

TRINITY COLLEGE DUBLIN

DOCTORAL THESIS

**Regional-scale structure and dynamics of the
crust, upper mantle and transition zone from
waveform tomography with massive datasets**

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Declaration of Authorship

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Abstract

Regional-scale structure and dynamics of the crust, upper mantle and transition zone from waveform tomography with massive datasets

by Nicolas Luca CELLI

In the last decade, the number of seismic stations deployed globally increased dramatically, allowing to construct tomographic models with increasing resolution. Seismic data coverage however, is not homogeneous across the globe, and many regions are sampled by a distribution of ray paths that is both lower than average and uneven. In this work, we compute regional-scale tomographic models from massive, global waveform datasets, that we optimise for Africa, South America and the South- and North Atlantic Oceans, where coverage is highly heterogeneous. To maximise coverage in the study areas, we assemble a very large dataset of both regional and global teleseismic waveforms, retrieving all freely available data in the Atlantic Ocean and surrounding continents. We then invert the waveforms using the Automated Multimode Inversion (AMI) of S-, Multiple S- and Surface waves. AMI produces a set independent linear equations with uncorrelated uncertainties for each source-receiver path, describing the path-average P- and S-wave velocity structure and dispersion curves within approximate sensitivity kernels. We then combine all the equations in a linear system and solve it for the 3D distribution of P- and S-wave velocities and 2Ψ azimuthal anisotropy in the crust, upper mantle and transition zone. Similarly, we combine all phase velocity dispersion curves and invert them to produce 2D phase velocity maps independently at different periods. Finally, we exploit the mutual consistency of our very large dataset to automatically identify and remove the least consistent measurements from both our 3D models and phase velocity maps. In order to obtain the best possible models, for each study area we compute a different model that is tuned to yield the best results in the region. In South America and the South Atlantic Ocean, we parametrise our 3D model SA2109 on a ~ 300 km triangular grid and fine tune its regularisation with the aid of spike tests to yield robust results across the area. In South America, we image a cratonic lithosphere that is more complex than previously proposed and identify the boundaries of the cratonic cores of the Paraná and Parnaíba cratons; we also retrieve clear images of the Nazca slab, the Pampean slab gap, the subduction of the Chile Rise and the southernmost end of the subduction. In the South Atlantic Ocean, we image the low velocities of the Mid-Ocean Ridge (MOR) and a number of hotspots.

Using age-averaging techniques, we are able to identify the different cooling of the oceanic lithosphere in different oceanic basins, as well as the seismic signature of the Tristan da Cunha hotspot, masked by the MOR in tomographic images. In Africa, our 3D model AF2109 reveals the presence of many smaller cratonic cores within the previously proposed boundaries of the West African and Congo cratons. Comparing the model with the outlines of cratons from field geology, we find that under the crust of the Angola, Tanzania and southwestern Kaapvaal shields, the craton lacks its lithospheric roots. By using a global dataset of diamondiferous kimberlites, we assess that in these areas, thick cratonic lithosphere was once present. Comparing the location and age of the kimberlites with our tomographic images, we infer that the Angola, Tanzania and Kaapvaal cratons lost a large portion of their cratonic lithosphere in the past 200 Million years (M.y). By combining the age of the youngest diamondiferous kimberlites in the area with Large Igneous Provinces and reconstructing the location of the cratonic erosion during the past 150 M.y., we infer that the erosion of the cratonic lithosphere followed the impact of mantle plumes on the cratons. In the Northeast Atlantic, we compute the 3D tomographic model NAT2019, which we parametrise over a very dense ~ 120 km grid. To tackle the very uneven coverage in the area, the model is regularised with 3D-varying coefficients that change in concert with data sampling. The model reveals the presence of a large, low-velocity anomaly, rooted under eastern Greenland in the transition zone, that rises, upwards and eastwards, towards Iceland. At 50-100 km depth, the low velocity anomalies distributes along the Reykjanes and Kolbeinsey Ridges. We interpret the low-velocity body as the Iceland Plume, captured by the MOR. Our model shows thinner lithosphere in the western part of the Northeast Atlantic, under the Greenland Plate. This is consistent both with the location of the ascent of the plume we image and the distribution of seamounts, more numerous on the Greenland Plate. Finally, we use the phase velocity dispersion curves produced by AMI to compute a set of 98 Love and Rayleigh phase velocity maps on a grid with average 225 km spacing that allows to extract densely spaced, robust dispersion curves anywhere on the globe. The data coverage at most periods is global, and the phase velocities, although computed independently for each period, vary smoothly across the produced maps. Compared to previous datasets, our maps sample the presence of strong lateral heterogeneities in the scale of few hundreds of kilometres. At short and intermediate periods, where the range of the sampled depths is narrowest, we can match the results to known geological structures such as cratons, orogenic belts and mid-ocean ridges. We then extract dispersion curves at all points of the model grid, that show, when clustered together, distinct trends for the oceanic and continental lithosphere. We tested the resolution of our dispersion measurements at a regional scale in northwestern Mongolia; at short periods, our dispersion curves sample the different structure between the Hangai Dome and the neighbouring Lake Region, ~ 500 km apart, with lower velocities underneath the Hangai Dome, in agreement with its higher elevation.

Preface

This thesis is based on a collection of two articles submitted to peer-reviewed journals and in accepted or published (chapters 3, 4), one manuscript in preparation for submission (chapter 5), and one manuscript that will contribute to a separate publication I will co-author (chapter 6).

The results presented were generated using programs written mostly in Unix Shell, C, C++, Fortran, and Python. All programs were compiled and run on Ubuntu Linux distributions.

All figures figures were produced using the Generic Mapping Tools (GMT, Wessel et al., 2013) and ParaView (<https://www.paraview.org>).

Statement of co-authorship

Chapter 3 has been accepted for publication in *Geophysical Journal International*. I obtained and processed the newest data since 2012, computed the tomographic models, analysed and interpreted them and made the figures. Andrew Schaeffer and Matteo Ravenna obtained part of the data; Andrew Schaeffer and Carmen Gaina contributed to the discussion and comments on the text. I wrote the manuscript with input from Sergei Lebedev, my co-supervisor at the Dublin Institute for Advanced Studies (DIAS).

Chapter 4 has been published in *Nature Communication* (Celli, N.L., Lebedev, S., Schaeffer, A.J. et al. African cratonic lithosphere carved by mantle plumes. *Nat Commun* 11, 92 (2020) doi:10.1038/s41467-019-13871-2). I obtained and processed the newest seismic data since 2012, analysed the kimberlite location dataset, computed the tomographic model, performed the analysis and interpretation of the results and made the figures. I wrote and ideated the manuscript with close input from Sergei Lebedev. Andrew Schaeffer obtained part of the seismic data and provided valuable discussion on the relationship between cratons and kimberlites. Carmen Gaina provided the plate reconstructions and valuable discussion on the interaction between hotspots and cratons.

Chapter 5 is in preparation for submission. I obtained and processed the newest seismic data since 2012, modified the inversion code for variable resolution, computed the tomographic model, analysed and interpreted the results and made the figures. Andrew Schaeffer obtained part of the seismic data and provided valuable input for discussion. I wrote the manuscript with input from Sergei Lebedev and Carmen Gaina.

Chapter 6 will not be presented as a separate publication, but its results will be used as input data for a separate publication by Javier Fulla, in preparation. I collected the dispersion data and obtained processed the newest ones since 2012, adapted the inversion code to work on source-station tomography using big data with the advice of Sergei Lebedev, analysed, interpreted the results, wrote the text and made the figures.

“Have You Tried Turning it Off and On Again?”

Roy Trenneman

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*To my mother Sara, who always encouraged me to pursue my
scientific career*

Chapter 1

Introduction

Seismology has been used as a tool for investigating the structure of the Earth's interior since the beginning of the 20th century. The pioneering work of early seismologists led to the discovery of the main discontinuities that divide the earth into inner and outer core, mantle and crust (Lehmann, 1936; Oldham, 1906; Mohorovicic, 1909) based on observations of seismic wave phases that matched reflections and refractions of at these—at the time only hypothesised—discontinuities. During that time, analytical relationships had been developed that define the dependence of body wave speed from travel-time measurements with depth (e.g. Weichert and Geiger, 1910). In the following decades, similar relationships were found by Takahashi (1955) and Takahashi (1957) for determining the depth dependence of surface waves from their phase velocity.

While the subdivision of the Earth into inner- and outer core, mantle and crust still stands and appears on virtually every modern basic Earth science textbook, the scientific community has been aware since the 1960s that a heterogeneous, anisotropic Earth was necessary to explain the complex seismic signals recorded by an ever-growing global network of seismic stations. A major leap forward in the interpretation of these signals was, according to Nolet (2008), a presentation by Japanese seismologist Keiti Aki (Aki, 1974) who described the recorded P-wave arrival times as an "image" of the Earth at depth, giving birth to what is known by today as seismic tomography. By combining many measurements in a single system of equations, seismic tomography is able to use their mutual consistency to solve for a model of the Earth's heterogeneities in 3D. At the same time, exploiting the dependency of seismic velocities on the properties of the material sampled by the propagating wave, these "seismic images" can be interpreted as representative of geological structures at depth, offering a result (an image) that is more easily interpretable compared to 1D velocity profiles.

If seismic tomography relies on the combination of seismic data, its efficacy depends strongly on the distribution and availability of seismic stations. Early tomographic models (e.g. Dziewonski, 1977; Masters et al., 1982; Woodhouse and Dziewonski, 1984; Nataf, Nakanishi, and Anderson, 1986) relied on relatively few seismic stations and were able to

sample heterogeneities in the scale of thousands of kilometres. From the 1990s in particular, the number of broadband and very broadband seismic stations deployed worldwide increased dramatically, allowing for the construction of a number of global models that were able to improve the resolution of the heterogeneities to the scale of hundreds of kilometres (e.g. Montagner and Tanimoto, 1991; Su, Woodward, and Dziewonski, 1992; Trampert and Woodhouse, 1995; Laske and Masters, 1996; Grand, Hilst, and Widyantoro, 1997). In later years, the availability of data became such, that today, models can be constrained by millions of seismograms (e.g. Amaru, 2007; Schaeffer and Lebedev, 2013).

The increase in seismic data coverage was also supported by the improvement in tomographic techniques. The first tomographic models, constructed with P-wave travel-time measurements, were able to sample the large-scale heterogeneities in the lower mantle (e.g. Dziewonski, 1977). In the following years, normal modes (Masters et al., 1982) and surface waves were used to model the upper mantle (Tanimoto and Anderson, 1984; Woodhouse and Dziewonski, 1984; Nataf, Nakanishi, and Anderson, 1986) and its anisotropy (Montagner and Nataf, 1986; Montagner and Tanimoto, 1991). Until the 1990s, tomographic models were relying mostly on ray theory and its surface wave equivalent, the Jeffreys-Wentzel-Kramers-Brillouin approximation (JWKB, Dahlen and Tromp, 1998) for the computation of synthetic seismograms, which considers the influence of scattering negligible. The study of the sensitivity of seismic waves (e.g. Wielandt, 1987; Li and Tanimoto, 1993) brought into focus the shortcomings of ray theory in the presence of strong heterogeneities. Improvement over ray theory have been made by describing path-average perturbations with accurate crustal models and approximate sensitivity areas (Nolet, 1990; Lebedev et al., 2005). By modelling the cross-branch coupling of normal modes of the earth, Li and Romanowicz (1996) were able to implement finite-frequency effects (wave scattering) on the sensitivity area of body waves in a tomographic model, starting the field of finite frequency tomography. In the following years, Bijwaard and Spakman (2000) implemented the effects of ray bending in a whole-mantle, P-wave tomographic model through a non-linear inversion process. At the same time, the different sensitivities of different seismic waves lead to the development of tomographic models that relied on the modelling of the full waveform, including data from both surface- and body waves—and in some cases normal modes—that were thus able to simultaneously sample, although with varying accuracy, the crust and large part of mantle (e.g. Megnin and Romanowicz, 2000; Lebedev and Hilst, 2008; Fichtner et al., 2009; Ritsema et al., 2011; Debayle and Ricard, 2012; Rickers, Fichtner, and Trampert, 2013; Schaeffer and Lebedev, 2013; French and Romanowicz, 2015). Similarly to body- and surface wave based models, these studies were computed using both ray- and finite frequency theory, with the former relying on larger datasets and increased coverage and the latter on more accurate modelling of the waveforms, yielding often comparable results (Trampert and Spetzler, 2006).

Modern tomographic models produced with different techniques can now count on thousands of seismic stations across the globe and efficient inversion algorithms to exploit them, allowing the imaging of heterogeneities in the upper mantle in the scale of few hundreds of kilometres. Thanks to the improved resolution, seismic tomography produced in the last decades astonishing images of the structure of the continents and oceans.

In the oceans, the long-wavelength structure of the oceanic lithosphere and underlying asthenosphere was explored by a large number of tomographic studies in the '90s (Zhang and Tanimoto, 1992; Zhang, Tanimoto, and Stolper, 1994; Tanimoto and Zhang, 1992; Silveira et al., 1998), that resolved both the cooling and thickening of the oceanic lithosphere and the underlying convecting mantle. At the same time, these images revealed the presence of localised low-velocity anomalies corresponding to the location of intraplate hotspots (Zhang and Tanimoto, 1992). In recent years, the deployment of more land stations and Ocean Bottom Seismometers (fundamental to improve on the relatively few source-station paths in the oceans) and improvement in waveform modelling techniques (important for the resolution of thin, deep features such as mantle plumes) brought into focus the presence of low velocity anomalies extending sometimes deep in the mantle under many hotspots and seamounts. On a global scale, Montelli et al. (2006) examined the seismic signature of 35 hotspots worldwide, making a catalogue of the ones that could be interpreted as deep mantle plumes using finite-frequency tomography. Years later, French and Romanowicz (2015) produced stunning images of broad, low-velocity anomaly columns, interpreted as mantle plumes, rising from the deep mantle under many hotspots and ranked them in terms of resolution. The number of tomographic studies targeting proposed mantle plumes however is countless (e.g. Wolfe et al., 1997; Allen et al., 1999; Bijwaard and Spakman, 1999; Ritter et al., 2001; Foulger et al., 2001; Rhodes and Davies, 2001; Debayle, L  v  que, and Cara, 2001; Allen et al., 2002; Ritsema and Allen, 2003; Weeraratne et al., 2003; Zhao, 2004; Pilidou et al., 2005; Montelli et al., 2006; Waite, Smith, and Allen, 2006; Koulakov, 2007; Montagner et al., 2007; Nolet, Allen, and Zhao, 2007; Sicilia et al., 2008; Chang and Van der Lee, 2011; Chen et al., 2015; Schl  mer et al., 2017; Steinberger et al., 2018; Civiero et al., 2018; Maguire et al., 2018), and differences between models motivated skepticism towards the capability to resolve such deep, thin features, with a number of studies proposing alternative explanations for the low-velocity anomalies observed beneath hotspots (e.g. Anderson, Tanimoto, and Zhang, 1992; Keller, Anderson, and Clayton, 2000; Foulger et al., 2001; Foulger and Pearson, 2001; Foulger, 2002; Foulger and Anderson, 2005). Many hotspots in fact locate in very remote areas and the scarce and uneven coverage of seismic data makes the resolution of the thin, column-like features expected for mantle plumes even

more challenging. To increase the local coverage, ad-hoc seismic networks have been deployed on a number of hotspots, e.g.: PLUME experiment in Hawaii (Laske et al., 2009); ICEMELT experiment in Iceland (Bjarnason et al., 1996); RHUM-RUM experiment in La Reunión (Barruol, Sigloch, and Group, 2017); Tristan da Cunha (Geissler, 2014). Many of the other, oceanic hotspots on the globe however are topped by only few stations and are sampled only by global or large-scale tomographic studies. It is striking however, that the structure—and, in some cases, very existence—of mantle plumes even under some of the well known hotspots like Iceland (Wolfe et al., 1997; Allen et al., 1999; Bijwaard and Spakman, 1999; Foulger et al., 2001; Allen et al., 2002; Rickers, Fichtner, and Trampert, 2013), Tristan da Cunha (Colli, Fichtner, and Bunge, 2013; Schlömer et al., 2017; Bonadio et al., 2018) and Afar (Sebai et al., 2006; Benoit, Nyblade, and VanDecar, 2006; Montagner et al., 2007; Sicilia et al., 2008; Chang and Van der Lee, 2011; Adams and Nyblade, 2011; Adams, Nyblade, and Weeraratne, 2012; Hansen, Nyblade, and Benoit, 2012; Civiero et al., 2018) is still debated, with uncertainties about the shape of the plumes, their source and even their number.

In the continents and in other regions well covered by seismic stations, tomography relying on P-wave travel-times achieved in the 1990s cell sizes of hundreds of kilometres, with models capable of extremely good resolution in the areas well-covered by ray paths (Spakman, Lee, and Hilst, 1993; Hilst, Widyantoro, and Engdahl, 1997; Bijwaard, Spakman, and Engdahl, 1998; Zhao, 2001). These models provided images of the subducting slabs (e.g. Spakman, Lee, and Hilst, 1993; Hilst, Widyantoro, and Engdahl, 1997) and upwelling beneath hotspots (e.g. Bijwaard, Spakman, and Engdahl, 1998) that crossed the mantle transition zone, providing evidence for the concept of whole mantle circulation. In the upper mantle, the presence of fast seismic velocities in the lithosphere under the Archean and Paleoproterozoic portions of the continents was well known since the 1980's (e.g. **Grand1984a**). The shape of these units however could be retrieved only with the development of models capable of resolving features of few hundreds of kilometres (Polet and Anderson, 1995), and their vertical layering, to which tomography is less sensitive, remains a matter of debate until today (Lebedev, Boonen, and Trampert, 2009; Eaton et al., 2009; Yuan and Romanowicz, 2017; Ravenna et al., 2018). Since the 2000's, tomographic models were finally able to resolve the relevant scales needed for imaging the architecture of the continents (e.g. Weeraratne et al., 2003; Fouch et al., 2004; Darbyshire et al., 2004; Lee and Frederiksen, 2005; Fishwick, Kennett, and Reading, 2005; Chevrot and Zhao, 2007; Darbyshire and Lebedev, 2009; Lebedev, Boonen, and Trampert, 2009; Begg et al., 2009; Fishwick, 2010; Legendre et al., 2012; Schaeffer and Lebedev, 2014; Dave and Li, 2016; Debayle, Dubuffet, and Durand, 2016; Petrescu et al., 2017; Lebedev et al., 2018). The deployment of dense networks like the USArray and the consequent increase in data provides the tools to identify separate cratonic units in areas where geological sampling is scarce and previous, smoother models resolved larger, single structures (e.g.

Greenland, Lebedev et al., 2018). In some of the less-well sampled cratons (e.g. Africa, Siberia, Antarctica and South America) especially, the very heterogeneous coverage imposes lower resolution on tomographic models, so that some of what are thought to be among the largest cratonic provinces are still relatively poorly constrained.

1.1 Motivation and outline

In both continents and oceans, large regions of the globe remain less sampled by seismic stations to today. In these areas, tomography is challenged by the scarcer and heterogeneous coverage. Since these areas are typically difficult to reach (e.g. Saharan desert, Amazon forest, Oceans...) independent geological information is limited as well, hindering the validation of the obtained results through comparison to exposed geological structures.

In these areas, in order to resolve the fine-scale details of the lithosphere and mantle, seismic tomography must be able to take advantage of the large number of seismic data available today, which requires considerable computational effort even with modern infrastructure. Methods based on the path-average approximation are typically capable of implementing larger amounts of data compared to inversions using the finite-frequency approach. While this comes at a cost in terms of accuracy in the waveform modelling, ray-based inversions compensate by better constraining the tomographic inversion (Trampert and Spetzler, 2006). In areas where data are both scarce and uneven, coverage is paramount, and the capability to process a big dataset of ray-based methods desirable. Among the ray-based inversion approaches, the Partitioned Waveform Inversion (PWI, Nolet, 1990) is particularly suited to invert large amounts of data. The PWI reduces the computational size of the inversion process by partitioning it into independent inversions for the average V_P and V_S constraints for each source-receiver path, followed by an inversion of all path-average equations for a 3D model of the Earth's structure. This process, refined and automated in the Automated Multimode Inversion (AMI, Lebedev et al., 2005), allows efficient extraction of structural information from millions of seismograms, and has been used to compute a number of global (Lebedev and Hilst, 2008; Schaeffer and Lebedev, 2013; Schaeffer, Lebedev, and Becker, 2016) and continental-scale (Lebedev and Nolet, 2003; Legendre et al., 2012; Schaeffer and Lebedev, 2014; Lebedev et al., 2018) tomographic models.

In this thesis, I investigate the structure and evolution of the lithosphere and underlying mantle on a regional scale by applying AMI to a very large database of over 1.5 million waveform fits and computing tomographic models that are fine-tuned to the region of interest. We apply this technique to areas with very heterogeneous coverage of seismic stations: South America, the South Atlantic Ocean, Africa and the North Atlantic

Ocean and surrounding continents. We also compute a set of global phase-velocity maps on densely parametrised grids from which we can extract broadband dispersion curves at any point on the globe.

By including new, regional data and carefully optimising each model to achieve the highest resolution possible in each area, we resolve details of the continental and oceanic lithosphere and mantle that were unseen in previous models.

Chapter 2 contains a summary of the theoretical framework necessary to understand the computation of the tomographic models discussed in the further chapters. Specifically, the first sections of the chapter starts from the formulation of the wave equation, the description of Body- and Surface waves and how these can be represented in terms of normal mode summation, providing thus the synthetic seismograms used for the waveform inversion. In the following sections, the main features of the inversion scheme used to construct the final models are outlined.

Chapter 3 describes the features and construction of the new model SA2019, generated from the application of AMI to a very large dataset combining global and regional measurements. We discuss the optimisation of the model to resolve the area covering the South Atlantic and South America, examine the challenges posed by the locally lower seismic data coverage and how these can be overcome through the use of waveform tomography with big data. We then discuss in detail the choice of regularisation parameters through quantitative analysis of resolution tests and its effects on the local structures. After this, the model is presented, its main features are outlined and it is compared to other S-wave tomographic models. We then present a detailed discussion of the new findings derived from the improved resolution on the outlines of cratons and mobile belts, and the relationships between volcanic areas, thinner lithosphere and structure of the subducting slab in South America. Finally we discuss the interconnected low-velocity patterns linking Mid-Ocean Ridges and Hotspots in the South Atlantic Ocean and use age-averaging techniques to separate the seismic signal of the cooling oceanic lithosphere from its local anomalies.

Chapter 4 combines a newly produced tomographic model of the region with a large, global Kimberlite dataset to gain insights on the evolution of the cratonic lithosphere in Africa. At first we examine the model and validate the new details on the structure of the African cratons against a compilation of geological information at the surface. After that, we use the present-day images of the cratons provided by our model and the constraints on their past presence from diamondiferous Kimberlites to highlight areas indicative of cratonic erosion. Finally, we discuss the possible mechanisms of cratonic erosion by using a combination of Kimberlite ages, hotspot locations, plate reconstructions, tomographic images and local tectonic history.

Chapter 5 presents NAT2019, a new model sampling the Iceland Hotspot from the crust to the transition zone, parametrised over an extremely dense model grid of approximately 120 km spacing laterally. The model is outlined at first, and is compared to both P- and S-wave tomographic models. We discuss then the correlation between the complex low-velocity body we observe and plume-related features at the surface, inferring its nature as a mantle plume. Using age-averaging over the model, we discuss the lower velocities confined west of the North Atlantic Ridge and compare them to a recent database of seamount features. We examine then the correlation of the low velocities rising under Iceland from the transition zone with evidences of past and present volcanism, discussing the interplay of the Iceland Plume, Greenland Cratons and the North Atlantic Ridge.

Chapter 6 presents the results from the inversion of the phase velocity measurements produced by AMI in a set of 98 global, fundamental mode phase velocity maps for both Rayleigh- and Love waves in a 20-360 s period range. We discuss the main features of the maps and compare them to previous datasets. We then extract a global set of densely distributed dispersion curves from the inverted maps and examine the capability to resolve local structure at any point on the globe. Finally, we discuss the local resolution of the phase velocity maps in a case study on the Hangai Dome in northern Mongolia.

Chapter 7 discusses the main conclusions of the work carried out in this thesis.

Chapter 2

Theoretical framework

The theory of seismic wave propagation and its uses in tomographic inversion is vast, and numerous comprehensive treaties have been written that cover these topics, e.g.: Dahlen and Tromp (1998), Aki and Richards (1980), Nolet (2008), and Stein and Wysession (2003). For the purpose of this thesis, in this chapter I will summarise the theoretical framework that casts the basis for the computation of the tomographic models discussed in the next chapters. Specifically, I will first discuss the elements constituting the wave equation and introduce some of the common assumptions on the medium traversed by the seismic waves. After that, I will describe the equations that govern the propagation of body waves. Following body waves, I will describe surface waves and how we can represent them in terms of normal modes of the Earth, effectively describing the computation of synthetic seismograms. Finally, I will describe the waveform inversion schemes used to produce our models, starting from the Partitioned Waveform Inversion (PWI) (Nolet, 1990) and then moving to the Automated Multimode Inversion (AMI) (Lebedev et al., 2005) and the following 3D linear inversion (Lebedev and Hilst, 2008; Schaeffer and Lebedev, 2013). If not expressly specified in the text, I refer the reader to two books that extensively treat the subject: Nolet (2008) and Dahlen and Tromp (1998). For further details on specific topics, I will refer directly to the relevant published academic papers.

2.1 Seismic waves sampling the Earth Structure

Earthquakes excite seismic waves that propagate through the Earth. Through seismometers, we can sample these waves and record on seismograms the resulting displacement (or its derivatives velocity and acceleration) at their location as a function of time. The amplitude and direction of the displacement depend on the characteristics of the earthquake source, the properties of the media traversed by the seismic wave from the source to the receiver and the sensitivity of the recording instrument. Assuming the knowledge of the receiver term, seismograms contain a wealth of information on the earthquake source and the structure of the Earth lying between it and the receiver. In order to understand the nature of the dynamic processes in the Earth and how these can be inferred from the study of seismic wave recordings, it is important to describe the relationships

between the forces acting in a medium and its properties.

2.2 Stress, strain and the equations of motion

We can describe the forces F acting on a generic 3D volume element dV by the stress tensor σ_{ji} , which describes the forces acting on the plane j in the direction i . In a system with orthogonal axes x_1, x_2, x_3 , and in a situation of equilibrium we can write:

$$F_i = \sum_{j=1}^3 \frac{\sigma_{ji}}{x_j} dV \quad (2.1)$$

By adding a volume force $f dV$, we induce a small displacement u in the element volume dV . By using Newton's law $F = ma$, we can relate displacement and its derivative, acceleration $a = \partial^2 u_i / \partial t^2$ to the density of the material ρ , obtaining what is known as the elastodynamic equation:

$$\rho \frac{\partial^2 u_i}{\partial t^2} = \sum_j \frac{\partial \sigma_{ji}}{\partial x_j} + f_i \quad (2.2)$$

With the source term f_i negligible when we are far from the source. A constant displacement u will result in a translation of the volume, and its rotation and deformation are given by its spatial derivatives $\partial u_i / \partial x_j$. In the case of pure rotation, $\partial_i u_j = -\partial_j u_i$. We can define thus a rotation tensor:

$$\Omega_{ij} = \frac{1}{2} \left(\frac{\partial u_j}{\partial x_i} - \frac{\partial u_i}{\partial x_j} \right) \quad (2.3)$$

And a symmetric strain tensor:

$$\epsilon_{ij} = \frac{1}{2} \left(\frac{\partial u_j}{\partial x_i} + \frac{\partial u_i}{\partial x_j} \right) \quad (2.4)$$

The diagonal elements of ϵ_{ij} indicate changes in volume, while the other, symmetrical ones indicate shear strain. Combined, the rotation and strain tensors fully describe the spatial derivatives of the displacement u .

2.3 Seismic wave equation

In a three-dimensional, perfectly elastic medium, stress in a single direction influences all components of the strain field, with a linear relationship between stress and strain controlled by the elastic tensor c_{ijkl} :

$$\sigma_{ji} = \sum_{kl} c_{ijkl} \epsilon_{kl} \quad (2.5)$$

The symmetries of stress and strain, together with other, additional ones (see Nolet, 2008 and references therein) simplify the elastic tensor from 81 to 21 elements. In an anisotropic, elastic medium, the elastodynamic equation becomes:

$$\rho \frac{\partial^2 u_i}{\partial t^2} = \sum_{jkl} \frac{\partial}{\partial x_j} \left(c_{ijkl} \frac{\partial u_l}{\partial x_k} \right) + f_i \quad (2.6)$$

If we simplify to the case of an isotropic, elastic medium, the elastic tensor depends only on the bulk modulus κ and the shear modulus μ , which are properties of the material.

$$\begin{aligned} c_{ijkl} &= \left(\kappa - \frac{2}{3}\mu \right) \delta_{ij}\delta_{kl} + \mu(\delta_{il}\delta_{jk} + \delta_{ik}\delta_{jl}) \\ \lambda &= \kappa - \frac{2}{3}\mu \end{aligned} \quad (2.7)$$

and the relationship between stress and strain becomes a function of the Lamé parameters λ and μ :

$$\begin{aligned} \sigma_{ji} &= [\lambda \delta_{ij}\delta_{kl} + \mu(\delta_{il}\delta_{jk} + \delta_{ik}\delta_{jl})] \epsilon_{kl} \\ &= \lambda \delta_{ij}\partial_k u_k + \mu(\partial_i u_j + \partial_j u_i) \end{aligned} \quad (2.8)$$

If we substitute into the elastodynamic equation:

$$\rho \frac{\partial^2 u_i}{\partial t^2} = \partial_i \lambda \partial_k u_k + \partial_j \mu (\partial_i u_j + \partial_j u_i) + \lambda \partial_i \partial_j u_k + \mu \partial_i \partial_j u_j + \mu \partial_j^2 u_i, \quad (2.9)$$

we can simplify using the curl $\nabla \times \mathbf{u}$ and divergence operators $\nabla \cdot \mathbf{u}$, and assuming small gradients in the Earth structure, such that $\nabla \lambda \approx 0$ and $\nabla \mu \approx 0$:

$$\rho \partial_j^2 \mathbf{u} = (\lambda + 2\mu) \nabla \nabla \cdot \mathbf{u} - \mu \nabla \times \nabla \times \mathbf{u} \quad (2.10)$$

Which is the equation of a seismic wave propagating in an isotropic, elastic, homogeneous medium. The homogeneous approximation is responsible for the elision of the first two terms of equation 2.9. We know by now that the Earth is a heterogeneous medium, but we can use the homogeneous approximation by assuming that the changes in material properties are small ($\nabla \lambda \approx 0$ and $\nabla \mu \approx 0$), with variations over distances much larger than the wavelength of seismic waves. This approximation is at the base of ray theory, which allows the presence discontinuities by using the reflection and transmission of seismic wave amplitudes.

2.4 Body waves

By taking equation 2.10 and applying the divergence $\nabla \cdot \mathbf{u} = \theta$ and curl $\nabla \times \mathbf{u} = \varphi$ operators, we obtain, respectively:

$$\begin{aligned}\rho \partial_t^2 \theta &= (\lambda + 2\mu) \Delta \theta \\ \frac{1}{\alpha^2} \partial_t^2 \theta &= \Delta \theta\end{aligned}\tag{2.11}$$

$$\begin{aligned}\rho \partial_t^2 \varphi &= \mu \Delta \varphi \\ \frac{1}{\beta^2} \partial_t^2 \varphi &= \Delta \varphi\end{aligned}\tag{2.12}$$

The values $\alpha = \sqrt{(\lambda + 2\mu)/\rho}$ and $\beta = \sqrt{\mu/\rho}$ describe thus the variations in displacement in time as a function of the Lamé parameters. They define thus the velocity of the propagating seismic wave.

Any vector field can be decomposed into a vector and scalar field. By applying this to the displacement $\mathbf{u}$ by using Helmholtz's theorem, we define:

$$\mathbf{u} = \nabla \Phi + \nabla \times \mathbf{\Psi},\tag{2.13}$$

and the curl and divergence:

$$\begin{aligned}\nabla \cdot \mathbf{u} &= \nabla^2 \Phi \\ \theta &= \Delta \Phi \\ \nabla \times \mathbf{u} &= -\nabla^2 \mathbf{\Psi} \\ \varphi &= -\nabla^2 \mathbf{\Psi}\end{aligned}\tag{2.14}$$

We can see that the potential Φ , travelling with speed α , has no rotation and is thus called the P(ressure)-wave potential, with α being the P-wave velocity. Conversely, the potential $\mathbf{\Psi}$ has no volume change, and it is called the S(hear)-wave potential, with β being the S-wave velocity. The S-waves have non-zero particle motion on the two axes perpendicular to the propagating direction, and are thus commonly decomposed into vertical, S_V , and horizontal S_H . Due to the spherical nature of the wavefront radiated from an earthquake source, body waves travel deep into the Earth and their amplitudes decay spherically with the radius r^{-1} . The path a body wave travels is controlled largely by the reflection and refraction of the waves at Earth's discontinuities in depth.

2.5 Surface waves

While in an infinite, homogeneous medium we would obtain only body waves, the existence of the Earth's free surface poses the conditions for the propagation of surface waves. Surface waves are generated by the interference of body waves in a layered

half-space. Similar to S-waves, interference of P and S_V waves generates the vertically-polarised Rayleigh waves and interference of S_H waves generates horizontally polarised Love waves. Since they are confined to the surface, surface wave amplitudes decay cylindrically with $\sqrt{r}^{-1}$, making them the most energetic arrivals on most seismograms. Their propagation near the surface implies also that they sample only the shallower structure of the Earth, although we will see that long-period surface waves provide valuable information on a large part of the mantle. We will see in the next section how travelling surface waves can be described by high-order standing waves known as the Earth's normal modes.

2.5.1 Normal modes

The existence of the Earth's free surface constrains the elastodynamic equations discussed in the previous chapters. In the case of a non-rotating Earth, the elastodynamic equations for an isotropic, elastic medium (equation 2.6) can be re-written in spherical coordinates (r, θ, ϕ) , in the time and frequency domain, respectively:

$$\begin{aligned}\rho \frac{\partial^2 \mathbf{u}}{\partial t^2} + \mathcal{L}\mathbf{u} &= \mathbf{f} \\ \mathcal{L}\mathbf{u} &= \omega^2 \rho \mathbf{u} + \mathbf{f}\end{aligned}\tag{2.15}$$

With the term $\mathcal{L}$ including the derivatives of displacement and the elastic tensor. The use of spherical coordinates makes the solution to the elastodynamic equations a harmonic function of time and longitude ($e^{i\omega t}$, $e^{im\phi}$, respectively), co-latitude (expressed by the Legendre function $P_l^m(\cos\theta)$) and the radius r , indicating the modes of vibration of the earth. Due to the non-negligible effects of gravity for very low frequencies, the source term $\mathbf{f}$ is retained. The displacement can be thus defined as:

$$\mathbf{u}(\mathbf{r}, t) = \exp(-i\omega_l^m t) {}_n\mathbf{u}_l^m\tag{2.16}$$

And the term ${}_n\mathbf{u}_l^m$ can be written in harmonic functions:

$$\begin{aligned}{}_n\mathbf{u}_l^m &= {}_nU_l(r) Y_l^m(\theta, \phi) \hat{\mathbf{r}} \\ &+ \nu^{-1} [{}_nV_l(r) \partial_\theta Y_l^m(\theta, \phi) + (\sin\theta)^{-1} W_l(r) \partial_\phi Y_l^m(\theta, \phi)] \hat{\boldsymbol{\theta}} \\ &+ \nu^{-1} [{}_nV_l(r) (\sin\theta)^{-1} \partial_\phi Y_l^m(\theta, \phi) - {}_nW_l(r) \partial_\theta Y_l^m(\theta, \phi)] \hat{\boldsymbol{\phi}}\end{aligned}\tag{2.17}$$

with $\nu = \sqrt{l(l+1)} \approx l + \frac{1}{2}$. The spherical harmonics representing longitude and co-latitude in the term $Y_l^m(\theta, \phi)$ are defined as:

$$Y_l^m(\theta, \phi) = (-1)^m \left[\frac{(2l+1)(l-m)!}{4\pi(l+m)!} \right]^{\frac{1}{2}} P_l^m(\cos\theta) \exp(im\phi),\tag{2.18}$$

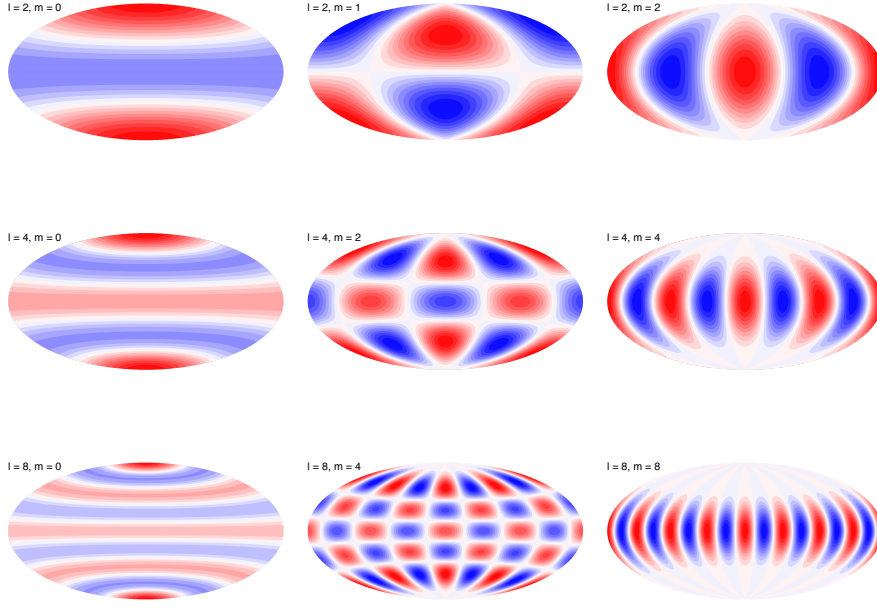


FIGURE 2.1: Plot of some of the $Y_l^m(\theta, \phi)$ spherical harmonics. From <http://www-udc.ig.utexas.edu/external/becker/sh/lmcomp.pdf>

with $P_l^m(\cos\theta)$ the associated Legendre functions. The real part of $Y_l^m(\theta, \phi)$ has thus $2m$ nodal lines in longitude and $l - m$ nodal lines in co-latitude (see examples in Fig. 2.1). The terms m and l and are commonly referred as the azimuthal and angular orders. The functions ${}_nU_l(r)$, ${}_nV_l(r)$ and ${}_nW_l(r)$ are the Earth's radial eigenfunctions and depend on the angular order l and the mode number n . Solving the Earth's eigenfunctions with the boundary conditions of the stress-free surface, the n modes exist only at a set of discrete frequencies ${}_n\omega_l^m$ (Nolet, 2008) (Fig. 2.2). The lowest eigenfrequency is called the fundamental mode, and the higher ones overtones. From the solution of the radial eigenfunctions, we can see how ${}_nU_l(r)$ and ${}_nV_l(r)$ are coupled and ${}_nW_l(r)$ is independent, indicating the presence of two separate types of modes.

The modes depending only on the torsion along $\hat{\phi}$ are called toroidal modes (${}_nT_l^m$), their radial dependency is dictated only by ${}_nW_l(r)$ and they depend only on the shear modulus μ . They thus do not exist in fluids or gases.

Modes oriented along $\hat{\theta}$ are called spheroidal modes (${}_nS_l^m$), they depend both on ${}_nU_l(r)$ and ${}_nV_l(r)$ and depend on both Lamé parameters.

2.5.2 From normal modes to surface waves

For high angular order l , spherical harmonics start to behave similarly to simple harmonic functions, except at the source poles. They create thus a pattern similar to a standing wave, with source at one of the poles. In this section we will thus see how normal modes can describe a travelling wavefront and thus the propagation of seismic waves.

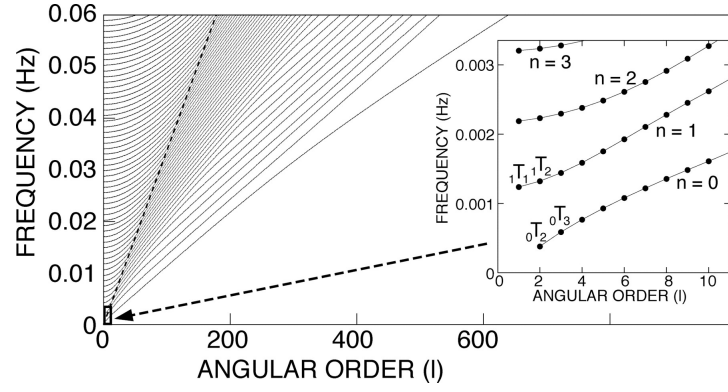


FIGURE 2.2: Frequency of torsional modes as function of l , computed from PREM (Dziewonski and Anderson, 1981). From Cowsik (2007).

For earthquakes, the double-couple source forces have at most two perpendicular nodal planes, so that the azimuthal order $m \leq 2$; joining this with the high l approximation, the asymptotic theory for the Legendre functions gives:

$$Y_l^m(\theta, \phi) \xrightarrow{l \gg 1} \frac{1}{\pi \sin^{1/2} \theta} \cos \left[\nu \theta - \frac{\pi}{4} + \frac{m\pi}{2} \right] e^{im\phi} \quad (2.19)$$

And its derivatives along θ and ϕ :

$$\begin{aligned} \partial_\theta Y_l^m(\theta, \phi) &\xrightarrow{l \gg 1} -\frac{\nu}{\pi \sin^{1/2} \theta} \sin \left[\nu \theta - \frac{\pi}{4} + \frac{m\pi}{2} \right] e^{im\phi} \\ \partial_\phi Y_l^m(\theta, \phi) &\xrightarrow{l \gg 1} \frac{im}{\pi \sin^{1/2} \theta} \cos \left[\nu \theta - \frac{\pi}{4} + \frac{m\pi}{2} \right] e^{im\phi}. \end{aligned} \quad (2.20)$$

Since ν depends directly on the very large l , it becomes the controlling factor on the motion in the θ and ϕ directions. By looking at equation 2.17, we can see how the motion in θ is governed by ${}_n V_l(r)$ and the one in direction ϕ by ${}_n W_l(r)$. Because of the dominance of ν , we can neglect the terms not multiplied by ν . If, for simplicity, we consider only the vertical displacement, in the time domain we can re-write equation 2.16 as:

$$u_r(\theta, \phi, t) = \sum_{m,l,n} Y_l^m(\theta, \phi) {}_n A_l^m \exp(i_n \omega_l t - {}_n \alpha_l t) \quad (2.21)$$

Which describes the displacement u along the radius r in each location on the earth as the summation of all normal modes of the earth.

The term ${}_n A_l^m$ represents the amplitude of the displacement, and it includes the amplitude of ${}_n U_l^m(a)$ at the surface $r = a$ (with a the radius of the Earth), depending in turn on other terms. For surface waves, amplitude depends on the Source term S , the site term at the receiver and the propagation term. The source term depends, for each multiplet, both on the moment tensor and on the mode eigenfunctions ${}_n U_l(r)$, ${}_n V_l(r)$ and ${}_n W_l(r)$ at the source (see Nolet, 2008). Since the mode amplitude varies with $e^{im\phi}$ and a double-couple earthquake source has only two nodal lines, the source does not excite azimuthal

orders greater than 2. The amplitude of the surface wave can be described by:

$$A_n(\omega) = S_n(\omega) \times \frac{e^{i(\pi/4 - \phi\pi/2) - \nu\theta/2Q_n}}{\sqrt{8\pi\nu_n|\sin\theta|}} \times (U_{rri} V_{r,-i} W_r) \quad (2.22)$$

The term ${}_n\alpha_l$ in equation 2.21 describes the quality factor ${}_nQ_l$ of the mode:

$${}_n\alpha_l = \frac{{}_n\omega_l}{2{}_nQ_l}; \quad (2.23)$$

Which describes the decay of the amplitudes with the propagation of the surface wave and represents thus the attenuation term. The higher the frequency, the higher the attenuation. If we rotate the system by putting the origin at the epicentre, ϕ becomes the azimuth of a hypothetical source-receiver path ζ_s and θ becomes directly related to the epicentral distance Δ . If we also define a wavenumber k :

$$\begin{aligned} \Delta &= a\theta \\ k &= \frac{\nu}{a}, \end{aligned} \quad (2.24)$$

we can rewrite the spectrum of displacement as a function of the epicentral distance Δ and the azimuth ζ_s from the source:

$$u_r(\Delta, \zeta_s, \omega) = \sum_n A_n(\omega, \delta, \zeta_s) e^{ik_n(\omega)\Delta}, \quad (2.25)$$

and in the time domain:

$$u_r(\Delta, \zeta_s, t) = \sum_n \int A_n(\omega, \delta, \zeta_s) e^{ik_n(\omega)\Delta - \omega t} d\omega \quad (2.26)$$

The value of the integral is strongly oscillating and amplitudes will interfere destructively if A is a smooth function of the frequency, so that displacement will not be excited. If the exponent is stationary across frequencies however, it will contribute to the displacement:

$$\frac{d}{d\omega} [k_n(\omega)\Delta - \omega t] = 0, \quad (2.27)$$

or:

$$\frac{\Delta}{t} = \left(\frac{d\omega_n}{dk} \right) = \mathcal{U}_n \quad (2.28)$$

Where $\mathcal{U}_n$ has the dimension of a velocity and indicates the propagation of the non-zero displacement u , which appears thus as a travelling wavefront from the source to the antipode. It is known as group velocity. The toroidal and spheroidal modes in this system refer thus to a particle motion in the both vertical- and radial- and transverse direction with respect to the source-receiver path, respectively. They can be thus viewed as the propagation of Rayleigh and Love waves.

2.5.3 Sensitivity of surface waves

In the previous sections, we saw how high- l normal modes can be viewed as travelling wavefronts with wave packets propagating with a group-velocity $\mathcal{U}_n$. By observing the wavenumber $k_n(\omega)$, we can see how planes of equal phase travel with a velocity C_n for which $k_n\Delta - i\omega t$ is constant:

$$\frac{d\Delta}{dt} \equiv C_n(\omega) = \frac{\omega}{k_n(\omega)} \quad (2.29)$$

With $C_n(\omega)$ the phase velocity of a certain frequency. Measurements of $C_n(\omega)$ can be extracted from the wavenumber, relating to the source-station distance Δ . However, this indicates that our capability of measuring phase velocity is constrained by the presence and distribution of seismic stations, which is still sparse and uneven at present day.

We know from that surface waves propagate cylindrically, while body waves propagate spherically in the earth. The theory of ray-mode duality however shows a relationship between the two, indicating that surface waves result from interfering body waves. Recalling the elastodynamic equations in a far-source region, the Rayleigh's Principle relates variations in frequency ω to variations in the Earth model in terms of density and Lamé parameters (included in $\mathcal{L}$):

$$\begin{aligned} \omega^2 \rho \mathbf{u} &= \mathcal{L} \mathbf{u} \\ (\rho + \delta\rho)(\omega_k^2 + \delta\omega^2)(\mathbf{u}_k + \delta\mathbf{u}_k) &= (\mathcal{L} + \delta\mathcal{L})(\mathbf{u}_k + \delta\mathbf{u}_k) \\ \delta\omega^2 &= 2\omega_k \delta\omega = \frac{(\delta\mathcal{L} \mathbf{u}_k, \mathbf{u}_k) - \omega_k^2 (\delta\rho \mathbf{u}_k, \mathbf{u}_k)}{(\rho \mathbf{u}_k, \mathbf{u}_k)}. \end{aligned} \quad (2.30)$$

If we assume spherically symmetric perturbations, we can use the solutions of the Earth's eigenfunctions for Toroidal and Spheroidal modes to parametrise variations in frequency to variations in the seismic velocities V_s and V_p from a background reference model:

$$\begin{aligned} \delta\omega &= \int [K_\kappa \delta\kappa + K_\mu \delta\mu + K'_\rho \delta\rho] dr \\ &= \int [K_P \delta V_P + K_S \delta V_S + K_\rho \delta\rho] dr. \end{aligned} \quad (2.31)$$

If a surface wave can be seen as interfering body waves, then the turning point r_p of the diving body waves constituting it will give us information on its depth sensitivity. By equating the wavenumber $k_n(\omega)$ to the horizontal one for a body wave with travel time

phase ωT and ray parameter p , we can write:

$$\begin{aligned} k_S &= \frac{d(\omega T)}{d\Delta} = \omega p \\ p &= \frac{k_S}{\omega} = \frac{k_n(\omega)}{\omega} = \frac{1}{C_n(\omega)}, \end{aligned} \quad (2.32)$$

and:

$$\begin{aligned} p &= \frac{r_b}{V_S(r_b)} \\ C_n(\omega) &= \frac{a}{r_b} V_S(r_b), \end{aligned} \quad (2.33)$$

showing that the phase velocity can be directly related to the distribution of shear-wave velocity V_S . Under the turning point r_b , the energy decreases exponentially. For the fundamental mode however, the contribution of the part below the turning point is not negligible, and the ray-mode duality relationships differ from the ones for higher modes, with the fundamental mode travelling substantially slower than the other, higher modes. By using Rayleigh's Principle, we can define, for the each parameter m describing V_S , V_P or ρ :

$$\begin{aligned} (\delta k_n)_\omega &= -\frac{(\delta \omega)_k}{\mathcal{U}_n} \\ \left(\frac{\partial k_n}{\partial m} \right)_\omega &= -\frac{1}{\mathcal{U}_n} \left(\frac{\partial \omega_n}{\partial m} \right)_k \\ \left(\frac{\partial C_n}{\partial m} \right)_\omega &= -\frac{C_n}{k_n} \left(\frac{\partial k_n}{\partial m} \right)_k \end{aligned} \quad (2.34)$$

Which indicate the dependency of variations of phase and group velocity on variations of seismic properties of the chosen starting Earth model. In figure (Fig. 2.3), we computed $\partial C_n / \partial V_S$ for Rayleigh waves using the model AKM135 (Kennett, Engdahl, and Buland, 1995) in a wide 0-350 s period range for the fundamental mode and first 5 overtones. It is apparent that the complexity of the sensitivity pattern increases with the mode number. The fundamental mode is thus the simplest, and increasing periods are mostly sensitive to increasing depth. Higher modes on the other hand, are sensitive to greater depths compared to the fundamental mode at any given frequency.

2.6 Waveform inversion

So far, we discussed how to represent surface- and body waves in terms of normal modes. Most of the work carried out in this thesis relies on the inversion of recorded surface and body waves to find an Earth model that is able to successfully fit the observed wave phases. In the next sections I am going to discuss the inversion procedures that lay the foundations of the Partitioned Waveform Inversion (PWI, Nolet, 1990) and Automated Waveform Inversion (AMI, Lebedev et al., 2005), which have been used to compute the tomographic models discussed in the next chapters of this thesis.

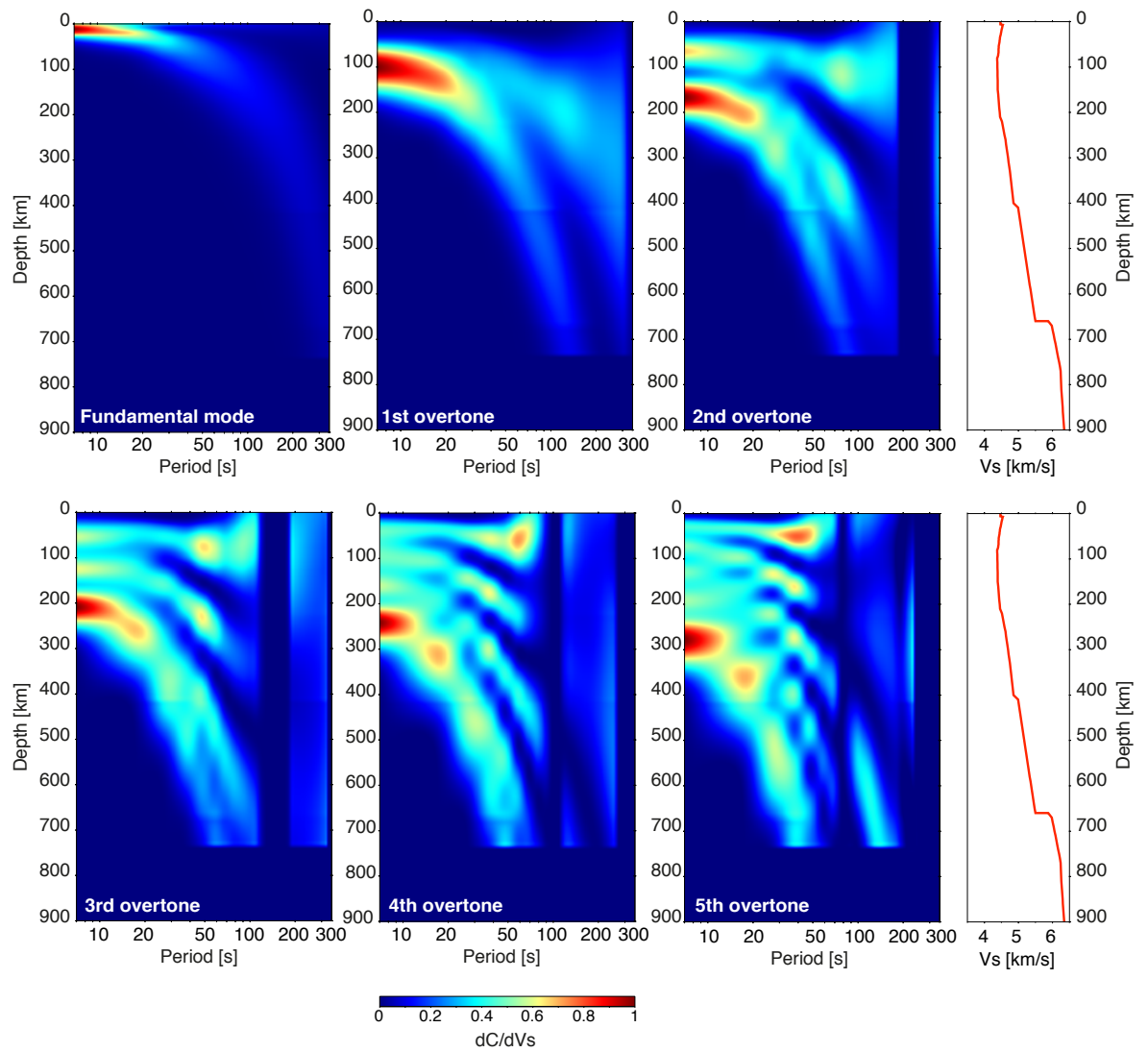


FIGURE 2.3: S-wave sensitivity kernels for model AKM135 (Kennett, Engdahl, and Buland, 1995), computed for the fundamental and first 5 higher modes of Rayleigh-wave phase velocity. The reference V_s model is shown to the right, in red.

In order to extract the information on the seismic properties of the Earth, we need to: compute synthetic seismograms from a background model through forward modelling; compare observed data with the synthetic ones; update the initial Earth model to match the fitting waveforms.

We carry out the forward modelling of the seismogram through the summation of normal modes of the Earth as described in the previous section using the Jeffreys-Wentzel-Kramers-Brillouin approximation (JWKB, Dahlen and Tromp, 1998). The JWKB is the equivalent of ray theory for surface waves, and it indicates that the effects of scattering is neglected. The boundaries of the JWKB approximation relate to the existence of heterogeneities of a size that can affect the travelling waveform. A typical example is the scattering of high frequency components of the seismogram that travel through the strongly heterogeneous crust, where scatterers capable of affecting the modelled wavelengths exist (Nolet, 2008; Lebedev et al., 2005; Schaeffer and Lebedev, 2013). The comparison between the observed and synthetic data is performed in the time domain through waveform fitting. The selection and windowing, filtering of the wave phases for which the JWKB is valid has been automated and optimised in AMI by Lebedev et al. (2005), allowing for the processing of large datasets. Finally, the update to the Earth model is carried out through a non-linear conjugate gradient method.

2.6.1 Waveform fitting

The synthetic seismograms are computed from modal summation, and for a certain station p , the seismogram will be:

$$u_p(\omega) = \sum_n A_{np}(\omega) e^{ik_n(\omega)\Delta_p}. \quad (2.35)$$

The JWKB approximation indicates that the phase measured along the path is solely resulting from the contribution of the initial phase at the source $\phi_0(\omega)$ and the accumulated phase over the source-station path $k_n(\omega)\Delta_p$. Since the complex amplitude term already includes the contribution of the initial phase, we can write the synthetic seismogram as:

$$u_p(\omega) = \sum_n A_{np}(\omega) \exp \left[i \int_0^{\Delta_p} k_n(\omega) d\Delta \right]. \quad (2.36)$$

Where we omitted the dependency on source-station azimuth. Starting with a simple, 1D model, we can write the perturbations in the model in terms of wavenumber:

$$k_n(\omega, \theta, \phi) = k_n(\omega) + \delta k_n(\omega, \theta, \phi). \quad (2.37)$$

If we define the kernel G_n as the partial derivatives of the wavenumber for a generic perturbed model parameter $\delta \mathbf{m}(r)$:

$$\int_0^{\Delta_p} k_n(\omega, \theta, \phi) d\Delta = k_n^0(\omega) \Delta_p + \int_0^a \int_0^{\Delta_p} \mathbf{G}_n(\omega, r) \cdot \delta \mathbf{m}(r) d\Delta dr \quad (2.38)$$

By applying the JWKB approximation, the kernel is invariant in the path and depends only on the radius r :

$$\delta k_n(\omega) \equiv \int_0^{\Delta_p} k_n(\omega, \theta, \phi) d\Delta - k_n^0(\omega) \Delta_p = \int_0^a \mathbf{G}_n(\omega, r) \cdot \left[\int_0^{\Delta_p} \delta \mathbf{m}(r) d\Delta \right] dr. \quad (2.39)$$

This allows us to parametrise the model in a simple set of layers, expressed in a set of (few) basis functions $\mathbf{h}_i(r)$. We can thus express the path-averaged perturbations as a set of M new model parameters γ_i , so that:

$$\frac{1}{\Delta_p} \int_0^{\Delta_p} \delta \mathbf{m}(r) d\Delta = \sum_{i=1}^M \gamma_i \mathbf{h}_i(r), \quad (2.40)$$

and:

$$\delta k_n(\omega) = \sum_{i=1}^M \gamma_i \int_0^a \mathbf{G}_n(\omega, r) \cdot \mathbf{h}_i(r) dr. \quad (2.41)$$

By removing the kernel G_n from the path integral, we reduce the number of parameters we are inverting for, making the inversion process more computationally efficient. In order to more easily separate the contributions of the different parts of the waveform, the misfit between the synthetic data u_p and the observed ones u_{obs} is performed in the time domain. We define the misfit function for the parameters γ_i as:

$$F(\gamma) = \int [R u_{obs}(t) - R u_p(t, \gamma)]^2 dt + \gamma \cdot C_\gamma^{-1} \gamma, \quad (2.42)$$

with R a time-frequency windowing operator that selects and weights the different parts of the seismogram—typically surface-, S- and multiple S waves—that we fit, and C_γ is the covariance matrix of the parametrised model. The design of the operator R is critical and it has been mostly improved for AMI (Lebedev et al., 2005). We will discuss its design in section 2.8. The non-linear inversion minimises $F(\gamma)$ to find an optimal set of model parameters. This is achieved by first computing the local gradients of $F(\gamma)$ numerically around a central point, then the further directions of steepest descent are found recursively and the minima are identified first along the descent direction, and finally with a local quadratic interpolation (Nolet, 2008; Polak and Ribiere, 1969).

This inversion scheme is at the base of both the PWI (Nolet, 1990) and AMI (Lebedev et al., 2005); its most refined and fully automatised version in AMI, which will be discussed in the next sections of this chapter, served as a base for the development of a number of tomographic models (Lebedev and Hilst, 2008; Legendre et al., 2012; Schaeffer and

Lebedev, 2013; Schaeffer and Lebedev, 2014; Lebedev et al., 2018), including the ones presented in this thesis.

2.7 Partitioned Waveform Inversion

The PWI was developed by Nolet (1990) to break down the computational cost of a large-scale tomographic inversion into more manageable sections. The partitioning consists of performing the non-linear inversion for each path, separately, resulting in path-average perturbations in the model parameters γ_i . These can be then easily stored and used in a later stage as data for a linear inversion solving for a tomographic model. In addition to breaking down the computational cost, the partitioning also allows for an easier addition of new data to the inversion: instead of having to re-invert the whole dataset for each tomographic model, only the new paths have to be inverted, and the linear tomographic inversion repeated with the increased dataset.

Although it can seem trivial to combine all equations in a linear system, surface wave dispersion is a property of the average of all layers $\mathbf{h}_i(r)$, which tells us that at least some of the i equations are strongly correlated. Using γ_i directly as data for the tomographic inversion and assuming independent data would thus lead to a very ill-posed problem. We can see from Fig. 2.4 how for two parameters γ_1, γ_2 , large variations in one parameter reflect in similar variations in the second one, with the acceptable solutions describing an ellipse in the $\gamma_1 - \gamma_2$ plane. If we rotate the parameters into a coordinate system aligned along the ellipse axes, we can define a new set of parameters η_1, η_2 for which large variations in η_1 do not cause large variations in η_2 , indicating that their errors are uncorrelated.

We can perform the rotation of the parameters by using the Hessian matrix of F , which consists of the second derivatives with respect to γ for the optimal value γ_{opt} and the transformation matrix $\mathbf{S}$:

$$H_{ij} = \frac{\partial^2 F}{\partial \gamma_i \partial \gamma_j} \quad (2.43)$$

$$\mathbf{H} = \mathbf{S} \mathbf{\Lambda} \mathbf{S}^T$$

where $\mathbf{S}$ diagonalises the Hessian matrix, so that we can define the transformed parameters:

$$\begin{aligned} \gamma &= \mathbf{S} \eta \\ \eta &= \mathbf{S}^T \gamma \end{aligned} \quad (2.44)$$

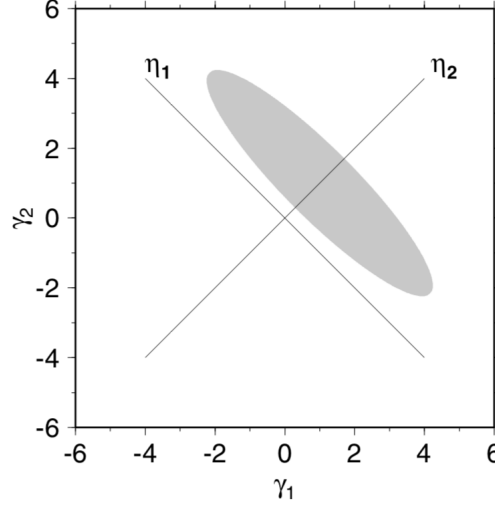


FIGURE 2.4: Example of the misfit function between two generic parameters γ_1, γ_2 . The grey-shaded area represents where a zone of hypothetical acceptable misfit. The lines represent the orientation of the new coordinate system η_1, η_2 . From Nolet (2008).

If we perform the Taylor expansion the misfit function and define a region of acceptable misfit ϵ as deviation from the optimal value $\Delta\gamma = \gamma - \gamma_{opt}$, we can define:

$$F(\gamma) \approx F(\gamma_{opt}) + \frac{1}{2} \Delta\gamma \cdot H \Delta\gamma$$

$$\frac{1}{2} \Delta\gamma \cdot H \Delta\gamma = \frac{1}{2} \Delta\eta \cdot S^T H S \Delta\eta = \frac{1}{2} \Delta\eta \cdot \Lambda \Delta\eta = \frac{1}{2} \sum_{i=1}^M \lambda_i \Delta\eta_i^2 < \epsilon \quad (2.45)$$

The error ϵ is determined by judging the influence of the errors in the data and approximations (ray theory), after which we can estimate an error $\Delta\eta_i$ for the new parameters η_i . The new, independent parameters must thus satisfy $|\Delta\eta_i| < \sqrt{2\epsilon/\lambda_i}$. Once we defined a new set of parameters, we can define a new basis function $g_j(r)$:

$$g_j(r) = \sum_{i=1}^M S_{ij} h_i(r) \quad (2.46)$$

such that:

$$\begin{aligned} \frac{1}{\Delta_p} \int_p \delta m(r) d\Delta &= \sum_{i=1}^M \gamma_i h_i(r) \\ &= \sum_{i=1}^M \sum_{j=1}^M S_{ij} \eta_j h_i(r) \\ &= \sum_{j=1}^M \eta_j g_j(r) \end{aligned} \quad (2.47)$$

The values that the radius can take seem to indicate a very large number of constraints, some of which are not independent. We can reduce these to a set of independent linear

equations by introducing a dual basis function $\tilde{\mathbf{g}}$ that is orthogonal to $\mathbf{g}$ so that:

$$\int_0^a \tilde{\mathbf{g}}(r) \cdot \mathbf{g}(r) dr = \delta_{ij} \quad (2.48)$$

We can thus rewrite equation 2.47 as:

$$\frac{1}{\Delta_p} \int_p \int_0^a \delta \mathbf{m}(r) \cdot \tilde{\mathbf{g}}_k(r) dr d\Delta = \eta_k \pm \Delta \eta_k, \quad (2.49)$$

which describes the final constraints on the Earth structure in term of perturbations with respect to a background models and uncertainty $\Delta \eta_k$. The dependence on a background model however poses issues in the case of using different background models, as it is desirable in, e.g. the highly heterogeneous crust. It is advisable thus to explicitly add the integration on the background model as part of the linear equation, allowing thus for the use of different background models for different constraints:

$$\frac{1}{\Delta_p} \int_p \int_0^a \delta \mathbf{m}(r) \cdot \tilde{\mathbf{g}}_k(r) dr d\Delta = \eta_k \pm \Delta \eta_k + \frac{1}{\Delta_p} \int_p \int_0^a \delta \mathbf{m}_0(r) \cdot \tilde{\mathbf{g}}_k(r) dr d\Delta. \quad (2.50)$$

2.8 Automated Multimode Inversion

AMI is a fully-automated waveform fitting and inversion algorithm. The generation of synthetic seismograms and the waveform fitting is based on the PWI described in the previous section, optimised and automated to detect the portions of the seismogram for which the JWKB approximation is valid and weight their different contribution. In addition to this, AMI includes the possibility of using 3D reference models (compared to the 1D of the previous implementations of the PWI) and the estimation of approximate sensitivity areas (compared to the ray-paths used in the PWI).

2.8.1 Linking waveforms to Earth models

As the PWI, AMI computes synthetic seismograms by means of modal summation. We can re-write equations 2.35, 2.37 in terms of local phase velocities and their perturbation from a background model:

$$\begin{aligned} C_n(\omega, \theta, \phi) &= C_m^{(0,p)}(\omega) + \delta C_m^p(\omega, \theta, \phi) \\ u_p(\omega) &= \sum_n A_{np}(\omega) e^{i\omega \Delta_p / C_n(\omega, \theta, \phi)} \\ &= \sum_n A_{np}(\omega) \exp \left[i\omega \Delta_p / (C_m^{(0,p)}(\omega) + \delta C_m^p(\omega, \theta, \phi)) \right]. \end{aligned} \quad (2.51)$$

As in the PWI, we can reduce the number of parameters by describing path-average phase velocity perturbations instead of local ones, which are in turn linked to perturbations in S- and P- wave velocity from a background model:

$$\begin{aligned}\overline{C_m^p(\omega)} &= \frac{1}{\Delta_p} \int_0^{\Delta_p} C_n(\omega, \theta, \phi) d\Delta - C_n^{(0,p)}(\omega) \\ &= \int_0^a \frac{\partial C_n^{(0,p)}(\omega)}{\partial \beta(r)} \overline{\delta \beta_p(r)} dr + \int_0^a \frac{\partial C_n^{(0,p)}(\omega)}{\partial \alpha(r)} \overline{\delta \alpha_p(r)} dr\end{aligned}\quad (2.52)$$

with the perturbations in P- and S- wave velocities $\overline{\delta \alpha_p(r)}$ and $\overline{\delta \beta_p(r)}$ from the background model. Phase velocities and their perturbations relate therefore the information contained in the waveforms to the Earth model.

2.8.2 3D reference models

One of the differences between AMI and the PWI resides in the use of reference models. Strong heterogeneities in the Earth (especially in the crust) make the choice of a single, 1D reference model a crude approximation of reality; by adding an explicit integration over the background model (equation 2.50), the PWI already allows for the choice of different models for different source-station paths, which improves substantially over the use of a single model. AMI goes further beyond the path-average approximation by computing the phase velocities derivatives with respect to P- and S- wave velocity (Frechét derivatives) on a 3D reference model, and then average them over approximate sensitivity areas (Lebedev and Hilst, 2008; Schaeffer and Lebedev, 2013). Including a 3D reference, the definition of the background phase velocities becomes thus:

$$C_m^0(\omega) = \int_{\theta} \int_{\phi} \overline{K(\theta, \phi)} C_m(\omega, \theta, \phi) d\theta d\phi, \quad (2.53)$$

with $\overline{K(\theta, \phi)}$ the sensitivity kernel describing the source-station sensitivity area, averaged over the filtered waveform (we will discuss them in detail in the next section). This allows for the easy implementation of a 3D background model. We can thus precompute the reference phase velocities and their Frechét derivatives for each node of the 3D reference model, for each mode and frequency. Precomputing the derivatives grants computational efficiency in the integration over the sensitivity areas, following the partitioning approach of the inversion method.

Since the strongest heterogeneities are in the crust, our 3D model is built as a combination of CRUST2.0 (Bassin, Laske, and G., 2000) with added topography and bathymetry, and a modified version of AK135 (Kennett, Engdahl, and Buland, 1995) for the mantle (Fig. 2.5). The model is parametrised on a triangular grid (Wang and Dahlen, 1995a) with approximate 28 km spacing (Schaeffer and Lebedev, 2013). The inclusion of an accu-

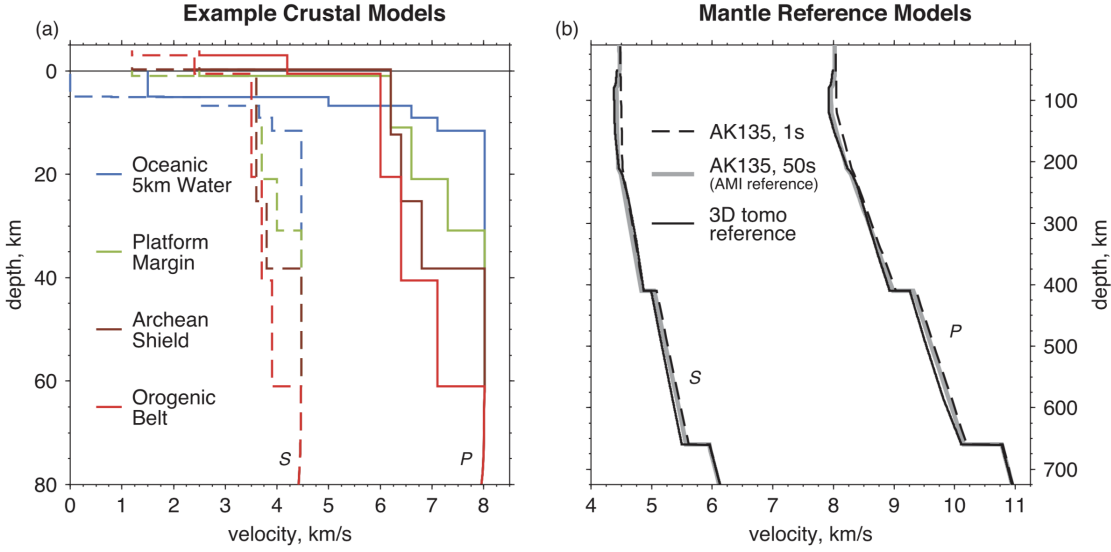


FIGURE 2.5: 3D P- and S-wave background models used in AMI: a) four example crustal models; b) mantle reference model used by AMI (grey) along other ones, including our own average used in the linear tomographic inversion. From Schaeffer and Lebedev (2013).

rate reference model avoids errors from the strong crustal heterogeneities in the model, which would map both in the crust and underlying mantle. We can then invert for the deviations from the crustal model itself, and although the resolution of long-period surface waves is not sufficient to accurately model the fine details of the upper crust, we can resolve the largest heterogeneities and prevent them from polluting the mantle structure.

2.8.3 Sensitivity kernels

AMI further improves on the ray-path approximation by substituting ray paths with approximate sensitivity areas. These areas are defined in terms of Fresnel zones, and include all nodes of the 3D reference model grid for which the phase shift between the source and the point and the source and receiver is less than $\pi/2$. Wang and Dahlen (1995b) defined the half-width of the " $\pi/2$ " perimeter as:

$$\delta = \sqrt{\frac{\lambda/2}{|\cot\Delta_s + \cot\Delta_r|}} \quad (2.54)$$

with Δ_s and Δ_r the distances of a point to the source and receiver, respectively. Only one kernel per seismogram is used, and λ is the average wavelength of the fundamental mode of the band-passed waveform fit. The kernels have a path-normal trapezoidal shape, with their amplitude decreasing linearly between $\delta/2$ and δ (Fig. 2.6). Since the integral over the path-normal is constant for each path, nodes at the extremes of the path (source and receiver) will have larger kernel amplitudes (different circle sizes in Fig. 2.6).

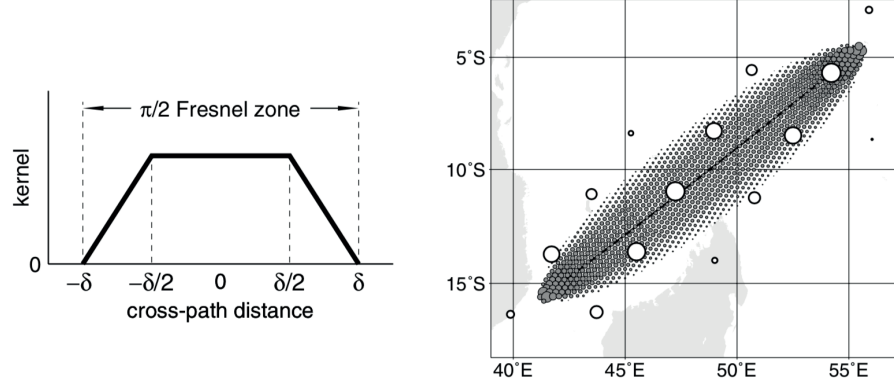


FIGURE 2.6: Sensitivity kernels $\overline{K(\theta, \phi)}$ used in AMI. Left: path-perpendicular cross-section through a kernel illustrating the trapezoidal shape. Right: example of a sensitivity kernel from Lebedev and Hilst (2008); grey circles show the nodes of the integration grid, white of the 3D tomographic model grid; both circle sizes are scaled with the weights of the knots in the area integral. From Lebedev and Hilst (2008).

2.8.4 Depth parametrisation and non-linear inversion

The inversion scheme in AMI follows the one described in the previous chapter for the PWI. In AMI, we can rewrite equations 2.40, 2.47 in terms of P- and S-wave velocities; in the case of S-waves:

$$\begin{aligned}
 \overline{\delta\beta^{(p)}(r)} &= \frac{1}{\Delta_p} \int_0^{\Delta_p} [\beta(\theta, \phi, \omega) - \beta^{(0,p)}(r)] d\Delta \\
 &= \sum_{i=1}^M \gamma_i h_i(r) \\
 &= \sum_{i=1}^M \eta_i g_i(r),
 \end{aligned} \tag{2.55}$$

With γ_i the original parameters and η_i the transformed parameters with uncorrelated uncertainties. In AMI, the basis functions $h_i(r)$ represent depth layers, are triangular and have apexes at 18 nodes for S-waves (7, 20, 36, 56, 80, 110, 150, 200, 260, 330, 410-, 410+, 485, 585, 660-, 660+, 809, 1008 km depth) and 10 nodes for P-waves (7, 20, 36, 60, 90, 150, 240, 350, 484, 585).

The equations with uncorrelated uncertainties for each waveform are then expanded and weighted on the horizontal sensitivity areas $\overline{K(\theta, \phi)}$, forming the sensitivity kernels $G_i^{\alpha, \beta}$. We can rewrite the independent linear equations 2.47 as:

$$\int_{\oplus} G_i^{\beta} \delta\beta(r) + G_i^{\alpha} \delta\alpha(r) d^3r = \eta_i \pm \Delta\eta_i, \tag{2.56}$$

with $i = 1, \dots, M$.

2.8.5 Near-nodal radiation check and time-frequency windowing

One of the strengths of AMI consists in its ability to select and weight time-frequency windows in which the JWKB approximation is valid in a fully automated way. At first, the earthquake radiation pattern for each source-station path, frequency and mode is computed and the ones with azimuth falling into a nodal band are discarded. Such paths are dominated by scattered energy and have increasing complexity towards the nodal bands (Lebedev et al., 2005), resulting in violations of the JWKB approximation. During the procedure, only frequency bands with at least half of the maximum amplitude are kept.

In order to break down the contributions of different parts of the waveform, AMI selects a set time-frequency windows defined as boxcars in time and Gaussian bandpass filters in frequency. This has the benefit of substantially simplifying the waveforms that are fitted and isolate the contributions of different wave phases to the inversion. Of the frequencies passing the near-nodal radiation criterion, only the ones with fundamental mode wavelengths such that $\lambda < \Delta/3$ are kept. A high frequency limit is set in relationship to the earthquake duration (Lebedev et al., 2005). The first and last time windows are set to be right before the S- and multiple-S wave arrival time and after the last fundamental mode arrival in all frequency bands.

The first Gaussian filter is computed at the lowest frequency in the allowed range and its signal-to-noise ratio (SNR) checked against a time window before the first P arrival. If the SNR is insufficient, the frequency is increased until a suitable SNR is found. For a suitable frequency band, waveform fitting is performed at first in the largest time-window available. If the misfit is less than 5%, the fit is successful. If this is true only in parts of the filtered seismogram, then the time window is either narrowed to include only the successful wave trains, or split in more windows, with each window containing at least one complete wave train. If a successful time-frequency window is found, the next frequency filter is added. For fundamental mode windows, fits are accepted only if they fit in 3 successive frequency filters, to avoid interference of the mode with its own coda. Finally, all successfully fitting time-frequency windows are inverted simultaneously to find a suitable $\overline{\delta\beta(r)}$.

2.8.6 Energy equalisation and phase weighting

In section 2.6.1, we defined the misfit function F in equation 2.42 as including a filtering and weighting operator R . In AMI, the equation for each time-frequency window i becomes:

$$F_i = \frac{\int_{t_1}^{t_2} w_i(t) (d_i(t) - f_i s_i(t))^2 dt}{\int_{t_1}^{t_2} w_i(t) d_i^2(t) dt}, \quad (2.57)$$

with d_i and s_i the observed and synthetic seismograms, respectively, t_1 and t_2 the beginning and end time of the i -th time window and $w_i(t)$ a weighting function. Amplitudes of the synthetics are dependent on the scalar moment of the event, that can have large uncertainties. AMI thus relies on the equalisation factor f_i to get rid of the influence of the amplitudes during waveform fitting, which is defined as:

$$f_i = \sqrt{\frac{\int_{t_1}^{t_2} d_i^2(t) dt}{\int_{t_1}^{t_2} s_i^2(t) dt}}. \quad (2.58)$$

As noted in the previous section, all N time-frequency windows are inverted together. Each window is characterised by a weight $w_i(t)$ that includes the weighting of wave train envelopes, wave types and data errors (see Lebedev et al., 2005 for more details) and multiplied by a frequency ratio accounting for phase shifts due to the perturbation of phase velocities. We can thus define the total misfit function as:

$$F_{tot}(\overline{\delta\beta(r)}) = \frac{\sum_{i=1}^N (\omega_{ref}/\omega_i)^2 \int_{t_1}^{t_2} w_i(t) (d_i(t) - f_i s_i(t))^2 dt}{\sum_{i=1}^N (\omega_{ref}/\omega_i)^2 \int_{t_1}^{t_2} w_i(t) d_i^2(t) dt}. \quad (2.59)$$

In Fig. 2.7, we can see an example AMI fit.

2.9 3D linear tomographic inversion

The linear equations with uncorrelated uncertainties provided by AMI (equation 2.56) provide robust constraints on the path-average Earth structure, and its computational efficiency allows to produce millions of waveform fits. We can then combine these fits in a large linear tomographic inversion solves for P- and S- wave velocities, moving thus from the path-average models to a full 3D model of the Earth seismic structure. We can then formulate the linear system as:

$$Am = d. \quad (2.60)$$

2.9.1 The data vector

In order to combine the equations, we scale the data η_i and the kernels G_i by their uncertainties $\Delta\eta$ for each path. This results in scaled data q_i^p and kernels J_i (we omit the kernel

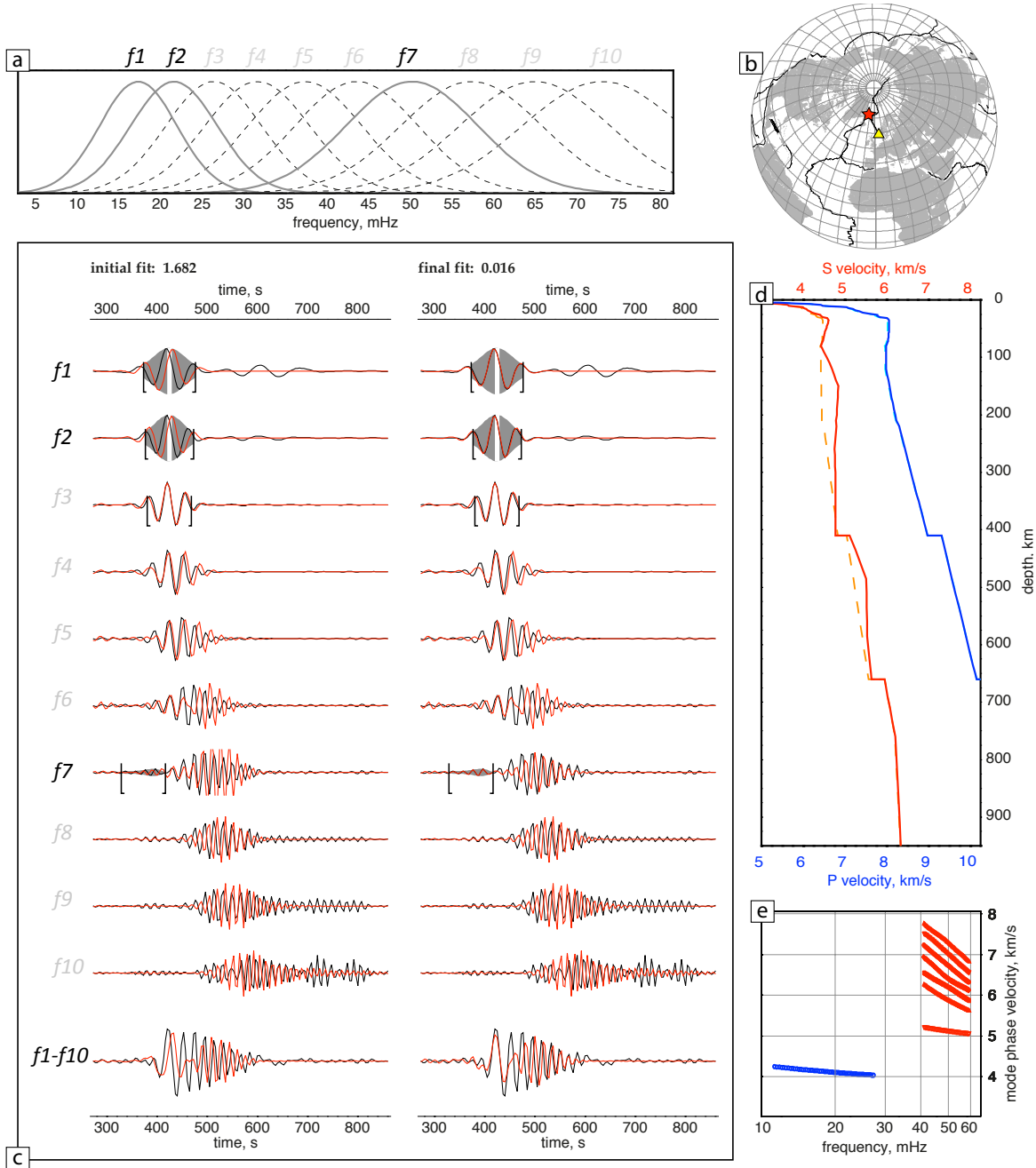


FIGURE 2.7: Example of a successful AMI fit for station Z6.NWG29 (southern Norway). a) Used gaussian frequency filters: the successfully fitting ones are shown as solid lines. b) Source (red star) station (yellow triangle) path map. c) Synthetic (red) and observed (black) seismograms filtered at the frequency bands shown in a) for both the initial fit (left) and after the non-linear inversion (right). Bottom seismograms includes all filtered frequency bands. Brackets indicate attempted waveform fit. Fitting time-frequency windows are highlighted in grey. d) P- (blue) and S-wave (red) velocity path-average 1D model resulting from the inversion; reference model is shown as dashed lines, computed model as solid. e) Phase velocity curves constrained by the fit: blue indicates the fundamental mode and red the overtones.

path superscript for ease of notation), for each path p and equation i :

$$\begin{aligned} q_i^p &= \frac{\eta_i^p}{\Delta\eta_i^p} \\ J_i^\beta &= \frac{G_i^\beta}{\Delta\eta_i^p} \\ J_i^\alpha &= \frac{G_i^\alpha}{\Delta\eta_i^p} \end{aligned} \quad (2.61)$$

Since the equations are parametrised over 18 and 10 vertical basis functions for S- and P-wave velocity, respectively, we obtain a large number of equation. In order to reduce this, only eigenvalues of the scaled parameters q_i^p above a certain eigenvalue threshold are needed by the tomographic inversion, resulting in an average of 3.5 equations retained for each path. The inclusion of S- and multiple-S waves determines a slower depth decay of the eigenvalues, meaning that waveform fits including additional phases will have a larger number of equations.

We can thus gather all the $p = 1, \dots, N$ source station paths containing the reduced, weighted equations $i = 1, \dots, M^*$ in a single data vector $\mathbf{d}$:

$$\mathbf{d} = \begin{bmatrix} d_1 \\ \vdots \\ d_k \end{bmatrix} = \begin{bmatrix} q_1^1 \\ \vdots \\ q_{M^*}^1 \\ \vdots \\ q_{M^*}^N \end{bmatrix}, \quad (2.62)$$

with k the number of data rows, N the number of seismograms and M^* the reduced number of equations per path.

2.9.2 The model vector

The model vector $\mathbf{m}$ is a global, 3D model. Vertically, the model is parametrised using the same set of basis functions used in AMI (18 for S- and 10 for P-wave velocities). Horizontally, we parametrise the model on a triangular grid constructed in the same way as the integration grid used in AMI (Wang and Dahlen, 1995a). The horizontal model grid is co-registered with the integration grid, although usually with much larger grid spacing. As the grids are co-registered, the equations can be easily interpolated from the fine integration grid onto the model grid.

The tomographic model is inverted for P- and vertically polarised S-wave velocities as well as 2 Ψ S-wave azimuthal anisotropy. The P-wave velocity term is included due to the Rayleigh wave sensitivity to it. The sensitivity of Rayleigh waves to compressional

waves, is however small, and Lebedev and Hilst (2008) found that they cannot be resolved independently and must be damped instead to the S-wave velocity structure, to which sensitivity is greater. The azimuthal anisotropy introduced in the linear system takes into account the azimuthal dependency of the phase velocities. If we assume the medium to be weakly anisotropic, Smith and Dahlen (1973) found that phase velocity can be described as an isotropic term with small π and $\pi/2$ perturbations:

$$\begin{aligned} C(\omega, \Psi, \theta, \phi) = & A_0(\omega, \theta, \phi) \\ & + A_1(\omega, \theta, \phi) \cos 2\Psi + A_2(\omega, \theta, \phi) \sin 2\Psi \\ & + A_3(\omega, \theta, \phi) \cos 4\Psi + A_4(\omega, \theta, \phi) \sin 4\Psi, \end{aligned} \quad (2.63)$$

with Ψ the azimuth of the path and $A_{0..4}$ the 3D S-wave velocity distributions. Rayleigh waves are mostly sensitive to 2Ψ and Love waves to 4Ψ anisotropic terms (Montagner and Tanimoto, 1991), so that in the models presented in this thesis, only the former ones are parametrised.

The model vector consist thus of four model parameters V_P, V_S, A_1, A_2 , the horizontal knots n_H and $n_V^P = 10$ P-wave basis functions and $n_V^S = 18$ S-wave basis functions. The model term $\mathbf{m}$ thus can be written as:

$$\mathbf{m} = \begin{bmatrix} V_S(i, j) \\ A_1(i, j) \\ A_2(i, j) \\ V_P(i, j) \end{bmatrix} = \begin{bmatrix} V_S(i, j) \\ \vdots \\ V_S(n_V^S, n_H) \\ A_1(i, j) \\ \vdots \\ A_1(n_V^S, n_H) \\ A_2(i, j) \\ \vdots \\ A_2(n_V^S, n_H) \\ V_P(i, j) \\ \vdots \\ V_P(n_V^P, n_H) \end{bmatrix} \quad (2.64)$$

with $j = 1, n_H$ and $i = 1, n_V$ for the appropriate set of basis functions (P or S). The model elements are ordered thus first by parameter, then by shell knot and finally over depth.

2.9.3 The sensitivity matrix and path weighting

The matrix A relate the data and the model vectors and represent thus, the sensitivity of the data to the model parameters. For k the number of data rows and l the total number of parameters, the sensitivity matrix will have a very large size $k \times l$. It consists of the scaled kernel functions $J_i^{\alpha, \beta}$. The scaled kernels are a combination of the transformed

basis functions g_i^p and the sensitivity areas $\overline{K^p(\theta, \phi)}$. Each row corresponding to a single equation and each sensitivity kernel $\overline{K^p(\theta, \phi)}$ covering only a small portion of the nodes on the globe, the sensitivity matrix turns out to be very sparse, with most of the columns for each row being zero. We exploit the sparsity of the matrix by storing only the non-zero elements to reduce the computational size of the problem.

Due to the large number of waveform fits, many equations will share a very similar source-station path. If these were included in the tomographic inversion as-is, they would dominate the sampled model nodes over, e.g. crossing single paths, biasing the inversion towards their information. To prevent that, a path-similarity weight is included for each seismogram. The weight is a sum of individual weights of seismogram similarity with all other seismograms. The individual weights are cosine tapers with amplitudes related to the distance between pairs of seismograms. After constructing the sensitivity matrix, we compute the path-similarity weights and scale the data with them, defining the new, scaled data vector d^s where equations with high path-similarity are down-weighted.

2.9.4 Inversion

The system resulting from the model and data vectors described in the previous sections is overparametrised. In order to stabilise the inversion, we need to regularise the system of equations. The regularisation is carried out by slight norm damping and lateral and vertical smoothing. These are incorporated into the linear system in the form of various operators in the sensitivity matrix A :

$$\begin{bmatrix} A \\ aI \\ b\Phi \\ c\Theta \\ d\Pi \end{bmatrix} m = \begin{bmatrix} d^s \\ 0 \\ 0 \\ 0 \\ 0 \end{bmatrix}. \quad (2.65)$$

The identity matrix I and the associated coefficient a define the norm damping. The Φ term minimises the difference $\delta V_P - \delta V_S$, damping their ratio to the reference model; being Rayleigh waves less sensitive to V_P , this effectively damps it to the better constrained V_S . The term Θ computes the lateral and vertical derivatives of the model m , defining the gradient damping. The lateral and vertical derivatives can be weighted separately, allowing different lateral and vertical gradient damping. The term Π penalises the difference between the model parameter on a single knot and the average of all its neighbours and defines thus the lateral smoothing.

The model is inverted by trying to minimise the weighted sum of data misfit, model norm and solution gradient, following the approach of Zielhuis and Nolet (1994):

$$\min \left[(A\mathbf{m} - \mathbf{d})^T (A\mathbf{m} - \mathbf{d}) \right], \quad (2.66)$$

with $A, \mathbf{d}^s$ being the hybrid, regularised system described in equation 2.65. Due to the large size of the problem, we replace the direct evaluation of equation 2.65 with the more efficient least-squares method developed by Paige and Saunders (1982) (LSQR).

Chapter 3

The upper mantle beneath the South Atlantic Ocean, South America and Africa from waveform tomography with massive datasets

3.1 summary

We present a tomographic model of the crust, upper mantle and transition zone beneath the South Atlantic, South America and Africa. Taking advantage of the recent growth in broadband data sampling, we compute the model using waveform fits of over 1.2 million vertical-component seismograms, obtained with the Automated Multimode Inversion of surface, S and multiple S waves. Each waveform provides a set of linear equations constraining perturbations with respect to a three-dimensional (3D) reference model within an approximate sensitivity volume. We then combine all equations into a large linear system and solve it for a 3D model of S- and P-wave speeds and azimuthal anisotropy within the crust, upper mantle and uppermost lower mantle. In South America and Africa, our new model SA2019 reveals detailed structure of the lithosphere, with structure of the cratons within the continents much more complex than seen previously. In South America, lower seismic velocities underneath the Transbrasilian Lineament (TBL) separate the high-velocity anomalies beneath the Amazon Craton from those beneath the São Francisco and Paraná Cratons. We image the buried portions of the Amazon Craton, the thick cratonic lithosphere of the Paraná and Parnaíba Basins, and an apparently cratonic block wedged between western Guyana and the slab to the west of it, unexposed at the surface. Thick cratonic lithosphere is absent under the Archean crust of the São Luis, Luis Alves and Rio de La Plata Cratons, next to the continental margin. The Guyana Highlands are underlain by low velocities, indicating hot asthenosphere. In the transition zone, we map the subduction of the Nazca Plate and the Chile Rise under Patagonia. Cratonic lithosphere beneath Africa is more fragmented than seen previously, with separate cratonic units observed within the West African and Congo Cratons, and with cratonic lithosphere

absent beneath large portions of Archean crust. We image the lateral extent of the Niassa Craton, hypothesized previously, and identify a new unit, the Cubango Craton, near the southeast boundary of the greater Congo Craton, with both of these smaller cratons unexposed at the surface. In the South Atlantic, the model reveals the patterns of interaction between the Mid-Atlantic Ridge (MAR) and the nearby hotspots. Low-velocity anomalies beneath major hotspots extend substantially deeper than those beneath the MAR. The Vema Hotspot, in particular, displays a pronounced low-velocity anomaly under the thick, high-velocity lithosphere of the Cape Basin. A strong low velocity anomaly also underlies the Cameroon Volcanic Line and its offshore extension, between Africa and the MAR. Subtracting the global, age-dependent V_s averages from those in the South Atlantic Basins, we observe areas where the cooling lithosphere is locally hotter than average, corresponding to the location of the Tristan da Cunha, Vema and Trindade hotspots. Beneath the anomalously deep Argentine Basin, we image unusually thick, high-velocity lithosphere, which suggests that its anomalously great depth can be explained, at least to a large extent, by isostatic, negative lithospheric buoyancy.

3.2 Introduction

The South Atlantic Ocean and the adjacent continents of South America and Africa share a fascinating geological history, but also a relatively poor sampling with seismic data. The South Atlantic Ocean formed from the breakup of the supercontinent Gondwana into Africa and South America during the Cretaceous (Torsvik et al., 2009; Seton et al., 2012; Gaina et al., 2013). The cores of both continents are formed by large Archean cratons (Cahen et al., 1984; Cordani and Sato, 1999) surrounded by Neoproterozoic mobile belts (Vail, 1985; Li et al., 2008; Begg et al., 2009; De Brito Neves and Fuck, 2013), and did not undergo major continental reworking until the continental breakup 130 Ma, initiating the formation of the South Atlantic Ocean (Torsvik et al., 2009; Moulin, Aslanian, and Unternehr, 2010; Seton et al., 2012). During the Cenozoic, the start of the Andean Subduction affected western South America, and voluminous intraplate magmatism, attributed to mantle plume(s) occurred in the Afar region, followed by rifting along the branches of the Afar triple junction (Wilson and Guiraud, 1992; Montagner et al., 2007). The emplacement of a number of continental Large Igneous Provinces (LIPs) over the past 250 m.y.—commonly attributed to the impingement of mantle plumes (Morgan, 1971; Wilson and Guiraud, 1992)—and extensive intraplate volcanism, in the form of continental flood basalts and widespread seamount chains, indicates the presence of very active mantle processes affecting the lithosphere beneath the area. However, our understanding of the crustal and lithospheric architecture of the region and how it has been affected by the underlying mantle processes is limited by the availability of both geophysical and geological data, hindered in the area by the presence of thick sedimentary covers, dense forests and large water masses.

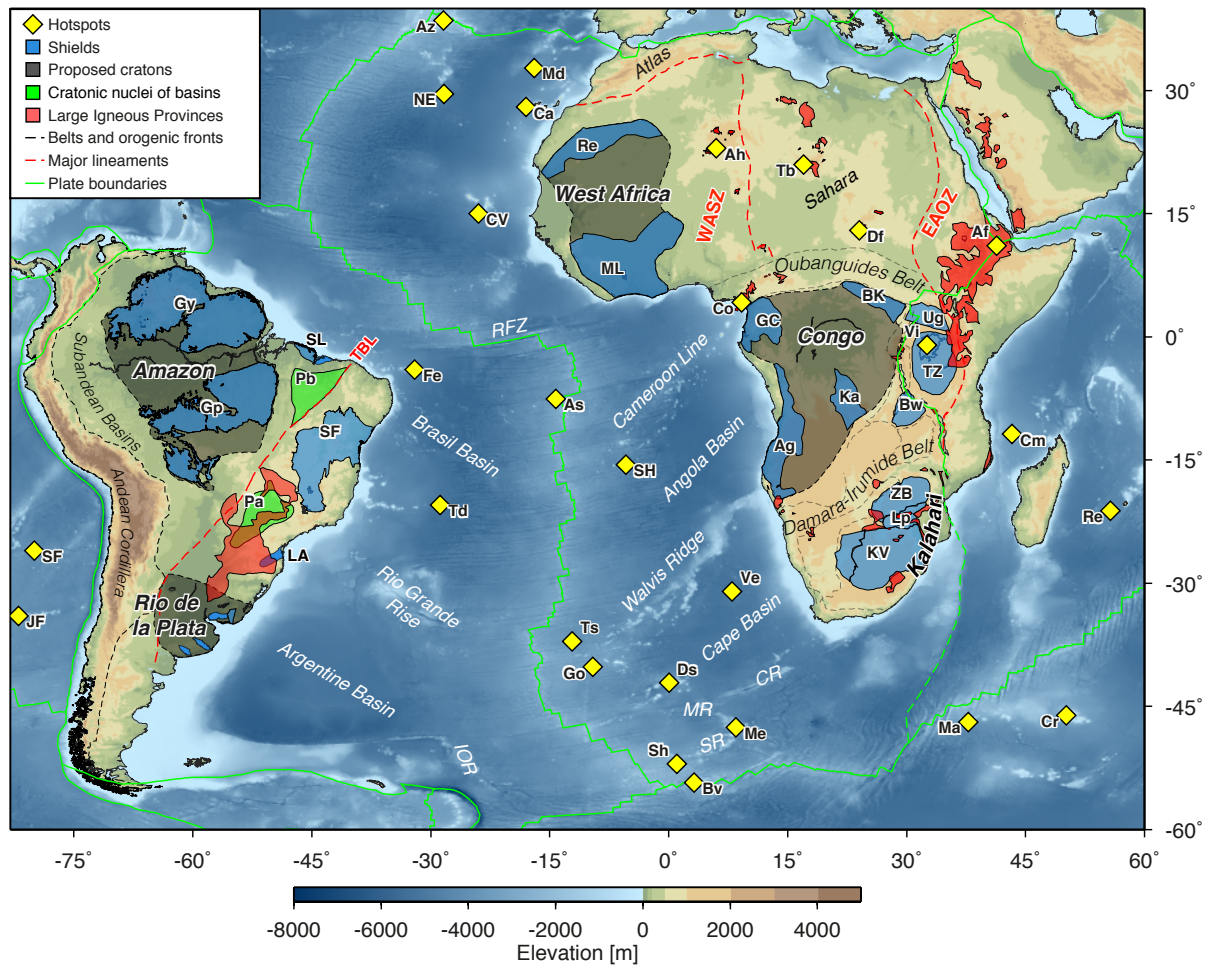


FIGURE 3.1: Tectonic map of the study area. Previously proposed extent of the Amazon, Rio de la Plata, West Africa and Congo cratons are shown as black transparencies, while the Kalahari Craton is shown as composed by the Zimbabwe, Kaapvaal and Limpopo units (Feng, Lee, and Assumpção, 2007; Begg et al., 2009; Sanchez Bettucci, Peel, and Oyhantcabal, 2010; Oyhantcabal, Siegesmund, and Wemmer, 2011; Brazilian Geological Survey et al., 2016). Known shields are shown in blue with black lettering: TZ, Tanzania; ZB, Zimbabwe; KV, Kaapvaal; SF, São Francisco; SL, São Luis; LA, Luis Alves; Re, Reguibat; ML, Man-Lèo; GC, Gabon-Cameroon; Ag, Angola; BK, Bomu-Kibali; Ug, Uganda; Bw, Bangweulu Block; Ka, Kasai; Lp, Limpopo Block; Gy, Guyana; Gp, Guaporé. Proposed extent of thick lithosphere under sedimentary basins is shown in green: Pa, Paraná; Pb, Paranaíba. Mobile belts are plotted in black dashed lines. Volcanic continental terranes (Modified from Gaina et al. (2013)) are plotted in red. Hotspots are shown as yellow diamonds: Az, Azores; Md, Madeira; Ca, Canary; NE, New England; CV, Cabo Verde; As, Ascension; SH, Saint Helena; Fe, Fernando de Noronha; Td, Trindade; Ve, Vema; Ts, Tristan da Cunha; Go, Gough; Ds, Discovery; Me, Meteor; Sh, Shona; Bv, Bouvet; Ma, Marion; CCr, Crozet; Cm, Comoro; Re, Reunion; Ah, Ahaggar; Tb, Tibesti; Df, Darfur; Af, Afar; Vi, Victoria; SF, São Felix; JF, Juan Fernandez. Oceanic features: RFZ, Romanche Fracture Zone; MR, Meteor Rise; IOR, Islas Orcadas Rise; CR, Cape Rise; SR, Shona Rise. Other main features: WASZ, West African Shear Zone; EAOZ, East African Orogenic Zone, TBL, Transbrasilian Lineament. Plate boundaries are shown in green (solid lines: verified; dotted lines: proposed).

Globally, Archean cratonic crust is normally underlain by thick, cold—and thus seismically fast—lithospheric mantle (Schaeffer and Lebedev, 2015). In the study area, previous seismic tomography yields images of broad regions of thick, high velocity lithosphere at the root of the West African, Congo, Kalahari Cratons in Africa (Fouch et al., 2004; Pasyanos and Nyblade, 2007; Begg et al., 2009; Fishwick, 2010) and Amazon and São Francisco Cratons in South America (Feng, Assumpção, and Van der Lee, 2004; Feng, Lee, and Assumpção, 2007; Chulick, Detweiler, and Mooney, 2013; Neto, Julià, and Schimmel, 2019). Surface geology data show Archean shields of smaller lateral extent (Fig. 3.1) (Wilson and Guiraud, 1992; Begg et al., 2009; Kachingwe, Nyblade, and Julià, 2015), but large portions of the cratonic Archean basement are unexposed, buried under the sedimentary cover. In South America, there is less agreement between geological and geophysical data on the less sampled western portion of the cratonic province, leaving open questions on the extent of the cratonic roots of Amazonia (Cordani and Sato, 1999; Feng, Assumpção, and Van der Lee, 2004; Feng, Lee, and Assumpção, 2007; Rizzotto and Hartmann, 2012; Chulick, Detweiler, and Mooney, 2013). Previous seismic studies also reported the presence of thick, high-velocity lithosphere underneath the Paraná and Parnaíba Basins, but the shape of these proposed cratonic blocks, unexposed at the surface, remains debated (Snok and James, 1997; Heintz, Debayle, and Vauchez, 2005; Feng, Lee, and Assumpção, 2007; Lebedev, Boonen, and Trampert, 2009; Chulick, Detweiler, and Mooney, 2013). Little agreement is also found about the extent of the Rio de la Plata and Luis Alves Cratons, exposed only in few, scattered locations (Sanchez Bettucci, Peel, and Oyhantcabal, 2010; Oyhantcabal, Siegesmund, and Wemmer, 2011; Santos et al., 2019) and notably absent in tomographic models (e.g. Feng, Lee, and Assumpção (2007), Schaeffer and Lebedev (2013), and Debayle, Dubuffet, and Durand (2016)). In northern Africa, low-velocity anomalies have been detected beneath the Saharan and East African magmatic provinces (Montagner et al., 2007; Sicilia et al., 2008; Reusch et al., 2010; Fishwick, 2010; Hansen and Nyblade, 2013; Debayle, Dubuffet, and Durand, 2016), but their shape and depth extent are still debated (Ebinger and Sleep, 1998; Reusch et al., 2010). In the South Atlantic Ocean, past intraplate magmatism can be identified by hotspot tracks on the oceanic lithosphere (Torsvik et al., 2009; Gaina et al., 2013). Topography and maps of residual basement depths reveal deviations from the bathymetry predicted by normal seafloor spreading models (Müller et al., 2008) in the location of hotspots and hotspot tracks (Figs. 3.1, 3.2), mainly due to the anomalously thick crust. The most prominent examples are the Walvis Ridge and Rio Grande Rise, the hotspot tracks related to the Paraná-Etendeka magmatism and the onset of the Tristan da Cunha mantle plume (Morgan, 1971; O'Connor and Duncan, 1990; O'Connor and Jokat, 2015; Gassmöller et al., 2016; Baba et al., 2017; O'Connor et al., 2018; Schlömer et al., 2017). The anomalous concentration of hotspots and basement depth anomalies in the south-eastern Atlantic Ocean (Sleep, 1990; Nyblade and Robinson, 1994; O'Connor et al., 2018) has been linked to the Large Low-Shear Velocity Province (LLSVP) at the Core-Mantle Boundary beneath (Lay,

2005; Torsvik et al., 2006). There is little agreement, however, regarding the number and origin of the plumes feeding the South Atlantic intraplate volcanism (Colli et al., 2014; French and Romanowicz, 2015; Hosseini, 2016; O'Connor et al., 2018; Colli et al., 2018).

Seismic tomography is a suitable tool for the investigation of the structure of the lithosphere and underlying mantle, as seismic velocities can constrain the composition and temperature of the rock in the Earth's interior. In our study area, however, coverage with seismic data is both relatively low and uneven, due to the few seismic stations available and the few, localised sources of strong seismicity. In both continents, where coverage is relatively denser, a number of regional- and continental-scale tomographic models have been computed: South America (Vandecar, James, and Assumpção, 1995; Vdovin et al., 1999; Van Der Lee, James, and Silver, 2001; Feng, Assumpção, and Van der Lee, 2004; Heintz, Debayle, and Vauchez, 2005; Feng, Lee, and Assumpção, 2007; Rocha, Schimmel, and Assumpção, 2011; Neto, Julià, and Schimmel, 2019) and Africa (Montagner et al., 2007; Pasyanos and Nyblade, 2007; Sicilia et al., 2008; Begg et al., 2009; Fishwick, 2010; Reusch et al., 2010; Guidarelli et al., 2011; Hansen, Nyblade, and Benoit, 2012; Hansen and Nyblade, 2013). Fewer regional models include or specifically focus on the South Atlantic Ocean (Vdovin et al., 1999; Heintz, Debayle, and Vauchez, 2005; Colli, Fichtner, and Bunge, 2013; James, Dalton, and Gaherty, 2014). Many global upper-mantle and whole-mantle models produced in recent years, however, show increasing resolution that makes them more and more comparable to regional or continental scale models (Grand, Hilst, and Widyanoro, 1997; Shapiro and Ritzwoller, 2002; Lebedev and Hilst, 2008; Ritsema et al., 2011; Lekić and Romanowicz, 2011; French, Lekic, and Romanowicz, 2013; Debayle and Ricard, 2012; Schaeffer and Lebedev, 2013; Chang et al., 2015; French and Romanowicz, 2014; Debayle, Dubuffet, and Durand, 2016).

In this paper, we present our new, S-wave velocity tomographic model SA2019, constructed using a very large global waveform dataset, including a significant amount of newly available data from the region. The improved data sampling and the effective extraction of structural information from the large waveform dataset by our inversions produce significantly improved resolution compared to previous models covering the study area, which allows us to address problems of regional tectonics and geodynamics across South America, the South Atlantic and Africa. In the following sections, we describe the construction of the model, assess it through resolution tests and comparisons with other state-of-the-art tomographic models, and discuss geodynamic inferences from our tomography.

3.3 Data and methods

Our new model SA2019 is computed as a global, azimuthally anisotropic shear-wave velocity model of the crust, upper mantle and transition zone. The model is specifically

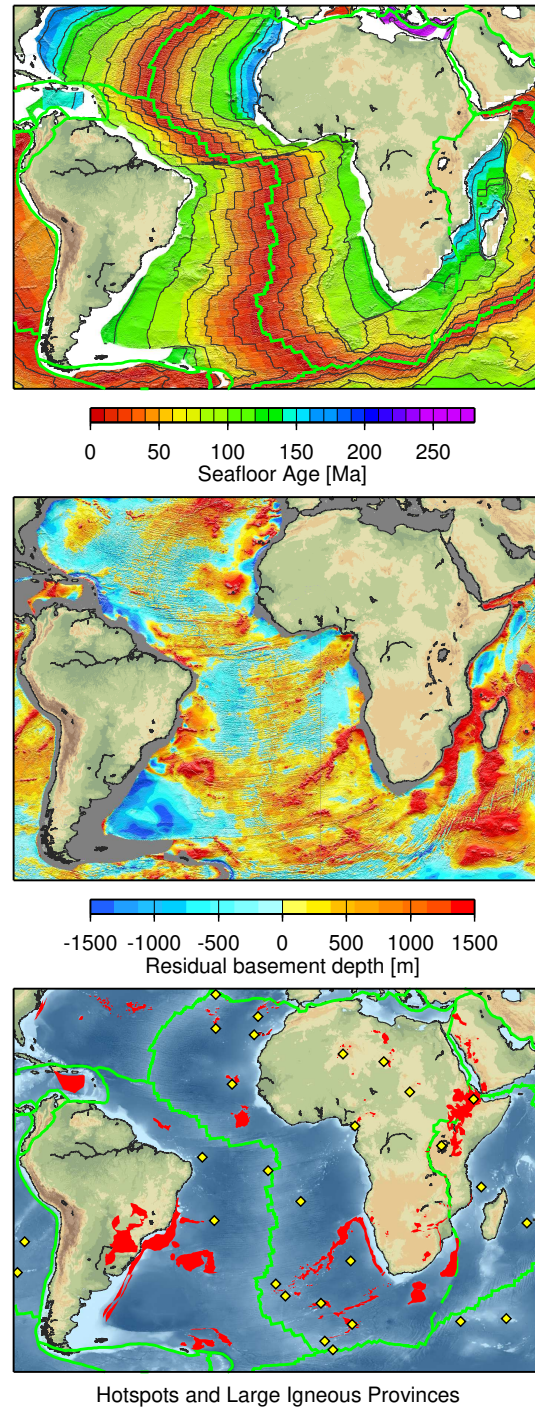


FIGURE 3.2: Top: age of the oceanic lithosphere derived from magnetic and other geophysical data, from Müller et al. (2008). Centre: Residual ocean basement depth map relative to the plate model GDH-1 (Stein and Stein, 1992), from Müller et al. (2008). Bottom: topographic map with Oceanic and continental LIP's contours (modified from Gaina et al. (2013)) overlaid in red and hotspot locations as yellow diamonds.

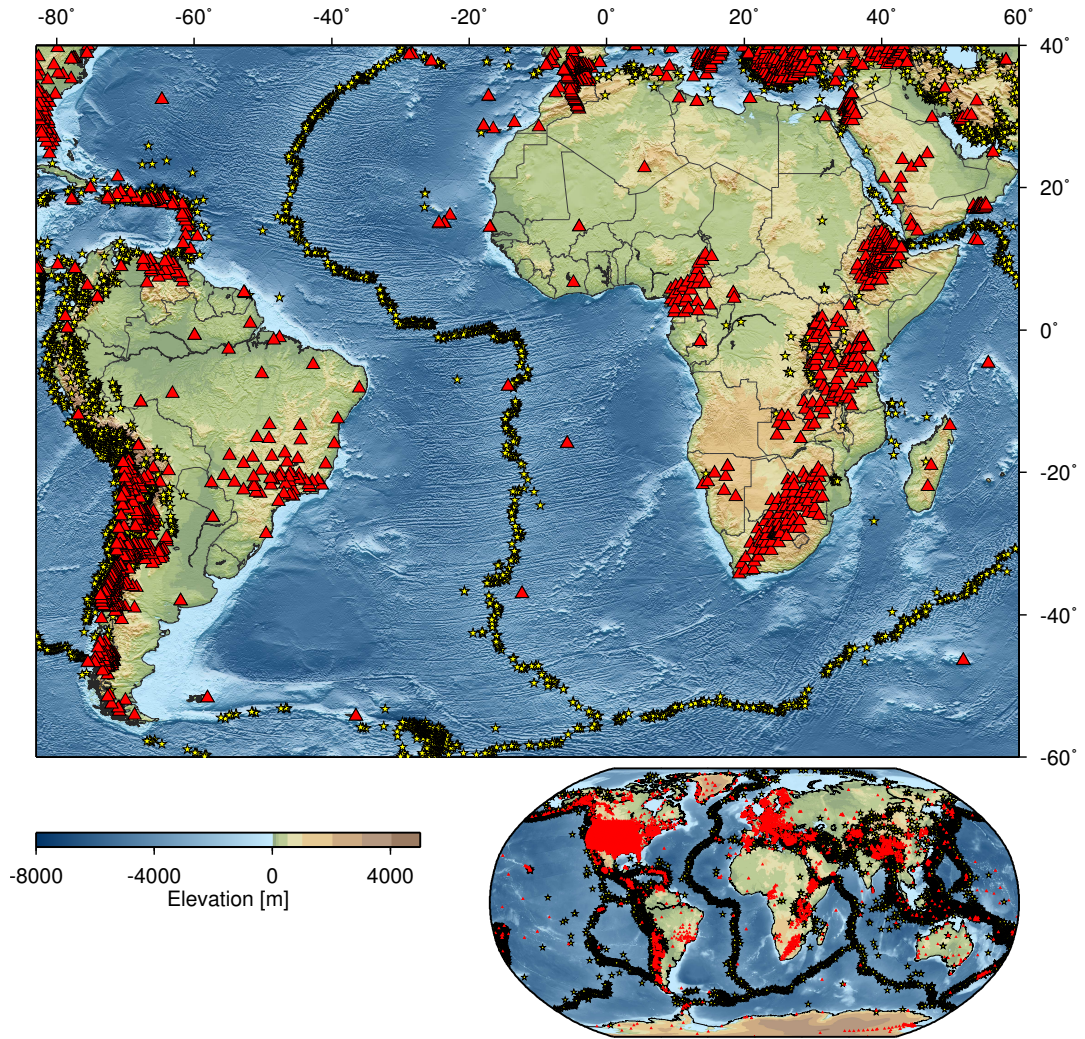


FIGURE 3.3: Earthquakes and broadband seismic stations within the study area and around the world. The station/event coverage is global and has been maximised in the Atlantic area by assembling all freely available seismic data. Events are plotted as yellow stars, seismic stations as red triangles.

focussed on the South Atlantic Ocean and neighbouring continents by means of maximising the data sampling of the region and by tuning the inversion parameters to optimise the resolution across the region. In this section we present the dataset constraining the model and briefly discuss the modelling procedures used to construct it (for a complete description of the methods, we refer to Lebedev et al. (2005), Lebedev and Hilst (2008), and Schaeffer and Lebedev (2013)).

3.3.1 Data

SA2019 is constrained by over 1.2 million vertical-component seismograms, waveform-fitted using the Automated Multimode Inversion (AMI) of surface and S-wave forms (Lebedev et al., 2005). The fits are derived from broadband, teleseismic earthquake

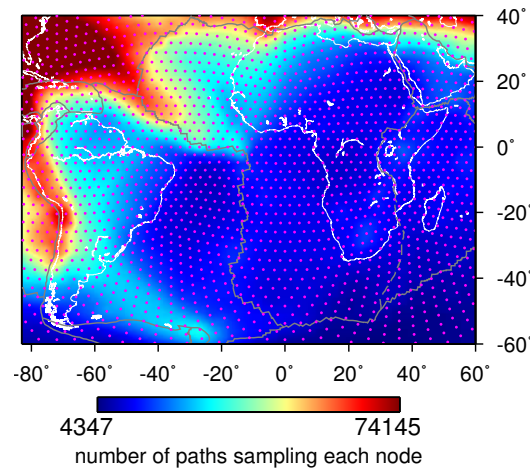


FIGURE 3.4: Map of the number of paths sampling each model grid node. Model grid nodes are plotted as magenta circles, coastlines and plate boundaries in white and grey (solid lines: verified; dotted lines: proposed). The plot highlights the good coverage of our model, showing that each model grid node is sampled at least by 4626 paths, which is the global minimum.

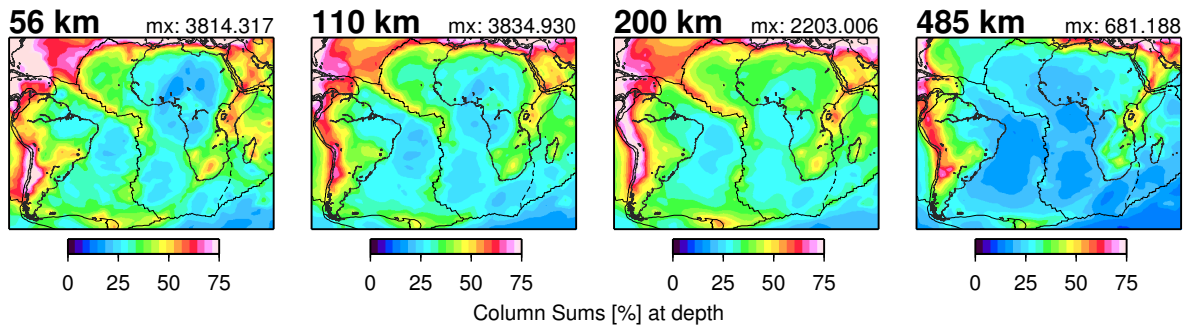


FIGURE 3.5: Plot of the sums of the columns of the sensitivity matrix at four selected grid node depths (56, 110, 200, 485) that represent most of the investigated depth range. Dark blue regions indicates zero sampling (nowhere on the globe) while white regions represent 75% of the maximum value in the study area. Column sums are an adimensional representation of the model relative sampling at each node (section 3.4.1).

recordings retrieved on a global scale, but with a regional focus, with all freely available recordings within the region being used. In total, 6360 stations and 27550 events worldwide are used (Fig. 3.3). We selected all earthquakes from the Harvard Centroid Moment Tensor Solutions catalog (Dziewonski, Chou, and Woodhouse, 1981; Ekström, Nettles, and Dziewonski, 2012) since 1994, and since 1990 for stations in particularly undersampled areas. Recordings with source-receiver distances between 500 and 18000 km were selected with a distance-varying magnitude threshold (Schaeffer and Lebedev, 2013).

3.3.2 Coverage

The data coverage of the region is uneven and sparser than to the north of it, in particular, in North America and Europe. Complementing the regional data with global data yields a highly redundant sampling of all investigated depths throughout the region, with every

node of the global model grid sampled by at least 4626 paths (Fig. 3.4). To quantify the coverage, we use the sums of the columns of the sensitivity matrix. The column sums reflect, for each grid node, the quantity of the data sampling each model parameter, the sensitivity of each waveform to structure, and the “path similarity weights” applied in order to even the data coverage by reducing the effect of bundles of similar rays (Lebedev and Hilst, 2008)(Fig. 3.5).

3.3.3 Inversion procedure

We use a three-step inversion procedure to compute the tomographic model. At first, we feed the pre-processed dataset (quality control, response correction) to AMI, which operates fully automated inversion of large waveform datasets. AMI computes synthetic seismograms through modal summation of the fundamental and first 20 higher spheroidal modes and performs waveform fitting of direct S-, multiple S-, and Rayleigh-wave phases to the real data within a set of time-frequency windows, with elaborate phase weighting. The result is a set of linear equations with uncorrelated uncertainties (Nolet, 1990) describing 1D average P- and S-wave velocity perturbations within the sensitivity volumes between the source and receiver with respect to a 3D reference model (Lebedev et al., 2005; Lebedev and Hilst, 2008). By simultaneously inverting waveforms of the S-, multiple S- and Rayleigh-wave phases, AMI extracts structural information across a wide range of depths, thanks to the different depth sensitivity of surface waves (the depth sampling range of which is determined by the period of the fundamental mode) and S- and multiple S-waves (which can be seen as a result of the interference of the higher modes and dive deep into the upper mantle and the lower mantle). The resulting waveform fits cover a very broad 10-450 s period range, with the bulk of the fits sampling periods of 20-250 s (Lebedev and Hilst, 2008; Schaeffer and Lebedev, 2013).

In the second step, the resulting equations are inverted together as a large linear system for the 3D distribution of P- and S-wave velocity deviations from the reference model and S-wave azimuthal anisotropy, via LSQR (Paige and Saunders, 1982). The model is parametrised over a triangular grid (Wang and Dahlen, 1995a) with an average 327-km inter-knot spacing and a depth parametrisation over 18 and 10 depth nodes for S- and P-wave velocities, respectively (S-wave velocities: 7, 20, 36, 56, 80, 110, 150, 200, 260, 330, 410-, 410+, 485, 585, 660-, 660+, 809 and 1007 km; P-wave velocities: 7, 20, 36, 60, 90, 150, 240, 350, 485 and 585 km). S-, multiple S- and Rayleigh wave phases are only weakly sensitive to P-wave velocity, and we strongly couple P wavespeeds to S-wave velocity during the inversion process. We invert for S-wave 2Ψ azimuthal anisotropy in order to prevent anisotropic heterogeneities from mapping into isotropic ones, but we leave the interpretation of anisotropy for future work. We use the same 3D reference model for both AMI and the 3D linear inversion. In the crust, we use the 3D model CRUST2 (Bassin, Laske, and G., 2000) smoothed across cell boundaries, with added topography

and bathymetry and with a depth-dependent correction to the seismic velocities, according to the 1D global average of our own tomography (Lebedev and Hilst, 2008). In the mantle, the reference is close to our own 1D global average. In the third step of our inversion procedure, we exploit the dataset redundancy and select only the most mutually consistent data, through an a posteriori outlier rejection procedure. From a first-iteration model, we compute the synthetic data by multiplication of the sensitivity matrix and the model vector. We then compare the synthetic and real data and discard the ones with the largest misfits, associated mostly with largest earthquake mislocations, station timing errors and diffraction effects unaccounted for in our modelling (Lebedev and Hilst, 2008; Legendre et al., 2012). For the smaller errors in the remaining data, previous tests have shown that even systematic moderate errors in the event locations do not substantially alter the tomographic image of a region covered with criss-crossing source-station paths (Lebedev, Nolet, and Hilst, 1997). The effects of diffraction on the recovery of anomalies by our waveform inversions have been tested on synthetic data computed for heterogeneous Earth models with fully numerical methods and shown to be small (Lebedev and Hilst, 2008; Qin et al., 2008). In this study, we kept 65 percent of the most mutually consistent data and used seismograms from ~ 770 thousand source-station paths to constrain our final model.

3.3.4 Resolution tests and the choice of regularisation parameters

Our linear 3D inversion is regularised by means of lateral smoothing, vertical gradient damping and a slight norm damping (Lebedev and Hilst, 2008). To account for the changes in data sampling with depth, all regularisation parameters are vertically scaled by multiplying coefficients.

We tested our regularisation parameters by constructing a synthetic model $\mathbf{m}_s$ with 200 m/s spikes in S-wave velocity anomalies at each depth node at selected locations in the model (Fig. 3.6). In order to tune our model regularisation, we performed a set of different inversions, varying both absolute values and depth-scaling coefficients of the regularisation parameters. We examined dV_s deviations from the input model in 1D horizontal and vertical profiles centred on the spike location, focussing on shape recovery. We used tests with widely spaced spikes in order to prevent interference of the output point-spread patterns from different input spikes (Rawlinson and Spakman, 2016). The absolute amplitudes of spike anomalies are strongly reduced in the output, due to the averaging of tomographic images (Lebedev and Nolet, 2003).

If not accurately scaled with depth, model regularisation can bias the retrieved anomalies. For example, shallower model nodes may absorb most of the signal from anomalies located at greater depths. To prevent such biases, we decrease the norm damping and increase the smoothing coefficients with depth. Fig. 3.6 shows the recovery of two example spikes with the optimal regularisation parameters and scaling factors: spike 1 is positioned at 110 km depth in a well sampled area, whereas spike 2 is located at 330 km

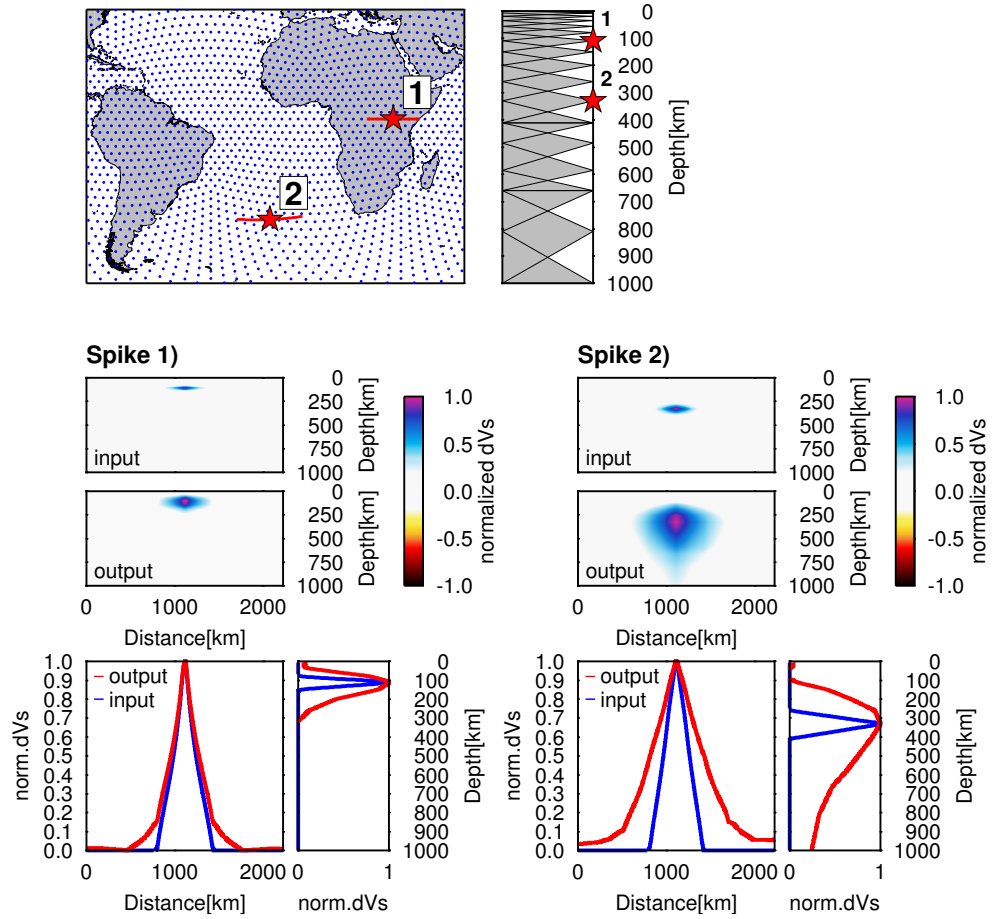


FIGURE 3.6: Synthetic spike tests for SA2019. The location of the spikes is shown on map as red stars. Nodes of the model grid are plotted in blue. Depth location of the spikes is shown on top of the basis functions in the top right panel. For each spike we plot a vertical cross-section through the input (top) and output (bottom) models. Location of the cross-section is shown on the map in red. The bottom plots show horizontal and vertical 1D S-wave anomaly profiles along the cross-section. Input profiles are shown in blue, output in red. All S-wave velocity anomalies are normalised to the maximum.

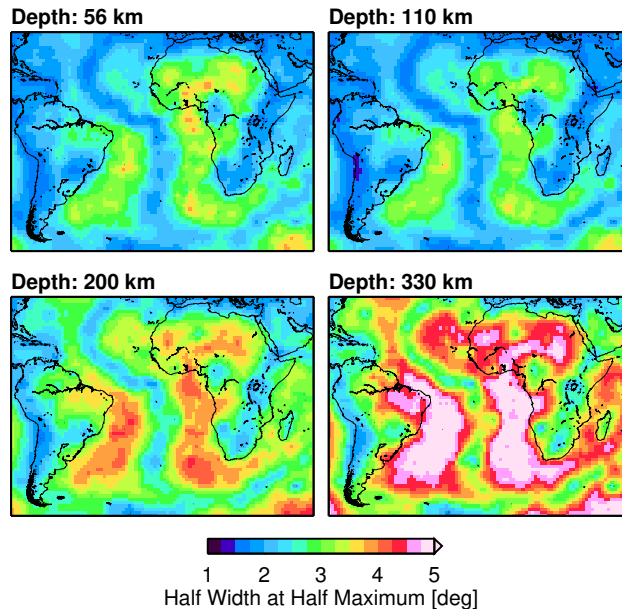


FIGURE 3.7: Lateral resolution map at four depths (56, 110, 200, 330 km) through the model. Lateral resolution is defined as the average of the Half Width at Half Maximum contours at each node, computed from synthetic spike tests at single nodes of the model grid.

in a less sampled area of the model. The output synthetic model shows good recovery of the shape of the input anomaly for spike 1 along both vertical and horizontal axes. The deeper spike 2 shows broader spreading of the velocity anomaly, as expected, but with the maximum of the anomaly at the correct depth. Small secondary maxima appear at shallow depth; their amplitude is probably negligible in the inversion of real data, which provides a very dense sampling of the top 200 km of the mantle.

We also used a series of synthetic spike tests with shifting spike patterns to estimate the lateral resolution of our model across the entire region. The averaging (resolution) length was computed as the Half-Width at Half-Maximum (HWHM) for spikes positioned at the nodes of the model grid. The model recovered for each of the spikes yielded an estimate of one line of the resolution matrix. In order to achieve a detailed HWHM map while retaining enough spacing between the spikes to correctly measure the signal tails, we spaced the 200 m/s spikes at a 5-node distance laterally, and repeated the process for a set of 20 spike-pattern models that, together, covered 67% of the model grid nodes. The process was repeated for the depths of 56, 110, 200, 330 km.

The variations of thus estimated lateral averaging length across the region at different depths are plotted in Fig. 3.7. The maps show HWHM values of 1.5-3 degrees in the lithosphere, with the largest values in West Africa and in the South Atlantic Basins, in agreement with the lowest relative coverage values highlighted by the column sums (Fig. 3.5). At 330 km depth, larger HWHM values of 4-5 degrees can be observed in the regions of poorer coverage; for very few of the worst recovered spikes (highest HWHM) at this depth, HWHM contours do not close due to the averaging lengths being sufficiently long

to cause interference between different spike lobes.

3.4 The model SA2019

In this section we discuss the general features of the model and compare it with other tomographic models sampling the area. A detailed, feature-by-feature discussion on specific areas of the model is given in the following sections.

3.4.1 Model description

We display our model as the isotropic S-wave velocity anomalies with respect to the 3D reference velocity model (Figs. 3.8, 3.9).

We identify the lithosphere of cratons as the fastest areas of the model, with S-wave velocity anomalies of 4.5% and greater at lithospheric depths (80-260 km depths, magenta colours on the maps). The depth extent of these high-velocity anomalies suggests that the cratonic lithosphere in both the African and South American continents reaches deeper than 200 km depth (e.g. West African, Congo, Zimbabwe and Amazon Cratons). The velocity and depth extent of these anomalies are in agreement with observations from the North American Craton, e.g. Grand and Helmberger (1984), Romanowicz (2009), and Schaeffer and Lebedev (2014), but their distribution appears to be more fragmented.

Areas of volcanism are underlain by low-velocity bodies of varying amplitudes, typically between -100 (e.g. Sahara volcanic provinces) and -400 m/s (Central Andes, Colombia, Afar). In the South Atlantic Ocean, SA2019 retrieves continuous, low-velocity anomalies beneath the Mid-Atlantic Ridge (MAR) with V_s as low as 100-150 m/s below global average at 56-110 km depths in most locations, and lower in locations where the ridge and hotspots are closest.

3.4.2 Comparison with other models

We compared our model with 5 other S-wave velocity models encompassing most or all of the study area: 3D2016_09Sv (Debayle, Dubuffet, and Durand, 2016), an updated version of the model SaCo2013 (Colli, Fichtner, and Bunge, 2013), SEMum2 (French, Lekic, and Romanowicz, 2013), S40RTS (Ritsema et al., 2011) and CUB (Shapiro and Ritzwoller, 2002).

The model 3D2016_09Sv is an upper-mantle S-wave velocity model constrained by multimode Rayleigh waves (Debayle, Dubuffet, and Durand, 2016). SaCo2013 is a regional-scale upper-mantle model for isotropic S-wave velocity, computed using the adjoint method (Fichtner, Bunge, and Igel, 2006); it focusses on the South Atlantic oceanic lithosphere and samples parts of Africa and South America. SEMum2 is a global, radially anisotropic mantle model derived from long period waveforms and group velocity maps. S40RTS is a

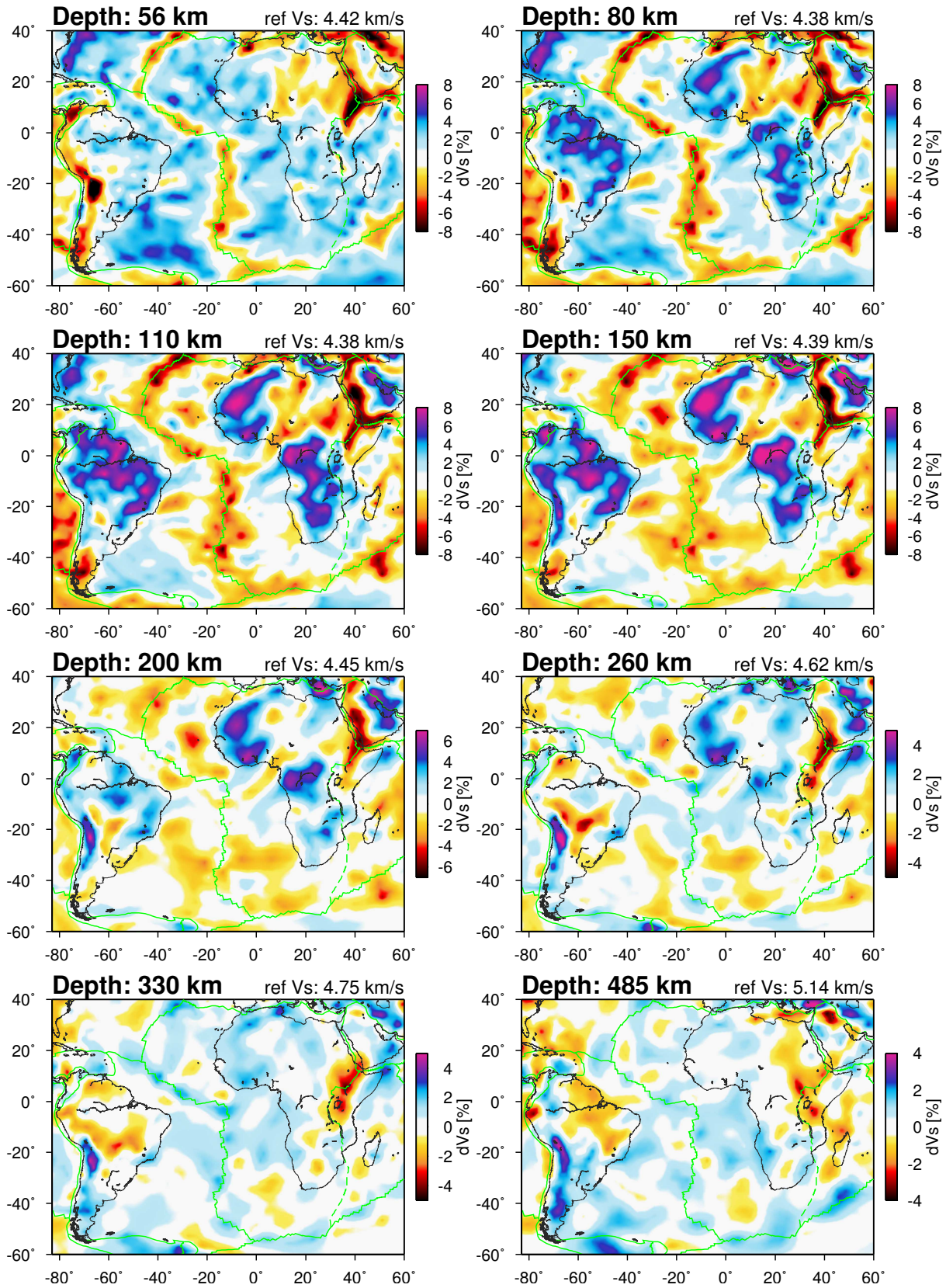


FIGURE 3.8: Map views of SA2019 S-wave velocity variations at eight depths (56, 80, 110, 150, 200, 260, 330, 485). Velocity perturbations are plotted in % with respect to the reference at depth (marked on the top right of each map). Major plate boundaries are plotted as green lines (solid lines: verified; dotted lines: proposed).

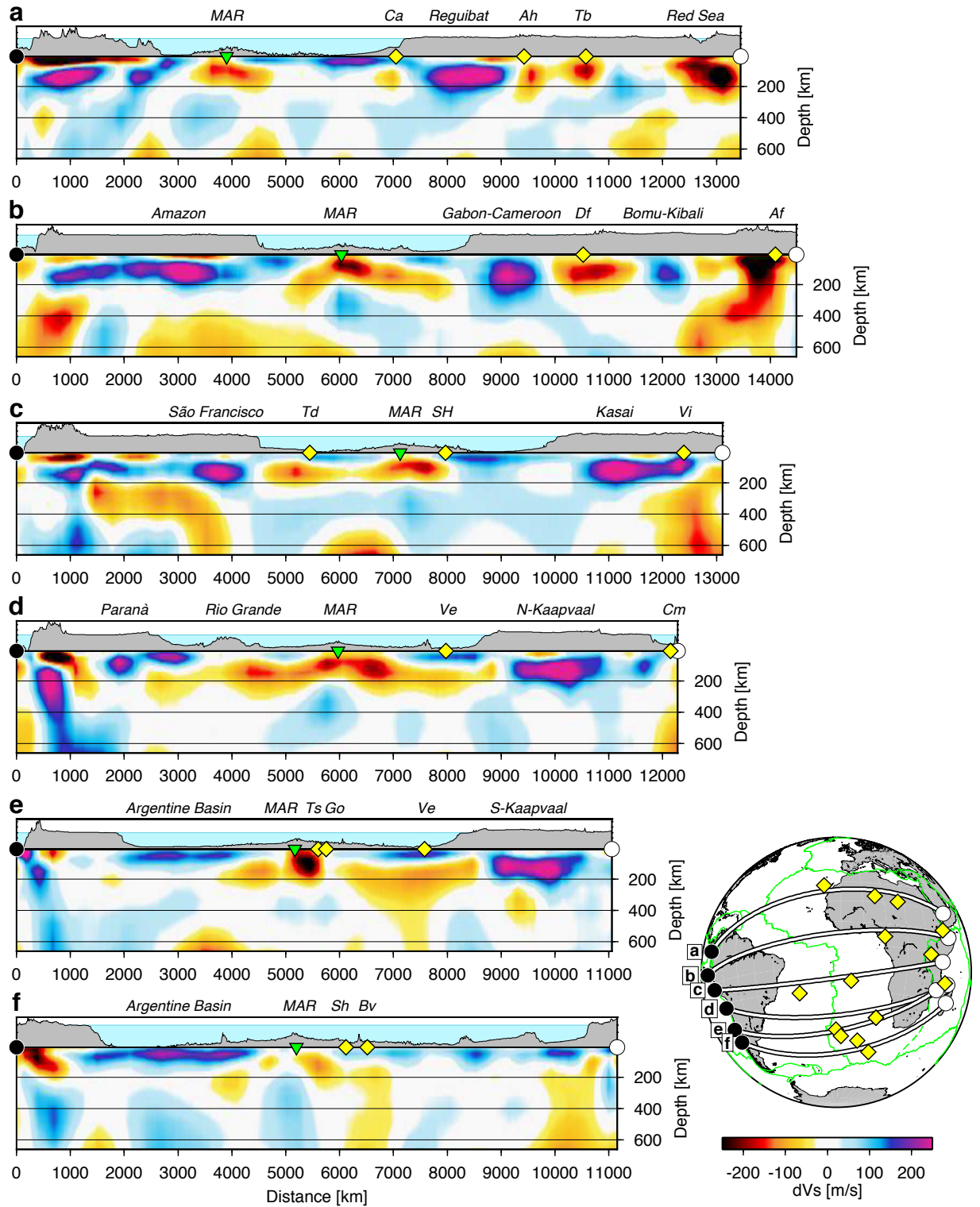


FIGURE 3.9: W-E cross-sections across the study area. Location of the sections (white), plate boundaries (green) and crossed hotspots (yellow diamonds) are shown at the bottom right. On the vertical sections, hotspots within 500 km from the section are shown as yellow diamonds, location of the Mid-Atlantic Ridge axis as green reverse triangles.

Depth: 100 km

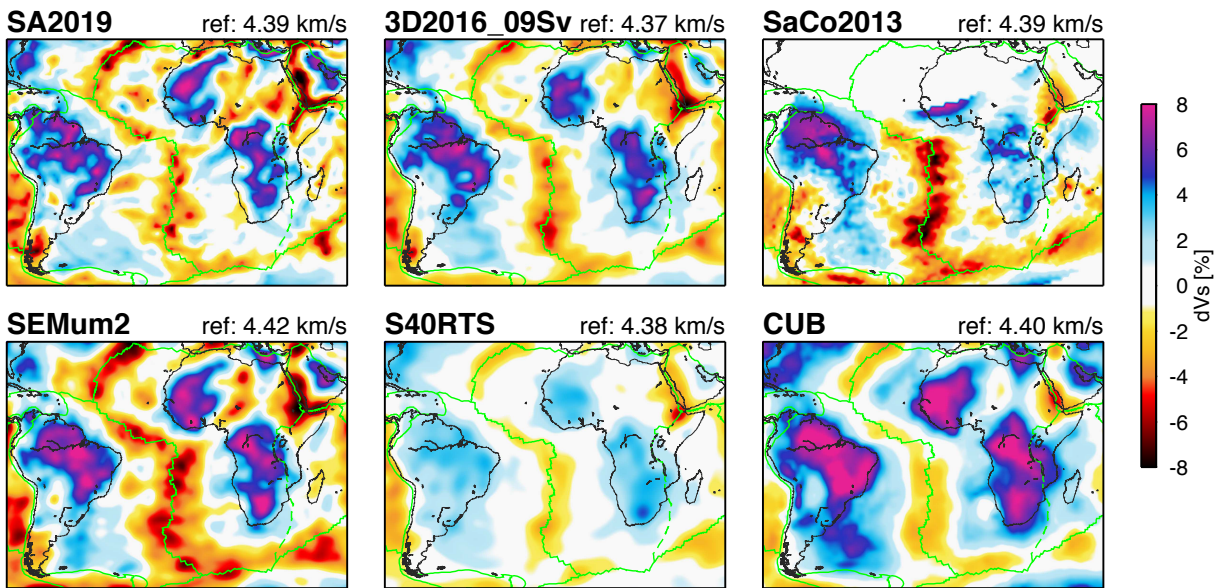


FIGURE 3.10: Comparison of SA2019 with five other tomographic models at 100 km depth: 3D2016_09Sv (Debayle, Dubuffet, and Durand, 2016), SaCo2013 (Colli, Fichtner, and Bunge, 2013), SEMum2 (French, Lekic, and Romanowicz, 2013), S40RTS (Ritsema et al., 2011) and CUB (Shapiro and Ritzwoller, 2002). All models are plotted as S-wave velocity anomalies with respect to the mean velocity at depth, shown on the top right corner of each model plot. Major plate boundaries are plotted as green lines (solid lines: verified; dotted lines: proposed).

Depth: 300 km

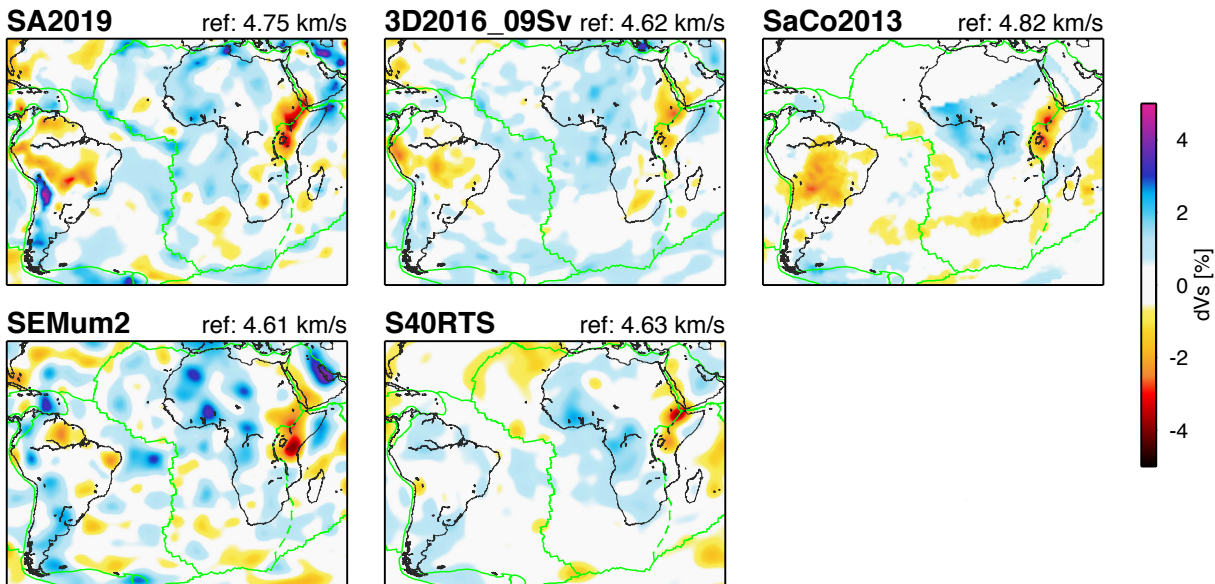


FIGURE 3.11: Comparison of SA2019 with four other tomographic models covering the area at 300 km depth. The model CUB is interpretable to 250 km depth only (Shapiro and Ritzwoller, 2002) and is therefore not shown at this depth. All models are plotted as S-wave velocity anomalies with respect to the mean velocity at depth, shown on the top right corner of each model plot. Major plate boundaries are plotted as green lines (solid lines: verified; dotted lines: proposed).

whole mantle global model constrained by both Rayleigh and body waves and spheroidal mode-splitting functions. Finally, CUB is a crustal and upper mantle radially anisotropic model computed using a Monte Carlo inversion of Rayleigh and Love dispersion measurements (Shapiro and Ritzwoller, 2002).

In order to compare the models, we removed the mean V_s at each depth and plotted the anomalies as $dV_s\%$ from that value (Figs. 3.10, 3.11). The models are consistent at larger scales. At smaller scales, SA2019 offers an improvement in resolution for many regional structures. At lithospheric depths (100 km), SA2019 shows improved definition of the craton and ridge boundaries compared to other global- and continental scale models (Fig. 3.10). Compared to 3D2016_09Sv, SA2019 shows steeper V_s gradients at the boundaries of cratons and low-velocity zones, while also recovering more structure and higher amplitudes for the mid-ocean ridges. In the deep upper mantle (300 km depth and greater), our model retrieves stronger, more focussed images of the seismic-velocity anomalies known from regional studies of the East African Rift System (Pasyanos and Nyblade, 2007; Bastow et al., 2008; Hansen and Nyblade, 2013; Civiero et al., 2016) and the Andean subduction zone (Fukao and Obayashi, 2013; Scire et al., 2014; Scire et al., 2017) compared to the other models (Fig. 3.11).

3.4.3 Structural resolution test

In order to further verify that the new, fine-scale structures we observe are not artifacts of uneven coverage, we computed a synthetic model simulating the large scale features (cratons and ridges) imaged by previous studies (Fig. 3.12). By inverting such synthetic model with the same regularisation parameters as in the inversion of real data, we can observe if the fine-scale structures we recover are real or, instead, a product of a distortion of the larger anomalies present in the Earth.

We simulated the Amazon, Rio de la Plata, West Africa, Congo and Kalahari Cratons by assigning 200 m/s S-wave velocity anomalies in the 7-260 km depth range. For the São Francisco craton, we assigned the same values but until 200 km depth and added a single -200 m/s spike at 260 km depth. We also simulated the low velocities of the Mid-Atlantic Ridge by assigning -200 m/s S-wave velocity anomalies at depths of 7-110 km depth around the ridge itself. Finally, we simulated the presence of deeper hot mantle upwellings underneath the East African Rift and the Tristan da Cunha Island by assigning -200 m/s S-wave velocity anomalies in a 7-260 km depth range underneath the features. The results show clearly that the distortion of the large features we simulate is small, with smoother gradients associated to smaller—and thus more difficult to recover—synthetic bodies, e.g. northern Kalahari and Tristan da Cunha. As expected, the deep, very small-scale, negative anomaly underneath the São Francisco Craton is smoothed out by the inversion. For the larger synthetic bodies, differences are small. This indicates that if the cratons were the broad anomalies previously postulated (e.g., (Begg et al., 2009; French,

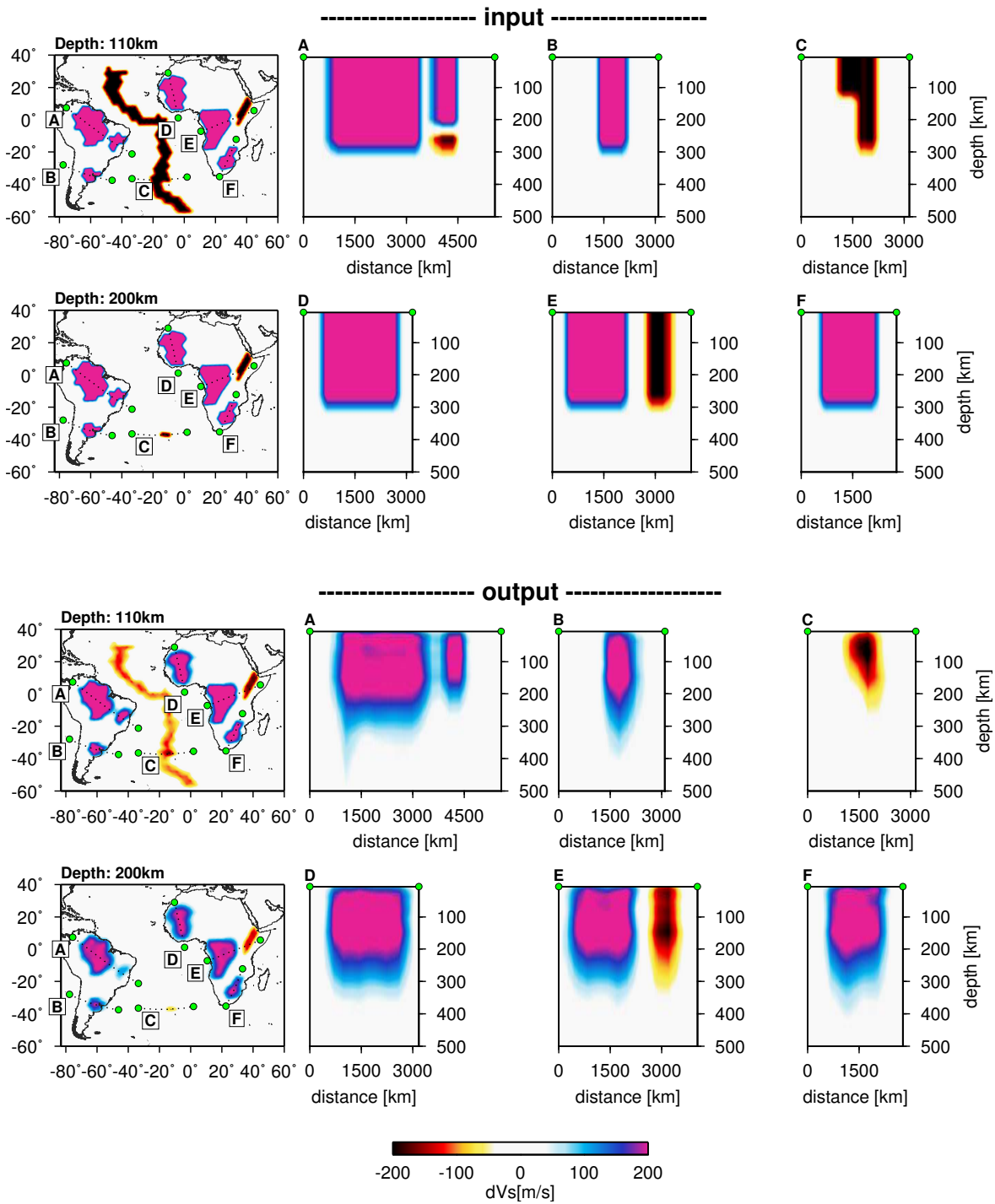


FIGURE 3.12: Synthetic model simulating realistic lithospheric structure. Top rows show the input synthetic structure; bottom panels show the inversion results. On the left panels, we plot horizontal slices through the model at 110 and 200 km. On the right panels, we plot vertical cross sections (A-F) through the model. Location of these cross-sections are plotted on the horizontal velocity maps.

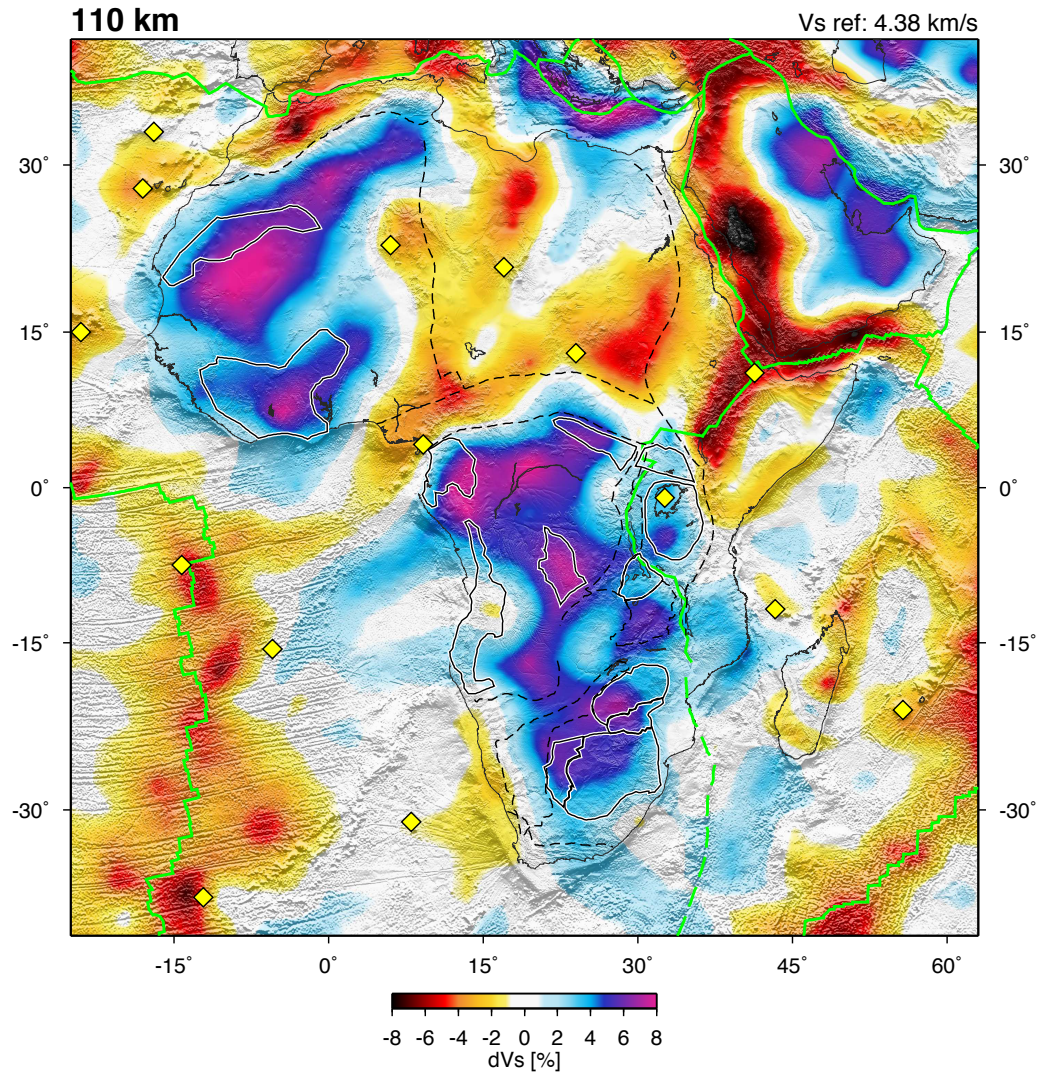


FIGURE 3.13: Plot of the model at 110 km beneath the Africa. Tectonic boundaries from Fig. 3.1 are overlaid for comparison, with Archean Shields as solid black lines and major lineaments and mobile belts as dashed lines. Hotspot locations are shown as yellow diamonds, major plate boundaries as green lines (solid lines: verified; dotted lines: proposed).

Lekic, and Romanowicz, 2013)), our model would correctly reconstruct their shapes. This confirms that the complex geometries we retrieve are real features.

3.5 Africa

The model SA2019 provides a detailed image of the lithospheric structure of the African continent. At a large scale, our model shows agreement with other recent models on the location of the major cratons and low-velocity zones (see section 3.4.2). At smaller scales, our model reveals finer details of the lithospheric structure of the continent, substantially improving on the previous studies. The African cratonic lithosphere and its evolution are discussed in detail in a separate publication (Celli et al., *in press*), and in this section we

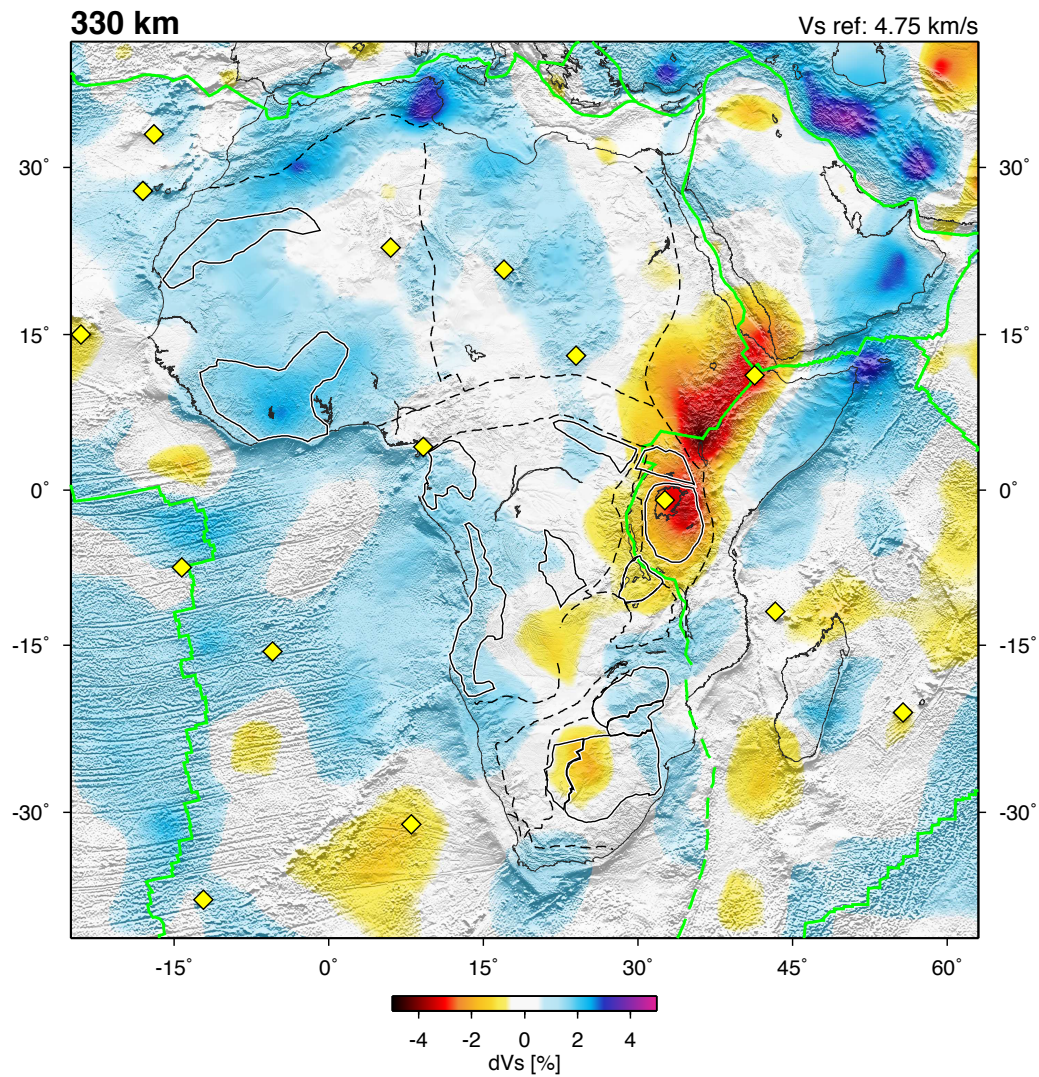


FIGURE 3.14: Plot of the model at 330 km beneath Africa. Tectonic boundaries from Fig. 3.1 are overlaid for comparison, with Archean Shields as solid black lines and major lineaments and mobile belts as dashed lines. Hotspot locations are shown as yellow diamonds, major plate boundaries as green lines (solid lines: verified; dotted lines: proposed).

outline some of the major features of SA2019 only briefly, for completeness.

At 110-150 km depth, our model samples high velocities underneath the Archean crust of the West African, Congo, Kalahari and Tanzanian Cratons (Fig. 3.13). In contrast to a number of previous models (e.g. Fig. 3.10, Begg et al. (2009) and Fishwick (2010)), SA2019 reveals a number of separate cratonic cores within the boundaries of the West African and Congo Cratons, instead of cratonic lithospheres extending beneath the entire cratons. Underneath the Irumide belt and near the southeastern boundaries of the Congo Craton, we image the Niassa Craton, hypothesized previously based on geological data Andreoli (1984) and De Waele et al. (2009) and identify a new unit, the Cubango Craton—both of these unexposed at the surface (Celli et al., *in press*).

In northern Africa, we image low-velocity anomalies beneath the Saharan and East African magmatic provinces (Fig. 3.13) (Montagner et al., 2007; Sicilia et al., 2008; Reusch et al., 2010; Fishwick, 2010; Hansen and Nyblade, 2013). In the Sahara, the location of velocity minima corresponds to the Ahaggar, Tibesti and Darfour hotspots. The low velocities are mostly confined within the Sahara Metacraton and the adjacent Oubanguides Mobile belt, West African Shear Zone and the East African Orogenic Front.

Beneath East Africa, strong low-velocity anomalies locate under the Ethiopian Rift Valley, Red Sea and Gulf of Aden. Under the Ethiopian Rift Valley, low-velocities are visible down to the deep upper mantle and transition zone (Fig. 3.14), indicative of a voluminous upwelling of hot mantle beneath large parts of the East African Rift.

3.6 South America

In South America, our model provides new insights into the structure of the continental lithosphere and underlying mantle. In the lithosphere and sublithospheric mantle, the model provides new information on the lateral and radial extent of the cratons—some of them unexposed at the surface—and the distribution of low velocities, both beneath the Andean Cordillera, active at present, and the mainly Precambrian continental platform (De Brito Neves and Fuck, 2013). In the upper mantle, our model samples the high velocity structures of the subducting Andean slabs, showing good agreement with other, body wave tomographic studies down to the transition zone (Fukao and Obayashi, 2013; Scire et al., 2014; Scire et al., 2017; Chen, Wu, and Suppe, 2019). In this section we present a feature-by-feature discussion on the structure of the South American lithosphere and underlying mantle.

3.6.1 Cratons

The South American continental lithosphere comprises several cratonic units, located primarily in the northern and eastern part of the continent and surrounded by a set of Neoproterozoic mobile belts (De Brito Neves and Fuck, 2013). The main cratonic units are the Amazon and São Francisco Cratons, the surface extent of which is marked by

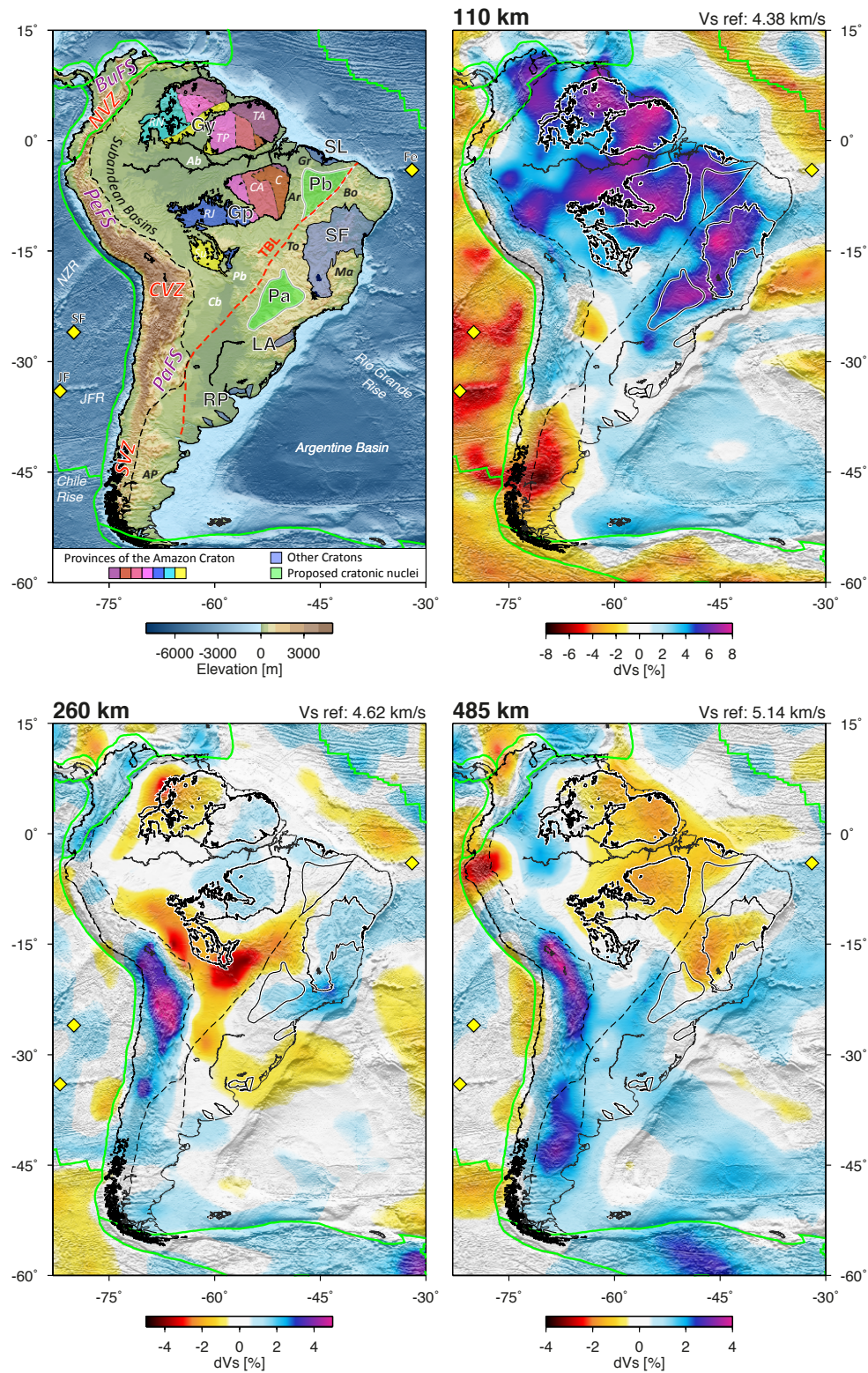


FIGURE 3.15: (Caption next page)

FIGURE 3.15: (Previous page) Top left: map of the main tectonic features of South America. Cratonic units: Gy, Guyana; Gp, Guaporé; SF, São Francisco; SL, São Luís; LA, Luís Alves; RP, Rio de La Plata. Provinces of the Amazon Craton (digitised from (Santos et al., 2008; Rizzotto and Hartmann, 2012) and applied to shield contours by (Brazilian Geological Survey et al., 2016): C, Carajás; CA, Central Amazon; TP, Tapajós-Parima; TA, Transamazon; RJ, Rondônia-Juruena; RN, Rio Negro; Su, Sunsás. Proposed extent of buried cratonic blocks (Brazilian Geological Survey et al., 2016): Pa, Paranapanema; Pb, Paranaíba. Sedimentary basins: Ab, Amazon; Pb, Pantanal; Cb, Chaco. Mobile belts: Gr, Gurupi; Ar, Araguaia; Bo, Borborema; To, Tocantins; Ma, Mantiqueira. Flat Slab sections: BuFS, Bucaramanga; PeFS, Peruvian; PaFS, Pampean. Andean volcanic zones: NVZ, CVZ, SVZ for Northern-, Central-, and Southern Volcanic Zone, respectively. Hotspots are shown as yellow diamonds: SF, São Felix; JF, Juan Fernandez; Fe, Fernando. Other labels: TBL, Transbrasilian Lineament; AP, Argentinian Platform. Major plate boundaries as green lines. Top right and bottom right and left: S-wave velocity anomaly at 110, 260 and 485 km depth under South America. Tectonic boundaries from top left panel are overlain for comparison.

large Archean and Paleoproterozoic crust (Fig. 3.1) (Cordani and Sato, 1999). In addition to these large cratonic regions, other cratonic nuclei, identified by the presence of small portions of Archean and Paleoproterozoic crust (Brazilian Geological Survey et al., 2016), are scattered in the north (São Luis Craton) and south east (Luis Alves and Rio de La Plata Cratons). Geophysical studies also inferred the presence of cratonic roots underneath the Paraná and Paranaíba Basins, even though no associated Archean or Proterozoic crust is exposed at the surface (Feng, Assumpção, and Van der Lee, 2004; Feng, Lee, and Assumpção, 2007; Rizzotto and Hartmann, 2012; De Brito Neves and Fuck, 2013; De Castro et al., 2014).

The dense coverage on the continent provided by the global and regional data constraining SA2019 allows us to determine to a new level of detail the shape and distribution of cratons in South America, highlighting previously unseen details in the structure of the main cratonic blocks as well as imaging the boundaries of unexposed units.

Amazon

The Amazon Craton is commonly identified as a single block centred on the Amazon River. It comprises the Archean and Proterozoic Guyana and Guaporé Shields (Fig. 3.15). Its provinces, identified and dated through a combination of geological and geochronological data (Santos et al., 2000; Santos et al., 2008; Rizzotto and Hartmann, 2012), show an approximately east-west pattern of accretion with the youngest terranes to the west of the craton, forming the Rio Negro, Rondônia-Juruena and Sunsás province (Fig. 3.15). The craton borders to the west with the Sunsás Belt, a Mesoproterozoic orogenic belt that has been interpreted as the southward extension of the Grenville Orogeny in North America (Santos et al., 2008), now covered by the cenozoic sediments of the Subandean Basins (Santos et al., 2008; Brazilian Geological Survey et al., 2016). To the east, the Amazon Craton borders to the Tocantins and Araguaia Fold Belts (De Brito Neves and Fuck, 2013; De Castro et al., 2014), which are part of the Neoproterozoic Brasiliano Orogeny and separate the Amazon from the southeastern cratonic domains, referred as “Brasiliano” or

"Extra-Amazonian" domains in De Brito Neves and Fuck (2013).

SA2019 reveals a complex high-velocity body (+8% dV_s) in the lithosphere under the proposed extent of the Amazon Craton (Figs. 3.15). We image high velocity roots under large parts of the Guyana and Guaporé Shields, mostly corresponding to the Carajás, Central Amazon, Transamazon and Tapajos-Parima provinces, while the younger Rio Negro, Rondônia-Juruena and Sunsás provinces are underlain by lower velocities. In the Amazon Basin, we identify the continuation of the craton under the thick sedimentary cover, which bridges the gap between the two sides of the Tapajos-Parima and Central Amazon provinces, indicating the presence of a large, buried part of the craton. At 200–330 km depth, an area of pronounced low velocities, centred under the Pantanal-Chaco Basin, underlays most of the Rondônia-Juruena and southern Sunsás provinces. Similarly, the Rio Negro Province in the north also lacks a lithospheric root, which is thus present only under portions of the shields that are older than 2 Ga (Carajás, Central Amazon, Transamazon and Tapajos-Parima provinces) (Santos et al., 2000; Santos et al., 2008; Rizzotto and Hartmann, 2012). This implies either that the youngest, western portion of the craton never developed a thick lithospheric root or that its root has been eroded. Possible mechanisms for the removal of the cratonic lithosphere include its reworking during the Sunsás orogeny (similarly to the "crumple zone" model proposed by Lenardic, Moresi, and Mühlhaus (2000)), or metasomatism induced by the hydrous fluids rising from the slab, as hypothesised by Kusky et al. (2014) for both the Wyoming and western Amazon Cratons. Under the the western Guyana shield, we image a strong low velocity anomaly in the asthenosphere. It lays beneath both the cratonic roots of the Tapajos-Parima and Transamazon provinces and the moderately high velocities of the Rio Negro province (Fig. 3.15). Where it underlays cratonic lithosphere, topography shows a steep increase (the Guyana Highlands, from 67–60W, Fig. 3.15), suggesting that the cratonic lithosphere may be pushed up by the anomalously hot mantle underneath, generating dynamic topography in the area. This low-velocity area has been imaged through body-wave tomography by Chen, Wu, and Suppe (2019), who interpreted it as the subducted Caribbean Large Igneous Province (Hoernle, Hauff, and Bogaard, 2004; Spikings et al., 2015). In the north-western tip of the continent, we image a high-velocity lithospheric body west of the Guyana Shield at 80–110 km depth, which protrudes to the west beyond the Cordillera at 110–150 km depth. This anomaly is strongest in the lithosphere, presenting velocities that are typical of cratons. At depths greater than 200 km, less pronounced fast anomalies follow the extent of the Cordillera, mapping the presence of the Bucaramanga Slab in the area. While the velocity anomaly within cratonic lithosphere and a slab can be similar, the location of the shallowest part of this high-velocity unit east of the Andes indicates that its nature cannot be solely explained by the slab, suggesting the presence of a buried cratonic unit. West of the Cordillera, the spatial overlap with the Bucaramanga Flat Slab casts doubts on the nature of the block, but its size—especially at 150 km depth—agrees more with a thick, fast lithospheric unit than a linear, slab-like feature.

Hypotheses have been proposed suggesting a western fragment of the Guyana shield extending underneath the Maracaibo block (Hilst et al., 1994; Cediél, Shaw, and Caceres, 2003); SA2019 images the presence of a cratonic block in this area, possibly wedged between the subducting Bucaramanga slab and the Guyana Shield.

Parnaíba and São Luis

The Parnaíba Basin is a thick sedimentary basin located between the mobile belts of the Neoproterozoic Brasiliano Orogeny (De Brito Neves and Fuck, 2013). The basin is bordered by the Araguaia Fold Belt to the west, the Gurupi Fold Belt to the north, the Tocantins Fold Belt to the south, and the Borborema Fold Belt to the east (De Castro et al., 2014). While the Araguaia and Borborema Fold Belts separate it from the large Amazon and São Francisco Cratons, the Gurupi Belt marks the separation of the basin from the small, Archean São Luis Craton (Nogueira et al., 2017). In the Parnaíba Basin, previous tomographic studies showed pronounced high velocities, indicating thick lithospheric root (Feng, Lee, and Assumpção, 2007; Brazilian Geological Survey et al., 2016) and suggesting the presence of cratonic lithosphere underneath the sediments. The São Luis Craton, by contrast, shows no fast, thick lithosphere in tomographic models, but presents well known cratonic crust (De Castro et al., 2014) and has been interpreted as a fragment of the West African Craton on the basis of geochronological and plate reconstruction studies (Klein and Moura, 2008).

In the Parnaíba Basin, our model shows a high-velocity anomaly under the northern half of the basin. The Parnaíba Craton appears to be connected to the Amazon and São Francisco Cratons at 110 km depth, but slices of the model at 80 and 150 km depth (Fig. 3.8) show markedly lower velocities under the Araguaia and Borborema Foldbelts, suggesting that it is an independent body, separated by narrow corridors, at the limits of our resolution from the neighbouring cratons. In agreement with previous studies (Feng, Lee, and Assumpção, 2007), we do not retrieve high velocities under the the São Luis Craton, indicating the presence of a cratonic root under the Parnaíba Basin and not under the São Luis Craton. Alternatively, the Parnaíba and São Luis may be distinctly different cratonic units, separated by an orogenic belt that is too narrow to be resolved by our tomography.

São Francisco and Paraná

According to plate-tectonic reconstructions (Seton et al., 2012), the São Francisco Craton is the South American counterpart of the Congo Craton. Its surface extent spans from the Atlantic Coast to the border of the Tocantins Belt in the west (Cordani and Sato, 1999), and from the Borborema Belt in the north to the Paraná Basin in the South. The lithosphere of the Paraná Basin is believed to hide a cratonic core, referred to as the Paranapanema block (Pimentel et al., 2000; Mantovani et al., 2005; Feng, Lee, and Assumpção, 2007). The Paraná Basin is also the location of the flood basalts of the Paraná Large Igneous Province

(Thiede and Vasconcelos, 2010), erupted at 135 Ma and believed to have marked the onset of the Tristan Da Cunha Mantle Plume (Morgan, 1983).

SA2019 shows high-velocity, cratonic lithosphere beneath both the São Francisco Craton and Paraná Basin. The Borborema province north of the São Francisco Craton is underlain by weak positive anomalies at lithospheric depths, which indicates that it lacks a thick cratonic root, in agreement with recent P-wave tomography identifying lower velocities in the region (Neto, Julià, and Schimmel, 2019). This is in agreement with the late Proterozoic age of its heavily sheared crust (Cordani and Sato, 1999; Ganade de Araujo, Weinberg, and Cordani, 2014). In the northern Paraná Basin, we observe thick lithosphere with pronounced high velocity anomalies within it, coinciding with the proposed location of the Paranapanema Block (Mantovani et al., 2005). In the southern Paraná Basin, we image weaker (3-4% dV_s) anomalies within a localised high velocity area. The presence of thick, cold lithosphere only in the northern part of the basin, together with the distribution of the Paraná LIP basalts suggests that the lithosphere in the south of the Paranapanema was either thin at the time of the LIP eruption, or that it was thinned by the plume causing the LIP.

Rio de La Plata and Luis Álv

South of the Paranapanema block, several small outcrops of cratonic crust define the evidence for the presence of the Luis Álv (De Brito Neves and Fuck, 2013) and Rio de La Plata Cratons (Sanchez Bettucci, Peel, and Oyhantcaval, 2010; Oyhantcaval, Siegesmund, and Wemmer, 2011). For both of the cratons, the large cenozoic cover (Brazilian Geological Survey et al., 2016) makes their lithospheric extent a matter of debate, with some studies proposing large extent of these (Sanchez Bettucci, Peel, and Oyhantcaval, 2010; Oyhantcaval, Siegesmund, and Wemmer, 2011; De Brito Neves and Fuck, 2013), or even the presence of a single mega craton including both units (Santos et al., 2019). Previous tomographic models (Feng, Lee, and Assumpção, 2007; Debayle, Dubuffet, and Durand, 2016; French and Romanowicz, 2015) typically recover a high velocity signal under the São Francisco and Paranapanema blocks, but not under the any of the proposed extents of the Rio de La Plata Craton, in disagreement with the geological interpretation identifying large, wide (and unexposed) cratonic bodies.

Our model does not retrieve a high-velocity signal under the proposed location of the Rio de La Plata and Luis Álv Cratons, in disagreement with their proposed large size (e.g. Oyhantcaval, Siegesmund, and Wemmer (2011) and Santos et al. (2019)). In order to verify that the absence of thick, cold lithosphere is not due to lack of coverage in the area, we computed a synthetic model simulating the cratonic root of the proposed extent of the Rio de La Plata Craton. Results of this synthetic structural test (Fig. 3.12) demonstrate that if a large, thick cratonic root of the proposed—or similar—size was indeed underlying the outcrops of the Rio de La Plata Craton, our model would be able to retrieve it. If the Craton would be as wide as proposed, the presence of Archean shields and the lack

of a high velocity lithospheric root at present day would suggest that the craton lost its root. While this could be explained by processes of cratonic erosion (section 3.6.1), it is notable that all small, rootless—according to seismic tomography—cratons, i.e. São Luis, Rio de La Plata and Luis Álvares, locate at the margin of the continent. This may indicate that these cratons have been somewhat reworked during the formation/disassembly of Western Gondwana, either by being fragments of the conjugate African cratons as suggested by Klein and Moura (2008) for the São Luis Craton, or possibly as reworked blocks in the collisional/extensional processes.

The Transbrasilian Lineament

The Transbrasilian Lineament (TBL) is a large, seismically active shear zone (Assumpção et al., 2004; Arthaud et al., 2008; De Brito Neves and Fuck, 2013; Curto et al., 2014). It runs within the Neoproterozoic fold belts of the Brasiliano Orogeny (Borborema, Tocantins) (Cordani and Sato, 1999; De Brito Neves and Fuck, 2013) in the North and continues between the Paraná and Pantanal-Chaco Basins in central South America, while its southern end is debated (Fig. 3.15) (Brazilian Geological Survey et al., 2016). According to plate reconstruction models (e.g. Seton et al. (2012)), the paleo-location of the northern end of the lineament coincides with the southern point of the West African Shear Zone (Begg et al., 2009), suggesting a common origin of the two shear zones (Klein and Moura, 2008). Previous tomographic studies identified lower velocities in the area, corresponding to the location of the seismicity related to the lineament (Feng, Lee, and Assumpção, 2007; Rocha et al., 2016).

Our model retrieves a pronounced low velocity anomaly under the TBL, its amplitude increasing towards the south-west. It terminates at a velocity minimum under the Pantanal Basin. This minimum shows low S-wave velocities down to 260-330 km depth. The substantial drop in S-wave velocity anomaly under the location of the TBL and the seismic activity in the area (Assumpção et al., 2004) show that the lineament continues to be to the present day a seismically active zone of hot mantle upwelling (Figs. 3.15, 3.16), following the areas of weak lithosphere formed during the Neoproterozoic Brasiliano Orogeny (De Brito Neves and Fuck, 2013).

3.6.2 The Andes and Patagonia

The Andean Cordillera marks the western boundary of the South American plate, where the Nazca and Antarctic plates subduct. The subduction is divided into different sectors, with the presence of areas of flat slab subduction (Bucaramanga, Peruvian and Pampean Flat Slabs), characterised by absence of volcanism, alternated by the Northern-, Central- and Southern Volcanic Zones (Ramos and Folguera, 2009). In the southern Andes, another volcanic gap (Patagonian) is present where the Chile Rise subducts. Models of crustal thickness (Meijde et al., 2015) agree on thickest crust to be centred around the

orogenic bend known as the Bolivian Orocline, with thinner crust in the northern and southern sectors of the Cordillera. In Patagonia, the scarcity of continental basement outcrops has generated a long-standing debate on its origin as an independent block (Martínez Dopico et al., 2011; Schilling et al., 2017). While Precambrian rocks have been sampled in the region, geochemical evidence show a wide erosion and refertilisation of its lithospheric mantle (Schilling et al., 2017).

Volcanism and low velocities

Under the whole Andean Cordillera, SA2019 shows consistently low S-wave velocity anomalies of -5 to -8% at shallow lithospheric depths (56, 80 km, Fig. 3.8). The velocity minima correlate well with the distribution of main volcanic fields in the Northern, Central and Southern Volcanic Zone (Ramos and Folguera, 2009) (Fig. 3.16). Under the Central Volcanic Zone, we image the lowest velocity anomalies in the continent, corresponding to the areas of most pronounced crustal thickening (Meijde et al., 2015). Beneath this area of low velocities, a thin, nearly-vertical low-velocity conduit is visible down to 330 km depth at the northern edge of the Pampean Flat Slab, suggesting a deeper origin for the strong, broad low-velocities we observe in the shallow lithosphere (Fig. 3.16). In Patagonia, we image low velocities underneath the Cordillera corresponding to the Southern Volcanic Zone (Ramos and Folguera, 2009). At 80 km depth and below, the velocities in the area show a clear minimum where the Chile Rise is subducted, forming the Chile Triple Junction (CTJ). The whole of Patagonia is thus underlain by a 150 km deep low velocity zone centred around the subducting ridge axis. While the subduction of the MOR can be interpreted to be the source for much of the partial melting in the area, the actual point where the ridge subducts—and where we image the lowest velocities—corresponds to the Patagonian Volcanic Gap. North of the CTJ, at 200-260 km depth, the signal from the subducted MOR disappears, with the high velocities of the slab appearing instead. Patagonia shows much greater uplift than the rest of the continent (Flament et al., 2015), with greatest elevation on the flanks of the CTJ. The topography has been proposed to be caused by the presence of a slab window under Patagonia (Guivel et al., 2003; Breitsprecher and Thorkelson, 2009), created by the northward movement of the Chile Rise burning through the subducting plate and generating positive dynamic topography, supported by geomorphology and geodynamic modelling (Guillaume et al., 2009; Pedoja et al., 2011; Flament et al., 2015). SA2019 indicates an area of thinner, hotter lithosphere extending approximately 1000 km north and south of the CTJ, underlaying the whole extent of the high topography in Patagonia. North of the CTJ, velocities are lower and we image the presence of the Nazca Slab at depth, while south of the CTJ we observe less strong low velocities and no slab. We can thus postulate that the lower velocities in northern Patagonia are due to the presence of hydrous fluids rising from the slab underneath, while south of the CTJ no slab is visible.

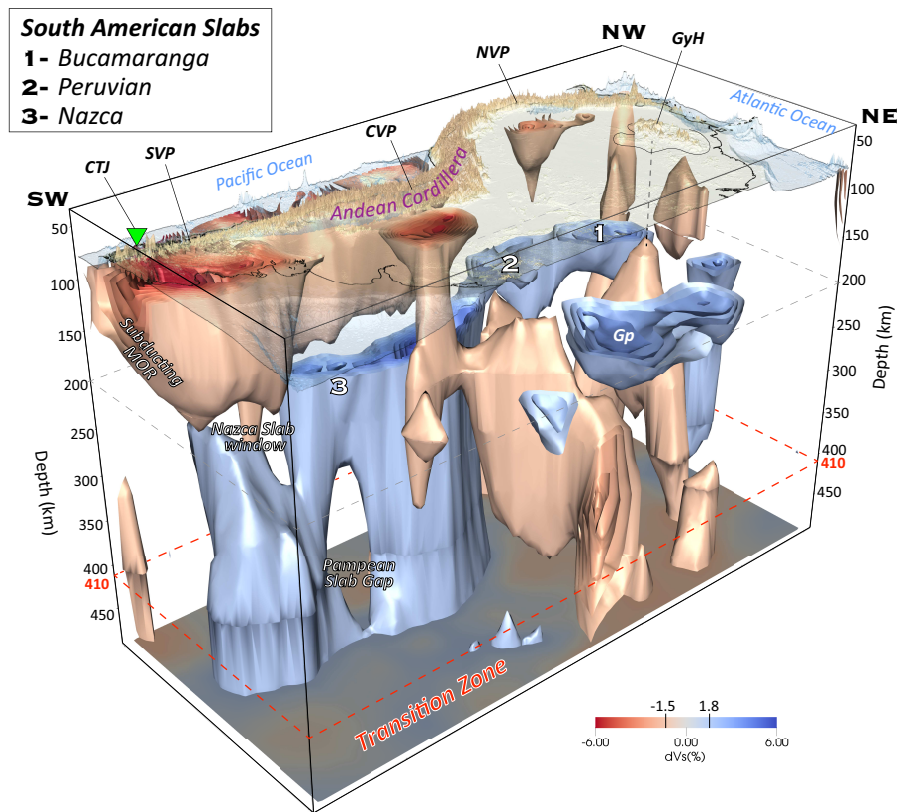


FIGURE 3.16: 3D structure of western South America from SA2019, shown as velocity anomaly isosurfaces. Negative dVs contours smaller than -1.5% are plotted between 80 and 485 km depth, while positive contours for dVs greater than 1.8% are plotted between 200 and 485 km depth only to facilitate view on the slab (200 km depth grid plotted in grey). A model slice cut at 485 km depth underlays the 3D plot. Topography is plotted on top as transparency; coastlines are shown in black. The Guyana Highlands area is drawn as a black line. Volcanic provinces of the Andes are labelled as SVP, CVP, NVP, meaning Southern-, Central- and Northern Volcanic Province, respectively. Other labels: CTJ, Chile Triple Junction; GyH, Guyana Highlands; Gp, Guaporé Craton.

Slabs

Starting at 150 km depth, our model samples the high velocity anomalies of the subducting slabs. At 150 km depth, we image a strong signal under the Bucaramanga, southern Peruvian and Nazca Slabs, with gaps corresponding to the Northern- and Southern Volcanic Zones (for a more thorough discussion on the Bucaramanga Slab and the Maracaibo Subplate, we refer to section 3.6.1). At 200-260 km, we retrieve a continuous image of the slab, with highest velocities around the Bolivian Orocline. The area of highest velocities corresponds to where the oldest—and thus thickest—portions of the Nazca Plate subduct (Capitanio et al., 2011), indicating that the greater S-wave anomalies show the subduction of a much thicker oceanic lithosphere. In the deeper upper mantle and mantle transition zone, we image the Bucaramanga and Peruvian Slabs in the north and the Nazca Slab in the South, separated from the thick, central slab by two slab gaps corresponding to the Peruvian and Pampean Flat Slabs (Ramos and Folguera, 2009) (Fig. 3.16). These gaps

correlate very well with the location of the subduction of the Nazca and Juan Fernandez Ridges and have been proposed to be the result of the Easter Island and Juan Fernandez Hotspots “burning” through the slabs (Portner et al., 2017). In the transition zone, the northern slabs appear to be much lower in velocity compared to the central and southern ones (Fig. 3.15). This can be explained once again by the age and corresponding thickness of the subducted oceanic lithosphere: the Bucaramanga and Peruvian Slabs are subducting much younger lithosphere, implying that its thickest portions (visible now over 1000 km east of the Andes at 485 km depth) have been subducted a longer time ago and spent more time in the mantle, possibly being reheated and showing thus less high velocities. South of the Pampean Slab Gap, we image the Nazca Slab underneath the slab window corresponding to the low velocities of the ridge (see section 3.6.2). Under the Antarctic Plate, SA2019 does not retrieve strong high velocities indicative of the presence of a subducting slab. This can be explained, at least partially, by the very low plate motion rates of the Antarctic Plate (Becker et al., 2015), which indicate in turn little or no subduction south of the CTJ.

3.7 South Atlantic Ocean

In the South Atlantic Ocean, our model reveals complex low-velocity anomalies associated with the mid-ocean ridge, the numerous hotspots and the thickness and anomalies of the cooling oceanic lithosphere. In the next sections we discuss the main features of the oceanic lithosphere and underlying mantle.

3.7.1 The South Atlantic Ridge

The South Atlantic Ridge is the southern hemisphere portion of the Mid-Atlantic Ridge. It extends throughout our model and represents the plate boundary between the African and South American plates. Global tomographic studies (Grand, Hilst, and Widyanoro, 1997; Shapiro and Ritzwoller, 2002; Lebedev and Hilst, 2008; Ritsema et al., 2011; Lekić and Romanowicz, 2011; Debayle and Ricard, 2012; French, Lekic, and Romanowicz, 2013; Schaeffer and Lebedev, 2013; French and Romanowicz, 2014; Chang et al., 2015; Debayle, Dubuffet, and Durand, 2016) sample the MAR, but only few regional studies focus specifically on the South Atlantic oceanic lithosphere and underlying mantle (Silveira et al., 1998; Debayle and Ricard, 2012; Colli, Fichtner, and Bunge, 2013; James, Dalton, and Gaherty, 2014). Most of these models sample continuous velocity lows under the ridge, with lower velocities in the proximity of the Tristan da Cunha and Ascension Hotspots (Fig. 3.10).

Our model recovers a strong, continuous low-velocity signal throughout the extent of the South Atlantic Ridge at all depths down to 200 km, where the ridge-related anomalies are not visible anymore (Fig. 3.17). The axial continuity of the low-velocity signal is interrupted at the Romanche Fracture Zone—an active transform fault that offsets the MAR

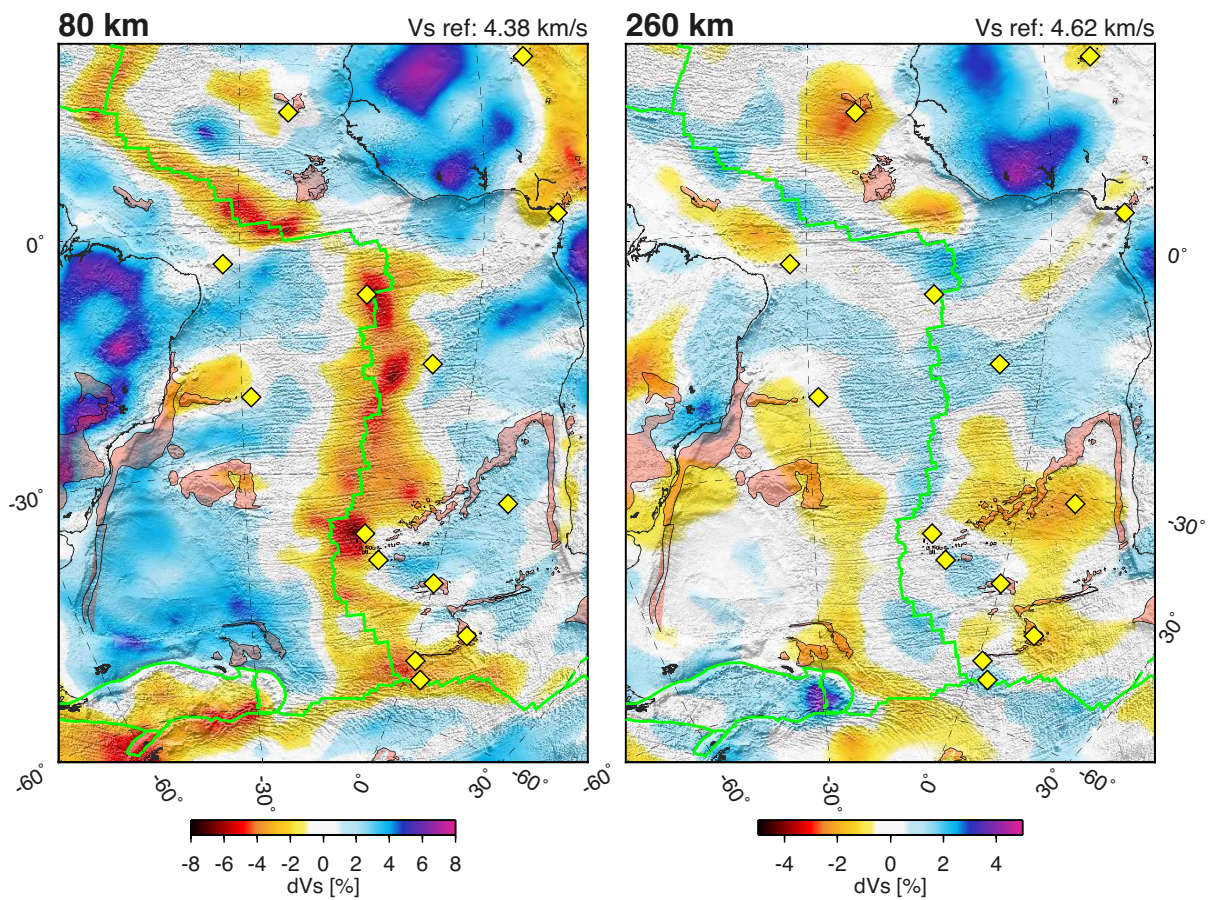


FIGURE 3.17: Plot of the model at 80 and 260 km beneath the South Atlantic Ocean. Oceanic and continental LIP's contours modified from Gaina et al. (2013) are overlaid in pink; hotspot locations are shown as yellow diamonds, major plate boundaries as green lines (solid lines: verified; dotted lines: proposed).

by 900 km—suggesting a disturbance in the mantle flow upwelling pattern caused by the shearing structure. At 80–110 km depth, SA2019 recovers a very clear and detailed pattern in the ridge structure, with many, localised regions of low velocity. Most S-wave-velocity minima in the central section of the MAR are consistently located east of the ridge axis (Figs. 3.17, 3.9), especially in the proximity of hotspots, e.g. the St. Helena and Tristan da Cunha Islands (hotspot features and their interaction with the Mid-Atlantic Ridge will be discussed in more detail in section 3.7.2). The close correlation between the location of the largest shifts from the ridge axis, areas of more pronounced low-velocity anomalies, and the presence of nearby active hotspots suggests a close relationship between the hotspots and the MAR.

3.7.2 Ocean Hotspots

Numerous recognised hotspots and seamounts are scattered across the South Atlantic Ocean, with an unusually high density in the Cape Basin (Thorne, Garnero, and Grand, 2004; O'Connor et al., 2018). The Tristan da Cunha Hotspot, in particular, has attracted a lot of attention due to its evident hotspot tracks and traces of onshore past magmatism in the form of LIPs (Morgan, 1983; Sleep, 1990; O'Connor and Duncan, 1990; Rohde et al., 2013; Schlömer et al., 2017; Bonadio et al., 2018). Low velocities have been detected under a number of Hotspots in different studies (Colli, Fichtner, and Bunge, 2013; French and Romanowicz, 2015; O'Connor et al., 2018), especially Canary and Cabo Verde (French and Romanowicz, 2015), but the seismic structure of the hotspots in the Cape Basin is still debated (Baba et al., 2017; Gassmöller et al., 2016; O'Connor et al., 2018; Schlömer et al., 2017; Bonadio et al., 2018).

SA2019 retrieves pronounced low velocity anomalies from the locations beneath oceanic hotspots, which we can divide in two broad groups by their position relative to the Mid-Ocean Ridge: hotspots close to the ridge (Azores, Ascension, Saint Helena, Tristan da Cunha, Gough, Shona, Bouvet, Marion) and intraplate hotspots, away from oceanic ridges (Madeira, Canary, New England, Fernando de Noronha, Trindade, Vema, Meteor, Discovery, San Felix, Juan Fernandez, Crozet, Reunion, Comoro). For the intraplate hotspots, the low seismic velocity contrasts starkly with the surrounding fast, older oceanic lithosphere, producing clear images of the presence of hot mantle down to 200–260 or, in some cases, 330 km depth (e.g. Cabo Verde, Vema, Trindade; fig. 3.17). The Discovery and Meteor Hotspots are an exception, as the model does not show low velocities beneath the location of these seamounts.

Near-ridge hotspots are more difficult to image due to their proximity to the low velocities of the MAR. They coincide with areas where the low-velocity anomalies from the South Atlantic Ridge are more pronounced (Fig. 3.19), suggesting either a lack of a sufficient lateral resolution of the tomographic image to separate the two features or the existence of an interaction between the magmatism comprising both the ridge- and

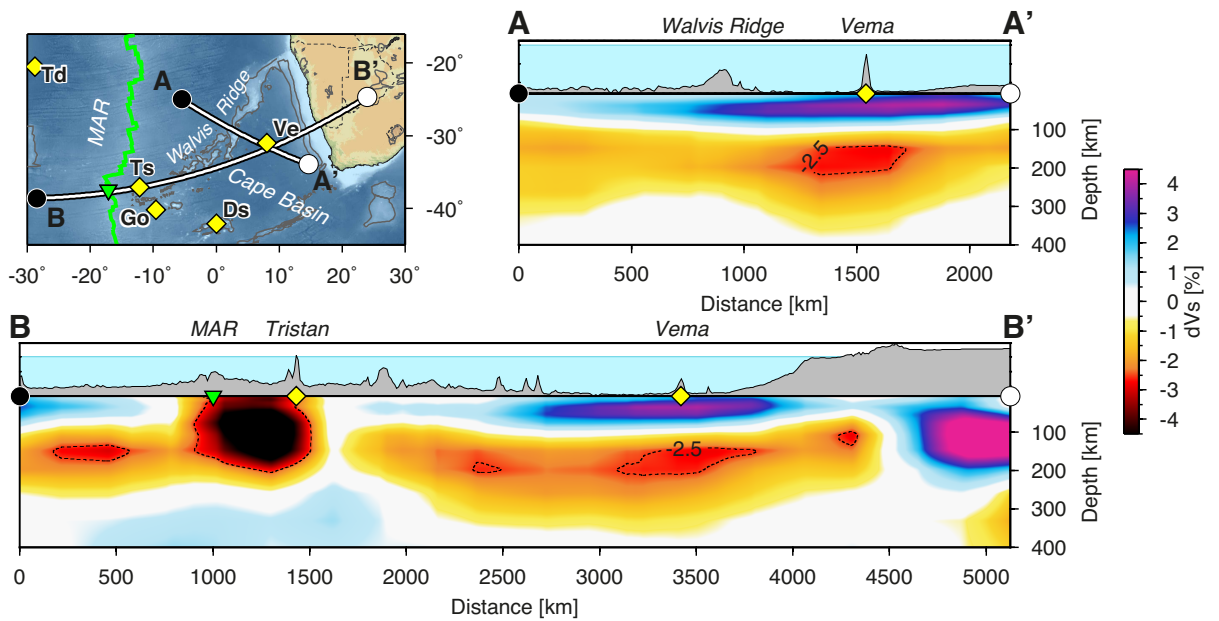


FIGURE 3.18: Cross-section through SA2019 in the Walvis Ridge, Tristan da Cunha and Vema hotspot area. S-Wave velocity anomaly contours of -2.5% are drawn in black to highlight areas of pronounced low velocities. Topography and bathymetry are plotted on top of the sections. Hotspots are shown as yellow diamonds, Mid-Ocean Ridges as green reverse triangles. In the map, Plate boundaries are plotted in green on top of topography and Oceanic and continental LIP's contours modified from Gaina et al. (2013) are overlaid in grey.

hotspot feeding systems (Plume-Ridge interaction (Schilling, 1991)). Between the Ascension and Saint Helena hotspots, the ridge anomalies deviate from the axis of the MAR towards the hotspot location. The Saint Helena hotspot, far from the MAR, locates on the eastern boundary of a strong velocity minimum (anomalies -6 to -7% at 110 km depth) that departs over 500 km east from the ridge axis, while Ascension Island, only 100 km west of the ridge, top a low-velocity anomaly that is centred under the ridge. The Azores and Bouvet islands both sit on top of Mid-Ocean Ridge triple junctions and show laterally extended, pronounced low-velocity anomalies at shallow depths, their signal fading rapidly at 260 km depth.

Tristan da Cunha and Vema

The Tristan da Cunha island is located 460 km east of the MAR, in the youngest part of the Cape Basin. It is believed to be the expression of the mantle plume that generated the Paraná-Etendeka LIP during the breakup of Gondwana (Morgan, 1971; O'Connor and Duncan, 1990; Rohde et al., 2013; O'Connor et al., 2018; Bonadio et al., 2018), with the associated hotspot tracks being the Walvis Ridge to the east and the Rio Grande Rise to the west. For the Walvis Ridge, geochronological data show a clearly increasing age progression towards the Etendeka LIP, further supporting its hotspot origin (O'Connor et al., 2018).

In the Tristan da Cunha area, SA2019 samples a broad velocity minimum between the

hotspot location and the South Atlantic Ridge, with velocities substantially lower than elsewhere beneath the ridge at 56 (-6 to -7% vs. -2 to -3%), 80 and 110 km depth (-8% vs. -6 to -7%) (Fig. 3.18). We infer that the strong low-velocity signal from the ridge partially masks the expression of the Tristan hotspot. To the north-west of the Walvis Ridge, in the Angola Basin, we image a low velocity minimum that does not correspond to any bathymetric feature. The location of this velocity low fits the area of plume-generated crust computed by numerical modelling of the mantle plume (Gassmüller et al., 2016), which extends northwards off the track of the Walvis Ridge. At 150-260 km depth, the velocity minimum shifts east of the Walvis Ridge, towards the location of the Vema Seamount, which is underlain by low-velocities down to 330 km depth (Fig. 3.18). The absence of lithospheric low velocities under the Vema hotspot can be related to the presence of thick, old lithosphere in the Cape Basin, preventing the hot material from rising from the asthenosphere beneath, as suggested by O'Connor et al. (2012) and O'Connor et al. (2018) for the case of the Discovery seamount. At 330 km depth, the low-velocity anomalies beneath Vema, Meteor and Cabo Verde hotspots are the only visible hotspot-related anomalies, making Vema one of the major hotspots in the South Atlantic Ocean.

Under the Walvis Ridge and Rio Grande Rise, at 56-80 km depth, the topographical expression of the two hotspot tracks is located on the boundary between the low velocities in the north and the high velocities of the Argentine and Cape Basins in the south. At 110-260 km, lower velocities underlay part of the Walvis Ridge, although these anomalies do not follow the location of the hotspot track: they connect the Vema hotspot to shallower velocity minima in the proximity of the ridge (Fig. 3.18).

Cameroon Volcanic Line and Saint Helena Rise

The offshore continuation of the Cameroon Volcanic Line (CVL) is a series of seamounts in the northern Angola Basin. Its furthest offshore portion, pointing towards the Saint Helena Island, is often referred as the Saint Helena Rise. The seamount chain protrudes from Cameroon towards the Saint Helena Hotspot, with dated seamounts following an approximately linear age progression (Maher et al., 2015). Approaching the continent however, the Saint Helena Rise overlaps with the CVL, with both off- and onshore volcanism being continuous and geochemically similar since 65 Ma (Fitton and Dunlop, 1985) and thus difficult to relate to a single hotspot track (Halliday et al., 1990; Marzoli et al., 2000; Reusch et al., 2010; Njome and Wit, 2014). The origin of the onshore part of the CVL has been ascribed to a number of different processes: a frozen mantle plume source in the upper mantle (Halliday et al., 1990); edge-driven convection based on tomography and SKS splitting (Reusch et al., 2010; De Plaen et al., 2014); flux from the Afar Hotspot from geodynamic modelling (Ebinger and Sleep, 1998); lithospheric instabilities from laboratory experiments (Milelli, Fourel, and Jaupart, 2012). Fairly uniform transition zone thickness has been interpreted as an additional proof that the volcanism cannot be related to a deep mantle plume (Reusch et al., 2011), and the lack of a strong

anisotropy (De Plaen et al., 2014) does not support the proposed flow from Afar. The Saint Helena Rise presents on the other hand a more clear, progressive set of volcanic ages that suggest a hotspot origin (Maher et al., 2015). Unlike for the onshore part, no previous tomographic study focussed on the offshore segment of the CVL, although most global or large regional models show a drop in seismic velocities in the vicinity of the seamount chain (Colli, Fichtner, and Bunge, 2013; French, Lekic, and Romanowicz, 2013; Debayle, Dubuffet, and Durand, 2016).

Close to the coast and onshore, low-velocity anomalies are visible from 80-260 km depth, with lowest velocities close to Mount Cameroon (Fig. 3.8). In agreement with the lack of a clear hotspot track and the uniform transition zone thickness (Reusch et al., 2011), these anomalies do not continue in the transition zone and seem to be limited to the shallow, hot asthenosphere neighbouring the fast anomalies of the Gabon-Cameroon craton. Offshore, at 80-260 km depth, SA2019 shows continuous, strong low-velocity anomalies following the seamounts of the Saint Helena Rise-CVL, connecting the anomalies of the Cameroon hotspot to the South Atlantic Ridge (Fig. 3.17). While the low-velocity corridor follows the seamount distribution close to Africa, it tilts to the north closer to the MAR, pointing to the Ascension Hotspot rather than Saint Helena, underneath a very wide area of high residual basement depth (Fig. 3.2). The continuous distribution of low velocities along the whole volcanic chain differs from what we image in more well-known hotspots (e.g. Tristan da Cunha, Cabo Verde), where we retrieve stronger, low velocities only at the location of the youngest, present-day volcanism. This structure is difficult to reconcile with a single hotspot origin. Combining our tomography with age information (Reusch et al., 2010; Maher et al., 2015), we can suggest a hotspot origin for the low-velocities we image under Saint Helena and to its west—with the MAR possibly pulling the hotspot to the west from under the island—generating the seamounts of the Saint Helena Rise. The low-velocity channel we image from the CVL to the MAR on the other hand must be ascribed to a different process, possibly indicating an area of thinner, hotter oceanic lithosphere. A possible explanation is that the hotter mantle in Cameroon—be it from a hotspot, edge-driven convection (Reusch et al., 2010; De Plaen et al., 2014), flowing from Afar (Ebinger and Sleep, 1998) or from lithospheric instabilities (Milelli, Fourel, and Jau-part, 2012)—heated and thinned the oldest oceanic lithosphere in the area, in agreement with recent offshore volcanism close to Africa (Reusch et al., 2010). This can have created a lithospheric corridor allowing for hot asthenosphere to flow from Cameroon to the MAR following the gradient of the younging oceanic lithosphere-asthenosphere boundary. While this hypothesis fits volcanic ages (Reusch et al., 2010; Maher et al., 2015), maps of residual topography (Fig. 3.2) and velocities imaged by our tomography (Fig. 3.19), the absence of seamount volcanism on top of this low-velocity corridor remains unexplained.

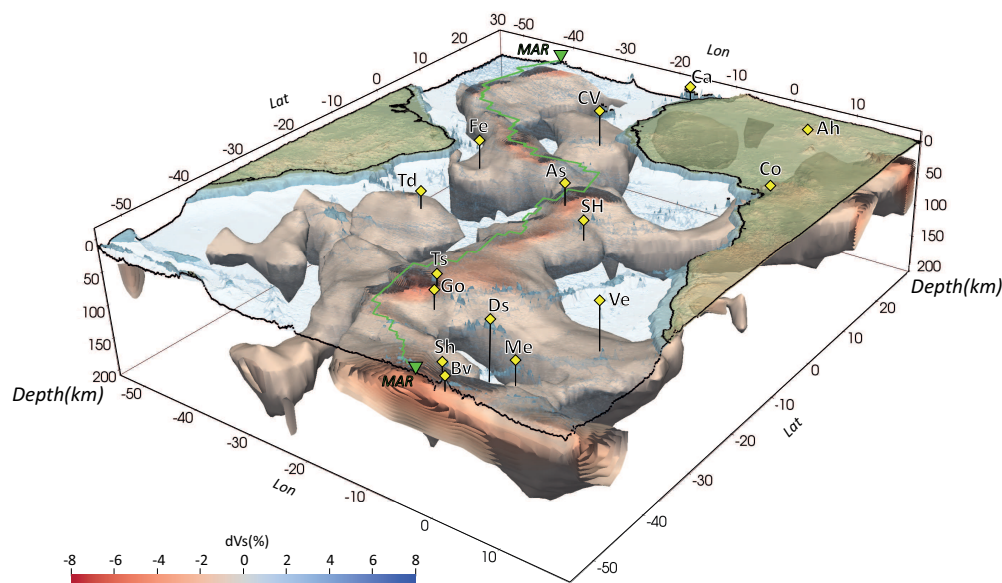


FIGURE 3.19: 3D structure of the South Atlantic low-velocity zones from SA2019 shown as velocity anomaly isosurfaces between 80 and 200 km depth. S-Wave Velocity anomalies contours of -1.5%. The Mid Ocean Ridge at surface is plotted in green. Hotspot are plotted at the surface as yellow diamonds, with sticks connecting to the velocity isosurface at depth to help perspective view: Ca, Canary; CV, Cabo Verde; As, Ascension; SH, Saint Helena; Fe, Fernando de Noronha; Td, Trindade; Ve, Vema; Ts, Tristan da Cunha; Go, Gough; Ds, Discovery; Me, Meteor; Sh, Shona; Bv, Bouvet; Ah, Ahaggar; Co, Mount Cameroon. Coastlines, topography and bathymetry from ETOPO1 are plotted on top as transparencies.

3.8 Cooling of the oceanic lithosphere

The oceanic lithosphere of the South Atlantic imaged by SA2019 is perturbed by a very complex system of low-velocity anomalies, related to many, localised upwellings of hot mantle material. Due to the presence of hotspots and different ridge spreading rates and geometries, the present state and evolution of the lithosphere can differ substantially between different oceanic basins, as we see from the complex low-velocity patterns imaged in the South Atlantic (Fig. 3.17). For an undisturbed evolving lithosphere, the process of thermal cooling and thickening has been proposed to follow either the half-space cooling (Parker and Oldenburg, 1973), or the “Plate” model (Stein and Stein, 1992).

By averaging our model over portions of the oceanic lithosphere of the same age (age data from Müller et al. (2008)), we obtain velocity profiles that reflect the average S-wave velocity structure at each age—which is in turn a proxy for the thickness and temperature of the lithosphere. Combining these profiles for different seafloor ages, we obtain a seismic image of the cooling oceanic lithosphere in the averaged area. When performing the age-averaging over the whole globe, lateral averaging removes the influence of local features, yielding a picture of the evolution of the global average, “normal” oceanic lithosphere with time (Fig. 3.20a). The change from the positive S-wave velocity anomalies of the oceanic lithosphere to the low-velocities of the asthenosphere roughly follows the 1200 C half-space cooling isotherm (Becker et al., 2014).

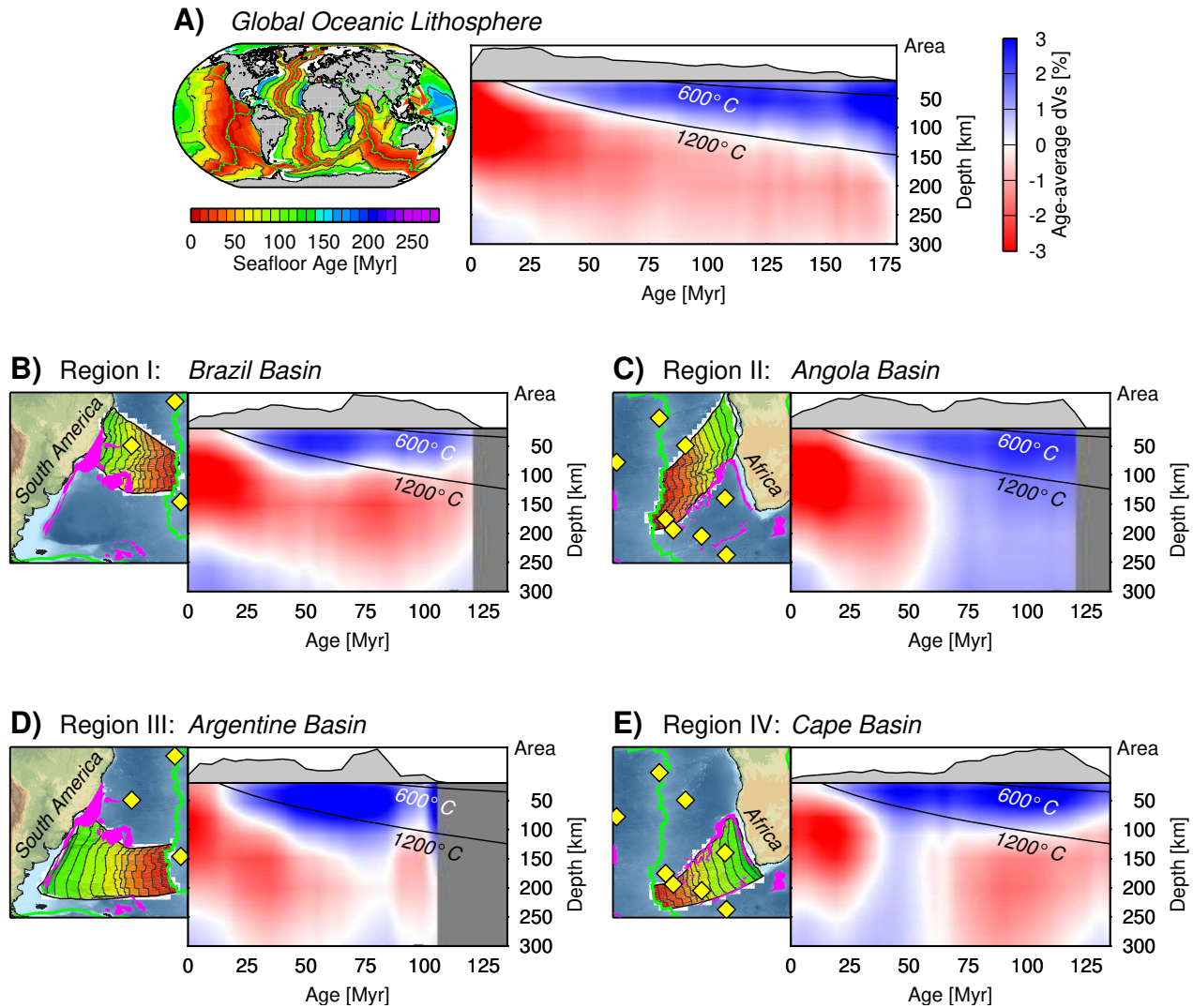


FIGURE 3.20: Age-average cross-sections through the cooling oceanic lithosphere. The averaging is computed for the whole globe (A), and four basins in the South Atlantic Ocean (B,C,D,E). For each region, a map shows the age of the oceanic lithosphere as from Müller et al. (2008) of the averaged area over topography and bathymetry, with hotspots as yellow diamonds, LIPs in magenta (Gaina et al., 2013) and plate boundaries in green. In the age-average cross-sections, the averaged area for each age bin is plotted on top of the plot, and the 1200 and 600 degrees isotherms (Becker et al., 2014) are plotted as black lines.

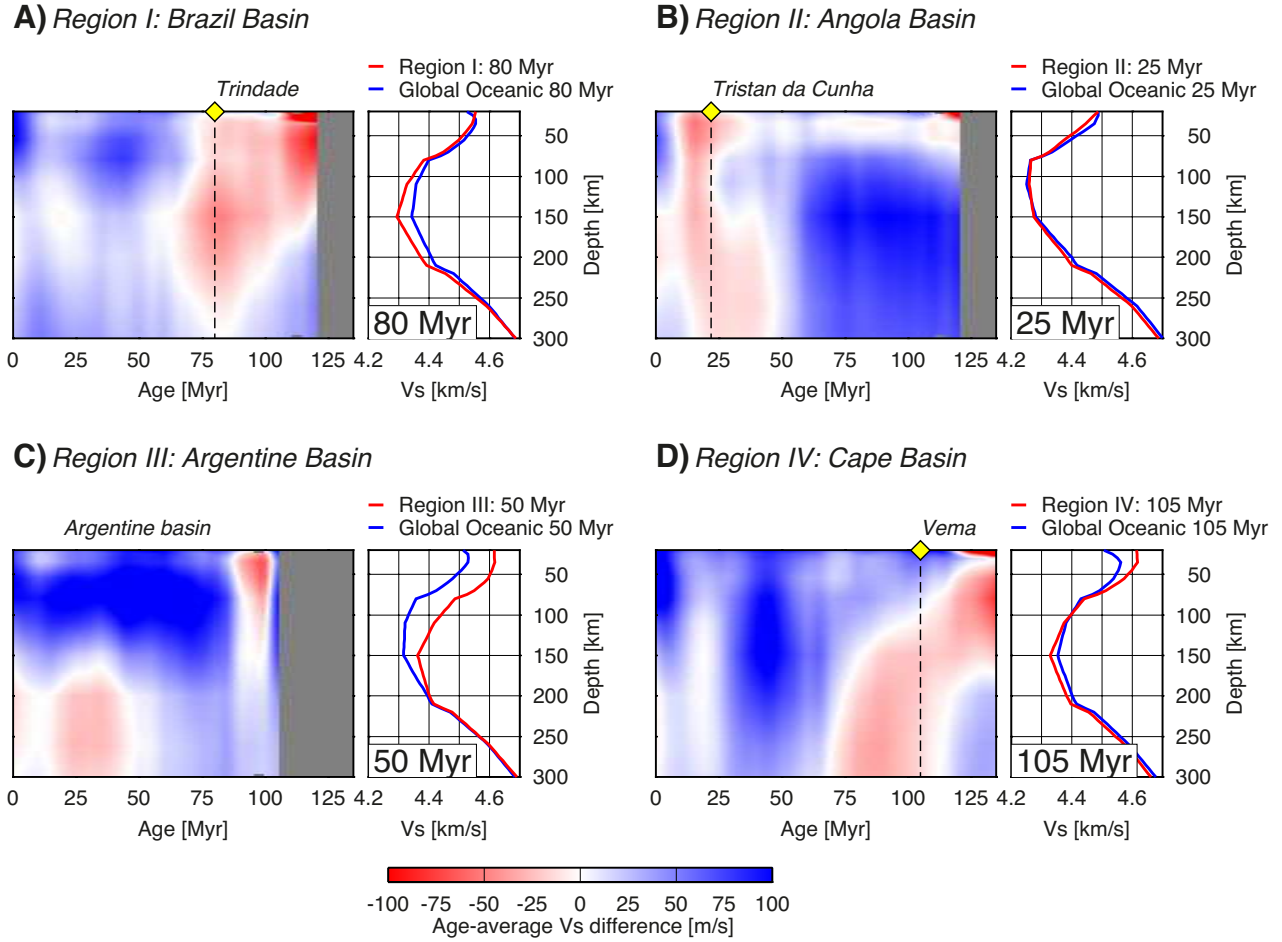


FIGURE 3.21: Local Age-average anomalies with respect to the global for four oceanic basins (A,B,C,D). For each basin, the left panels show the local age-average cross-section; features specific of the averaged area are plotted at the lithospheric age where they locate. Absolute S-wave velocity profiles for the ages of the features are plotted in the right panels: red is the regional average, blue is the global.

In order to analyse local anomalies in the cooling of the oceanic lithosphere, we averaged our model over smaller regions that present markedly different sublithospheric structures: the Brazil, Angola, Argentine and Cape Basins in the South Atlantic Ocean (Fig. 3.20b,c,d,e). When restricting the averaging to these smaller regions, we can observe very different age-average cross-sections in each of them, indicative of the presence of strong, local effects affecting the oceanic lithosphere. Our observations on a global scale show that undisturbed oceanic lithosphere, on average, thickens with age monotonically, including at ages over 100 m.y., but effects of local mantle anomalies have a large impact over the lithospheric structure of particular oceanic basins.

3.8.1 Anomalous cooling

Using the global and local age-average cross-sections, we can quantitatively compute local variations by subtracting the global, “normal spreading” age-average values from the local ones. The resulting cross-section isolates thus the seismic signature of local mantle processes from the one of the average cooling lithosphere (Fig. 3.21). In the Brazil Basin (Fig. 3.21a), we identify pronounced low velocities at 80 M.y.—corresponding to the age of the lithosphere under the Trindade hotspot—and earlier, suggesting that the hotspot affected the evolution of the oceanic lithosphere older than 80 M.y. in the basin. In the Angola Basin (Fig. 3.21b), we can image a clear low velocity “column” at 25 M.y. This structure is the same age as the Tristan da Cunha island (included in the averaging area), indicating that what we see is the averaged seismic signature of the hotspot in the lithosphere and upper mantle. The hotspot signal is masked by the ridge in tomographic images, but is successfully isolated from it by using the differential age-averaging. In the Argentine Basin (Fig. 3.21c), we can identify the presence of anomalously thick, fast lithosphere under most of the basin, which coincides with the negative residual basement depth (Müller et al., 2008), suggesting that at least a large part of the bathymetry anomaly can be explained by lithospheric buoyancy. In oceanic lithosphere of age 90-110 M.y., we can also observe the effect of the tip of the subducting Chile Rise reaching the back of the basin from under the continent (Fig. 3.17). In the Cape Basin (Fig. 3.21d), we observe the very strong low-velocity anomaly underneath the ages corresponding to the location of the Vema seamount (105 M.y. old lithosphere). Complementing the tomographic image, the age-average residuals depict more clearly the anomaly beneath Vema, also highlighting a possible link between its deep signature and the low-velocity anomalies found offshore South Africa (Fig. 3.17).

3.9 Conclusions

Our new model SA2019 provides an improvement in resolution over the previously published tomographic models of the region thanks to the improved coverage and the tuning of regularisation coefficients with synthetic tests, yielding sharper images of both continental and oceanic lithosphere, which can be “ground-truthed”, in a number of instances, by independently derived geological information.

In South America and Africa, the complexity of the cratonic lithosphere revealed by the new model is in stark contrast with the images of the large, monolithic cratons proposed in the past. In Africa, multiple cratonic cores lie within the previously proposed boundaries of the West Africa and Congo Craton, and we image the Niassa and Cubango cratons, buried under thick sediments and not previously identified. In South America we image the shape of both the buried and exposed portions of cratonic lithosphere within the boundaries of the Amazon Craton. We also image the cratonic roots of the Parnaíba

Basin and Paranapanema Block, unexposed at the surface. In the north-west of the continent, we image a possible new cratonic block buried under the Cenozoic sediments of the Subandean Basins and possibly extending underneath the Maracaibo subplate. We also image the lack of a cratonic root underneath the small fragments of cratonic crust of the São Luis, Rio de La Plata and Luis Alves Cratons. Between cratonic blocks, we image clear low-velocity patterns corresponding to large trans-continental shear zones originated at the time of the assembly of Gondwana (Transbrasilian Lineament), acting as regions of active mantle upwelling still at present day. In north-west South America, we image the presence of sublithospheric low-velocity anomalies under the western Guyana Shield, explaining the high elevation of the Guyana Highlands in terms of dynamic topography.

Thanks to the new resolution of our model, we also retrieve accurate information on the subduction under South America. In the north, we image the Peruvian and Bucaramanga Flat Slab. In the central Andes, we image the Nazca Slab and the Pampean Slab Gap. We also image a low-velocity conduit rising from the deep upper mantle under the Central Volcanic Province. In the southern Cordillera, we see that the lithosphere is dominated by the low velocities of the subducted Chile Rise, creating a slab window. South of the Chile Triple Junction, we image no slab under the subducting Antarctic Plate.

In the South Atlantic Ocean, we image complex patterns in the low-velocities underneath the MAR. SA2019 identifies many strong, localised anomalies underneath the ocean hotspots in the area, tying links between the ones closest to the MAR and the signal from the Ridge itself. The shallow lithospheric signal of the Tristan da Cunha hotspot in particular is masked by the one from the MAR, indicating that the hotspot has a relatively small seismic signature. By contrast, we image strong, pronounced low-velocity anomalies underneath the intraplate Cabo Verde and Vema hotspots. We image low-velocities under the Cameroon Volcanic Line-Saint Helena Rise, connecting the MAR to the Cameroon hotspot. The large picture we can get from the new images of the South Atlantic Ocean shows thus a very complex and deeply interconnected pattern of low-velocity anomalies that crosses the ocean both underneath the ridge and the cold oceanic basins.

The global design of SA2019 allows us to estimate the age-average seismic signal of the cooling of the oceanic lithosphere, which, on average, thickens with age. By subtracting the seismic signature of the average spreading oceanic lithosphere from local, basin-scale averages, we can quantitatively isolate a thin, vertical cooling anomaly in the Angola Basin at 25 M.y., which we can relate to the signature of the Tristan da Cunha Hotspot. With the same process, we better image the vertical structure of the Vema and Trindade hotspots and the highly anomalous thick lithosphere underneath the Argentine Basin.

3.10 Acknowledgements

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Chapter 4

African cratonic lithosphere carved by mantle plumes

4.1 Abstract

How cratons, the ancient cores of continents, evolved since their formation over 2.5 Ga ago is debated. Seismic tomography can map the thick lithosphere of cratons, but its resolution is low in sparsely sampled continents. Here we show, using waveform tomography with a large, newly available dataset, that cratonic lithosphere beneath Africa is more complex and fragmented than seen previously. Most known diamondiferous kimberlites, indicative of thick lithosphere at the time of eruption, are where the lithosphere is thin today, implying surprisingly widespread lithospheric erosion over the last 200 Ma. Large igneous provinces, attributed to deep-mantle plumes, were emplaced near all lithosphere-loss locations, concurrently with or preceding the loss. This suggests that the cratonic roots foundered once modified by mantle plumes. Our results imply that the total volume of cratonic lithosphere has decreased since its Archean formation, with the fate of each craton depending on its movements relative to plumes.

4.2 Introduction

The mantle roots of cratons are thought to have been coupled to the overlying crust since their Archean formation and stabilization (Carlson, Pearson, and James, 2005). Cratonic mantle lithosphere is compositionally buoyant, thick (over 200 km), cold and mechanically strong, which is probably what enabled the cratons to survive intact for over 2.5 Ga (Griffin et al., 2003). The occasional destruction of the cratonic mantle lithosphere is well documented, but its mechanisms are debated. It has been attributed to the effects of fluids and metasomatism caused by subduction or rifting and extension adjacent to the cratons (Griffin et al., 1998; Menzies, 2007; Lee, Luffi, and Chin, 2011; Kusky et al., 2014; Dave and Li, 2016; Liao et al., 2017; Wenker and Beaumont, 2017). It has also been suggested that cratonic lithosphere can be destroyed by interaction with thermo-chemical plumes rising from the deep mantle (Sobolev et al., 2011; Guillou-Frottier et al., 2012; Wang, Van

Hunen, and Pearson, 2015; Hu et al., 2018), and recent tomographic models are consistent with the thinning of cratons by plumes in parts of the Baltic Shield, Greenland and the Siberian Craton (Legendre et al., 2012; Lebedev et al., 2018). It remains unclear, however, if a significant proportion of the original volume of the cratonic lithosphere (as it was at, say, the Archean-Proterozoic boundary) may have been eroded or if, instead, the great majority of cratons are preserved.

Seismic tomography detects present-day cratonic lithosphere by anomalously high seismic velocities at and around 100-200 km depths (Fig. 4.9), with these anomalies mainly due to the anomalously low temperatures within the thick lithosphere. Diamondiferous kimberlites and lamproites yield evidence for the existence of the characteristically thick cratonic lithosphere at the time of their emplacement (Boyd, Gurney, and Richardson, 1985). Taken together, the evidence from tomography and kimberlites can offer insights into the temporal evolution of the cratonic lithosphere (McKenzie and Priestley, 2007; Faure et al., 2011). This requires, however, tomography with resolution at the relevant, regional tectonic scales and sufficiently large kimberlite databases.

The African continent is composed primarily of Precambrian terranes, assembled in the Late Neoproterozoic-Early Paleozoic Pan-African orogeny (Cahen et al., 1984; Begg et al., 2009). Three major cratons identified in Africa are the West African, Congo and Kalahari Cratons (Fig. 4.1a), with the smaller Tanzanian Craton located east of Congo (Begg et al., 2009). A number of Large Igneous Provinces (LIPs)—large-scale volumes of both intrusive and extrusive igneous rocks (Coffin and Eldholm, 1994)—were emplaced in Africa over the last 200 million years (My). The Central Atlantic Magmatic Province (Wilson, 1997) (CAMP) at 200 Ma and Paraná-Etendeka (Morgan, 1983) at 135 Ma accompanied or preceded the opening of the central and southern Atlantic Ocean, respectively. The emplacement of the Karroo LIP (Giuliani et al., 2014) at 180 Ma pre-dated the onset of seafloor spreading between Africa and Antarctica at 170 Ma (Klimke et al., 2018). More recently, abundant volcanism has accompanied the development of the East African Rift System (EARS) (30 Ma to present (Ebinger and Sleep, 1998)).

Most tomographic models of Africa show broad high-velocity anomalies beneath its three major cratons (e.g. (Begg et al., 2009; Fishwick, 2010), 4.9). Until recently, however, seismic data coverage in much of Africa has been sparse, limiting the tomographic resolution.

Here, we assemble now available broadband seismic data from new stations in different parts of Africa, which significantly improve the data sampling, and combine them with a very large global dataset. We use waveform inversion to extract structural information from surface and regional S and multiple S waves recorded on the seismograms. The resulting tomographic model AF2019 shows an African lithosphere that is much more complex and fragmented than seen previously. The increased resolution of the imaging makes possible a quantitative joint analysis of the seismic and kimberlite data, revealing continual evolution of Africa's cratons over the last 200 million My.

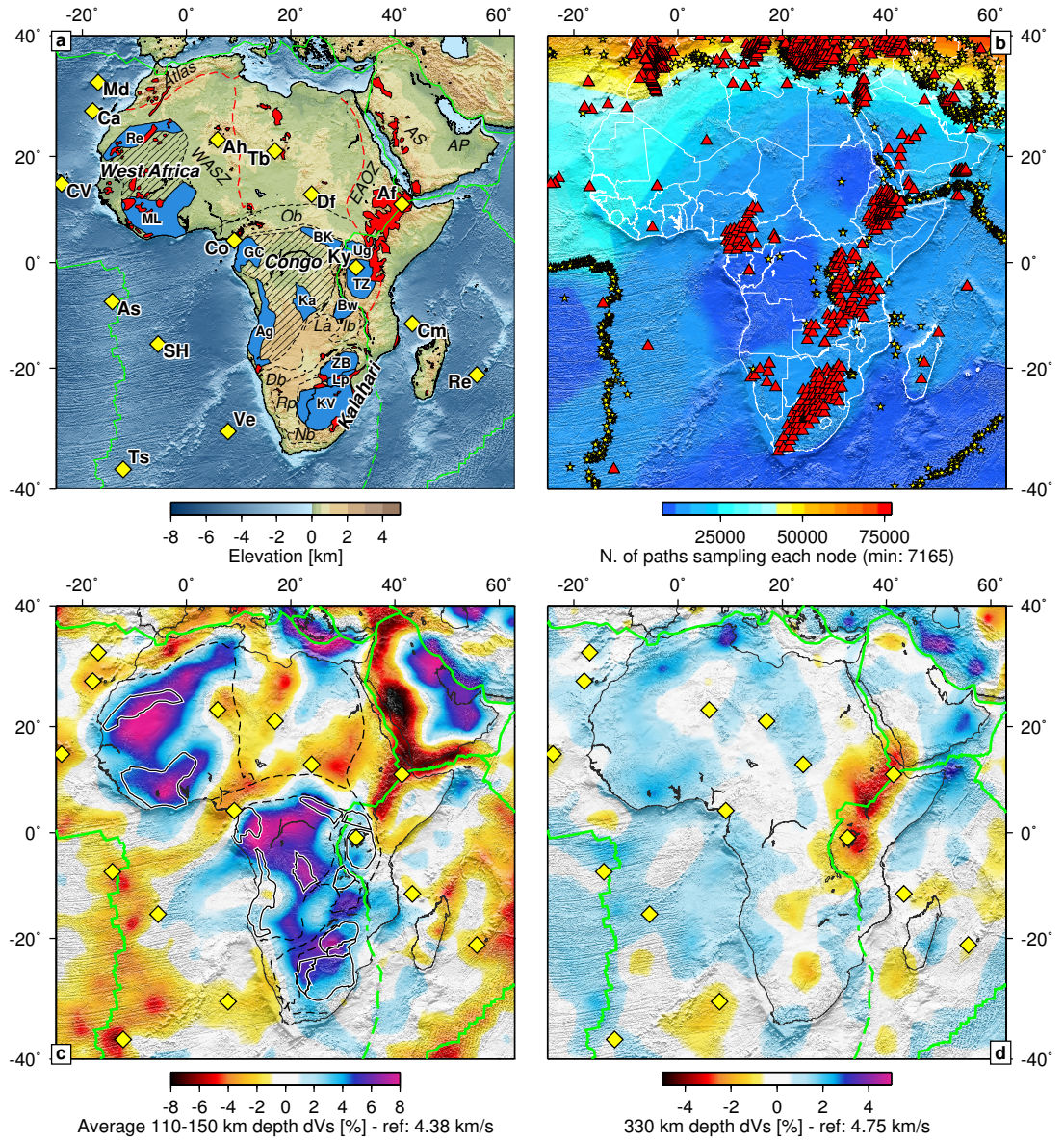


FIGURE 4.1: Main tectonic features, seismic data coverage and our tomography of Africa. a) Cratons and other primary features. Archean shields are plotted in blue: Re, Reguibat; ML, Man-Lèò; GC, Gabon-Cameroon; BK, Bomu-Kibali; Ug, Uganda; TZ, Tanzania; Ka, Kasai; AG, Angola; Bw, Bangweulu Block; ZB, Zimbabwe; Lp, Limpopo Block; KV, Kaapvaal. Mobile belts are plotted in black dashed lines: Ob, Oubanguides; Ib, Irumide; Db, Damara; Nb, Namaqua; La, Lufilian Arc; Rp, Rehoboth Province. Large Igneous Provinces and Volcanics are plotted in red, hotspots as yellow diamonds: Md, Madeira; Ca, Canary; As, Ascension; SH, Saint Helena; Ve, Vema; Ts, Tristan da Cunha; Cm, Comoro; Re, Reunion; Ah, Ahaggar; Tb, Tibesti; Df, Darfur; Af, Afar; Ky, Kenyan. Other features: WASZ, West African Shear Zone; EAOZ, East African Orogenic Zone; AS, Arabian Shield; AP, Arabian Platform; Atlas, Atlas Mountains. b) seismic stations (red triangles) and events (yellow stars) used in tomography, plotted on the hit-count map of data sampling. c) Average shear-wave speed (V_S) in the 110-150 km depth range, with geological features as in (a). d) V_S at 330 km depth.

4.3 Results

4.3.1 Tomography of the African Upper Mantle

Our new, upper-mantle model of Africa and surroundings is constrained by a global dataset of waveform fits of over 1.2 million vertical-component, broadband seismograms, including the newly available data from Africa (Fig. 4.1b, 4.6a). Despite the improvements, the station coverage in Africa remains uneven and relatively sparse. For this reason, it was essential to include the global data, which contained source-station pairs that sampled the Africa region only partially but yielded information complementary to that from the African stations alone. The regularisation of the model was tuned using extensive regional spike tests (section: 4.5). The new regional data, the addition of the global data, and the area-specific regularisation resulted in a substantial improvement in resolution across Africa, compared to previously published global and regional tomographic models (Begg et al., 2009; Fishwick, 2010; Schaeffer and Lebedev, 2013) (Fig. 4.9). The model is global and contains Shear-wave velocity (V_S) distributions beneath other continents and oceans as well; there it is similar to the published models SL2013sv and SL2013NA (Schaeffer and Lebedev, 2013; Schaeffer and Lebedev, 2014).

High-velocity anomalies associated with the cold cratonic lithosphere dominate the model at 100-200 km depths (Fig. 4.1c, 4.7, 4.8). Their depth extent varies from one craton to another. Underneath the EARS, a pronounced low-velocity anomaly extends from Afar in the north to Tanzania in the south and from near the surface down to the deep upper mantle and transition zone, the bottom of the model (Fig. 4.1c-d) (Pasyanos and Nyblade, 2007; Bastow et al., 2008). Major low-velocity anomalies underlie the northern margins of the Red Sea and the Gulf of Aden (on the Arabian Plate, where the majority of the volcanoes are located) (Chang and Van der Lee, 2011); these anomalies extend down to 200-260 km depth (Figs. 4.7, 4.8).

Low velocity anomalies are also present beneath the lithosphere of southern Africa and adjacent ocean basins to the southwest and east of it (Fig. 4.1d), indicating hot asthenosphere beneath the African Superswell (Nyblade and Robinson, 1994). This hot asthenosphere is likely to be responsible for at least a part of the anomalously high elevation observed.

Comparing our new model with those published previously (Fig. 4.9), we observe consistency at larger scales, with all models showing pronounced high velocities beneath the three major cratons. At smaller scales, the higher resolution of our new tomography brings into focus the deep structural variations relating to regional tectonic features. For example, there is no smearing of the high-velocity anomalies, characteristic of cratonic lithosphere, into the Atlantic Ocean (which casts doubt on the notion—put forward previously and based on earlier, smoother tomographic models—that cratonic lithosphere of the Congo and other cratons extends westward beneath the Atlantic Ocean (Begg et al., 2009)). The new data available today allow us to resolve sharper boundaries of the

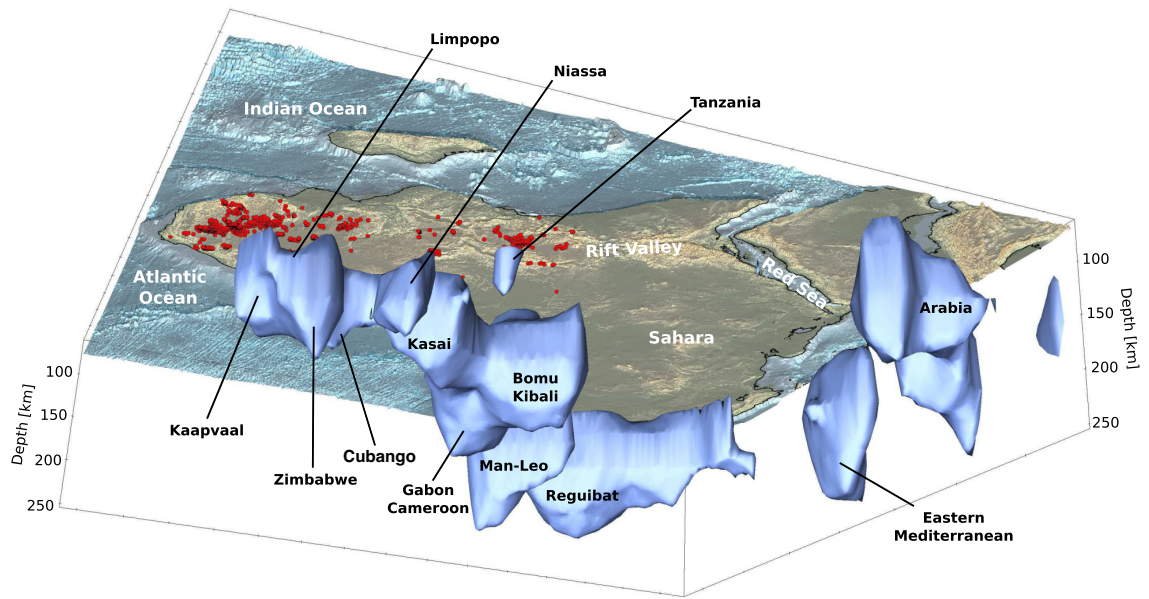


FIGURE 4.2: Three-dimensional representation of cratonic lithosphere in the tomographic model. The view is looking up and to the northwest from beneath the southern Indian Ocean. The $+5\%$ δV_S contour, plotted between 80 and 260 km depths, encloses the nuclei of the cratonic lithosphere. The bottom of the nuclei is not the lithosphere-asthenosphere boundary (that would be closer to 0% δV_S), but thicker nuclei do indicate thicker cratonic lithosphere. The Eastern Mediterranean anomaly comprises the thick, Triassic oceanic lithosphere Seton et al., 2012 and a portion of the Hellenic subducting slab. Subduction is also seen beneath and to the northeast of Arabia. Kimberlite and lamproite locations are shown as red circles.

cratonic lithosphere (Fig. 4.9), located primarily within the outlines of Africa's known major cratons but with substantial complexity and fragmentation, and with a number of separate cratonic blocks outside of them (Fig. 4.1c). Many of the features we discuss could be seen, in a smoother form, in some of the earlier models, in particular SL2013 (Schaeffer and Lebedev, 2013). This is to be expected: an increase in resolution adds detail, rather than changing the image entirely. The higher resolution provided by our new model AF2019 is essential in that it reveals the kimberlite-craton relationship that could not be identified using previous models. In southern Africa, for example, the Kalahari Craton in SL2013 is generally where we see it now but it is smoother and broader. The western boundary of the West African and Congo Cratons are also more sharply defined in the new AF2019 than in SL2013 and other previous models.

In Fig. 4.2, we plot the highest-velocity (and, by inference, lowest-temperature) cores of the cratonic mantle lithosphere using 3D surfaces of positive 5% V_S anomaly. This threshold isolates velocities characteristic of cratonic lithosphere according to global tomography (Schaeffer and Lebedev, 2015) and, alternatively, to temperature estimates from samples from cratonic mantle lithosphere (McKenzie, Jackson, and Priestley, 2005) and conversion of the temperatures to seismic velocities (Agius and Lebedev, 2013). The bottom of these cores is not the Lithosphere-Asthenosphere Boundary (LAB). However,

thicker (and colder) cores do indicate where the lithosphere is the thickest (Fig. 4.7), as expected from the relationship between the lithospheric thickness and temperature given by realistic geotherms (Eeken et al., 2018; Garber et al., 2018). In the thick cratonic lithosphere, the increase of temperature with depth is relatively slow and the LAB can be expected to be marked by only a subtle change in the slope of the depth dependence of temperature and seismic velocity (Ravenna et al., 2018). For this reason, direct estimates of the LAB depth from seismic tomography models are ambiguous, unless thermodynamic modelling including seismic data or models is performed (Fullea et al., 2012). For the purpose of discriminating whether or not the characteristically cold, thick cratonic lithosphere is present beneath a location, the 5% V_S anomaly is an effective threshold. Our results and inferences, however, are not dependent on this particular number and also hold with a 4% or 4.5% threshold (Supplementary Note 1, Fig. 4.12).

4.3.2 Cratonic lithosphere beneath Africa

Known Archean shields occupy only portions of the recognized African cratons (Fig. 4.1a). Archean basement is likely to extend beyond these shields' boundaries, but its complete extent is unknown due to the overlying sedimentary cover. Globally, there is a strong correlation between the locations of high-velocity cratonic lithosphere and the Archean crust above (Schaeffer and Lebedev, 2015). The presence of thick mantle lithosphere, as revealed by our tomography, thus shows where Archean crust is likely to be present as well, even if unexposed at the surface. We can also identify a number of locations with Archean crust but without cratonic mantle roots, which implies their erosion in the course of the cratons' evolution.

In West Africa, our model shows two major high-velocity bodies beneath the Man-Lèò and Reguibat Shields. Between the two lithospheric roots, the lithosphere is somewhat thinner and, thus, warmer, suggesting the existence of two, separate lithospheric units. In the westernmost parts of both the Man-Lèò and Reguibat Shields, the cratonic roots are absent.

Beneath the Congo Craton, previously identified as a single, broad, high-velocity anomaly (Begg et al., 2009; Fishwick, 2010; Schaeffer and Lebedev, 2013) (Fig. 4.9), we image three distinct, high velocity blocks of cratonic mantle lithosphere, with different thickness beneath the Gabon-Cameroon, Bomu-Kibali and Kasai Shields (Begg et al., 2009) (Figs. 4.1c, 4.2). Between these blocks, the lithosphere is also cratonic, but thinner than within them. By contrast, the Angolan Shield in the west does not have any cratonic mantle lithosphere beneath it, apart from its northernmost tip.

Near the southeastern boundary of the Congo Craton, we image another relatively thick, high-velocity lithospheric block, located either just within or just outside the Congo Craton, depending on the definition of the boundary (Jelsma et al., 2018; Begg et al., 2009). Covered by Phanerozoic sediments, the block is characterised by higher topography compared to the neighbouring Owambo Basin to the west and the part of the Damara Belt

that lies to the east of it. Diamondiferous kimberlites (Figs. 4.3, 4.4, 4.5), high P-wave velocities detected previously beneath the eastern part of the block (Yu et al., 2017) and gravity and heat flow data (Leseane et al., 2015) provide further evidence for the presence of cratonic lithosphere beneath this unit. Because this cratonic block underlies the Cubango River basin, we identify it as the Cubango Craton. Our tomography shows that the Cubango Craton is a few hundred km wide and forms a distinct thick-lithosphere unit within the Congo Craton (Figs. 4.1c, 4.2, 4.4, 4.5).

In eastern Africa, a small, relatively thin high velocity root under the Tanzania Craton is underlain by a pronounced low-velocity anomaly associated with the EARS, as seen previously in regional tomography (Weeraratne et al., 2003). South of Tanzania, crustal geology is complex and the definition and lithospheric age of tectonic units are debated (Andersen and Unrug, 1984; Andreoli, 1984; De Waele et al., 2009). Our imaging enables us to identify and map the previously unknown lateral extent of the cratonic lithosphere.

The Bangweulu Block, south-southwest of the Tanzania Craton, displays reworked Archean rocks and has been considered a craton (Borg and Shackleton, 1997; De Waele et al., 2009) but is a product of Proterozoic geodynamic evolution (Andersen and Unrug, 1984). Our model shows no cratonic root beneath this block. A recent regional seismic study (O'Donnell et al., 2013) also shows no high velocities, and the high electrical resistivity detected near the southern boundary of the block (Sarafian et al., 2017) is thus likely to represent thick lithosphere of a unit to the south.

South of Bangweulu, we map, for the first time, the hidden Niassa Craton of Archean age, proposed previously to underlie some of the younger rocks of the Southern Irumide Belt based on geological data (Andreoli, 1984; De Waele et al., 2009). High seismic velocities (O'Donnell et al., 2013) and electrical resistivities (Sarafian et al., 2017) have recently been detected in regional studies, but the lateral extent of the anomalies remained unknown. Our results show that the thick, high-velocity lithosphere of the Niassa Craton, unexposed at the surface, extends as much as 500 km across.

The lithosphere of the Niassa Craton may have played a key role in the localisation of the deformation associated with the southward propagation of the East African Rift System (Koptev et al., 2018). Around the Tanzania Craton, the Eastern and Western Rift branches have developed along the eastern and western boundaries of this mechanically strong block. Further south, the EARS continues as the Malawi Rift, situated along the eastern boundary of the Niassa Craton (see the plate boundary in Fig. 4.1c). The Niassa Craton is thus likely to have determined the location of the rift and contributed to the complexity of the EARS morphology.

In southern Africa, the thickest lithosphere is beneath the western Zimbabwe Craton and Limpopo Belt, with thinner cratonic roots present beneath the northern, central and north-western Kaapvaal Craton (James et al., 2001). Cratonic lithosphere is also present beneath the Paleoproterozoic Kheis-Okwa-Magondi Belt west-northwest of Kaapvaal (Begg et al., 2009), consistent with it having Archean basement (Begg et al.,

2009), and beneath central and northern parts of the adjacent Rehoboth Province (Muller et al., 2009), the location of the previously proposed, unexposed Maltahohe Craton (Anhaeusser, 1990; Begg et al., 2009). At shallow mantle-lithosphere depths, the Limpopo Belt stands out with relatively low V_S between the Moho and ~100 km (Ravenna et al., 2018). Cratonic lithosphere is notably absent beneath the southern and eastern Kaapvaal and north-eastern Zimbabwe Cratons.

The Arabian Plate separated from Africa only around 25 million years ago, with the opening of the Red Sea and the Gulf of Aden (Stern and Johnson, 2010). Thick cratonic lithosphere underlies the Arabian Platform in the eastern part of the plate (Fig. 4.1c), implying unexposed Archean crust—not known from geological data—beneath its thick sediments. The high-velocity anomaly extends north-east just across the main Zagros Thrust, but most of it is well within Arabia (Regard et al., 2010) and, therefore, is due to cratonic lithosphere rather than subduction. The south-eastern margin of the craton is just west of the Oman Mountains. The deep boundary of the Arabia Platform thus determines both the surficial boundary between the Persian Gulf and the Gulf of Oman and the structural boundary between the adjacent Zagros and Makran (Regard et al., 2010) subduction zones. Our results also indicate that the low-elevation Arabian Platform in the east and the high-elevation Arabian Shield in the west of the Arabia Plate have distinctly different lithospheres and asthenospheres (thick and thin, cool and hot, respectively), suggesting that the east-west elevation increase is primarily isostatic, in contrast with a dynamic origin postulated for it in some previous studies (Daradich et al., 2003).

Given the Precambrian age of most of Africa (Cahen et al., 1984; Begg et al., 2009), the complexity of its lithospheric architecture is remarkable. African cratonic lithosphere is highly fragmented, compared to the vast cratonic domains in North America, northern Eurasia or Australia (Fig. 4.10). Average V_S at 100-150 km depth beneath Africa is lower than the global continental average, whereas the lateral V_S gradients are higher than average (Fig. 4.9), indicating greater lithospheric heterogeneity and fragmentation. This points to an extensive reworking of the lithosphere. Another key observation is that a significant proportion of the Archean cratonic crust in Africa is not underlain by characteristically thick cratonic lithosphere. Beneath the western Reguibat and Man-Lèò and nearly the entire Angola Shield, and beneath southern Kaapvaal and north-eastern Zimbabwe cratons, the mantle roots are missing, which implies their erosion by mantle processes.

4.4 Discussion

Lithospheric erosion can be mapped by comparing the present extent of cratonic lithosphere, evidenced by tomography, and its past extent, evidenced by diamond-bearing

kimberlites and lamproites. These rocks serve as proxies for the existence of thick cratonic lithosphere at the time of their eruption, because the pressure-temperature conditions of the diamond-stability field require the presence of thick (>150 km) lithospheric roots (Boyd, Gurney, and Richardson, 1985; Levinson, Gurney, and Kirkley, 1992; Pearson, Canil, and Shirey, 2003). The minimum depth of the origin of kimberlites (~ 150 km) is well established, based on diamond stability and experimental petrology (Mitchell, 1986; Wyllie, 1980; Boyd, Gurney, and Richardson, 1985). It is also well established that a great majority of kimberlites come from the deep lithosphere (with some, possibly, even deeper (Ringwood et al., 1992)) and that at these depths, temperatures that are sufficiently low to create the conditions for the diamond stability field imply a very thick lithosphere, which occurs only in cratons (Boyd, Gurney, and Richardson, 1985; Pearson, Canil, and Shirey, 2003; Artemieva, 2009). Kimberlites provide estimates of both the geothermal gradient and composition in the lithospheric mantle at the time of their eruption and have been used extensively to map the thick Archean lithosphere beneath the cratons of Africa (Begg et al., 2009).

Comparing the distribution of 1606 kimberlite and lamproite samples in Africa (Faure et al., 2011) with V_S anomalies at 110-150 km depth, we observe that at least half of the kimberlites are not in the areas of high-velocity anomalies associated with the thick lithosphere of cratons ($>5\%$, Figs. 4.1,4.2,4.3,4.4). This is in contrast with North America, for example, where most kimberlites are on thick cratonic lithosphere (Faure et al., 2011; Schaeffer and Lebedev, 2014) (Fig. 4.10). This implies that a significant proportion of African cratonic lithosphere (including, at least, the southern Kaapvaal, western Man-Lèò, Angola and Tanzania Cratons) has been destroyed or substantially thinned since the kimberlite eruptions, during the last 200 My.

It has been suggested previously that kimberlites tend to erupt near the edges of cratonic blocks (McKenzie and Priestley, 2007; Faure et al., 2011), with implications for both the origin of kimberlites and their utility as representative deep-lithosphere samples. The lithospheric craton boundaries can be mapped from our high-resolution tomography using either a threshold value or the gradient of δV_S . Regardless of how we define the boundaries, kimberlites in Africa do not demonstrate a preferential location, either closer to cratonic lithosphere boundaries or within cratonic interiors (Fig. 4.12). Many other diamondiferous kimberlites and lamproites are well outside of today's thick cratonic lithosphere, indicating lithospheric erosion.

What processes can result in the destruction of the cold, strong and compositionally buoyant roots of cratons is a matter of debate (Lee, Luffi, and Chin, 2011). Fluids, melt infiltration and metasomatism can weaken and re-fertilize the lithospheric mantle, so that it can then be recycled into the convecting mantle. Proposed mechanisms that could promote this include subduction (Kusky et al., 2014; Liao et al., 2017), rifting and stretching (Wenker and Beaumont, 2017) and mantle plumes (Sobolev et al., 2011; Lebedev et al.,

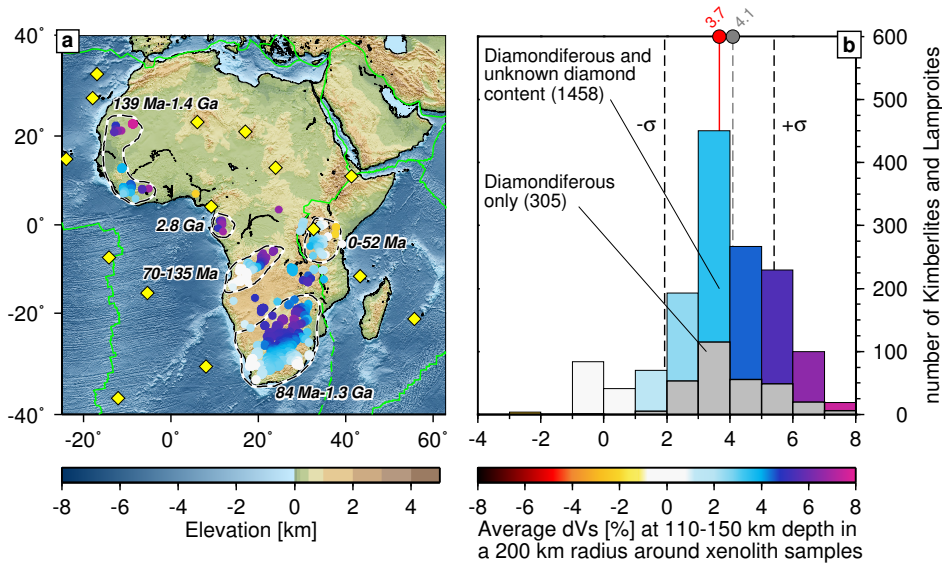


FIGURE 4.3: Seismic velocity in the lithosphere beneath kimberlites and lamproites. The emplacement age ranges from Tappe et al., 2018 are indicated on the map for the different areas (dashed lines). Values of the average δV_S over the 110-150 km depth range and within a 200-km-radius circle around each Kimberlite sample are plotted on the map (a) and as a histogram (b). δV_S colour scale is as in Fig 1c. The circle sizes show the lateral averaging area. Average δV_S across all diamondiferous and unknown-diamond-content kimberlites and lamproites is shown with a red line and dot (b), with the standard deviation also shown. Average δV_S for diamondiferous samples only is shown with a grey line and dot. Both groups present the same pattern: most are not on thick cratonic lithosphere at present. Elevation and bathymetry are plotted on panel a; plate boundaries are plotted in green.

2018; Hu et al., 2018). The destruction of the North China and Wyoming Cratons' lithosphere, for example, has been attributed to hydrous melts rising from subducting slabs (Menzies, 2007; Kusky et al., 2014; Dave and Li, 2016). In Africa, the diamondiferous kimberlite ages (Tappe et al., 2018) require the presence of cratonic lithosphere until ~150 Ma beneath the western Man-Lèò Craton, ~130 Ma beneath the Angolan Shield and ~84 Ma beneath southern Kaapvaal Craton, whereas the most recent subduction in the vicinity was during the Pan-African Orogeny around 500 Ma (Kröner and Stern, 2004), which rules out the subduction-related mechanisms.

Lithospheric stretching associated with rifting may facilitate melt infiltration and lithospheric refertilization, possibly resulting in the destruction of cratonic lithosphere. In Africa, rifting and craton-lithosphere loss happened in temporal and spatial proximity in some but not all cases. The most recent rifting episode relatively close to southern Kaapvaal Craton occurred during the ~170 Ma breakup of southern Gondwana (Seton et al., 2012), but the presence of thick cratonic lithosphere is evidenced by kimberlites until much later, up to 84 Ma (Tappe et al., 2018). More importantly, the rifting was many hundreds of kilometres away from some of the parts of southern Kaapvaal with root loss. Rifting and associated extension are thus unlikely to have caused its lithosphere's destruction.

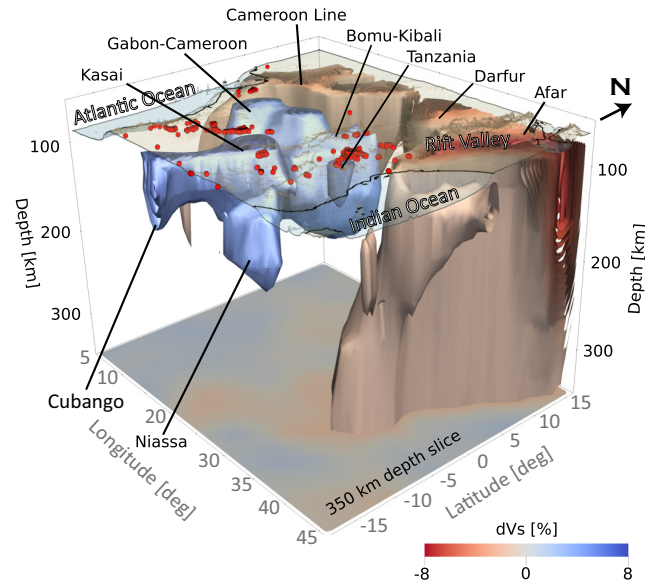


FIGURE 4.4: Three-dimensional view of hotspots and cratons in central-eastern Africa. High velocity isosurfaces at +5% and low-velocity ones at -1.5% are plotted in the 80-350 km depth range. The 350 km δV_S map view is plotted at the depth. Kimberlite and lamproite sample locations are shown as red circles.

The opening of the southern and central Atlantic Ocean was preceded by rifting at the western margins of the Congo and West African cratons. However, the fate of the cratonic lithosphere adjacent to the resultant continental margin is markedly different from one location to another. The Angolan Shield lost its cratonic lithosphere almost entirely, but the southwestern Gabon-Cameroon Shield just to the north of it has not lost any.

African cratonic lithosphere must therefore have been eroded by a different process—most likely, interaction with thermo-chemical mantle plumes (Sobolev et al., 2011; Lebedev et al., 2018; Legendre et al., 2012). LIPs, their origin commonly attributed to mantle plumes (Sobolev et al., 2011; Wilson, 1997; Morgan, 1983; Ebinger and Sleep, 1998) are co-located across Africa with the cratonic lithosphere destruction that we identify. These include the Central Atlantic Magmatic Province (CAMP, 200 Ma), with magmatism on and near the West African Craton; Karroo (180 Ma) on the Kalahari Craton; Etendeka (130 Ma) on the Angola Shield; and Afar-EARS (30 Ma-present), reaching the Tanzania Craton.

We are, at present, witnessing on-going craton destruction in Tanzania, where hot asthenosphere attributed to the Kenyan Plume (George, 1998) is in direct contact with what remains of the lithospheric keel of the craton (Koptev et al., 2015)(Fig. 4.4). Evidence for the past existence of thick cratonic lithosphere is given by diamondiferous kimberlites aged up to 52 Ma, but younger samples (e.g. Igwisi Hills Kimberlite, 0.012 Ma (Tappe et al., 2018)) are barren. This is consistent with Cenozoic erosion of the Tanzania Craton lithosphere and the loss of the pressure-temperature conditions necessary for diamond stability.

The Angolan Shield kimberlites indicate the presence of a thick cratonic root up to 124–135 Ma (Tappe et al., 2018), roughly the time of the Paranà-Etendeka LIP emplacement (135 Ma (Morgan, 1983)). At present, the Angolan Shield has no cratonic lithosphere, in stark contrast with the part of the Congo Craton just to the north, located away from the LIP and with cratonic root intact (Figs. 4.1, 4.2). This suggests that the Angolan Shield lithosphere was eroded by the Tristan da Cunha Plume, thought to have caused the Paranà-Etendeka LIP (Morgan, 1983).

In southern Kaapvaal, the ~180 Ma Karroo LIP (Duncan et al., 1997; Giuliani et al., 2014) overlies the eroded part of the craton. The presence of diamondiferous kimberlites up to 84 Ma (Tappe et al., 2018) indicates that cratonic lithosphere initially survived the impact of the LIP-causing plume but was, eventually, recycled into the convecting mantle ~100 My later. The lithosphere loss appears to have coincided with a pronounced uplift at the southern margin of the craton at ~80 Ma, evidenced by apatite fission track data (Wildman et al., 2017). In the western Man-Lèò Craton, the emplacement of the CAMP at ~200 Ma also preceded the loss of the cratonic keel, by at least 50 Ma.

Fig. 4.5 shows a map of the thick cratonic lithosphere beneath Africa today, as inferred from our high-resolution tomography. It also indicates where cratonic lithosphere has been eroded—this is where diamondiferous kimberlites are not on thick lithosphere at present (red). In addition to known diamondiferous kimberlites, we also plot those with unknown diamond content in pink (for many of these, the presence of diamonds is not listed for commercial reasons—it is well known, for example, that there are diamonds in Angola). Kimberlite ages (Tappe et al., 2018) indicate that the erosion we see has occurred over the last 200 My.

For two example locations of craton-lithosphere erosion (Angola and southern Kaapvaal), we trace their movements according to Africa's absolute plate motion in recent plate-tectonic reconstructions (Seton et al., 2012; Müller et al., 2016; Torsvik and Cocks, 2017) (Fig. 4.5). The tracks show that the cratons with missing lithosphere were, at the time of the LIP emplacement, in close proximity to the hotspots that are now beneath the South Atlantic. The differences in the tracks illustrate the uncertainty in the plate motion of Africa; we also note that the hotspots themselves are not necessarily stationary over 100 My time scales (Seton et al., 2012). For the Angolan Shield, there is no evidence for thick cratonic lithosphere after the Etendeka LIP emplacement, when the craton was above the Tristan da Cunha Hotspot. The lithosphere of southern Kaapvaal Craton, in contrast, was lost while far from any hotspot, after 84 Ma. This confirms that, in this case, the cratonic root was recycled into the convecting mantle ~100 My after the plume impingement.

Our observations imply that the impact of a mantle plume on cratonic lithosphere can weaken and modify it sufficiently so that it is eroded and recycled into the convecting mantle, possibly enhancing and accelerating pre-existing metasomatic weakening processes (Wang, Van Hunen, and Pearson, 2015). The lithosphere loss can occur

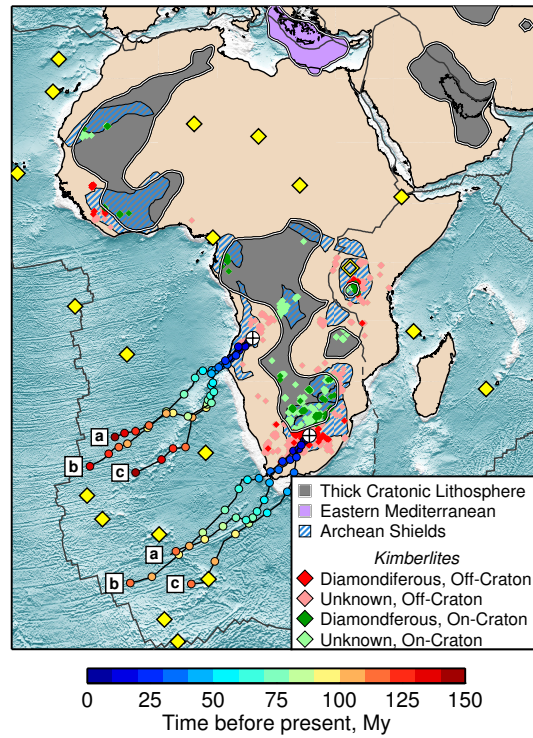


FIGURE 4.5: Preserved and eroded cratonic lithosphere beneath Africa and Arabia. The thick lithosphere present today is shown in grey. The grey areas include all locations with shear-velocity anomaly exceeding +5% in the 80-150 km depth range, which indicates cold, thick cratonic lithosphere (except in eastern Mediterranean where the high velocities show a part of the Hellenic slab and the exceptionally thick, Triassic oceanic lithosphere). Geologically mapped Archean shields Begg et al., 2009 are shown with blue and white stripes. Kimberlites and lamproites atop thick lithosphere at present are distinguished from those atop thinned lithosphere. The latter (red, pink) are indicators of lithospheric erosion. Locations of two now eroded cratons (white crossed circles) are reconstructed back in time following three different plate-tectonic reconstructions: a Torsvik and Cocks, 2017, b Müller et al., 2016, c Seton et al., 2012. Present-day hotspot locations are shown as solid yellow diamonds; the southern, Tanzania end of the elongated EARS anomaly—as empty diamond.

concurrently with or shortly after interaction with the plume (as in the case of Angola and Tanzania) or a few tens of million years later (~ 50 My for western Man-Lèò and ~ 100 My for southern Kaapvaal Cratons).

Not all cratonic lithosphere close to LIPs gets eroded. The lateral extent of the root-loss zone depends upon pre-existing lithospheric structure and how the plume interacts with it. Most root-loss areas we identify are elongated, a few hundred kilometres wide and stretching along the boundaries of the remaining cores of cratons with intact lithosphere. How plumes weaken cratonic lithosphere and how it is then removed by concurrent or subsequent convection processes is an important outstanding problem for future research.

In order to constrain the time of the craton-lithosphere destruction, we focussed on cratons with diamondiferous kimberlites. Root loss, at some point after the craton formation, can also be inferred in other areas, where Archean cratonic crust occurs but is

not underlain, at present, by thick mantle lithosphere (Fig. 4.5), for example, eastern Kalahari and north-western Congo Cratons (adjacent to the Karoo and Cameroon Line basalts, respectively).

In a recent paper, Hu et al. (Hu et al., 2018) also considered possible effects of plumes on the lithosphere of Africa. Using inferences from older, smooth tomographic models (Ritsema et al., 2011) and from the evolution of topography, they proposed that the depleted, buoyant lower part of the cratonic lithosphere beneath parts of Africa's cratons was removed by plumes but then replaced, fairly rapidly, by a new, fertile, dense lower lithosphere, so that the lithosphere remained thick but changed in its composition. A broader inference from this hypothesis is that cratonic roots are episodically removed or thinned by mantle dynamics but then re-grow, regaining their characteristic thickness.

The new evidence from our high-resolution tomography, based on much more data than previous tomography of Africa, confirms the erosion and pin-points its locations. It also shows, however, that the thinning of the lithosphere by plumes is permanent and irreversible, in contrast to the hypothesis of Hu et al. (Hu et al., 2018).

A further inference from our results is that the weakened, thin lithosphere beneath cratons affected by the lithospheric erosion is likely to be vulnerable to reworking in the next orogenic cycle (Begg et al., 2009). The parts of cratons with eroded lithosphere thus have reduced chances of survival for a geologically long time.

The presence of hot, positively buoyant asthenosphere can be expected to increase the surface elevation (Hu et al., 2018). Seismic velocities we observe in the asthenosphere and transition zone beneath southern Africa (Fig. 4.1d) are lower than elsewhere beneath the continent, indicating higher temperature, which can account for at least some of the higher elevation of southern Africa at present.

We conclude that Africa has lost a substantial proportion of its cratonic lithosphere over the last 200 My. During this time, Africa was moving slowly across an area with numerous plumes (Torsvik et al., 2010), which appear to have eroded its cratons, resulting in a more complex and fragmented cratonic lithosphere distribution than in Eurasia, North America or Australia. More generally, this implies that the total volume of cratonic lithosphere globally must have decreased substantially since its Archean formation, with the fate of each craton depending on its plate-tectonic movements and its luck in dodging mantle plumes.

4.5 Methods

4.5.1 Waveform tomography

Our azimuthally anisotropic, S-wave speed, tomographic model is constrained by over 1.2 million seismograms, waveform-fitted using the Automated Multimode Inversion (AMI) (Lebedev et al., 2005) of surface- and S-wave forms. The global waveform dataset is from recordings at 6360 seismic stations, using 27550 earthquakes in total (Fig. 4.6).

Our model is focussed on the Africa region, with the data coverage in this region maximised (all freely available broadband data were included) and the regularisation tuned to optimise the resolution in Africa. The global data complements the regional dataset and ensures dense sampling of the entire Africa and surroundings. Every model grid node of the model is sampled by at least 7767 paths (Fig. 4.1b). Our model-grid nodes have the same coordinates at different depths, so that the number of paths hitting a node does not change with depth, but the structural sensitivity of the data varies from node to node in 3D, as can be seen in the variations of the sums of the columns of the sensitivity matrix (Fig. 4.6). The earthquakes are taken from the catalogue of Centroid-moment tensors from the Global Centroid-moment-tensor (GCMT) project (Ekström, Nettles, and Dziewonski, 2012), since 1994 and since 1990 for selected stations. Source-receiver distances are between 500 and 18000 km.

After preprocessing (quality control, response correction) our waveform inversion procedure comprises three steps. First, we invert the seismogram waveforms using the well-established AMI (Lebedev et al., 2005). AMI computes synthetic seismograms by mode summation and performs waveform fitting of S-, multiple S- and fundamental-mode surface waves to the real data within multiple time-frequency windows, with elaborate time-frequency and phase weighting. The result is a set of linear equations with uncorrelated uncertainties (Nolet, 1990) describing average, depth-dependent S- and P-wave velocity perturbations within 3D sensitivity volumes between the source and receiver with respect to a 3D reference model (Lebedev et al., 2005; Lebedev and Hilst, 2008). In the second step, the equations are inverted together as a large linear system for the 3D distribution of P- and S-wave velocities and S-wave azimuthal anisotropy (Nolet, 1990; Lebedev and Hilst, 2008), using LSQR (Paige and Saunders, 1982). The model is parametrised using a triangular grid with an average 327 km inter-knot spacing and with a depth parametrisation over 18 and 10 triangular basis functions for S- and P-wave velocities, respectively (S-wave velocities: 7, 20, 36, 56, 80, 110, 150, 200, 260, 330, 410-, 410+, 485, 585, 660-, 660+, 809 and 1007 km; P-wave velocities: 7, 20, 36, 60, 90, 150, 240, 350, 485 and 585 km). We invert for S-wave azimuthal anisotropy to ensure anisotropy does not map into isotropic heterogeneities but leave its interpretation for future work. Our 3D reference model comprises CRUST2 (Bassin, Laske, and G., 2000) for the crust, with added topography and bathymetry, and our own 1D average for the upper mantle (Lebedev and Hilst, 2008). In the third step of our inversion procedure, we exploit the data redundancy to select only the most mutually consistent data by means of a posteriori outlier analysis. From an initial 3D model, we compute the synthetic data by matrix multiplication of the model and the sensitivity matrix. We then compare the synthetic and real data and discard the data with the largest misfits. This amounts to the selection of the most mutually consistent data and is effective at removing the data with the largest errors, in particular in the source location and origin time. In this study, we selected the most mutually consistent ~770 thousand seismogram waveform fits to constrain our final

model. Our linear 3D inversion is regularised by means of Laplacian lateral smoothing, vertical gradient damping and slight norm damping (Lebedev and Hilst, 2008). Depth-dependent regularisation parameters were tuned using synthetic spike tests with S-wave velocity anomalies at each depth node at selected locations on the model (Figs. 4.13, 4.14), to make sure that the maximum of the broadened output anomaly was at exactly the depth of the input spike, for each depth node. Figs. 4.13, 4.14 show example spike tests using the final regularisation parameters and scaling factors. We chose these examples to represent both best and worst case scenarios in our model. Vertical δV_S profiles across the anomalies show coincident maxima for the input and retrieved synthetic models. Fig. 4.15 shows an example of a targeted resolution test (its designed detailed in Supplementary Note 2), confirming, in particular, that if the lithosphere of Africa's cratons was not fragmented, as we observe, then the fragmentation would not appear in the model as an artifact.

4.5.2 Kimberlites and lamproites database

We used a global compilation of 4244 kimberlite sample locations (Faure et al., 2011). The database includes kimberlites and lamproites that are diamond bearing, barren and with unknown diamond content. In our analysis, we used African samples that were either diamond-bearing or with unknown diamond content, aiming to include samples for which the database gives no information on the diamond content but which are from well-known diamondiferous regions (e.g. Angola, Tanzania). To avoid the sampling bias generated by pipes containing many samples, we re-sampled the data on a 0.03-degree grid, with samples in the same cell counting as one datum (Supplementary Note 3, Fig. 4.11). We complemented the spatial information from this database with age information on confirmed diamondiferous kimberlite occurrences from (Tappe et al., 2018).

4.6 Acknowledgements

We thank Maya Kopylova, Balz Kamber, Dave Chew and Emma Tomlinson for discussions on the origin and composition of kimberlites and of cratons. We acknowledge the Incorporated Research Institutions for Seismology (IRIS; <http://www.iris.edu>), the GEON Global Seismic Network (<https://geofon.gfz-potsdam.de>) and Observatories and Research Facilities for European Seismology (<http://www.orfeus-eu.org>) for providing the data used in this study. We acknowledge Africa Array (Nyblade et al., 2011) for creating important parts of the dataset used in this study. The maps and cross-sections were implemented with the Generic Mapping Tools (GMT, (Wessel et al., 2013) and Paraview (<https://www.paraview.org>). This work was supported by the Science Foundation Ireland (SFI) grant 13/CDA/2192, with additional support from grant 16/IA/4598, co-funded by SFI, the Geological Survey of Ireland, and the Marine Institute and from the SFI grant 13/RC/2092, co-funded under the European Regional Development Fund.

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4.7 Supplementary information to Chapter 4

4.7.1 Supplementary Note 1

We observe that kimberlites and lamproites in Africa do not tend to be located close to the boundaries of the cratonic mantle lithosphere, as proposed previously (McKenzie and Priestley, 2007; Faure et al., 2011). We compared the location of the kimberlite and lamproite samples with those of the cratonic boundaries, mapped the average S-wave velocity anomalies in the 110-150 km depth range and, alternatively, using Vs gradients (Fig. 4.10). We defined the boundary areas as all points within a 200-km wide band just outside the craton boundary (defined using different, alternative shear-velocity anomaly thresholds) (Fig. 4.10a). In order to test the robustness of the patterns, we repeated the comparison with a different definition of the boundary area (+/-150 km from the boundary, 300 km width in total) (Fig. 4.10b). These wide boundary regions, with alternative different widths and differently defined boundary lines, account for differences in possible definitions of cratonic boundaries and also for the finite resolution of our tomography. With all tested margins, no preferential distribution of the samples within the boundary regions is observed, indicating that cratonic margins do not appear to control their distribution (Fig. 4.10a-f). Comparing the kimberlite and lamproite distributions to the gradients of the S-wave velocity anomalies, we can also observe that most of the samples occur where the gradients are low (Fig. 4.10g,h,i), lower than continental average. This confirms that the African kimberlites and lamproites are not distributed preferentially at or near the craton-lithosphere boundaries.

4.7.2 Supplementary Note 2

To verify that the lithospheric fragmentation we observe is not caused by artifacts due to insufficient or uneven data coverage, we conducted a series of resolution tests, including spike tests (Figs. 4.12, 4.13) and a targeted structure test (Fig. 4.14). In the structure test, the synthetic input model simulated the broad anomalies beneath the major cratons seen in earlier tomographic models. A velocity anomaly of 250 m/s was assigned to all model nodes within the broad-craton boundaries (Begg et al., 2009) in the 7-250 km depth interval. A velocity anomaly of -200 m/s was assigned to the nodes beneath the Ethiopian Rift Valley in the 7-250 km depth range, in order to simulate the East African Rift Zone. The results (Fig. 4.14) show accurate recovery of both the shape and amplitude of the anomalies, with minor differences in the simulated Kalahari craton, whose northern part in the test was very narrow. The tests show that if the cratonic lithospheres were the broad, monolithic features seen in earlier, smoother models, based on smaller datasets, then our tomography would have retrieved them as such.

4.7.3 **Supplementary Note 3**

Because some kimberlite pipes yield numerous samples, the locations of these pipes may get excessive weight in the distributions we examine. We have thus re-sampled the kimberlite and lamproite data on grids and tested a range of different grid cell sizes, including 0.03, 0.1 and 0.5 degree cell sizes (Fig. 4.15). As the cell size increases beyond 0.03 degrees, not only samples from the same pipes but also those from different, closely-spaced pipes get averaged together, given the width of a pipe typically a few hundred meters (Kjarsgaard, [2007](#)). This leads to a bias, with excessive weight for isolated samples. The patterns we observe, however, are robust and present both without the resampling and with resampling with various cell sizes: most diamondiferous kimberlites and lamproites in Africa are not in areas with the high shear velocity anomalies in the lithospheric depth range characteristic of thick cratonic lithosphere.

4.7.4 **Supplementary Figures**

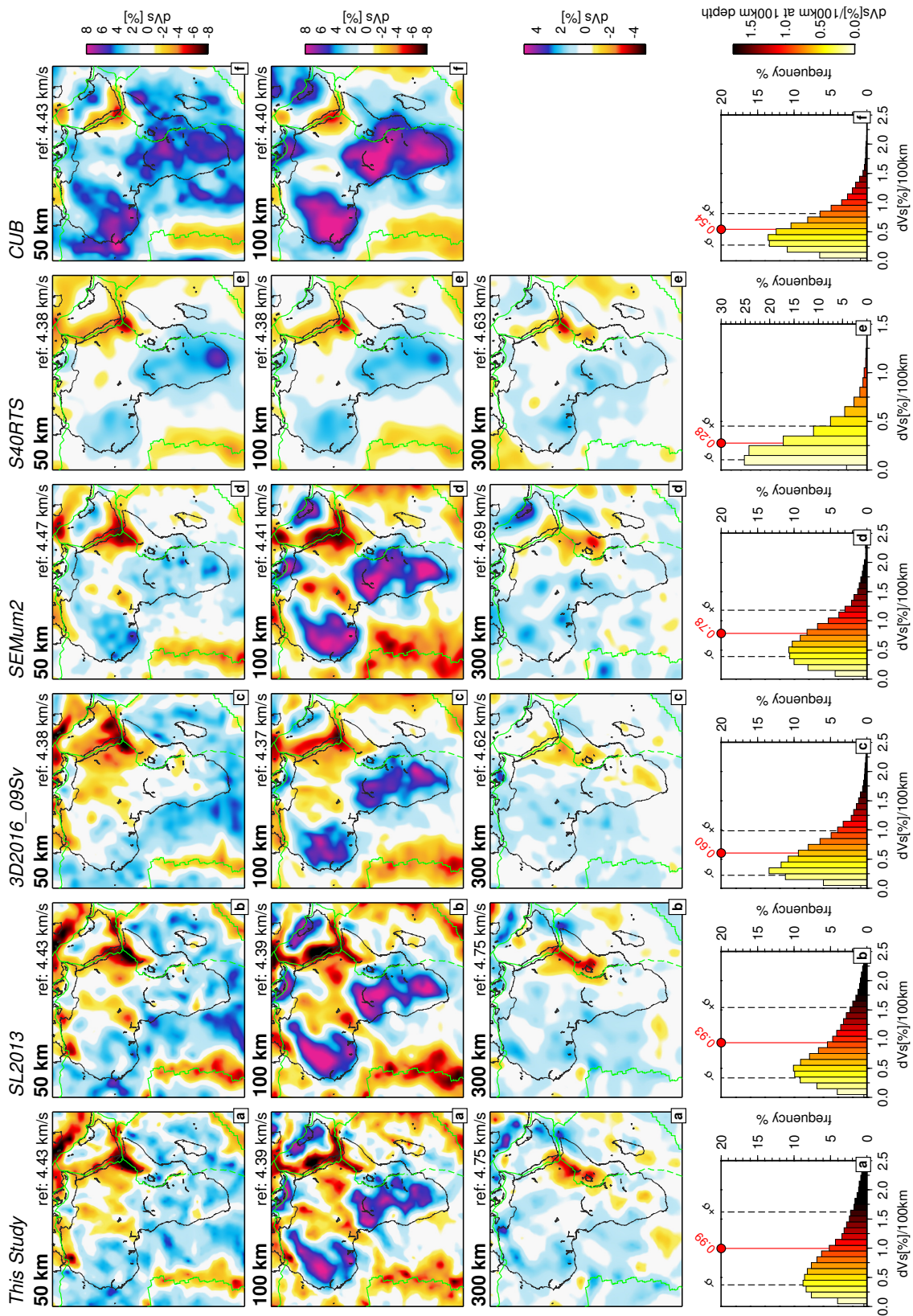


FIGURE 4.6: Caption next page.

FIGURE 4.6: (Previous Page) Comparison of our tomographic model AF2019 with other tomographic models. Tomographic mapviews at 50, 100 and 300 km depth are shown for the following models: a) AF2019; b) SL2013 (Schaeffer and Lebedev, 2013), c) 3D2016_09Sv (Debayle, Dubuffet, and Durand, 2016); d) SEMum2 (French, Lekic, and Romanowicz, 2013); e) S40RTS (Ritsema et al., 2011); f) CUB (Shapiro and Ritzwoller, 2002). All models are plotted as S-wave velocity anomalies with respect to the mean velocity at depth, shown on the top right corner of each model plot. Major plate boundaries are plotted as green lines (solid lines: verified; dotted lines: proposed). The bottom row shows histograms of the S-wave velocity anomaly gradients in continental Africa; averages are plotted in red, standard deviations as dashed black lines.

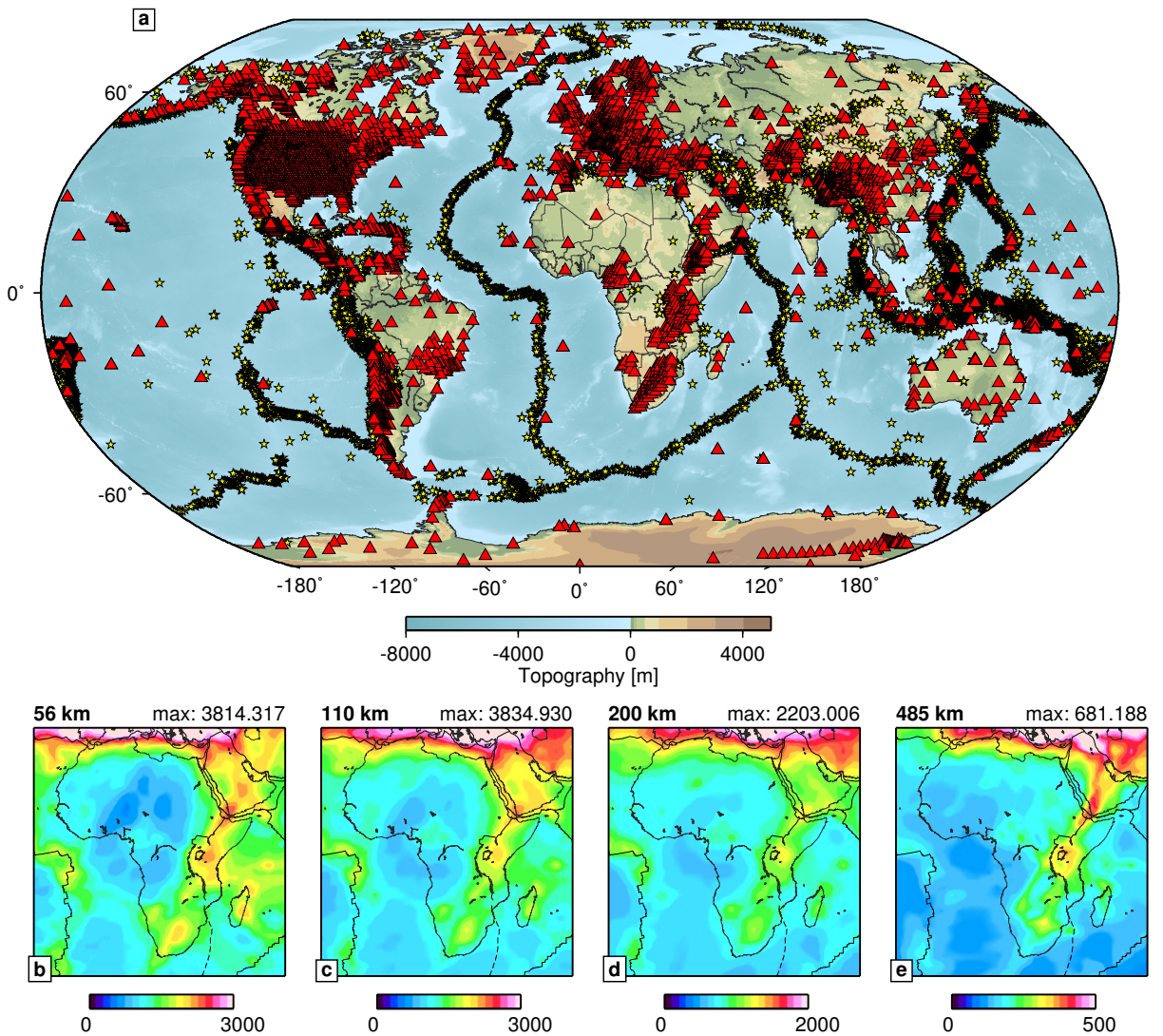


FIGURE 4.7: The data sampling we used to compute our tomographic model. a): Global distribution of seismic stations (red triangles) and earthquakes epicentres (yellow stars) used in our waveform tomography. b-e): The sums of the columns of the sensitivity matrix at four selected grid node depths (56, 110, 200, 485) that represent most of the depth range investigated. The column sums yield an adimensional measure of the relative sampling of the structure at each node by our waveform data set.

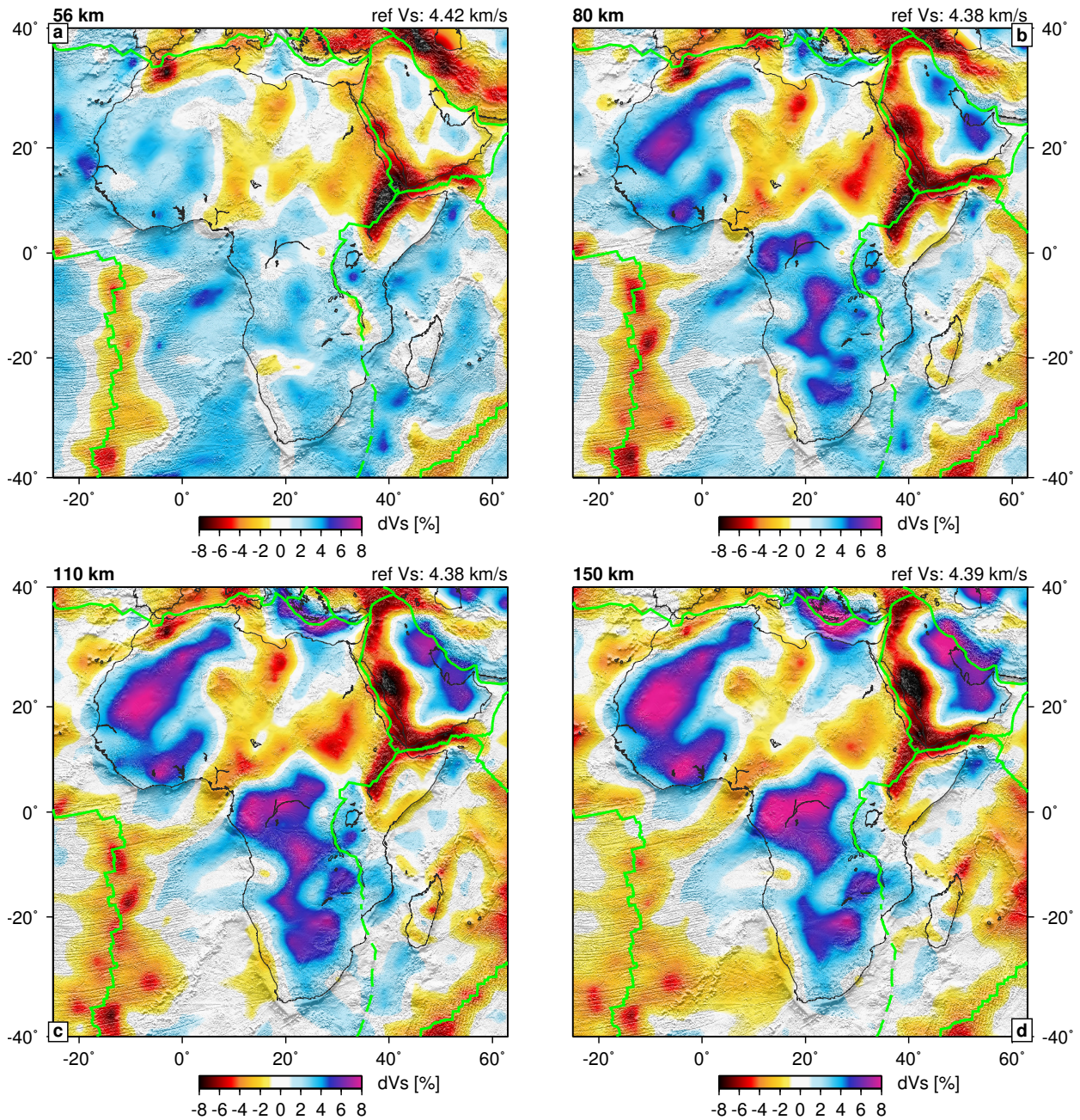


FIGURE 4.8: S-wave velocity anomalies at four depths in the shallow upper mantle: a) 56 km; b) 80 km; c) 110 km; d) 150 km. The depth is indicated above each panel on the left, with the reference velocity on the right. Topography is superimposed as shading.

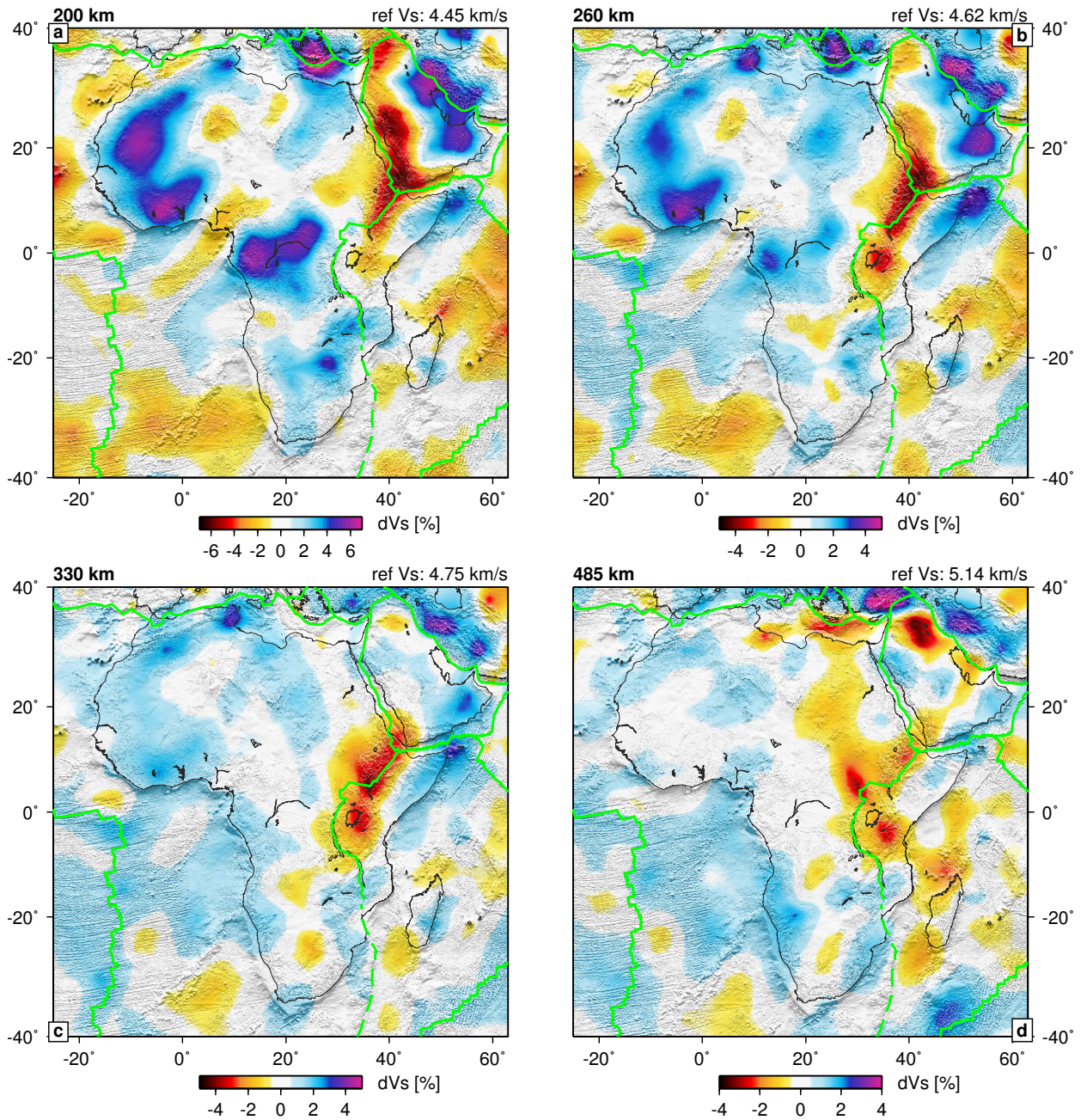


FIGURE 4.9: S-wave velocity anomalies at four depths in the deep upper mantle and transition zone: a) 200 km; b) 260 km; c) 330 km; d) 485 km. The depth is indicated above each panel on the left, with the reference velocity on the right. Topography is superimposed as shading.

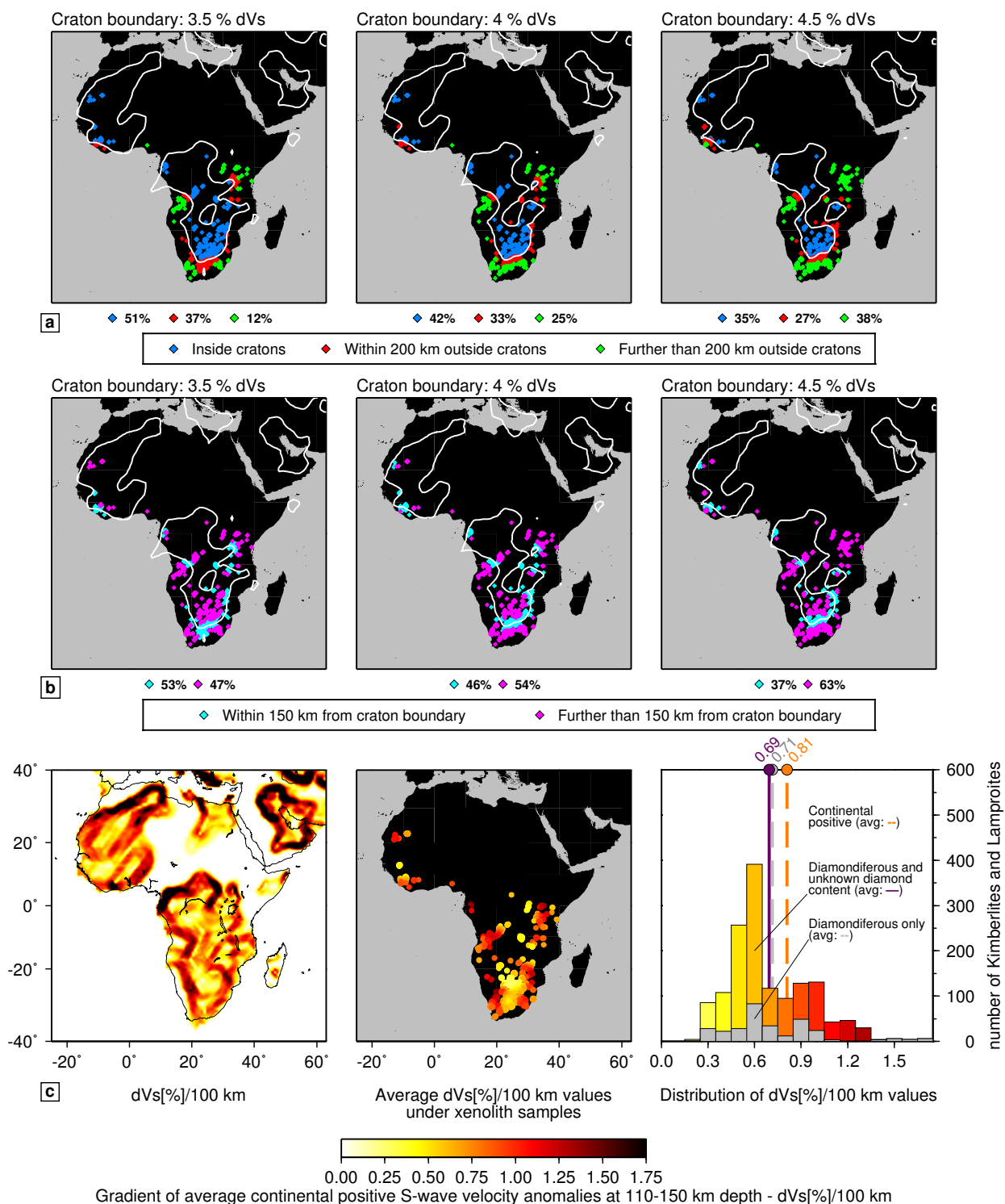


FIGURE 4.10: Caption next page.

FIGURE 4.10: (Previous page) Kimberlites and lamproites in Africa do not tend to be located close to the boundaries of the cratonic mantle lithosphere. This is shown by the kimberlite and lamproite locations relative to cratonic boundaries, defined either using S-wave velocity anomaly contours (a, b), or the gradient of the average S-wave velocity anomaly in the lithosphere (c). In panel a), samples are grouped as: inside craton boundaries (blue); within 200-km wide area just outside the craton boundary (green); further than 200 km outside the craton boundary (red). In panel b), samples are grouped as: within ± 150 km from a craton boundary (300-km wide area around the boundary) (cyan) and further than 150 km from a craton boundary (magenta). For each contour, percentages of each population are indicated at the bottom of the plot. Panels c) show the gradient of the S-wave velocity anomaly averaged over the 110-150 km depth range (for clarity, only areas with positive dVs are included), its values under the kimberlite and lamproite samples, and the histogram of these values for diamondiferous and diamondiferous plus "unknown samples". The average gradient values for all positive-anomaly areas, for diamondiferous plus unknown samples, and for diamondiferous samples only are indicated in orange, purple and grey, respectively.

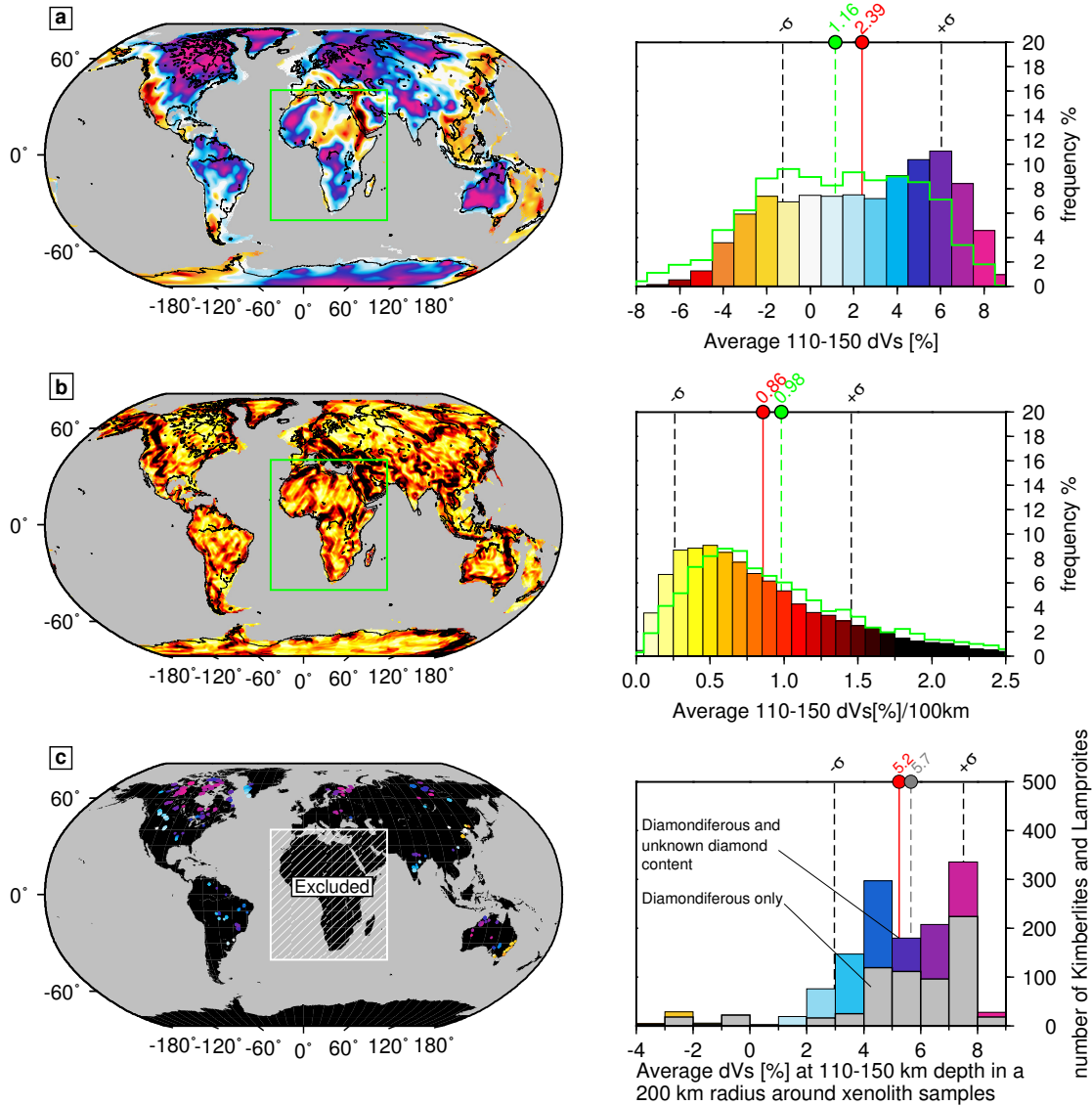
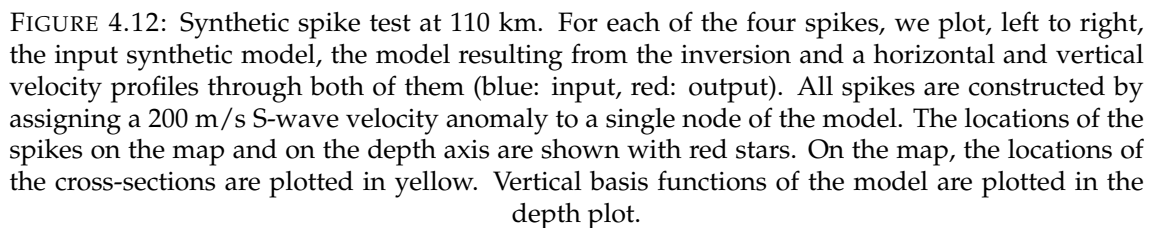


FIGURE 4.11: Continental lithosphere and kimberlites outside Africa. dVs average at 110-150 km depth (a), its gradient for the continental regions (b) and its average value in a radius of 200 km around non-African kimberlite samples only (c). In the histograms a) and b), coloured bins represent the global values, whereas the green bins are for Africa only. Average values for the whole globe and Africa are shown in red and green, respectively. Standard deviations for all continents are shown as black dashed lines. The extent of the "Africa" area is shown as green boundaries on the map views a), b) and as a white striped area in map c). In the histogram c), coloured bins are for kimberlite and lamproites that can be either diamondiferous or of unknown diamond content, whereas the grey bins are for diamondiferous samples only.



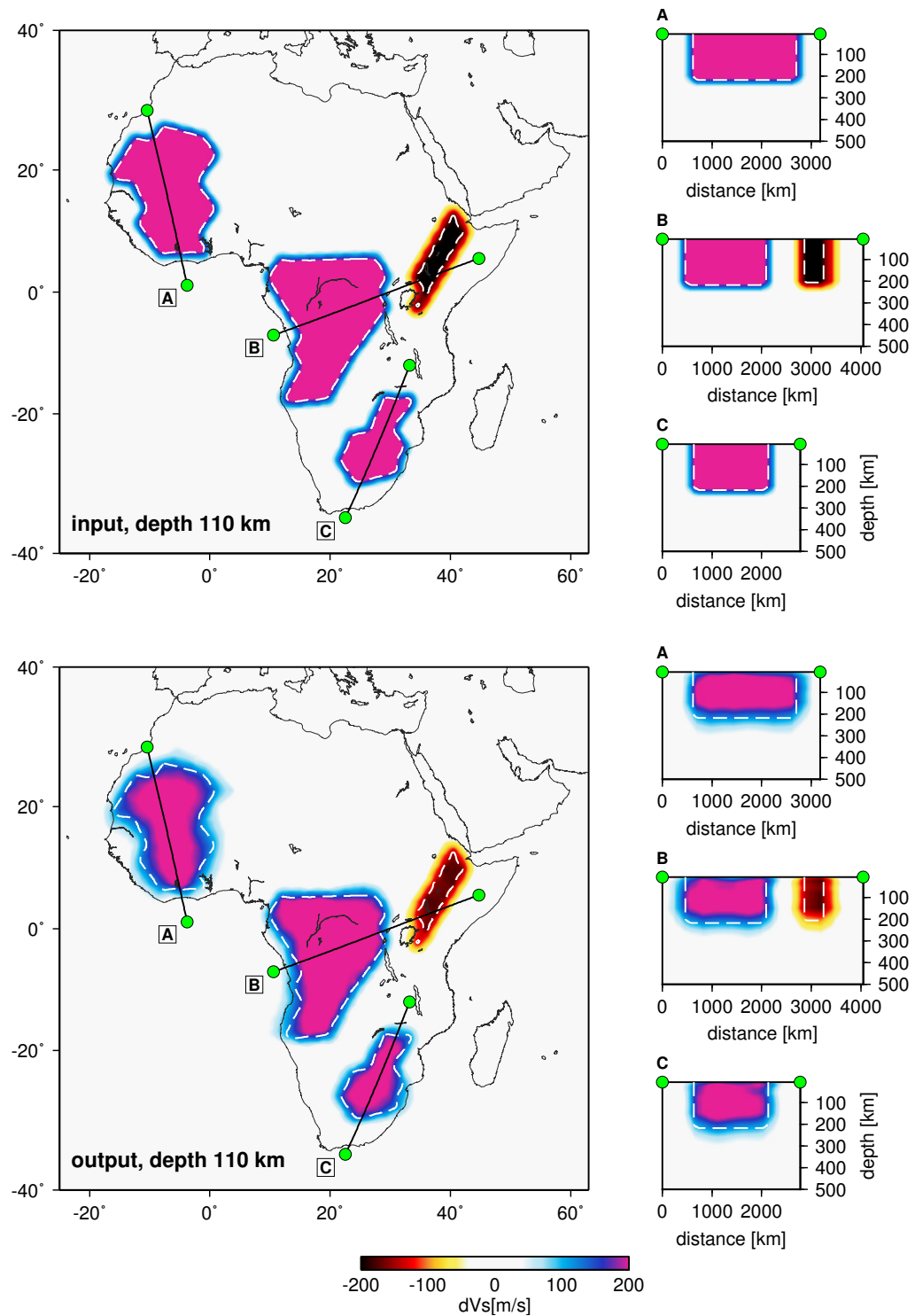


FIGURE 4.14: Synthetic structure test simulating broad anomalies beneath the major cratons and the Ethiopian Rift Valley as seen in previous tomographic models. Top model shows the input synthetic structure, bottom model shows the output of the inversion. Both models are shown both as 110 km depth horizontal slice (left panels) and three vertical cross-sections cutting through the synthetic structures (right panels). Location of the cross-sections is plotted on the horizontal slices. Contours of the synthetic input are plotted as dashed white lines.

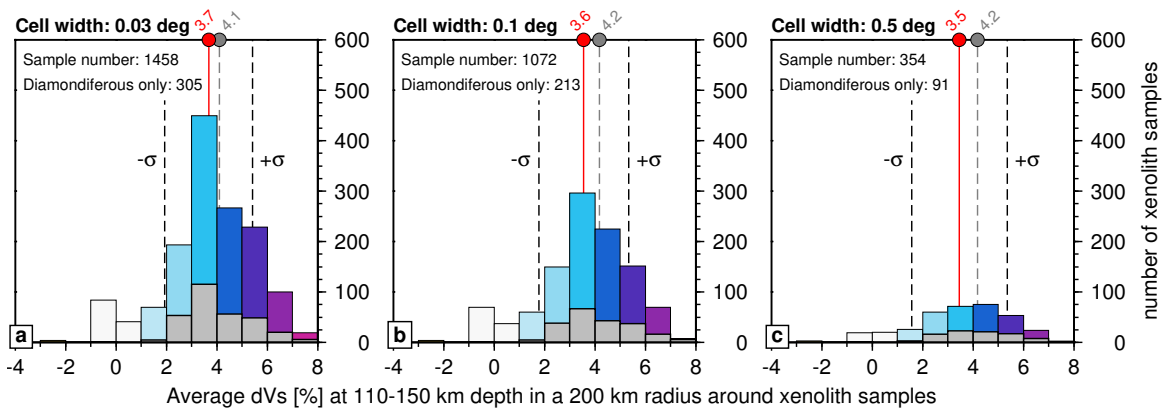


FIGURE 4.15: The effect of resampling the kimberlite and lamproite data on a grid, so as to avoid a sampling bias in the histograms due to numerous samples coming from some of the kimberlite pipes. Results of the resampling with the grid cell sizes of 0.03, 0.1 and 0.5 degrees are shown in the left, middle and right panels, respectively. The histograms are computed using average S-wave velocity anomaly over the 110-150 km depth range, also averaged across a circle of a 200 km radius around each sample. The combined diamondiferous and unknown-content samples are shown as coloured bars, with their average plotted as red line and circle. Diamondiferous samples only are plotted as grey bars, with their average plotted as grey line and circle. The number of samples remaining after the regridding is specified in the top left corner of each panel.

Chapter 5

Iceland Plume rising from beneath Greenland imaged by waveform tomography

5.1 Abstract

Iceland and the entire Northeast Atlantic around it are characterised by abundant intraplate volcanism, anomalously high topography and, in many places, anomalously thick basaltic crust. This has been attributed to the hot Iceland Plume rising from the deep mantle, but the structure and the very existence of the plume are debated. Using seismic waveform tomography with very large datasets, including abundant, recently collected data in the region, we image a large, low-velocity body located beneath central-eastern Greenland in the mantle transition zone (410-660 km depth) and rising, upwards and eastwards, to below Iceland near the surface. We interpret this body as the Iceland plume, ascending from under Greenland and captured by the Mid-Atlantic Ridge. The ascent of the plume beneath the western part of the Northeast Atlantic is consistent with the thinner lithosphere there, as revealed by our tomography, and much more abundant seamounts, compared to the eastern part of the basin.

5.2 Introduction

The Iceland Mantle Plume is believed to have affected the evolution of the entire Northeast Atlantic region, producing huge volumes of thickened crust and the intraplate basalts of the North Atlantic Igneous Province (NAIP) (Saunders et al., 1997), but its structure and its very existence remain unclear. Seismic tomography is sensitive to the compositional and temperature variations of the mantle, and can detect the primarily thermal anomalies expected within mantle plumes (Goes, Cammarano, and Hansen, 2004). While early body-wave studies reported a vertically elongated, low-velocity anomaly under Iceland (Wolfe et al., 1997; Allen et al., 1999; Ritsema, Van Heijst, and Woodhouse, 1999; Allen et al., 2002; Bijwaard and Spakman, 1999; Zhao, 2004), there is little agreement

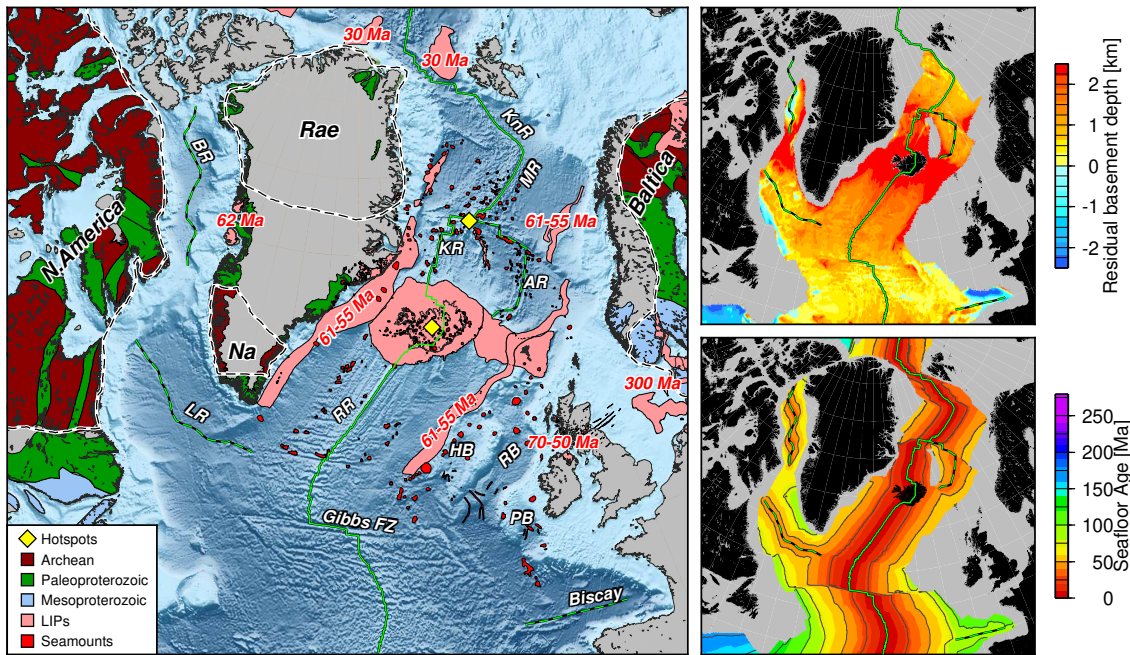


FIGURE 5.1: a) map of the study area. Geological features are plotted on top of topography. LIPs are from (Johansson, Zahirovic, and Müller, 2018), seamounts from (Horn et al., 2017), age of the cratons from (Whitmeyer and Karlstrom, 2007; Bogdanova et al., 2008; Kumar et al., 2007). LIP ages (Wilkinson et al., 2017) are indicated in red, cratons in black. Oceanic features are labelled in white. Ridges: RR, Reykjanes; KR, Kolbeinsey; AR, Aegir; MR, Mohns; KnR, Knipovich; LR, Labrador; BR, Baffin Bay. Basins: HB, Hatton; RB, Rockall; PB, Porcupine. Plate boundaries are plotted in green, inactive ridges as dashed lines. b) Map of residual basement depth (Müller et al., 2008). c) Age of the oceanic lithosphere (Müller et al., 2008).

on how broad the inferred hot anomaly is (Allen et al., 2002; Pilidou et al., 2005; Delorey, Dunn, and Gaherty, 2007; Rickers, Fichtner, and Trampert, 2013), where it rises from (Foulger et al., 2001; Bijwaard and Spakman, 1999; Nolet, Allen, and Zhao, 2007), whether it is a mantle plume at all (Foulger and Pearson, 2001; Keller, Anderson, and Clayton, 2000; Foulger, 2002) and if so, how many plumes there are (Steinberger, 2000; Ganerød et al., 2010; Glišović and Forte, 2019).

Iceland is located on the only portion of the Mid Atlantic Ridge (MAR) above sea level, and is characterised by anomalously thick crust (Bjarnason and Schmeling, 2009), high topography (Torsvik et al., 2015) and flood basalts (Fig. 5.1) with chemical signatures (Rare Earths, minor and major elements and radiogenic isotopes (Schilling, 1973; Elliott, Hawkesworth, and Grönvold, 1991; Fitton et al., 1997; Saunders et al., 1997; Korenaga and Kelemen, 2000; Fitton et al., 2003)) that set them apart from Mid Ocean Ridges, together with NAIP volcanics. Its anomalous features have been long related to a ridge interacting with anomalously hot mantle, often interpreted as a rising mantle plume on the basis of geochemistry (Langmuir et al., 1978; Fitton et al., 1997; Hanan et al., 2000; Fitton et al., 2003), seismic tomography (Wolfe et al., 1997; Bijwaard and Spakman, 1999; French and Romanowicz, 2015) and numerical modelling (Ito, Lin, and Gable,

1996; Barnett-Moore et al., 2017; Glišović and Forte, 2019). However, the widespread, complex distribution of the NAIP volcanics (Saunders et al., 1997; Ganerød et al., 2010; Wilkinson et al., 2017) and related uplift (Jones et al., 2002) is difficult to reconcile with the classic view (Morgan, 1971) of a single, narrow, vertical plume (Steinberger et al., 2018). This, together with the varying depth resolution (Shen et al., 1998; Keller, Anderson, and Clayton, 2000; Foulger et al., 2001; Wolfe et al., 2002) of available tomographic models, lead to alternative proposed models invoking the contribution of other plumes under Jan Mayen (Rickers, Fichtner, and Trampert, 2013) and the Azores (Ganerød et al., 2010; Glišović and Forte, 2019), or an upper mantle, non-plume origin of the thermal anomaly under Iceland (Korenaga and Kelemen, 2000; Foulger and Pearson, 2001; Foulger, 2002; Foulger and Anderson, 2005).

Recent tomographic images (Amaru, 2007; Rickers, Fichtner, and Trampert, 2013; Hosseini, 2016) revealed new, more complex—and sometimes conflicting—structures in the low velocities under Iceland that depart from the classic plume view. In the upper mantle, branches of the Iceland Plume have been proposed to depart far from Iceland following viscous fingering (Schoonman, White, and Pritchard, 2017), reaching southern Norway and the Celtic Sea (Rickers, Fichtner, and Trampert, 2013; Arrowsmith et al., 2005) at present day, and towards Baffin Bay during the Paleogene (Steinberger et al., 2018). In the deep mantle, the plume has been proposed to tilt to the south-east (Rickers, Fichtner, and Trampert, 2013; French and Romanowicz, 2015), possibly linking Iceland to the African Large Low Shear Velocity Province (LLSVP) at the core-mantle boundary (Steinberger et al., 2015; Yuan and Romanowicz, 2017; Barnett-Moore et al., 2017). In spite of recent advances in seismic station coverage, the plume structure in the upper mantle and its interaction with the MAR remain unclear. Anomalies in a number of tomographic images (Rickers, Fichtner, and Trampert, 2013; Arrowsmith et al., 2005; Krischer et al., 2018) have been interpreted as consistent with a plume under Iceland and the Eurasian-Plate part of the Northeast Atlantic. Yet, the distribution of recent intraplate volcanism (Gaina et al., 2016; Horni et al., 2017) indicates that a great majority of intraplate seamounts have been forming, instead, on the Greenland plate.

5.3 Tomography of the Northeast Atlantic

In order to investigate the Iceland Plume and its influence on the surrounding region, we compute NAT2019 (Fig. 5.2, a new, azimuthally anisotropic, S-wave tomographic model of the lithosphere, upper mantle and transition zone (TZ) under the North Atlantic region using waveform inversion (Lebedev et al., 2005; Lebedev and Hilst, 2008). Our model combines over 1.2 million global and regional teleseismic waveform fits (Lebedev et al., 2005) from over 27000 events and 6000 stations. For the fundamental mode

surface waves, global measurements provide the long periods constraining the deep upper mantle, while regional ones carry the higher-frequency information on the fine-scale details of the crust, mantle lithosphere and shallow asthenosphere, respectively. S and multiple S waves yield structural information on the heterogeneity in the deep upper mantle and transition zone. Compared to previous models by our group (Schaeffer and Lebedev, 2013; Lebedev et al., 2018), NAT2019 is constrained by a significant amount of new, regional data in the study area and has a resolution that varies in concert with data sampling, implemented using regularisation coefficients that scale in 3D with the variations of data sensitivity (section 5.5.2). Compared to models computed using spectral element, waveform simulations (Rickers, Fichtner, and Trampert, 2013; Krischer et al., 2018), our inversion, relying on the JWKB approximation (Dahlen and Tromp, 1998), utilises a dataset that is larger by an order of magnitude. The large amount of data provides significant additional ray-crossing that is key for the imaging of the North Atlantic Ocean and Greenland, where, compared to the well-sampled continents, data are scarcer.

At long wavelengths, NAT2019 agrees with previous global and regional models (Rickers, Fichtner, and Trampert, 2013; Schaeffer and Lebedev, 2013; French and Romanowicz, 2015; Amaru, 2007; Lebedev et al., 2018), showing low velocities beneath the MAR and high velocities beneath the North American and Greenland Cratons and the Baltic Shield (Fig. 5.2, S3, S4). In Greenland, our model retrieves and improves on the sharpness of the boundaries of the Rae and North Atlantic cratons, recently identified as separate (Lebedev et al., 2018). Under the ocean, compared to recent models (Rickers, Fichtner, and Trampert, 2013; Krischer et al., 2018), we identify more heterogeneous structure along the MAR close to Iceland. Most importantly, our model images a connection between the shallow velocity minima in Iceland and along the MAR to the ones in the TZ.

In the upper mantle, seismic velocities under the Reykjanes, Kolbeinsey and Mohns ridges are anomalously low. On average, Mid-Ocean Ridges on the globe display S-wave anomalies of -3 to -5% close to the ridge axis in the shallow upper mantle (Fig. S7). The velocities on the MAR north and south of Iceland are comparatively much lower (approximately -8% dVs and lower) and extend over broader regions. Such velocities are indicative of decompression melting over broad areas, inconsistent with a slow-spreading MAR (Niu and Hékinian, 1997), but consistent with the presence of an upwelling of hotter mantle. Under the V-shaped Reykjanes Ridge, the very low velocities locate symmetrically around the ridge axis and are confined to the upper mantle, rapidly fading at depths greater than 330 km. Under the Kolbeinsey and Mohns Ridge and next to Jan Mayen, velocity minima are located systematically west of the MAR in the entire upper mantle depth range. The extinct Aegir Ridge is not underlain by low-velocity anomalies. On the other, western side of Greenland, low-velocity anomalies are present deeper than 200 km under the extinct Baffin Bay- but not under the Labrador Ridge. In the TZ, the

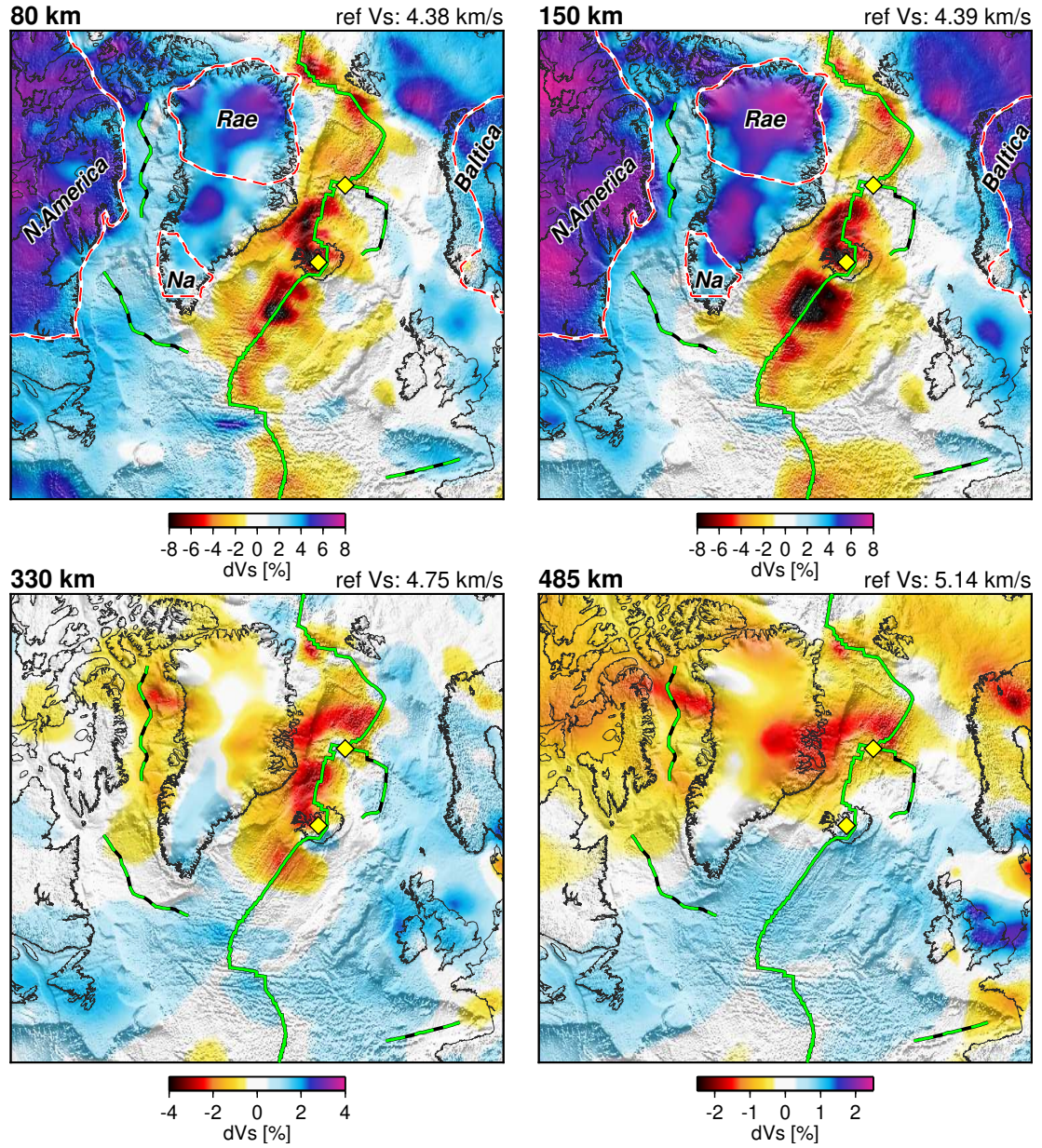


FIGURE 5.2: S-wave velocity anomaly of NAT2019 tomographic model at 80, 150, 330 and 485 km depth. Values in % from the reference, indicated on the top right of each panel. Plate boundaries are plotted in green, inactive ridges as dashed lines.

lowest velocities locate under central eastern Greenland, from where they connect to the ones both under the North Atlantic and Baffin Bay in the upper mantle.

Our images indicate the presence of a large volume of anomalously hot mantle, rooted under eastern Greenland in the TZ, that rises to the upper mantle and is deflected both west and east by the lithospheric keels of the North Atlantic and Rae Cratons, with the larger, eastern branch directly connecting to the anomalously low velocities under Iceland and the MAR (Fig. 5.4).

North of Iceland, we observe very low velocities throughout the upper mantle (reds and blacks in Fig. 5.2) that, at depths greater than 200 km, locate mostly west of the MAR. By averaging NAT2019 over oceanic lithosphere of the same age for the Greenland and European portion of the North Atlantic, we can see that, on average, the lithosphere and asthenosphere under the Greenland plate is slower for most ages (Fig. 5.4). By subtracting the global average, "normal cooling" signal from local averages, we can see how the Greenland and European oceanic plates are markedly hotter and colder than the average cooling lithosphere, respectively. Under the Greenland plate, the whole mantle is anomalously slow from the lithosphere down to the TZ, implying that the large low-velocity body we observe today has affected the cooling of the oceanic lithosphere in the whole western North Atlantic. Thinner lithosphere and hotter asthenosphere would cause a widespread decompression melting below the lithosphere-asthenosphere boundary (LAB), and indeed recent compilations of intraplate basalts show a larger amount of seamounts on the Greenland Plate (Horní et al., 2017; Gaina et al., 2016). By counting only the seamounts on the oceanic lithosphere, 205 locate under the Greenland plate, compared with only 104 under the European plate (Fig. 5.4). This suggests that the distribution of intraplate volcanism is largely dictated by the upwelling of hot mantle we observe.

South of Iceland, the V-shaped ridges formed by the Reykjanes Ridge shows a deviation from the asymmetry we image to the north, being the only portion of the MAR displaying symmetric low-velocity anomalies (Fig. 5.2). The ridge has long been proposed as an area of plume-ridge interaction, with a hot mantle wedge flowing horizontally from Iceland (Schilling, 1973; Delorey, Dunn, and Gaherty, 2007; Benediktsdóttir et al., 2012). Our tomographic images show that the Reykjanes Ridge represents the southernmost tip of the large hot mantle anomaly, following its SW tilt. The tilt of the eastern branch in the shallow upper mantle thus appears to follow the MAR, possibly indicating that the hot mantle, prevented from flowing south-east by the ridge, flows instead southwards along its axis, channelling along its thinner lithosphere (Ito, Lin, and Gable, 1996). Azimuthal anisotropy, much weaker under the Reykjanes Ridge than anywhere else on the MAR (Fig. S6), supports this hypothesis, suggesting the presence of contrasting directions in the area, i.e.: ridge-perpendicular flow from the spreading MAR, ridge-parallel flow from the channelled hot mantle and possibly a small vertical component from the rising hot mantle that would map in radial anisotropy, unaccounted for in our model.

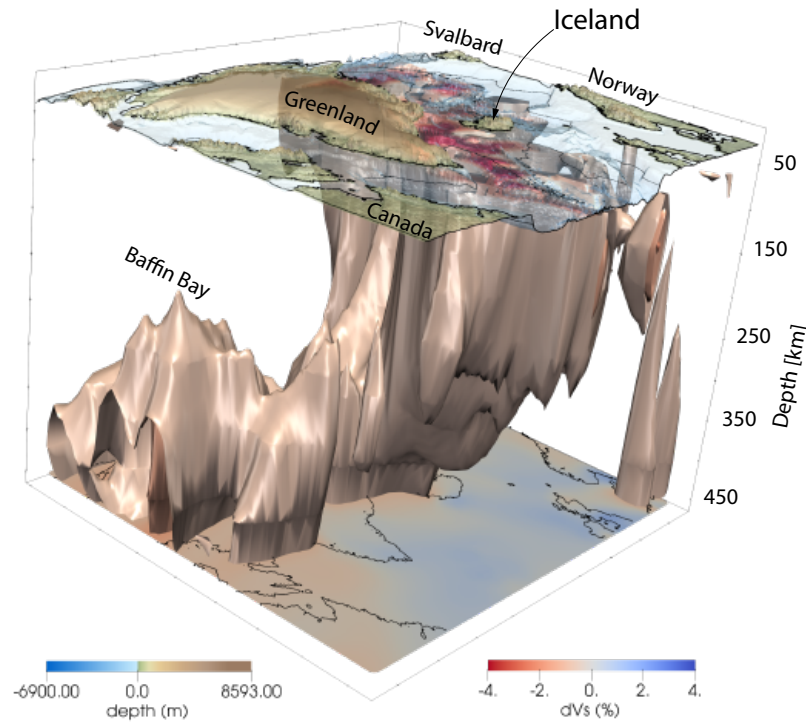


FIGURE 5.3: 3D Plot of the Iceland Plume as -0.75% velocity anomaly contour between 56 and 485 km depth, vertically exaggerated. Topography is superimposed as transparency; the -2000 contour is shown as black line. The plot is underlain by the 485 km depth slice through the model, with coastlines on top.

5.4 Discussion

The imaged complex, large scale low-velocity body connects the anomalously low velocities close to Iceland and the MAR to the ones deep in the TZ, suggesting an upwards continuity that we interpret as the seismic expression of the Iceland Plume. While its shape differs from the classic view of a narrow, vertical mantle plume, its complexities reconcile many previously proposed and apparently contrasting observations.

Earlier (Wolfe et al., 1997; Allen et al., 1999; Ritsema, Van Heijst, and Woodhouse, 1999; Allen et al., 2002; Bijwaard and Spakman, 1999; Zhao, 2004) and recent (Rickers, Fichtner, and Trampert, 2013; Schaeffer and Lebedev, 2013; Krischer et al., 2018; Rychert, Harmon, and Armitage, 2018) seismic studies sample low velocities in the upper mantle under Iceland, interpreting it as a plume. Other studies however opposed this hypothesis arguing that the anomaly is confined to the upper mantle (Foulger et al., 2001; Foulger

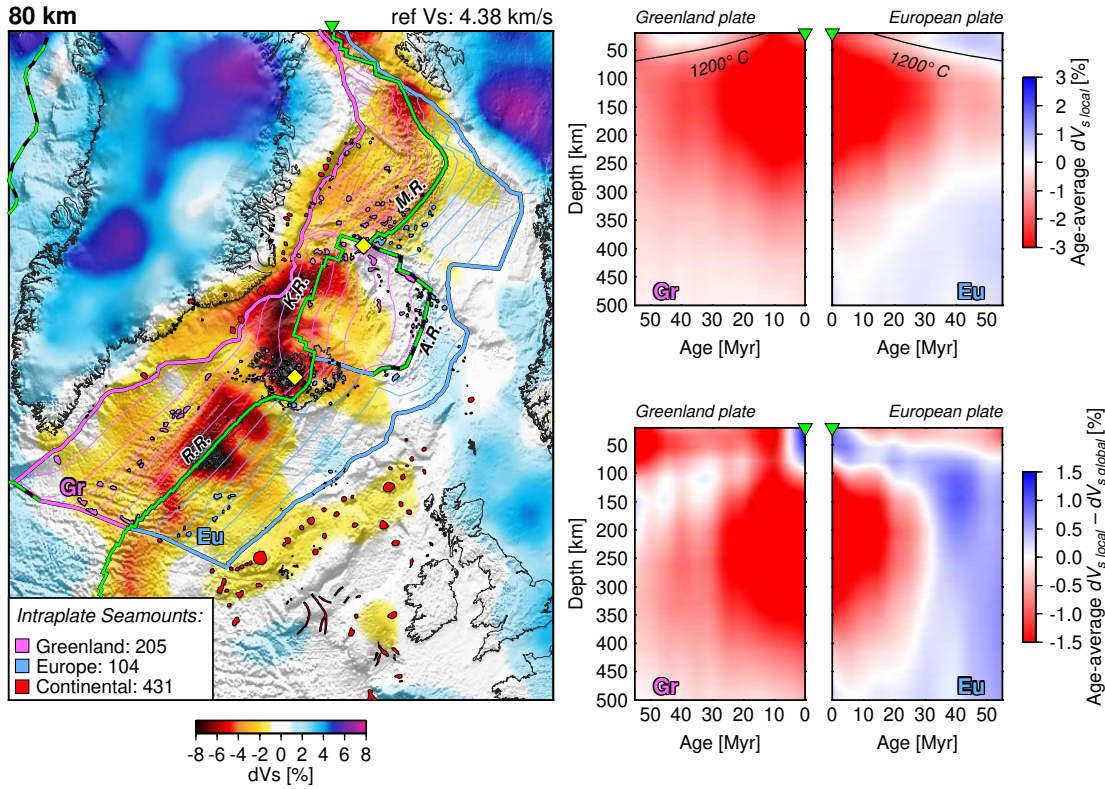


FIGURE 5.4: Evidence for differences between the western and eastern North Atlantic. A slice at 80 km depth through NAT2019 is shown in the main panel. Lithospheric age contours every 25 Ma and boundaries of the Greenland and European oceanic plates are plotted in pink and pale blue, respectively. Past (dashed) and present (solid) plate boundaries are plotted in green and marked as green triangles at plot boundaries. Intraplate oceanic seamounts from (Horní et al., 2017) are coloured based on the plate they locate on. Continental Seamounts are plotted in red. Top right panels show the age-average S-wave velocity anomaly under the Greenland and European oceanic plates, with 1200 deg isotherms from (Becker et al., 2014). Bottom right panels show the difference between the local age-averages and the global average.

and Pearson, 2001). Our model indicates that low velocities under Iceland and the Reykjanes Ridge indeed represent the tip of the plume, but which is also confined in the area to the upper mantle, flowing along the ridge axis (Ito, Lin, and Gable, 1996).

In the deeper upper mantle and TZ, our images are consistent with the ones provided by P-wave tomography (Amaru, 2007) (Fig. 5.8). Earlier models however (Bijwaard and Spakman, 1999) indicate Iceland as the plume centre in the upper mantle in a somewhat arbitrary location among very widespread low-velocity anomalies, which then connect to the lower mantle plume to the south of Iceland (Bijwaard and Spakman, 1999; Shen et al., 1998). The plume root we image in the TZ is more than 900 km NW of Iceland, consistently with more recent body-wave tomography (Amaru, 2007) (Fig. 5.8b), and in an area where the TZ is reportedly thinner (Kraft et al., 2018). This indicates that if the plume continues south of Iceland in the lower mantle, a shift of approximately 1000 km is required at the bottom of the TZ. While previous studies (Shen et al., 2002; Bijwaard and Spakman, 1999) already proposed a—smaller—tilt of the plume and recent tomography

in the deep mantle pictures a variously tilted conduit (Hosseini, 2016), such a large lateral shift at the bottom of the TZ remains unexplained.

Our model's complex plume structure well matches the complex distribution of the NAIP volcanism: by supposing the plume was of similar size at 60-50 Ma, we can explain the distribution of the quasi-simultaneous volcanic provinces in western and Eastern Greenland (Ganerød et al., 2010; Wilkinson et al., 2017) in terms of flow equally split by the keels of the Greenland cratons, centred over the plume root (O'Neill, Müller, and Steinberger, 2005). Similarly, Greenland slowly moving north-west over a very wide plume can explain the progressively waning LIP volcanism in western Greenland after 60 Ma (Wilkinson et al., 2017) by an increasingly larger portion of the plume being redirected instead east by the cratonic keel. Under central Greenland, where the lithosphere is now thinnest (Fig. 5.2, S5), the lack of records of the past presence of thick cratons prevents us from determining whether the lithosphere was thin before the plume onset or was thinned by the plume itself, as proposed by a recent study (Steinberger et al., 2018). While this information will be important for understanding of the past mantle flow, in both cases this corridor very likely contributed to localise the flow under the areas of most intense recorded volcanism during the Paleogene.

From our new inferences, much of the plume-related hot mantle is confined to the Greenland plate. On the continental European margin, previous studies (Arrowsmith et al., 2005; Rickers, Fichtner, and Trampert, 2013) identified two fingers of hot mantle departing from Iceland and reaching under the Irish Sea and Southern Norway, well correlating with episodes of past volcanism and uplift of the North Atlantic- and British Tertiary Igneous Provinces (NAIP and BTIP) mostly during the Paleocene (Jones et al., 2002; Rickers, Fichtner, and Trampert, 2013; Cogné et al., 2016; Łuszczak, Persano, and Stuart, 2018). Under these areas, our model shows regions of zero velocity anomaly west of Britain and under southern Norway. This indicates that there, the lithosphere is thinner and hotter than under the North Sea Basin and Great Britain which show moderately positive δV_s at 56-200 km depth. Although differences in inversion techniques, coverage, inversion parameters and reference model may account for the variations between tomographic models (Figs. 5.7, 5.8), the lack of strong, low velocities we observe at present day indicates a moderately hot mantle in the area that differs from the strong anomalies recorded under the Greenland plate and MAR. Correlation between plume-related structures and more recent, Neogene uplift that would provide evidence for recent plume activity in the area is controversial (Koptev et al., 2017; Łuszczak, Persano, and Stuart, 2018). But how was the plume able to flow southwards and create the BTIP during the Paleocene, while being at present day confined west of the MAR? Before the onset of seafloor spreading, the continental margin was closed, putting in contact lithospheres of comparable thickness (Steinberger et al., 2018). Hot asthenosphere may then have

flown across the closed continental margin and possibly following pre-existing structures (Koptev et al., 2017), eroding the continental lithosphere and hence causing the widespread decompression melting, volcanism and uplift in the Paleocene. With the onset of seafloor spreading, 56 Myr ago, the presence of a corridor of much thinner, oceanic lithosphere and the upwelling of the MAR (Morgan, 1978; Sleep, 1997) would have gradually imposed a strong barrier against the SE flow of the mantle. The much shallower oceanic LAB and the convection cells of the spreading ridge could possibly have captured the plume (Ito, Lin, and Gable, 1996), leading to the asymmetrically hot mantle we observe today on the two sides of the MAR.

5.5 Supplementary material to chapter 5

5.5.1 Method

The model NAT2019 is computed from a database of over 6 million vertical component seismic waveforms. These are fitted using the Automated Multimode Inversion (AMI) (Lebedev et al., 2005) of surface-, S- and multiple S-wave forms, providing over 1.2 million waveform fits. The model construction is global, but the data coverage in the study area has been maximised, obtaining all freely available data in the North Atlantic region. Earthquakes are taken globally from the Global Centroid-moment-tensor (GCMT) catalogue (Ekström, Nettles, and Dziewonski, 2012) since 1994, with a distance-magnitude varying threshold (Schaeffer and Lebedev, 2013). As a result, each node of the final tomographic model is constrained by at least 3807 fits globally and by over 25000 in the study area (Fig. 5.5). Under the most well-sampled areas (e.g. USA and western Europe), the number of waveform fits sampling each node can be as high as 226521.

After download, we quality control the waveforms against clipping and gaps, ensuring that the whole waveform—and especially the surface wave train—is fully recorded; after that, each trace is response corrected and downsampled to 1 Hz.

The successfully retrieved traces undergo a three-step inversion procedure. At first, we invert the seismogram using AMI (Lebedev et al., 2005). AMI computes synthetic seismograms by modal summation and fits the surface, S- and multiple S-waves to the observed ones within a complex set of weighted time-frequency windows. By minimising the misfit in the time domain, AMI produces for each fit a set of linear equations with uncorrelated uncertainties (Nolet, 1990) describing the average S- and P-wave velocity variations from a 3D reference model within approximate sensitivity areas (Lebedev et al., 2005). In the second step, all equations are inverted jointly for the distribution of P- and S-wave velocities and 2Ψ S-wave azimuthal anisotropy in 3D (Nolet, 1990; Lebedev and Hilst, 2008) using LSQR (Paige and Saunders, 1982). We parametrise the tomographic model using a very dense triangular grid (Wang and Dahlen, 1995a) of approximate 120 km knot spacing and 18 and 10 triangular basis functions for S- and P-wave velocities,

respectively (S-wave velocities: 7, 20, 36, 56, 80, 110, 150, 200, 260, 330, 410-, 410+, 485, 585, 660-, 660+, 809 and 1007 km; P-wave velocities: 7, 20, 36, 60, 90, 150, 240, 350, 485 and 585 km). Perturbations of the model parameters are computed with respect to the same 3D model used in AMI, which is a combination of CRUST2 (Bassin, Laske, and G., 2000) with added topography and our own 1D mantle average (Lebedev and Hilst, 2008). In the third and final step of our inversion procedure, we identify and remove outliers exploiting our large data redundancy. Outliers are mostly related to errors in the source location and station timing (Legendre et al., 2012), and figure as highly anomalous values that the tomographic inversion focusses at the location of the problematic station or event to minimise data misfit. In order to identify these outliers, we first compute an initial tomographic model m_i , which will be polluted by such data errors. We then compute the synthetic data d_i by multiplying the initial model by the sensitivity matrix A . Finally we compare the original d_o s and synthetic d_i data and discard the ones with the largest misfit.

5.5.2 Multi-resolution regularisation

The linear inversion is regularised by horizontal and vertical smoothing and a slight norm damping. Typically, a single coefficient determines the amount of regularisation applied, which will affect variously sampled nodes differently. A choice of norm damping suitable for imaging a well-sampled lithosphere will result in overdamping of deeper, less sampled nodes, lowering the amplitudes of the anomalies; conversely, the less sampled—and thus more susceptible to errors—greater depths would benefit from more smoothing, undesired in the well sampled lithosphere. This issue is prominent in global tomographic imaging due to the very strong variations in 3D sampling (Fig. 5.6).

In order to account for differently sampled nodes, we scale regularisation with data sampling in 3D. This is implemented by scaling the regularisation coefficients with the sums of the columns of the sensitivity matrix. Column sums are a-dimensional quantities that contain information on the number of sensitivity areas sampling the node, the sensitivity kernel weights and the path-similarity weights, making them good indicators of data sensitivity. We scale norm damping proportionally to the column sums, while smoothing is scaled inversely. This way, well sampled nodes will be more damped and less smoothed, while poorly sampled ones will be less damped but smoothed more. The regularisation coefficients for both smoothing and norm damping are thus scaled as follows:

$$f(i) = f_0 * f_s(i) \quad (5.1)$$

$$f_s(i) = \frac{cs(i)/cs_{max} + \epsilon}{\epsilon + 1} \quad (5.2)$$

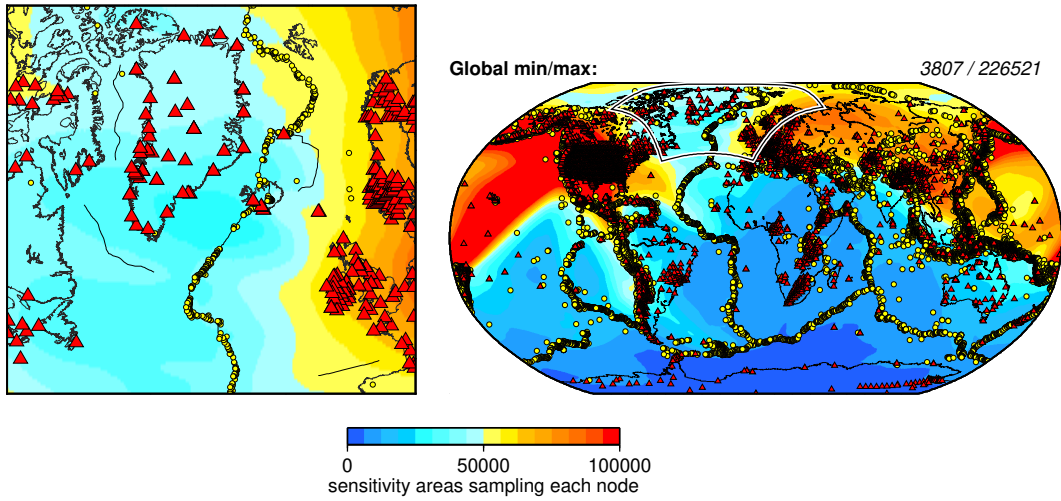


FIGURE 5.5: Data coverage for the NAT2019 tomographic model, zoomed in the study area (left) and globally (right), represented by the number of approximate sensitivity areas sampling each node. The location of the study area on the globe is shown as black and white lines. The shape—and thus the counting—of the sensitivity areas is depth-invariant. Stations are shown as red triangles, events as yellow circles. Past and Present-day plate boundaries area shown in black.

With $f(i)$ the regularisation factor for the i – th node of the model grid, f_0 the initial, unscaled regularisation factor, $f_s(i)$ the scaling factor and $cs(i)$ the column sums. ϵ determines how much the scaling is effective, and is chosen empirically. For $\epsilon = 0$, the scaling factor is equal to the normalised column sums, yielding the largest scaling. Increasing ϵ will progressively reduce the effect of the column sum scaling in a non-linear way (Fig. 5.13).

While our intention is to make the regularisation take sampling into account, it is advisable to choose non-zero ϵ values for norm-damp scaling to tame the effect of data errors and inconsistencies in the model. For smoothing scaling, the same rule applies to prevent over-smoothing of greater depths resulting from the very large values of $1/f_s(i)$ (Fig. 5.13).

The effects of different ϵ values in 3D is shown in Fig. 5.14 at 56, 150 and 485 km depth. Lower ϵ values allow for greater variation in the scaling factor in 3D, but also induce generally lower (proportional) or higher (inversely proportional) average scaling, indicating that the choice of absolute regularisation coefficients and their scaling are intertwined.

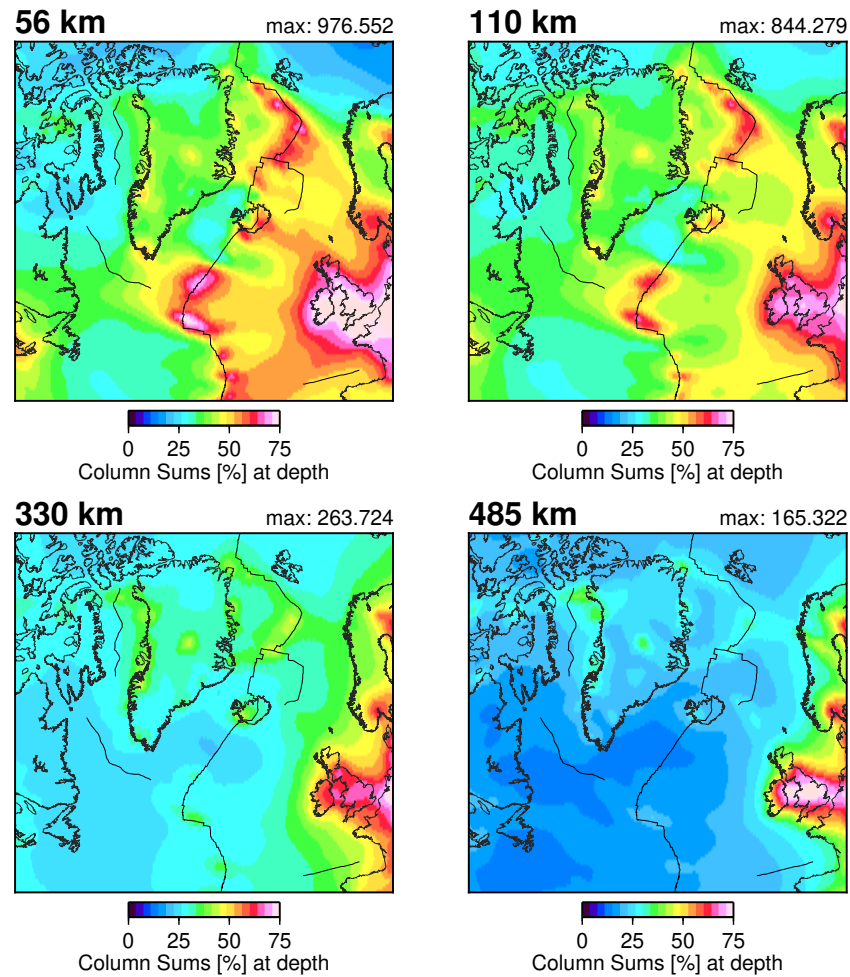


FIGURE 5.6: Model sensitivity for NAT2019 at four node depths. The sensitivity is represented by the sums of the columns of the sensitivity matrix, in percentage from the absolute max. The column sums include the effects of kernel weights, path weights and the number of paths sampling each node. Past and Present-day plate boundaries area shown in black.

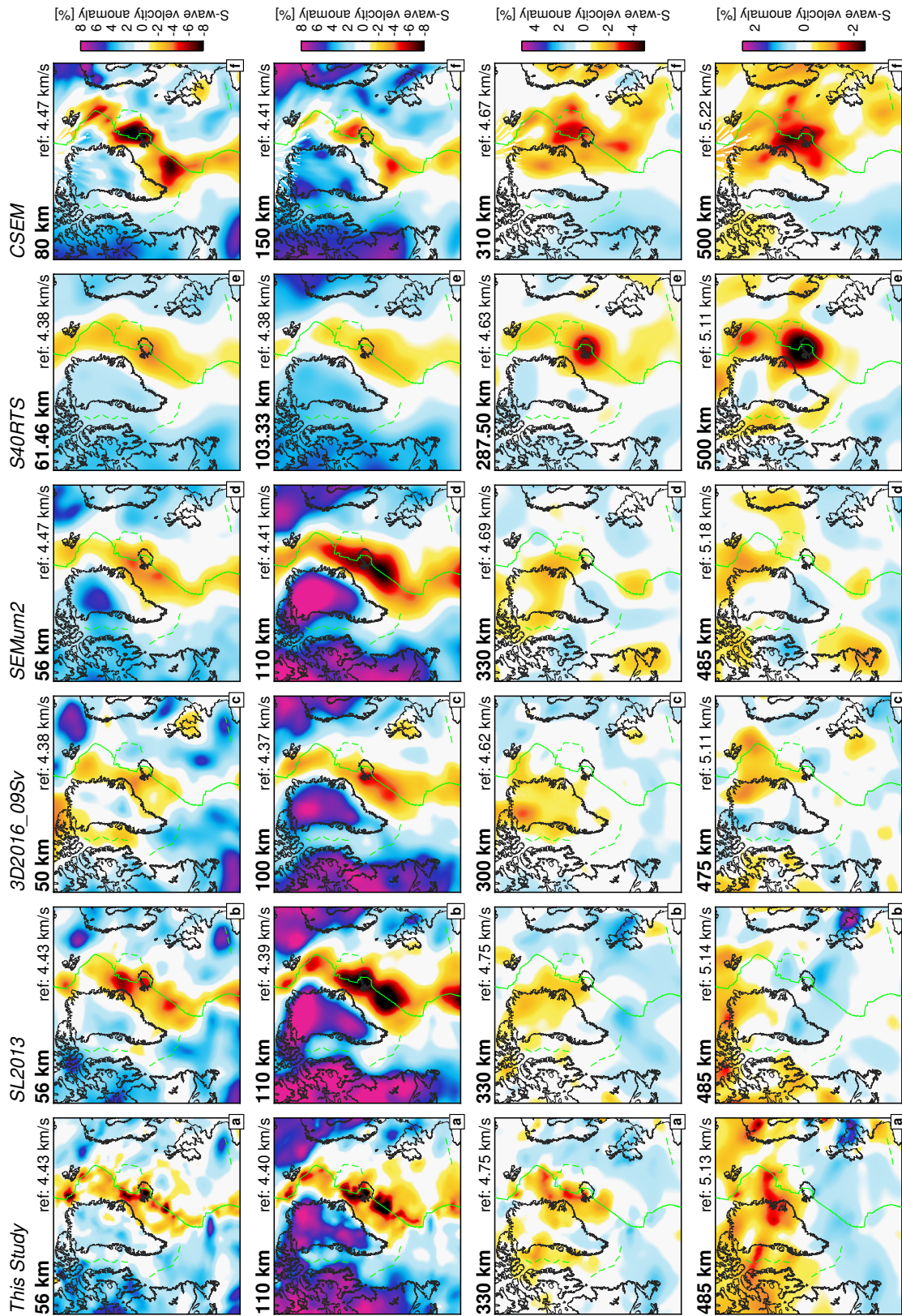


FIGURE 5.7: (Caption next page.)

FIGURE 5.7: Comparison between a) NAT2019 and other S-wave tomographic models in the shallow and deep lithosphere, asthenosphere and transition zone: b) SL2013 (Schaeffer and Lebedev, 2013), c) 3D2016Sv (Debaille, Dubuffet, and Durand, 2016), d) SEMum2 (French, Lekic, and Romanowicz, 2013), e) S40RTS (Ritsema et al., 2011), CSEM (Fichtner et al., 2018). Each model is plotted at the depth node closest to 50, 100, 300 and 485 km depth and is plotted with respect to the average at depth. Reference velocity is plotted on the top right of each panel. Past (dashed) and present (continuous) plate boundaries are plotted in green.

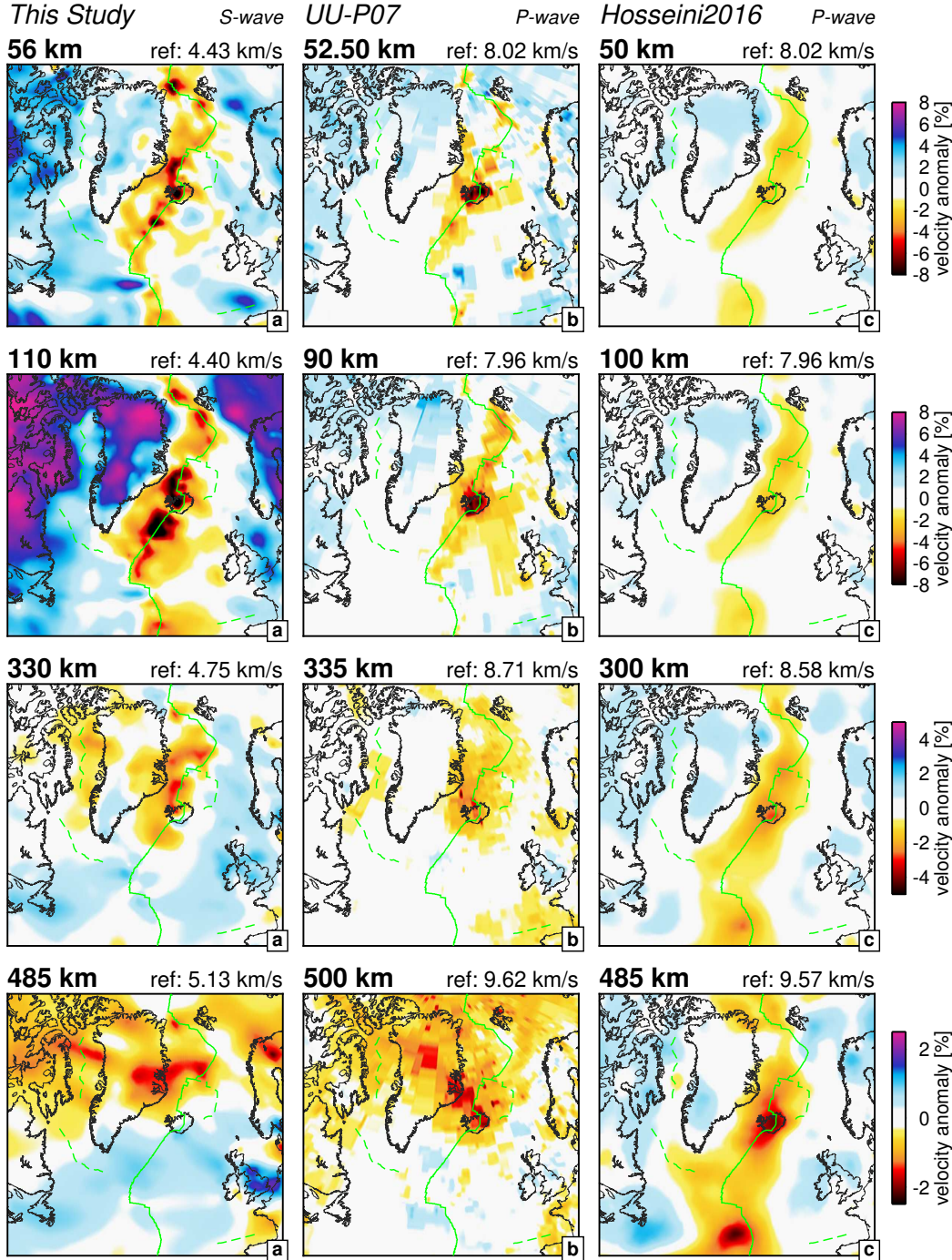


FIGURE 5.8: Comparison between a) NAT2019 and other P-wave tomographic models in the shallow and deep lithosphere, asthenosphere and transition zone: b) UU-P07 (Amaru, 2007) and c) Hosseini2016 (Hosseini, 2016). Each model is plotted at the depth node closest to 50, 100, 300 and 485 km depth. P-wave models are plotted with respect to AK135 (Kennett, Engdahl, and Buland, 1995), while NAT2019 is plotted with respect to the average at depth. Reference velocity is plotted on the top right of each panel. Past (dashed) and present (continuous) plate boundaries are plotted in green.

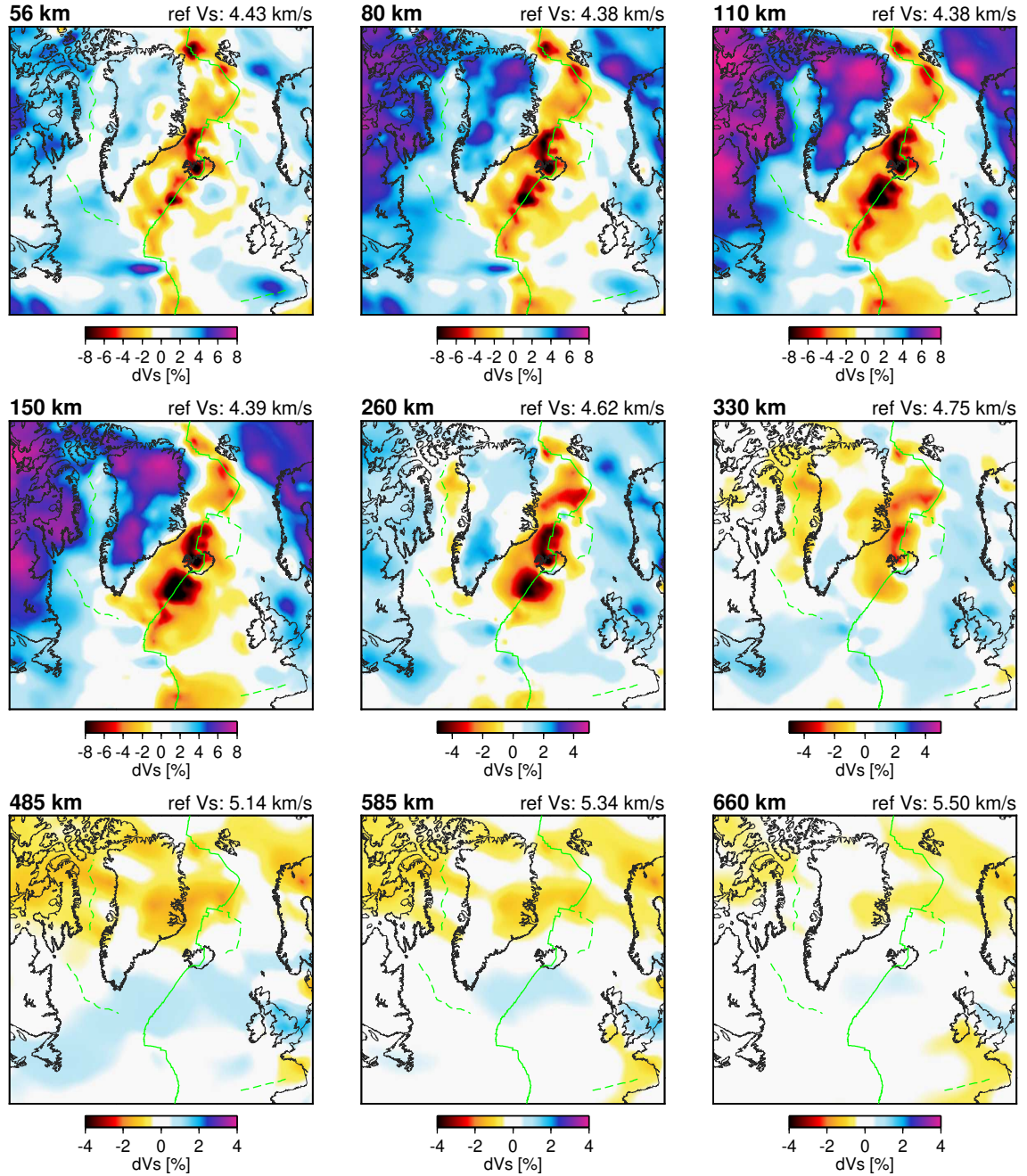


FIGURE 5.9: Vertical component, S-wave velocity anomaly slices through NAT2019 at 56, 80, 110, 150, 200, 260, 330, 485, 585 and 660 km depth. Reference velocity is plotted on the top right of each panel. Past (dashed) and present (continuous) plate boundaries are plotted in green.

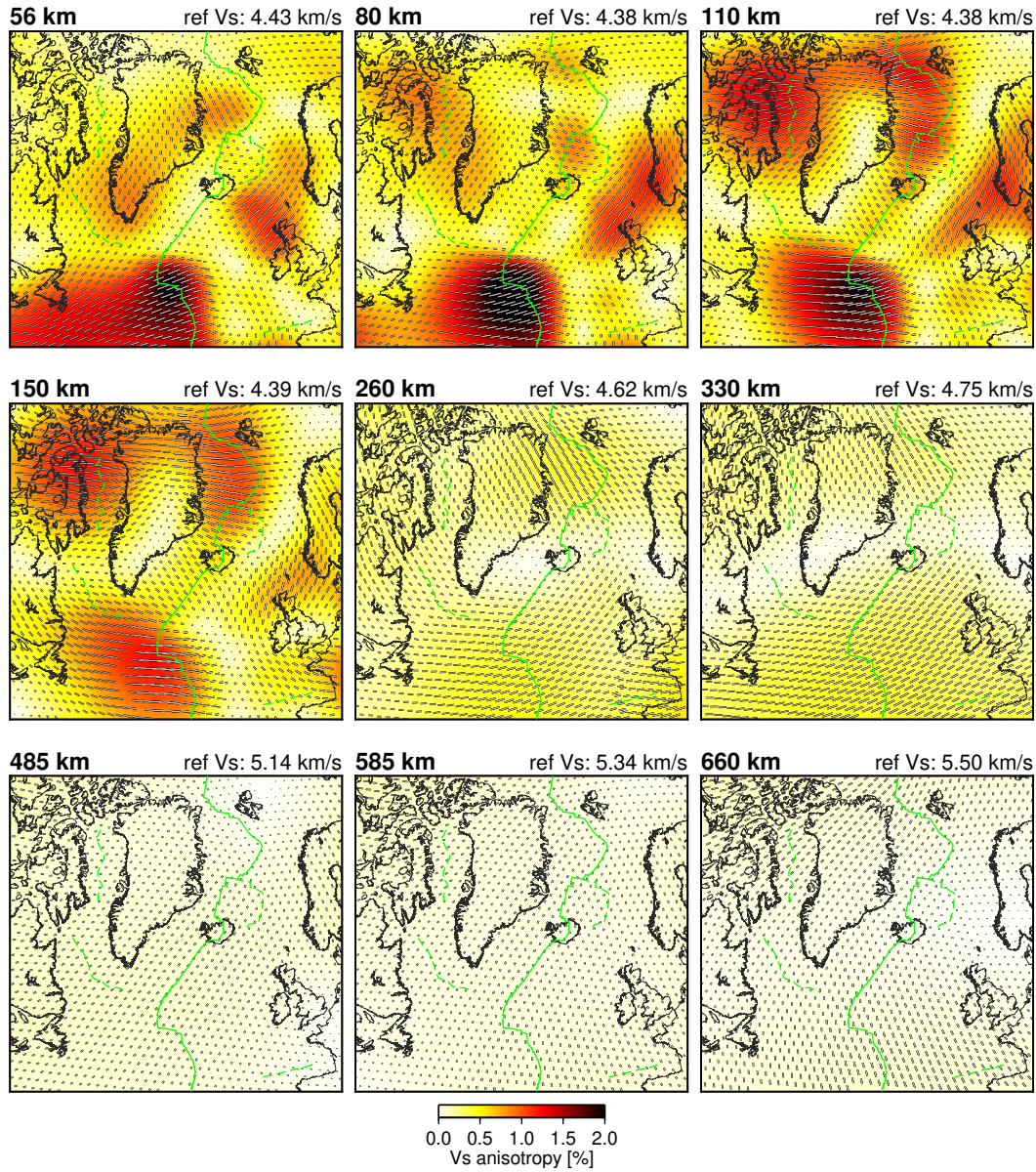


FIGURE 5.10: 2 Ψ S-wave azimuthal anisotropy from NAT2019 at 56, 80, 110, 150, 200, 260, 330, 485, 585 and 660 km depth. Reference velocity is plotted on the top right of each panel. Grey sticks mark the orientation and the amplitude of the fast axis. Past (dashed) and present (continuous) plate boundaries are plotted in green.

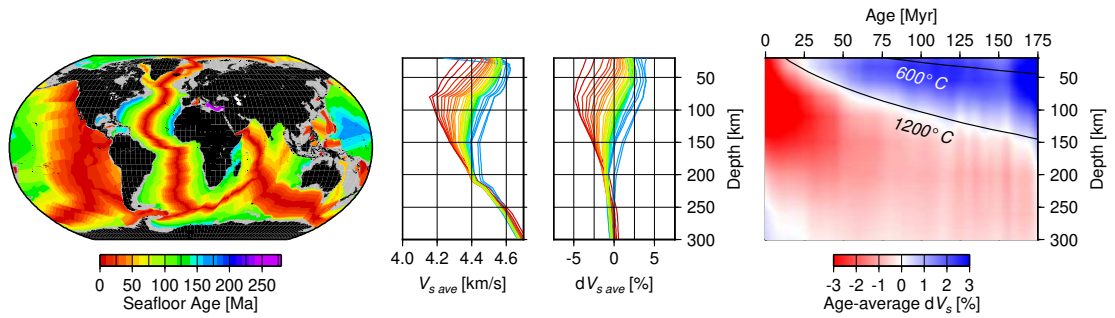


FIGURE 5.11: Global oceanic age-average measurements from NAT2019. Left: global ages of the oceanic lithosphere, from (Müller et al., 2008). Middle: V_s and V_s anomaly profiles from NAT2019 averaged under each lithospheric age bin in a 5 Myr interval. Right: age-average cross-section obtained combining the V_s anomaly profiles in the middle panel along age. The 600 and 1200 degrees isotherms from (Becker et al., 2014) are plotted on top as black lines.

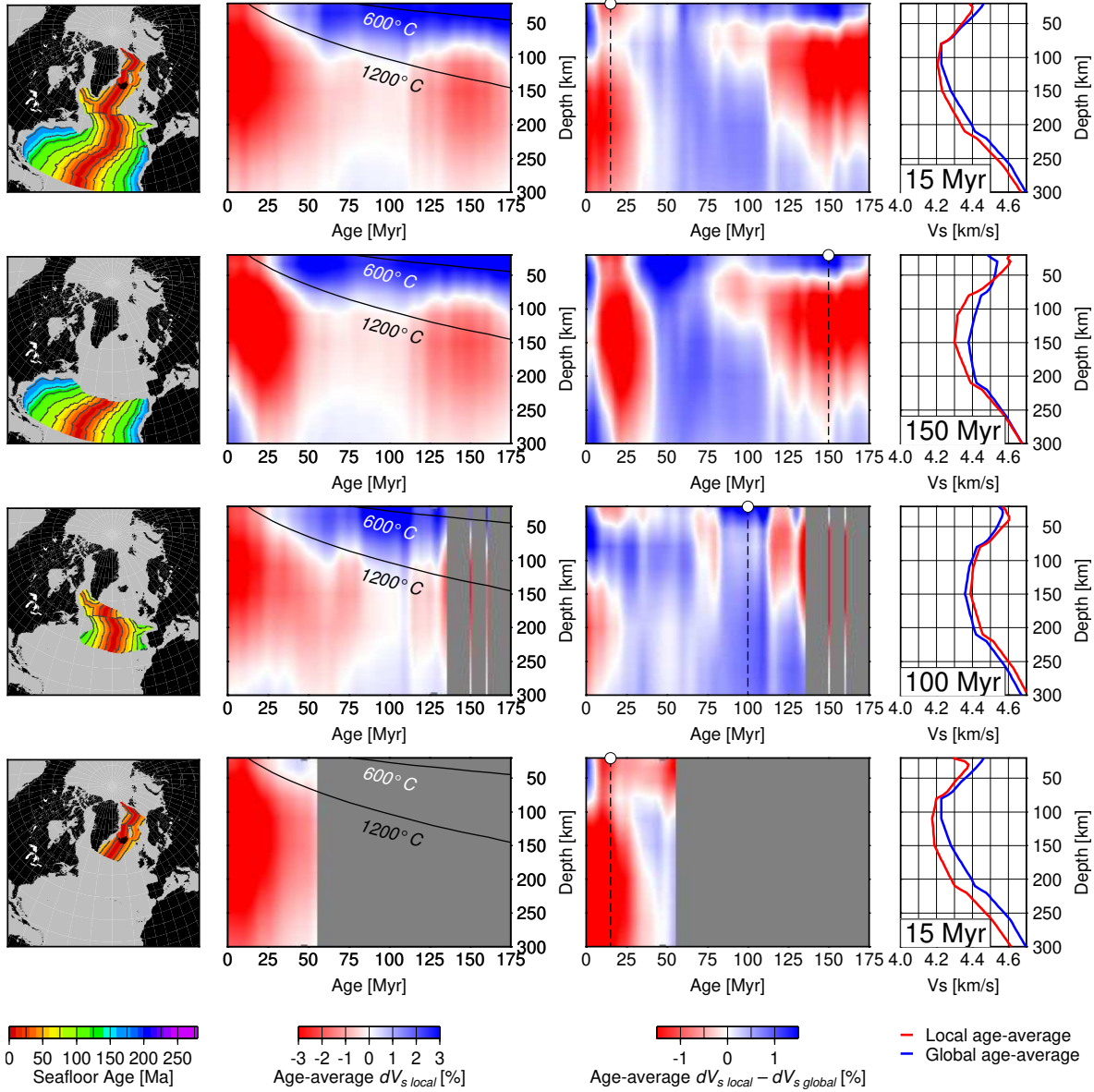


FIGURE 5.12: Age-average cross-sections through the oceanic lithosphere in four subregions of the North Atlantic Ocean. For each subregion, we show, left to right: the lithospheric age of the age-averaging area; the local age-average cross-section; the local age-average cross-section minus the global age-average; a selected average cross profile for a specific oceanic age interval both locally and globally. The 600 and 1200 degrees isotherms from (Becker et al., 2014) are plotted on top of the age-average cross-sections as black lines.

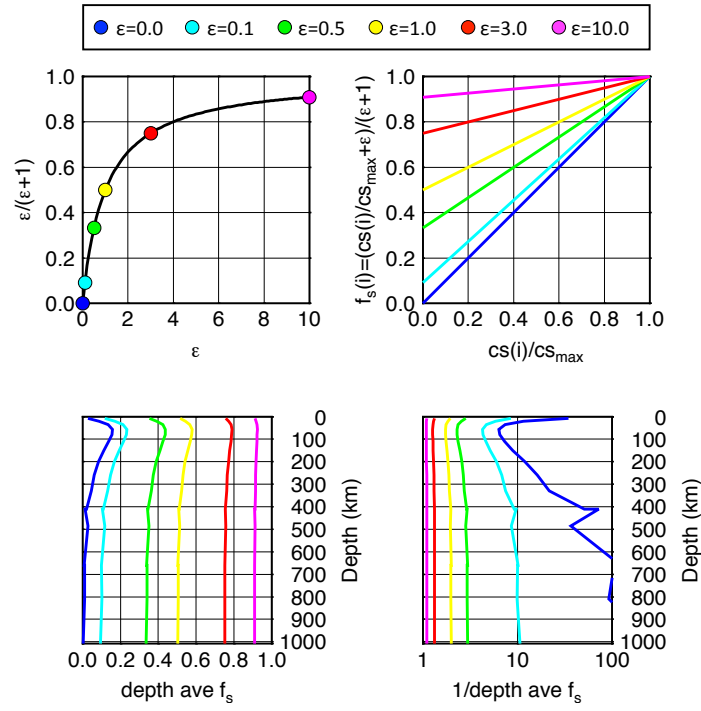


FIGURE 5.13: Analysis of the scaling of the regularisation with the sums of the column of the sensitivity matrix for $\epsilon = 0, 0.1, 0.5, 1, 3$ and 10 . Top left: relationship between the chosen ϵ and the normalised one affecting the regularisation scaling. Top right: effect of different choices of ϵ on the scaling factor $f_s(i)$ for different column sums $cs(i)$ values. Columns sums are normalised to the absolute maximum. Bottom row: depth-average scaling factor $f_s(i)$ (left) and inverse scaling factor (right) for different ϵ values.

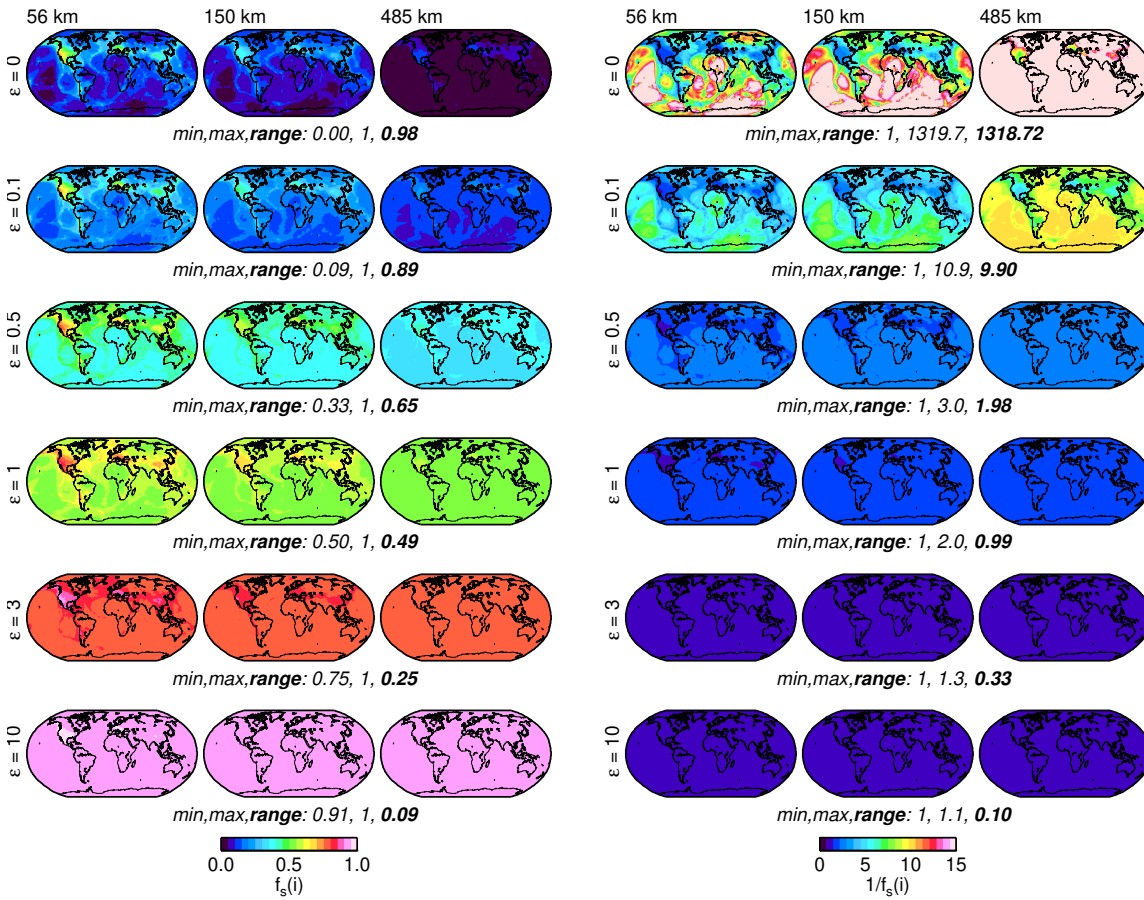


FIGURE 5.14: Example of the effects of the choice of ϵ on the proportional- (left, used for norm damp scaling) and inverse (right, used for smoothing scaling) scaling factor $f_s(i)$ for $\epsilon = 0, 0.1, 0.5, 1, 3$ and 10 , top to bottom. For each ϵ , $f_s(i)$ is plotted at 56, 150 and 485 km depth. Under each row, the minimum, maximum and range over all depths of $f_s(i)$ are reported.

Chapter 6

Global, fundamental-mode phase velocity tomography

Abstract

Surface-wave phase velocity dispersion measurements provide valuable information on the structure of the Earth's upper mantle, but direct measurements are limited by the distribution of seismic stations, leaving large part of the globe uncovered. Here we present a global dataset of 98 fundamental mode, Rayleigh and Love phase velocity maps in the 20-460s period range computed using the Automated Multimode Inversion (AMI) of millions of seismic waveforms. Thanks to the dense global coverage we can parameterise our model with a dense triangular grid with an average 225 km inter-knot spacing. The laterally continuous distribution of phase velocities and the smooth variations across separately inverted periods allow extraction of Love- and Rayleigh dispersion curves at any point on the Earth's Surface. This allows us to produce robust dispersion measurements in areas that are otherwise poorly sampled by seismic stations.

6.1 Introduction

Surface-wave phase velocities are sensitive to the average vertical shear-wave velocity structure along source-station sensitivity areas, with different frequencies and modes of oscillation sampling different depths. For the Rayleigh and Love fundamental mode, increasing wavelengths sample increasing depths, while higher modes have more complex sensitivity patterns and are, on average, more sensitive to greater depths at each given frequency (Fig. 2.3). In order to provide information on the vertical variations in the Earth's structure, phase velocities are commonly inverted for the distribution of S-wave velocity (V_s) within depth layers. The inversion process however is both underdetermined and non-unique and phase velocities represent thus a more "pure" type of seismic data that can be more readily inverted together with other geophysical observables that constrain variations in the Earth's structure to reduce the uncertainties on the local S-wave velocity structure (e.g.: Julià et al., 2000; Fullea et al., 2012).

The distribution of seismic stations around the globe however is uneven, and until recently, phase velocity measurements were either limited to the well-sampled continents, or to long-period measurements coming from large source-station distances, capable of resolving only long-wavelength structures. The recent growth in the number of modern, broadband seismometers dramatically improved the coverage of seismic data, enabling the construction of global dispersion models with increasing coverage and period ranges (Laske and Masters, 1996; Ekström, 2011; Durand, Debayle, and Ricard, 2015). In this study, we gathered over 2.2 million fundamental- and higher-mode Rayleigh and Love phase velocity curves over the whole globe obtained by the Automated Multimode Inversion (AMI) (Lebedev et al., 2005). AMI automatically extracts constraints on P- and S-wave velocity variations from millions of teleseismic waveforms, which have been used to produce a number of well-tested recent global- and regional scale 3D S-wave tomographic models (Lebedev and Hilst, 2008; Schaeffer and Lebedev, 2013; Schaeffer and Lebedev, 2014; Lebedev et al., 2018). As a byproduct of the waveform-fitting procedure, AMI measures phase velocities for all modes that contribute to each waveform fit. While phase velocity measurements from AMI have been used for array-based cross-correlation measurements (e.g. Agius and Lebedev, 2013; Bonadio et al., 2018), our very large dataset allows us to use these measurement to compute global phase-velocity maps via tomographic inversion for both Rayleigh and Love waves. Our dataset includes almost 1 million higher mode measurements, but in this study we will discuss the result from the inversion of the fundamental-mode data only, leaving the inversion of higher modes for future work. The global distribution of phase velocity measurements allows us to extract dispersion curves across all sampled periods at any point on the globe, providing broad-band data in areas with otherwise scarce data coverage.

6.2 Data

Our dataset consists of over 1.17 million Rayleigh- and 300,000 Love-wave fundamental mode phase-velocity curves, spanning a 10-460 s total period range. The curves are obtained by the Automated Multimode Inversion (Lebedev et al., 2005) of waveforms of earthquake recordings, coming from 6,242 stations and 25,496 events worldwide. We used all earthquakes from the Harvard Centroid Moment Tensor Solutions catalog (Ekström, Nettles, and Dziewonski, 2012) since 1994, with some selected data from 1990, with source-receiver distances between 500 and 18000 km, and with magnitudes above a distance-dependent threshold (Schaeffer and Lebedev, 2013).

6.3 Inversion procedure

The advantage of the multi-mode waveform inversion is that both the fundamental and higher mode (surface-wave and body-wave) contributions to the waveform are modelled. Accurate fundamental-mode measurements can be made even when the fundamental mode interferes with higher modes. In order to determine which modes contribute to the waveform fits, we introduce phase velocity perturbations of 1% for each mode and fitting frequency window. After this, we recompute the perturbed synthetic waveforms and evaluate their misfit with the initial ones. Only modes that, if perturbed, result in a substantial increase of the misfit (quantified using an empirical threshold) can be considered measured by the waveform inversion and are retained (Lebedev et al., 2005; Lebedev and Hilst, 2008). The low misfit thresholds and anti-cycle-skip precautions used by AMI (Lebedev et al., 2005) ensure that errors of the measurement of the phase of the signal are negligible. The main sources of errors in the dataset are the uncertainties in the hypocentre locations and origin times of the events and the unaccounted-for fundamental-mode diffraction (Legendre et al., 2012).

We invert the fundamental mode phase velocities for maps at each period, independently. In order to obtain a dense sampling of the frequency range of the measurements, we inverted for maps at 98 different periods in the 10-460 s range, with a logarithmic sample-interval increase. At each period, all phase-velocity measurements are inverted together in a large linear system (Lebedev and Hilst, 2008; Schaeffer and Lebedev, 2013) for the 2D distribution of phase velocity and its 2Ψ and 4Ψ azimuthal anisotropy. Each map is parameterised over a dense triangular grid (Wang and Dahlen, 1995a) with an average 225 km inter-knot spacing. The average over all the phase velocity measurements at the period is used as the reference. In order to ensure consistency in between periods, we vary the regularisation on each phase velocity map by taking into account the different data sampling. For the purposes of this study, we invert for 2Ψ and 4Ψ azimuthal anisotropy to allow the model to explain anisotropic heterogeneities, so that they do not map into isotropic ones.

In order to minimise the impact of errors in the measurements on phase-velocity maps, we exploit the data redundancy of the dataset and select only the most mutually consistent data through an a-posteriori outlier rejection procedure. At each period, we use the first-iteration phase-velocity map to compute synthetic data (by multiplication of the sensitivity matrix and the model vector). We then compare the synthetic and real data and discard the measurements with the largest misfit. In this study, we kept thus a selected, most mutually consistent 70% of the initial dataset and computed the final phase-velocity maps with these "de-noised" data.

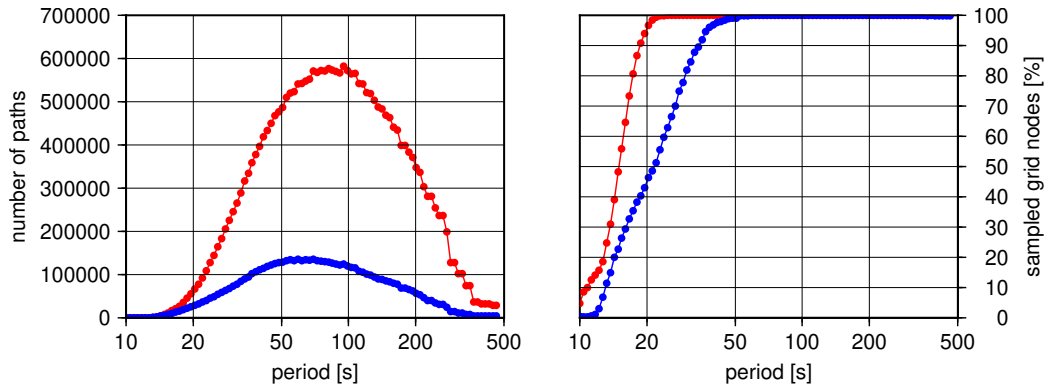


FIGURE 6.1: number of paths (left) and percentage of the model grid nodes covered (right) for each period by the final phase velocity maps, after outlier rejection, for both Rayleigh (red) and Love waves (blue).

6.4 The models

The resulting set of 98 phase velocity maps sample the whole globe. Periods of 50-100 s are sampled by the largest number of paths, and for most of the period range, phase velocity maps cover the whole globe (Fig. 6.1). However, most of the measurements at very short periods are obtained from shorter paths. As a result, periods shorter than 20 s for Rayleigh and 40 s for Love present gaps in the coverage that correspond to areas where short source-station paths—which constrain the highest frequencies—are absent (Figs. 6.3, 6.4). For periods greater than 20- and 40 s for Rayleigh and Love, respectively, phase velocities are constrained by longer source-station distances, allowing for a complete, global data coverage.

Compared to the Rayleigh global phase velocity studies of Ekström, 2011, our models agree on the location of the major features of the mantle such as low velocities under the ridges and orogenies and high velocities under thick oceanic lithosphere and cratons (Fig. 6.2). If we zoom in however (e.g. in the North Atlantic, as shown in the bottom of Fig. 6.2), our maps show greater detail in the distribution of velocity anomalies both in the oceans and continents. Features like the very low velocities close to Iceland and Afar and the boundaries of the North American Craton show very clear, sharp boundaries that match previous tomographic images and geological structures (see, e.g. Schaeffer and Lebedev, 2014; Rickers, Fichtner, and Trampert, 2013; Begg et al., 2009). Africa and Greenland are continents sampled by the least amount of seismic data; in the location of the Greenland and West African Cratons, our model retrieves separate high-velocity cores where previous images show a single, smooth unit, indicating the gain in resolving power obtained by the addition of the new data. In the following subsections we will discuss observations on the 2D tomographic models by grouping them by period ranges.

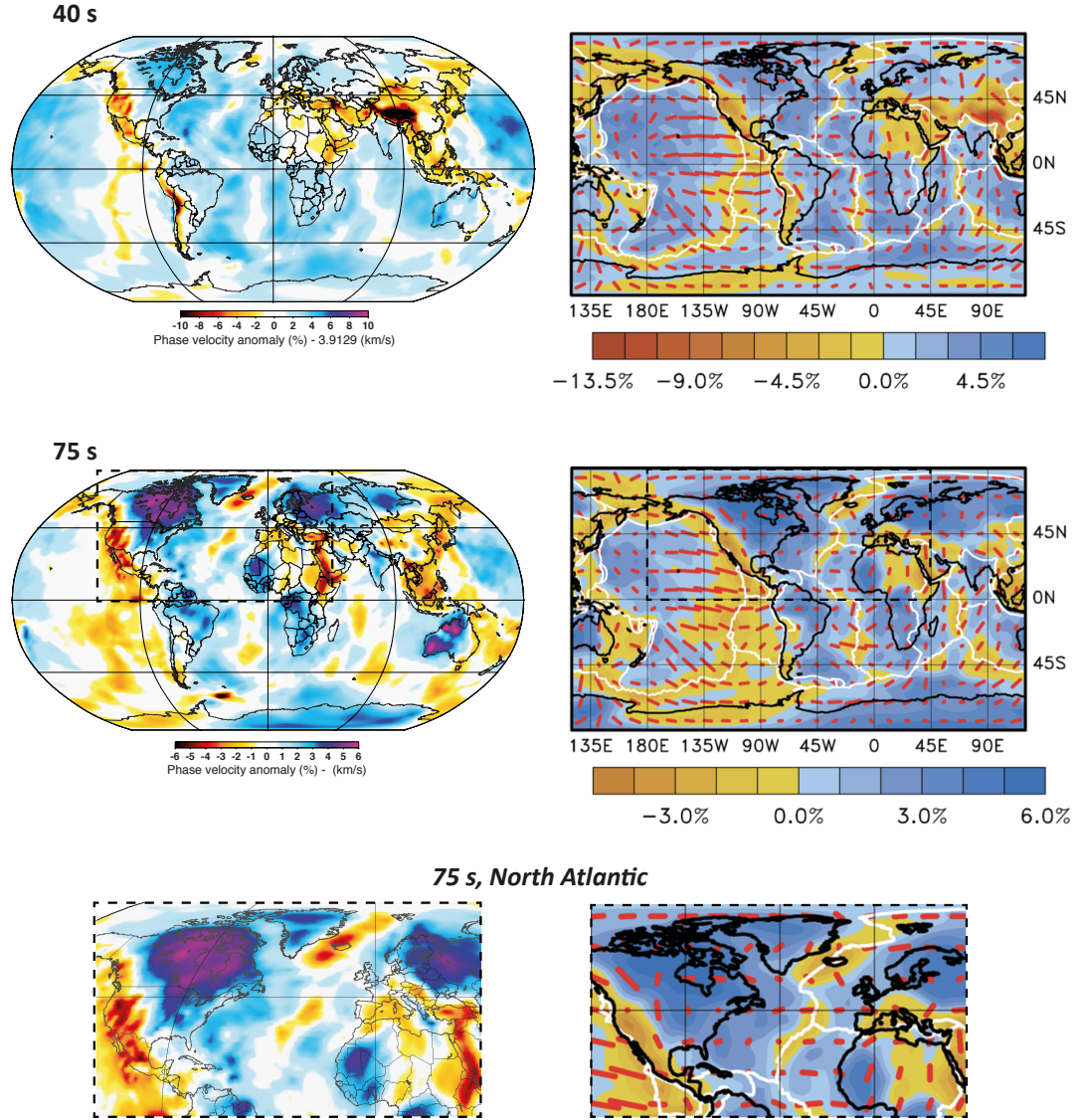


FIGURE 6.2: Comparison of phase velocity anomaly maps for Rayleigh waves between this study (left) and Ekström, 2011 (right) for periods of 40 and 75 s. Anomalies are with respect to the inversion reference and PREM (Dziewonski and Anderson, 1981) for this study and the maps from Ekström, 2011, respectively. The bottom panels show a zoomed view of the 75 s maps; the zoomed area is shown as dashed lines.

6.4.1 Short periods

For periods lower than 20 and 40 s for Rayleigh and Love, respectively, the oceans and Africa present coverage gaps. In these areas, the sampled portions correctly report high velocities in the oceans, as expected in the oceanic lithosphere (Figs. 6.5, 6.6). In the Americas, Eurasia and Australia, the source-station distribution allows for a good coverage of the continents. At 20- and 30 s our maps identify low velocities of the continental crust, with minima corresponding to major orogenic belts (Tibet, Mongolia, Andes, North American Cordillera, Zagros and Caucasus), in agreement with evidence for the presence of thickened crust (e.g.: Royden et al., 1997; Meijde et al., 2015).

6.4.2 Intermediate periods

At periods between 50- and 150 s, fundamental mode Rayleigh waves are mostly probing the lithosphere in the continents and both lithosphere and asthenosphere in the oceans. Our maps at 60 s show very high velocities corresponding to the location of the North American, East European, Siberian, Amazon, São Francisco, West Africa, Congo, Kalahari and Australian Cratons (Fig. 6.5). High velocities, most prominent in 60 s Love-wave maps, are visible in the oldest oceanic basins, indicating the sampling of the cooling oceanic lithosphere. We image the highest velocities in the oceans in the western Pacific, corresponding to some of the oldest—and thus thickest and coldest—oceanic lithosphere on earth (Müller et al., 2008). Under mid-ocean ridges, we image continuous low velocities, which are both lowest and most widespread under the fast-spreading ridges (Müller et al., 2008) of the Pacific and SW Indian Ocean. Areas of particularly pronounced low-velocities underlay major hotspots (Iceland, Azores, Eifel, Cabo Verde, Tristan da Cunha, Afar, Hawaii, Samoa, Macdonald, Pitcairn, Galapagos, Kerguelen).

6.4.3 Long periods

At long periods (larger than 150 s), fundamental-mode surface waves sample a large depth interval, spanning from the lithosphere to the transition zone (Fig. 2.3). Although Rayleigh-wave sensitivity at these periods peaks in the deep upper mantle, the wide depth sampling makes visual interpretation of the phase velocity maps difficult. Non-vertical structures expected in this depth range, such as slabs, are averaged out in the phase velocity maps by the simultaneous sampling of the portions of the mantle immediately above and below. At 350 s, only the slabs in SE Asia are visible. We image strong, low-velocities under the Afar hotspot and East African Rift Valley, in agreement with the presence of a large-scale vertical upwelling in the area (Ebinger and Sleep, 1998). Similar low velocities are also visible in SW North America under the shallowest portions of the slab. Finally, very high velocity nuclei locate under the Sahara desert, South-eastern Brazil and Southern France. We interpret these as probably due station timing errors

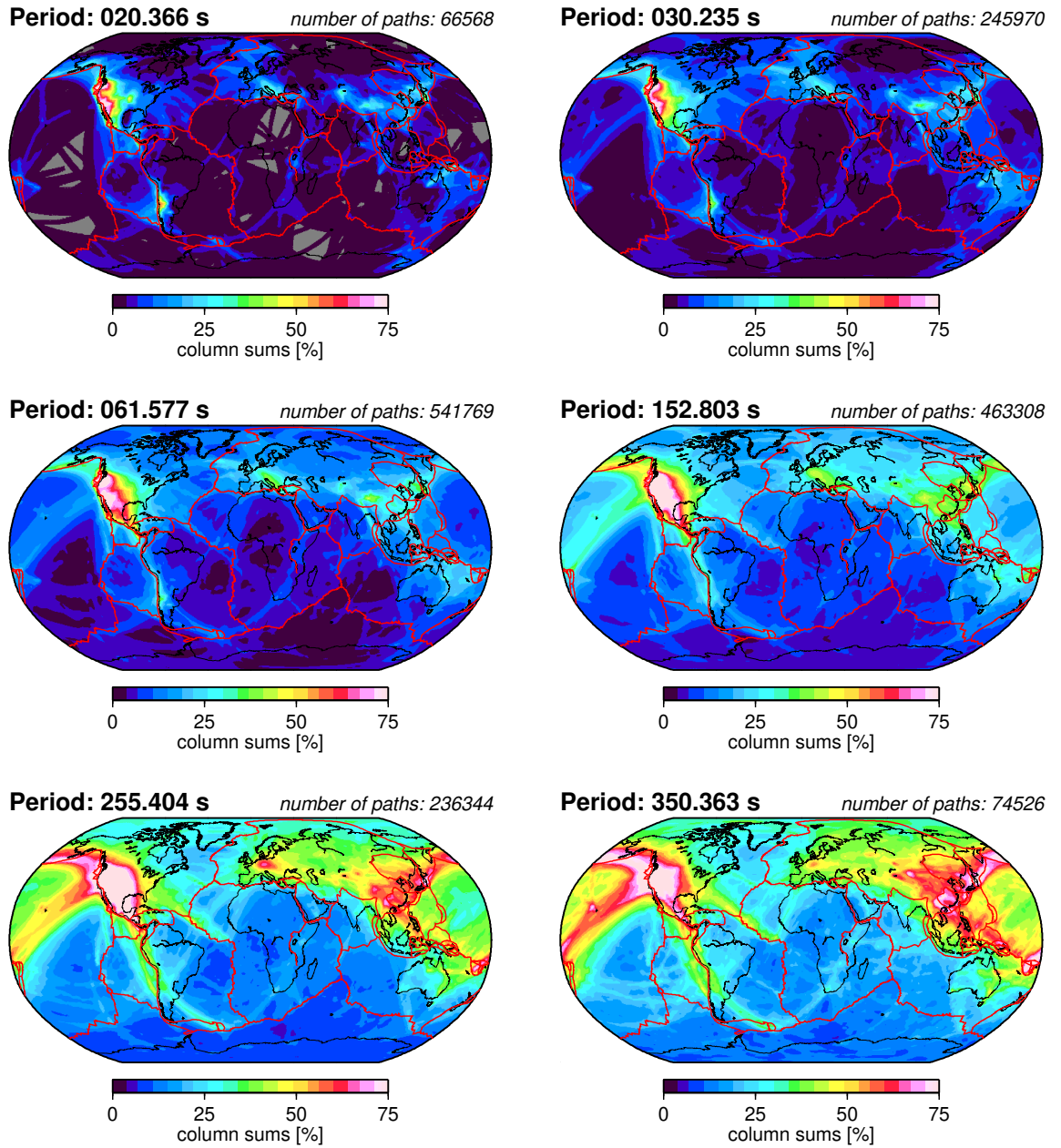


FIGURE 6.3: Sums of the column of the sensitivity matrix for Rayleigh waves at six sample periods. Values are in percentage to the maximum at each period. The total number of paths is reported at the top right of each plot.

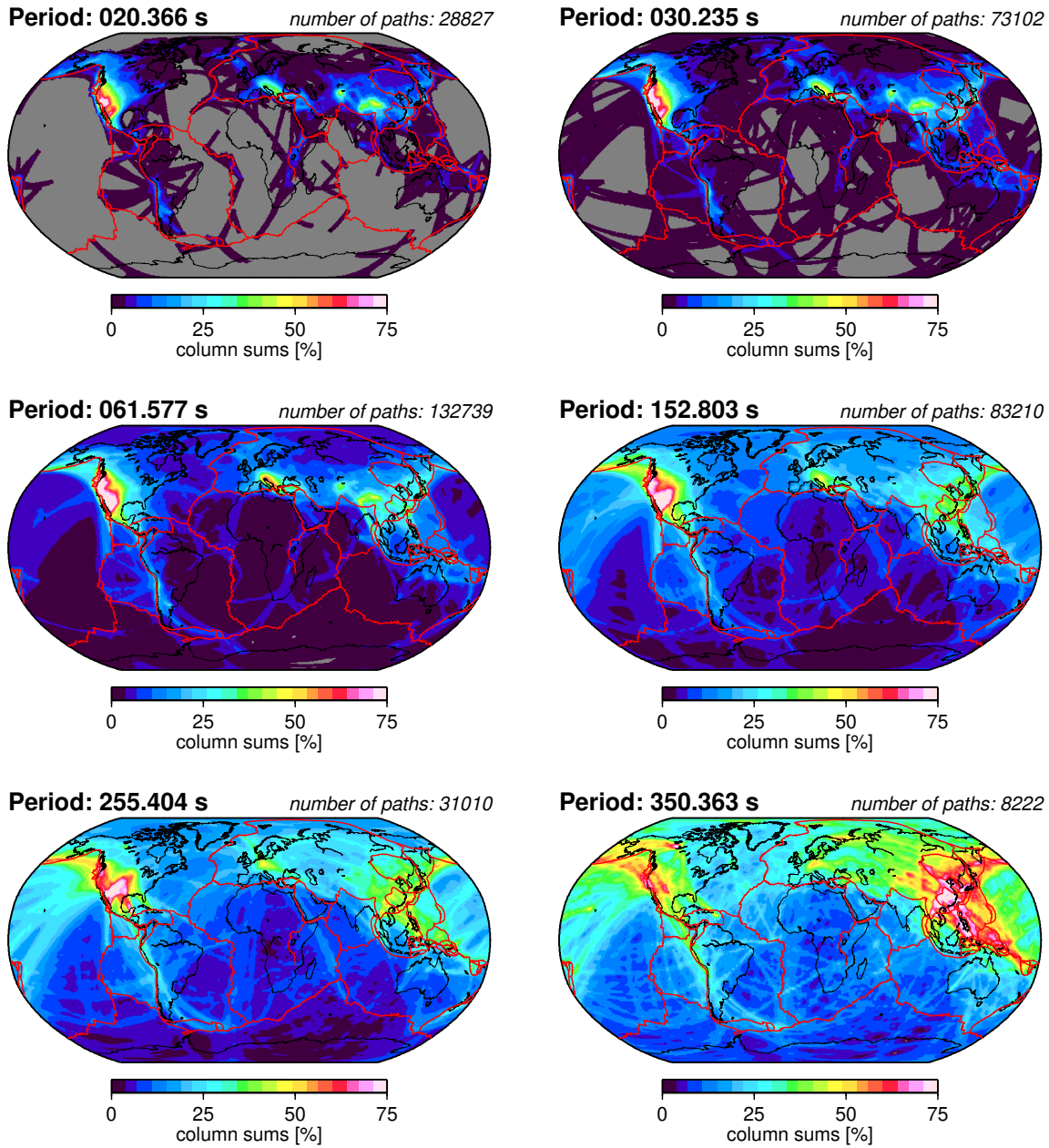


FIGURE 6.4: Sums of the column of the sensitivity matrix for Love waves at six sample periods. Values are in percentage to the maximum at each period. The total number of paths is reported at the top right of each plot.

(Legendre et al., 2012), focussed under the station by the tomographic inversion to minimise data misfit.

6.4.4 Global dispersion curves

The resulting set of phase velocity maps allows us to extract phase velocity dispersion curves at any point on the Earth's surface. Phase velocities at each grid node of the triangular model grid yield a dataset of 12,252 dispersion curves, covering the globe with an approximately even node spacing. By stacking all Rayleigh curves together, we can observe two different trends in the distribution of phase velocities (Fig. 6.7). Selecting curves under dry ("Continental") and wet areas ("Oceanic") only, we are able to approximately separate curves sampling continental and oceanic lithospheres. While this is a crude approximation, it provides clearly distinct trends for the two regions, indicating that our curves can provide robust average values for distinct lithospheric types. Dispersion curves extracted under continental lithosphere, are on average anomalously slow for periods shorter than 40 s and fast for longer ones with respect to the average, reference phase velocity. This fits the lower velocities generally associated with the thicker and slower continental crust, and the cold and fast cratonic lithosphere observed by global tomographic models (Schaeffer and Lebedev, 2015). Exceptions to the "dry/wet" approximation (e.g. the continental—but submerged—Barents Sea) map into the distributions as areas of secondary maxima. The curves plotted in Fig. 6.7 show the full sampled period range.

6.5 Local application: the Hangai Dome

We tested the validity of our measurements on a regional scale in the region of the Hangai dome. The Hangai dome is an anomalously high plateau in central-northern Mongolia, adjacent to the Baikal Rift Zone, in turn bordering with the Siberian Craton to the North. The origin of the Baikal Rift is debated, and while geophysical imaging largely agrees on the presence of a mantle upwelling, it is unclear whether it is an active mantle plume (Chen et al., 2015) or a result of rifting (Lebedev, Meier, and Hilst, 2006; Fullea et al., 2012). The Hangai dome and the neighbouring Sayan Mountain Range are constellated with recent volcanism (Fig. 6.8) (Hunt et al., 2012), and although recent studies demonstrated that the average topography in the region can be explained solely by local isostasy (Fullea et al., 2012), large lateral elevation differences exist between the volcanically active Hangai dome and the neighbouring Lake Region (dashed lines in Fig. 6.8).

In this section, we use our global Rayleigh and Love phase velocity maps to overcome the relatively poor sampling of seismic data in the region and identify possible lateral differences in the mantle structure that can explain the distribution of volcanism

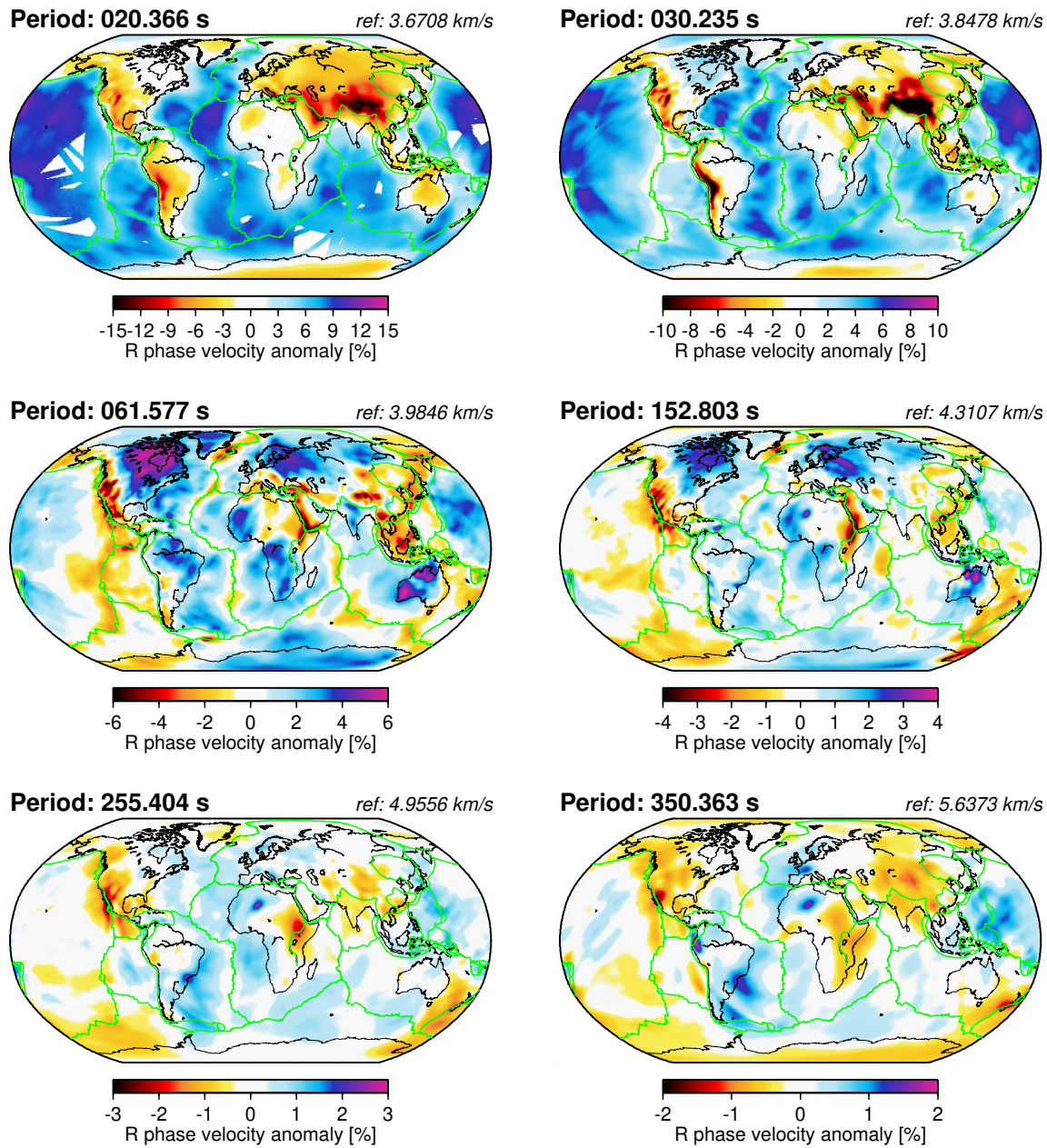


FIGURE 6.5: Phase velocity anomaly maps for Rayleigh waves at six sample periods. Anomalies are with respect to the inversion reference, which is the global average phase velocity of all individual source-station measurements and is reported on the top right of each plot.

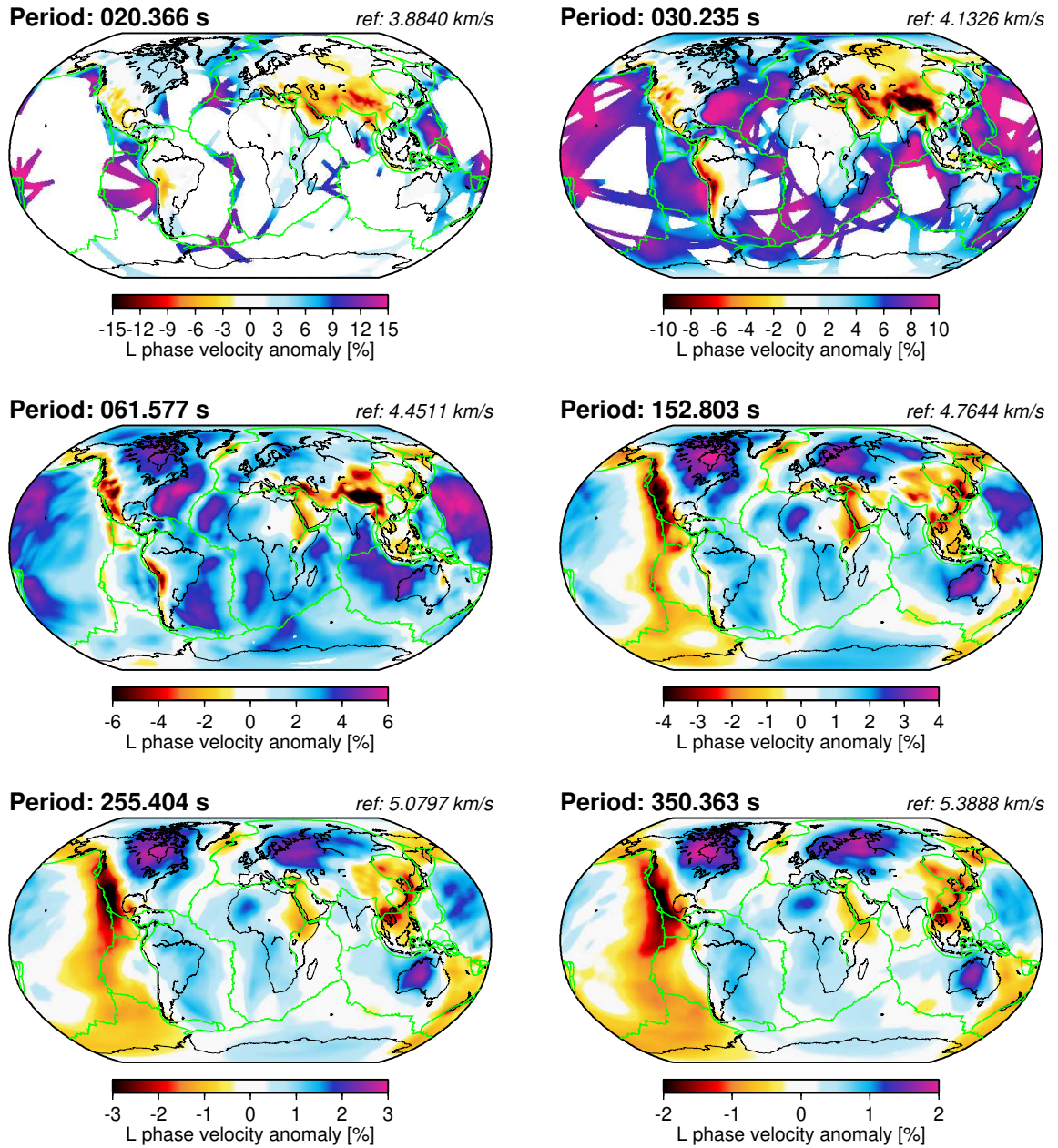


FIGURE 6.6: Phase velocity anomaly maps for Love waves at six sample periods. Anomalies are with respect to the inversion reference, which is the global average phase velocity of all individual source-station measurements and is reported on the top right of each plot.

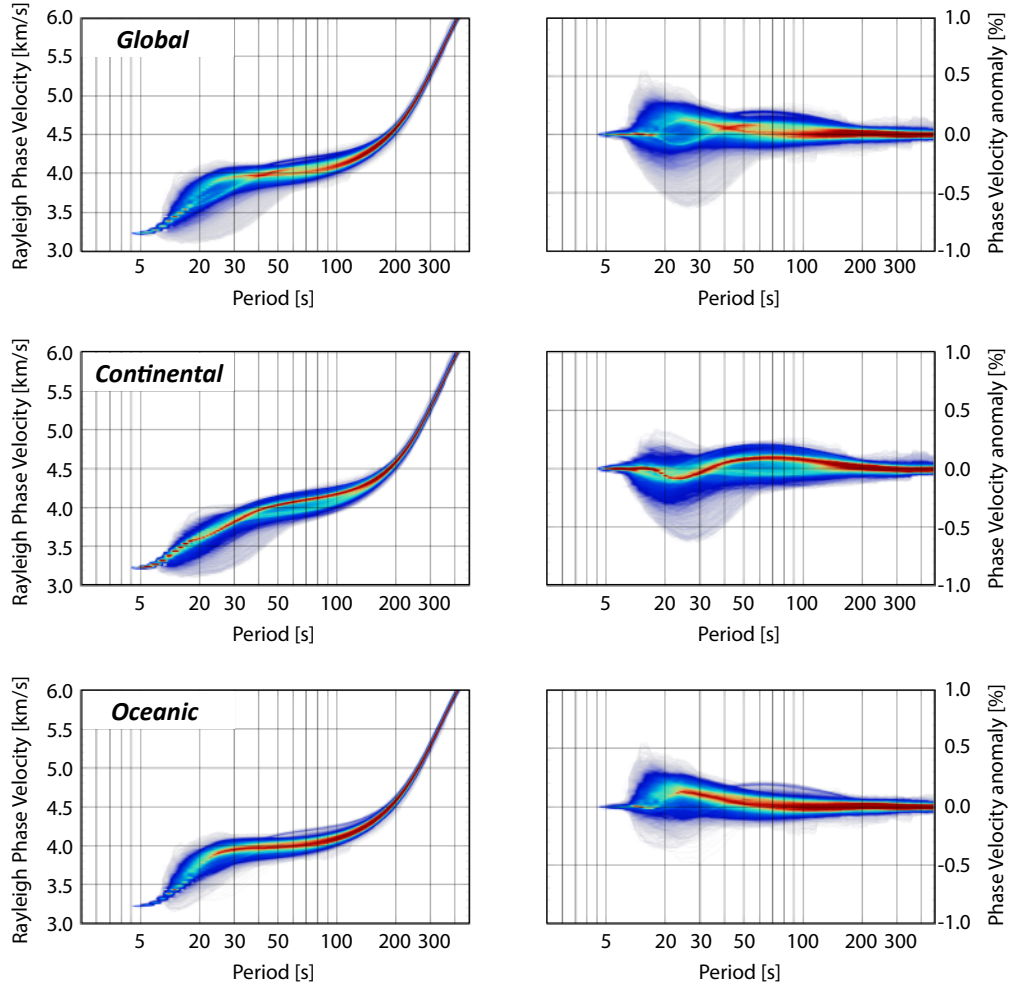


FIGURE 6.7: Distribution of Rayleigh dispersion curves extracted from our phase velocity maps, shown both as absolute phase velocities (left) and phase velocity anomaly with respect to the reference at each period (right). Top panels show the distribution of velocities for all curves. Middle and bottom panels show curves under continental and oceanic areas, respectively. In each panel, the distribution is normalised to the maximum.

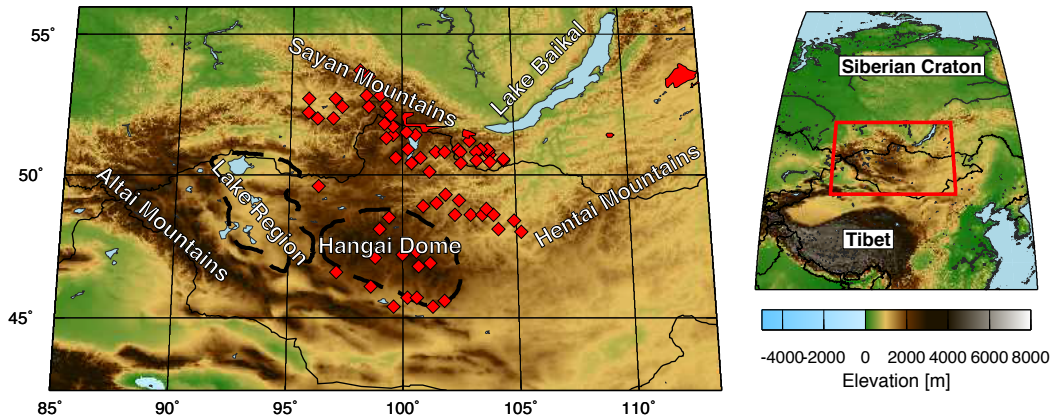


FIGURE 6.8: Elevation map of the study area and its location in eastern Eurasia. Location of recent volcanism is plotted as red squares. The areas of the Hangai Dome and Lake Region are drawn as black dashed lines.

and topography in the area. At a large scale, both Rayleigh and Love phase velocity maps sample pronounced high velocities north of the Baikal Rift, indicating the presence of the Siberian Craton (Fig. 6.9). South of the rift, low velocities characterise the Mongolian orogenic belts. To the south of the study areas, extremely low phase velocities underlie the Tibetan plateau. In and around the Hangai Dome, low velocities are mostly localised under the dome and south-eastern Mongolia at short periods (e.g. 20s map in Fig. 6.9). At 50 s, the velocity low moves NW, underlying both the Hangai Dome and the Lake Region. At longer periods, the velocity minima fade into a wider, moderate low-velocity area that extends south of the Baikal Rift.

In order to obtain robust estimates of phase velocity under the Hangai dome and Lake Region, we extracted dispersion curves every 0.5 degrees under the two areas (Fig. 6.10) and averaged them (Fig. 6.11). On average, phase velocities both under the high Hangai dome and the low Lake Region are much lower than an average continental value, represented in Fig. 6.11 by the 1D model AK135 (Kennett, Engdahl, and Buland, 1995). However, the averaged curves are very similar in between the two areas at medium- and long-periods. At 20-30 s, average phase velocities for the Hangai Dome are substantially lower than under the Lake Region. Our observations suggest that while the region is underlain by a major mantle anomaly, the Hangai dome seems to differ mostly in the shorter periods, indicating a shallow source for the observed topographic differences.

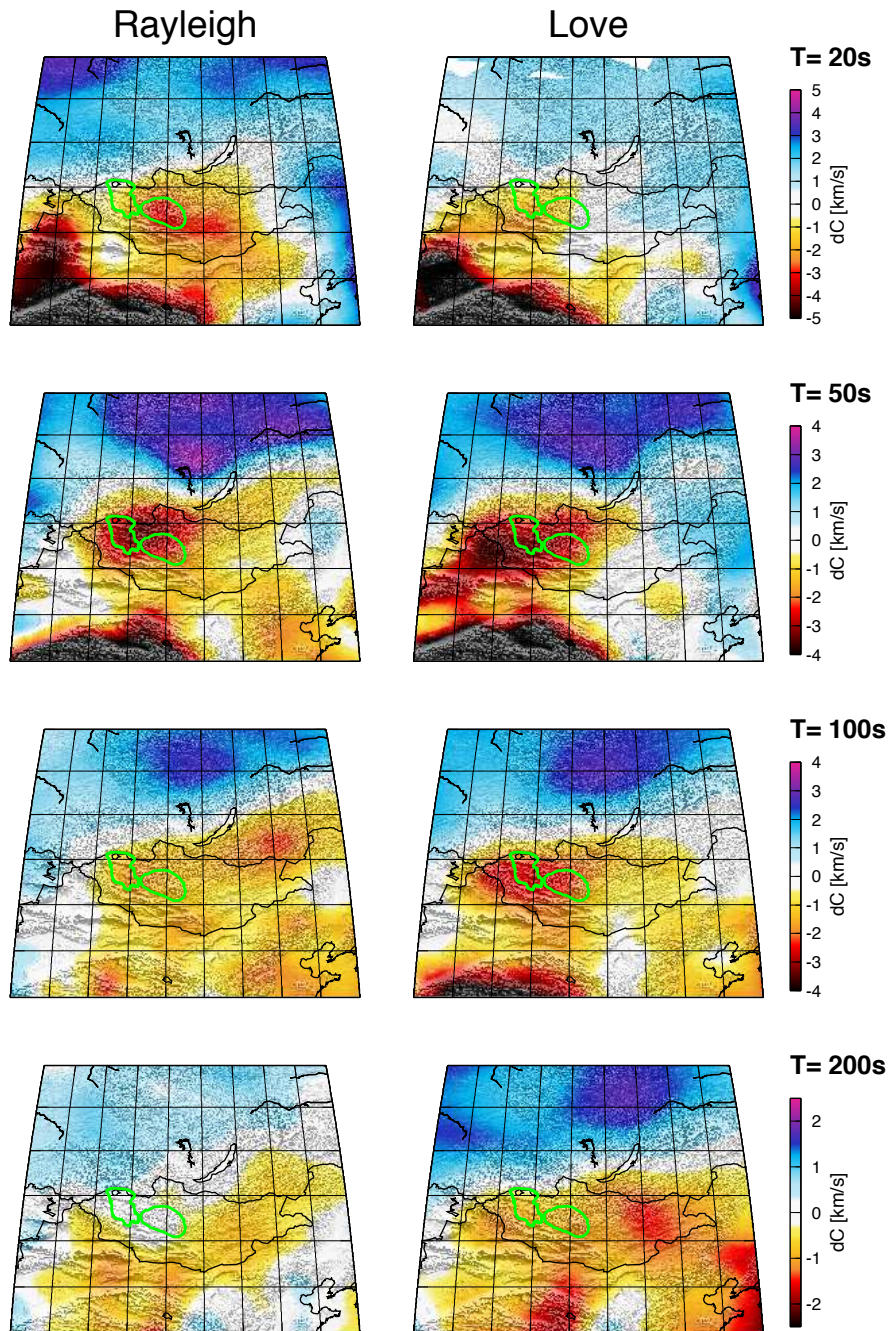


FIGURE 6.9: Phase velocity maps for the study area at periods of 20, 50, 100 and 200s. Rayleigh and Love phase velocities are plotted left and right, respectively. Location of the Hangai Dome and Lake Region are drawn in green.

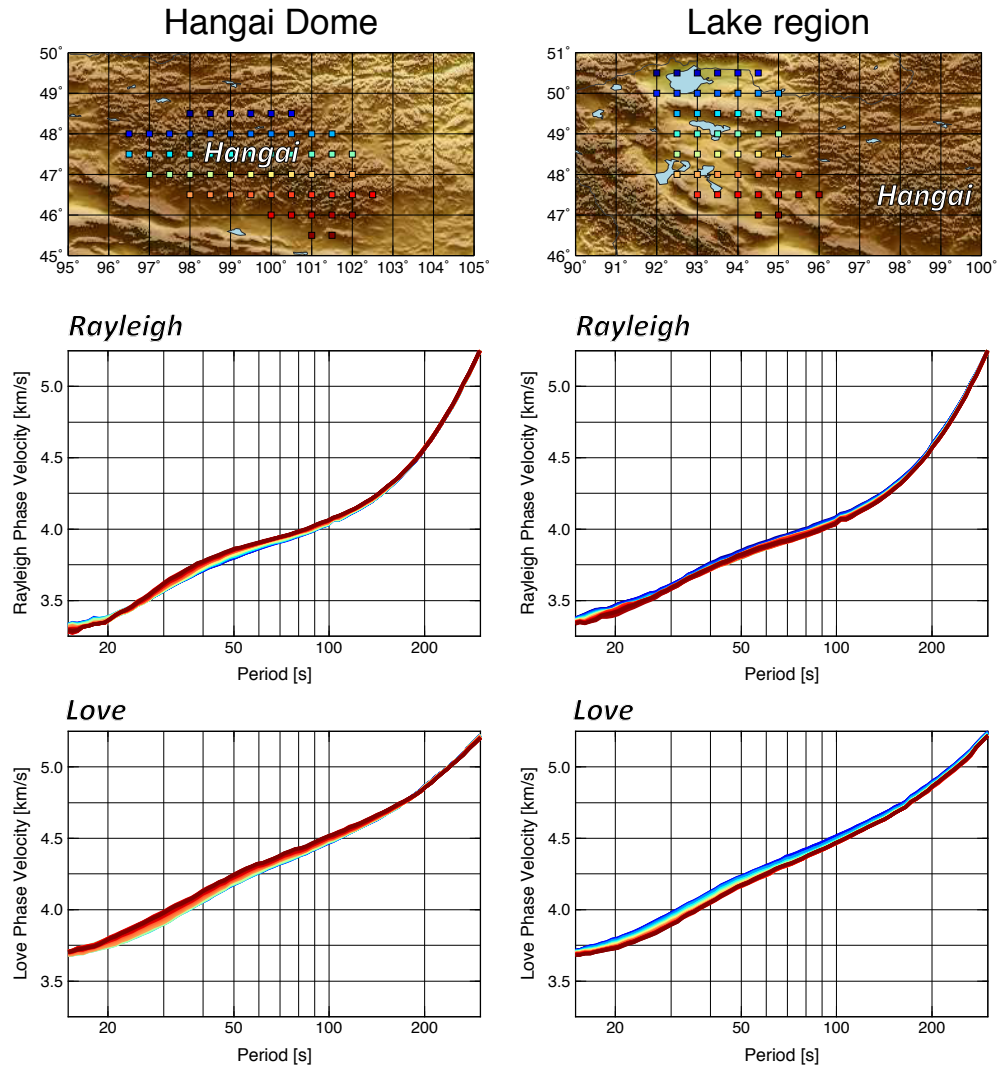


FIGURE 6.10: Dispersion curves extracted from the phase velocity maps for the Hangai Dome (left) and Lake Region (right). The location where the curves are extracted are plotted as squares on top of topography, colour-coded by latitude. Rayleigh waves are plotted in the middle panels, Love waves in the bottom ones. Colours correspond to the ones in the maps.

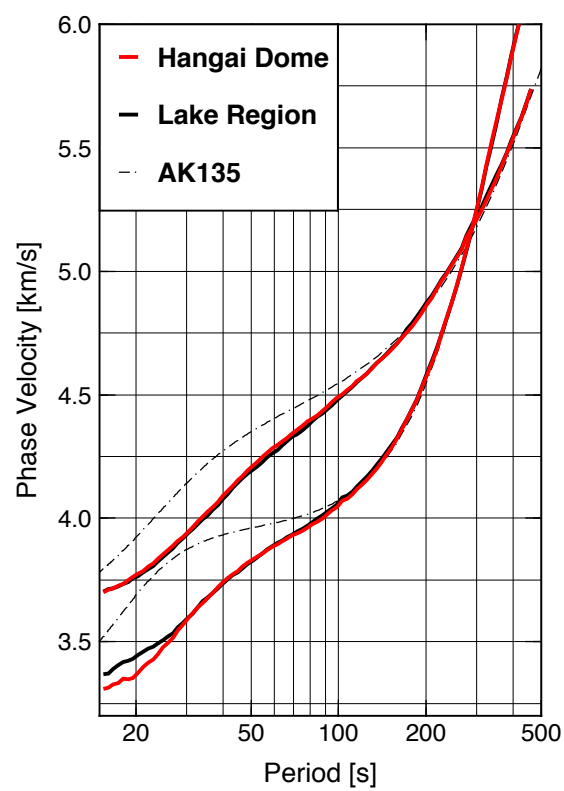


FIGURE 6.11: Average Rayleigh and Love dispersion curves for the Hangai Dome, the Lake Region and AK135. The curves are the average of the ones shown in Fig. 6.10.

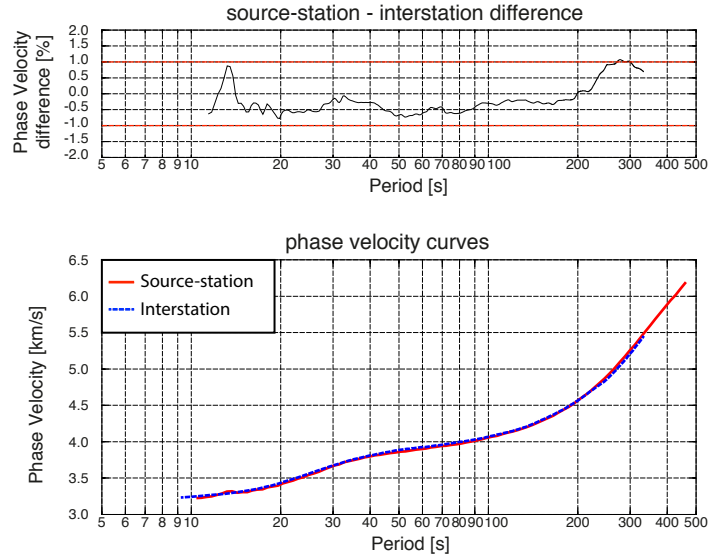


FIGURE 6.12: Comparison between the Rayleigh-wave dispersion curves derived from our phase-velocity maps and the one derived from inter-station interferometry between stations TLY and ULN. Difference in percentage is plotted on top, while the actual curves are plotted on the bottom.

6.5.1 Comparison to inter-station measurements

In order to validate the extraction of dispersion curves on a small scale, we compared our results (source-station) with dispersion curves obtained through inter-station cross-correlation measurements (Lebedev, Meier, and Hilst, 2006) between the Global Seismographic Network stations TLY (Talaya) and ULN (Ulaanbaatar), separated by 500 km. While source-station measurements offer robust dispersion at intermediate- to long periods, inter-station measurements provide better results at higher frequencies (Lebedev, Meier, and Hilst, 2006). Both methods however overlap in a broad frequency range, allowing for cross validation. In order to compare the two measurements, we sampled our Rayleigh phase velocity maps every 0.5 degrees on the great circle path connecting the two stations and computed the average. By comparing the resulting phase velocities (Fig. 6.12), we can see that velocities differ by approximately 0.5% or less on most of the periods, showing appreciable consistency. At very short periods the differences increase, in agreement with the more accurate modelling of high frequencies obtained from inter-station interferometry (Lebedev, Meier, and Hilst, 2006). For very long periods, the opposite is true, with the low frequencies being better constrained by the long paths used in the source-station method.

6.6 Conclusions

We are able to provide a set of global, fundamental mode phase velocity maps in a very broad 20-460 s period range for both Rayleigh and Love waves. The maps result from the inversion of a very large global dataset, which allows us to have a distribution of phase velocities which is smooth and continuous both laterally and across periods. We can hence extract dispersion curves at any point on the globe, that can be then readily inverted for the distribution of S-wave velocities with depth. The distribution of phase velocity curves, extracted under each model grid node, shows clearly separate trends under oceanic and continental areas, indicating that our model can accurately identify major differences in lithospheric structure. Thanks to the very large coverage, we can go one step further and resolve features on a finer scale than previous global phase velocity models (Dalton and Ekström, 2006; Ekström, 2011; Durand, Debayle, and Ricard, 2015). Globally, the phase velocity maps sample the seismic signature of the cooling oceanic lithosphere, hot (slow) at mid-ocean ridges and cold (fast) in the oldest oceanic basins. In the continents, areas of thick continental crust are highlighted by low seismic velocities, while in the lithosphere, thick, cold cratonic blocks show fast velocities. Under both oceans and continents, we image slower phase velocities underneath hotspot areas. Regionally, we tested the validity of our maps in northern Mongolia, where both plume and plate origin have been suggested to explain the presence of the anomalously high and volcanically active Hangai Dome. By both evaluating the phase velocity maps locally and extracting and averaging the dispersion curves under the area, we observe slower velocities in whole central northern Mongolia. Under the Hangai Dome, velocities are lower at short periods compared to the neighbouring, low-topography Lake Region. Our results show that our maps can be successfully used in the context of regional studies, yielding valuable information even in areas where local studies are otherwise limited by the presence of seismic stations.

Chapter 7

Conclusions

The Automated Multimode Inversion of teleseismic waveforms is an efficient tool for the extraction of structural information from millions of seismograms. Global tomographic models computed using the equations produced by AMI (Lebedev and Hilst, 2008; Schaeffer and Lebedev, 2013) are able to constrain the structure of the crust, upper mantle and transition zone worldwide, but different regions can have very different data sampling and thus require different parameterisation and regularisation to produce the best possible models.

In this thesis we retrieved and inverted all freely available regional waveform data in Africa, South America, the South and North Atlantic Oceans and surrounding continents and combined them with a global waveform dataset, obtaining a very large database of over 1.2 million vertical- (Rayleigh) and 300.000 horizontal (Love) component waveform fits. By inverting these equations in a large, linear system, and carefully choosing its parameterisation and regularisation, we produce tomographic models that are global, but are optimised to yield the best results on the area of interest, where coverage is maximised.

In chapters 3, 4 and 5, we computed regional-scale 3D tomographic models of P- and S-wave velocity and S-wave azimuthal anisotropy of the crust, upper mantle and transition zone using the linear equations produced by AMI from our database of ~ 1.2 million vertical component seismograms. In chapter 6 we use instead the phase velocity dispersion curves computed by AMI from both our vertical- and horizontal component databases of waveform fits. We then invert the curves to compute a set of 98 global phase velocity maps in a 20-360 s period range for both Rayleigh- and Love waves.

7.1 Tackling the unevenness of seismic coverage

In all of the study areas, seismic data coverage is very heterogeneous across the models. Oceanic (the South Atlantic Ocean in chapter 3 and the North Atlantic Ocean in chapter 5) and inaccessible (the Amazon Basin in chapter 3, Sahara desert in chapter 4 and Greenland in chapter 5) regions present much lower coverage with ray paths compared to the densely instrumented Andes, Eastern Africa, North America and Europe.

In order to produce models that are consistent across the study area, in chapter 3, 4 we chose the regularisation parameters based on the results of spike tests. While it is possible to choose the optimal regularisation based on the results from the inversion of real data, its effects on differently sampled areas can be difficult to identify within the complexity of real structures. By inverting a synthetic model with spikes distributed evenly in the region of interest, we quantitatively assess the effect of the regularisation in 3D and choose the set of parameters that reconstruct the signal more accurately for all spikes across the whole study area. In addition to this, assessing the spike width on each model grid knot provides an a-posteriori quantitative measure of the lateral resolution of the model. In chapter 5, the lateral variations in coverage between the North Atlantic and the surrounding continents are even more pronounced. In order to tackle this, we fine tuned a multi-resolution regularisation scheme that varies the regularisation coefficients in concert with data sampling. By relaxing the norm damping and increasing the smoothing with the decrease in seismic coverage, well-sampled areas like western Europe receive minimum smoothing but maximum damping, to avoid biasing the inversion towards them. In less sampled areas, the opposite is true, preventing the overdamping of less sampled portions on of the model, while regularising them through the increased smoothing. This allows us to parametrise our model very densely, with the multi-resolution regularisation smoothing areas that are less constrained.

7.2 New images of the structure of the cratonic lithosphere

In chapter 3 and 4, we present models that sample the lithosphere of Africa and South America. Both continents present large outcrops of cratonic crust, and previous, smoother tomographic models (e.g. Ritsema et al., 2011; Begg et al., 2009; Feng, Lee, and Assumpção, 2007) provided images that suggested the presence of few, large cratonic units in the lithosphere similar to the ones observed in North America and eastern Europe. Our models SA2019 (chapter 3) and AF2019 (chapter 4) however reveal a cratonic lithosphere that is much more complex than previously thought. In South America, the lithospheric roots of the Guyana and Guaporé Shields extend beyond the contours drawn from surface geology, underlaying part of the Amazon Basin, and are absent underneath their youngest, western provinces. In Africa, we image the presence of separate cratonic units within the boundaries of the West Africa- and Congo Cratons. We retrieve the presence of seismically fast lithosphere, typical of cratons but unexposed at the surface, under the Paraná and Parnaíba basins and the proposed location of the Niassa Craton in southeast Africa. On the other hand we also image little or no seismically fast lithosphere under the location of Archean shields such as the Luís Alves, São Luis and Rio de La Plata in South America and Angola, southeastern Kaapvaal and Tanzania in Africa.

7.3 Evidences of cratonic evolution

This new, exciting and somewhat puzzling view of the most ancient and stable cores of the continents can only partially be explained by an improvement of the tomographic models. In chapter 4, comparing the new, sharp boundaries of the cratonic lithosphere to independently derived geological data (Begg et al., 2009) indicate areas where there are evidences for cratonic crust, but no matching fast lithosphere is found, suggesting its erosion through time. In order to map this erosion, we compare the structure of the cratonic lithosphere produced by our model with proxies for its past presence in the form of diamondiferous kimberlites. Our results show that in the past 200 M.y., many of the African cratons lost a large part of their lithosphere. Comparing areas of inferred loss of the cratonic lithosphere with the distribution of Large Igneous Provinces and Plate reconstructions, we infer that erosion follows the impact of Mantle Plumes on the cratons.

7.4 Hotspot-lithosphere interactions

Throughout this thesis, we image the presence of low-velocity anomalies in the mantle underneath a large number of hotspots. The sensitivity of our data does not allow us to identify whether these are deep mantle plumes sourced in the lower mantle, but we retrieve the presence of low velocity anomalies that penetrate the transition zone in a number of locations. Strong low velocity anomalies under Cabo Verde and Vema (chapter 3), Afar and the East African Rift System (chapter 4) and Iceland (chapter 5) can be visible as deep as 485 km depth. Although we interpret these as mantle plumes, the images we retrieve differ greatly from the classic view of plumes as thin, vertical structures (Morgan, 1971), revealing instead the presence of broad, complex low-velocity areas, not dissimilarly from that proposed by French and Romanowicz (2015). The shape of the plumes we image seems to be strongly influenced by the surrounding structures: in east Africa, the low velocities appear to form a vertical "curtain" that follows the plate boundary of the EARS; in Iceland, the low velocities are rooted under eastern Greenland in the mantle transition zone and rise, upwards and eastwards, towards Iceland at 100 km depths and shallower, where they distribute along the Reykjanes and Kolbeinsey ridges; in the South Atlantic, the low velocities under the Vema Seamount seem to flow westwards towards the South Atlantic Ridge and Tristan da Cunha at 150 km dept and shallower. The structure of mantle plumes in the upper mantle seems thus to be guided by the presence of lithospheric structures. Moreover, we can observe an interaction between hotspots and spreading centres that creates complex low-velocity systems in which the contribution of passive and active upwelling to the seismic image is difficult to discern (Figs. 3.19, 4.4, 5.3).

7.5 Age-averaging of the oceanic lithosphere

In chapters 3 and 5, we compute age-average profiles of the lithosphere in the local basins and, exploiting the global design of our models, globally. Being the oceanic lithosphere cooling and thickening with age (Stein and Stein, 1992), these profiles are indicative of the cooling state of the oceanic lithosphere. We find that locally, different oceanic basins show very different average cooling. By correlating these with the presence of hotspots, we find that the lithosphere of the basins is on average anomalously slow (hot) close to hotspots. We then use the global age-average as representative of the "normal" average cooling and subtract it from the local age-averages; the difference shows where the local lithosphere does not follow normal cooling. This way, we quantitatively separate the signal of the average spreading lithosphere from the anomalous ones. In Iceland, this evidences that the lithosphere west of the North Atlantic Ridge is hotter than the average, indicating that the plume is at present mostly confined under the Greenland plate. In the South Atlantic Ocean, the differential cooling clearly shows the presence of weak low velocities corresponding to the Tristan da Cunha hotspot, masked by the mid-ocean ridge in the tomographic images.

7.6 Local phase velocity curves from global tomography

In chapter 6, we compute global, fundamental mode phase velocity maps at 98 periods for both Rayleigh and Love waves. Compared to previous work (Ekström, 2011), our maps show an improvement in the resolution of known geological structures. The very wide 20-360 s period range and the global coverage allow us to extract phase velocity dispersion curves everywhere on the globe, producing a dataset that can be readily used to invert for the local seismic structure. The advantage of our dataset is that it is consistent throughout the globe and across a wide period range, containing information in areas otherwise poorly sampled by seismic data. By testing the local validity of our phase velocity maps in northwestern Mongolia, where few seismic stations are present, we see that we are able to resolve anomalies in the scales of few hundred kilometres.

7.7 Final remarks

The tomographic models we computed sample features on a broad range of scales, from the narrow Nazca slab and the details of the African and South American cratons, to the large and deep upwelling under the East African Rift Zone and Iceland. Resolving the structure of the lithosphere and underlying mantle at a regional scale, we learn that some features we thought to be thousands of kilometres large, like the West Africa and Congo Cratons, have a surprisingly small seismic signature, while other structures that were proposed to be narrow, like the Iceland Plume, appear as very large and complex

in our tomographic models. The increased resolution implies sharper lateral velocity gradients, that allow us to quantitatively compare tomographic models to the wealth of geological information available from literature, both ground-truthing our observations, and gaining insight on the temporal evolution of the features of the subsurface we image at present.

7.8 Future work

The work presented in this thesis unveils an unexpectedly complex Earth structure, opening up a number of possibilities for future projects. In addition to this, the resolution of the models presented in this work is limited by both data distribution and inversion techniques. Here I will briefly summarise some of the research possibilities that I see in the immediate future to improve on our modelling and analyse its results in further detail.

7.8.1 Multi-resolution parameterisation

Following on the work presented in chapter 5 with the multi-resolution regularisation, implementing a 3D-varying parameterisation would allow to compensate for the very differently sampled areas of the globe. Models computed with both regularisation and parameterisation that vary in 3D could image very different regions with a degree of accuracy that reflects the model sensitivity. This could be achieved by using an a-priori measure of model sampling and resolution, as e.g. the column sums used to scale regularisation in chapter 5, or the resolution maps computed in chapter 3. Another perhaps more interesting approach could be to vary the parameterisation during the 3D inversion, using for instance voronoi cell parameterisation and trans-dimensional Monte Carlo methods; this however would require to add to the large memory footprint of the sensitivity matrix the need for a very large number of inversions, raising much larger computational issues.

7.8.2 3D oceanic age-averaging

In chapter 3 and 5 I extensively used the averaging of V_s profiles along oceanic age (age-averaging) to separate ridge- and hotspot-related average features in the ocean basins. By going a step further, it would be possible to remove the global age-average V_s , indicative of normal cooling, from the 3D model, obtaining a 3D model of the residual V_s anomalies. Such a model could indicate not only features such as hotspots or anomalously thick oceanic lithosphere, but also give us insights on the differences in thermal structure between oceans with sometimes wildly different spreading regimes. Finally, identifying the average normal cooling for each ocean would allow to compute more accurate age-average V_s residuals, extracting even the more feeble anomalous signals from the models.

7.8.3 Correlation between cratons and geological proxies

In chapter 4, we correlated out tomographic images with kimberlites as proxies for cratons in Africa. In other areas of the globe however, the distribution of kimberlites is largely unknown due to the uneven geological sampling (e.g. South America, Antarctica). By gathering a compilation of other high-pressure minerals such as garnets, it would be possible to broaden our analysis to other areas of the globe and possibly improve on the ones we analysed. In addition to this, using a variety of proxies for the presence of cratonic lithosphere in the past could help us to understand the mechanisms with which the cratonic lithosphere is recycled into the convecting mantle.

7.8.4 Joint, multimodal inversion of 1-station and 2-station dispersion curves

Phase velocity maps presented in chapter 6 rely solely on teleseismic measurements computed by AMI, which are mostly sensitive to periods greater than 20 s. By augmenting the dispersion curves database with data from inter-station cross-correlation (Lebedev, Meier, and Hilst, 2006), it would be possible to broaden the frequency band to periods < 10 s and add additional data in already sampled but poorly constrained short periods. This would allow us to image the details of the crust, poorly constrained by AMI measurements only. However, the different sensitivity of the two datasets to each periods needs to be taken into account, and the data weighted accordingly. Additionally, chapter 6 presents the inversion of fundamental-mode measurements only; by computing phase velocity maps with higher-mode dispersion curves—already produced by AMI—it would be possible to gather information on the deep upper mantle and transition zone. The structural information from the crust to the mantle transition zone could then be extracted by jointly inverting the augmented fundamental-mode phase velocity maps with the higher-mode ones for the distribution of S-wave velocities in 3D.

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