



Plume-lithosphere interaction beneath southwestern Africa – Insights from multi-mode Rayleigh wave tomography

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ARTICLE INFO

Keywords:

Waveform inversion

Rayleigh wave

Fundamental and higher-mode

Craton

Etendeka continental flood basalt

Plume-lithosphere interaction

ABSTRACT

A stationary mantle plume heats the overlying lithosphere from below, generating melts and causing lithospheric erosion. When the lithosphere moves away from the plume, it can be re-established by melt residues that have been since attached, partly due to cooling, at the base of the lithosphere. The Etendeka flood basalt region in northwest Namibia used to overlie the Tristan da Cunha mantle plume in the Early Cretaceous and thus provides an ideal place to examine plume-lithosphere interaction. Here we determine the upper mantle shear wave velocity structure of southern Africa down to 400 km depth by waveform inversion of multi-mode Rayleigh waves in order to find the imprints left by the Tristan da Cunha mantle plume. Thick lithosphere with high shear wave velocities is observed beneath the Congo and Kalahari Cratons, extending down to 200 km depth with the largest thickness beneath the Limpopo Belt. Along the landfall of the Walvis Ridge and Damara Belt, our model reveals a thick lithosphere down to 100–200 km depths. The thick lithosphere seems to constitute the Congo Craton, but the thick Congo cratonic lithosphere is extended farther south than observed in most of the previous models. We propose that the lithosphere along the landward extension of the Walvis Ridge was affected by magmatic processes related to the plume, but has been partially reconstructed since then by melt depletion and successive cooling. A high velocity anomaly beneath the northwest coast of Namibia down to a depth of ~80 km coincides with distribution of the Etendeka basalt at the surface and the seaward-dipping reflectors at the continental margin. At sub-lithospheric depths, a low-velocity anomaly (at 100–200 km depth) and a high-velocity anomaly (at 250–350 km depth) can be detected. The origin of these anomalies could be related to an ongoing edge-driven convection process.

1. Introduction

The lithosphere architecture of southern Africa is an assembly of several stable Archean cratonic fragments stitched together with Proterozoic Pan-African orogenic belts. The major blocks are the Kalahari and the Congo Cratons separated by the Damara Belt, a Pan-African orogenic belt developed around 550 Ma (e.g. Begg et al., 2009) during the closure of an ocean basin (Fig. 1). The Kalahari Craton itself is constituted, in its core, by the Kaapvaal Craton (assembled before 3.0 Ga), which collided with the Zimbabwe Craton to the north at 2.0–2.8 Ga, forming the Limpopo orogenic belt (e.g. de Wit et al., 1992). Along the southern margin of the Kalahari Craton, the Namaqua-Natal metamorphic belt was formed later until about 1.0 Ga (Eglington, 2006). In

this paper we refer to the Kalahari Craton as its Archean core that includes the Kaapvaal and Zimbabwe Cratons and the Limpopo Belt, although the original Kalahari Craton extended over a much larger area during the Paleoproterozoic, covering most of southern Africa and northern Antarctica (Jacobs et al., 2008). Being a mosaic of different cratons, the lithospheric constituents of southern Africa are considered to be among the most stable and oldest cratonic regions of the world (Linol et al., 2015).

The lithosphere in this region has been dramatically modified by the final stage of the Gondwana break-up in the Early Cretaceous. Occurrences of diamondiferous kimberlites indicate a 150–200 km thick lithosphere at time of eruption (Boyd et al., 1985). Evidences show, however, that the lithosphere in southern Africa is not of the original

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<https://doi.org/10.1016/j.tecto.2022.229587>

Received 29 January 2022; Received in revised form 9 September 2022; Accepted 16 September 2022

Available online 23 September 2022

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state, but has encountered significant reworking. The widespread volcanic eruptions in the western margin of southern Africa indicate magmatic processes in the lithosphere (Fig. 1). The most elevated topography, developed since the late Mesozoic, is isostatically supported by sub-crustal lithospheric mantle (Hu et al., 2018). Layered seismic anisotropy in the cratonic lithosphere suggests different origin. While a frozen-in anisotropy exists in the upper layer, the lower layer is polarized with alignment to the Cenozoic mantle flow (Hu et al., 2018), suggesting a reconstructed lower lithosphere.

The break-up of South America from Africa led to the opening of the South Atlantic Ocean. The process of lithosphere weakening that allows the final breaking is still under debate (Brune et al., 2013). Several mechanisms have been proposed to explain the break-up, ranging from plate boundary forces, mantle drag and gravitational potential. Another factor, though not the prerequisite, is the mantle plume, which could have helped to trigger the rifting (Brune et al., 2013). The arrival of a plume head at the base of the lithosphere often involves the production and release of large amounts of melts which intrude into the lithosphere and partly erupt at the Earth's surface. This is evident in the present day by the relic of a large igneous province (LIP) that includes both the Paraná (in South America) and the Etendeka (in northern Namibia) continental flood basalt (CFB) provinces, which are associated with the Tristan da Cunha hotspot and with the opening of the South Atlantic Ocean (Peate, 1997; Coffin et al., 2006; Sebai et al., 2006; see Fig. 1). The Walvis Ridge represents the track of the plume with age progression of basaltic rocks (Gibson et al., 2005; O'Connor et al., 2012; Fromm et al., 2017) between the Etendeka CFB province and the current position of the Tristan da Cunha hotspot, whereas no continuous hotspot track links to the Paraná CFB province in the South America except the Rio Grande Rise (Fig. 1). Northern Namibia and the Walvis Ridge are, therefore, an ideal place to investigate the role of the plume-lithosphere interaction during the break-up of the South Atlantic Ocean (Gibson et al., 2005; O'Connor et al., 2012; Fromm et al., 2017). Given the large extent of the Etendeka (and Paraná) CFB province, the mantle lithosphere in northwestern Namibia might have been also significantly modified by the mantle plume. An open question is, whether the effects

of the plume could have survived in the deep lithosphere for 130 million years. Temperature differences may disappear (e.g., Ito and Lin, 1995), while compositional differences may persist (e.g., Calvert et al., 1995; Tesauro et al., 2020). A further question is whether the break-up of Gondwana (South America and Africa) was triggered by the mantle plume or if mantle plume emplacement is just a secondary effect that took place during the opening of the South Atlantic Ocean.

Previous seismic tomography studies mapped the thickness of the lithosphere beneath southern Africa. Although all models show a thick lithosphere beneath the major cratons, due to sparse and uneven distribution of seismic stations in South Atlantic and Africa, the mantle lithosphere was better constrained beneath the Kaapvaal Craton where dense station arrays were available. The Kalahari Craton is identified as a thick and high seismic velocity feature in seismically imaged lithosphere models, although its thickness varies in different models derived from various methods. Global and regional seismic tomographic models using teleseismic body waves have identified a lithosphere keel to a depth of >250 km (Woodhouse and Dziewonski, 1984; Ritsema et al., 1999; Ritsema and van Heijst, 2000; James et al., 2001; Romanowicz and Gung, 2002; Kustowski et al., 2008; Begg et al., 2009; Ortiz et al., 2019; White-Gaynor et al., 2020; Boyce et al., 2021). The thickness of the lithosphere may be overestimated by seismic body waves due to vertical smearing along ray paths, while surface waves are more sensitive to the vertical velocity variations (Priestley and Tilmann, 2009). In most recent surface wave tomographic models (Li and Burke, 2006; Sebai et al., 2006; Priestley et al., 2008; Fishwick, 2010; Adam and Lebedev, 2012; Emry et al., 2019; Celli et al., 2020) the lithosphere thickness of the Kaapvaal Craton is estimated to be about 200 km. The thickness of the Congo Craton is less well constrained than that of the Kalahari Craton. As illustrated in Figs. S1 and S2 and also shown by Celli et al. (2020), the shape of the thick Congo cratonic mantle lithosphere, e.g., its southern and western extent, varies among different tomographic models. In some models (e.g., Sebai et al., 2006; Pasyanos and Nyblade, 2007; Emry et al., 2019; Celli et al., 2020), thick lithosphere was imaged beneath the southern margin of the Congo Craton and towards the Etendeka CFB and Walvis Ridge regions, whereas in other models the

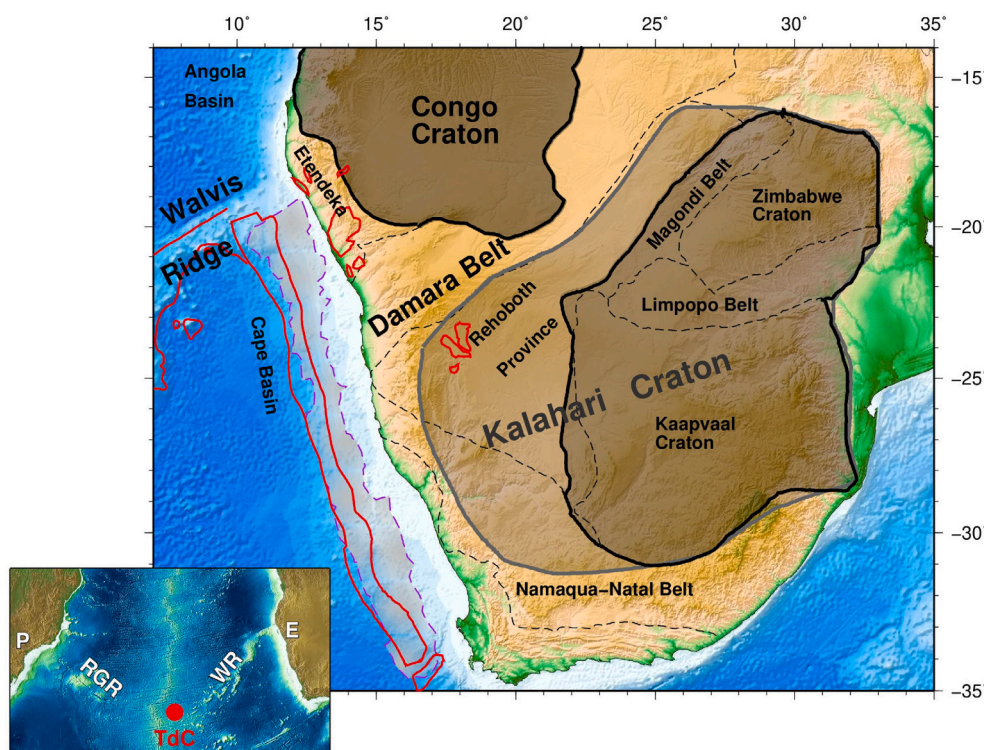


Fig. 1. Topographic map of southern Africa with major tectonic units (after Yu et al., 2015). Black dashed lines depict major tectonic boundaries. Thick dark shaded areas with black solid outlines mark the Congo Craton and the Archean core of the Kalahari Craton, including mainly the Kaapvaal and Zimbabwe Cratons (Jacobs et al., 2008). The light shaded area with gray outline roughly marks the Palaeoproterozoic Kalahari Craton. Red lines indicate the Large Igneous Provinces (LIP) with ages <150 Ma (Coffin et al., 2006). The gray shaded area offshore with magenta dashed outline marks the regions of seaward dipping reflectors (SDR) (Koopmann et al., 2014). The inset map shows the geographic locations of the Tristan da Cunha (TdC) hotspot, Walvis Ridge (WR), Rio Grande Rise (RGR), Etendeka (E) and Paraná (P) continental flood basalt area in southern African and South America, respectively. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

thick cratonic lithosphere beneath the Congo Craton is limited farther to the north and east. The deep lithosphere structures beneath the Etendeka CFB and Walvis Ridge regions still remain elusive.

Recent seismic studies using active or passive sources observed seismic anomalies in the crust and upper mantle related to the plume-lithosphere interaction along the Walvis Ridge. [Fromm et al. \(2015, 2017\)](#) observed a thick crust (~30 km) with high-velocity lower crust beneath the Walvis Ridge by wide-angle reflections and refractions, which is confined by the continent-ocean transition (COT), and interpreted this as a result of magmatic underplating by plume related melts. The crustal thickness increases further landwards along the Walvis Ridge to ~45 km beneath the Etendeka CFB region ([Heit et al., 2015](#)). However, these studies concluded that the impacted area is too small for a mantle plume to trigger the continental break-up. [Yuan et al. \(2017\)](#) interpreted the observed anomalously early arrivals of the mantle transition zone discontinuity phases in the Walvis Ridge and Etendeka CFB areas as being caused by high-velocity anomalies in the upper mantle and associated them to the modification of the lithosphere by the plume in which the lithospheric mantle was thinned during the continental break-up by thermal erosion and thickened again by melt depletion.

In this work we integrated the recently deployed seismic networks in the region around Tristan da Cunha ([Geissler et al., 2016](#)) and in Northwest Namibia ([Heit et al., 2010](#)) with other available permanent and temporary stations in southern Africa (including parts of the Kalahari Craton and the southern tip of the Congo Craton) and the Atlantic Ocean and performed a Rayleigh wave tomographic inversion by

waveform modeling. In addition to the fundamental mode surface waves, we also modeled the higher modes up to the fourth order. In this way, the vertical resolution has been greatly increased down to a depth of 400 km. Our results image the lithospheric structure beneath southern Africa in unprecedented detail and provide key results that could help improve our understanding of the mechanisms by which mantle plume can modify the overlying lithosphere.

2. Data and method

2.1. Waveform data

We use the Rayleigh waveforms from vertical component seismograms to model the fundamental and higher modes up to the fourth order. Including higher modes in the waveform modeling can provide additional sensitivity for larger depths (e.g., [Pandey et al., 2014](#)). Waveform data from >150 permanent and temporary stations ([Fig. 2](#)), operational at different periods between 1997 and 2014, have been requested from GEOFON or IRIS data centers. The time periods for the permanent stations are from 2006 to 2014 for networks AF, G, GE, GT, II and IU (except some AF stations which starts in 2010). The temporary stations include the SASEK experiment deployed across the Kaapvaal Craton (XA:1997–1999; [Silver, 1997](#)), the GEOFON Namibia Temporary Network, consisting of five broadband stations, deployed in the Damara Belt (XC:1998–1999; [Kind, 2000](#)), the Walvis Ridge passive-source seismic experiment WALPASS (6A:2010–2012; [Heit et al., 2010](#)) and the Tristan da Cunha OBS experiment (TdC:2012–2013; [Geissler et al.,](#)

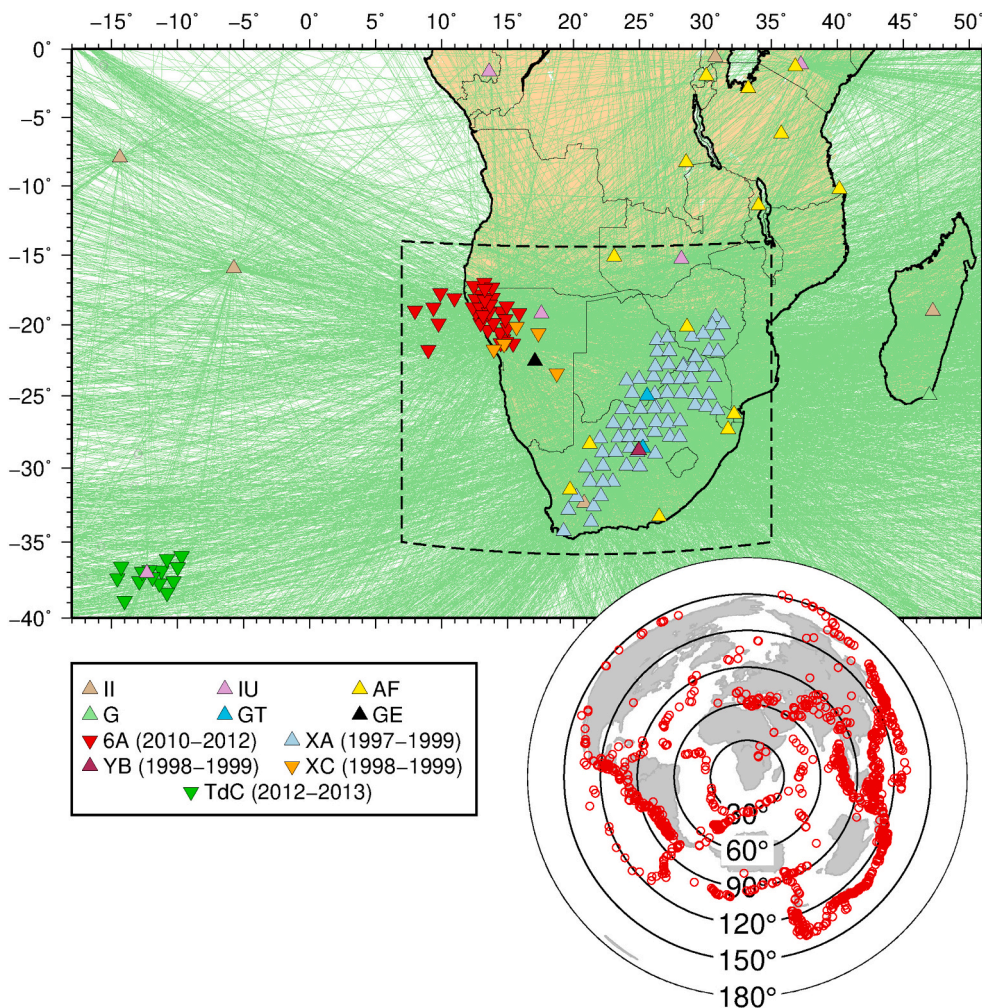


Fig. 2. Map of temporary and permanent stations used in this study. Green lines denote the surface wave great circle paths from earthquakes to stations, which are finally selected and used in the inversion. Legend box shows the FDSN network codes for the permanent and temporary stations and the Tristan da Cunha (TdC) network. Dashed lines outline the study area. The inset map shows the distribution of teleseismic earthquakes used in the surface wave analysis. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

2016). Compared to previous surface wave tomographic studies (e.g., Priestley et al., 2008), the present work includes the local WALPASS and Tristan da Cunha seismic networks, which enables us to achieve a dense path coverage in the key study area (Figs. 2 and S3). The Global CMT earthquake catalog was used to generate the event list and to obtain the source information. Earthquakes with magnitude (M_w) larger than 5 and epicentral distance larger than 700 km were selected. Waveform data are requested in full Seed volume with a length of 2 h. Instrument responses were not removed from the waveforms but were used in calculating synthetic seismograms. Since our method is fully automated, we do not check visually the waveforms at this stage. Our automated approach applies strict selection criteria before inversion.

2.2. Tomographic inversion

The technique used for constructing a 3-D SV velocity model has two distinct stages and was previously employed in a number of regional scale surface wave tomography studies (e.g., Debayle, 1999; Debayle and Kennett, 2000; Pilidou et al., 2004; Priestley et al., 2006; Pandey et al., 2014). The first stage involves modeling each waveform for a 1-D path averaged SV wave velocity model from a source to a receiver. This is done using an automated waveform inversion algorithm (Debayle, 1999), which is based on the approach of Cara and L  v  que (1987), where a set of secondary observables is used to model each multimode Rayleigh waveform. In the second stage we combine all these 1-D velocity models to achieve a final 3-D velocity model. Our model covers an area from latitude of 14  S to 35  S and longitude of 7  E to 35  E and has a regular horizontal grid of 1      1   and vertical grid of 25 km.

2.2.1. Inversion for 1-D path-average models

In comparison with the strongly nonlinear problem of inverting waveforms directly, these secondary observables have only a mild nonlinear dependence on the model parameters. As shown by Cara and L  v  que (1987) this technique minimizes the dependence on the starting model and reduces the number of iterations needed to find a 1-D depth-dependent model, which predicts waveforms compatible with the observed surface wave seismogram. Cara and L  v  que (1987) define a set of observed and synthetic secondary observables derived by cross-correlation with single mode synthetics. The single mode synthetics are computed assuming a reference 1-D crustal and upper mantle model for the fundamental mode and each higher mode up to the fourth order. A set of secondary observables are derived from cross-correlation of the recorded seismogram with pure-mode synthetic seismograms calculated for each of the fundamental and four higher modes. This cross-correlation with pure-mode synthetics reinforces the part of the actual signal that is close to the pure-mode synthetics. In parallel a set of synthetic secondary observables is derived by cross-correlation of a full synthetic seismogram with the same pure-mode reference synthetic seismograms. The resulting observed and synthetic cross-correlograms are filtered at different frequencies using Gaussian filters, and then their envelope functions are calculated. Then, one value at each significant local maximum and two values at the flanks of the observed envelopes are selected as a set of actual secondary observables. Once the envelope-based actual secondary observables are fitted with the synthetic envelopes, the inversion proceeds by fitting the phase of the cross-correlogram at the maximum of the envelope function. At each iteration, the full synthetic seismogram is updated to match the actual waveform. The waveform fitting procedure is automatic for each seismogram. The period range used here is 50–160 s for the fundamental and the first four higher Rayleigh wave modes, depending on their signal-to-noise ratio (SNR). At each period, the SNR is deemed adequate if the ratio between the maximum amplitudes of the envelopes of the signal and noise is larger than 3. The signal is evaluated in five frequency bands centered at periods of 50, 70, 90, 120, and 160 s, respectively. After the waveform inversion, we check that convergence is achieved and that the final synthetics fits the actual waveform (i.e. the residual between the actual

and synthetic waveform is small). Noisy waveforms are systematically rejected, as they cannot be explained with reasonable path-average 1D SV models (Debayle, 1999).

The inversion is performed for the upper mantle structure with the crustal structure being fixed. The 1-D model used for the calculations of the reference pure-mode synthetics and for the complete synthetic seismogram at the first iteration includes path average crustal models from the 3-D global 3SMAC crustal model (Nataf and Ricard, 1996) and the modified PREM model (Dziewonski and Anderson, 1981) for the mantle, where the original 220 km discontinuity is smoothed out (see Priestley et al., 2006). The stress displacement functions necessary to build the initial synthetic seismograms are computed using the reflectivity code from Takeuchi and Saito (1972). Source parameters are taken from the Global CMT catalog (Dziewonski et al., 1981; Ekstr  m et al., 2012). The inversion is considered successful if the final synthetic waveform matches well the observed seismogram and if the inversion converges towards a unique and stable velocity model. The automated selection of secondary observables along with the waveform inversion schemes are discussed in detail by Debayle (1999). The output of this inversion scheme is a 1-D path average shear wave velocity model for the upper mantle along each great circle path. In this study, we obtained 7248 1-D path averaged models, corresponding to 1234 earthquakes. Fig. S4 shows the distribution of path lengths. For this data set we achieved >200 paths crossing each 5      5   for the entire study area (Fig. S3).

Fig. S5 shows some examples of waveform inversion. We follow closely the criteria for acceptance of waveform fit given in Debayle (1999). We isolate multi-modes Rayleigh waves using a group velocity window between 7 and 3 km/s. We first check that the energy of the residual signal (i.e., difference between the actual and synthetic seismograms) is small. We compute the ratio between E_{resfin} , the energy of the residual seismogram at the last iteration, and E_{actual} , the energy of the actual seismogram. Energies are summed over the signals for group velocities between 6 and 3.5 km/s and the ratio $E_{res/act} = E_{resfin}/E_{actual}$ is computed, which is considered as the misfit. We also compute an energy reduction parameter $R_e = (E_{resinit} - E_{resfin})/E_{resinit}$, where $E_{resinit}$ is the energy of the initial synthetic seismogram, also summed for group velocities between 3.5 and 6 km/s. The waveform fitting is accepted if $E_{res/act}$ is smaller than 0.3 or if R_e is larger than 90%. Our experience shows that these conditions ensure a very good fit of the actual waveforms. Full details of the waveform inversion approach and of the criteria used at each step of the process are given in Debayle (1999).

This inversion approach requires two basic assumptions. The first assumption is that the observed seismograms can be fully represented in terms of multi-mode surface waves that propagate independently, and the second assumption is that they do so along great circle paths. The necessary condition for the validity of the first assumption is that the medium should be varying smoothly (Woodhouse, 1974) and for the second that the lateral velocity variations should not be too large. Kennett (1995) established that the validity of the path-averaged approximation holds good at periods >30 s and for surface waves propagating beneath continents at regional distances (with typical path lengths of less than ~4000 km). For the case where surface waves cross major structural boundaries (continent–ocean transition), which is the case in our present study, Kennett (1995) suggests to use the path-average approximation only at periods >50 s. Previous studies (Tramert and Woodhouse, 1995; Yoshizawa and Kennett, 2002) also confirm that for the fundamental and first higher modes, off-great circle propagation can be neglected at periods >40 s and at epicenter-station distances smaller than 10,000 km. In this study, we choose a period range between 50 and 160 s in order to work safely with the great circle approximation and expect that the propagation or site effects are neglected. However, our study area is far from any seismically active zone, only nearly half of the paths of the 1-D path averaged model (3411) are shorter than 10,000 km (Fig. S4). In order to increase the path coverage, we introduced paths longer than 10,000 km (>90  

acknowledging the fact that longer paths involving a larger Fresnel zone are more susceptible to off-great-circle deviations and multi-pathing. To counterbalance this effect, we have compared our final velocity model with the model considering only shorter paths ($<90^\circ$, Fig. S6) and show that the longer paths do not significantly bias our tomographic model (see Section 4 Results). Therefore, for the final interpretation, we have used the model with longer paths, as we believe that the benefit of the additional ray coverage is larger than the bias due to off-great-circle propagation.

2.2.2. Inversion for 3-D velocity model

The tomographic inversion scheme we used is developed by Debayle and Sambridge (2004) to be applied to large datasets such as ours. This scheme is an extension of the continuous regionalization algorithm of Montagner (1986) using the inversion approach of Tarantola and Valette (1982). This algorithm produces both the isotropic component of 3-D SV-wave velocity heterogeneity and the azimuthal anisotropy. We have undertaken inversions with or without azimuthal anisotropy. We found that the isotropic part is very robust and not significantly altered by the extraction of azimuthal anisotropy. However, our azimuthal coverage does not allow us to properly resolve the azimuthal anisotropy everywhere in the study area. For this reason, we only discuss the isotropic part of the model. Involving anisotropy in the inversion may help us to avoid potential biases by neglecting azimuthal anisotropy.

A laterally smooth model is obtained by imposing correlation between neighboring points with the use of a Gaussian covariance function, defined by a scale length (L_{corr}) and a standard deviation (σ). L_{corr} defines the distance to which adjacent points of the model are correlated and acts as a spatial filter which controls the horizontal smoothness. Whereas σ controls the amplitude of the model perturbation in the Earth structure that is permitted in the inversion. The value chosen for L_{corr} is 200 km, which means that our model is smoothed over wavelengths of ~ 600 km (Debayle et al., 2016) corresponding to the distance between two points on each side of the maximum, with amplitude of approximately one third of the maximum. The width of L_{corr} centered around each ray path ensures a full coverage of the study area with a large redundancy. In addition, this value allows us to exploit the information contained in our shortest period (50 s) Rayleigh waves (Sieminski et al., 2004), while keeping a relatively smooth model. The a priori standard deviation is set to $\sigma = 0.05$ km/s according to the expected values at regional (Debayle and L  v  que, 1997; Debayle et al., 2001; Nishimura and Forsyth, 1989) and global scales (Debayle and Ricard, 2012). In Fig. S7 we visualize the effects of choosing different regularization parameters (L_{corr} and σ) on velocity models at 100 km depth and show that our final choice of parameters is appropriate.

3. Resolution and error assessments

3.1. Effects of crustal models

The method mentioned in the previous section, the tomographic inversion consists of two stages and there are diverse factors which can influence the reliability of the model. In the first stage of calculating the 1-D path-average model, artefacts can arise from errors in the assumed crustal model and source parameters.

In the waveform modeling, the global 3-D 3SMAC crustal model (Nataf and Ricard, 1996) is used as an a priori model. Because of the imposition of the a priori crustal model, anomalies above the Moho are not constrained by the surface wave data but simply reflect the a priori model. Anomalies immediately below the Moho are in principle resolved, but may suffer from artefacts if there are discrepancies between the actual and assumed crustal structures. This trade-off has been evaluated by Debayle and Kennett (2000) and is likely to be small at depths significantly larger than the Moho depth. Previous studies (Debayle and Kennett, 2000; Pilidou et al., 2004; Priestley et al., 2008) suggest that the effects of crustal corrections with different crustal

models, e.g., the 3SMAC or the CRUST2.0, are indistinguishable at depths larger than 100 km and are minor at shallower depths. This is well explainable, as the crustal thickness is averaged over the entire path through the model and therefore the difference is small for different standard Earth's models. Debayle and Kennett (2000) showed that the effect produced by a 10 km variation in crustal thickness on a SV velocity model with a path-average crustal thickness of 32 km is within the a posteriori error bars (<0.05 km/s) in the uppermost 100 km and is negligible below 100 km depth. In most of our study area, the Moho depth is between 30 and 40 km with small variations (e.g., Heit et al., 2015; Fromm et al., 2017; Boyce et al., 2021) beneath the African continents, while it is located at shallower depths beneath the thinner oceanic crust. Also, the method we adopted for the waveform inversion is based on Cara and L  v  que (1987), which holds a weak dependency on the starting model. Therefore, we can safely start the inversion from a unique upper mantle model (i.e., a smooth version of the PREM model) with a crustal part from the 3SMAC model adapted to each path. Regarding the influence of errors in the source parameters, we believe that a good (redundant) path density and azimuthal coverage, like the ones we obtained for this work, are a basic requirement for reducing those errors.

3.2. A Posteriori error assessment

For the second stage, where realization of a 3-D model is achieved from combining all 1-D path-averaged models using a continuous regionalization method, we computed the a posteriori error on the tomographic model based on Tarantola and Valette (1982) and adapted it to the continuous regionalization by Montagner (1986). The maps and cross sections in the 3-D distribution of the a posteriori error are shown in Fig. 3. It is well known that the a posteriori error, which is obtained from the square root of the diagonal terms of the a posteriori covariance matrix, is a useful tool to quantify model uncertainties. The a posteriori covariance matrix C_m is related to the a priori covariance matrix C_{mo} by $C_m = (I - R)C_{mo}$, where I is the identity matrix and R is the resolution matrix. Therefore, the error estimates depend on the a priori covariance matrix. We note that it is also true for R , and any resolution test will depend on the choices made for the correlation length and a priori standard deviation. Provided that our a priori choices are reasonable, the a posteriori errors are a very useful tool to guide the interpretation of the seismic model. Regions where the a posteriori error is close to the a priori error (i.e. $\sigma = 0.05$ km/s) can be regarded as regions of poor resolution (R close to 0) as shown in dark gray regions at the border of the map. As a result of the dense path coverage and rich higher-mode content of the data set, the a posteriori error is nearly ± 0.04 km/s for most of the region to a depth of 300 km. This means that S-wave perturbations larger than ± 0.04 km/s ($\sim 0.9\%$ considering an average velocity of 4.4 km/s) are significant at the 68% confidence level and that perturbations greater than ± 0.08 km/s ($\sim 1.8\%$) are significant at the 95% confidence level. We indicate the 0.9% and 1.8% contour levels on the inverted velocity perturbations as visual guide to areas of significant heterogeneity at 68% and 95% confidence levels, respectively.

3.3. Checkerboard tests

The image accuracy is not quite easy to quantify, however, traditional checkerboard tests are capable of providing at least a qualitative measurement of the ability of a particular data set in resolving the associated existing features. In usual practice of checkerboard tests, both high and low velocity anomalies are placed next to each other. To further test the smearing effects, both in the horizontal and vertical directions, we placed the high and low velocity anomalies separately in a neutral background zone and observe if the anomalies are recovered after the inversion. We conducted a number of checkerboard tests to examine the ability of the selected dataset to recover velocity anomalies of different sizes. Fig. 4 shows the test with seismic anomalies extending

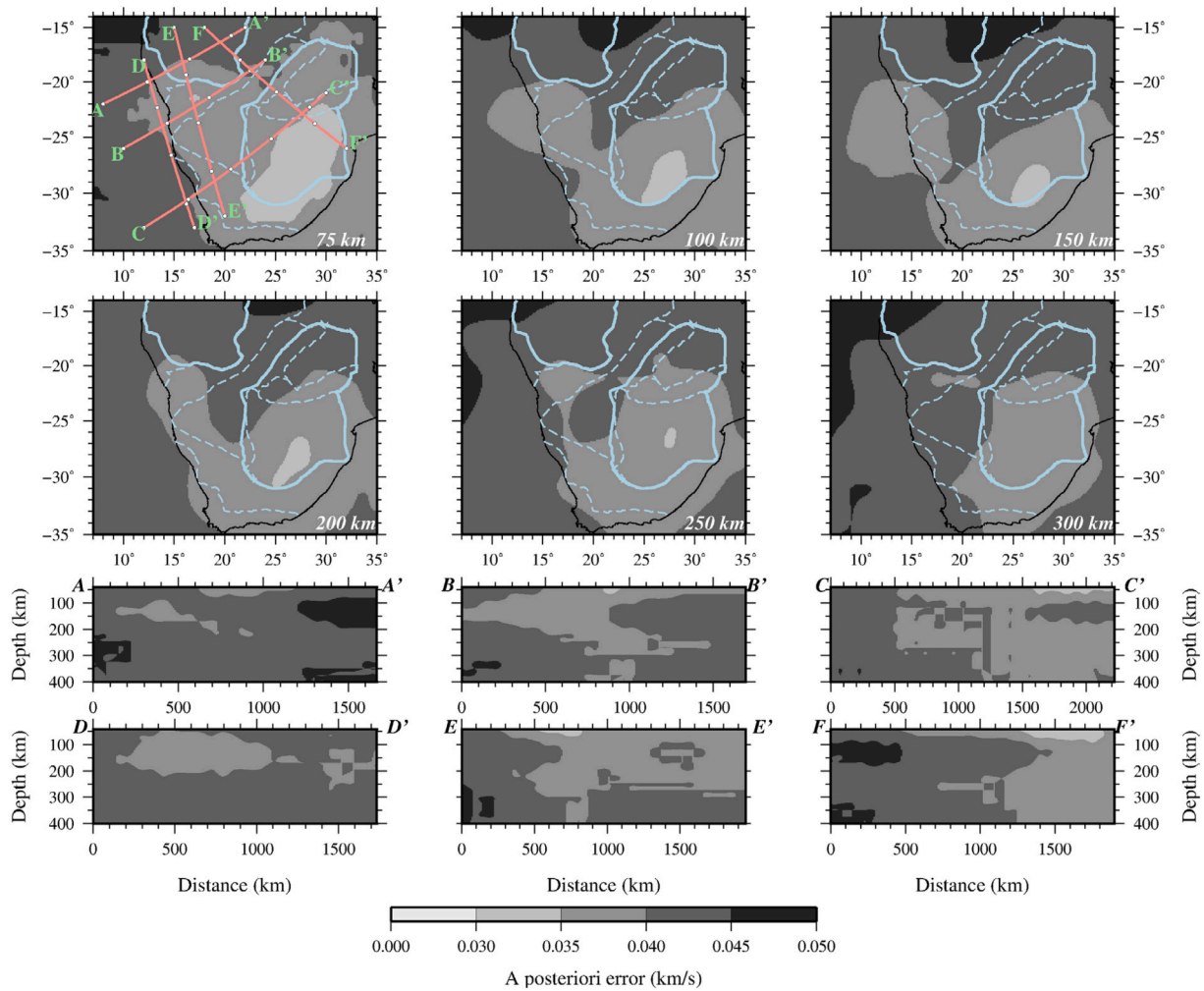


Fig. 3. A posteriori errors at depths of 75, 100, 150, 200, 250 and 300 km (upper 6 panels) and along 6 cross sections (lower 6 panels). Pink lines at the map of 75 km depth denote the locations of the cross sections. The locations of the cross sections are the same as shown in Fig. 7 for the 3D velocity model. The correlation length (L_{corr}) is 200 km, the same as used for the inversion. In most of the regions the errors are within the range of 0.035–0.040 km/s or less. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

over 500×500 km horizontally in the middle of the map and 100 km vertically. Alternating high and low velocity anomalies with magnitudes of $\pm 6\%$ are spread over the entire volume and are separated by ~ 500 km wide 0 % anomalies. We calculated synthetic Rayleigh wave seismograms for the same ray paths, source parameters, and frequency contents as in the observed data and carried out the same inversion procedure. At shallow depth (< 200 km) the input model can be almost completely recovered. At depths of 200–400 km nearly half of the magnitude of the anomalies can be recovered. The synthetic test shown in Fig. 4 gives an intuitive representation of the ability to recover a particular model from our ray coverage and a priori choices at the scale length of the individual blocks. However, as shown by L  v  que et al. (1993) such a test does not demonstrate that other synthetic models with larger anomalies will necessarily, or even equally, be well retrieved. Keeping this in mind, we performed further tests with grid size of 500×500 km horizontally and 50 km vertically (Fig. S8) and similarly with 250×250 km horizontally and 100 km (Fig. 5) or 50 km (Fig. S9) vertically. We conclude that in the $500 \times 500 \times 100$ km and $500 \times 500 \times 50$ km tests, all the anomalies can be well recovered. The $250 \times 250 \times 100$ km test can recover most of the features, but the $250 \times 250 \times 50$ km test failed to resolve the depths > 200 km.

3.4. Anomaly recovery tests

We performed anomaly recovery tests (Figs. S10, S11 and 6) to further evaluate the reliability of the anomalies, observed and interpreted in this study, and to assess the horizontal and vertical smearing. We designed input models with simplified anomalies approximately representing the main observations (see next sections) and simulated the inversions, similar to the checkerboard tests, based on the actual distribution of the ray paths from the real data. In the first test, we use one single high-velocity anomaly beneath the coastal area at depths between 250 and 350 km (Fig. S10) to test the detectability and the lateral smearing of the anomaly. In the second test, we add a low-velocity anomaly above the high-velocity anomaly at depths between 100 and 200 km to observe the interaction of the two anomalies (Fig. S11). In the third test, we added two more high-velocity anomalies at depths of 70–200 km (Fig. 6). The outcome of all these tests shows that all these anomalies are recoverable by our data with some extent of lateral smearing, e.g., between the two anomalies representing the Congo and Kaapvaal Cratons. The tests are a useful guide for the interpretation.

4. Results

The final isotropic 3-D SV velocity perturbation model is visualized by six horizontal sections (Fig. 7) and six vertical cross sections (Fig. 8).

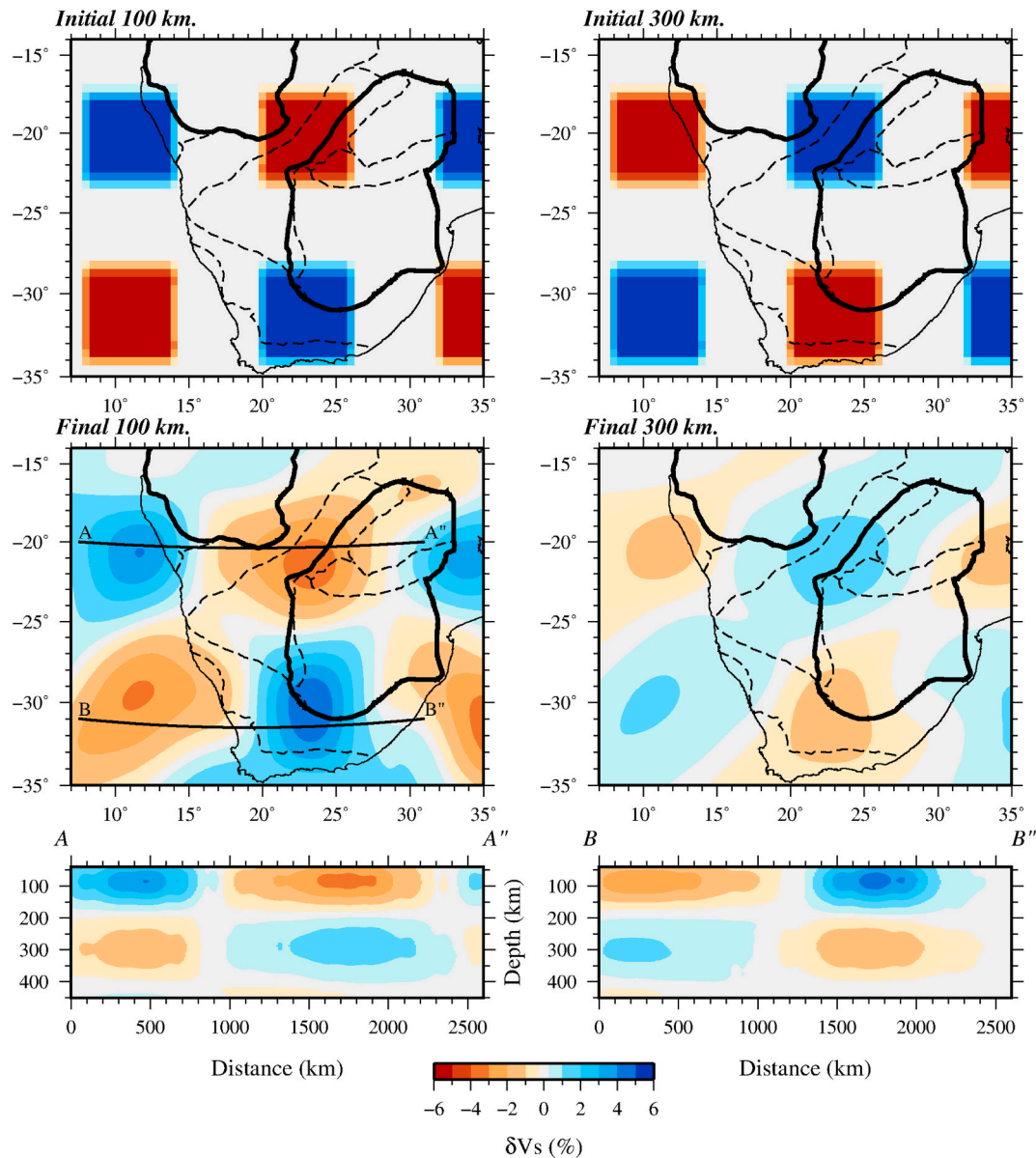


Fig. 4. Horizontal and vertical slices of the checkerboard resolution test. Alternating high and low velocity perturbations with a size of 500×500 km in the horizontal and 100 km in the vertical direction and a magnitude of $\pm 6\%$ are separated by zero-anomaly background in the input model.

The corresponding absolute velocity model is shown in Figs. 9 and 10. As discussed in Section 2.2.2, the current path coverage and distribution are not sufficient to reliably resolve the azimuthal anisotropy, although it was taken into consideration for constructing the 3-D model. Therefore, only the isotropic SV wave velocity is interpreted in this paper. We note that anisotropic tomography for southern Africa was conducted by Sebai et al. (2006) and Adam and Lebedev (2012) with different approaches. We indicate the confidence levels of SV perturbations of 0.9% and 1.8% on the cross-sections in Fig. 8 as visual guide to areas of significant heterogeneity at 68% and 95% confidence levels, respectively.

4.1. Horizontal sections

The depth slice at 75 km exhibits velocity anomalies in the shallow lithosphere, where the influence of the crustal model is negligible. Overall, the study area is of high velocity, including the offshore area. Anomaly H3 represents the high-velocity upper lithosphere of the Kaapvaal Craton. Along the western African coast where the Walvis Ridge southwards meets the continental margin, the lithosphere is

characterized by high velocity (H1) (Figs. 7a and 9a). Another high-velocity body can be seen in the south between 27 and 29°S. The slices at 100 and 150 km depth are dominated by high velocity anomalies of the Congo (H2) and Kalahari (H3) Cratons. With respect to the boundary of the Congo Craton, the high velocity anomaly extends farther to the west beneath the Etendeka CFB province. The high velocity anomalies beneath the cratons become weaker at ~ 200 km depth but they are still visible and concentrate in the region beneath the Limpopo Belt. This depth marks the lower boundary of the high velocity anomaly (H3). A low-velocity anomaly at depths of 100–200 km (L1) and a high-velocity anomaly at depths of 250–350 km (H4) can be seen offshore Namibia beneath the Walvis Ridge.

4.2. Vertical crosssections

Figs. 8 and 10 show six selected vertical cross sections spanning over major tectonic units. Three cross sections (AA', BB' and CC') are in the SW-NE direction, whereas the other three cross sections (DD', EE' and FF') are in the NW-SE direction. The exact locations of these cross

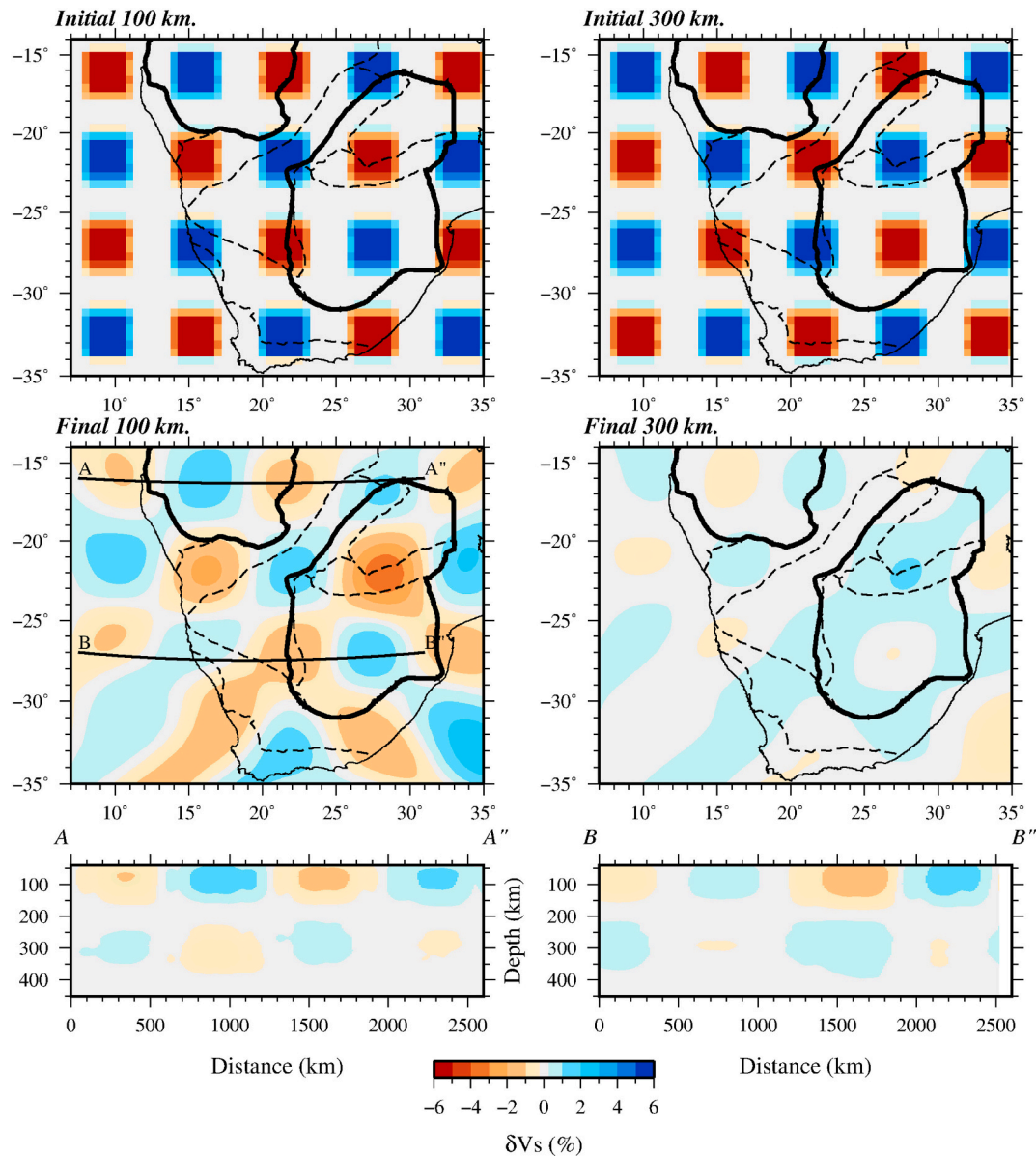


Fig. 5. Same as Fig. 4 but for anomalies of 250×250 km horizontal size.

sections are shown in the 75 km depth slice in Figs. 7a and 9a.

Section AA' extends from the Walvis Ridge in the ocean through the Etendeka CFB area on land to the Congo Craton. Beneath the Walvis Ridge, relatively high velocities (H1) are visible in the uppermost mantle to a depth of ~ 80 km. To the east beneath the continent, a pronounced high velocity anomaly (H2) appears. It becomes thicker and stronger and also more prominent down to a depth of < 200 km beneath the Congo Craton. At greater depths beneath the offshore area, the low-velocity anomaly L1 with a confidence level of 95% and the high velocity anomaly H4 with a confidence level of 68% are clearly visible.

Section BB' starts from the Cape Basin and extends eastward along the Damara Belt. Compared to cross section AA', the general features of cross section BB' are similar, but a sharp demarcation of the mantle anomalies between the ocean and continent can be clearly observed. Whereas the offshore high velocity anomaly H1 remains the same, anomaly H2 is clearly separated from anomaly H1 and increases its thickness sharply at distances of 700–800 km. The deep mantle anomaly, H4, extends slightly beneath the coastal area of the continent.

In the south, Section CC' starts from the southern part of the Cape Basin and enters the continental region at the Namaqua-Natal Belt and

crosses the Kalahari/Kaapvaal Craton. The high velocity feature beneath the cratonic regions is present to a depth of > 200 km. Unlike the previous two cross sections (AA' and BB'), the pronounced shallow offshore anomaly (H1) is missing and the deeper high velocity anomaly (H4) is replaced by a wide low velocity anomaly.

The distribution of anomalies H1, L1 and H4 along the coast can be well visualized by cross-section DD', which is nearly north-south oriented and is located offshore along the continental margin. This section starts from the Walvis Ridge passing through the Cape Basin. The shallow mantle high-velocity anomaly (H1) is thicker beneath the Walvis Ridge than south of it and is underlain by deeper anomalies (L1 and H4).

Section EE' covers the entire coastal area of the continent starting from the Congo Craton, crossing the Damara Belt and ending at the Namaqua-Natal Belt. The high velocity mantle anomaly (H2) is prominent beneath the Congo Craton and the Damara Belt. The deep mantle anomaly (H4) is much less visible.

Section FF' was defined to cross both the Congo and the Kalahari Cratons. The cross section is dominated by anomalies H2 and H3, reaching a maximum depth of ~ 200 km beneath the Kalahari Craton.

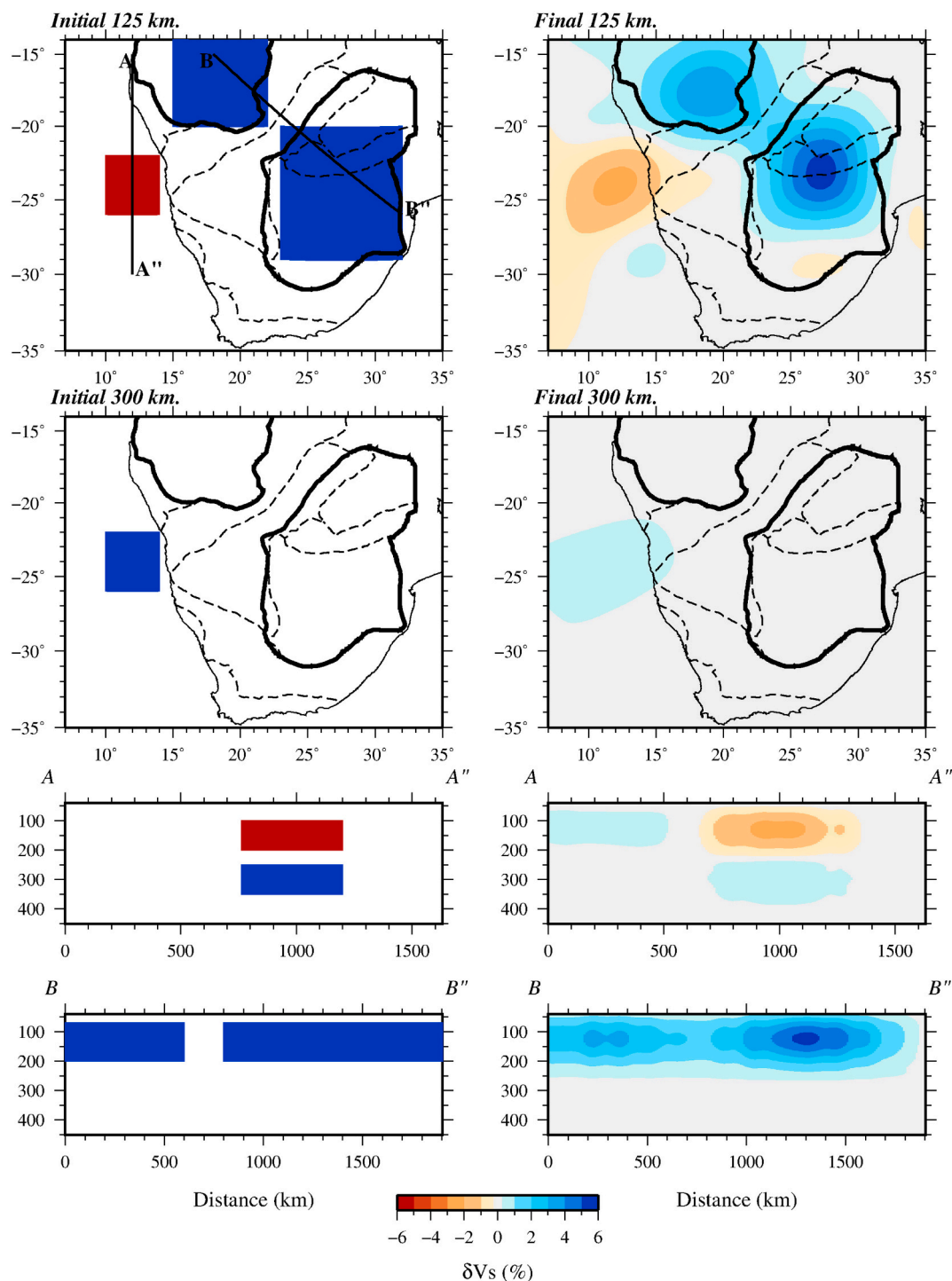


Fig. 6. Horizontal and vertical slices of an anomaly test. High and low velocity perturbations are designed to approximately simulate our observed mantle anomalies. A magnitude of $\pm 6\%$ is given to each anomaly.

5. Discussion

As shown above, we observed shear wave velocity anomalies at different depths in the mantle lithosphere (100–200 km depth) and sublithosphere (250–300 km depth). In Fig. 11 we summarize the observations and propose possible interpretations. These anomalies may have been induced by different tectonic and dynamic processes or represent an integrated result of a single process.

5.1. Plume-lithosphere interaction

The terms lithosphere and asthenosphere were originally defined with reference to rheology, with the lithosphere essentially behaving as elastic solid floating on and moving over the asthenosphere which is deforming as a viscous fluid (Barrell, 1914). Later on, additional terms like thermal, chemical or seismic lithosphere have been introduced (Anderson, 1995; Eaton et al., 2009), with the seismic lithosphere being defined as the high-velocity lid overlying a low-velocity asthenosphere as it is used in the present work. The base of the lithosphere has been

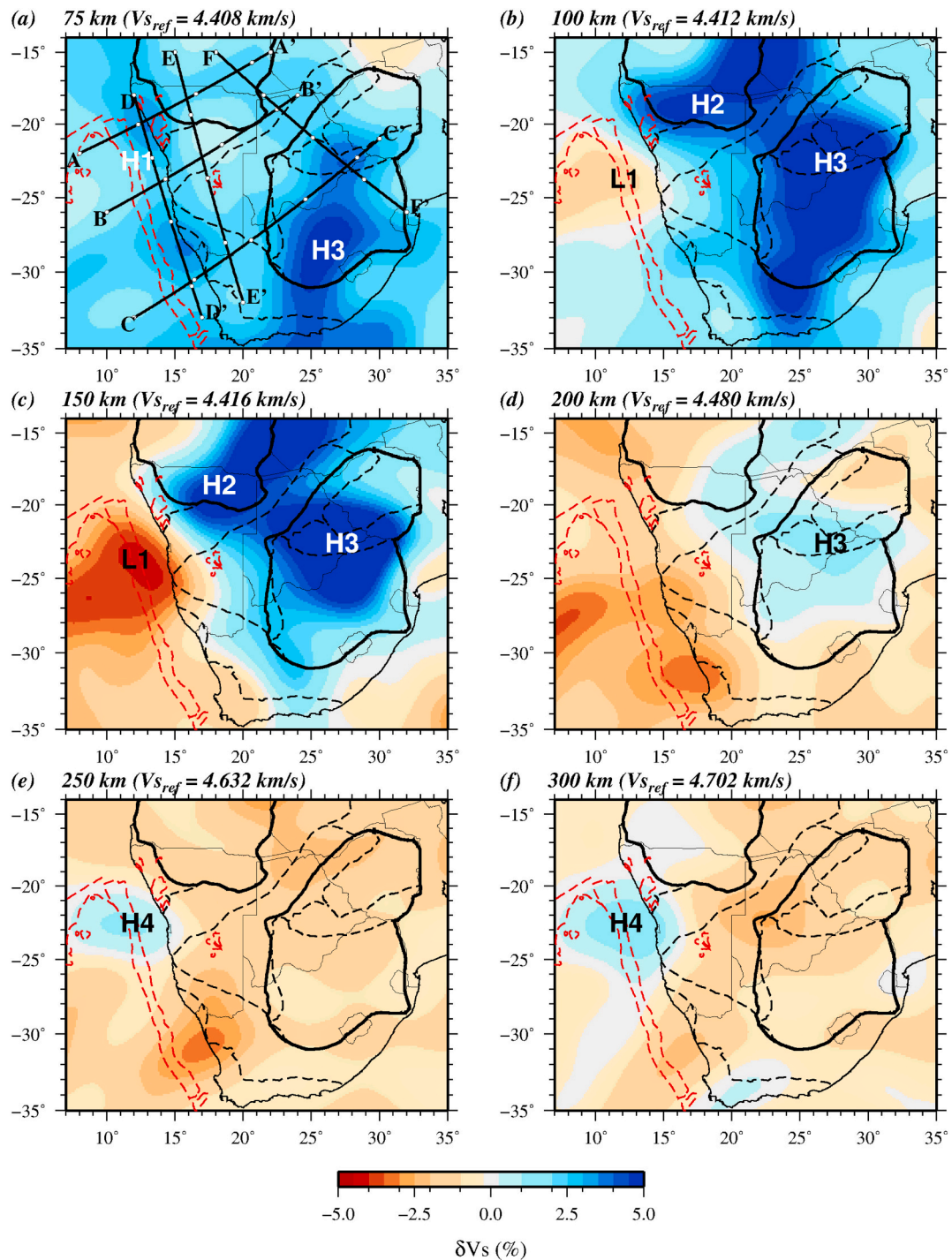


Fig. 7. Horizontal sections of SV velocity perturbations at depths of 75, 100, 150, 200, 250 and 300 km. The great circle lines drawn on the 75 km depth slice indicate the locations of the six cross sections shown in Figs. 8 and 10. The velocity perturbations are referenced to the smoothed PREM model (Dziewonski and Anderson, 1981). The black and red lines are the same as in Fig. 1. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

defined in different ways by modeling seismic velocity profiles from surface wave data (Eaton et al., 2009). In this study, we compare velocity profiles with the smoothed PREM model and identify the mantle lithosphere by high-velocity perturbations over the reference model. In most cases, the base of the high-velocity mantle lid coincides with the largest negative velocity gradient and is in general agreement with the base of the lithosphere defined by thermal gradient changes (e.g. Priestley et al., 2008). Our study area covers different cratonic blocks, mobile belts, oceanic domains, as well as a hotspot track along the

Walvis Ridge. The 3-D SV velocity model at depths of 100–200 km (Figs. 7b-d and 9b-d) exhibits variations in the lithospheric thickness beneath different tectonic regions and potential imprints of the Cretaceous mantle plume.

In our study area, the thickness of the lithosphere generally agrees with the global map of LITHO1.0 (Pasyanos et al., 2014), as well as with previous surface wave studies (Li and Burke, 2006; Priestley et al., 2008; Fishwick, 2010; Ravenna et al., 2018; Emry et al., 2019; Celli et al., 2020; White-Gaynor et al., 2021). In Figs. 7b-d and 8b-d, high-velocity

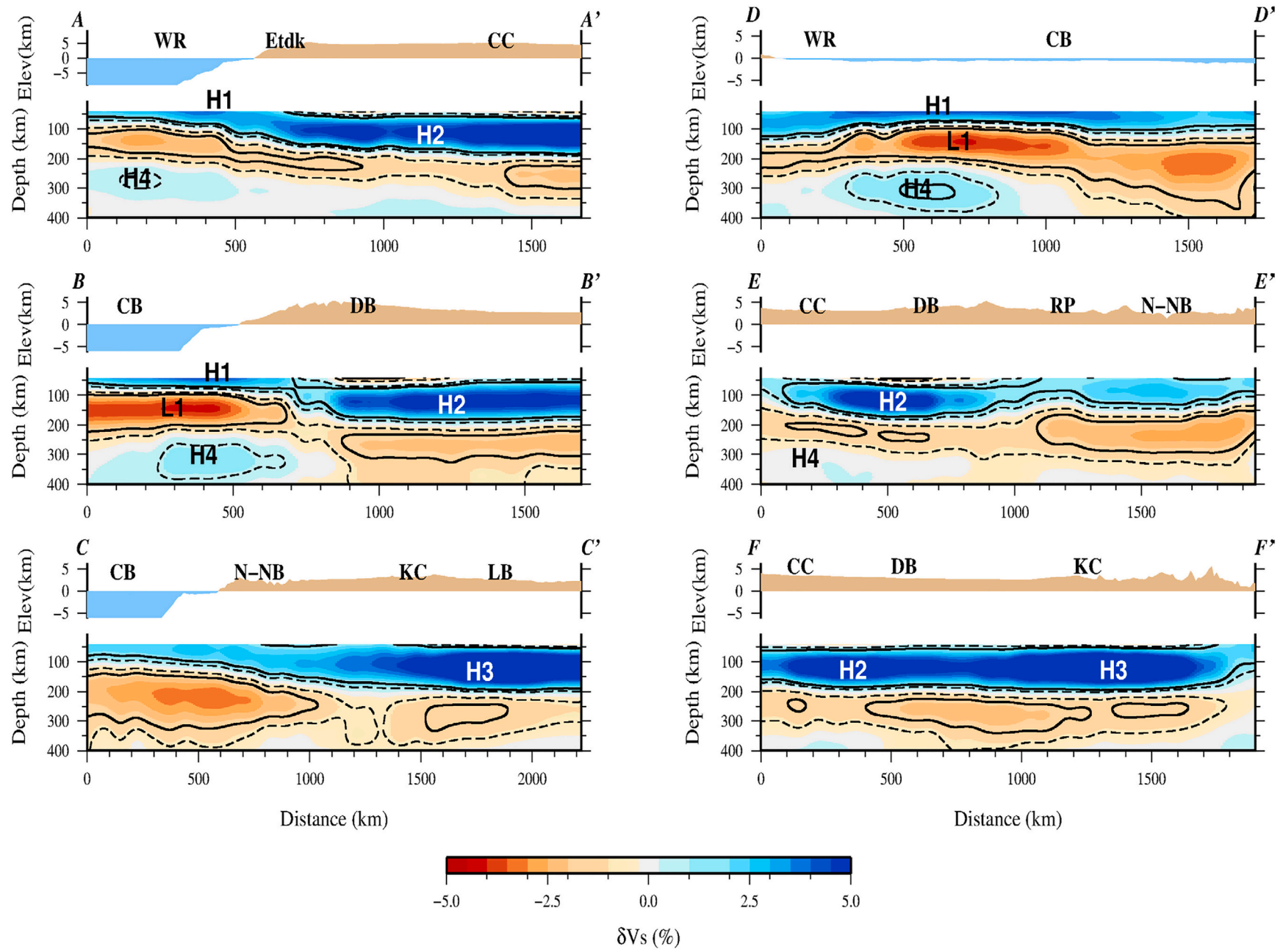


Fig. 8. Cross-sections of SV velocity perturbation along great circle lines AA', BB', CC', DD', EE' and FF'. Cross section locations are shown in Fig. 7. The velocity perturbations are referenced to the smoothed PREM model (Dziewonski and Anderson, 1981). Surface topography is plotted on top of each cross section with major tectonic units indicated. Contour lines surround the areas whose anomalies (i.e., deviations from the reference model) are significant at least at the 68% (dashed lines) and 95% (solid lines) confidence levels. Abbreviations: WR, Walvis Ridge; Etdk, Etendeka; CC, Congo Craton; CB, Cape Basin; DB, Damara Belt; RP, Rehoboth Province; N-NB, Namaqua-Natal Belt; KC, Kaapvaal Craton; LB, Limpopo Belt.

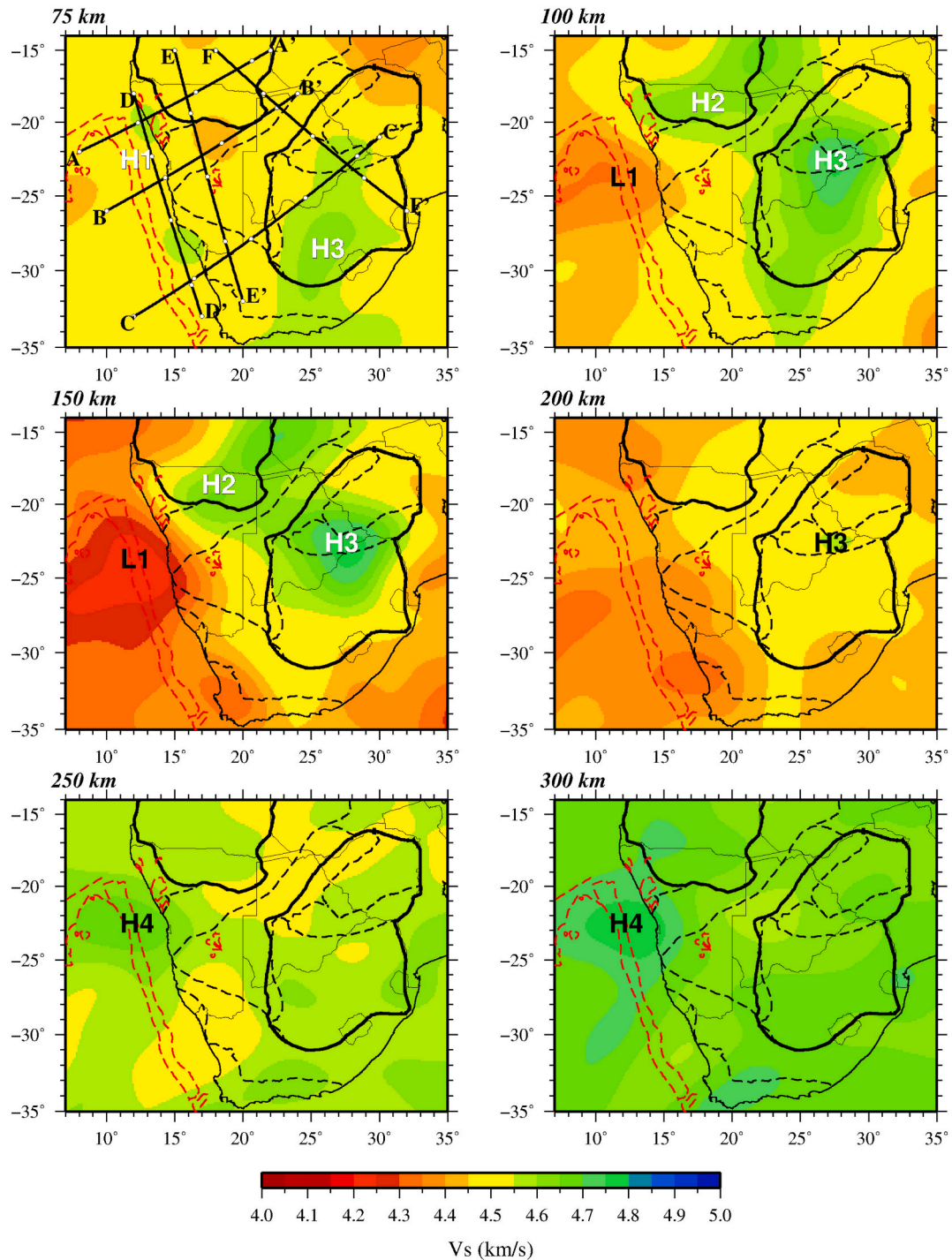


Fig. 9. Same as Fig. 7 but for absolute velocity.

anomalies H2 and H3 represent the thick cratonic mantle lithosphere beneath the Congo and Kalahari Cratons, respectively, extending to a depth of ~200 km. Compared with previous models (e.g., Priestley et al., 2008; see Fig. S13), the thick Congo Craton lithosphere, indicated by high-velocity anomaly H2, extends farther south beneath the Damara Belt and farther west towards the Walvis Ridge. As also observed by previous studies mentioned above, a thick lithosphere keel underlies the eastern part of the Kalahari Craton (anomaly H3), which covers the Kaapvaal and Zimbabwe Cratons and the Limpopo Belt in between. The thickest lithosphere is found beneath the Limpopo Belt with a thickness of ~200 km. Adam and Lebedev (2012) also obtained a thickest lithosphere beneath the Limpopo Belt using surface wave tomography, but

with larger thickness values, as interpreted by Ravenna et al. (2018). In their study, the lithospheric thickness increases from ~200 km in the central and south Kaapvaal Craton to ~300 km beneath the Limpopo Belt. The difference in the absolute thickness values may be due to different methods used to identify the base of the lithosphere from the velocity profiles. Our estimate is in good agreement with Li and Burke (2006), Priestley et al. (2008) and Fishwick (2010). Similar lithospheric configuration was also observed by other recent surface wave studies (Emry et al., 2019; Celli et al., 2020; White-Gaynor et al., 2021). Anomalies H2 and H3 are separated by the Damara Belt at 100 km depth and seem to merge at larger depths of ~200 km (Figs. 7–10), however this connection could be an effect of smearing as shown by the anomaly

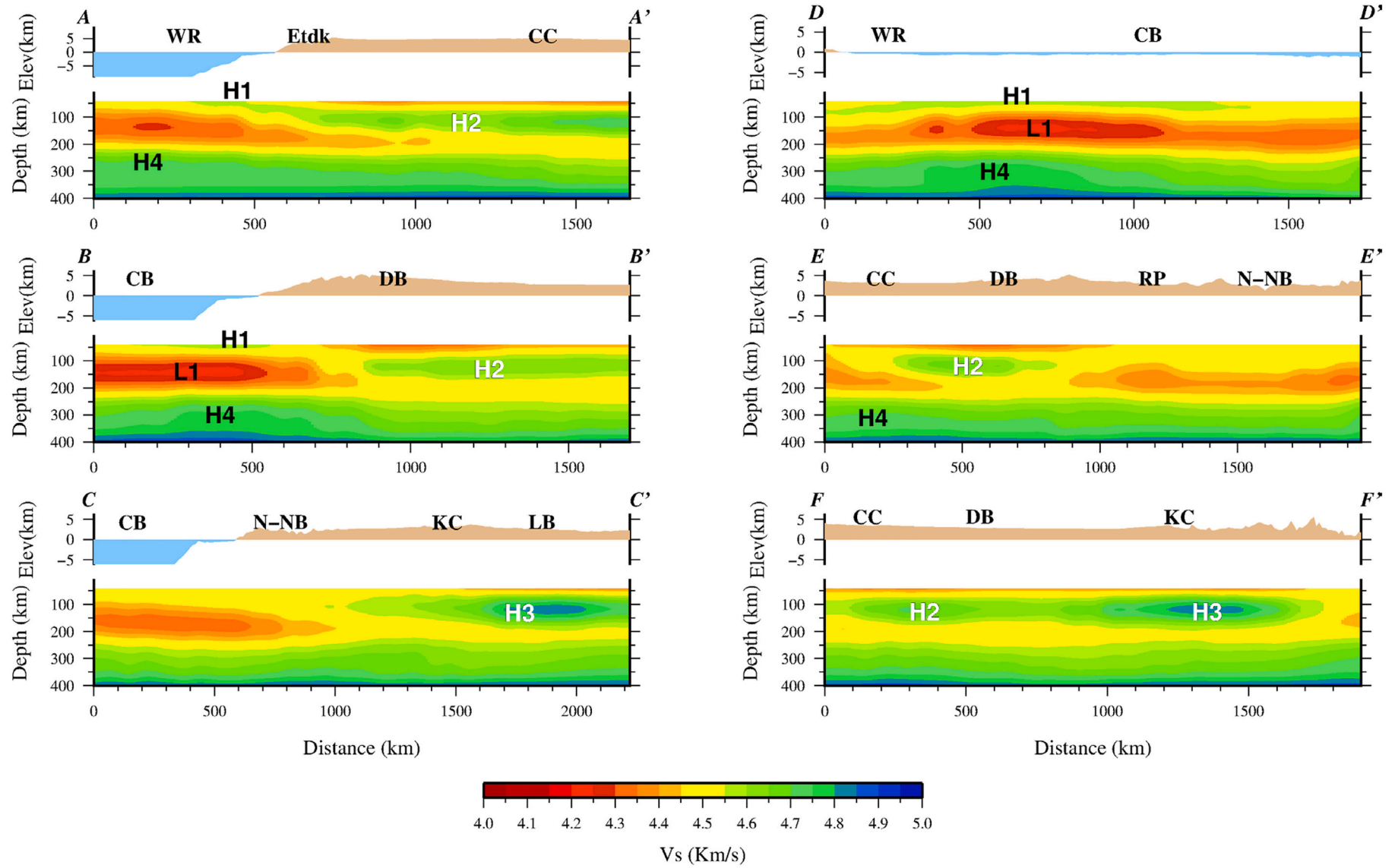


Fig. 10. Same as Fig. 8 but for absolute velocity.

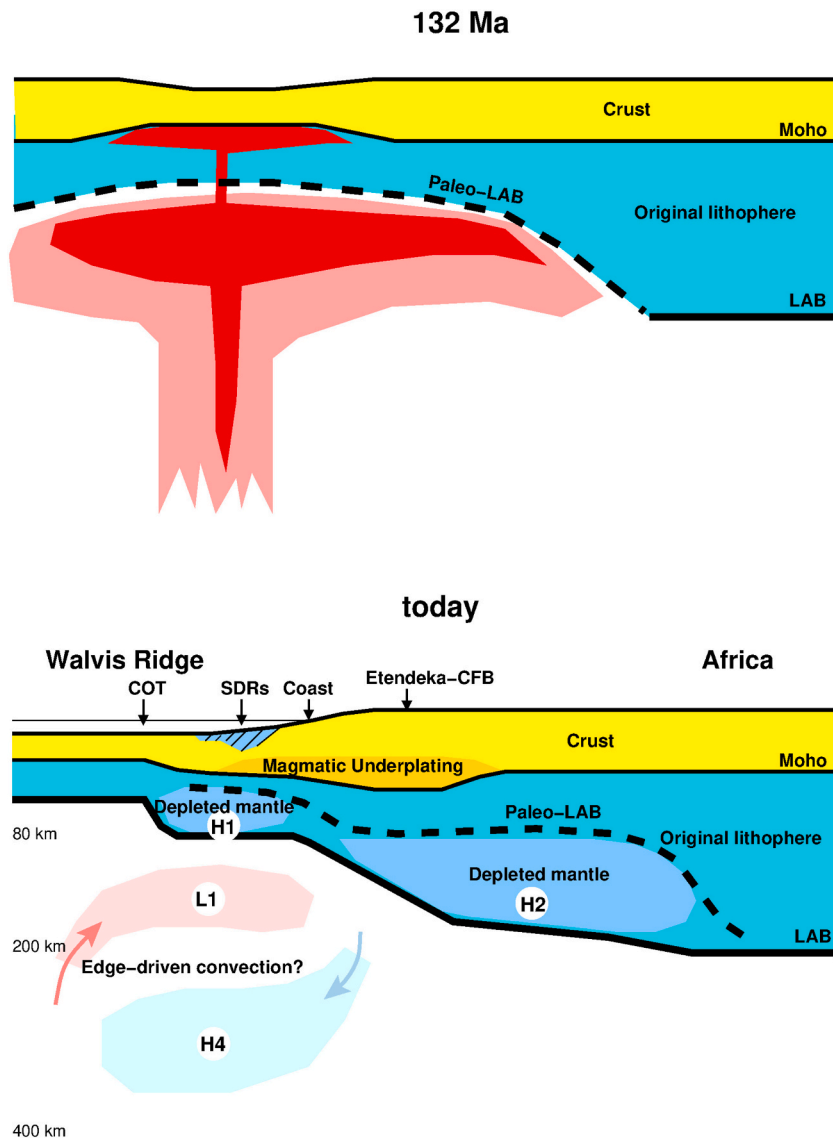


Fig. 11. Interpretation cartoon along the trajectory of the hotspot track from Walvis Ridge to Damara Belt. Upper panel shows the scenario at 132 Ma when the mantle plume arrived at the base of the lithosphere. The originally thick lithosphere was eroded and destroyed by the plume, to the depth depicted by the Paleo-LAB (Yuan et al., 2017). Melts produced by the thermal plume were accumulated at the base of the crust and mantle lithosphere. Lower panel shows the present-day conditions, where a thick lithosphere is reconstructed by melt residues (depleted mantle). Magmatic underplating at the base of the crust beneath the Etendeka CFB area is observed by Fromm et al. (2015, 2017) and Heit et al. (2015). Observed velocity anomalies (L1, H1, H2, H4) are indicated. COT: continent-ocean transition, SDRs: seaward dipping reflectors, CFB: continental flood basalt.

recovery test (Fig. 6).

Most of the previous surface wave models focus on the Kaapvaal Craton, where dense seismic station networks were available, providing a good resolution in the mantle lithosphere (e.g., Li and Burke, 2006; Priestley et al., 2008; Emry et al., 2019). The thick lithospheric keel of the Congo Craton is limited farther north to the craton centre, which is well separated from the Kalahari Craton by the Damara Belt (e.g., Priestley et al., 2008; Fishwick, 2010). The recent surface wave models of Emry et al. (2019) and Celli et al. (2020) appear to show that the thick lithosphere of the Congo Craton starts already from the landward extension of the Walvis Ridge and partly meets the Kalahari Cratons at depth (Figs. S1 and S2). In the present work, the inclusion of the WALPASS network added more surface wave paths and therefore increased the resolution in the region. We have tested the results we obtained by removing the stations from the WALPASS array (Fig. S12) and found that anomaly H2 becomes less pronounced in the study area. The Congo Craton is imaged by high-velocity anomalies limited to the areas farther north and east. In Fig. S13, we compare our observations with the thick lithospheric root seen by Priestley et al. (2008). We picked the contours of lithospheric thickness of 150 km from Priestley et al. (2008) and superpose them on the horizontal sections at depths of 100 and 150 km from the inversion including and excluding the WALPASS data, respectively. The distribution of the thick lithosphere agrees

reasonably well when the WALPASS data are not included (Figs. S13c, d). However, by including the WALPASS data, the high velocity anomaly H2 stands up markedly beneath the southern edge of the Congo Craton and the Damara Belt, close to the landfall of the Walvis Ridge. The recent surface wave models of Emry et al. (2019) and Celli et al. (2020) also seem to show that the deep lithosphere of the Congo and Kalahari Cratons partly meet at depth.

Evidence for the lithospheric layering and implication for a thick mantle lithosphere beneath the landfall of the Walvis Ridge is provided by receiver functions. Yuan et al. (2017) observed early arrivals of P-to-S converted phases of both the two main mantle transition discontinuities (the 410 and 660) by ~ 2 s beneath the Damara Belt and the southern Congo Craton in northwest Namibia and inferred a higher velocity upper mantle and hence a thicker mantle lithosphere there, which is consistent with the thick lithosphere (anomaly H2) observed in this study. The anomaly of the 410 and 660 arrivals disappears to the east beneath Tsumeb, suggesting that the implied thick lithosphere is a regional feature and is separated from the Congo Craton. This gap in the thick lithosphere of H2 can not be resolved by all existed surface wave tomographic models, which also implies that the gap may be small. Yuan et al. (2017) also observed a negative velocity gradient within the lithosphere at a depth of 80 km and interpreted it as a thin layer of crystallized melts at the paleo lithosphere-asthenosphere boundary (paleo-

LAB) formed by the interaction of the Cretaceous mantle plume with the formerly thick lithosphere. They assumed that the lithosphere was thinned to ~80 km at time of continental break-up when the plume arrived and has been gradually re-thickened, mainly by incorporation of melt residues. The paleo-LAB (Yuan et al., 2017), indicated in Fig. 11, represents the base of the thinned lithosphere at the time of continental break-up at 132 Ma (Renne et al., 1996). The present-day lithosphere was reconstructed by the remnant plume melts and eroded mantle lithospheric materials, which was chemically differentiated and re-attached to the base of the previously thinned lithosphere. These materials have a higher seismic velocity than the ambient mantle due to melt depletion and dehydration (Hall and Kincaid, 2003). The implication was, however, based on indirect observation of the mantle transition zone variation. Here we propose that the thick lithosphere beneath the Etendeka flood basalt areas and the landfall of the Walvis Ridge (anomaly H2) provides a direct evidence for the re-thickening of the lithosphere.

The plume-lithosphere interaction is two-fold. The pre-existing lithosphere was thinned by erosion by the mantle plume (Sobolev et al., 2011) to the depth of the paleo-LAB, but may subsequently be re-thickened by accretion of depleted peridotitic materials after extraction of melt. Melts from large thermal plumes have contributed to the formation of original cratonic lithosphere in the Precambrian (e.g., Arndt et al., 2009). This process can be active throughout geological time that also helps to recratonize the previously destroyed continental mantle lithosphere (Liu et al., 2021). Isotope studies of mantle xenoliths and magma from the Damara Belt show distinctive chemistry that can be attributed to deep decompression melting of pyroxenite-bearing materials in the Tristan da Cunha plume head (Owen-Smith et al., 2017) and the source of metasomatic enrichment of the previously depleted mantle lithosphere is related to the plume head at the time of continental breakup (le Roex and Class, 2014). Melt-induced dehydration increases both the viscosity and the seismic velocity of the residual mantle. Depleted residual mantle, poor in Fe and rich in Mg, is less dense and significantly more viscous than the original mantle, therefore, remains stable when attached to the base of the lithosphere and becomes part of the overlying rheological lithosphere (Hall and Kincaid, 2003). In this way, a viscous boundary that is effectively equivalent to a thickened lithosphere is created (Hall and Kincaid, 2003). The numerical modeling of Liu et al. (2021) provides a scenario of recovery of a lithospheric root destructed by mantle plume. They show that a 200 km thick lithosphere can be completely reconstructed in 400 M.y., whereas in 163 M.y. much of the root is already recovered. The process may also apply to our study area although the tectonic situation is different from the model where a lithospheric hole was modeled. The speed of the lithosphere growth may vary depending on tectonic settings and model parameters.

We therefore interpret the high-velocity thick lithospheric keel beneath the southern Congo Craton and the Damara Belt as the result of the plume-lithosphere interaction in the Cretaceous. Because mantle shear wave velocity is more dependent on temperature than on chemical composition (Schutt and Leshar, 2006), melt depletion alone is not sufficient to significantly alter the mantle velocity. Karato and Jung (1998) show that water significantly reduces seismic wave velocities through anelastic relaxation and thus provide a connection between low water content and high seismic velocity, where loss of volatiles in the depleted mantle occurs during plume melting. Furthermore, the attachment of the buoyant, depleted mantle builds a thick thermal boundary layer with the overlying original lithosphere and hinders convective heat transfer from the asthenosphere below and thus speeds up thermal cooling of the lithosphere.

The plume-lithosphere interaction has also left imprints in the crust and mantle lithosphere. High-velocity anomalies have been observed in the lower crust offshore along the coast by previous studies and were interpreted as magmatic underplating induced by the plume (e.g., Fromm et al., 2015, 2017). Strong variation in the crustal thickness is observed along the landfall of the Walvis Ridge, where the crust

significantly thickens landwards from ~30 km at the coast to ~45 km beneath the Etendeka continental flood basalt area where the Vp/Vs ratio reaches a high value of ~1.89 (Heit et al., 2015). These anomalies were interpreted as magmatic underplating induced by the mantle plume melts at the base of the crust during the break-up of Gondwana. The crust is thinner towards the offshore area where the Moho depths are mostly <30 km (Fromm et al., 2015).

The high-velocity anomaly along the coast (H1) coincides spatially with the large igneous province (LIP; Coffin et al., 2006) and the seaward dipping reflectors (SDR; Koopmann et al., 2014). The SDR is a ubiquitous feature of volcanic passive margins and is commonly interpreted as magmatism associated with continental rifting and break-up. Fromm et al. (2017) analyzed a number of wide-angle reflection profiles from the Walvis Ridge to the Cape Basin and found a thick oceanic crust with 22–28 km thickness beneath the ridge and high velocity lower crust beneath a wide area of the continent-ocean transition zone. The high velocities are interpreted as representing magma intrusions or underplating at the onset of rifting triggered by the plume head. The H1 high-velocity anomaly (Figs. 7a and 9a) is located in the same area, where the high velocity lower crust was observed by Fromm et al. (2017). This evidence could indicate a common origin, however, in our images, anomaly H1 is observed at greater depths in the mantle, at 50 km or deeper (Figs. 7a and 9a) and disappears at 100 km depth (Figs. 7b and 9b). We interpret anomaly H1 to represent melt residue associated with the SDRs and lower crustal magmatic underplating.

The plume-lithosphere interaction is evidenced in the vicinity of the Walvis Ridge by the magmatic underplating at the base of the crust, the SDRs and associated high-velocity anomaly in the uppermost mantle, and the thick mantle lithosphere there. Although it is not clear how far the plume-lithosphere interaction may extend and where the boundary between the plume modified lithosphere and the original lithosphere is located, observed anomalies are rather localized and suggest that the modified mantle lithosphere is likely restricted in the NW Namibia coastal areas around the Etendeka CFB areas along the inland projection of the Walvis Ridge. The limited lateral extent may imply that the plume was not the major trigger for the continental break-up.

5.2. Sublithospheric anomalies

The low-velocity anomaly L1 at depths of 100–200 km and the high-velocity anomaly H4 at depths of 250–350 km, below the mantle lithosphere, are enigmatic. The spatial distribution of these anomalies is restricted to a small area, with a diameter of <500 km, close to the Walvis Ridge. They are located mostly beneath the offshore area and are separated from the overlying lithosphere (Figs. 8a, b, d and 10a, b, d). According to the checkerboard tests (Figs. 4 and 5) and the anomaly recovery tests (Figs. S10, S11 and 6), the magnitude of these anomalies, especially the deeper one (H4) might be underestimated, so that the actual velocity contrast of both anomalies could be larger than the ones shown in the model.

Similar features have also been observed in previous tomographic studies. A tomography model (King and Ritsema, 2000) exhibits high-velocity anomalies at sublithosphere depths beneath both coasts of the South Atlantic Ocean in Africa and South America, extending to the mantle transition zone. The anomalies are preserved in the model obtained by including data from surface waves and normal modes (Ritsema et al., 2011). Some indications of the low velocity anomaly L1 can also be found in the model of Emry et al. (2019), where sublithosphere low velocity anomalies are visible beneath the western coastal areas of southern Africa. Anomaly H4 is, however, not visible, possibly related to the lack of resolution capacity of their model in this region.

Pronounced seismic velocity anomalies beneath the Atlantic passive margins are puzzling. They can be caused by currently active mantle processes and/or be related to the former Cretaceous mantle plume or continental break-up event. Hu et al. (2018) provide an explanation in the view of plume-lithosphere interactions, in which the high-velocity

anomaly (H4) may represent the remnant of the original craton lithosphere, that detached from the overlying lithosphere triggered by the Cretaceous mantle plume during the South Atlantic break-up. They proposed that the delaminated lithosphere, being depleted in dense basaltic components, was more buoyant and viscous than the surrounding mantle which allowed it to flow along with the African plate without sinking into the deep mantle (Hu et al., 2018). However, it is still difficult to explain how the delaminated cratonic roots could be able to survive unaffected by mantle convection for 130 myrs and to reconcile this hypothesis in order to explain the low-velocity anomaly above (L1).

An alternative idea associates these anomalies to active mantle processes caused by the edge-driven convection (EDC), as proposed by King and Anderson (1998). The hypothesis was initially proposed for the study area based on a body-wave tomography study (King and Ritsema, 2000) that revealed high-velocity anomalies beneath the South Atlantic coasts of Africa and South America at sub-lithospheric depths extending down to the mantle transition zone and interpreted them as the effects of EDC. Beneath a lithosphere step, where the thickness of the lithosphere changes abruptly, as in the case along the western African coast, a small-scale mantle convection cell is created causing local downwelling of cold lithospheric material into the deeper upper mantle and upwelling of hot asthenospheric materials (King and Anderson, 1998). The EDC might be therefore the dominant mechanism along the continent-ocean transition beneath the western African coast, where the lithosphere thickness changes abruptly from ~200 km onshore to <100 km offshore. In our study area, the lithosphere thickness decreases rapidly from the continental domain beneath the Congo Craton and the Damara Belt towards offshore south of the Walvis Ridge facilitating the EDC. Further south beneath western South Africa, the Paleoproterozoic Kalahari Craton does not maintain a thick lithosphere keel, probably due to the lithosphere thinning suffered during the Gondwana break-up or the interaction of the lithosphere with the Cretaceous mantle plume (Celli et al., 2020). Therefore, the EDC, if it exists, should occur farther east beneath the western margin of the Kaapvaal Craton. However, we do not observe significant velocity variations in this region, probably because the increase of the lithosphere thickness from west to east is gradual (section C-C'). The EDC cell may be larger and involve greater depth range, as seen by King and Ritsema (2000), however, this is beyond the resolution of our model.

We include the EDC convection cell in the interpretation cartoon (Fig. 11), as the model can explain both the high-velocity anomaly H4 and the overlying low velocity-anomaly L1. However, we accept that there could be other models to explain the observations, like the plume-induced lithospheric delamination model of Hu et al. (2018), or other possible delamination/foundering processes, which might currently be under way. More data and multidisciplinary approaches will be needed to finally clarify these questions.

6. Conclusions

We provide a 3-D shear-wave velocity model of the lithosphere beneath southern and southwestern Africa including the offshore areas by modeling multi-mode Rayleigh waves. All observed anomalies can be well explained by the interaction of a mantle plume with the overlying lithosphere in the Cretaceous and the reestablishment of the craton lithosphere since then. The high-velocity anomaly in the uppermost mantle offshore Namibia around the Walvis Ridge, interpreted as magma-depleted and dehydrated lithosphere, may represent the melting zone that developed during the final continental break-up stage, which spatially coincides with the SDRs. We observe a thick lithosphere keel beneath the southernmost Congo Craton and the Damara Belt, which may represent the depleted mantle. This part of lower lithosphere was probably thermally eroded by the arrival of the mantle plume in the early Cretaceous and was finally restored by extraction of melts from the emplaced plume head in a process that possibly also included

dehydration. Beneath the western coast of Namibia, a low-velocity anomaly at 100–200 km depth and a high velocity anomaly at 250–350 km depth are observed, which are clearly separated from the overlying lithosphere. These anomalies can be explained by the effect of the edge-driven convection that may occur at sharp boundaries of the lithospheric thickness variation.

CRediT authorship contribution statement

Shantanu Pandey: Investigation, Writing – original draft, Writing – review & editing. **Xiaohui Yuan:** Conceptualization, Investigation, Data curation, Writing – original draft, Writing – review & editing, Visualization, Funding acquisition. **Eric Debayle:** Methodology, Software, Writing – review & editing. **Wolfram H. Geissler:** Conceptualization, Data curation, Writing – review & editing, Funding acquisition. **Benjamin Heit:** Data curation, Writing – review & editing.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

Data will be made available on request.

Acknowledgements

This work was supported by the Deutsche Forschungsgemeinschaft in the frame of SPP SAMPLE and the Senatskommission für Ozeanographie. Equipment was provided by the Deutscher Geräte-Pool für Amphibische Seismologie (DEPAS) at Alfred Wegener Institute Bremerhaven (AWI) and Deutsches GeoForschungsZentrum Potsdam (GFZ). RV Maria S. Merian was used for the OBS deployment and recovery. The field experiment was partly supported by the GFZ, AWI and the Geological Survey of Namibia. Eric Debayle was supported by the French ANR JIGSAW2 ANR-20-CE49-0001-01. Waveform data for temporary and permanent stations were provided by the GEOFON and IRIS data centers. The model computations and checkerboard resolution test were performed on the High-Performance Computing (HPC) system at Indian Institute of Geomagnetism (IIG), New Panvel, Navi Mumbai. We thank Christoph Sens-Schönfelder for the technical assistance in checking the OBS clock, Nicolas Luca Celli for providing the tomographic model, the editor and three anonymous reviewers for their very constructive reviews.

Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.tecto.2022.229587>.

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