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1. Introduction

The Paraná-Etendeka continental flood-basalt (CFB) province is one of the classical large igneous provinces, emplaced about 132 Ma [Renne et al., 1996]. The voluminous, mostly basaltic extrusions are thought to be caused by a mantle plume, whose surface expression can be tracked through the aseismic Walvis Ridge (Figure 1) to the present-day Tristan da Cunha (TdC) hotspot in the center of the South Atlantic Ocean [Richards et al., 1989]. However, the role of plume-lithosphere interaction during continental breakup and the opening of the South Atlantic Ocean is poorly understood.

Our study area (Figure 1) includes parts of the Congo and Kalahari cratons, the Damara mobile belt separating the two cratons, and the Kaoko mobile belt along the coast. This portion of Gondwanaland preserves rocks that represent the interaction between the Congo and Kalahari cratons in Africa with those in the Rio de la Plata craton in South America [Prave, 1996]. The Damara mobile belt was developed during the closure of an ocean basin at 550 Ma between the Congo craton and the Kalahari craton. The Kaoko belt is one of the coastal arms of the Damara orogen and consists of deformed Neoproterozoic rocks [Begg et al., 2009]. Continental rifting in our study area commenced in the Early Cretaceous [Begg et al., 2009] and is thought to be responsible for structural variations in the lithosphere and the upper mantle at the present-day continental margin in northwestern Namibia [e.g., Brune et al., 2013].
The study area, together with southern and eastern Africa, is characterized by anomalously high topography, called the African Super swell. The causes for the high elevations are currently not well understood but have been related to mantle processes such as lithospheric heating or dynamic topography [Nyblade and Robinson, 1994; Braun et al., 2014]. These processes are likely related to the African Superplume originating from one of the two global large low shear velocity provinces (LLSVPs) in the lowermost mantle [King and Ritsema, 2000; Garnero et al., 2007]. Torsvik et al. [2006] showed that most large igneous provinces, including Paraná- Etendeka, form above the border of LLSVPs. Therefore, Burke et al. [2008] proposed that these borders are plume-generation zones in the lowermost mantle. The TdC hotspot, located near the South Atlantic spreading ridge, is thought to be a surface expression of a mantle plume with its source in the lower mantle [e.g., Courtillot et al., 2003]. The mantle plume is interpreted to be responsible for the development of the Walvis Ridge and the Etendeka continental flood-basalt province [e.g., Torsvik et al, 2003]. The age progression of basaltic rocks along the 3000 km long Walvis Ridge is thought to result from plume-lithosphere interaction [e.g., O’Connor and Duncan, 1990; O’Connor et al., 2012]. Global seismic tomography failed to consistently observe the plume in this region throughout the entire mantle [e.g., Ritsema and Allen, 2003]. The missing plume signature in the lower mantle can be attributed to the ray theory used in most studies ignoring finite-frequency wave propagation effects, e.g., wave front healing [Nolet and Dahlen, 2000] and also to sparse distribution of seismic stations in the South Atlantic and surrounding continents. Nevertheless, based on a recent global seismic tomographic model using full waveform inversion [French and Romanowicz, 2015], the TdC mantle plume has been classified as clearly resolved.

While global seismic tomography [e.g., Ritsema and Allen, 2003; Begg et al., 2009; Schaeffer and Lebedev, 2013; French and Romanowicz, 2015] revealed large-scale seismic structures in the mantle beneath the South Atlantic region and constrained the thickness of the lithosphere [e.g., Pasyanos et al., 2014;
Steinberger, 2016], these studies do not provide the sufficient resolution in our study area, required for investigating the plume-lithosphere interaction, largely due to the lack of local seismological networks. Within the frame of a German collaborative research program (SAMPLE), a number of passive-source and controlled-source seismic experiments have been conducted onshore and offshore northwestern Namibia. The Walvis Ridge passive-source seismic experiment (WALPASS) has been operated from 2010 to 2012 and covered a large area of the continental-ocean transition [Heit et al., 2010] (Figure 1). The network has been designed to study the seismic structure of the crust and mantle lithosphere by diverse seismic methods and to constrain crust and mantle modifications caused by plume-lithosphere interaction at the initiation of a hotspot track.

Heit et al. [2015] studied crustal structure with receiver functions and found that the region where the Walvis Ridge intersects the African continent, is associated with overthickened crust, up to 45 km, and a high Vp/Vs ratio. These characteristics have been interpreted as evidence for magmatic underplating at the base of the crust, induced by plume activity. Controlled-source seismic data have also detected a high-velocity lower crust both onshore and offshore Namibia, which correlates well with the Etendeka basalt region [Ryberg et al., 2015; Fromm et al., 2015]. In this work, we focus on the structure of the lithosphere and the upper mantle down to the mantle transition zone (MTZ) and visualize the lithosphere-asthenosphere boundary (LAB). The depth of the LAB is an important parameter to decipher plume-lithosphere interaction during the Paraná-Etendeka CFB eruption in South America and Africa. The plume impingement upon the lithosphere and the continental rifting can cause lithospheric reworking and change the thickness of the lithosphere. We also present arrival times for phases converted at the 410 and 660 km MTZ discontinuities (abbreviated as 410 and 660, respectively). Arrival times of phases converted at the 410 and 660 are influenced by shallower velocity perturbations along the ray path and therefore carry useful information on the upper mantle seismic structure. By investigating the structure of the lithosphere under northwestern Namibia we aim to elucidate the role of the plume in reworking the mantle lithosphere as well as the geo-dynamic evolution of the Walvis Ridge following the breakup of the Gondwana continent.

2. Data and Methodology

The WALPASS seismic network consisted of 28 onshore stations in northwestern Namibia and 12 OBS stations offshore (see Figure 1 and Heit et al. [2010, 2015] for more network details and station coordinates). We have performed a receiver function study with teleseismic P and S waves (hereafter called PRF and SRF, respectively). In this study, only onshore stations are used since the offshore network does not provide useful receiver functions due to high noise level in the records. Data from four previous temporary broadband stations in the study area and a permanent station TSUM are also included in this study.

The aim of this work is to study the LAB and upper mantle discontinuities with both PRF and SRF. The PRF method is best suited for studying the Moho and the MTZ discontinuities. Converted phases from discontinuities in the depth range of 100–200 km where the LAB usually resides are often masked by crustal reverberations [Yuan et al., 2006]. In SRF, multiples are automatically separated from the primary conversions, therefore, the SRF method is potentially better suited to analyze the LAB. However, due to interference with some other phases and the existence of overcritical incidence, the quality and quantity of useful SRF data are limited [see, e.g., Yuan et al., 2006; Kind et al., 2012, for a review]. While the PRF method is well established, there are a number of technical challenges specific to SRF: (1) S waves are characterized by lower frequencies than P waves due to strong S wave attenuation in the mantle, resulting in low resolution. Therefore, the SRF method cannot resolve fine crustal structures, but it may be more useful for observing gradient transition zones, which are generally thought to be the case for the LAB. (2) The S-to-P converted waves can only be observed for teleseismic events in limited distance ranges, as they do not occur at post-critical incidence angles. The distance range, in which S-to-P conversions can be observed, depends mainly on the depth of the discontinuity. For observation of the LAB at depths up to 200 km, the useful distance range for SRF is about 60–85° for the S phase and 85–115° for the SKS phase. In addition, as also stated below, the S and SKS waveform data are noisier than those of the P waves, therefore, earthquakes with larger magnitudes are required. These limitations result in a smaller number of available SRFs compared to that of the PRFs. (3) Since S waves are not the first arrivals, they arrive within the P wave coda and may interfere with multiple mantle P waves. At distances of 80–90°, there is strong interference among the S, SKS, and
ScS phases. Therefore, Wilson et al. [2006] suggest to limit the distance range to less than 75°8 to avoid wave field interference.

Here we used earthquakes with magnitudes (mb) greater than 5.5 and at epicentral distances between 30°8 and 95°8 for calculation of PRF, while those with epicentral distances between 60° and 85° (S phase) and between 85° and 115° (SKS phase) were used for SRF (both S and SKS receiver functions are referred to as SRF). A band-pass filter of 1–30 s was applied to all seismograms prior to any other processing. Waveform data were visually inspected and manually selected, mainly according to (but not strictly following) signal/noise ratios, both on vertical and horizontal components. Seismograms exhibiting strong interference with other phases (e.g., for SRF, interference among S, SKS, and ScS phases at distances between 80°8 and 90°8) were discarded. Approximately 130 events were selected for both the PRF and SRF analysis (Figure 2c). The PRF computation was performed following the approach described by Yuan et al. [1997], while the SRF analysis was performed using the approach outlined by Kumar et al. [2006] and Yuan et al. [2006]. Coordinate rotation and deconvolution were then performed to compute PRF and SRF. The Z-N-E-component seismograms were rotated into the P-SV-SH system, where P-to-S and S-to-P conversions are polarized along the SV and P directions, respectively, and, therefore, are separated from their mother phases. Theoretical back azimuth and incidence angle were used for PRF, while an optimal incidence angle was chosen for SRF, with which the amplitudes on the SV component within a small window around the S wave arrival will be minimized by rotation. After rotation, we use a time-domain spiking deconvolution method to generate both PRF and SRF, which includes two steps. In the first step, an inverse filter is created by applying a least

![Figure 2.](image-url)
squares method to minimize the difference between the observed waveform and the desired delta-like spike function on the component, where the mother phase (P or S) is polarized. In the second step, a PRF or SRF is generated by convolving the inverse filter with the component, where the converted phases (P-to-S or S-to-P) are polarized. There is a practical difference between PRF and SRF. P-to-S phases arrive after the P wave, whereas S-to-P phases are precursors to S. Furthermore, an S-to-P conversion from the same discontinuity has a different sign than that of a P-to-S conversion. Therefore, an additional step in the SRF processing is to reverse the time axis and to change the sign of amplitudes, so that SRF can be directly compared with PRF. In total, we have obtained 2000 PRFs and 1300 SRFs.

We transform PRF and SRF to depth domain by the common conversion point (CCP) stacking method [Dueker and Sheehan, 1998; Kind et al., 2002]. Each amplitude in receiver function time series is projected to space within a Fresnel zone around the conversion point, calculated using a background model. Because there is no well-constrained P and S wave velocity model available for the study area, the global IASP91 reference model [Kennett and Engdahl, 1981] is used for time-depth conversion. This model provides a good approximation for the crustal thickness [Heit et al., 2015], but it may deviate from actual structure in the upper mantle. As shown in the next sections, the deviation of mantle structure from the 1-D model beneath our study area has resulted in arrival time anomalies of the MTZ conversions. By using the IASP91 model in our analysis, we can exactly quantify the time variations and infer seismic anomalies in the upper mantle.

Since the ray paths of S-to-P waves are flatter than the incident S wave, a critical incidence angle exists at certain depths, which prohibits S-to-P conversions at depths greater than that associated with the critical angle [Kind et al., 2012]. Most S-to-P conversions occur in the shallow upper mantle. Depending on epicentral distance, the path coverage reduces with depth below 200 km and disappears below 400 km. P-to-S converted waves, however, have much steeper incidence angles. These conversions can occur at much greater depths than the S-to-P conversions, thereby providing sampling of the entire upper mantle, including the MTZ. Therefore, the SRF profiles will be displayed from surface to 300 km depth, while the PRF profiles will be displayed to 800 km depth.

3. Results

Figure 2 shows stacked PRF and SRF at each station after moveout corrections for P-to-S (Figure 2a) and S-to-P (Figure 2b) converted phases, respectively. In each figure, stations are sorted from south to north. We applied the bootstrap resampling analysis to estimate the uncertainty in the receiver function stacks. For each station, a subset of 60% of total PRF or SRF traces is randomly selected to generate a sample stack. We repeated this process 100 times and then calculated the mean and standard deviation (σ) of the 100 generated sample stacks. The mean receiver function of each station is plotted along with a range of ±2σ from the mean, which can be considered as a rough estimate of the uncertainty (Figures 2a and 2b). The Moho converted phase (positive amplitudes at times between 3 and 6 s) is well recognized in PRF (Figure 2a) and SRF (Figure 2b) traces at most stations. Following the Moho conversion, a significant converted phase with negative polarity can be clearly detected in both PRF and SRF sections at many stations. The arrival time of this negative phase varies between 6 and 11 s. Compared to the Moho conversion, the negative phase shows more variations in both amplitude and arrival time. At some stations, the negative phase is weak, whereas at other stations, it is characterized by a complex signal with multiple troughs. The negative phase is clear and well separated from the Moho conversion, therefore, it is unlikely a sidelobe introduced by data processing and rather represents a robust signal from a different discontinuity in the subsurface. To test the azimuthal variation of this phase, we stacked the SRFs as a function of back azimuth (Figure 2d). However, except some variations in arrival time, we did not find strong azimuthal variation in phase polarity, suggesting that this negative phase mainly results from a boundary conversion with a relatively strong seismic impedance contrast, rather than from an interface of azimuthal anisotropy.

In Figure 3, we show three CCP-stacked SRF profiles, down to 300 km depth, where a pronounced negative discontinuity is observed at mantle depths between 70 and 120 km.

Two west-east profiles are presented: one located in the northern portion of our study area along the landfall of the Walvis Ridge (A-A’) and the other located in the southern portion near the southern edge of the seismic array (B-B’). A third profile (C-C’) runs parallel to the coast, approximately 100 km inland. All three profiles display prominent signals, including a positive phase at ~40 km depth followed by a negative
A low-velocity layer at ~80 km depth. The positive phase has been interpreted as the Moho conversion [Heit et al., 2015]. The negative phase shows more vertical variation and scatter, and it is interpreted as a relic feature generated by mantle plume-lithosphere interaction, where the thick lithospheric root was removed by the ascending plume (see next section for detailed discussion). We label this phase the paleo-LAB. Below the border between Angola and Namibia (Figure 3a, profile A-A'), it is located at 80 km depth in the east (14.5°E), but it deepens to 110 km depth beneath the Walvis Ridge. At a shallower depth (60–70 km), another negative phase is visible along the western portion of the profile (11.5–12.5°E). Along profile B-B' (Figure 3b), the paleo-LAB phase is observed at similar depths as those seen along profile A-A', but this phase displays more scatter. At the eastern termination of this profile, the paleo-LAB is found at 120 km depth. Along the north-south profile (C-C'), the paleo-LAB phase appears to be at a depth of 80–90 km (Figure 3c). In all three profiles, this low-velocity layer can be recognized as a strong converter. At greater depth a very weak negative phase appears to be visible at 170–180 km below profiles B-B' and C-C'. This much weaker phase does not seem to be as uniformly distributed as the shallower one.

In Figure 4, we present CCP-stacked PRF images down to a depth of 800 km. These are shown along the same profiles as those shown in Figure 3 for the SRF to facilitate comparison between the results. As mentioned before, the PRF also contains crustal reverberations, which usually dominate the depth range of the LAB, potentially masking any converted phases from the base of the lithosphere. Along all three profiles in Figure 4, the crustal multiples can be seen at depths between 100 and 200 km. The strongest conversion is associated with the Moho, at about 30–40 km depth [Heit et al., 2015]. Negative conversions can be clearly identified beneath the Moho, at similar depths as those seen in the SRF in Figure 3. These negative phases in the PRF, however, are less continuous compared to the SRF. Along profile A-A', two negative conversions
are discernible in the west (12–13°E), similar to Figure 3a. The shallower conversion is immediately beneath the Moho, at a depth of 60–70 km. The deeper interface is inclined toward the west, from depths of 80 to 120 km. Along the other two profiles, the negative conversions are located at depths of 70–80 km and show more variation in both amplitude and depth. The 410 and 660 km MTZ discontinuities can also be clearly observed. Using the IASP91 reference model, these boundaries are positioned at depths of 395 and 645 km, respectively, beneath nearly the entire study area, except in the northeast beneath permanent station TSUM (Figure 4b). There is a clear step of ~15 km along both discontinuities beneath profile B-B' at ~17°E. To the east of this point, the MTZ discontinuities appear at their expected depths (i.e., 410 and 660 km, respectively). Along profiles A-A' and C-C', these two discontinuities arrive consistently earlier and are found at apparent depths of 395 and 645 km, respectively.

Figure 4. (a–c) CCP stacked images of PRF within 100 km of each profile shown in plot (d). Profile locations are indicated in Figure 4d. PRF amplitudes converted within a Fresnel zone are stacked and a spatial smoothing window of 40 km in the horizontal and 4 km in the vertical directions are applied. Positive and negative amplitudes are color-coded red and blue, respectively. Converted phases from the Moho, the 410 and the 660 are clearly identified and marked in plot (a). Crustal multiples dominate the depth range between 100 and 200 km, masking possible converted phases from the LAB. At shallower depths, beneath the Moho, a negative phase indicates a low-velocity layer, which is interpreted as the paleo-LAB. All these phases are marked by dashed lines in each plot and labeled in plot (a). Early arrivals of the MTZ discontinuities are seen along all profiles. Along profile B-B', the MTZ discontinuities appear to switch back to expected depths beneath the easternmost portion of the profile. (d) Map of piercing points at depths of 35 km (red crosses), 410 km (green crosses), and 660 km (blue crosses) along with the locations of the three CCP profiles.
4. Discussion

4.1. The Paleo-LAB—Relic of the Thermal Erosion of the Lithosphere

We observe, both in PRF and SRF, a pronounced negative discontinuity in the shallow upper mantle at a depth less than 100 km beneath most of the study area. The discontinuity has a rugged topography, with strong lateral depth variation, and is characterized in some places by multiple interfaces at different depths. This discontinuity is too shallow to be interpreted as the present-day LAB, since global seismic tomography tends to show a thicker lithosphere beneath north Namibia [e.g., Pasyanos et al., 2014; Steinberger, 2016], although the tomographic resolution suffers from the scarcity of local seismic stations. A new surface wave tomography model, based on data recorded by the local WALPASS network, reveals a wide-spread high-velocity anomaly, covering most of northern Namibia and extending to a depth of 180 km (Pandey, personal communication). Our observation of the early arrival of both the 410 and 660 discontinuities further corroborates the existence of a thick and seismically fast lithosphere beneath our study area (this will be discussed in section 4.2).

Taking into account the history of mantle plume emplacement and continental breakup in our study area, we propose the following scenario. We suggest that the negative-polarity discontinuity at < 100 km depths is actually the remnant of mantle plume-lithosphere interaction. We refer to this negative conversion as the paleo-LAB and suggest that it originated as a consequence of the continental breakup, which was preserved during postrifting processes. The paleo-LAB could represent the depth that has been reached by the top of the plume after having eroded the originally thick lithospheric root [Sobolev et al., 2011]. It can be expected that large parts of the plume melts accumulate at the base of the remaining lithosphere, forming a thin layer of intensive melting. Some of the magmas may have intruded to the base of the crust leading to magmatic underplating. Thickening of the continental crust with elevated Vp/Vs ratio by magmatic underplating has been observed in the landfall area of the Walvis Ridge [Heit et al., 2015; Ryberg et al., 2015]. After cooling, the melts accumulated in the thin layer became crystallized. The discontinuous seismic velocity reduction could be achieved by a combination of consolidated plume melts as pyroxenites below the paleo-LAB [i.e., Rader et al., 2015] with intensively depleted and dehydrated peridotites above. The negative conversion observed at ~80 km depth should represent the upper boundary of this intensively crystallized melts. The lower boundary is not clearly visible in our data. A possible explanation is that this layer is too thin (less than a few kilometers) to allow both boundaries to be clearly resolved. Alternatively, the lower boundary is a gradient transition, as melt concentration reduces with increase of depth. In addition, the seismic velocity reduction within this thin layer has no significant influence on the overall seismic velocity in the entire lithosphere.

Additional indicators for thermal erosion of the lithospheric mantle beneath the study region have been derived by a xenolith study, which was conducted just to the south of our array. By analyzing garnet lherzolite xenoliths from kimberlite pipes at the southern border of the Damara Belt, Boyd et al. [2004] argued that the lithosphere must once have been as thick as that beneath the Kalahari craton (>200 km). They estimated that the formation depth of melting required to produce the peridotites in this region to be about 100 km (pressure <3 GPa), which are believed to reflect regional igneous activity associated with or following the thermal erosion of an originally thicker lithosphere.

On a broader scale, the Paraná CFB province is the South American counterpart to the Etendeka CFB province, and pronounced negative conversions have also been observed at similar depths (80–120 km) beneath Paraná. These negative conversions were interpreted as the LAB [Heit et al., 2007]. This interface below Brazil likely has a common origin with our interpreted paleo-LAB in Namibia. The thickness of the present-day lithosphere beneath the Paraná CFB region estimated by surface wave tomography is larger (about 150 km by Feng et al. [2007], about 200 km by Steinberger [2016], and up to about 250 km by Pasyanos et al. [2014]). Surface wave studies suffer from the same problem of station paucity in South America as in southern Africa, resulting in significant variability in estimating the lithospheric thickness among these studies. Nevertheless, these studies tend to indicate that the lithosphere beneath the Paraná is thicker than 150 km. We therefore suggest that the shallower interface at depths of 80–120 km seen by receiver functions [Heit et al., 2007] in eastern Brazil could also reflect the paleo-LAB.

As an alternative interpretation, we discuss whether the observed discontinuity could be described as a midlithospheric discontinuity (MLD).
In some cratonic regions, negative MLDs, have been observed at depths of around 100 km, much shallower than what would be expected for a cratonic LAB (see Rader et al. [2015] for a collection of observations). The definition of an MLD is rather diffuse, and every discontinuity that occupies middle portions of the lithosphere can theoretically be called an MLD. An MLD may represent lithospheric layering, characterized by a reduction of seismic velocities beneath it or it may be caused by variation in anisotropic properties [Yuan and Romanowicz, 2010; Sodoudi et al., 2013]. As a relatively sharp boundary, the MLD cannot be explained by a thermal gradient. For stable continents, Karato et al. [2015] proposed subsolidus deformation mechanisms that could be responsible for the sharp velocity drop at the MLD as well as a weaker seismic interface at LAB depths. In their model, the MLD is the result of elastic accommodation due to grain-boundary sliding; therefore, its depth is sensitive to temperature and water content. Selway et al. [2015] indicate that the grain-boundary sliding estimate of Karato et al. [2015] would not lead to a sufficient velocity decrease to explain observations. They also rule out anisotropy variations. Instead, they suggested that the existence of an MLD could be attributed to long-period filtering that combines multiple small signals into a single layer. They also suggest that these layers might be rich in amphibole and have local or regional extent in some cases, but they rule out the existence of such a layer on a global scale. Rader et al. [2015] postulated that the MLD may represent a layer of crystallized melt that was formed as the lithosphere cooled down. They attribute the MLD to a chemical interface related to the paleo-intersection of a volatile-rich solidus and progressively cooling lithosphere, and they suggest that this phase may be related to the early evolution of a young lithosphere.

As stated previously, our study area is located at the border of a craton and has experienced significant modifications due to lithospheric erosion and plume-lithosphere interaction. In other words, the area was not stable enough to enhance elastic accommodation as in the case suggested by Karato et al. [2015] for MLD development. It is also important to note that our study area was deformed by subduction and collision during the Neoproterozoic [de Wit et al. 2008] and by extension of the continental lithosphere during the Cretaceous prior to emplacement of the plume and breakup [Heine et al., 2013]. Hence, the alternative interpretation of the discontinuity as an MLD seems problematic.

4.2. Early Arrival of the MTZ Discontinuity Phases Implying High Seismic Velocities in the Upper Mantle

The 410 and 660 km discontinuities define the top and bottom of the MTZ. The boundaries are generally thought to reflect mineralogical phase transformations of olivine to higher-pressure minerals, which create sharp vertical gradients in density and seismic velocities. The 410 marks the transition from olivine to wadsleyite, and the 660 marks the transition from ringwoodite to perovskite (also known as bridgmanite) plus magnesiowüstite. Experimental studies have shown that both transformations are subject to temperature variation and have Clapeyron slopes of opposite signs (reviewed by Helffrich [2000]). In the absence of other effects, such as changes in mantle composition, water content, or metastability caused by either fast-decending cold subduction or hot upwelling plume, a lateral increase in temperature can cause the 410 km discontinuity to deepen and the 660 km discontinuity to rise (or vice versa). It is expected that a temperature perturbation of 80°C may result in a 10 km variation in the transition zone thickness [Helffrich, 2000]. However, arrival times of the mantle discontinuity phases can also be significantly influenced by the average velocities in the crust and upper mantle. As a result, the two discontinuities may apparently be shifted in the same direction in time series receiver functions. For example, receiver function profiles traversing the central Tibetan plateau have shown a simultaneous apparent deepening of the MTZ discontinuities from south to north Tibet, which has been interpreted as evidence for low velocities in the upper mantle beneath northern Tibet [Kind et al., 2002; Zhao et al., 2010].

Beneath our study area, the 410 and 660 are clearly observed in the PRF migrated sections (Figure 4); however, they are imaged at apparent depths of about 395 and 645 km, respectively, which is 15 km shallower than global average values. At 17°E, both discontinuities jump to greater depths (Figure 4b, profile B-B'). Toward the eastern end of profile B-B’, beneath permanent station TSUM, both discontinuities are observed at depths of 415 and 662 km, respectively. Since receiver functions are time series, the time-depth conversion in the CCP stacking profiles depends on the velocity model used. However, the MTZ thickness, defined by the depth difference between 410 and 660, is independent of the upper mantle velocity variations. Our results indicate a constant MTZ thickness, implying that there are no major variations in the MTZ.
temperature. Hence, the apparent changes in their depths in the same direction and with roughly the same magnitude must be a result of velocity variations in the upper mantle.

In Figure 5, we show a waveform comparison of receiver function stacks for the WALPASS network versus that for permanent station TSUM. Prior to stacking, all receiver functions were moveout-corrected for a reference slowness of 6.4 s/degree. Station TSUM provides 20 years of high quality data and allows for reliable identification of the MTZ discontinuities. We performed the same bootstrap resampling analysis as previously done in Figure 2 to estimate the uncertainty of the waveform stacks. In Figure 5, we plotted the mean PRF along with the error bounds, defined by $\pm 2\sigma$ from the mean. It is clearly observed that the 410 and 660 P-to-S conversions arriving 1.5 s earlier at the WALPASS network (42.6 and 66.9 s) than at station TSUM (44.6 and 68.3 s). The arrival times of the two transition zone discontinuities at TSUM are close to those predicted by the IASP91 reference model (44.1 and 68.1 s). We also note that the arrival times are referred to the reference slowness used for the moveout correction and that these time values will vary for other slowness values.

In the western part of our study area, where rifting and continental breakup occurred, changes in lithospheric thickness, composition, and hydration state are expected to produce higher velocities in the upper mantle that could lead to anomalous arrivals of the MTZ conversions. However, it is unclear from our data, where the required high-velocity anomaly would be located. Early arrivals of the MTZ phases can be related to velocity increases in the lithosphere or to an increase in the lithospheric thickness. Alternatively, the required high velocity could also reside in the sublithospheric mantle. In Figure 6, we try to quantify the effect of possible velocity perturbations on the arrival times of the MTZ discontinuity phases. Since the 410 and 660 change simultaneously, we focus on the P410s arrival time in our analysis. We consider two depth ranges for our calculation: an approximate depth range for the lithosphere (0–200 km) and a depth range encompassing the entire upper mantle (0–400 km).

Cammarano et al. (2003) showed that seismic velocities in the mantle lithosphere are more sensitive to temperature variations than to composition changes. The estimated sensitivities of $V_p$ and $V_s$ to temperature at a depth of 200 km are $-0.75 \pm 0.15\%$ and $1.30 \pm 0.30\%$ per 100°C, respectively [Cammarano et al., 2003], resulting in $\Delta \ln V_s/\Delta \ln V_p$ (ratio of $S$ wave velocity perturbation versus $P$ wave velocity perturbation) of 1.73. In our calculation, we use the IASP91 model as our initial model, and we use a $\Delta \ln V_s/\Delta \ln V_p$ of 1.73 to perturb the $V_p$ and $V_s$. We find that a 1.5 s early arrival of the P410s phase can be acquired by a 2.9% and 5% increase in $V_p$ and $V_s$, respectively, within the lithosphere or by a 1.5% and 2.6% in $V_p$ and $V_s$, respectively, in the entire upper mantle (Figure 6a).

A possible scenario is that most of the study area is underlain by a thick mantle lithosphere with increased seismic velocities similar to intact cratonic areas. A thick lithosphere beneath northern Namibia seems to be confirmed by global seismic tomography studies [Pasyanos et al., 2014; Steinberger, 2016]. Additionally, a wide-spread high-velocity anomaly, extending beneath much of northern Namibia and extending to...
180 km depth is also observed by a new surface wave tomography model developed with data from the local WALPASS seismic network (Pandey, personal communication). In the same model, high velocities seem to extend much deeper to more than 300 km depth.

To verify our model calculation, we reproduced the PRF cross section B-B' by using a varying background model (Figure 6b) to account for seismic anomalies in the upper mantle. We used the IASP91 model beneath the eastern portion (station TSUM), while beneath the eastern portion (WALPASS), a modified model is used as indicated by the red circle in Figure 6a, where Vp and Vs in the upper 200 km are increased by 2.9% and 5%, respectively. The resulted profile shows that the seismic anomalies in the upper mantle are completely accounted for and both the 410 and 660 are imaged at their expected depths. Similar result can be obtained by using another model involving the entire upper mantle, as indicated in Figure 6a by the blue circle.

In Figure 7, we compare global 410 arrival times with lithospheric thickness. The global map of lithospheric thickness is from the LITHOS1.0 model [Pasyanos et al., 2014]. The 410 arrival times are taken from global studies of the MTZ [Li et al., 2003; Shen et al., 2014]. Thick lithosphere is found beneath most cratonic regions, including our study area (Figures 7a and 7c). While the 410 time variations also contain information about topography along the top of the MTZ, the correlation between the 410 arrival times and the thickness of the lithosphere is discernible. Throughout most of the world, the 410 conversion arrives earlier where the lithosphere is thicker or vice versa (Figure 7a). A similar 410 phase time anomaly can be observed within the Kaapvaal craton in southern Africa, where a thick lithospheric root exists (Figure 7c). At one station, the 410 time anomaly is twice as large compared to nearby stations, which may suggest a lateral variation in either the lithospheric thickness or seismic velocities. Gao et al. (2002) observed similar arrival time anomalies of the MTZ discontinuity phases with data from a temporary seismic array in South Africa. Using the IASP91 reference model, they located the 410 and 660 at average depths of 396 and 641 km, respectively (Figure 7c). Located between northwestern Namibia and the Kaapvaal craton, station TSUM is characterized by a very small time anomaly that arrives 0.5 s later.

Variations of the 410 arrival times can be induced by topographic variations along the top of the MTZ as well as by upper mantle velocity perturbations. We used Clapeyron slopes for the 410 and 660 to estimate...
the possible topography along the top of the MTZ to remove any associated effects on the global 410 arrival times. We measure the differential times between the 410 and 660 and use them as a proxy for MTZ thickness. We then correct for topographic variation along the 410 using the 410 arrival time data, assuming Clapeyron slopes of \(1.3\) and \(2.2\) MPa/K for the 410 and 660, respectively [Helffrich, 2000]. Clapeyron slope is the partial derivative of pressure (\(P\)) with respect to temperature (\(T\)), \(\partial P/\partial T\). The relationship between variations in pressure and depth follows \(\partial P/\partial T = \rho g \partial z/\partial T\), where \(\rho\) is the density, \(g\) is the gravitational acceleration, and \(z\) is the depth. Using the Clapeyron slopes given above and densities in the IASP91 model, the MTZ thickness variations can be accommodated by topography along the top and bottom of the MTZ by 57% and 43%, respectively, assuming similar temperature perturbations at the two discontinuities, as is the case within mantle plumes and beneath subduction zones. We hence assign 57% of the MTZ thickness variation to the 410 topography and 43% to the 660 topography (with opposite signs). The remaining

Figure 7. (a) Global distribution of the 410 phase arrival times [Li et al., 2003; Shen et al., 2014] superimposed on a map of lithospheric thickness from the LITHOS1.0 model [Pasyanos et al., 2014]. In general, thick lithosphere (blue) correlates with early arriving 410 phases (circles). Phase arrival times are predicted by the IASP91 model. The anomaly associated with the WALPASS array is indicated by the red circle. We note that cratonic South Africa is also characterized by early 410 arrivals. (b) Global 410 arrival times versus lithospheric thickness. The observed times \(t_{410s}\) are corrected for topography along the top of the MTZ (see text for further details). The red solid line indicates the linear regression of the 410 times, while the red dashed lines mark \(\pm 2\) s from the regression line. (c) Enlarged map of southern Africa. Station TSUM (small yellow cross) is located between two regions (northwestern Namibia and the Kaapvaal craton) underlain by thick lithosphere. The average 410 arrival times for the WALPASS (WP) and the Kaapvaal (KV) networks [Gao et al., 2002] (yellow circles) are plotted at their central locations.
Variabilities of the 410 arrival times, as shown in Figure 7b, should mainly reflect the velocity perturbations in the upper mantle.

A diagram of the 410 arrival time versus lithospheric thickness (Figure 7b) shows a roughly linear relationship albeit with very large data scatter. Most of the 410 times fall within a range of ±2 s from the linear regression. Possible reasons for the large scatter include measurement error of 410 times (usually less than 0.5 s), errors in the lithospheric thickness map and lateral variations of seismic velocities in the upper mantle, which are not considered in the calculation. In addition, there are also errors in the correction of the 410 topography described above. Topography of the MTZ discontinuities does not necessarily follow the same theoretical Clapeyron slopes used here, as differences in the mineral composition also have effects on the MTZ topography.

In our data, the 410 arrival time anomaly disappears to the east beneath station TSUM, where the lithospheric thickness was previously estimated to be 120 km [Kumar et al., 2007]. TSUM is located at the southern edge of the Congo craton, which is poorly constrained at lithospheric depths. Although there are no measurements of the 410 arrival times within the Congo craton, the similar 410 time anomaly observed in the Kaapvaal craton [Gao et al., 2002] may imply that northwestern Namibia is underlain by thick lithosphere, similar in thickness to that observed beneath the Kaapvaal craton. Station TSUM is located between the plume-affected northwestern Namibia area to the west and the cratonic units to the east. Furthermore, the correlation of the 410 time at TSUM with that predicted by the IASP91 reference model suggests that the seismic structure beneath TSUM can be well represented by a model with an intermediate lithospheric thickness. In some global tomographic models, the thick lithosphere beneath TSUM may possibly reflect the poor resolution due to the scarcity of seismic stations in this region, and hence, a gap (i.e., a region of thin lithosphere) beneath TSUM might not be resolved by these models. The LITHO1.0 model [Pasyanos et al., 2014] displays more details than other existing global lithospheric thickness models and features an isolated region of thick lithosphere beneath our study area with thinner lithosphere beneath TSUM and further to the east.

As an alternative explanation, variations in arrival times from the MTZ discontinuities may also be partially caused by high velocities in the asthenosphere. This scenario could indicate a regional downwelling of colder material at the step in lithosphere thickness along the continent-ocean transition, possibly initiated by edge-driven convection [King and Anderson, 1998]. High-velocity anomalies of 1–2% have been observed with seismic tomography beneath the Atlantic coasts of Africa and South America at asthenospheric depths and have been interpreted as small-scale, edge-driven convection [King and Ritsema, 2000]. High-velocity sublithospheric structure beneath the western margin of the African continent could partly contribute to the early arrivals of the MTZ phases. This edge-driven convection should terminate west of TSUM, where the MTZ arrival time anomalies disappear.

4.3. Melt Depletion and Dehydration—Reconstruction of a Thick Lithosphere

The observations discussed above likely imply the existence of a thin lithosphere at the time of continental breakup and that of a thickened lithosphere today.

In this section, we first investigate the hypothesis that the increasing depth of the LAB is simply generated due to continuous lithospheric cooling using a 1-D numerical modeling approach (Figure 8). In order to do so we solve the 1-D heat equation accounting for radiogenic crustal contributions (1.7 μW/m³ in the upper crust and 2 μW/m³ in the lower crust [Artemieva and Mooney, 2001]). The thickness of the upper and lower crusts is assumed to be 20 and 25 km, respectively [Heit et al., 2015; Ryberg et al., 2015]. The temperature of the LAB is assumed to be between 1150°C and 1300°C. Thermal diffusivity is set to 10^-6 m²/s. Boundary conditions of 0°C at the surface and 1475°C at 250 km depth correspond to a mantle potential temperature of 1350°C and an adiabatic gradient of 0.5 K/km. The initial condition assumes that the plume head eroded the lithosphere to a depth of 80 km occurring so rapidly that the temperature profile of the remaining lithosphere remained undisturbed. During subsequent thermal equilibration, it is assumed that the plume does not have any influence since (i) the temperature anomaly due to the plume head has been diffused or (ii) the plume tail is far away from our region of interest. Results of the numerical modeling (Figure 8) show that after thermal erosion, the LAB cannot reach the expected depth of 200 km by conductive cooling of the lithosphere alone. For instance, if the LAB originally had a depth of 80 km, the present-day LAB should not be deeper than 135–175 km, depending on the equivalent temperature assumed in the
calculation. In a more realistic setting that includes delayed cooling due to small-scale convection, the present-day lithospheric thickness would be even less.

The cartoon in Figure 9 summarizes our observations and preferred interpretation. The observed apparent topography of the 410 and 660 mirrors the present-day topography of the LAB. A thick lithosphere beneath our study area may either reflect a pre-existing feature present at the time of breakup or one that developed after the plume-lithosphere interaction. Both the MTZ arrival times and the LITHO1.0 model of Pasyanos et al. [2014] indicate that the region of thick lithosphere in northwestern Namibia seems to be rather localized and does not extend much beyond our study area at the landfall of the Walvis Ridge (Figure 7c). Instead, it appears to be detached from the Kalahari and Congo cratons which could be an indication of mantle plume influence. Interestingly, the LITHO1.0 model also shows thick lithosphere beneath the Paraná flood basalts; however, the opposite is observed beneath the younger Deccan traps, where the lithosphere is about 100 km thick. We suggest that the lithosphere beneath northwestern Namibia was thinned in the Cretaceous by a mantle plume, marked by the depth of the paleo-LAB, and that the lithosphere has increased the thickness over the past 132 million years.

Melting from a large thermal plume may contribute to the formation of thick lithosphere [e.g., Arndt et al., 2009]. High plume temperatures lead to high degrees of melting and melt extraction. The pre-existing lithosphere could have first been eroded by the plume [Sobolev et al., 2011] to the depth of the interpreted paleo-LAB, but subsequently, it may have thickened again due to accretion of depleted peridotitic material after the extraction of melt. Melt-induced dehydration increases both the viscosity and the seismic velocity of the residual mantle. Depleted residual mantle has lower density and is significantly more viscous than the original mantle plume material (or than normal upper mantle), therefore, it would have become part of the overlying rheological lithosphere [Hall and Kincaid, 2003]. In this way, a viscous boundary that is effectively equivalent to a thickened lithosphere could have been created [Hall and Kincaid, 2003].

Plume melting produces a highly melt-depleted, dehydrated, buoyant, and viscous chemical boundary layer that could potentially result in the formation of a thick lithospheric keel. The effects of thermal erosion and lithospheric thinning as well as subsequent formation of depleted mantle lithosphere should be reflected as a

Figure 8. 1-D numerical solution of the heat equation. Model results illustrate that lithosphere, which was thinned to 80 km due to thermal erosion by a plume cannot reach the expected LAB depth of ~200 km by conductive cooling of the lithosphere alone. (a) Initial condition (grey) and modeled present-day temperature profile (red). (b) Time-dependent evolution of 1150°C isotherm (orange) and 1300°C isotherm (green), which are used as end-member proxies for the thermal LAB.
seismic velocity increase beneath the area [e.g., Fernández et al., 2009; O’Reilly et al., 2009]. Numerical modeling by Fernández et al. [2009], integrating data from petrology and mineral physics with geophysical data, suggests that the excess magma generation associated with the TdC plume resulted in more depleted mantle and is responsible for variations in the depth distribution of P wave and S wave velocities when compared to standard mantle composition.

Because seismic velocities in the mantle are more dependent on temperature variations than on chemical composition, melt depletion alone is insufficient to produce the high mantle velocities imaged beneath cratons or to cause significant velocity variations [Schutt and Lesher, 2006]. However, loss of volatiles in the depleted mantle may occur during plume melting as shown by Karato and Jung [1998] who provide a connection between low water content and high seismic velocity. They state that water significantly reduces seismic wave velocities through anelastic relaxation. Therefore, plume melting and melt extraction will have an effect to increase seismic wave velocities through anelastic relaxation. Therefore, plume melting and melt extraction will have an effect to increase seismic wave velocities after cooling through dehydration of anhydrous minerals. The depleted residual plume materials are more buoyant; hence, they may be stably attached to the thermally eroded lithosphere, forming the thick lithosphere seen today.

5. Conclusions

Receiver functions indicate a strong widespread negative conversion at an average depth of ~80 km beneath northwestern Namibia. This discontinuity is interpreted as a paleo-LAB that formed during the CFB
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