithospheric depths—temperature is the main factor influencing *S*-velocity variation throughout the Brazilian Northeast. Moreover, commonly accepted compositions for the mantle seem to have little effect on velocity anomaly above 400 km depth, making composition of secondary importance with respect to temperature (Cammarano *et al.* 2003). Although Afonso *et al.* (2010) warn that seismological studies alone cannot separate compositional from thermal anomalies, and that the ratio of density to shear wave should be used instead in a periodotitic upper mantle, velocity anomalies continue to be interpreted in terms of temperature variations (e.g. Goes *et al.* 2000; James *et al.* 2004; Goes & Yoshizawa 2005) and further utilized to infer lithospheric strength (Poupinet *et al.* 2003; Bastow *et al.* 2005). Typically, seismic sources located at teleseismic distances from the recording stations are considered in seismic tomography studies, resulting in a near-vertical sampling of the uppermost mantle and crust. Near-vertical incidence results in an excellent lateral resolution of the velocity variations within the tomographic volume, but introduces vertical (direction of the ray) smearing of the anomalies making the assessment of the depth extent of anomalies difficult. Seismic transmission tomography does not have the power either to resolve seismic discontinuities. In spite of all these limitations, seismic tomography remains the prime tool for imaging the Earth's upper mantle on a regional scale.

In this paper, we developed tomographic images for the upper mantle under the Borborema Province of NE Brazil from the inversion of relative *P*-wave traveltime residuals measured at seismic stations in the region. The relative residuals were developed following the multichannel cross-correlation approach of VanDecar & Crosson (1990), to be then inverted for slowness perturbations and station correction terms following the approach of Rawlinson *et al.* (2010). We considered all available data recorded until 2015 December at broad-band and short-period stations in the region, resulting in a total data set with 1912 measurements for a combined network of 58 stations unevenly spread within the study area.

Our results display a marked velocity contrast at about 7°S latitude that separates slow velocity anomalies (-0.10 km s^{-1}) to the north from fast velocity anomalies to the south (0.12 km s^{-1}). The boundary closely coincides with a geologically mapped Precambrian shear zone—the Patos Lineament—which also separates thin from thick crustal types in the Borborema Province (Almeida *et al.* 2015; Luz *et al.* 2015a,b). Moreover, the anomaly can be traced down to depths over 200 km, revealing it as a lithospheric-scale feature. At shallow depths ($\sim 100 \text{ km}$), the tomographic slices reveal a large low-velocity anomaly (-0.16 km s^{-1}) immediately east of the MQA, roughly coinciding with the surface projection of a marked geoid anomaly in the region (Ussami *et al.* 1999). We argue that the velocity contrast across the Patos Lineament reflects a change in lithospheric strength, from weak (north) to strong (south), consistent with the tectonic framework proposed by Luz *et al.* (2015b), and that the shallow low-velocity anomaly results from high temperatures associated to the ponding of laterally flowing plume material upwelling at a distant mantle plume under the Paraná basin of SE Brazil.

2 TECTONICS AND POST-BREAKUP EVOLUTION

2.1 Tectonic setting

The Borborema Province was characterized by Almeida *et al.* (1981) as a geological and structural domain located in the northeast region of Brazil, limited by the São Francisco Craton to the South, the Parnaíba Basin to the West, and a number of sedimentary basins along the continental margin to the north and east (see Fig. 1). It consists of a basement of Palaeoproterozoic rocks with scattered Archean nuclei that underlie metamorphic supracrustal rocks with ages spanning the entire Proterozoic era (Vaucheret *et al.* 1995; Neves 2003; Van Schmus *et al.* 2008). The Province was mostly structured during the Brasiliano/Pan-African orogeny, at the end of the Neoproterozoic, resulting in a complex orogenic system affected by deformational, metamorphic, and magmatic processes (Santos *et al.* 2008). The predominant view is that the Borborema Province originated from the amalgamation of several small continental fragments, microplates, and island arcs that were scattered among the large landmasses that formed West Gondwana: the West African-São Luiz craton to the North and the Congo-São Francisco craton to the South (De Brito Neves & Cordani 1991; De Sá, Macedo *et al.* 1992; Cordani *et al.* 2003). The Province is pervaded by numerous shear zones (Fig. 1), the most prominent of which being the Transbrasiliano Lineament—outcropping in the NW corner—and the Patos and Pernambuco Lineaments—roughly parallel to each other and coincident with latitudes 7°S and 8°S, respectively. Those lineaments are continental-scale features that can be traced into the African continent in palaeogeographic reconstructions (Trompette 1994; Santos *et al.* 2008; Oliveira *et al.* 2010) and are sometimes regarded as the boundaries of the tectonically independent blocks that assembled to form the Province.

In the Mesozoic, the evolution of the Borborema Province was marked by extensional events that eventually led to continental breakup and the opening of the Atlantic Ocean. Following Matos (1992, 1999), Late Jurassic extensional events would have initially led to the formation of a number of half-grabens along

three main axis of deformation: (i) Gabon–Sergipe–Alagoas, (ii) Recôncavo–Tucano–Jatobá, and (iii) Cariri–Potiguar. Deformation would have initially progressed mostly along the Gabon–Sergipe–Alagoas and Recôncavo–Tucano–Jatobá trends to later shift west and continue along the Cariri–Potiguar trend. The development of the equatorial margin would have eventually aborted the rifting process along the Cariri–Potiguar trend, and allowed for the deposition of thick sedimentary sequences along the other two axis of deformation. With the separation of the West Africa craton from the São Luiz cratons along the equatorial margin, rifting along the Recôncavo–Tucano–Jatobá would have aborted and rifting would have continued along the Gabon–Sergipe–Alagoas trend. The end result would have been the formation of a number of marginal basins along the Brazilian East passive margin and the presence of a number of aborted rift basins in the continental interior.

After continental breakup, episodes of intraplate volcanism and uplift characterized the evolution of the Province. Uplift has its main expression in the high elevations of the Borborema Plateau (Fig. 1), a topographic feature located in the eastern half of the Borborema Province with maximum elevations of ~1200 m (e.g., Morais Neto *et al.* 2009; Oliveira & Medeiros 2012). Tectonic uplift of the Borborema Plateau was originally postulated from stratigraphic ages inferred from the non-fossiliferous deposits of the Serra dos Martins formation, which are found in high-altitude mesas scattered throughout the northern plateau (see Fig. 1). Apatite and zircon fission-track analysis set a maximum Palaeogene age for the sediments, while indirect relationships with Cenozoic volcanics revealed a minimum age of ~25 Ma, implying plateau uplift happened in Cenozoic times (Menezes *et al.* 2003; Jardim de Sá *et al.* 2005). Volcanism consists of small volumes of alkaline rocks that are arranged along two mutually orthogonal magmatic alignments: the FNMA, mostly offshore and trending EW, and the MQA, on shore and trending in the NS direction (Mizusaki *et al.* 2002; Knesel *et al.* 2011). K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dates are 26–34 Ma for Mecejana volcanism (Mizusaki *et al.* 2002) and 22–2 Ma for volcanic rocks at the Fernando de Noronha archipelago (e.g. Knesel *et al.* 2011), delineating a clear age progression from west to east along the alignment; K–Ar ages for MQA volcanism, on the other hand, range from 80 to 30 Ma (see Mizusaki *et al.* 2002), expanding to 93–7 Ma when $^{40}\text{Ar}/^{39}\text{Ar}$ dates are considered (Knesel *et al.* 2011), displaying no clear age progression along the alignment. The expanded age range for MQA volcanism suggests a time overlap of up to 15 Ma in magmatic activity along both alignments, as well as a temporal superposition with postulated Cenozoic uplift in the Borborema Plateau.

Apatite fission-track analysis on granitic–gneissic samples from the Plateau, on the other hand, revealed the presence of two main cooling events (Morais Neto *et al.* 2009): one in the Late Cretaceous (100–90 Ma), and one in the Neogene (20–0 Ma). The Neogene event was related to a climatic change that accentuated erosion of pre-existing topography and resulted in the rugged topography that characterizes the Plateau today and the deposition of the Barreiras Formation. The Late Cretaceous event, on the other hand, was interpreted as reflecting the onset of uplift and denudation of regional, permanent topography in the Plateau, which would have followed shortly after continental breakup along the east passive margin. In this case, overlap with MQA magmatism would have been minimal and plateau uplift would have likely pre-dated intraplate volcanism in the Province.

2.2 Post-breakup evolution

One of the earliest models to explain intraplate volcanism and uplift in the Province invoked the presence of a deep mantle plume under the Borborema Plateau (Jardim de Sá *et al.* 1999; Jardim de Sá 2001). From analysis of Cenozoic deformational patterns in the Province, the ‘reasonable geographic overlap’ between the Borborema Plateau and the Cenozoic volcanism, and the time overlap between the onset of Cenozoic denudation (Barreiras Formation) and alkali volcanism, the authors concluded that thermal doming (due to a mantle plume) was the most likely mechanism of uplift. The plume hypothesis, however, is inconsistent with the lack of age progression along the MQA, the small-volume and long-lived character of its magmatism, and the direction of movement of the South American Plate (Almeida *et al.* 1988; Knesel *et al.* 2011), as well as not necessary for explaining the amount of partial melt, temperature and depth ranges from which the source magmas were derived (Silveira 2006). Moreover, a recent receiver function study in the Province found no evidence for a thermal perturbation in the upper-mantle transition zone under the Borborema Plateau (Pinheiro & Julià 2014). Thermal doming was independently postulated from analysis of residual geoid anomalies (Ussami *et al.* 1999), where a local maximum in the northeastern most corner of the Province (obtained after removing the 10th degree geopotential model EGM96) was observed. Plots of geoid anomaly with respect to topography were consistent with Pratt isostasy with compensation depths of ~150 km so the authors concluded that the Borborema Province is located on top of a low-density (thermal) body that might be locally contributing to tectonic uplift in the region. Interestingly, the maximum anomaly values seem to be displaced to the north with respect to the Plateau and coincide with MQA volcanism.

Knesel *et al.* (2011) associated intraplate volcanism in the Province to an active, small-scale convection cell located at the edge of the continent. Numerical modelling has demonstrated that a small-scale convection cell can develop at a near-vertical thermal boundary such as the continent-ocean transition (King & Anderson 1998). More specifically, calculations show that when the size of the thermal contrast between the hotter mantle underlying the continent and the colder mantle underlying ocean is less than 0.1 per cent of the background temperature, a small-scale convection cell develops with a downwelling sense of flow next to the vertical lithospheric boundary. According to Knesel *et al.* (2011), these attributes fit well within seismic velocity models for the upper mantle beneath Eastern Brazil and the Atlantic Ocean (King & Ritsema 2000; Zhou 1996). They further propose that subcontinental lithosphere would be metasomatized and that entrainment of the lowermost portions of this metasomatized lithosphere into the downwelling flow would enhance fertility and trigger melts that would account for the plume-like characteristics of the Cenozoic basalts. Moreover, elastic modelling of Bouguer admittance functions for the Borborema Province by Oliveira & Medeiros (2012) revealed the need for a negative buried load at the base of the crust under the Borborema Plateau. They proposed that the buried load might be the result of a 4–5 km thick layer of mafic material that underplated the plateau’s crust in the Cenozoic, from melts that would have originated from the same small-scale convection cell postulated in Knesel *et al.* (2011). The postulated mafic layer would have provided the necessary buoyancy to isostatically elevate the Plateau to its present position.

The models above describe possible geodynamic scenarios for coeval volcanism and uplift in the Borborema Province; some models, however, view these two intraplate phenomena as resulting from

independent processes. Morais Neto *et al.* (2009), for instance, identified two cooling events in the region from apatite fission-track analysis on rock samples collected in the Plateau. A younger, Neogene event was interpreted as resulting from climate change, while an older, Late Cretaceous event was related to the onset of uplift in the Plateau. Numerical modelling of continental stretching and breakup by Kuszniir & Karner (2007) had already demonstrated that lateral flow of lithospheric material into the adjacent hinterland may arise from stretching of the lithosphere, and generate uplift of 400–600 m over a timescale of 60–100 My and over lateral distances of 400–600 km (after thermal re-equilibration). Morais Neto *et al.* (2009) thus related Plateau uplift to the opening of the South Atlantic and, as the average topography of the Borborema Plateau is ~1000 m, they also invoked mafic underplating from the Saint Helena and Ascension plumes to explain the remaining 400 m of topography.

Most recently, Luz *et al.* (2015b) determined *S*-wave velocity–depth profiles from the joint inversion of receiver functions and surface wave dispersion for several seismic stations in the Borborema Province. They concluded that there are two crustal types in the eastern Province: (i) the thick crustal type, including stations located in the southern Plateau (south of the Patos Lineament), with thicknesses around 36–38 and a 4–5 km thick layer of mafic material in the lowermost crust; and (ii) the thin crustal type, including stations in the northern Plateau (north of the Patos Lineament) and the Sertaneja Depression, with thicknesses around 30–32 km and lacking the layer of mafic cumulates. Furthermore, the thin crustal type displays an intracrustal discontinuity that is absent in the thick crustal type. The intracrustal discontinuity had been previously imaged in a receiver function study by Almeida *et al.* (2015), and interpreted as a detachment zone that developed during Mesozoic stretching of the original Brasiliano crust. Luz *et al.* (2015b) argue that the layer of mafic cumulates belongs to the original Proterozoic crust that made the Brasiliano orogen, and that its preservation in the thick crustal type—along with the absence of a detachment zone—demonstrates the southern Plateau is rheologically stronger. The comparatively weaker, thin crustal type would have thus developed after delamination of the mafic layer during Brasiliano compression and crustal stretching during Mesozoic extension, resulting in a thin crust with an intracrustal detachment zone and no mafic layer. The thinned crust would have then subsided to form the Sertaneja Depression, remaining at a lower elevation with respect to the relatively undeformed, thicker Plateau crust after thermal re-equilibration. The Sertaneja Depression would have originally included today's northern Plateau, which would be the only portion of the Borborema Plateau that would have actually experienced tectonic uplift.

3 SEISMIC STATIONS, EVENT SELECTION AND DATA PROCESSING

3.1 Seismic stations and event selection

The database for this study was constructed from seismic waveforms recorded at 58 seismic stations belonging to a number of seismic networks in the Borborema Province and neighboring regions (Fig. 2). The combined network included: (i) 14 broad-band stations belonging to the *Rede Sismográfica do Nordeste* (RSISNE), a permanent monitoring network for the NE Brazil region that has been in operation since 2011; (ii) six broad-band stations that operated during 2007–2009 under the *Institutos do Milênio* initiative; (iii) seven broad-band stations and 22 short-period stations

belonging to a temporary network deployed under the *Instituto Nacional de Ciência e Tecnologia em Estudos Tectônicos* (INCT-ET), in operation between 2011 September and 2013 May; (iv) eight broad-band stations that were deployed under the Borborema Deep Electromagnetic and Seismic experiment and (v) the Global Seismographic Network station RCBR, in operation since 1999. The networks were mostly overlapping in time and had at least one common station (RCBR).

The data set consisted of *P* waveforms recorded at the combined seismic network originating from seismic sources at epicentral distances between 30° and 95° and magnitudes 4.6 mb and above. This epicentral distance range includes *P*-ray paths with turning points in the lower mantle that avoid interference with upper-mantle transition zone triplications. Hypocentral locations were taken from the online catalogue of the National Earthquake Information Center website of the United States Geological Survey (<https://earthquake.usgs.gov/earthquakes/search/>), resulting in the selection of 101 seismic sources. However, as the selected sources provided an uneven azimuthal distribution of events biased towards the Nazca–South America subduction zone, some events were removed to balance the azimuthal coverage. We also added teleseismic *P* waveforms for PKPdf core phases, by selection of sources with epicentral distances between 150° and 180° and magnitude 5.5 mb and larger. This resulted in 20 additional seismic events contributing with steeply incident ray paths to our final data set. The final distribution of teleseismic sources for this study is displayed in Fig. 2.

3.2 Relative traveltimes picking

Our observations consist of *P*-wave relative residuals developed from the 1912 waveforms selected for analysis. Relative residuals are preferred because they decrease possible biases from outside the target volume during tomographic inversion (e.g. Evans & Achauer 1993). The relative residual between the *i*th event and the *j*th station (rr^{ij}) is defined as

$$rr^{ij} = t^{ij} - (1/n_i) \sum_{j=1}^{n_i} t^{ij}, \quad (1)$$

where t^{ij} is the difference (or residual) between the observed and calculated traveltimes and n_i is the number of stations recording the *i*th event. Relative residuals are free of uncertainties in event origin time and, when applied to waveforms that are highly coherent across the network, can be accurately retrieved from picking of a common reference feature in the seismic recordings (see e.g. Zandt 1981). Moreover, to minimize picking errors for the reference feature, relative residuals were developed using a variation of the multichannel cross-correlation approach of VanDecar & Crosson (1990). In its original form, this approach first measures relative delay times between all station pairs that recorded a given event through cross-correlation, to then determine relative arrival times at each station by minimizing the measured relative delay times in a least-squares sense. The relative arrival times are constrained to have zero network average during minimization, so they are relative to the network average. In this study, we developed the relative residual measurements through the ‘multichannel phase cross-correlation’ scheme (MCPCC) of Schimmel *et al.* (2003), in which a phase cross-correlation function is utilized to cross-correlate waveforms between station pairs. The phase cross-correlation function (Schimmel 1999) uses a time-series built from the instantaneous phase of

Figure 2. Upper left: location of the 58 seismic stations used in this study. All stations were equipped with broad-band seismometers, except for the INCTE-ET sites marked with white triangles, which were equipped with short-period seismometers. Upper right: distribution of the control grid nodes through the study volume. The control nodes are interspaced at 0.4° in the horizontal directions, and 40 km in the vertical direction. Bottom: location of the 121 seismic sources selected for analysis. The stars represents the localization for the 101 events used to get the P phase and the circles shows the localization for the 20 events used to find the PKP $_{df}$ phase.

the analytic signal and, as amplitude information is dropped, waveform similarity is not biased by the large amplitude portions within the correlation window.

Fig. 3 displays an example of the MCPCC picking procedure through a 5.5 mb event that occurred on 2011 July 5 in the South American Plate (72.2° W, 35° S, 32 km depth). The epicentral distance to the centre of the network is 49° , approximately. Fig. 3(a) shows the vertical recordings for the eight broad-band stations that simultaneously recorded this event, aligned relative to the theoretical P -wave arrival time estimated from the IASP91 global earth model (Kennett & Engdahl 1991). Note the considerable misalignment among the waveforms—likely reflecting important variations in velocity under the stations—and the large discrepancies (up to ~ 3 s) between the onset of P -wave energy and the theoretical traveltimes, probably resulting from a combination of deviations of the true structure with respect to IASP91 and location errors (Schimmel *et al.* 2003). To determine relative residuals, seismograms were bandpass filtered between 0.6 and 1.6 Hz to enhance waveform coherency, and the first prominent positive peak was hand-picked and utilized as the reference feature. Interstation delays were then determined through the MCPCC approach, utilizing cross-correlation windows of 2 s centred on the hand-picked reference times, and

relative arrival times were obtained through the minimization approach of VanDecar & Crosson (1990). Fig. 3(b) shows the bandpass filtered waveforms aligned according to the relative arrival times, demonstrating the visual accuracy of the MCPCC approach in correctly aligning the waveforms. The correctness and accuracy of the alignment is further demonstrated in Fig. 3(c), where the filtered seismograms are superimposed. For this event, the standard deviation is 12 ms and relative delays range from 3.47 ± 0.59 s between stations NBCL and NBMO, to 6.20 ± 0.59 s between stations NBPV and NBPA. MCPCC was applied to a minimum of five waveforms per event, where one waveform was always recorded by the reference station RCBR.

In general, all data and correlation settings, which included frequency bands for cross-correlation, window lengths and position, were individually selected for each event. Commonly, frequencies for P waveforms ranged between 0.6 and 2.0 Hz, while for PKP $_{df}$ waveforms ranged between 0.4 and 1.5 Hz. Uncertainties depend on data quality. For P -wave measurements, most of the standard deviations were typically less than 15 ms, although uncertainties of about 100 ms were sometimes obtained for very poor-quality data. For PKP $_{df}$ core phase measurements, standard deviations were typically about 10 ms or less. When short-period recordings were

Figure 3. (a) Raw vertical component broad-band recordings for a 5.5 mb teleseismic event located in the South American Plate. The traces are aligned with respect to the theoretical *P*-wave arrival time, as predicted by IASP91. (b) Same waveforms aligned with respect to their relative arrival time, bandpass filtered in the 0.6–1.6 Hz frequency band. (c) Filtered traces in (b) superimposed.

involved, the broad-band recording was simulated by replacing the instrument response with that of a RSISNE broad-band instrument.

The measured arrival times were corrected for station elevation using the theoretical ray incidence with an average *P*-wave velocity of 5.8 km s^{-1} , and relative residuals were developed from the corrected arrival times. Overall, we developed relative residuals for 1567 *P* and 345 *PKP*_{df} arrivals, totaling 1912 measurements. The *P*- and *PKP*_{df}-wave relative residual distributions of the final data set are shown in Fig. 4, revealing standard deviations of 0.449 s for both data sets. A Gaussian distribution function is superimposed to the histograms in Fig. 4, demonstrating that the relative residuals deviate only slightly from the Gaus-

sian distribution, and validating their optimization using a least-squares criterion. The limited geographical distribution of seismic sources and receivers often leads to highly irregular sampling of the subsurface by the recorded seismic energy (Rawlinson *et al.* 2010). To balance the distribution of sources around the receivers, some events were removed from the original data set. In particular, events with backazimuths between 180° and 360° and with standard deviation equal or above 15 ms were not utilized during the inversion process. The final data set provided a better balance of sources around the study area. The azimuthal distribution of the *P*-wave relative residuals is shown in the rose diagram displayed in Fig. 5.

Figure 4. Relative residual histograms from *P*-wave seismograms. *PKPdf* arrivals are illustrated in white, and the distribution of all phases is shown in black. A Gaussian distribution (thin red solid line) has been superimposed to assess the relative residual pattern.

4 TOMOGRAPHIC INVERSION

To develop tomographic images from teleseismic *P*-wave data, we used the inversion scheme of Rawlinson *et al.* (2010). The inversion is based on the Fast Marching Method (FMM), an efficient eikonal solver that simultaneously computes seismic traveltimes throughout an entire wave front, and the subspace method of Kennet *et al.* (1988), an inversion scheme that reduces computational efforts through projection into a lower dimensional parameter space. The combination provides stable and robust results even in strongly varying media (Rawlinson *et al.* 2010).

4.1 Model parametrization

The target volume spans 635.5 km in depth, and extends laterally between -12.6° and -1° in latitude and -45° and -33.4° in longitude. Those boundaries were chosen to ensure that all rays enter the target volume from its bottom boundary, with no penetration from the sides. The 3-D structure within the target volume is parametrized through a mosaic of cubic B-spline volume elements, the values of which are controlled by a mesh of velocity control points in spherical coordinates. The control points are interspaced at 0.4° in the horizontal directions, and 40 km in the vertical direction (Fig. 2). The choice of the grid spacing depends on the available *a priori* information, the amount and distribution of data and the capabilities of the inversion routine (Rawlinson & Sambridge 2003). In our study, grid spacing was chosen to be able to detect structures with minimum scale lengths of about 80 km (see Figs 9–12).

4.2 Mathematical background

Predicted relative residuals for the velocity model are obtained from traveltimes between the base and the surface of the target volume calculated through the FMM algorithm. The FMM was originally utilized for migration of coincident reflection data (Sethian

& Popovici 1999; Popovici & Sethian 2002), and to compute reflection and refraction phases in layered media (Rawlinson & Sambridge 2004a,b). In seismic tomography, it was first utilized by Rawlinson *et al.* (2006) to investigate the crust and upper mantle under Tasmania using teleseismic *P*-wave recordings. The FMM tracks the first-arrival wave front as an evolving surface of constant traveltimes defined along a narrow band of gridpoints, where the traveltimes values are updated after solving the eikonal equation using an upwind, mixed-order difference scheme. The traveltimes are then constructed in a downwind fashion from known narrow-band, upwind values (Rawlinson *et al.* 2006), which ensures that the update of the eikonal equation is consistent with the direction of the information flow. The narrow band is initialized from the bottom surface of the local model with initial traveltimes provided by the ak135 global reference model.

The inverse problem is formulated as an optimization problem where the objective function

$$S(\mathbf{m}) = (\mathbf{g}(\mathbf{m}) - \mathbf{d}_{\text{obs}})^T \mathbf{C}_d^{-1} (\mathbf{g}(\mathbf{m}) - \mathbf{d}_{\text{obs}}) + \epsilon (\mathbf{m} - \mathbf{m}_0)^T \mathbf{C}_m^{-1} (\mathbf{m} - \mathbf{m}_0) + \eta \mathbf{m}^T \mathbf{D}^T \mathbf{D} \mathbf{m} \quad (2)$$

is minimized in a least-squares sense. In eq. (2), $\mathbf{m}$ is the vector of unknown model parameters (coefficients of the cubic B-splines), $\mathbf{g}(\mathbf{m})$ are the predicted relative residuals, $\mathbf{d}_{\text{obs}}$ is the vector containing the observed relative residuals, $\mathbf{C}_d$ represents the data covariance matrix, $\mathbf{C}_m$ the model covariance matrix, $\mathbf{m}_0$ is a reference model and $\mathbf{D}$ is a first-order smoothing matrix. The first term in the right-hand side of eq. (2) represents the weighted misfit between observations and predictions; the second term represents a damping regularization, where the solution vector $\mathbf{m}$ is constrained to approach the reference model $\mathbf{m}_0$; and the third term implements a smoothness constraint that penalizes rapid variations among adjacent model parameters. Error estimates for the measured relative residuals are used to fill the diagonal elements in $\mathbf{C}_d$, while the

Figure 5. Rose diagram for the 101 teleseismic P -wave events developed in this study. The radial axis indicates number of events. Note the preferential sampling of our data set in the NE-SW direction, with a heavy bias towards the South American subduction zone.

diagonal elements in C_m are set to 0.1 km s^{-1} , according to expected model uncertainty. However, since the damping factor is varied to tune the inversion, the *a priori* velocity uncertainty we choose is relatively unimportant (Rawlinson *et al.* 2006). The relative contribution of each term is controlled through the damping (ε) and the smoothness (η) parameters, which are determined empirically after analysis of the corresponding trade-off curves (see Section 4.3).

As $g(\mathbf{m})$ is a non-linear operator, the objective function is optimized iteratively after linearization, with successive applications of a mixed-order FMM scheme to solve the forward problem and a subspace inversion method to solve the linearized problem. To solve the inverse problem, a subspace inversion scheme, developed in Kennett *et al.* (1988) is used. The subspace inversion method works by projecting the quadratic approximation of $S(\mathbf{m})$ onto an n -dimensional subspace of the full model space (Rawlinson *et al.* 2006). Singular value decomposition is used to produce an orthonormal basis and, for large-dimensional subspaces, identify and remove the unnecessary basis vectors at each iteration. The model update is then obtained by the inversion of a relatively small square matrix.

4.3 P -wave tomography results

The tomographic results for the Borborema Province are displayed in Figs 6 and 7 as velocity perturbations, where red (warm) colours

refer to slower-than-mean velocity perturbations, while blue (cold) colours refer to faster-than-mean perturbations. Fig. 6 contains a number of horizontal slices at selected depths with the seismic stations and the main geological features superimposed, while Fig. 7 displays a number of cross-sections along several lines of constant latitude and longitude. The final solution model reduces the data root mean square (rms) by 30 per cent, from 471.51 to 335.25 ms, which corresponds to a data variance reduction from 0.222 to 0.114 s^2 .

The images in Figs 6 and 7 were obtained using a damping parameter of $\varepsilon = 30$ and a smoothness parameter of $\eta = 0.00001$. These regularization parameters were determined through the trade-off curves between rms misfit and model roughness, and rms misfit and model variance, respectively. First, the damping parameter was set to $\varepsilon = 1$ and the smoothness parameter η was varied (Fig. 8a), from which a value of $\eta = 0.00001$ was chosen as offering the best trade-off between minimizing the rms and producing the smoothest model. Next, the smoothing parameter was set to $\eta = 0.00001$ and the damping parameter ε was varied (Fig. 8b), yielding a value of $\varepsilon = 30$ satisfying the data set to nearly the same level as the smaller values. The first step was then repeated fixing the damping parameter to $\varepsilon = 30$ (Fig. 6c), confirming a $\eta = 0.00001$ for the smoothness parameter as an appropriate choice. The convergence of the inversion process was examined through inspection of the data rms reduction as function of iteration number. As no significant rms reduction was observed after five iterations, the inversion process

Figure 6. Horizontal slices through the 3-D target volume at several key depths, showing velocity perturbations obtained after inversion of *P*-wave relative residuals. Triangles mark the locations of the seismic stations, while the black, solid lines mark the boundaries of the main geological features. P.B. stands for Parnaíba Basin; S.F.C., São Francisco Craton; A.B., Araripe Basin; R.T.J, Recôncavo-Tucano-Jatobá; B.P., Borborema Plateau; P.L., Patos Lineament; Pb.L., Pernambuco Lineament; S.D., Sertaneja Depression; Po.B., Potiguar Basin; MQA, Macau-Queimadas Alignment and FNMA, Fernando Noronha-Mecejana Alignment. Point heat-flow measurements from Hamza *et al.* (2005) are superimposed to the 150 km depth slice.

was stopped to avoid mapping data errors and inconsistencies into structure.

Inspection of the tomographic results at lithospheric depths (100–150 km), reveals that the Patos Lineament marks the separation between slow velocities to the north and fast velocities to the south throughout the eastern half of the Province. This is further

demonstrated in the two cross-sections cutting across the Borborema Plateau in the NS direction (Figs 7g and h), showing a sharp transition from fast to slow velocities across 6.5°S latitude. The fast velocity anomaly is especially well developed under the southern Borborema Plateau, which previous works have suggested might constitute a rheologically strong tectonic block (Santos *et al.* 2014;

Figure 7. EW (top) and NS (bottom) cross-sections cutting through the 3-D target volume, obtained from inversion of P -wave relative residuals. The inset in the upper left corner displays the location of the cross-sections. Note the presence of a shallow low-velocity anomaly under the northeastern most corner of the Province.

Luz *et al.* 2015b). In the western half, alternating regions of fast and slow velocity seem to be separated by the main shear zones that pervade the Province, especially north of the Patos Lineament. A prominent low-velocity anomaly is observed right under the northeastern most corner of the continent at 100 km depth (Figs 6a and 7a). The western border of this anomaly coincides pretty closely with the MQA, while the southern border seems to coincide with the east–west trending Patos Lineament. The vertical cross-sections reveal this anomaly does not exceed 150 km depth (Figs 7a and e).

At asthenospheric depths (200–400 km), inspection of the horizontal slices in Fig. 6 reveals a stripe of low-velocity material trending in the NE–SW direction, with two lateral offshoots in the EW direction approximately coinciding with the Patos Lineament

and the Araripe basin. Most of the teleseismic sources selected for this study sample the underlying mantle along the NE–SW direction (recall Fig. 5); however, we think this NE–SW trend is well resolved (see resolution tests in Section 5). Slow velocities are also observed under the equatorial margin, reaching about 100–150 km inland from the border of the continent. Our tomographic slices at 200–400 km depth also suggest fast velocities might be present further south and west, under the São Francisco craton and the Parnaíba basin, respectively. Although this possibility cannot be ruled out, resolution tests (see Section 5) suggest that seismic anomalies are not well resolved south of 10°S latitude and west of 40°W longitude.

Figure 8. Trade-off curves utilized to determine damping and smoothing during the tomographic inversion. (a) Damping is set to $\epsilon = 1$ and smoothness is varied to find an optimal value of $\eta = 0.00001$; (b) smoothness is fixed ($\eta = 0.00001$) and damping is varied to determine an optimal value of 30 and (c) damping is set to 30 and smoothness is varied to confirm the value found in (a).

5 RESOLUTION TESTS

To investigate the robustness of the tomographic features described in the previous section, a number of resolution tests have been carried out. Those tests include the traditional checkerboard test, along with a spike or special geometry test focusing on specific features identified within the target volume. We always use source and receiver distributions that mimic the corresponding distributions of the observational data set in order to predict the pattern of relative residuals for a given structure.

The first resolution test considered in this study is the checkerboard test, in which the input model consists of alternating regions of high and low velocities. Regions in which the checkerboard pattern is correctly recovered can be regarded as well resolved. For the test, we used a 100 km length-scale for the alternating pattern in all directions, with a separation of about 40 km between anomalies. The synthetic traveltimes ‘data’ were contaminated with Gaussian noise of zero mean and 50 ms standard deviation. Horizontal slices for the input and the recovered models are shown in Fig. 9, and vertical cross-sections are shown in Figs 10 (E-W slices) and 11 (N-S slices). The horizontal slices show that the checkerboard pattern is well recovered north of latitude 10°S and east of longitude 40°W, although the size of the predicted anomalies degrades with depth due to vertical smearing along the teleseismic ray paths (Figs 9 b and c). The alternating pattern is reasonably recovered down to 400 km

depth within this region, albeit showing an apparent smearing along the predominating western ray paths (recall Fig. 5) and degrading the size of the anomalies at large depths.

For the second resolution test, the input model consisted of a low-velocity stripe at asthenospheric depths trending in the NE-SW direction and a shallower low-velocity anomaly in the northeastern most corner of the Borborema Province, mimicking the main features identified in the previous section. The anomalies, however, are disconnected in the input model. Also, as with the checkerboard test, the synthetic teleseismic ‘data’ were contaminated with Gaussian noise of zero mean and 50 ms standard deviation. The low-velocity stripe was centred at depths of 250 km, while the low-velocity anomaly at the corner of the continent was centred at 150 km depth. Realize that we are more interested in testing the recovery of the overall geometry and shape of the anomalies, rather than the correct amplitudes. The anomalies seem to be well recovered laterally at all depths (Fig. 12), although vertical resolution is poor due to vertical smearing (Fig. 13). Note that the NE-SW trending stripe and the shallower (corner) anomaly appear connected in the inverted model (Figs 12b and 13b, left). Finally, because of the steep ray paths and the decreasing number of crossing rays at shallow depths, the elongation of the anomalies appears to be larger in the upper portion of the recovered model.

Figure 9. Input (left) and recovered (right) checkerboard models utilized in the resolution test. We used a 100 km length-scale for the alternating pattern in all directions, with a separation of about 40 km between anomalies. A Gaussian noise of zero mean and 50 ms of standard deviation is used to contaminate the synthetic ‘data’. The input pattern is well recovered north of latitude 10°S and east of longitude 40°W.

6 IMPLICATIONS FOR POST-BREAKUP EVOLUTION

The most important features revealed by the tomographic images of the Borborema upper mantle developed in this study are: (i) the existence of a relatively fast lithospheric mantle under the southern Borborema Plateau, when compared to the surrounding regions (including the northern Plateau); (ii) the presence of a marked, shallow (< 150 km) low-velocity anomaly under the northeastern most corner of the Borborema Province and (iii) the presence of a NE-SW trending stripe of low-velocity material at asthenospheric

depths. These three observations have important implications for models of (intraplate) plateau uplift and volcanism proposed for the Borborema Province.

6.1 Intraplate volcanism

As described in Section 2, post-breakup volcanism is arranged along two magmatic alignments: the FNMA, mostly offshore and trending in the EW direction, and the MQA, located on shore and trending in the NS direction. A regional analysis of the spatial and temporal

Figure 10. Input (left) and recovered (right) checkerboard models for synthetic tests: E-W cross-sections from (a) 6.2 S, (b) 8 S and (c) 9.5 S. Typical smearing can be seen where the distribution of the stations is poor and at depths below 400 km.

distribution of K/Ar ages for magmatic and volcanic rock samples in the Borborema Province by Mizusaki *et al.* (2002) concluded that radiometric ages younger than 80 Ma resulted from hotspot activity. The authors noted that, according to petrographic and geochemical data, those rocks likely formed during tectonic processes related to the westward drift of the South American Plate after splitting from the African plate, and that they represented the remains of the large mantle plume (St. Helena) that induced continental breakup and opening of the equatorial Atlantic. They also noted that the most pronounced hotspot would have evolved along the FNMA, which displays a westward trend along the chain. A mantle plume origin for the MQA volcanism had also been proposed by Jardim de Sá *et al.* (1999), in spite of its known lack of age progression along its trend and its orthogonal orientation with respect to plate movement (Almeida *et al.* 1988). Moreover, analysis of geochemical data had shown that MQA magmatism could be equally well explained through metasomatic processes in the lower lithosphere followed by decompression melting (Silveira 2006).

Based on $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations, Knesel *et al.* (2011) noted that volcanic activity in the continent ranged between 93 and 7 Ma and that it did not stop prior to the onset of volcanism on Fernando de Noronha, demonstrating a time overlap up to 15 Ma between both magmatic alignments. The authors argued that the extended duration, small volume, and lack of clear age progression of the on-shore volcanism suggests that it is more likely the result of edge-driven convection rather than a mantle plume. According to numerical modelling (King & Anderson 1998), a small-scale convection cell may form in the sublithospheric mantle, next to rapid (step-like) changes in lithospheric thickness, when temperature contrasts across the near-vertical boundary are small enough. Downward flow is expected next to the boundary, while flow in the upward direction would be observed at distances between 100 and 600 km downstream. According to Knesel *et al.* (2011), entrainment of the lowermost portions of a metasomatized

Figure 11. Input (left) and recovered (right) checkerboard models for synthetic tests: N-S slices from (a) 36.4 W, (b) 38 W and (c) 39.7 W.

lithospheric mantle into the edge-driven downwelling would enhance fertility and trigger melts that would account for the plume-like characteristics of the Cenozoic basalts. Such a flow would also explain the lack of age progression along the MQA and, as a small-scale convection cell would travel with the plate, it would also account for the long-lived character of the on-shore volcanism.

Mapping of transition zone thickness under the Province with receiver functions has already demonstrated the absence of deep thermal anomalies that could be related to the presence of a mantle plume (Pinheiro & Julià 2014). Our tomographic images at 100 km depth now demonstrate the presence of a marked low-velocity anomaly under the northeastern most corner of the Province, bound by the Macau-Queimadas magmatic alignment to the west and by the Patos Lineament to the south (Fig. 6a). The location of this low-velocity anomaly closely coincides with the local geoid anomaly of 10 m reported by Ussami *et al.* (1999). The negative polarity of the observed anomaly, its shallow character, and the geographical overlap of its western border with the MQA suggest that this

anomaly might represent the seismic signature of the small-scale convection cell postulated by Knesel *et al.* (2011). Numerical modelling predicts edge-driven convection to dominate sublithospheric flow for peak-to-peak temperature contrasts of up to a few degrees, tolerating a background asthenospheric flow from continent to ocean under Couette flow boundary conditions (King & Anderson 1998). The vertical cross-section at 6°S latitude (Fig. 7a) shows an asthenospheric (250 km depth) *P*-velocity change across the MQA of $\sim 0.1 \text{ km s}^{-1}$, which would translate into a temperature contrast of 100 °C (Cammarano *et al.* 2003). Realize that a 0.1 km s^{-1} velocity contrast is likely an underestimation, as damping during the tomographic inversion will tend to make velocity variations smooth. This temperature contrast is about 1 order of magnitude larger than required for edge-driven convection, and suggests that large-scale, eastward asthenospheric flow is more likely to dominate.

Alternatively, Sleep (2003) proposed that intraplate volcanism in northeast Brazil younger than 85 Ma could be explained through lateral sublithospheric flow. The source of such flow would be

Figure 12. Input and recovered (left and right, respectively) model with P -wave velocity anomalies for synthetic tests. Depth slices for: (a) 100 km, (b) 150 km and (c) 250 km. The images show a good lateral resolution for all depths. However, a vertical smearing is found in the output image at 150 km (b, left).

a plume tail now centred under the Paraná basin of SE Brazil, more than 1500 km away from the Borborema Province, which has been imaged in other tomographic studies (Vandecar *et al.* 1995; Schimmel *et al.* 2003). Although the seismic anomaly had been initially interpreted as a fossil mantle plume related to the voluminous flood basalts that are interfingered with the basin's sediments, Sleep (2003) related such anomaly to a different plume—the Paraná plume—that would have crossed the Amazon rift basin at about 85 Ma and reached the Brazilian northeast through lateral flow along sublithospheric channels. Moreover, the laterally flowing magmas

would have ponded in front of the São Francisco craton's thick lithosphere, which would have behaved as a lithospheric dam. The dam would have eventually been breached on the northern and southern edges, reaching the continental margins and forming the Fernando de Noronha and Martin Vaz hotspot tracks.

Interestingly, our tomographic images at 200–450 km depth reveal a low-velocity channel in the asthenospheric mantle trending in the NE–SW direction, bordering the Borborema plateau to the west. We propose that this low-velocity channel is the seismic sig-

Figure 13. Input and recovered (left and right, respectively) model with P -wave velocity anomalies for synthetic tests. E-W slices from the Fig. 12 for: (a) 6° S and (b) 6.5° S. The vertical smearing make the anomalies seems connected in the output image (b).

nature of the lateral flow postulated in Sleep (2003). Moreover, if our interpretation were correct, the tomographic images suggest the flow has two EW trending offshoots under the Araripe basin and along the Patos Lineament, respectively, and that it spreads along the equatorial margin upon reaching the Potiguar Basin. The vertical cross-sections in Fig. 7 additionally suggest that the shallow

(100 km depth) low-velocity anomaly might be connected to this lateral flow, which would indicate ponding of plume material at the edge of the continent. Realize that although the numerical experiments presented in Figs 12 and 13 do show that the connection might be an artefact resulting from smearing along the ray paths, they do not preclude the anomalies from being actually connected.

Nonetheless, more detailed tomographic images are needed in order to resolve the continuity of the shallow and deep low-velocity anomalies.

Independent evidence for our postulated channel of lateral asthenospheric flow has been sought in published compilations of surface heat flow (Hamza & Muñoz 1996; Hamza & Silva Dias 2003; Hamza *et al.* 2005; Davies 2013). All compilations roughly agree that the Borborema Province is a region of relatively higher heat flow when compared to the surrounding geological provinces; however, those studies targeted lateral variations on either global or continental scales, and do not seem to have enough resolution to resolve lateral variations on the same scale that our tomographic images resolve lateral S -velocity variations. Nonetheless, a detailed comparison of point heat-flow measurements (Hamza *et al.* 2005) with our 200 km depth tomographic slice (Fig. 6b) reveals a general superposition of 40–70 mW m⁻² values with fast-velocity anomalies and of 70–100 mW m⁻² values with slow-velocity anomalies. However, important discrepancies are also observed. Low heat-flow values of 20–40 mW m⁻², for instance, are found within the slow-velocity anomaly region under the north easternmost corner of the continent. Independent constraints on subsurface temperature for the Borborema Province could in principle be obtained from Curie isothermal depths derived from spectral analysis of magnetic data (Correa *et al.* 2016). Although relatively shallow Curie depths are found in the NW portion of the Province, consistent with slow velocities observed in our tomographic slices, no clear correlation is observed for the remaining of the Province at any depth. Low resolution due to large spatial windows (150 × 150 km) required to build the radial magnetic spectra probably make a small-wavelength comparison difficult. Finally, and perhaps not surprisingly, a visual correlation of Curie depths with the point heat-flow measurements of Hamza *et al.* (2005)—including the 20–40 mW m⁻² values—is also not fully satisfying.

Another source for independent validation of our postulated asthenospheric channel is seismic anisotropy, as strain can lead to the alignment of its fast anisotropic axis in the direction of flow (e.g. Zhang & Karato 1995). Unfortunately, the only SKS splitting estimates available in the Province (Bastow *et al.* 2015) seem to either be dominated by fossil anisotropy in the lithosphere or yield null results due to the cumulative effect of travelling through anisotropic layers with orthogonally oriented fast axes. Nonetheless, another SKS splitting study for the SE and Central Brazil regions (Assumpção *et al.* 2006) has provided seismic evidence for lateral asthenospheric flow around the southern border of the São Francisco craton, demonstrating that channelling of sublithospheric mantle under the Brazilian lithosphere is indeed plausible.

6.2 Plateau uplift

Regarding uplift of the Borborema Plateau, as reviewed in Section 2, traditional explanations fall into one of the following two categories: (i) permanent topography, in which the high topography is explained through isostatic compensation due to thickening of the underlying crust (Morais Neto *et al.* 2009; Oliveira & Medeiros 2012) and, (ii) dynamic topography, in which the high topography is explained through thermal doming due to either a mantle plume (Jardim de Sá *et al.* 1999) or a hot, low-density body in the upper mantle (Ussami *et al.* 1999). Alternatively, it has recently been proposed that the Borborema Plateau's high topography is not the result of tectonic uplift, but differential stretching of the lithosphere due to the stronger rheology of the southern Plateau (Luz *et al.* 2015b).

As stated before, fast- and low-velocity anomalies observed in tomographic images of the upper mantle are commonly interpreted in terms of temperature variation (Goes *et al.* 2000; James *et al.* 2004; Goes *et al.* 2005). However, variations in seismic velocity can also result from changes in chemical composition, temperature, partial melts, and/or fluids. This question was thoroughly investigated in Artemieva *et al.* (2004), who assessed the relative contributions of thermal and non-thermal effects on global tomographic maps of seismic S velocity (V_S) and attenuation (Q_S) through comparison with maps of the thermal structure of the continental upper mantle. In their study, they reported that the sign of V_S (and Q_S) usually inversely correlates with lithospheric temperature, and that theoretical V_S^T (and Q_S^T) values calculated solely from temperature anomalies are capable of explaining observed variations in V_S (and Q_S) in 50 per cent of the continental areas at 100 km depth. Moreover, a detailed inspection of the comparison maps shows that most of the Brazilian northeast falls within that 50 per cent region, although the eastern half of the Borborema Province seems to have S velocities a bit (~ 2 per cent) above the theoretical values. In any case, Artemieva *et al.* (2004) warn about interpreting small-scale anomalies due to different lateral resolution between temperature and tomographic maps.

Creep processes in the lower lithosphere can be modelled through a viscoelastic medium where viscosity depends exponentially on pressure and the inverse absolute temperature (Turcotte & Schubert 1973). Thus, for rocks kept at a given pressure (depth), an increase in temperature will rapidly decrease viscosity, reduce the depth at which viscoelastic relaxation occurs, and make the lithosphere more easily deformable through thermally activated creep processes. The tomographic slices developed in this study at 100–150 km depth (Figs 6a and b) demonstrate that the southern Plateau is underlain by relatively fast lithospheric mantle. Fast velocities actually continue across the southern border of the plateau and into the São Francisco craton but, as discussed in Section 5, seismic structure is not well resolved south of 10°S latitude (Fig. 9). Using a thermal interpretation of the velocity anomalies (Artemieva *et al.* 2014), and further relating temperature to rheology (Turcotte & Schubert 1973), we can conclude that the southern Plateau constitutes a colder, rheologically stronger lithospheric block within the Borborema Province. Thus, if our thermal interpretation of the velocity anomalies is correct, the tomographic images developed for the Borborema Province favor the differential stretching model proposed by Luz *et al.* (2015b) to explain the elevated topography of the southern Borborema Plateau.

The origin of the relatively high elevations also observed in the northern Plateau, nonetheless, remains unexplained. The vertical cross-section along latitude 6°S (Fig. 7a) shows that the low-velocity anomaly at 150 km depth under the northern Plateau might be connected with the postulated lateral flow bordering the southern Plateau to the west. This suggests thermal doming due to a hotter-than-average mantle might be the origin for tectonic uplift in the northern Plateau. However, the entire lithospheric mantle right above the postulated lateral flow seems to have been thermally heated, and only the region between the southern Plateau and the Araripe basin displays an elevated topography. Indeed, Oliveira & Medeiros (2012) noted a non-zero areal integral of free-air gravity anomalies throughout the Province, which indicates its topography might have a dynamic component. Dynamic topography due to thermal doming along the postulated lateral mantle flow might be responsible for the dynamic component of uplift on a regional scale, but cannot explain the detailed, short-wavelength, topography.

7 CONCLUSIONS

Teleseismic *P*-wave tomography has been used to investigate the seismic structure of the lithospheric and sublithospheric mantle beneath the Borborema Province of NE Brazil. Our tomographic images reveal a low-velocity anomaly close to the western border of the Borborema Plateau that might represent a lateral flow originating from a distant Paraná plume in SE Brazil. We suggest that intraplate volcanism results from ponding of this lateral flow at the corner of the continent, and that this low-velocity channel might be responsible for a dynamic component of the long-wavelength Borborema Province's topography. The tomographic images also reveal that the lithosphere north of the Patos Lineament is slower than that south of the lineament, supporting the differential stretching model of Luz *et al.* (2015b) as a valid explanation for the elevation of the southern Borborema Plateau. Why the northern Plateau is also at high elevation, however, remains enigmatic.

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REFERENCES

- Afonso, J.C., Ranalli, G., Fernández, M., Griffin, W.L., O'Reilly, S.Y. & Faul, U., 2010. On the V_p/V_s correlation in mantle peridotites: implications for the identification of thermal and compositional anomalies in the upper mantle, *Earth planet. Sci. Lett.*, **289**(3–4), 606–618.
- Almeida, F.F.M., Carneiro, C.D.R., Machado, D.L. & Dehira, L.K., 1988. Magmatismo pós-paleozóico no Nordeste oriental do Brasil, *Rev. Bras. Geociências*, **18**(4), 451–462.
- Almeida, F.F.M., Hasui, Y., de Brito Neves, B.B. & Fuck, R.A., 1981. Brazilian structural provinces: an introduction, *Earth-Sci. Rev.*, **17**(1–2), 1–29.
- Almeida, Y.B., Juliá, J. & Frassetto, A., 2015. Crustal architecture of the Borborema Province, NE Brazil, from receiver function CCP stacks: implications for Mesozoic stretching and Cenozoic uplift, *Tectonophysics*, **649**, 68–80.
- Artemieva, I.M., Billien, M., Lévêque, J.J. & Mooney, W.D., 2004. Shear wave velocity, seismic attenuation, and thermal structure of the continental upper mantle, *Geophys. J. Int.*, **157**(2), 607–628.
- Assumpção, M., Heintz, M., Vauchez, A. & Silva, M.E., 2006. Upper mantle anisotropy in SE and Central Brazil from SKS splitting: evidence of asthenospheric flow around a cratonic keel, *Earth planet. Sci. Lett.*, **250**(1–2), 224–240.
- Bastow, I.D., Julia, J., do Nascimento, A.F., Fuck, R.A., Buckthorp, T.L. & McClellan, J.J., 2015. Upper mantle anisotropy of the Borborema Province, NE Brazil: implications for intra-plate deformation and sub-cratonic asthenospheric flow, *Tectonophysics*, **657**, 81–93.
- Bastow, I.D., Nyblade, A.A., Stuart, G.W., Rooney, T.O. & Benoit, M.H., 2008. Upper mantle seismic structure beneath the Ethiopian hot spot: rifting at the edge of the African low-velocity anomaly, *Geochem. Geophys. Geosyst.*, **9**(12).
- Bastow, I.D., Stuart, G.W., Kendall, J.M. & Ebinger, C.J., 2005. Upper-mantle seismic structure in a region of incipient continental breakup: northern Ethiopian rift, *Geophys. J. Int.*, **162**(2), 479–493.
- Brito Neves, B.B. & Cordani, U.G., 1991. Tectonic evolution of South America during the late Proterozoic, *Precambrian Res.*, **53**(1–2), 23–40.
- Cammarano, F., Goes, S., Vacher, P. & Giardini, D., 2003. Inferring upper-mantle temperatures from seismic velocities, *Phys. Earth planet. Inter.*, **138**(3), 197–222.
- Cordani, U.G., Brito-Neves, B.B. & D'Agrella-Filho, M.S., 2003. From Rodinia to Gondwana: a review of the available evidence from South America, *Gondwana Res.*, **6**(2), 275–283.
- Correa, R.T., Vidotti, R.M. & Oksum, E., 2016. Curie surface of Borborema province, Brazil, *Tectonophysics*, **679**, 73–87.
- Davies, J.H., 2013. Global map of solid Earth surface heat flow. Geochemistry, *Geophys. Geosyst.*, **14**(10), 4608–4622.
- Descamps, F., Trampert, J. & Snieder, R., 2002. Anomalies of temperature and iron in the uppermost mantle inferred from gravity data and tomographic models, *Phys. Earth planet. Inter.*, **129**(3–4), 245–264.
- De Sá, E.F.J., Macedo, M.H., Fuck, R.A. & Kawashita, K., 1992. Terrenos proterozóicos na Província Borborema e a margem norte do Cráton São Francisco, *Braz. J. Geol.*, **22**(4), 472–480.
- Evans, J.R. & Achauer, U., 1993. Using the ACH method: theory and application to continental-scale studies, in *Seismic Tomography: Theory and Practice*, Vol. **319**, 319–360, Chapman and hall.
- Goes, S., Govers, R. & Vacher, P., 2000. Shallow mantle temperatures under Europe from P and S wave tomography, *J. geophys. Res.*, **105**(B5), 11153–11169.
- Goes, S., Simons, F.J. & Yoshizawa, K., 2005. Seismic constraints on temperature of the Australian uppermost mantle, *Earth planet. Sci. Lett.*, **236**(1), 227–237.
- Hamza, V.M., Dias, F.J.S., Gomes, A.J. & Terceros, Z.G.D., 2005. Numerical and functional representations of regional heat flow in South America, *Phys. Earth planet. Inter.*, **152**(4), 223–256.
- Hamza, V.M. & Dias, F.J.S., 2003. Functional representation of regional heat flow in South America—implications for the occurrence of low-temperature geothermal resources, *Trans.-Geotherm. Resour. Counc.*, **27**, 615–618.
- Hamza, V.M. & Muñoz, M., 1996. Heat flow map of South America, *Geothermics*, **25**(6), 599–646.
- Hansen, S.E., Nyblade, A.A. & Benoit, M.H., 2012. Mantle structure beneath Africa and Arabia from adaptively parameterized P-wave tomography: Implications for the origin of Cenozoic Afro-Arabian tectonism, *Earth planet. Sci. Lett.*, **319**, 23–34.
- James, D.E., Boyd, F.R., Schutt, D., Bell, D.R. & Carlson, R.W., 2004. Xenolith constraints on seismic velocities in the upper mantle beneath southern Africa, *Geochem. Geophys. Geosyst.*, **5**(1).
- Jardim de Sá, E.F., 2001. Tectônica cenozóica na margem equatorial da Província Borborema, Nordeste do Brasil (A contribuição da Geologia Estrutural no continente), in *VIII Simpósio Nacional de Estudos Tectônicos II International Symposium on Tectonics, Recife, Annals*, p. 25–28.
- Jardim de Sá, E.F., Matos, R.M.D., Moraes Neto, J.M., Saadi, A., Pessoa, Neto & OC., 1999. Epirogenia Cenozóica na Província Borborema : síntese e discussão sobre os modelos de deformação associados, *Simp. Nac. Estudos Tect., Lençóis Bahia, Anais*, **5861**, 58–67.
- Jardim de Sá, E.F., Souza, Z.S., Vasconcelos, P.M., Saadi, A., Galindo, A.C., Lima, M.G. & Oliveira, M.J.R., 2005. Marcos temporais para a evolução cenozóica do Planalto do Borborema, in *X Simpósio Nacional de Estudos Tectônicos IV International Symposium on Tectonics*, pp. 160–162, SBG, Curitiba, Abstracts CD, SBG, Curitiba, Abstracts CD.
- Kennett, B.L.N. & Engdahl, E.R., 1991. Traveltimes for global earthquake location and phase identification, *Geophys. J. Int.*, **105**(2), 429–465.
- Kennett, B.L.N., Sambridge, M.S. & Williamson, P.R., 1988. Subspace methods for large inverse problems with multiple parameter classes, *Geophys. J. Int.*, **94**(2), 237–247.
- King, S.D. & Anderson, D.L., 1998. Edge-driven convection, *Earth planet. Sci. Lett.*, **160**(3), 289–296.

- King, S.D. & Ritsema, J., 2000. African hot spot volcanism: small-scale convection in the upper mantle beneath cratons, *Science*, **290**(5494), 1137–1140.
- Knesel, K.M., Souza, Z.S., Vasconcelos, P.M., Cohen, B.E. & Silveira, F.V., 2011. Young volcanism in the Borborema Province, NE Brazil, shows no evidence for a trace of the Fernando de Noronha plume on the continent, *Earth planet. Sci. Lett.*, **302**(1), 38–50.
- Kusznir, N.J. & Karner, G.D., 2007. Continental lithospheric thinning and breakup in response to upwelling divergent mantle flow: application to the Woodlark, Newfoundland and Iberia margins, *Geol. Soc. Lond. Spec. Publ.*, **282**(1), 389–419.
- Luz, R., Julià, J. & Nascimento, A.F., 2015b. Crustal structure of the eastern Borborema Province, NE Brazil, from the joint inversion of receiver functions and surface wave dispersion: Implications for plateau uplift, *J. geophys. Res.*, **120**(5), 3848–3869.
- Luz, R.M.N., Julià, J. & Do Nascimento, A.F., 2015a. Bulk crustal properties of the Borborema Province, NE Brazil, from P-wave receiver functions: Implications for models of intraplate Cenozoic uplift, *Tectonophysics*, **644**, 81–91.
- Matos, R.M.D., 1992. The northeast Brazilian rift system, *Tectonics*, **11**(4), 766–791.
- Matos, R.M.D., 1999. History of the northeast Brazilian rift system: kinematic implications for the break-up between Brazil and West Africa, *Geol. Soc. Lond. Spec. Publ.*, **153**(1), 55–73.
- Menezes, M.R.F., Morais Neto, J.M., Szatmari, P. & York, D., 2003. Relações cronológicas entre o vulcanismo Macau e a Formação Serra do Martins, com base na datação Ar/Ar do plug basáltico "Serrote Preto" (RN, Nordeste do Brasil), in *IX Simpósio Nacional de Estudos Tectônicos III International Symposium on Tectonics, Búzios, Annals*, pp. 246–249.
- Milani, E.J. & Ramos, V.A., 1998. Orogenias paleozóicas no domínio sul ocidental do Gondwana e os ciclos de subsidência da Bacia do Paraná, *Rev. Bras. Geociências*, **28**(4), 473–484.
- Mizusaki, A.M.P., Thomaz-Filho, A., Milani, E.J. & De Césaro, P., 2002. Mesozoic and Cenozoic igneous activity and its tectonic control in north-eastern Brazil, *J. South Am. Earth Sci.*, **15**(2), 183–198.
- Morais Neto, J.M., Hegarty, K.A., Karner, G.D. & Alkmim, F.F.D., 2009. Timing and mechanisms for the generation and modification of the anomalous topography of the Borborema Province, northeastern Brazil, *Mar. Petrol. Geol.*, **26**(7), 1070–1086.
- Neves, S.P., 2003. Proterozoic history of the Borborema province (NE Brazil): Correlations with neighboring cratons and Pan-African belts and implications for the evolution of western Gondwana, *Tectonics*, **22**(4).
- Oliveira, E.P., Windley, B.F. & Araújo, M.N., 2010. The Neoproterozoic Sergipano orogenic belt, NE Brazil: a complete plate tectonic cycle in western Gondwana, *Precambrian Res.*, **181**(1), 64–84.
- Oliveira, R.G. & de Medeiros, W.E., 2012. Evidences of buried loads in the base of the crust of Borborema Plateau (NE Brazil) from Bouguer admittance estimates, *J. South Am. Earth Sci.*, **37**, 60–76.
- Pinheiro, A.G. & Julià, J., 2014. Normal thickness of the upper mantle transition zone in NE Brazil does not favour mantle plumes as origin for intraplate Cenozoic volcanism, *Geophys. J. Int.*, **199**(2), 996–1005.
- Popovici, A.M. & Sethian, J.A., 2002. 3-D imaging using higher order fast marching traveltimes, *Geophysics*, **67**(2), 604–609.
- Poupinet, G., Arndt, N. & Vacher, P., 2003. Seismic tomography beneath stable tectonic regions and the origin and composition of the continental lithospheric mantle, *Earth planet. Sci. Lett.*, **212**(1), 89–101.
- Rawlinson, N., Pozgay, S. & Fishwick, S., 2010. Seismic tomography: a window into deep Earth, *Phys. Earth planet. Inter.*, **178**(3), 101–135.
- Rawlinson, N., Reading, A.M. & Kennett, B.L., 2006. Lithospheric structure of Tasmania from a novel form of teleseismic tomography, *J. geophys. Res.*, **111**(B2).
- Rawlinson, N. & Sambridge, M., 2003. Seismic traveltome tomography of the crust and lithosphere, *Adv. Geophys.*, **46**, 81–199.
- Rawlinson, N. & Sambridge, M., 2004a. Multiple reflection and transmission phases in complex layered media using a multistage fast marching method, *Geophysics*, **69**(5), 1338–1350.
- Rawlinson, N. & Sambridge, M., 2004b. Wave front evolution in strongly heterogeneous layered media using the fast marching method, *Geophys. J. Int.*, **56**(3), 631–647.
- Ritsema, J., Nyblade, A.A., Owens, T.J., Langston, C.A. & VanDecar, J.C., 1998. Upper mantle seismic velocity structure beneath Tanzania, east Africa: implications for the stability of cratonic lithosphere, *J. geophys. Res.*, **103**, 21201–21213.
- Rocha, M.P., Schimmel, M. & Assumpção, M., 2010. Upper-mantle seismic structure beneath SE and Central Brazil from P-and S-wave regional traveltome tomography, *Geophys. J. Int.*, **184**, 268–286.
- Santos, A.C., Padilha, A., Fuck, R.A., Pires, A.C., Vitorello, I. & Pádua, M.B., 2014. Deep structure of a stretched lithosphere: Magnetotelluric imaging of the southeastern Borborema Province, NE Brazil, *Tectonophysics*, **610**, 39–50.
- Santos, T.J.S., Fetter, A.H., Hackspacher, P.C., Van Schmus, W.R. & Neto, J.N., 2008. Neoproterozoic tectonic and magmatic episodes in the NW sector of Borborema Province, NE Brazil, during assembly of Western Gondwana, *J. South Am. Earth Sci.*, **25**(3), 271–284.
- Schimmel, M., 1999. Phase cross-correlations: Design, comparisons, and applications, *Bull. seism. Soc. Am.*, **89**(5), 1366–1378.
- Schimmel, M., Assumpao, M. & VanDecar, J.C., 2003. Seismic velocity anomalies beneath SE Brazil from P and S wave travel time inversions, *J. geophys. Res.*, **108**(B4).
- Sethian, J.A. & Popovici, A.M., 1999. 3-D traveltome computation using the fast marching method, *Geophysics*, **64**(2), 516–523.
- Silveira, F.V., 2006. Magmatismo cenozóico da porção central do Rio Grande do Norte, NE do Brasil, *Dissertation, Universidade Federal do Rio Grande do Norte, Natal*.
- Sleep, N.H., 2003. Fate of mantle plume material trapped within a lithospheric catchment with reference to Brazil, *Geochem. Geophys. Geosyst.*, **4**(7).
- Trompette, R. & Carozzi, A.V., 1994. Pan-African Brasileiro aggregation of South America and Africa, *Geology of Gondwana (2000–500 Ma)*, Rotterdam, Balkema, p. 350.
- Turcotte, D.L. & Schubert, G., 1973. Frictional heating of the descending lithosphere, *J. geophys. Res.*, **78**(26), 5876–5886.
- Ussami, N., Molina, E.C. & Medeiros, W.E., 1999. Novos Vnculos sobre a Evoluo Térmica da Margem Continental Leste do Brasil, In *National Symposium on Tectonic Studies and International Symposium on Tectonics VII Simpósio Nacional de Estudos Tectônicos, Lençóis - BA. Resumos Expandidos*. Salvador - BA : Sociedade Brasileira de Geologia, **7**, 20–23.
- VanDecar, J.C., 1991. Upper-mantle structure of the Cascadia subduction zone from non-linear teleseismic travel-time inversion, *Doctoral dissertation*, <https://digital.lib.washington.edu/researchworks/handle/1773/6804>.
- VanDecar, J.C. & Crosson, R.S., 1990. Determination of teleseismic relative phase arrival times using multi-channel cross-correlation and least squares, *Bull. seism. Soc. Am.*, **80**(1), 150–169.
- VanDecar, J.C., James, D.E. & Assumpção, M., 1995. Seismic evidence for a fossil mantle plume beneath South America and implications for plate driving forces, *Nature*, **378**(6552), 25.
- Van Schmus, W.R., Oliveira, E.P., Da Silva Filho, A.F., Toteu, S.F., Penaye, J. & Guimarães, I.P., 2008. Proterozoic links between the Borborema Province, NE Brazil, and the Central African Fold Belt, *Geol. Soc. Lond., Spec. Publ.*, **294**(1), 69–99.
- Vaucher, A., Neves, S., Caby, R., Corsini, M., Egydio-Silva, M., Arthaud, M. & Amaro, V., 1995. The Borborema shear zone system, NE Brazil, *J. South Am. Earth Sci.*, **8**(3–4), 247–266.
- Zandt, G., 1981. Seismic images of the deep structure of the San Andreas fault system, central Coast Ranges, California, *J. geophys. Res.*, **86**(B6), 5039–5052.
- Zhang, S. & Karato, S.I., 1995. Lattice preferred orientation of olivine aggregates deformed in simple shear, *Nature*, **375**(6534), 774.
- Zhou, H.W., 1996. A high-resolution P wave model for the top 1200 km of the mantle, *J. geophys. Res.*, **101**(B12), 27791–27810.