**Crustal and upper mantle
structure of Botswana: is
Botswana rifting?**

Islam Fadel

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UNIVERSITY OF TWENTE.

ITC

FACULTY OF GEO-INFORMATION SCIENCE AND EARTH OBSERVATION

**CRUSTAL AND UPPER MANTLE STRUCTURE OF
BOTSWANA: IS BOTSWANA RIFTING?**

DISSERTATION

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the degree of doctor at the University of Twente,
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Islam Fadel
born on April 4, 1986
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To my family

Summary

Our knowledge of the Earth's crust and upper mantle is derived largely from geophysical observations (e.g., seismology, gravity, etc.). Constraining the physical conditions of the outer layers of the Earth is crucial to improving our fundamental understanding of the part that directly influences our societies. This will help us to predict and mitigate the challenging geohazards e.g., earthquakes and volcanoes. Moreover, it will increase our capability of targeting and extracting the natural resources.

Despite the exponentially increasing volume of high-quality geophysical data emerging from dense seismological networks, wide regions, including Africa, still suffer from the lack of seismological coverage. Moreover, the lack of seismological data restricts our ability to fully exploit the state-of-the-art potential field data from recent satellite missions e.g. GOCE. This is due to the dependency of potential field data on seismological models as crucial a priori information. This thesis focuses on these two components; the first is to bring insight into the crustal and upper mantle structure of Botswana that is poorly covered by seismological studies, and the second investigates two methodological developments to integrate seismological and gravity data.

In the first part of the thesis, the crustal and upper mantle structure of Botswana was investigated using new data from the recently deployed NARS-Botswana temporary seismological network. Receiver function technique was used to derive the crustal thickness map of Botswana. Moreover, surface wave tomography from ambient noise and earthquakes was used to derive the 3D shear wave velocity structure of Botswana's lithosphere. The results showed three distinctive features: 1) the deep sedimentary basins, 2) the extension of the East African Rift System (EARS) in the upper mantle and 3) the cratonic blocks. The Passarge and Nosop basins have a remarkable low-velocity signature in the crustal structure of Botswana that extends deeper than 10 km. At the upper mantle, the EARS appears as the lowest shear wave velocity anomaly in Botswana at the northern border with a thin crustal thickness. Low-velocity expressions extend from the EARS until the borders of the Kalahari and Congo cratons. Moreover, the low-velocity extensions seem to corrode the margins of the cratons by low-velocity finger-like struc-

tures. The low-velocity expressions penetrate the crust at weak zones causing noticeable earthquake activity at the Okavango Rift Zone and the border of the Kaapvaal Craton. The extensional mechanisms of the earthquakes, and especially the large 6.5 Mw earthquake of April 3, 2017, suggest rifting in both areas with a possible connection through central Botswana that is supported by the thin crust and high V_p/V_s ratio. The Kaapvaal and the Zimbabwe cratons show high shear velocities and deep keels that extend deeper than 200 km depth. Two deep cratonic roots are observed in the northwest at the margin of the Congo Craton and in the southwest at the Rehoboth Province. The margin of the Congo Craton shows a deep cratonic keel as well, which extends deeper than 200 km depth. The cratonic root at the Rehoboth Province supports the presence of a buried microcraton.

In the second part of the thesis, we introduced two different strategies to integrate seismological and gravity observations. The object-based approach is a good option in case of the availability of 3D seismic velocity models from regional or local observations. The approach gives the opportunity to extract the 3D subsurface structure from any available 3D seismic tomography model in a semi-automatic way. This provides a reproducible and robust extraction process that reduces the subjectivity of the interpretation of the 3D subsurface models. The extracted objects can be later inverted for their density contrast using a linear least-square inversion scheme. However, the estimated density contrasts resulting from the inversion are sensitive to minute variations in the 3D model. This suggests that the density contrasts of the objects should be constrained by boundary values that can be derived from the seismic velocities using empirical relations or mineral physics using thermodynamical modelling based on Gibbs free energy minimization. The second strategy is based on jointly inverting seismological and gravity observations. we developed a novel approach that, for the first time, enable the joint inversion of seismic surface wave and the GOCE gravity gradients. The approach showed promising results using synthetic models. Further application using real data will open the door to derive new generation of lithospheric models that can image both density and seismic velocities by combining GOCE and seismological data.

To summarize, this thesis brings insight into the crustal and upper structure mantle of Botswana. The updated crustal thickness map and the 3D shear wave velocity model provided a crucial knowledge about the extension of the East African Rift System, the distribution of the cratonic segments, and the extent of the sedimentary basins. Moreover, the two developed approaches to integrate seismological and gravity information open the door to exploit the full potential of the state-of-the-art GOCE gravity data. We hope that the new findings and the developed approaches will help us to better understand the geodynamics of Africa and facilitate the imaging of the density structure of the Earth.

Samenvatting

Onze kennis van de aardkorst en de bovenmantel is grotendeels afgeleid van geofysische waarnemingen (bijvoorbeeld seismiek, zwaartekracht, etc.). Het vaststellen van de fysische eigenschappen van de buitenste lagen van de aarde is cruciaal om ons fundamentele begrip te verbeteren, omdat de buitenste laag van de Aarde direct onze maatschappij beïnvloed. Deze kennis helpt ons om betere voorspellingen te maken van natuurrampen, zoals aardbevingen en vulkaanuitbarstingen, en maatregelen te kunnen treffen die het risico verminderen.

Ondanks de exponentiële groei van hoge kwaliteit geofysische gegevens door het verschijnen van een hoge dichtheid van seismologische netwerken, zijn sommige gebieden, waaronder Afrika, nog altijd verstoken van seismologische metingen. Tegelijkertijd beperkt dit gebrek aan seismologische meetgegevens onze mogelijkheden om de vele state-of-the-art gegevens van nieuwe satelliet missies, zoals GOCE, ten volle te benutten. De meetgegevens zijn cruciaal omdat ze als a priori informatie functioneren voor seismologische modellen. Dit proefschrift richt zich op beide elementen: Het eerste deel richt zich op het verkrijgen van meer inzicht in de seismologische structuur van de aardkorst en bovenste aardmantel in delen van Botswana die nog niet verkend waren. Het tweede deel beschrijft twee nieuw ontwikkelde methodes om zwaartekracht en seismologisch waarnemingen te integreren.

In het eerste deel van het proefschrift, is de structuur van de aardkorst en bovenste deel van de aardmantel van Botswana onderzocht met nieuwe gegevens van het NARS-Botswana tijdelijke seismologisch netwerk dat recentelijk geïnstalleerd is. De receiver functie techniek is gebruikt om een kaart van de aardkostdikte voor Botswana af te leiden. Verder is de oppervlaktegolftomografie van de omgevingsruis en aardbevingen gebruikt om de 3D structuur van de transversale golfsnelheid van de lithosfeer te bepalen. De resultaten laten 3 kenmerken zien: 1) diepe sedimentbekkens, 2) het spreiden van het Oost Afrikaans Rift systeem (EARS) in de bovenste aardmantel, en 3) de cratons. De Pasarge en Nosop bekkens hebben een uitgesproken lage-snelheid kenmerk in hun aardkost structuur, die dieper is dan 10 km. In de bovenste aardmantel is de EARS is zichtbaar in de laagste transversale snelheidsanomalie aan

de noordelijke grens van Botswana, alwaar de aardkorst dun is. Lage-snelheid kenmerken strekken zich uit van de EARS tot aan de grenzen van de Kalahari en Congo cratons. Tegelijkertijd lijken de lage-snelheid kenmerken de grenzen van de cratons te corroderen in de vorm van vingerachtige structuren. De lage-snelheid kenmerken penetreren de aardkorst op zwakke plekken en veroorzaken een waarneembare aardbevingsactiviteit in de Okavango Rift zone en de grens van het Kaapvaal craton. Het mechanisme van aardbevingen die door het oprekken van de aardkorst veroorzaakt worden, en dan met name de grote 6.5 Mw aardbeving van 3 april 2017, suggereert een verdere ontwikkeling van de rift in beide gebieden, met een mogelijke verbinding door centraal Botswana. Deze veronderstelling wordt ondersteund door het waarnemen van een dunne aardkorst en een hoge V_p/V_s verhouding. De Kaapvaal en de Zimbabweaanse cratons laten hoge snelheden zien en een wortel van het gebergte van meer dan 200 km dik. Twee diepe cratonische wortels zijn waargenomen in het noordwesten aan de rand van de Congo craton en in het zuidwesten van de Rehoboth provincie. De rand van het Congo craton laat ook een diepe wortel zien van meer dan 200 km diep. De wortel van de Rehoboth provincie ondersteunt het concept dat er een ondergegraven microcraton aanwezig is.

In het tweede deel van de thesis hebben we twee verschillende strategieën geïntroduceerd die zwaartekracht en seismische waarnemingen integreren. In het geval dat 3D seismische snelheid modellen van regionale of lokale waarnemingen beschikbaar zijn, zijn object-gebaseerde methodes een geschikte keuze. Deze methodes geven de mogelijkheid om semiautomatisch een 3D ondergrond structuur af te leiden van een 3D seismisch tomografisch model. Dit is een reproduceerbare en betrouwbare extractie dat subjectiviteit in de interpretatie van 3D ondergrond modellen vermindert. Objecten kunnen later geïnverteerd worden door het verschil in dichtheid in een lineaire least-squares inversie te gebruiken. Echter, de zo verkregen dichtheidsverschillen zijn gevoelig voor kleine variaties in het 3D model. Dit suggereert dat de verkregen dichtheidsverschillen gelimiteerd zouden moeten worden door randvoorwaarden te stellen die uit seismische snelheden afgeleid kunnen worden, hetzij door empirische relaties of door mineralenfysica door thermodynamisch modelleren gebaseerd op Gibbs free energy minimalisering. Een tweede strategie is gebaseerd op een gezamenlijk inverteren van seismische en zwaartekracht waarnemingen. We hebben een nieuwe methode ontwikkeld die, voor het eerst, een inversie van seismische golven en GOCE zwaartekracht gradiënten mogelijk maakt. Wanneer toegepast op synthetische modellen, geeft deze methode goede resultaten. Een toepassing op echte modellen zal in de toekomst de mogelijkheid geven om een nieuwe generatie lithosfeer modellen af te leiden die zowel zwaartekracht als seismische snelheid modellen verkregen door het combineren van GOCE data en seismische data inzichtelijk te maken.

Samengevat geeft dit proefschrift een nieuw inzicht in de aardkorst en het bovenste deel van de mantel van Botswana. De vernieuwde kaart van de aardkostdikte en het 3D model van de transversale golfsnelheid verschaft cruciale kennis over het oprekken van de Oost Afrikaanse Rift systeem, de posities van cratonische segmenten en de grootte van de sedimentatiebekkens. De twee ontwikkelende methoden om seismologische en zwaartekracht informatie te integreren opent deuren om het potentieel van GOCE zwaartekrachtmetingen ten volle te exploiteren. Wij hopen dat deze nieuwe bevindingen en de ontwikkelde methoden het in kaart brengen van de structuur van de aarde zal faciliteren en ons zal helpen om het begrip van de geodynamica van Afrika te vergroten.

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Introduction

1

1.1 Scientific Summary

This thesis has a technical development component on integration of satellite gravity and seismological data, and a practical application component on crust and upper mantle structure for Botswana. Botswana is one of the least studied regions of Africa and the world. It covers the Congo Craton in the northwest and the Kalahari Craton in the east with in between a series of mobile belts and deep sedimentary basins that developed during geologic times. However, there is limited and sparse information about the subsurface structure in the area that can describe the extent and the interaction between these different structural settings. In the first part of this thesis, I will explore the 3D subsurface structures and the tectonic settings of the crust and upper mantle. This will be developed in three phases. Firstly, an improved Moho map will be produced using receiver functions. Secondly, the crustal structure will be revealed by ambient noise surface wave tomography. Finally, a 3D shear wave velocity of the crust and upper mantle will be developed using ambient noise and earthquake surface wave dispersion measurements. In the second part of the thesis, I will present a methodological development for the cooperative inversion of surface wave information and GOCE gravity total fields and gradients. This will include two strategies. The first is a sequential inversion approach of the satellite gravity measurements and 3D shear velocity tomography model using a 3D object-based image analysis technique that was applied to the East African Rift System. The second is a joint inversion approach that will be presented and evaluated using synthetic tests.

1.2 Introduction in African geodynamics

Africa is a vast continent with large cratonic fragments and orogenic belts only in the extreme north and south. For the last two billion years, the movements of the craton nuclei have been persistently epeirogenic, giving a structural pattern of broad basins separated by irregular swells where plateaus and swells have been intermittently uplifted and denuded. Special in this case is the southern part of Africa which is influ-

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enced by the African superswell (Figure 1.1). The African superswell is an extraordinary uplift of the African continent, particularly its southern half; southern Africa lies, on average, a full kilometre above sea level, with seemingly anomalous uplifts extending well into the South Atlantic Ocean. The superswell is a relatively recent phenomenon, probably developed between 5 and 30 million years ago. A proposed cause of the superswell is a mantle plume, though this hypothesis is controversial and the origin of the superswell remains an area of active research (Nyblade and Robinson, 1994; Globig et al., 2016).

Of particular interest are the cratonic margins and some intracratonic domain boundaries, which have played a significant role in the tectonics of Africa. Archaean cratons are the stable remnants of Earth's early continental lithosphere, and their structure, composition and survival over geological time make them unique features of the Earth's surface (James et al., 2001). The lithospheric architecture of Africa consists of several Archean cratons and smaller cratonic fragments, stitched together and flanked by younger fold belts (Begg et al., 2009). These boundaries along the larger cratons have been the source of localised successive cycles of extension, rifting, and renewed accretion. They are underlain by a geochemically depleted, rigid, and mechanically robust lithospheric mantle; these cratonic roots have steep sides, extending in some cases to ≥ 300 km depth. Beneath smaller cratons (e.g., Kaapvaal) extensive refertilization has reduced the lateral and vertical extent of strongly depleted lithospheric mantle (Begg et al., 2009). The cratons, therefore, play a crucial role in African tectonics, and better understanding of their thickness and extent will provide insight into earlier and ongoing mantle dynamics.

Another major feature contributing to African geodynamics is the East African Rift System (EARS). This rift system comprises several discrete rift sectors with and without magmatic surface expressions. The EARS developed after the onset of the Miocene (20-25 Ma) (Ebinger, 2005). It has eastern and western rift branches that surround the Tanzania Craton (Figure 1.1). The rift extends to the south until the border of Botswana, and the Okavango Rift Zone (ORZ) in northern Botswana is widely accepted to be its southwestern terminus (Kinabo et al., 2007). However, the subsurface expression of this terminus is unclear and whether the EARS actually ends at the ORZ without further southern continuation is still a hot research topic.

Botswana is one of the least studied areas in Africa. Although the area is rich in mineral resources and especially diamond (Schlüter, 2008), the 3D subsurface structure and the deep tectonic settings are poorly resolved. Botswana covers two main cratonic provinces; the Kalahari Craton to the east and southeast, consisting of the Kaapvaal and Zimbabwe cratonic segments, and the Congo craton in the northwest. Little is known about the extent of the three cratons, their structure and the

Figure 1.1 Topographic map of Africa with the blue contour outlining Botswana. The black lines represent coast lines. WRB and ERB are the western and eastern rift branches, respectively. TZO is the Tanzania Craton. ORZ is the Okavango Rift Zone.

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geodynamical role of the mobile belts in between. Earthquake activity can be found in many parts of the country but is mainly occurring in the north and east of the country, in both mobile belt areas (like the Okavango Rift Zone in the north-northwest) but also the cratons (like the northwestern Kaapvaal). To date, several studies (e.g., Fouch et al., 2004; Yang and Ritzwoller, 2008; Kgaswane et al., 2009; Li, 2011; Adams and Nyblade, 2011; Youssof et al., 2015) have investigated the southeastern part of Botswana. However, the crust and upper mantle structure of the rest of Botswana are poorly resolved since only a few studies explored these regions (e.g., Wright and Hall, 1990; Hutchins and Reeves, 1980; Miensoopust et al., 2011a; Khoza et al., 2013a,b; Muller et al., 2009; Jones et al., 2013). The southwestern part of the country remains a bit of an enigma with a possible hidden microcraton underneath the Rehoboth province but for which the actual structure remains elusive (Begg et al., 2009; Wright and Hall, 1990). Therefore, there are fundamental questions about the crustal and upper mantle structure that need to be addressed, including:

- The crustal thickness variations across Botswana and their relation to the tectonic settings.
- The nature and the extent of the East African Rift System in Botswana, in particular, where it actually ends.
- The cause behind the noticeable earthquake activity in the north-western Kaapvaal block.
- The deep settings of the boundary between the Rehoboth Province and the Kheis belt and the northwestern Kaapvaal block.
- The velocity structure of the Rehoboth Province.

Since November 2013, a temporary array of seismometers was deployed in Botswana to address the open questions about the crustal and upper mantle structure. This NARS-Botswana network consists of 21 broadband seismic stations distributed over the whole country. It strategically covers the different geological and tectonic units and complements the existing stations (Fig. 1.2). It fills a gap in station coverage for southern Africa, extending both west and south from previous temporary seismic deployments in Zambia, Zimbabwe and South Africa. In the first part of the thesis, we will use the data from NARS network to explore the crustal and upper mantle structure underneath Botswana. Firstly, we will estimate the crustal thickness using seismological receiver functions (RFs). Then, seismic surface waves obtained from ambient noise will be used to explore the crustal structure. Finally, seismic surface wave dispersion measurements obtained from ambient noise and earthquake data will be used to derive the 3D shear wave velocity structure underneath Botswana.

Seismic techniques can provide detailed images of seismic velocity variations. However, they have limited sensitivity to density. Recent stud-

Figure 1.2 The NARS-Botswana seismological array. The yellow triangles are NARS stations. The cyan triangle is the Global Seismological Network station LBTB. The green triangle is the AfricaArray station MAUN.

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ies have shown that density is required to distinguish between compositional and temperature effects within the Earth and therefore represents a crucial physical parameter for solid Earth geophysics. The Rayleigh surface wave is well-known to be sensitive to density, albeit only weakly. However, estimating density using surface waves suffers from strong nonlinearity (Takeuchi and Saito, 1972). Recent pioneering work by Blom et al. (2017) has shown that full waveform inversion can theoretically be used to recover the subsurface density structure regardless of the small contribution of the density to the seismic wave field. Moreover, they showed that adding gravity information to the joint inversion process does not improve the inversion process on the global scale (maximum depth in their study is 3000 km depth) due to the shallow nature of the sensitivity kernels of the gravity data and the inherent non-uniqueness in the potential field inversion. On the crustal and upper mantle scale, Xing et al. (2016) showed that an improper shallow density structure could lead to errors in shear velocity models at middle and lower crustal depths. Based on the previous findings, it is clear that there is a need for a methodology that can constrain the density structure at crustal and upper mantle depths. Recent studies found novel approaches to combine gravity with seismic data to estimate the subsurface density structure on exploration scale (Moorkamp et al., 2011) and for the crust and the uppermost part of the mantle (Maceira and Ammon, 2009). In the second part of the thesis, we will therefore explore different strategies to estimate density structure using global gravity data derived from the recent GOCE satellite mission. The GOCE satellite was a unique satellite (the GOCE mission ended on 13.11.2013) that was equipped with a three-axes gradiometer and flying at an altitude of 260 km and less. GOCE provided the most detailed measurements of Earth's gravity from space ever by acquiring gravity gradients, i.e., the three-dimensional second derivatives of the gravitational potential. Firstly, we will describe a novel object-based approach to use 3D shear wave seismic velocity models to constrain the inversion of the gravity data. Then, we will apply this methodology to the East African Rift System. Finally, we will present a novel inversion technique that was designed to jointly invert seismic group and phase dispersion measurements, and satellite gravity and gravity gradient data to invert for the 3D shear-wave velocity and density structure.

1.3 Research objectives

There are two general objectives of the thesis. Firstly, exploring the crustal and upper mantle structure of Botswana. Secondly, exploring the possible strategies to integrate seismic surface waves with GOCE gravity and gravity gradients. The following objectives will be addressed:

1. Mapping crustal thickness variations across Botswana.

2. Estimating group and phase velocity maps using ambient noise.
3. Integration of interstation phase velocity maps obtained from earthquakes and ambient noise, and inversion of these data to image the 3D crustal and upper mantle shear wave velocity structure underneath Botswana.
4. Designing a sequential inversion methodology to use 3D shear wave velocity models to constrain the inversion of GOCE gravity measurements.
5. Designing a joint inversion scheme of surface wave phase and group velocity and GOCE satellite gravity measurements.

1.4 Structure of the thesis

The thesis consists of in total ten chapters. Apart from the introduction, study area, data and methods, and synthesis chapters, six chapters have either been published or submitted in peer-reviewed ISI journals (chapter three, four and seven) or are ready for submission (chapter five, six and eight)

- Chapter 1 presents the general research framework of the thesis and the research objectives. It also includes the description of the thesis structure.
- Chapter 2 gives a review of the tectonic setting of Botswana in addition to a brief review of recent geophysical studies across the country.
- Chapter 3 presents the NARS-Botswana array and the installation process in addition to a brief summary of the data and methods used throughout the thesis.
- Chapter 4 presents the receiver function study using data from the NARS array in Botswana. The results show variations in crustal thickness and V_p/V_s ratio across Botswana. The results also compared to previously published receiver function studies in the region.
- Chapter 5 is about ambient noise surface wave tomography of Botswana where the crustal structure is explored from group and phase velocity maps.
- Chapter 6 provides the primary results of the surface tomography study of Botswana. The chapter presents the 3D shear velocity model using integrated surface wave dispersion data obtained from ambient noise and earthquakes.
- Chapter 7 presents a novel implementation of Object-Based Image Analysis (OBIA) in 3D geophysical modelling in which 3D subsurface structure is extracted from the 3D shear wave velocity model of the East African Rift System.

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- Chapter 8 implements OBIA to reduce the non-uniqueness of satellite gravity inversion using the a priori information from the 3D shear wave model of the East African Rift System.
- Chapter 9 describes a methodological development of joint inversion of surface waves and satellite gravity measurements.
- Chapter 10 gives a summary of the thesis and potential future directions for research.

Botswana tectonics and geophysical studies

In this chapter, I will give a succinct overview of the main geological units and previous regional geophysical studies of Botswana. The lithospheric structure of Botswana encompasses diverse tectonic units that play a key role in the geodynamics of Africa. Botswana covers two stable cratonic blocks, the Kalahari Craton in the east and southeast and the Congo Craton in the northwest. In between these blocks, there are mobile belts and sedimentary basins that developed through a series of amalgamation and rifting processes (Figure 2.1 and Table 2.2). In the following sections, the main tectonic units will be described briefly. After that, an overview will be given of the main geophysical studies of Botswana.

2.1 Tectonic units

2.1.1 Cratons

2.1.1.1 Kalahari Craton

The composite Kalahari Craton consists of two cratonic blocks, the Kaapvaal Craton and Zimbabwe Craton, with in between the Limpopo mobile belt, which is an exotic crustal block (Begg et al., 2009).

The eastern to southeastern part of the Kaapvaal Craton was first created by intra-oceanic obduction of hydrated oceanic crust followed by granite formation around 3.7-3.2 Ga (Thomas et al., 1993). The oldest crust of this ancient core has an age of 3.7-3.4 Ga (Cahen et al., 1984; Thomas et al., 1993).

The central to southern part of the Kaapvaal Craton was formed due to amalgamation and cratonization of oceanic terrains around 3.3-3.1 Ga. The oldest crust of the Zimbabwe Craton, in its southeastern part, has an age of 3.5-2.8 Ga (Rollinson, 1993; Smith et al., 2009).

During the mid-Archean (3.1-2.9 Ga), rifting occurred and a continental margin developed along the western and northern borders of the Kaapvaal Craton. This was followed by arc- or continent-continent collision around 2.7 Ga (de Wit et al., 1992). Little is known about the western late-Archean block of the Kaapvaal Craton that is separated from the

Figure 2.1 (A) Map showing the main tectonic units adopted from Singletary et al. (2003) and McCourt et al. (2013) together with stations of the NARS-Botswana network (yellow triangles), GSN station LBTB (blue) and AA station MAUN (green). The gray transparent area represents the Okavango Rift Zone. (B) Sedimentary thickness map inferred from aeromagnetic data (modified from Pretorius (1984)) with earthquake distribution (IRIS service using all available catalogs between January 1900 to June 2017). The central Botswana magnitude 6.5 earthquake of 3 April 2017 and its main aftershocks are shown in red.

Cratons	<p>Formation of Kaapvaal Craton 3.7-3.2 Ga</p> <p>Formation of a major part of the Rehoboth Province around the Archean nuclei and collision with the Kaapvaal Craton 2.2-1.9 Ga.</p>	<p>Oldest rocks recorded in the Congo Craton 3.2-2.99 Ga.</p> <p>Amalgamation of the Congo and São Francisco cratonic units ~ 2.05 Ga.</p>	<p>Collision of eastern and western Kaapvaal cratonic blocks forming Colesberg Lineament 2.93-2.88 Ga.</p> <p>The break up of Africa and South America causing the splitting of Congo and São Francisco cratons ~130 Ma.</p>
Mobile belts	<p>Formation of Limpopo belt due to collision between Zimbabwe and Kaapvaal Craton 2.7-2.5 Ga.</p> <p>Kheis Orogeny including reactivation of the zone between Kheis and Kaapvaal Craton ~1.75 Ga.</p> <p>The initiation of the Ghanzi-Chobe Belt in Botswana as a rift basin due to extensional forces as a consequence of the Kibaran Orogeny collisional event a long Namaqua-Natal Belt in South Africa (1.1-1.3 Ga).</p>	<p>Magondi Basin formation ~2.16-2.0 Ga.</p> <p>Reactivation of the Kheis province and its amalgamation to the Rehoboth province in the Kibaran orogeny ~1.2 Ga.</p>	<p>Eburnean orogeny including the amalgamation of Kheis-Okwa-Magondi Belt 2.0-1.8 Ga.</p> <p>Formation of the Damara-Ghanzi Chobe Belt during the Damara Orogeny ~870-550 Ma.</p>
Sedimentary basins	<p>The formation of the Passarge and Nosop sedimentary basins during Neoproterozoic and early Paleozoic.</p>		

Figure 2.2 Table summarizing the main tectonic events that influenced the tectonic units in Botswana.

2. Botswana tectonics and geophysical studies

central mid-Archean Kaapvaal blocks by the magnetic Colesberg Lineament (de Wit et al., 1992).

2.1.1.2 Congo Craton

The southwestern border of the Congo Craton is located in the northwestern part of Botswana (Figure 2.1). The Congo Craton and São Francisco Craton in South America are widely accepted to be amalgamated together as a single cratonic block around 2.05 Ga until the break-up of Africa and South America around 130 Ma (Ernst et al., 2013). The Congo Craton consists of Archean and Paleoproterozoic rock units, where the oldest block at the northwestern end of the craton was formed around 3.2-2.99 Ga (Begg et al., 2009; Ernst et al., 2013).

2.1.1.3 Rehoboth Province

The Rehoboth Province is a poorly resolved region that extends from southwestern Botswana to western Namibia (Begg et al., 2009; van Schijndel et al., 2014). A major part of the province aggregated around an Archean nucleus during the Paleoproterozoic around 2.2-1.9 Ga (Van Schijndel et al., 2011) and then collided with the Kaapvaal Craton at 1.9 Ga (Luchs et al., 2013). In Botswana, the upper crustal part is overlain by the Nosop basin with sediments up to 15 km thick (Pretorius, 1984; Wright and Hall, 1990). The Rehoboth Province under the Nosop basin has been considered as an ancient micro-craton (the Maltahohe Craton) (Begg et al., 2009) or as a deeply seated extension of the Kaapvaal Craton, as interpreted by Wright and Hall (1990) using deep seismic profiling.

2.1.2 Mobile Belts

A number of mobile belts are caught between the previously mentioned cratonic blocks. The reworking along the edges of these cratonic blocks, the amalgamation, and the rifting events started in Archean times and they extend till today with the current incipient rifting in the Okavango Zone. In the following subsections, I will provide a brief description of the mobile belts in Botswana.

2.1.2.1 Limpopo Belt

The Limpopo Belt was formed as a result of a collisional event around 2.7 Ga during the Archean consolidation of the Zimbabwe and Kaapvaal cratons (McCourt and Vearncombe, 1992; Smith et al., 2009). The metamorphism of the older Archean rocks reached its peak around 2.56 Ga in the northern zone and around 2.7 Ga in the southern zone (Begg et al., 2009). The belt may have been reactivated between 2.06 and 2.0 Ga as a consequence of events in the Kheis-Okwa-Magondi Belt and the intrusion of the Bushveld complex in South Africa (2.05 Ga) (Begg et al., 2009).

2.1.2.2 Kheis-Okwa-Magondi Belt

The Kheis-Okwa-Magondi Belt is assumed to be a Paleoproterozoic belt that was accreted on the northwestern border of the Kalahari Craton due to the Eburnean Orogeny around 2.0-1.8 Ga (Thomas et al., 1993; Begg et al., 2009). The Magondi Basin was formed on the western edge of the Zimbabwe Craton around 2.16-2.0 Ga during an intracratonic extension that occurred during the early Proterozoic. Then, it was amalgamated during the Eburnean Orogeny with the Okwa-Kheis Belt to the west of the Archean blocks. After that, the margin of the Kheis Province and the Kaapvaal Craton was deformed during the Kheis Orogeny around 1.75 Ga. Finally, the Kheis Province was reactivated and amalgamated with the Rehoboth Belt during the Kibaran Orogeny around 1.2 Ga ago (Thomas et al., 1993).

2.1.2.3 Damara-Ghanzi Chobe Belt

The Damara-Ghanzi-Chobe Belt was formed during the Damara Orogeny which was a part of the Neoproterozoic Pan-African tectono-thermal event around 870-550 Ma due to the collision between the Kaapvaal and Congo cratons ((Haddon, 2005, section 3.5.3.2), and (Kröner and Stern, 2004)). The Okavango Rift Zone is located within the Damara-Ghanzi-Chobe belt, and is considered as the southwestern terminus of the East African Rift System. The Okavango Rift Zone is widely accepted as an incipient rift system (Modisi et al., 2000; Kinabo et al., 2008).

2.1.3 Passarge and Nosop sedimentary basins

Two sedimentary basins are caught between the Kheis-Okwa Magondi and Damara-Ghanzi-Chobe belts. Their locations reflect the gradual chronological tectonic development in Botswana moving from old cratons to mobile belts and ending by sedimentary basins caught between the mobile belts.

The Passarge basin in central Botswana is considered to be a part of the Ghanzi-Chobe Belt that is bordered by the Kalahari Suture Zone (KSZ) in the south. The KSZ is a Paleoproterozoic major thrust zone that is related to the formation of the Kheis-Magondi Belt, and it mainly separates the Archean blocks from the Proterozoic belts (Reeves, 1978). It was reactivated during the Mesoproterozoic and Neoproterozoic as a rift dipping to the northwest. During Neoproterozoic and early Paleozoic times, the Passarge basin was formed and filled by a thick layer of sediments from the Ghanzi Group which reaches down to 15 km depth (Haddon, 2005; Pretorius, 1984) (Figure 2.1-B).

The Nosop Basin in the southwest of Botswana is part of the Rehoboth terrane with a sedimentary thickness that reaches up to 15 km (Pretorius, 1984) (Figure 2.1-B). It was formed by the accumulation of Nama Group overlaying the Ghanzi Group sediments (Key and Ayres, 2000).

2.2 Previous Geophysical Studies in Botswana

In this part, I will give a brief review of the recent geophysical studies that were carried out within or including part of Botswana. I will start with the seismological studies, and after that magnetotellurics, gravity and magnetics, and then I will end with the recent joint interpretation and inversion effort for the study area.

Figure 2.3 Regional tectonic map showing the distribution of the recent geophysical measurements within or including part of Botswana over the major tectonic terrains in southern Africa modified from Leseane et al. (2015).

2.2.1 Seismological Studies

An early fundamental finding on Botswana tectonics was obtained from a seismological study by Reeves (1972) about rifting in Kalahari. The study reported a high level of seismicity within the Okavango Rift Zone

(ORZ), and it was the first to suggest rifting in this area as an extension of the East African Rift System.

Three decades later, a number of studies started the investigation of the eastern and southeastern part of Botswana, covering parts of the Kaapvaal Craton and Limpopo Belt. The Southern Africa Seismic Experiment (SASE), a temporarily seismological network (April 1997 to July 1999, Carlson et al. (1996)), and the Africa Array initiative (Nyblade et al., 2008) allowed imaging of the crustal and upper mantle structure of southern Africa in high resolution (e.g., James et al., 2001; Nguuri et al., 2001; Fouch et al., 2004; Yang and Ritzwoller, 2008; Li, 2011; Adams and Nyblade, 2011; Delph and Porter, 2015; Youssof et al., 2015) (Figure 2.3).

More recently, a temporary (2013-2015) array was deployed across the ORZ as part of the Seismic Arrays for African Rift Initiation (SAFARI) project (Gao et al., 2013b) (Figure 2.3). SAFARI brought new insights into the geodynamic development of the ORZ, its crustal and upper mantle structure, and anisotropy of the upper mantle (Yu et al., 2015a,b,c, 2017). The seismic results produced puzzling observations regarding the nature of the geodynamic forces driving ORZ given the lack of deep thermal anomalies in the upper mantle suggested by the normal thickness of the mantle transition zone in the area (Yu et al., 2015b). On the other hand, a 4-5 km thinned crust with high V_p/V_s ratio was reported by Yu et al. (2015c). Recent P-wave tomography by Yu et al. (2017) showed evidence for a deep root of the Congo Craton on the border of the ORZ. Moreover, they showed the presence of a low-velocity anomaly underneath the ORZ that was interpreted as a result of decompression melting due to lithospheric thinning. The different pieces of evidence support a passive rifting process due to differential movement of the cratonic blocks facilitated by fluids that reactivate a pre-existing weak zone within the Damara Belt. This is supported by geodetic observations (Malservisi et al., 2013), receiver functions (Yu et al., 2015b), and P-wave tomography (Yu et al., 2017). However, Yu et al. (2015b) also observed an anomalously thin mantle transition zone on the edge of their model beneath central-west and central Botswana that is associated with a positive thermal anomaly. The limited network coverage did not allow further investigation of this anomaly.

The previous seismological networks have been limited to the northern and southeastern parts of Botswana. Only a single active seismic study, to my knowledge, investigated the southwestern part of Botswana, the area of the Nosop Basin, using deep seismic profiling (Wright and Hall, 1990). The results showed that the Nosop Basin is very deep, filled with up to 15 km of sediments. The high velocity underneath the sedimentary basin was interpreted as a signature of the extension of the Kaapvaal Craton. The limited geophysical data coverage, especially of seismic observations, has been the main obstacle to understand the tectonic setting in central and southwestern of Botswana, in particular in

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relation to the lower crust and upper mantle structure. The main aim of this thesis is to provide an image of the subsurface structure beneath Botswana with the nationwide NARS-Botswana seismological network that will be described in the next chapter.

2.2.2 Magnetotelluric (MT)

The Southern African Magnetotelluric Experiment (SAMTEX) is a large project carried out between September 2003 to June 2008 (Figure 2.3) (Jones et al., 2009). The main aim of the project is to image the crustal and upper mantle 3D electrical structure in South Africa, Botswana, and Namibia. The project collected data from 750 MT sites. The results from the project gave new insight into the electrical structure of southern Africa. Several studies were carried out using the MT data across the borders or partially covered parts of Botswana. However, the data of central Botswana are not yet published.

Muller et al. (2009) processed SAMTEX data from a 1400 km long profile from Namibia through Botswana to South Africa. Their 2D MT inversion results gave first insight into the deep electrical structure of the Rehoboth province. The results show that the Rehoboth province has a similar electrical structure as the western Kaapvaal block with a thermal lithosphere of around 180 km, although xenolith data suggest that it may have been thicker prior to the Mesozoic eruption of the kimberlites. Miensopust et al. (2011a) carried out a 2D MT inversion to image the crustal and upper mantle electrical structure in northeastern Botswana using a 2D profile of SAMTEX data that striking through the Zimbabwe Craton, the Magondi belt and the Ghanzi-Chobe Zone. Their results show the presence of two middle to lower crustal conductors within the Magondi belt. A resistive but relatively thin (about 180 km) lithosphere is found for the Ghanzi-Chobe Zone, whereas the thickness of the Zimbabwe Craton is approximately 220 km. The study also found a resistive crustal structure within the Okavango Dyke Swarm located within the Limpopo belt. Khoza et al. (2013a) imaged the Limpopo Belt with 2D MT inversion. They reconcile the MT and metamorphic data within the tectonic framework of subduction, collision and transpression. Khoza et al. (2013b) used 3D MT to image the area between northeastern Namibia to southwestern Botswana and reported a resistive thick (250 km) and cold Congo cratonic block that extends into Botswana. On the other hand, the Damara Belt and Ghanzi-Chobe Zone are thinner and more conductive than the adjacent Congo Craton.

2.2.3 Gravity and Magnetics

A compilation of aeromagnetic and gravity data across Botswana was presented by Hutchins and Reeves (1980) and Reeves and Hutchins (1982). Their maps provided fundamental knowledge about the tectonic units within the country. Leseane et al. (2015) used aeromagnetic and

gravity data to explore the thermal as well as the Moho depth structure of the incipient Okavango Rift Zone (ORZ). They converted their Curie Point Depth map, obtained by 3D magnetic inversion, into heat flow in order to map the thermal variations of the ORZ. 2D spectral analysis of gravity data allowed estimation of Moho depth. The results show high heat flow ($> 50 \text{ mWm}^2$) and shallow Curie Point Depth values ($< 15 \text{ km}$) combined with a thin continental crust ($\sim 30 \text{ km}$). Their results were interpreted in terms of the MT study by Khoza et al. (2013b) that highlights the passive nature of the rifting process of the ORZ. They concluded that hot mantle fluids facilitate the rifting process by upward movement into the weaker zone of the crust.

2.2.4 Joint interpretation and inversion effort.

The SAMTEX MT experiment and the growing number of seismological arrays (Fishwick, 2010) opened the doors for a combined interpretation effort between the different geophysical datasets available. Jones et al. (2013) established a linear velocity-conductivity relationship for southern Africa using a 3D shear wave tomography model from seismic surface wave analysis and a resistivity image of the 100 km depth map obtained from SAMTEX MT data. Their linear relation was able to explain 80% of the resistivity and shear wave velocity structure at 100 km depth for southern Africa. Moreover, Fullea et al. (2011) applied integrated geophysical and petrological modelling on 1D profiles for the Rehoboth Province and the southwestern block of the Kaapvaal Craton in which they incorporated elevation, heat flow and MT and xenolith data.

NARS-Botswana, Data, and Methods

3.1 The NARS-Botswana Seismological Network

This section describes the NARS-Botswana seismological array. The text for this chapter is largely based on the descriptions of the array and the installation process as provided on the official NARS Botswana website (NARS-Website, 2013) and the figures are from the website.

NARS-Botswana is a temporary seismological array that has been deployed in Botswana to image the crustal and upper mantle structure. The network consists of 21 broadband seismic stations distributed over the whole country. It strategically covers the different geological and tectonic units. Moreover, it complements the existing stations MAUN from AfricaArray (AA), and LBTB from the Global Seismic Network (GSN) (Fig. 3.1 and Table 3.1) and its spacing is such that it is complementary to the resolution of the satellite gravity data from the GOCE satellite. The network is a natural extension of previous temporary AfricaArray deployments in Zambia, Zimbabwe, Namibia and South Africa. It fills a gap in station coverage for southern Africa, extending both west and south from these previous temporary seismic deployments. In the following subsections, a brief description is given of the different installation phases and the instruments used.

3.1.1 Site selection and preparation

The network uses broadband surface seismometers. Site selection was therefore based on standard criteria (Trnkoczy et al., 2002). Any site should limit as much as possible the influence of cultural noise (villages, roads, human activity) but also provide proper safety and shielding for temperature. It should be (relatively) easy to reach for installation and servicing and provide a stable underground for the seismometer. A potentially good site for a broadband seismic station is therefore on hard rock and far away from the cultural noise. However, in most places in Botswana, largely covered by desert sands, places have soft and

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Figure 3.1 The NARS-Botswana seismological array.

Table 3.1 The NARS-Botswana stations and their locations.

Station	Location	Sensor	Latitude (°N)	Longitude (°E)	Altitude (m)	Installation (year.daynr)
NE201	Sekoma	Trillium-120p	-24.51535	23.93279	980	2013.324
NE202	Lokgwabe	Trillium-120p	-24.11352	21.78230	1153	2014.011
NE203	Kole	Trillium-120p	-22.99307	20.19555	1313	2014.015
NE204	Xaudum	STS-2	-18.53956	21.33822	1060	2015.077
NE205	Selinda	Trillium-120p	-18.62089	23.50048	961	2015.241
NE206	Kasane	Trillium-120p	-17.80017	25.16189	1006	2015.239
NE207	Qangwa	Trillium-120p	-19.52957	21.17400	1094	2015.327
NE208	Khwee	Trillium-120p	-21.94641	25.44691	1083	2014.044
NE209	CKGR	Trillium-120p	-21.40355	23.77177	1005	2015.073
NE210	Groot-Laagt	STS-2	-21.36192	21.21571	1198	2015.071
NE211	Kacgae	Trillium-120p	-22.85382	22.20679	1153	2013.328
NE212	Kaudwane	Trillium-120p	-23.38039	24.66079	1038	2014.046
NE213	Phepeng	Trillium-120p	-25.47553	22.85729	1030	2013.327
NE214	Gumare	STS-2	-19.38931	22.16246	985	2015.326
NE215	Bottlepan	STS-2	-18.78417	25.19601	1035	2015.324
NE216	Phuduhudu	STS-2	-20.19568	24.53719	956	2014.040
NE217	Borolong	Trillium-120p	-21.09968	27.33424	1047	2015.236
NE218	Sowa	Trillium-120p	-20.56285	26.21785	941	2014.041
NE219	Moremi	Trillium-120p	-22.56967	27.44682	911	2015.235
NE220	Lephepe	Trillium-120p	-23.36303	25.85961	1020	2013.053
NE221	Mmakgori	Trillium-120p	-25.81185	24.80085	1158	2013.320

sandy soil, and due to the remoteness of most sites the possibility of the equipment theft or vandalism is a great concern. The installation team chose to install most stations on the property of primary schools which have 24/7 guarding and are well distributed across the country, taking the higher ambient noise for granted (Figure 3.2). Primary schools in Botswana are located on large plots, and we found at most places relatively quiet locations sufficiently away from school activities. All locations are away from main roads. Most areas in Botswana have thick sandy soil, and only a few of them have hard rock or clay. We have developed our own design of the station sites to overcome the lack of outcrops in the country (see next section).

Figure 3.2 Example of a good site for station NE218 (Sowa) in a school garden.

3.1.2 Construction

As mentioned before, Botswana has very few locations with hard rock outcrops; most locations are covered with a few 100 meters of sandy soils. We have applied two different siting approaches. At the sandy locations, we used our "Flow-Tek water tank" setup as illustrated in Figure 3.3. A steep pit of 2-meter depth is dug fitting the size of a 2500 litre water tank (Flow-Tek Botswana), and a 15 cm layer of concrete is poured at the bottom of the pit. The tank, from which the middle part of the bottom is cut out, is placed in the pit and extra concrete is poured through the main hole at the top. Thus, a bottom thickness of 25 cm of solid concrete, or more, is obtained which provides a good coupling with the earth and a stable basis for the seismometer. To avoid signals due to thermal expansion, no iron is used. The concrete sticks to the

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underground and by covering the tank bottom slab, and all sides till the top of the tank, a watertight construction is created. On top and bottom, the tank tube holes are used for air ventilation and covered with gauze against animals (Figure 3.3).

Figure 3.3 Schematic cartoon showing the configuration of the station setup.

The sensor is placed in a bottomless bucket, directly on the concrete floor. After adjusting the sensor (Figure 3.4), the bucket is filled with sandbags for extra temperature isolation, and they can easily be removed in case of a sensor problem (Figure 3.3 and 3.4). The tank inside working space is 140 cm in diameter and more than 150 cm in height. This space is enough for all instruments and batteries, and still comfortable to work in (Figure 3.5).

To avoid people climbing on the roof and to reduce temperature effects, a cover is placed on the exposed top of the tank, taking away all direct sunlight and rain on the seismometer housing. The tank main hole lid has screw-thread with a ring and is therefore easy to lock with a padlock (Figure 3.6).

In those cases where the ground is too hard to get in, a small shelter

Figure 3.4 Illustration of the sensor setup in to the left and the bottom-less bucket that contains the sensor on the right.

Figure 3.5 Illustration of the data logger preparation to the left. On the right, an overview inside the tank shows the sand bags cover on the sensor and the enough space for the instruments.

Figure 3.6 Illustration of the station roof and ventilation pipes to the left. In the middle, illustration of the lid with ring and padlock. On the right illustration of the enough space for a person to come in and out from the tank to setup and maintain the instrument.

Figure 3.7 Illustration of the shelter setup to the left and the sensor setup inside the shelter to the right.

from stones is built. Hollow stones and a double roof are used to have at least some temperature isolation (Figure 3.7). The base is a concrete slab that is constructed on top of the exposed bedrock. To ensure proper contact the top part of the weathered rock and thin soil layer (if present) are taken away before construction of the concrete slab. The rest of the installation is similar to the installation of the subsurface water tank. To reduce temperature influences, the STS-2 sensor (the most sensitive to temperature variations) was only used in the tank setup and not for surface installations in a house.

3.1.3 Positioning

The seismometer should be oriented as close as possible along the geographical NS direction. A magnetic compass measures the direction to the magnetic pole, but has to be corrected for the local declination. At station locations with metal in the construction and/or subsurface, a magnetic compass is impossible to use and therefore we use a Gyro-compass in combination with a magnetic compass. First we measure the magnetic directions at, at least, two different places to be sure that the presence of metal in the ground did not disturb the measurement. Then the gyro is exactly aligned along the compass and adjusted for declination. Now the gyro points exactly to the geographical NS direction (Figure 3.8). The gyro can then be brought down into the tank for proper orientation of the sensor.

3.1.4 Power

Some of the stations are installed in remote locations (natural parks) or far away from an access point to a reliable source of power. It was therefore decided to be independent of the local electricity net for the whole network, and therefore all sites are powered by solar energy. The total power consumption of a NARS station is 15 Watt. Therefore, a panel

Figure 3.8 Illustration of the positioning process at different locations to the left and the gyro compass used for the measurements to the right.

Figure 3.9 The 140 Watt solar panel used for the stations to the left and the intelligent charge controller that regulates the power and takes care for battery overcharging to the right.

of 140 Watt is installed (Figure 3.9-A) in combination with 2 x 105 Amp Hour deep-charge batteries. When fully charged, it will keep the station running for at least five days in case of overcast. An intelligent charge controller regulates the power and prevents overcharging of the batteries (Figure 3.9-B). The data logger is connected directly to the batteries and not at the controller load. A GSM modem is connected to the datalogger for communicating station status to the central processing unit at the Department of Geological Survey of the Republic of Botswana (DGS) headquarters in Lobatse. Depending on the battery charge level, the data logger power monitor will shut down or reboot the data logger system. This will avoid uncontrolled power switch off by the charge controller what could lead to a data logger disk crash. The solar panels are mounted on poles made of old drill pipes provided by DGS. The panel is mounted in a slot frame and closed with a padlock. On top of the frame, the GPS modem is mounted. Cables are routed inside the drill pipe and go below the ground into the Flow-Tek tank with instruments.

3.1.5 The Team

Without the effort of Arie van Wettum, the great support of our colleagues at the DGS, and the logistic support of AfricaArray (AA), it would not have been possible to set up this seismic network.

Figure 3.10 Arie van Wettum in the middle surrounded by Team-A from DGS.

3.2 Data and Methods

This thesis consists of two parts in which several methods were used. In the first part, receiver functions and seismic surface waves were used to explore the crustal and upper mantle structure of Botswana. Receiver function analysis is a well-established method to map the discontinuities underneath a seismic station. On the other hand, seismic surface waves obtained from ambient noise interferometry and the two-station method (for teleseismic earthquakes) are well suited to investigate the 3D crustal and upper mantle seismic velocity structure between stations. In the second part of the thesis, cooperative inversion strategies will be used to integrate seismological information and satellite gravity measurements with sequential and joint inversion approaches. In this section, the datasets and methods will be briefly described.

Figure 3.11 The different seismological networks and their stations that were used in this thesis plotted together with the regional tectonic boundaries based on Leseane et al. (2015).

3.2.1 Data

The work in this thesis depends mainly on two different datasets, seismological measurements from the African networks and gravity measurements from the Gravity Field and Steady-State Ocean Circulation Explorer (GOCE) satellite. In the following subsections, these two datasets will be briefly described.

3.2.1.1 Seismological data

The seismological dataset consists mainly of the data provided by NARS-Botswana but also from four other networks: 1) the AfricaArray (AF) network, 2) the SAFARI network in Botswana, 3) the AfricaArray-Namibia (AF-Nam) network and 4) the Global Seismological Network (GSN) stations in southern Africa. The NARS-Botswana data was used for the receiver function study in Chapter 4 and the ambient noise analysis in Chapter 5. For the construction of the 3D shear wave velocity model of the crust and upper mantle in Botswana, additional data from the previously mentioned networks were also used to increase the spatial coverage and to improve the resolution of the model. Figure 3.11 shows the distribution of the seismic stations used in this thesis.

3.2.1.2 Gravity data

In the second part of the thesis, satellite gravity data from the Gravity field and Ocean Circulation Explorer (GOCE) are used. With the launch of the GOCE (Floborghagen et al., 2011; Drinkwater et al., 2003) in 2009 the Earth science related efforts in gravity got another boost (van der Meijde et al., 2013). After the time-lapse and long-wavelength studies from the Gravity Recovery and Climate Experiment (GRACE, Tapley et al.

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(2004)) a new sensor was available for determination of the Earth's gravity field and geoid with high accuracy and spatial resolution. Equipped with a three-axes gradiometer and flying at an altitude of 260 km and less GOCE provided (the GOCE mission ended on 13.11.2013) the most detailed measurements of Earth's gravity from space ever by acquiring gravity gradients, i.e., the three-dimensional second derivatives of the gravitational potential. For solid Earth sciences GOCE data have made a unique contribution, providing insights that would not have been possible otherwise. Research has largely focussed on those areas where GOCE really made a difference because of the lack of terrestrial and airborne gravimetry data, like in Africa and South America. The primary application of GOCE data in those, and other, areas is crustal studies, such as recovery of Moho or intra-crustal discontinuities (van der Meijde et al., 2013).

Figure 3.12 The Bouguer anomaly map of Botswana based on the GOCO5c model derived from GOCE data.

Gravity models are based either on satellite measurements alone or on a combination of satellite, airborne, and ground-based gravity meas-

urements, often combined with altimetry data for oceanic regions. Up to now, four releases (R1 to R4) of GOCE gravity field models have been computed within the framework of the ESA project "GOCE High-Level Processing Facility" (HPF). In the frame of GOCE HPF, three different methods and processing philosophies are applied for gravity field modeling: the direct approach (DIR; Bruinsma, 2010), the time-wise approach (TIM; Pail et al., 2010), and the space-wise approach (SPW; Migliaccio et al., 2010). While the DIR models are combination models including CHAMP (Challenging Minisatellite Payload; Reigber et al., 2002), GRACE and satellite laser ranging (SLR) data (R3 and R4), all four releases of the TIM models are based solely on GOCE data and are independent of any other gravity field prior information. After R2, the SPW approach has been redefined to provide gravity gradient grids mainly for geophysical users. A detailed overview of the three approaches is provided in Pail et al. (2011). Details on standards and background models can also be found in the GOCE Standards document (EGG-C, 2010). Bouguer anomaly map of Botswana extracted from GOCO5c model that is purely based on GOCE observation is presented in Figure 3.12.

The accuracy of the models has been thoroughly validated by comparison with external data sources and reference gravity field models which do not contain GOCE information. The comparison of the R1 models with the combined model EGM2008 (Pavlis et al., 2012), which contains also terrestrial data, already revealed the high quality of GOCE data but also a sub-optimal data combination procedure exploited in the production of EGM2008 (Hashemi Farahani et al., 2013). It adds high-resolution gravity field information in regions, where no or only low-accuracy terrestrial gravity field information exists such as South America, central Africa, the Himalaya region, or Antarctica (van der Meijde et al., 2013, Figure 3.13).

This shows that for regions with sparse data coverage the GOCE data significantly increases opportunities to obtain improved or new knowledge on these regions. But the question is how to tackle the inherent non-uniqueness. Parametrization of an inversion is often a knowledge driven process, but for data sparse areas also a biased and subjective process. Creating models based on an interpolation between point locations of known crustal thickness, or extrapolation of geological information from other (similar) geological provinces, is logical from a knowledge driven concept but not necessarily correct. Because it is knowledge driven we tend to believe these models, they are close to what we believe is realistic, but there is no way of validating the quality of these models. For data sparse regions it is therefore close to impossible to decide what the added value is of a certain model. Differences in modelling techniques and data used, seismological or gravity or a mix, can both lead to differences in the final models. Assessing the impact of a specific data source or specific modelling technique is mathematically possible but giving it a physical quality value is not. Part of this thesis will address

this uncertainty and non-uniqueness.

Figure 3.13 Gravity anomaly differences (mGal) to EGM2008 in North America (top row) and South America (bottom row), for TIM_R1 (left column) and TIM_R4 (right column). From: van der Meijde et al. (2013).

3.2.2 Seismological methods

3.2.2.1 Receiver functions

Receiver function analysis is a well-established method to retrieve the subsurface discontinuities underneath a seismic station. Receiver functions (RFs) are time series that contain P-to-S conversions from discontinuities under a seismic station. They are often used to estimate the crustal thickness which is an important parameter of the geodynamic setting.

The analysis starts with isolating the 3-component earthquake records, i.e. the P wave signal with its coda for each earthquake, from the continuous data. Then, the north-south and east-west component signals are rotated to radial and transverse orientations. After that, RFs are calculated by deconvolving the vertical component from the radial and transverse components. The deconvolution process effectively removes both the instrument response and the source time function, keeping only

the P-to-S conversions from the discontinuous near-station structure (Ammon et al., 1990; van der Meijde et al., 2003). Figure 3.14 illustrates the different steps of the RFs analysis.

More specifically, in this thesis the receiver functions were calculated using earthquakes at epicentral distances of 30-105 degrees and magnitudes ≥ 5.5 . The earthquake signals were windowed by a 150 s time window around the first P-wave arrival time determined using the IASP91 model (Kennett and Engdahl, 1991). Then, the data was band pass filtered and rotated to radial and transverse components. Finally, the rotated components were deconvolved from the vertical one using time-domain iterative deconvolution (Ligorria and Ammon, 1999).

The resulting RFs can be modelled with different approaches, e.g., H-K analysis (Zhu and Kanamori, 2000), linear inversion (Ammon et al., 1990), and stochastic inversion methods (Sambridge, 1999a,b). I used the well-established H-K stacking technique that finds Moho depths. Further details about the H-K technique will be provided in Chapter 4.

3.2.2.2 Ambient noise

Ambient noise analysis is a technique that recently has become fundamental in passive seismic imaging. It overcomes the major limitations related to the dependency on wavefield generation by high-cost explosions or randomly distributed earthquakes.

The ambient noise technique emerged from pioneering applications in acoustics (e.g., Weaver and Lobkis, 2001; Lobkis and Weaver, 2001) and seismics (e.g., Campillo and Paul, 2003). At the start of this millennium, several studies showed that the elastic response of the Earth could be retrieved by extracting the Green's function from a random isotropic or diffuse wavefield. A fully diffuse field is composed of waves with random phases and amplitudes traveling in all directions. It can be used to obtain the Green's function between any two receivers by cross-correlation, as if one of them is a virtual source and the other one the receiver (Wapenaar, 2004; Shapiro and Campillo, 2004). In practice, for regional surface wave studies, the diffuse wavefield is usually generated by an inhomogeneous noise source distribution, i.e. violating the assumption of an isotropic wavefield. However, Yang and Ritzwoller (2008) carried out several synthetic tests that showed that an empirical Green's function emerged from long-term ambient noise time series.

In this thesis, ambient noise tomography is used to image the crustal structure of Botswana from fundamental mode Rayleigh wave group and phase velocities. Ambient noise tomography can be described as a two-step process that will be described briefly in the following paragraphs.

Figure 3.14 Schematic cartoon explaining the different steps of RF analysis starting with the P wave ray path (A) and a recorded 3-component earthquake signal (B). (C) shows a synthetic (noise free) radial RF as theoretically might be obtained by time domain iterative deconvolution of earthquake data, and (D) the corresponding subsurface model. The solid lines of the ray paths for (C) represent P wave segments, and the dashed lines S wave segments.

The first step of ambient noise tomography is to obtain (vertical component) empirical Green's functions from (vertical component) noise recordings and then to estimate the inter-station Rayleigh wave group and phase velocities from these Green's functions. We adopted the methodology of Poli et al. (2013) to extract the empirical Green's functions for all available inter-station paths. The data were corrected for instrument response, bandpass filtered, decimated, and subjected to spectral whitening. For each station pair, the cross-correlation function was calculated for 4-hour segments, excluding segments with amplitudes larger than ten times the standard deviation of the windowed data, and stacked over periods of three months. Figure 3.15 shows the high-quality cross-correlations estimated from all station pairs of the NARS-Botswana network. The estimated empirical Green's functions were then subjected to Automatic Frequency-Time Analysis (AFTAN) to estimate the inter-station

Figure 3.15 Vertical component noise cross-correlations for all station pairs of the NARS-Botswana network. The clear Rayleigh waves illustrate the high quality of the data.

Figure 3.16 Example of Automatic Time-Frequency Analysis for the cross-correlation of NARS-Botswana stations NE201 and NE218 stations (blue).

fundamental mode Rayleigh wave group and phase velocity dispersion curves (Bensen et al., 2007). The phase and group velocity dispersion curves were determined for periods between 3 and 35 seconds for each of the three-monthly stacks. Then, for each station, the three-monthly dispersion curves were averaged to obtain the final curve together with its standard deviation.

The second step is the inversion of the dispersion curves into group and phase velocity maps at various periods. I inverted the inter-station dispersion curves into group and phase velocity maps on a 1° by 1° grid for periods of 3-35 seconds using the method of Barmin et al., (2001).

The method depends on constructing the tomographic maps by minimising an objective function that considers three terms: 1) data misfit, 2) smoothness of the maps, and 3) the norm of the perturbation from the reference model. The contribution of each term to the overall objective function is controlled by a weighting factor. We varied the weights of these terms systematically until we obtained acceptable results considering the resulting data misfit, the final resolution, and the perturbation norm. For further details about the method, the reader is referred to Barmin et al. (2001) and Levshin et al. (2005).

3.2.2.3 Earthquake Helmholtz Tomography

Earthquake surface waves carry sufficient energy to allow direct estimation of group and phase velocities. In this thesis, I used the Automated Surface Wave Phase Velocity Measuring System (ASWMS) developed by Jin and Gaherty (2015) to determine phase velocity maps for periods between 30 and 100 seconds.

ASWMS measures the frequency dependent phase delays between nearby station pairs by multi-channel cross-correlation and determines phase velocities and propagation directions via the Helmholtz equation. Here ASWMS is applied to the vertical component earthquake records to directly derive Rayleigh wave phase velocity maps. The method is described in detail in Jin and Gaherty (2015), here I only give the most important steps of the procedure. The cross-correlation of windowed Rayleigh wave signals of two nearby stations quantifies their phase and amplitude difference. The cross-correlation is narrow-band filtered around the central periods under investigation to estimate the phase delays and corresponding phase delay times. Initial (apparent) phase velocity maps are constructed using the Eikonal equation, i.e. solely based the gradients of the inter-station phase delay times. The amplitudes of the windowed and narrow-band filtered Rayleigh wave auto-correlations at each station provide an approximation of the amplitude power at each central period. These amplitudes are therefore used to correct the initial (apparent) phase velocity maps for the effects of non-plane wave propagation (such as multipathing) using the Helmholtz equation. For more details see Jin and Gaherty (2015).

In this thesis, I used earthquake records between June 2016 to May 2017, with magnitudes $M_w > 6$, and at a distances of 30 degrees or more from the centre of the study area to conform to the plane wave assumptions. I used all earthquakes recorded by more than ten stations that resulted in 469 earthquakes to finally perform the earthquake Helmholtz tomography. The group and phase velocity maps from ambient seismic noise and earthquake Helmholtz tomography were used to derive the 3D shear wave velocity structure beneath Botswana.

Figure 3.17 The distribution of the earthquakes that were used for the earthquake Helmholtz tomography.

3.2.2.4 1D Inversion

The group and phase velocity maps provide the local variations of group and phase velocity at each grid point. Group and phase velocity curves at each grid point were jointly inverted to produce the local 1-D shear wave velocity structure for that location. The inversion of all grid points then results in a 3-D model of the crust and upper mantle. I used an earlier version of the Mineos software for the forward calculation of the fundamental mode Rayleigh wave group and phase velocities (Woodhouse and Doornbos, 1988). The inversion was carried out using iterative damped least squares optimisation based on the Levenberg-Marquardt algorithm similar to Greve et al. (2014). The 1D inversion aimed to minimise an objective function f (depending on measured data $\mathbf{d}$ and 1-D subsurface model $\mathbf{m}$) that consists of three misfit terms: the data misfit, the vertical smoothness, and the norm of the perturbation from the reference model following the equation:

$$f(\mathbf{d}, \mathbf{m}) = \| (G(\mathbf{m}) - \mathbf{d}) \|_2^2 + \alpha \| \text{Model smoothness} \|_2^2 + \beta \| \text{Model perturbation} \|_2^2 \quad (3.1)$$

where the first term represents data misfit. $G(\mathbf{m})$ is the calculated dispersion data for model $\mathbf{m}$, and $\mathbf{d}$ are the measured dispersion curves. The model smoothness term is calculated using the first difference of the 1-D model following Julià et al. (2000) and α is a parameter controlling the contribution of the model smoothness term to the objective function. The model perturbation term is calculated using the difference between the 1-D reference model and the estimated model, and β is a parameter controlling the contribution of the norm of the perturbation from the reference model to the objective function.

In this study, I used an interpolated version of AK135 as a starting model as used by Kgaswane et al. (2009) in South Africa. I carried out a test to estimate the influence of the starting model on the inversion

3. NARS-Botswana, Data, and Methods

Figure 3.18 This figure shows a test that was carried out to investigate the influence of the starting model on the inversion results. (A) shows the different starting models used in the test. (B) shows the inversion results using the different starting model in (A) where we can see that the differences between the 1-D retrieved models are minor.

results using four different starting models (Figure 3.18 A-B). The test used the average group and phase velocity curves between 3-35 s across Botswana derived from seismic ambient noise analysis to estimate the average 1-D crustal structure. The inversion results show that the 1-D models are close to each other (Figure 3.18-B). Therefore, I concluded that the influence of starting model on the inversion results is minor.

For this thesis, I inverted the group (3-30 s) and phase (3-100 s) velocity dispersion curves that are sensitive to the crust and upper mantle. An example of a 1-D inversion for a single grid point is presented in Figure 3.19. Further details about the 1-D inversion will be provided in Chapter 6.

Figure 3.19 The designed inversion scheme to jointly invert the 1D Rayleigh wave group and phase velocity dispersion curves. (A) shows the reference, measured, and inversion dispersion curves for one of the locations in Botswana. (B) shows the Rayleigh wave group and phase initial data misfit and the final inversion data misfit. (C) shows the starting reference model and the final inversion model.

3.2.3 Inversion methods

The cooperative inversion concept was introduced by Lines et al. (1988) and was defined as the approach in which we can estimate subsurface models that can fit different geophysical measurements. They divided the concept into two main schemes: sequential and joint inversion. The sequential inversion scheme depends on inverting multiple geophysical measurements separately. The inversion result from one method is used as an input or initial model for the next inversion step(s) of the other geophysical measurements. On the other hand, the joint inversion scheme inverts the multiple geophysical datasets simultaneously based on a coupling relation that relates the different measured physical properties empirically or structurally with each other. This coupling relation conceptually assumes that the rock properties and/or the boundaries between the subsurface bodies have a similar or a common response in these datasets. In other words, they have a coupling response which can be addressed empirically or physically. However, both sequential and joint schemes have advantages and disadvantages. In general, the main drawback of the joint inversion scheme is that the coupling assumption between the different geophysical methods is not valid for some subsurface environments, while the sequential scheme gives the freedom for

the inversion results to vary between the different geophysical inversion steps. On the other hand, the joint inversion approach results in a unique solution, while the sequential inversion has an added freedom in the subsequent inversion step(s) that increases the non-uniqueness of the solution (Gallardo and Meju, 2011).

Although the cooperative inversion schemes can lead to a better 3D subsurface models, there are difficult challenges that impede the development in this field. One of the central and crucial problems is quantifying the uncertainty in the produced models to evaluate them. The uncertainty in the cooperative inversion techniques can be caused by (Gallardo and Meju, 2011):

- (1) the inaccuracy in the data measurements and the limited band widths,
- (2) the incomplete knowledge about the complex physical system under investigation, and
- (3) the error in the theoretical and mathematical formulation of the physical systems to predictive modeling and the discretization of the system.

Different authors have treated these sources of uncertainty (e.g. De-Volder et al. (2002)) in either probabilistic (e.g. Alkhatib and Schuh (2007) and Cardiff and Kitanidis (2009)) or deterministic approaches (e.g. Vasco (2007) and Meju (2009)) to estimate the errors in the produced models. However, it is still an open question if the mentioned approaches can realistically address the uncertainty in the large scale applications of cooperative inversion schemes (Gallardo and Meju, 2011). An additional challenge is a necessity to powerful computational resources (Gallardo and Meju, 2011) which was the reason that some developed joint inversion schemes were not able to fully function due to the limited computational power (e.g. Maceira and Ammon (2009), Khan et al. (2015), and Basuyau and Tiberi (2011)). In this part of the thesis, two strategies will be introduced to integrate seismological and gravity data.

The motivation for the work came from the emerging large amount of high-quality seismological data from different networks and the global coverage gravity and gravity gradients measurements provided by recent GOCE satellite mission. New methodologies are needed to exploit the full capability of both datasets and to enhance the 3D subsurface models.

In this part of the thesis, two strategies will be introduced to integrate seismological and gravity, including the gradients, data. The two approaches can function under different conditions. The first strategy uses a sequential inversion approach that works when a 3D shear wave velocity model is available. The second strategy is based on a joint inversion approach given that dispersion measurements are available. In this case, both surface wave dispersion maps and gravity measurements can jointly be used to illuminate the 3D subsurface structure. In the follow-

ing subsections, a brief description of the two strategies is introduced. This part of the thesis is mainly methodological, therefore, to prevent repetition, full descriptions of the two methodologies are provided in chapter 6, 7, and 8.

3.2.3.1 Sequential inversion

The sequential inversion strategy was developed based on object-based philosophy. The approach uses a 3D shear wave velocity model obtained by surface wave inversion (Adams et al., 2012), that was used as the input for an object-based gravity inversion. The method can be divided into two main steps. The first step is the classification of the subsurface structures into objects using reproducible and subjective criteria based on object-based classification approach that will be introduced in chapter 6. The second step is the inversion of the gravity total field data to retrieve the density structure of the extracted objects using least-squares object-based inversion method (Schmidt et al., 2011). A detailed explanation of the methodology will be given in Chapter 7.

3.2.3.2 Joint inversion

The joint inversion strategy is developed to simultaneously invert surface wave dispersion measurements with gravity and gravity gradients. The developed method constitutes of different components:

- 3D gravity forward calculator using Tesseroids (Uieda et al., 2016).
- 3D surface wave forward calculator based on 1D solver that is earlier version of Mineos code (Woodhouse and Doornbos, 1988).
- Damped least squares inversion scheme.

The joint inversion method was implemented in a Python framework. A synthetic test was designed to investigate the capability of the developed methodology to retrieve the crustal structure using Rayleigh waves group and phase velocity jointly with satellite gravity total field and gradients measurements at the satellite altitude of 220 km similar to GOCE gravity measurements. The developed methodology and the synthetic test are fully explained in Chapter 9.

Crustal structure and dynamics of Botswana

4

Abstract

We present the first nationwide crustal thickness and Vp/Vs maps for Botswana based on the analysis of new receiver functions of NARS-Botswana network integrated with previous receiver function results in Botswana. Using H-K analysis, we found crustal thickness values ranging from 34 km for the Okavango Rift Zone to 50 km at the border between the Magondi Belt and the Zimbabwe Craton. For stations with significant sediment cover, a sediment correction was applied based on sequential H-K stacking. We observed distinct differences for the Kaapvaal craton. The northwestern part has a high Vp/Vs ratio typical of a predominantly mafic composition, indicating the extension of the Bushveld mafic complex. On the other hand, the western part with a Vp/Vs of 1.67 is felsic, probably as a result of delamination caused by rifting processes. Further to the west, we found a crustal thickness of 42 km and a Vp/Vs ratio of 1.76 for the Nosop Basin. These values are similar to other cratonic regions in Botswana, suggesting the presence of a buried craton as proposed by previous studies. We confirm a relatively thin crust (~ 34-39 km) and high Vp/Vs ratio (~1.84) underneath the Okavango Rift Zone (ORZ) found by previous receiver function studies. Furthermore, we found a thin crust (37 km) and high Vp/Vs ratio (1.84) in central Botswana underneath the Passarge Basin. The location is close to the location of the 2017 large magnitude extensional Botswana earthquake, suggesting that the rift initiation process extends further south than commonly assumed.

4.1 Introduction

The crustal structure in southern Africa mainly consists of cratons with mobile belts in between. The detailed structure of crust and upper

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mantle is still unknown due to limited seismic station coverage in large parts of southern Africa, in particular in Botswana. Botswana is located in between three main cratonic provinces: the Kaapvaal Craton in the south and southeast, the Zimbabwe Craton in the east and northeast, and the Congo Craton in the northwest. Little is known about the extent of the three cratons and the structure and geodynamical role of the mobile belts in between.

To date, only a few studies (e.g., Nguuri et al. (2001); Yang and Ritzwoller (2008); Kgaswane et al. (2009); Li (2011); Adams and Nyblade (2011); Youssof et al. (2015)) investigated the crust in the southeastern part of Botswana based on data from the SASE temporary seismological network (Carlson et al., 1996). Recently, a new temporary seismic line was deployed across the Okavango Rift Zone (ORZ) as part of the Seismic Arrays for African Rift Initiation (SAFARI) (Gao et al., 2013a). Partly overlapping with these studies are just a few studies using other geophysical techniques: deep seismic profiling for the Nosop Basin (Wright and Hall, 1990), potential field data analysis (Hutchins and Reeves, 1980; Leseane et al., 2015), magnetotelluric (MT) from the recent Southern African Magnetotelluric Experiment (SAMTEX) in the northern (Miensopust et al., 2011a; Khoza et al., 2013a,b), and southwestern (Muller et al., 2009) part of Botswana, as well as combined with regional seismic tomography for Southern Africa by Jones et al. (2013).

Since November 2013, a temporary array of seismometers is deployed in Botswana. This NARS-Botswana network (NARS-Website, 2013) consists of 21 broadband seismic stations distributed over the whole country. It strategically covers the different geological and tectonic units and complements the existing stations MAUN from AfricaArray (AA), and LBTB from the Global Seismic Network (GSN) (Figure 4.1). It fills a gap in station coverage for southern Africa, extending both west- and southwards from previous temporary seismic deployments in Zambia, Zimbabwe and South Africa. In this study, we will present crustal thicknesses and V_p/V_s ratios for Botswana using receiver functions analysis from the new data delivered by the NARS-Botswana network and link the results to the geodynamics of the region, with particular focus on the relation between cratons and mobile belts, and the possible existence of rifting in the central part of Botswana.

4.2 Botswana geological and tectonic history

The crust of Botswana encompasses diverse tectonic features (Figure 4.1) with a mix of cratons and mobile belts. The cratonic areas cover the Archean Kalahari Craton with the Kaapvaal and Zimbabwe cratonic blocks in the south and east of the country, and the Congo Craton in the northwest. In between, a series of mobile belts is found: the Limpopo, Kheis-Okwa-Magondi, and Damara belts (Key and Ayres, 2000). The late Archean Limpopo Belt, reworked during the Proterozoic, is located in the east of Botswana and sutures the Kaapvaal and Zimbabwe cratons (Barton

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Figure 4.1 Overview of the NARS-Botswana network. The green lines indicate the tectonic boundaries after Singletary et al. (2003). The yellow triangles are NARS stations used in this research, blue is GSN station LBTB, and green is AA station MAUN. The grey transparent contour outlines the Okavango Rift Zone after Yu et al. (2015c). The two light blue dashed lines comprise the Kalahari Suture Zone (KSZ) after Haddon (2005).

et al., 2006; Begg et al., 2009). The largest belt is the composite Kheis-Okwa-Magondi Belt that is bounded by the Kalahari Suture Zone (KSZ). It runs from southwest to northeast Botswana. The KSZ is a Paleoproterozoic major thrust zone that separates the Archean blocks to its southeast (Kalahari Craton, Zimbabwe Craton, Limpopo and Kheis-Okwa-Magondi belts) from the Proterozoic Damara, Rehoboth and Ghanzi-Chobe belts towards the north and west (Reeves, 1978). The Damara Belt is a complex Neoproterozoic belt that bounds the southeastern margin of the Congo Craton. It consists of a succession of sediments that are highly deformed and metamorphosed during the Pan-African Orogeny (Wright and Hall, 1990; Key and Ayres, 2000; Haddon, 2005). Bounding the Damara Belt towards the southeast is the Ghanzi-Chobe Belt that consists of tightly folded meta-sediment sequences (Wright and Hall, 1990). At the northern tip, cutting across the Damara and Ghanzi-Chobe Zone, is the incipient Okavango Rift, often considered to be the southwestern terminus of the East African Rift System (EARS). Within the belt, the Passarge basin has up to 15 km thick Ghanzi-Group sediments, which were deposited

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during Neoproterozoic and early Paleozoic times (Pretorius, 1984). One of the poorer defined regions is the Rehoboth Province which extends from southwestern Botswana to western Namibia (Begg et al., 2009; van Schijndel et al., 2014). A major part of the province aggregated around an Archean nuclei during the Paleoproterozoic (Van Schijndel et al., 2011). In the southwest, the upper part of the crust is formed by the Nosop basin. It is filled with thick Nama Group and Ghanzi Group sediments that reach up to 15 km depth (Pretorius, 1984; Wright and Hall, 1990). Underneath it, an ancient micro-craton (the Maltahohe craton) may exist (Begg et al., 2009), but it has also been interpreted as an extension of the Kaapvaal Craton (Wright and Hall, 1990). Further details on the tectonic evolution and composition of the crust for each of the provinces are provided with the findings of crustal thickness and V_p/V_s ratio in the Results and Discussion section.

4.3 Data and Methods

4.3.1 Data

NARS-Botswana is a temporary seismological network of consisting 21 broadband stations deployed across Botswana (Figure 4.1). The main aim of the project is to image the crustal and upper mantle structure (for more information about the NARS seismological network the reader is referred to the NARS-Website (2013)). In this study, we used data collected from these stations, between November 2013 and August 2016, as well as from GSN station LBTB and AA station MAUN. The reader is referred to Figure S1 in the supplementary material for an overview of the data used in this study.

4.3.2 Receiver Functions

P-wave receiver functions (RFs) are time series that contain P-to-S conversions from discontinuities under a seismic station. They are mainly used for estimating the crustal thickness and velocity relations (V_p/V_s ratios). Crustal thickness and V_p/V_s ratio are crucial parameters for understanding the tectonic evolution, the geodynamical processes and the composition of the crust. RFs are derived from earthquake records at a seismic station by deconvolving the vertical component from its radial and transverse components. This, theoretically, removes both the instrument response and source imprint from the recorded signal. The remaining signal contains only direct P-to-S conversions at discontinuous structure underneath the station, and the corresponding reverberations (Ammon et al., 1990; van der Meijde et al., 2003).

In this study, P-wave receiver functions were calculated for each station. Earthquakes with epicentral distances of $30^\circ - 105^\circ$ and magnitudes ≥ 5.5 were used. The earthquake signals were windowed between -30 to +120 seconds relative to the P-wave arrival time. The theoretical arrival

time was determined using the IASP91 reference model (Kennett and Engdahl, 1991). Then, the data was filtered using a Butterworth band pass filter with corner frequencies of 0.05 and 1.25 Hz. For high frequency RFs, used for sedimentary thickness estimation as described later, the corner frequency limits were set 0.05 and 2.5 Hz. After that, the data was rotated to radial and transverse components. Finally, the rotated components were deconvolved with the vertical one using a time-domain iterative deconvolution (Ligorria and Ammon, 1999). In the deconvolution process a further Gaussian filter is applied, which acts as a low-pass filter (Ligorria and Ammon, 1999). We have mostly applied Gaussian widths of 2.5 and 5.0, corresponding to the previously used low-pass cut-offs of roughly 1.25 and 2.5 Hz, respectively.

The receiver functions are then quality controlled, passing through three tests. Firstly, a least square misfit criterion was used to judge the quality of the RFs, where for each event the radial RF was convolved with the vertical component to retrieve the original radial one. If the retrieved signal has at least an 85% fit with the original, the RF was considered for further processing. Secondly, RFs with signal-to-noise ratio (S/N) larger than 3.0 were selected, where the S/N ratio was calculated using:

$$S = |A_r| / |A_t| \quad (4.1)$$

where A_r is the maximum amplitude on the radial RF and A_t is the maximum amplitude on the transverse RF. Finally, a visual check was done to ensure the quality of the RFs, and to exclude the ones with weak or anomalous first P-wave arrivals.

4.3.3 H-K analysis

H-K stacking is a well-established technique to obtain estimates of the Moho depth and Vp/Vs ratio of the crust using RFs (Zhu and Kanamori, 2000). The technique is based on searching the parameter space of all possible combinations of crustal thickness (H) and Vp/Vs ratio (K) within predefined ranges to find the maximum value of the following objective function:

$$S(H, K) = \sum_{j=1}^N w_1 r_j(t_1) + w_2 r_j(t_2) - w_3 r_j(t_3) \quad (4.2)$$

where N is the number of RFs at each station, $r_j(t)$ is the amplitude function at time t of the j^{th} RF, t_1 , t_2 , and t_3 are the travel times of the Moho converted phases Ps , $PpPs$, and $PsPs + PpSs$ calculated for the given values of H and K, and w_j are the weights of the three phases where $w_1 + w_2 + w_3 = 1$. Full details of this method can be found in (Zhu and Kanamori, 2000). The travel time calculations of the different converted phases require an average Vp velocity for the crust. In this study, Vp = 6.5 km/s was used based on Durrheim and Green (1992)).

In this research, the H-K method searched depth intervals between 20 to 60 km depth with 0.2 km increments and Vp/Vs ratios between

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Figure 4.2 (A-C) H-K analysis for station NE201, and (D-I) sequential H-K analysis for station NE205. Figure (A) shows the H-K stacking results for station NE201, (B) its RFs ordered by back-azimuth and (C) ordered by epicentral distance. Figures (D-F) present the H-K results for the sedimentary layer for station NE205 using RFs with Gaussian width of 5. Figures (G-I) present the results for the crustal thickness using RFs with a Gaussian width of 2.5. The green lines in the RF figures indicate the times of the converted phases where P_s phase is the first, $PpPs$ second, and $PsPs + PpSs$ third phase to arrive. The green shaded areas highlight the time windows of each phase.

1.65 to 1.95 with 0.0025 increments. The weights used were $w_1 = 0.6$ for Ps , $w_2 = 0.3$ for $PpPs$, and $w_3 = 0.1$ for the $PsPs + PpSs$ phases. However, for some stations, different weights ($w_1 = 0.6$, $w_2 = 0.2$, and $w_3 = 0.2$, or $w_1 = 0.7$, $w_2 = 0.3$, and $w_3 = 0.0$) were used based on the clarity of the Moho reverberations of the signal (for details see Table T1 in the supplementary material).

In a few cases, an apparent delay of the first P-arrival, due to a thick layer of sediments, was noticed. This apparently delayed first P-arrival is a composite peak of the original first P-arrival combined with the direct conversion at the base of the sediment layer. For these stations we applied sequential H-K stacking (Yeck et al., 2013). It is a modified H-K stacking approach that first uses high-frequency RFs (Gaussian width = 5.0) to estimate the thickness and Vp/Vs ratio of the sedimentary layer. We assumed Vp = 3.0 km/s for the sedimentary basin, using a depth range of 0-10 km with a step size of 0.2 km, and a Vp/Vs ratio range of 1.65 to 2.7 with a step size of 0.0025. The estimated depth and Vp/Vs ratio of the sedimentary layer are then used as a correction term for a second H-K search on the RFs with Gaussian width of 2.5 to find the thickness and Vp/Vs ratio of the crust. Further details on the technique can be found in Yeck et al. (2013).

Our uncertainty estimates for crustal thickness and Vp/Vs ratio are based on the approach by Youssof et al. (2013). Uncertainties are obtained by varying the Vp velocity by ± 0.2 km/s from the assumed value average crustal Vp velocity of 6.5 km/s providing upper and lower bounds of our results. It should be noted that this is a formal uncertainty corresponding to a variation in Vp velocity only. Due to complex structure, the epistemic uncertainties can be larger.

4.4 Results and Discussions

All stations in this study were subjected to conventional H-K stacking analysis. However, stations NE202, NE205, NE211, NE212, NE216 and NE220 showed a noticeable delay in the first peak (Figure 4.2D-I, supplementary figures S5, S13, S14, S18.1, and S22) that required sequential H-K analysis. Nevertheless, only for stations NE202, NE205 and NE212, significantly different results were obtained, considering the uncertainty margins, compared to the conventional H-K analysis. Therefore, sequential H-K results are only presented for NE202, NE205 and NE212. Figure 4.2A-C shows a representative example of the receiver function analysis using the standard H-K stacking technique applied to station NE201. The modelled theoretical Ps , $PpPs$, and $PsPs + PpSs$ arrival times, based on the optimum H and K values, show good agreement with the Moho conversion and reverberations on the measured receiver functions. Figure 4.2D-I shows a representative example of sequential H-K analysis, applied to station NE205. The average uncertainties in crustal thickness and Vp/Vs ratio for all stations are calculated to be in the range of ± 1.5

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Table 4.1 The results of the H-K analysis for the different tectonic regions.

Tectonic region	Station	Crustal thickness (km)			Vp/Vs			No. of Rfs
		Lower bound	Solution	Upper bound	Lower bound	Solution	Upper bound	
Western Zimbabwe Craton	NE217	37.6	40	40.6	1.78	1.78	1.81	18
	NE218	47.2	49	50.4	1.69	1.69	1.71	21
	NE219	40.8	42	43.4	1.71	1.72	1.72	20
Northwestern Kaapvaal Craton	LBTB	40.4	42	43.8	1.73	1.75	1.75	156
	NE201	34.8	36	37.8	1.80	1.81	1.82	16
	NE212	37.8	39	40.6	1.74	1.74	1.76	16
	NE220	37.8	39	41.0	1.81	1.82	1.83	12
Western Kaapvaal Craton	NE213	37.4	39	40.4	1.66	1.67	1.67	24
	NE221	38.0	39	40.6	1.66	1.67	1.67	22
Congo Craton	NE204	36.8	38	39.4	1.78	1.79	1.79	27
	NE207	36.8	38	39.4	1.70	1.71	1.71	11
Magondi Belt	NE208	38.8	40	41.8	1.67	1.69	1.69	24
	NE216	36.2	37	38.4	1.75	1.75	1.75	
Kheis Belt	NE202	37.4	39	40.2	1.70	1.71	1.72	20
Okwa Block	NE211	42.4	44	45.0	1.72	1.72	1.72	21
Magondi Belt	NE208	38.8	40	41.8	1.67	1.69	1.69	14
	NE219	40.8	42	43.4	1.71	1.72	1.72	20
Damara Belt	NE214	42.2	44	45.2	1.72	1.73	1.73	13
Ghanzi-Chobe zone	Maun	38.8	40	41.8	1.86	1.87	1.88	18
	NE205	37.0	38	39.6	1.81	1.81	1.81	14
	NE206	36.6	38	39.0	1.69	1.69	1.69	20
	NE210	44.6	46	47.6	1.81	1.81	1.81	18
	NE215	33.2	34	36.0	1.72	1.74	1.74	6
Passarge Basin	NE209	36.0	37	38.2	1.82	1.84	1.84	14
Nosop Basin	NE203	41.0	42	44.2	1.75	1.76	1.76	17

km and ± 0.02 , respectively. The final results for all stations are given in Table 4.1.

In the following, the RF results are grouped by their tectonic provinces moving from the Archean Cratonic blocks to the Archean and Proterozoic mobile belts and ending by the late Proterozoic and Phanerozoic sedimentary basins. This classification agrees with the spatial distribution of the tectonic units where the cratonic blocks in the east and northwest are surrounded by the mobile belts and the sedimentary basins are caught at the center of the mobile belts. We will discuss the results by firstly focussing on our RF results from NARS-Botswana (Table 4.1 and Figure 4.3 A-B) and then compare them with the results from the SAFARI (Yu et al., 2015c) and SASI (Youssof et al., 2013) networks. Figures 4.3-C and 4.3-D show all available RF results.

4.4.1 Cratons

4.4.1.1 Western Zimbabwe Craton

For NARS-stations on the Zimbabwe Craton (NE217, NE218, and NE219) we found crustal thicknesses varying between 40 and 49 km and Vp/Vs ratios between 1.69 and 1.78, with the thickest crust at the northwestern tip where the craton borders the Magondi Belt (Figure 4.3). Previous RF studies have found that the Moho beneath undisturbed Archean cratons in southern Africa is generally sharp (Nguuri et al., 2001; James et al., 2003), with a 35-40 km thick crust that is of felsic (Nair et al., 2006) or

intermediate (Youssof et al., 2013) composition. Here, we interpret the Vp/Vs ratio in terms of composition based on the crustal rock study of Holbrook et al. (1992), where low (< 1.78), intermediate ($1.78 < \text{Vp/Vs} < 1.81$) and high Vp/Vs (> 1.81) are characteristic of felsic, intermediate and mafic crustal rocks, respectively.

For station NE217, we found a crustal thickness of 40 km and Vp/Vs of 1.78, close to the results for the undisturbed central Zimbabwe Craton of Nguuri et al. (2001) ($H=35\text{-}40$ km), Nair et al. (2006) ($H\approx 38$ km, $\text{Vp/Vs}\approx 1.73$) and Youssof et al. (2013) ($H=35\text{-}38$, $\text{Vp/Vs}=1.74\text{-}1.79$). Moving south, towards NE219, and northwest, towards NE218, the crustal thickness increases towards the Limpopo and Magondi belts, respectively. This agrees with previous studies that have shown that post-Archean reworked belts in southern Africa have a thick crust of 45-50 km (Nguuri et al., 2001; James et al., 2003).

Station NE219 is located on the border between the Limpopo Belt and the Zimbabwe Craton. The Limpopo Belt was formed by the collision between the Zimbabwe and the Kaapvaal craton (~ 2.7 Ga) (Begg et al., 2009). For NE219 we found a 42 km thick crust, indicating a gradual increase of crustal thickness from the Zimbabwe Craton towards the Limpopo Belt (Youssof et al., 2013). The Vp/Vs value of 1.72 agrees with the values reported by Youssof et al. (2013) (1.72-1.74) and Nair et al. (2006) ($\text{Vp/Vs}\approx 1.74$). This confirms that the Vp/Vs values of the southern part of the Limpopo Belt are similar to those observed for undisturbed cratonic blocks.

For station NE218, on the border between the Zimbabwe Craton and the Magondi Belt, we found the largest crustal thickness among the NARS-Botswana stations and a Vp/Vs ratio of 1.69. The value of 49 km crustal thickness closely matches the 50.5 km of Youssof et al. (2013) for nearby station SA70 (Figure 4.3 C-D). Nguuri et al. (2001) and James et al. (2003) explained the thick crust as a result of post-Archean reworking of the margin of the Zimbabwe Craton (~ 2 Ga). This reworking has produced a more complex or gradual Moho generating weak converted phases as reported by many receiver function studies (Nguuri et al., 2001; James et al., 2003; Nair et al., 2006; Youssof et al., 2013). The observed thick crust may be further affected by the collisional event between the Zimbabwe and Congo Craton during the Pan-African Damara Orogeny (Gray et al., 2008). In general, our values show a good spatial correlation with the results from the SAFARI and SASE networks (Figure 4.3). However, we found a lower Vp/Vs ratio of 1.69 at NE218 than the value of 1.75 for station SA70 by Youssof et al. (2013), but it agrees with the felsic nature of the reworked belts surrounding the Zimbabwe Craton.

4.4.1.2 Kaapvaal Craton

For the Kaapvaal Craton (stations LBTB, NE201, NE212, NE213, NE220, NE221) we observe variations in crustal thickness between 36 and 42 km. The Vp/Vs ratios fall in two distinct blocks suggesting different

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Figure 4.3 Crustal thickness (km) and Vp/Vs ratio results for all RFs. Results from NARS-Botswana as circles, SASE as triangles (Youssof et al., 2013), and SAFARI as diamonds (Yu et al., 2015b). Multiple observations per location were plotted on top of each other to facilitate comparison. The background colors indicate crustal thickness and Vp/Vs ratio obtained by nearest neighbour interpolation with a search radius of 120 km using Generic Mapping Tools (GMT) (Wessel et al., 2013). The red star indicates the magnitude 6.5 earthquake that occurred on the 3rd of April 2017.

compositions. The northwestern Kaapvaal Craton block has Vp/Vs values ranging between 1.74 and 1.82, and the western Kaapvaal Craton block has Vp/Vs values of 1.67, on average (Table 4.1 and Figure 4.3 A-B).

For the NARS stations in the northwestern part of the Kaapvaal craton (LBTB, NE201, NE212, and NE220), we found crustal thicknesses of 36-42 km and Vp/Vs ratios of 1.74-1.82 (Table 4.1 and Figure 4.3 (A-C)). These crustal thickness values are similar to those of the undisturbed segment of the western Zimbabwe Craton (35-40 km), although station LBTB has a slightly thicker crust of 42 km. The Vp/Vs estimates are higher than those of the undisturbed western Zimbabwe cratonic segment. The higher Vp/Vs ratio may be linked to mafic crustal intrusions as a result of the major intracratonic Bushveld intrusion event at ~ 2.05 Ga. The Bushveld complex is located in north-central Kaapvaal Craton in South Africa with large surface outcrops. It is the largest layered mafic intrusion within the crust across the globe (Von Gruenewaldt et al., 1985). Nair et al. (2006, and references therein) proposed that this mafic intrusion may have further extended several hundred kilometers to the west. This proposed extension coincides with the location of the northwestern part of the Kaapvaal Craton in this study.

RFs of station LBTB were previously processed by Nair et al. (2006) and Youssof et al. (2013). Our results are obtained from 130 high-quality RFs from earthquakes with magnitudes > 6.0 . The RFs were binned based on ray parameter between 0.004 and 0.008 s/km with a step size of 0.0025 s/km to prevent bias of the results towards specific earthquake clusters at specific distances (Figure S4.1 and S4.2 in the supporting information). Our crustal thickness estimate of 42 km agrees with Nair et al. (2006) (41.4 km) and Youssof et al. (2013) (41.5 km). However, our Vp/Vs ratio of 1.75 is significantly different from Nair et al. (2006) (1.81) and Youssof et al. (2013) (1.82). Nair et al. (2006) used only 31 RFs and Youssof et al. (2013) reported large uncertainty bounds (1.76-1.87). Our Vp/Vs result is close to the lower uncertainty bound of Youssof et al. (2013). However, we have a smaller uncertainty in our Vp/Vs estimate (1.73-1.75) due to the larger number of high-quality RFs and the binning scheme used in our study.

The stations on the western part of the Kaapvaal craton (NE213, NE221) both give a crustal thickness of 39 km and a very low Vp/Vs ratio of 1.67 (Table 4.1, and Figure 4.3 A-B). The low Vp/Vs suggests absence of a lower crustal mafic layer due to the delamination of the lower crust and upper mantle caused by rifting during the Proterozoic (Durrheim and Green, 1992; Niu and James, 2002; Nair et al., 2006; Youssof et al., 2013). This is supported by analysis of xenolith samples from the lower crust that show a lack of mafic granulites (Schmitz and Bowring, 2003). Furthermore, Kgaswane et al. (2009) reported, based on a joint inversion of RFs and Rayleigh waves, relatively low velocities in this region, particularly in the lower part of the crust underneath station SA37 of SASE network.

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4.4.1.3 Congo Craton and transition to Damara Belt

In the northwestern part of Botswana two NARS stations (NE204 and NE207) are located at, or close to, the border of the Congo Craton according to the tectonic interpretation of Singletary et al. (2003) (Figure 4.1). These stations both show a crustal thickness of 38 km, whereas the V_p/V_s ratio varies between 1.71 and 1.79. The three northernmost stations of Yu et al. (2015c) (B12-B14) show a significantly thicker crust of ~ 46 km for the eastern edge of the Congo Craton (Figure 4.3 C-D). The most likely interpretation is that these stations are probably not on the Congo Craton but fall within the reworked Damara Belt, and the thick crust can be related to the collisional event between the Congo and Kaapvaal craton during the Pan-African Damara Orogeny (Gray et al., 2008). For station NE214, in the Damara Ghanzi-Chobe Belt at the border with the Congo Craton and the Okavango Rift region, we found a crustal thickness of 44 km and a V_p/V_s ratio of 1.73. The estimated values are comparable to the values of Yu et al. (2015c) and are indicative of a reworked belt of felsic composition (similar to that of the craton) (Figure 4.3 C-D). Our results show good agreement with those from Yu et al. (2015c) for station B11 (44.3 km and 1.72 for backazimuths 0-45 and 225-360 degrees) and B09 (41.8 km and 1.71).

4.4.2 Mobile Belts

4.4.2.1 Kheis-Okwa-Magondi Belt

The Kheis-Okwa-Magondi Belt is assumed to be a Paleoproterozoic belt that was accreted on the northwestern border of the Kaapvaal Craton due to the Eburnean Orogeny (~ 2.0 -1.8 Ga) (Thomas et al., 1993; Begg et al., 2009). The Magondi basin was formed on the western border of the Zimbabwe Craton (~ 2.16 -2.0 Ga) during an intracratonic extension that occurred during the early Proterozoic. Then, it was amalgamated during the Eburnean orogeny with the Okwa-Kheis belt to the west of the Archean blocks. After that, the margin between the Kheis province and the Kaapvaal Craton was deformed during the Kheis Orogeny (~ 1.75 Ga). Finally, the Kheis province was reactivated and amalgamated with the Rehoboth belt during the Kibaran orogeny ~ 1.2 Ga ago (Thomas et al., 1993).

This complex history is reflected in the results for the stations in this belt. The NARS stations NE208 and NE216 are located at the borders of the Magondi belt and give crustal thicknesses of 37-40 km and V_p/V_s ratios of 1.69-1.75 (Table 4.1, and Figure 4.3 A-B). The V_p/V_s values are similar to the results for stations B02-B04 from Yu et al. (2015c) and are possibly linked to a paleocollisional zone of felsic composition (Figure 4.3 C-D). However, the crustal thickness values are different, especially for station NE216 and B04. The results from NE216 will be further discussed in the sedimentary basin section together with station NE209 in the Passarge basin.

The RFs of the station located in the Kheis belt (NE202) required sequential H-K analysis. They showed a strong sedimentary effect (as identified by the large apparent shift of the first peak and the strong reverberation at 3 s in Figure S5 in the supplementary material). This effect is particularly visible for the backazimuth range between 180-360 degrees (Figure S5 A-F in the supplementary material). RF analysis for this backazimuth range did not produce coherent results and has been left out of the analysis. All results and interpretation are therefore based on RFs with 0-180 backazimuth (Figure S5 G-L in the supplementary material). Sequential H-K was applied and resulted in a sediment thickness of 0.8 km and a Vp/Vs ratio of 2.21. Finally, a crustal thickness of 39 km and Vp/Vs ratio of 1.71 were estimated using the sediment properties as time adjustment. The depth estimate did not significantly vary when we included other backazimuths. However, the Vp/Vs ratio varied significantly. This is most likely related to spatially varying sediment thickness (the station is at the border of the thick sedimentary Nosop basin which is sampled by RFs for events with western backazimuths). The strong backazimuthal effects could not be successfully modelled, remaining for a future study when more events from various backazimuths will be available.

The Okwa Block is an isolated inlier of the Precambrian basement in the Kalahari Desert that was mapped using gravity and magnetic data of Botswana (Hutchins and Reeves, 1980). It contains the Okwa Basement Complex that crops out at the northwestern margin of the Kaapvaal Craton. Using geochronological data, Mapeo et al. (2006) showed that the Okwa Complex was emplaced rapidly at 2.056 ± 3 Ga, simultaneously with the Bushveld Complex. However, the deep extension of the block was not defined due to the lack of seismological observations in this area. Station NE211 in the Okwa Block shows a relatively thick crust of 44 km, and a Vp/Vs ratio of 1.72 comparable to the Kheis Belt (Table 4.1, and Figure 4.3 A-B). The slightly thicker crust can be linked to the post-Archean reworking on the boundary of the cratonic regions during the Eburian Orogeny (~ 2 Ga) (James et al., 2003; Youssef et al., 2013).

4.4.2.2 Damara-Ghanzi-Chobe Belt

The Damara-Ghanzi-Chobe Belt formed during the Damara Orogeny which was part of the Neoproterozoic Pan-African tectono-thermal event between ~ 850 -550 Ma, due to the collision between the Kaapvaal and Congo cratons. The Okavango Rift Zone, which is considered to be an incipient rift system, is located within the Damara-Ghanzi-Chobe Belt.

The Ghanzi-Chobe Belt is a part of this composite belt. It was formed initially as a rift basin that was filled with volcano-sediments that were later deformed during the Pan-African Damara Orogeny forming a fold and thrust belt (Modie, 2000). The Okavango Rift Zone, which is considered to be an incipient rift system, is located within the Damara-Ghanzi-Chobe belt. It has spatially varying crustal thickness and Vp/Vs values (Table 4.1, and Figure 4.3 A-B). In the southwestern part, for sta-

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tion NE210, we find a thick crust of 46 km and a relatively high Vp/Vs ratio of 1.81. Intermediate crustal thicknesses of 38-40 km are found in the central part, for stations MAUN and NE205, with relatively high Vp/Vs ratios of 1.81-1.87 on the border of the Okavango Rift. The north-eastern part of the belt is characterized by a thin crust of 34-38 km (NE206 and NE215) and the lowest Vp/Vs ratios for this belt (1.74-1.68). Together with the results from Yu et al. (2015c), we observe high a Vp/Vs ratio at the central part of the belt, even extending to the central part of the Passarge Basin (Figure 4.3 C-D). This may be an indication of partial melting in the lower crust, facilitating the rifting process and causing the noticeable earthquake activity in the Okavango rift zone as mentioned by Yu et al. (2015b). The presence of either hot fluids ascending from the mantle through weak lithospheric zones or melt is further supported by the shallow Curie depth of 8-15 km from 3D inversion of magnetic data (Leseane et al., 2015).

The analysis for station MAUN, at the southern border of the Okavango Rift, deserves special attention due to strong backazimuthal RF variability. To our knowledge, two independent studies have reported crustal thickness values for this station. Using H-K stacking Kachingwe et al. (2015) found a crustal thickness of 44 km from RFs, whereas they found 38 km from a joint inversion of RFs and surface wave dispersion curves. Yu et al. (2015c) processed their RFs grouped in different backazimuths. The $0^\circ - 45^\circ$ together with the $225^\circ - 360^\circ$ backazimuths resulted in a thickness of 37.5 km and a Vp/Vs ratio of 1.88, whereas the $45^\circ - 225^\circ$ backazimuth range resulted in a thickness of 43.6 km with a Vp/Vs ratio of 1.85. Yu et al. (2015c) reported that the variation in the results by Kachingwe et al. (2015) is mainly due to azimuthal dependency. Our results for MAUN agree with Yu et al. (2015c) with RFs clearly showing azimuthal variations indicative of anisotropy or dipping crustal layers under the station. However, we were only able to get a reliable result for RFs from back-azimuths of $180^\circ - 360^\circ$, giving a crustal thickness of 40 km and a Vp/Vs ratio of 1.87 (Figure S3 D-F in the supporting information). For back-azimuths of $0^\circ - 90^\circ$, we found many RFs with *Ps* conversions from the Moho with reversed polarities and inconsistent H-K results. The H-K analysis of all RFs provided a crustal thickness of 43 km with 1.80 Vp/Vs but this was strongly affected by the unreliable contribution of the RFs in the $0^\circ - 90^\circ$ back-azimuth (Figure S3 A-C in the supporting information).

4.4.3 Sedimentary Basins

The Passarge basin covers the Ghanzi-Chobe Belt, and is bordered by the Kalahari Suture Zone (KSZ) in the south (Figure 4.1). The KSZ is a Paleoproterozoic major thrust zone related to the formation of the Kheis-Magondi Belt and separates it from the Ghanzi-Chobe Belt and the Nosop Basin (Reeves, 1978). During Neoproterozoic and early Paleozoic times, the Passarge basin was formed by significant downthrow of sediments

that presently reach a thickness of up to 15 km (Haddon, 2005; Pretorius, 1984).

The H-K analysis of station NE209 showed two possible solutions. The first solution shows a crust thicker than 49 km and Vp/Vs value lower than 1.65 (Figure S11.1 in the supplementary material). The second solution shows a thin Moho of 37 km with relatively high Vp/Vs ratio of 1.84 (Figure S11.2 in the supporting information). We prefer the second solution since the first one has an unrealistic Vp/Vs ratio. Therefore, we reduced the maximum allowed depth range from 20-60 to 20-45 km for this station to avoid the solution with large crustal thickness and focus on the shallow local maximum of the H-K stacking function. However, because of the ambiguity of the solution, it is important to mention that the RF analysis may have a high uncertainty that goes beyond the formal error given in Table 4.1.

For station NE216 on the boundary of the Passarge basin, we initially obtained a crust thicker than 44 km and an unrealistically low Vp/Vs ratio (lower than 1.65, Figure S18.1 in the supporting information). After investigating the RFs, we found that the solution was biased due to unclear $P_sP_s + P_pS_s$ converted phases. Therefore, we changed the weights of the Moho converted phases from (6, 3, 1) to (7, 3, 0) to obtain a stable result with a Vp/Vs ratio of 1.75 and a crustal thickness of 37 km (Figure S18.2 in the supporting information). Comparing our crustal thickness estimate for NE216 with results for the nearby SAFARI stations B03 and B05, we observe that (Yu et al., 2015c) found larger crustal thicknesses with values of around 40-42 km (Figure 4.3 C-D). Similarly large variations have been observed among nearby stations in Botswana, e.g., between station B10 (40 km) and B12 (46 km) on the border of Congo Craton (Yu et al., 2015c) (Figure 4.3 C-D). On the other hand, the Vp/Vs ratio of 1.75 is close to, and within the uncertainty bounds of, the Vp/Vs estimates for B04 and B05 (1.76-1.79).

The relatively thin crust and high Vp/Vs ratio observed for the Okavango Rift Zone extending underneath the Passarge Basin (NE216) can indicate continuation of the rift initiation region until the border of the Kalahari Craton. This southward extension is further supported by the large magnitude earthquake (Mw 6.5) on the 4th of April 2017 at the border of the Kalahari Craton (Figure 4.3 C-D). The earthquake was located at ~ 30 km depth with an extensional mechanism that suggests rifting in this area (USGS, 2017). Future analysis using ambient noise surface wave tomography for the NARS-Botswana network will provide additional evidence for the proposed southward extension of the rift zone.

The Nosop Basin in the southwest of Botswana, which covers part of the Rehoboth Province, has a thick sedimentary sequence that reaches down to 15 km (Pretorius, 1984; Key and Ayres, 2000). The Rehoboth Province has an Archean nucleus and a major part of the province was formed during the Paleoproterozoic (2.2-1.9 Ga) (Van Schijndel et al., 2011). We found a crustal thickness of 42 km and a Vp/Vs ratio of 1.76 for station NE203 in the northwestern part of the Nosop basin (Table 4.1,

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and Figure 4.3 A-B). The relatively thick crust and the intermediate Vp/Vs value are similar to the observations at LBTB station in the Kaapvaal Craton suggesting a cratonic composition. A cratonic composition for this region has been proposed before but is based on two conflicting possible scenarios. It might be associated to the buried Maltahohe micro-craton, as suggested by Begg et al. (2009), or to an extension of the Kaapvaal Craton under the basin, as suggested by Wright and Hall (1990). Although there are similarities in thickness and Vp/Vs ratio between Kaapvaal and the crust underneath the Nosop Basin (NE203), our preferred scenario is that it is a separate craton. The Kheis-Okwa Belt and western end of the Kaapvaal Craton are significantly different in composition (with very low Vp/Vs ratios around or below 1.7), thereby separating the two cratonic blocks. Moreover, the deep sediments in the Nosop Basin suggest a different geodynamic setting with significant subsidence, which is not observed on the Kaapvaal craton. Note, however, that we only have one station in the Nosop Basin and another on the border with the Kheis-Okwa Belt, and strong conclusions can therefore not be made. Regional surface wave tomography can possibly provide more evidence for either of the two scenarios.

4.5 Summary and conclusions

We presented the first nationwide crustal thickness estimates for Botswana from receiver function analysis. We identified different tectonic regimes with significantly different crustal structures.

The cratonic regions (Kaapvaal, Zimbabwe, and Congo craton) are on average 37-42 km thick and have Vp/Vs ratios corresponding to intermediate to mafic composition, becoming more felsic towards the borders with the mobile belts. The main exception is western Kaapvaal with a Vp/Vs ratio of 1.67 that is indicative of the absence of a mafic lower crust, probably due to delamination of the lower crust and upper mantle caused by rifting in the Proterozoic as has been suggested by previous studies.

The mobile belts show an overall thicker crust with the exception of the Damara-Ghanzi-Chobe Belt which is affected by rifting processes. The thickest crust, of 49 km, is observed at the boundary between the Magondi Belt and the Zimbabwe Craton. The highest Vp/Vs ratio, of 1.87, is found in the northeast of Botswana for station MAUN that may be associated with incipient rifting at the Okavango Rift Zone. Another possible region of incipient rifting is found for the Passage Basin in central Botswana, which is characterized by a thin crust together with a high Vp/Vs ratio.

The Kheis-Okwa-Magondi Belt shows variable crustal thicknesses, with 40 km for the Magondi Belt (station NE208), 44 km for the Okwa Block, and 39 km for the Kheis Belt. The Vp/Vs ratio of this composite belt has relatively small values of 1.69-1.72, characteristic of the felsic nature of the belts surrounding the Kalahari Craton. The southern part

of the belt, the Kheis block, separates the western Kaapvaal Craton from another cratonic block that has been identified underneath the Nosop Basin. Previous studies suggested that this might be either an extension of the Kaapvaal craton or the so-called buried Maltahohe micro-craton. Our results are more supportive of the latter theory.

4.6 Acknowledgments

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4.7 Supporting information for: Crustal structure and dynamics of Botswana

4.7.1 Introduction

With this document we provide the H-K stacking results. The weights for all stations are given in Table T1. Then, Figure S1 presents the recorded data used in this study. After that, Figure S2 presents the stacked receiver functions per station and organized based on the tectonic province. Finally, Figures S3-S23 present the H-K results for each station where (A) shows H-K stacking results, (B) shows RFs ordered by back-azimuth, and (C) shows RFs ordered by epicentral distance. In case of sequential H-K analysis, S5 and S14, (A) shows H-K stack for the sedimentary layer, (B) shows RFs (Gaussian width = 5.0) ordered by back-azimuth, (C) shows RFs (Gaussian width = 5.0) ordered by epicentral distance, (D) shows H-K stack for crustal thickness, (E) shows RFs (Gaussian width = 2.5) ordered by back-azimuth, and (F) shows RFs (Gaussian width = 2.5) ordered by epicentral distance.

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Table T1.

Station	Latitude	Longitude	Crust weights	Sediment weights
LBTB	-25.02	25.60	0.6, 0.3, 0.1	
Maun	-19.90	23.53	0.6, 0.3, 0.1	
NE201	-24.52	23.93	0.6, 0.3, 0.1	
NE202	-24.11	21.78	0.6, 0.3, 0.1	0.2, 0.5, 0.3
NE203	-22.99	20.20	0.6, 0.3, 0.1	
NE204	-18.54	21.34	0.6, 0.3, 0.1	
NE205	-18.62	23.50	0.6, 0.3, 0.1	0.2, 0.3, 0.5
NE206	-17.80	25.16	0.6, 0.3, 0.1	
NE207	-19.53	21.17	0.6, 0.3, 0.1	
NE208	-21.95	25.45	0.6, 0.2, 0.2	
NE209	-21.40	23.77	0.6, 0.3, 0.1	
NE210	-21.36	21.22	0.6, 0.3, 0.1	
NE211	-22.85	22.21	0.6, 0.3, 0.1	
NE212	-23.38	24.66	0.6, 0.3, 0.1	0.1, 0.3, 0.6
NE213	-25.48	22.86	0.6, 0.3, 0.1	
NE214	-19.39	22.16	0.6, 0.3, 0.1	
NE215	-18.78	25.20	0.6, 0.3, 0.1	
NE216	-20.20	24.54	0.7, 0.3, 0.0	
NE217	-21.10	27.33	0.6, 0.3, 0.1	
NE218	-20.56	26.22	0.6, 0.3, 0.1	
NE219	-22.57	27.45	0.6, 0.3, 0.1	
NE220	-23.36	25.86	0.6, 0.3, 0.1	
NE221	-25.81	24.80	0.6, 0.3, 0.1	

4. Crustal structure and dynamics of Botswana

Figure S1

4.7. Supporting information for: Crustal structure and dynamics of Botswana

Figure S2

4. Crustal structure and dynamics of Botswana

Figure S3 MAUN

H-K analysis of station MAUN shows azimuthal variation of the receiver functions. Figure (A), (B), and (C) show the H-K stacking results for all RFs. Figures (D), (E), and (F) are the final results using RFs within $180^{\circ} - 360^{\circ}$ back-azimuth.

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Figure S4.1 LBTB

H-K analysis of all receiver functions of LBTB.

Figure S4.2 LBTB

H-K analysis of LBTBs' receiver functions binned by ray parameter.

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Figure S5 NE202

H-K analysis of station NE202 showing azimuthal variations of the receiver functions. Figure (A) to (F) show the H-K stacking results for all RFs. Figures (G) to (F) are the final results using RFs within $0^\circ - 180^\circ$ back-azimuth.

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Figure S6 NE203

Figure S7 NE204

Figure S8 NE206

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Figure S9 NE207

Figure S10 NE208

Figure S11.1 NE209

This is the result for station NE209 without reducing the depth range from 20-60 to 20-45 km.

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Figure S11.2 NE209

The result is obtained for station NE209 after reducing the depth range from 20-60 to 20-45 km.

Figure S12 NE210

Figure S13 NE211

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Figure S14 NE212

Figure S15 NE213

4.7. Supporting information for: Crustal structure and dynamics of Botswana

Figure S16 NE214

Figure S17 NE215

Figure S18.1 NE216

Results for station NE216 with the phase weights (6, 3, 1) showing that the result goes beyond the HK search limit.

4. Crustal structure and dynamics of Botswana

Figure S18.2 NE216

Figure S19 NE217

Figure S20 NE218

4.7. Supporting information for: Crustal structure and dynamics of Botswana

Figure S21 NE219

Figure S22 NE220

Figure S23 NE221

Ambient noise tomography of Botswana

Abstract

We used data of the NARS-Botswana broadband seismic network to construct the first Rayleigh wave group and phase velocity maps of Botswana using ambient seismic noise. Vertical component continuous noise records from 23 stations were cross-correlated to obtain empirical Green's functions between all possible station pairs. The interstation Rayleigh wave dispersion curves, measured by frequency-time analysis, were inverted to group and phase velocity maps in the period range from 3 to 30 s. The maps show a striking correlation with known geological units in Botswana. The deep sedimentary Passarge and Nosop basins show up as short-period low-velocity anomalies distinct from the surrounding Archean and Proterozoic blocks without a thick sedimentary cover. The Okavango Rift Zone exhibits low group and phase velocities over the entire period range, suggestive of a low shear velocity crust. At longer periods, with sensitivity to the lower crust, the Okavango low-velocity anomaly connects to another low-velocity anomaly in central Botswana. The nature of this anomaly is unclear, but may be related to incipient lower crustal rifting under central Botswana, extending the East African Rift System a few 100 kilometers further southwest. This interpretation is also supported by the nearby lower crustal magnitude 6.5 extensional event of 3 April 2017. Finally, the dispersion maps require high shear velocities in the lower crust/upper mantle beneath the sedimentary Nosop basin. These high-velocities can be related to the cratonic basement of the Rehoboth Province.

5.1 Introduction

The African continent comprises several Archean cratons surrounded by younger Proterozoic fold belts. Understanding the reworking of these cratonic blocks by accretion and rifting is fundamental to understand

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the formation of the continent and the evolution of its crust. In southern Africa, Botswana is crucially located to investigate the transitions between the various units; it has the Kalahari Craton in the south and east and the Congo Craton in the northwest (Fig. 5.1A). The Kalahari Craton was formed by the 2.7-2.5 Ga collision of the Zimbabwe and Kaapvaal Cratons, with the Limpopo Belt, a separate crustal block, caught in between (Begg et al., 2009). During the Paleoproterozoic this composite craton grew to the west and north by the Magondi-Okwa-Kheis Belt and the Rehoboth province (Jacobs et al., 2008; van Schijndel et al., 2014). Associated with the formation of Gondwana, the Damara Belt and Ghanzi-Chobe zone developed due to the Neoproterozoic collision between the Kalahari and Congo Cratons (Khoza et al., 2013b, and references therein). The Rehoboth Province and adjacent areas are presently covered by a thick sedimentary layer extending to more than 10 km in the Nosop and Passarge Basins (Pretorius, 1984). This hampers studies of the nature and margins of the Rehoboth province and adjacent mobile belts, although there is some evidence that Rehoboth has an Archean crustal component (Van Schijndel et al., 2011; van Schijndel et al., 2014). Thus, whereas the general framework of the tectonic evolution of southern Africa is largely understood, many aspects are still unresolved.

NARS-Botswana (Fig. 5.1A) is a temporary broadband seismic network specifically designed to image the crustal and upper mantle structure of Botswana (NARS-Website, 2013). It is the first seismic deployment covering the entire country with 21 well distributed stations that were installed since November 2013.

Here, data of the NARS-Botswana network, together with those of station LBTB of the Global Seismological Network (GSN) and Africa-Array (AA) station MAUN were used to derive Rayleigh wave group and phase velocity maps of Botswana using ambient seismic noise.

5.2 Data and Methods

5.2.1 Interstation group and phase velocity measurements

Rayleigh wave phase and group velocity measurements were obtained from continuous vertical seismograms recorded by NARS-Botswana, LBTB and MAUN between November 2013 and August 2016. We adopted the methodology of Poli et al. (2013) to extract the phase and group velocity measurements for 253 interstation paths. The data were corrected for instrument response, filtered from 1 to 50 s period, decimated to 1 second, and subjected to spectral whitening. For each station pair, the cross-correlation function was calculated for 4-hour segments, excluding segments with amplitudes larger than 10 times the standard deviation of the windowed data, and stacked over 3 months periods. Then, an automatic frequency-time analysis was applied (Bensen et al., 2007) with a signal-to-noise ratio threshold of 10 to extract the phase and group velocity dispersion curves for periods between 3 and 30 seconds from the

Figure 5.1 (A) The NARS-Botswana network with yellow triangles indicating the station locations. GSN station LBTB is in blue and AA station MAUN in green. Green lines indicate the tectonic boundaries adopted from Singletary et al. (2003) and McCourt et al. (2013). The gray transparent area represents the Okavango Rift Zone. (B) Sedimentary thickness map inferred from aeromagnetic data (modified from Pretorius (1984)) with earthquake distribution (IRIS service using all available catalogs between January 1900 to June 2017). The central Botswana magnitude 6.5 earthquake of 3 April 2017 and its main aftershocks are shown in red.

3-month stacks. The dispersion curves per interstation path were then averaged to obtain the final curves with their standard deviation. Figure 5.2 shows examples of the dispersion curves for the Nosop Basin (NE202-NE203) and the Kaapvaal Craton (LBTB-NE201). The group and phase dispersion curves for the Nosop basin show low-velocities corresponding to the thick sedimentary basin while the curves in the Kaapvaal Craton show relatively high-velocities corresponding to the cratonic crust.

5.2.2 Construction of group and phase velocity maps

We inverted the interstation dispersion data for group and phase velocity maps on a 1° by 1° grid using the method of Barmin et al. (2001). A penalty function is minimized depending on data misfit, model smoothness and perturbation to the reference model. The spatial smoothing is governed by weight α and Gaussian smoothness width σ . The perturbation to the reference model, or norm damping, is controlled by weight β and the path density contribution which is governed by parameter λ . We varied the inversion parameters systematically considering the data misfit and model resolution to find optimum values for the parameters and finally selected $\alpha = 100$, $\sigma = 100$ km which is approximately equal to the cell size, $\beta = 1.0$ and $\lambda = 0.5$. Figure 5.3 presents the phase

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Figure 5.2 Phase and group velocity dispersion of interstation paths LBTB-NE201 within the Kaapvaal Craton (green) and NE202-NE203 within the Nosop Basin (blue).

velocity inversion results for a period of 22 seconds. The resolution map shows that the maximum resolution is approximately 1° . Figure 5.4 gives an overview of the final results with group maps (Fig. 5.4A-F) and phase velocity maps (Fig. 5.4H-M) for periods of 5, 10, 15, 20, 25 and 30 seconds. Figures 5.4G and 5.4N show the corresponding sensitivity kernels for model AK135 (Kennett et al., 1995) modified with a 5 km sedimentary layer, a midcrustal boundary at 15 km and crustal thickness of 35 km. The sensitivity kernels show that the main resolving power of the group and phase velocity measurements is within the 60 km depth range.

Figure 5.3 Phase velocity inversion results for a period of 22 s with (A) the measured interstation path velocities, (B) the resolution in km, and (C) the final inversion result.

5.3 Results and Discussion

The phase and group velocity maps presented here give us first insights into the overall crustal structure of Botswana. The 5 s group and phase velocity maps roughly correlate with the sedimentary thickness map of Pretorius (1984) inferred from aeromagnetic data (see Fig. 5.1B). The Nosop basin in the southwest has the lowest velocities, keeping very low values up to 15 s group and 10 s phase velocity. This is consistent with the very large sedimentary thickness of more than 10 km found by Pretorius (1984), and it is also in agreement with deep seismic sounding data for the Nosop basin (Wright and Hall, 1990). Additional, albeit less pronounced, low-velocity areas in the 5 s group and phase velocity maps are found in central Botswana and in the northwest. The low-velocity area in central Botswana roughly matches the region of the 500 m sedimentary thickness contour of Pretorius (1984) (Fig. 5.1B), again suggesting a sedimentary effect. The area in the northwest corresponds to the Damara Belt plus part of the Ghanzi-Chobe zone. It includes the Okavango Rift Zone (ORZ), an incipient rift that is widely accepted as the terminus of the southwestern branch of the East African Rift System (EARS). We observe that the group and phase velocities in this region remain low at longer periods, indicative of low crustal shear velocities.

Relatively high group and phase velocities are obtained around central Botswana, roughly starting at the Magondi Belt, continuing towards the Zimbabwe Craton, the Limpopo Belt and ending with the southern part of the Kaapvaal Craton in Botswana. At short periods (~ 5 s) the pattern largely corresponds to the areas with little or no (≤ 250 m) sediments. At longer periods high group and phase velocities are still found for these regions of Archean (Kaapvaal, Zimbabwe, Limpopo) or Paleoproterozoic (Magondi) age, representing relatively high-velocity crust.

The low-velocity area in central Botswana is enigmatic. It is present at both short and long periods. At short periods it can be explained as an effect of the sedimentary cover, but its effect should diminish with increasing periods, as is observed for the Rehoboth Province with the (thicker) sediments of the Nosop basin on top. Instead, at periods longer than 20 s it remains and connects to the low group and phase velocities in the northwest, related to the ORZ. This suggests lower than average shear velocities in the upper crust, potentially extending to larger depths. The anomalous character of the region is further indicated by the recent magnitude 6.5 earthquake of April 3, 2017. It was located in the lower crust at the eastern margin of the low-velocity anomaly at the border between the Kaapvaal Craton and the Magondi Belt. It had an extensional source mechanism, typical for rifting events in Africa (USGS, 2017) and was located in the lower crust. Note that large earthquakes are very rare in Botswana (Craig et al., 2011), only very few earthquakes were reported in this region since 1900 (see Fig. 5.1B). Combining the information about low shear wave velocities in the lower crust, and the link of the anomaly at depth to the ORZ, we suggest that the EARS extends further

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south than previously thought and that we see potential incipient rifting in the lower crust of central Botswana.

Additional small scale features can be recognized within the high-velocity cratonic areas. A weak linear low-velocity anomaly linking up with the central low-velocity anomaly can be identified in the southeast. It is visible in the group velocity maps up to a period of 25 s (most clearly at a period of 20 s) and it is also faintly visible in the phase velocity maps to 20 s. It potentially is an extension of the Colesberg Magnetic Lineament in South Africa which marks the suture between the Late and Mid-Archean Kaapvaal blocks resulting from a subduction-related collision (2.7-2.9 Ga) (Schmitz et al., 2004; Youssof et al., 2013). Interestingly, the area is also characterized by noticeable earthquake activity (Fig. 5.1B).

The maps at longer periods show a low-velocity anomaly in the north-east at a latitude of $\sim 20^{\circ}\text{S}$ and with an $\sim \text{EW}$ orientation. It is most clearly visible in the 25-30 s phase velocity maps and matches the location of a conductive object at approximately 20-45 km depth found by magnetotelluric profiling. Miensopust et al. (2011b) suggest that it may have a different composition, but the anomaly may also be attributed to the presence of fluids.

The high phase velocities at periods of around 30 s measured between stations in the southwest (NE203, NE202, NE213) require high shear velocities for the Rehoboth Province in the uppermost mantle and/or lower crust. This may be associated to the Archean origin of the Rehoboth Province as suggested by deep seismic profiling (Wright and Hall, 1990) and geochronological data (van Schijndel et al., 2014). Note that the region is at the border of our network where the resolution of our tomographic maps is low. Additional data from stations surrounding Botswana and the inclusion of longer period measurements using earthquake data will improve the current data set and will enable the future construction of a full 3D shear wave velocity model.

5.4 Conclusions

Rayleigh wave group and phase velocity maps for periods between 3 and 30 s were determined using high quality data from the temporary NARS-Botswana network. The results give first insights into the crustal structure across the country, showing remarkable variations within the crust. The Passarge and Nosop sedimentary basins dominate the short period group and phase velocity maps with low-velocity anomalies. The longer period maps, with sensitivity to middle and lower crust, show a clear distinction between the Archean blocks in the south and east and the younger mobile belts blocks to the west. The maps also suggest an extension of the Colesberg Magnetic Lineament across the northwestern block of the Kaapvaal Craton, recognized as a faint lower velocity anomaly. Moreover, a linear low-velocity feature in the northeast, crossing the Magondi Belt and linking up with the low-velocities of the ORZ in

5.4. Conclusions

Figure 5.4 (A-F) Group velocity maps at periods of 5, 10, 15, 20, 25, and 30 s. (G) Group velocity sensitivity kernels at 5-30 s periods calculated for model AK135 with a modified crust (black solid line). (H-M) Phase velocity maps at periods of 5-30 s. (N) Phase velocity sensitivity kernels at 5-30 s periods for the modified AK135 model (black solid line). The black dashed line represents the Colesberg Magnetic Lineament. The red star indicates the magnitude 6.5 earthquake of the 3rd of April 2017.

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the northwest, may be related to another lower crustal weak zone. Most enigmatic is the low-velocity anomaly in central Botswana. Its anomalous character is confirmed by the occurrence of the magnitude 6.5 earthquake of 3 April 2017 at its eastern end, a region that was hitherto not known for any significant earthquake activity. The nature of the structure is unclear, but the combination of low shear wave velocities in the lower crust, and the link of the anomaly at depth to the ORZ, might hint at incipient rifting in the lower crust of central Botswana. This would imply that the EARS might extend further south than previously thought, not ending at the ORZ but actually continuing a few hundred km further southwest.

5.5 Acknowledgments

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Crustal and upper mantle shear wave velocity structure of Botswana

6.1 Introduction

In the previous chapters we have inferred the crustal (velocity) structure for Botswana, which has lead to new insights into the geodynamic setting of Botswana. We found indications for potential rifting processes in the lower crust of Botswana, at ~ 500 km further southwest of where we previously thought that rifting ends in Botswana. Our models went, however, only to maximum 50-60 km depth, and conclusive evidence could not be drawn from the analysis. To improve the depth range of our ambient noise group and phase velocity tomographic maps from the previous chapter to full lithospheric depths we extend the analysis here with earthquake surface wave data. We thereby aim to retrieve information at lithospheric depths of the margins of the two cratonic provinces the Kalahari Craton in the east (consisting of the Kaapvaal and Zimbabwe Craton cratonic segments) and the Congo Craton in the northwest and the series of mobile belts in between. The analysis will focus on the extension of the East African Rift System (EARS) that is composed of several discrete rift sectors with and without the magmatic surface expressions on the surface (Ebinger, 2005). The rift system is believed to extend to the south until the border of Botswana. However, on 3rd of April 2017, an earthquake with magnitude M_w of 6.5 occurred in Botswana around 500 km southwest of the supposed EARS terminus (Kinabo et al., 2007). It happened at the border of the Kaapvaal Craton, had an extensional mechanism, and was located far away from any tectonic plate boundary or known rifting zone. To image the lithospheric structure and visualize possible deep rifting processes, we used data from 51 seismological

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stations in Botswana and its surrounding to provide the first insight into the shear-wave lithospheric structure of Botswana using Rayleigh wave group and phase velocity dispersion measurements.

Figure 6.1 (A) Tectonic map of the region together with the distribution of seismic stations used in this study. (B) The geological units in Botswana. (C) Sedimentary thickness map with earthquake distribution. The central Botswana magnitude 6.5 earthquake of 3 April 2017 and its aftershocks are shown in red.

6.2 Seismological Data

We used data recorded by 51 stations from four seismic networks (Figure 6.1-A). The largest part of the data are obtained from the NARS-Botswana network (NARS-Website, 2013). NARS-Botswana has been deployed since November 2013 and consists of 21 seismic stations that are strategically distributed over the different tectonic units in Botswana. In addition to the NARS-Botswana stations, we used 17 stations from the Seismic Arrays for African Rift Initiation (SAFARI) network that was deployed between May 2012 until June 2014 (Gao et al., 2013a). The focus of the SAFARI network was to image the crustal and mantle structure underneath the Okavango Rift Zone. Furthermore, we used 10 stations from the AfricaArray Namibia network that has been recording since May 2015. Finally, we used 2 stations from the Global Seismological Network (GSN) (Butler et al., 2004) and one station from the AfricaArray network (Nyblade et al., 2008). Except for the NARS-Botswana data, all data were downloaded from the Incorporated Research Institutions for Seismology (IRIS) web service.

6.3 Methodology

We used two complementary methods to derive the 3-D shear wave velocity structure beneath Botswana: 1) ambient noise tomography (Bensen et al., 2007) where inter-station Rayleigh wave group and phase velocity measurements are obtained from ambient noise data at relatively short periods (3-30 s), and 2) Helmholtz tomography (Jin and Gaherty, 2015) that yields Rayleigh wave phase velocity maps between 30 and 100 s from teleseismic earthquakes. The combination of group (3-30 s) and phase (3-100 s) velocity maps provides a dataset that is used to infer the crustal and upper mantle shear wave velocity structure. The two methods are briefly described in the following subsections.

6.3.1 Ambient Noise Tomography

Ambient noise tomography has recently become a well established surface wave imaging technique that uses ambient noise recordings. The Green's function response between two receivers is estimated from the cross-correlation of long term seismic noise, thus effectively turning one of the receivers into a virtual source (Shapiro and Campillo, 2004; Shapiro et al., 2005). Pioneering work by Shapiro et al. (2005) has shown that the crustal structure can be imaged by tomography of short-period (5-20 s) surface waves obtained from ambient noise records.

Here we used the same methodology as in chapter 5 for the ambient noise analysis of Botswana. However, we updated the group and phase velocity maps using new data and additional stations in Botswana and Namibia. We adopted the methodology of Poli et al. (2013) to retrieve the

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Figure 6.2 The noise cross-correlation of NE201-NE218, its inter-station path, and FTAN without and with phase matched filter.

inter-station empirical Green's functions. The vertical component continuous records from all stations were corrected for instrument response, filtered (1-50 s period), down sampled to 1 s, and finally subjected to spectral whitening. Then, the records between all available station pairs were segmented into 4-hour time windows. We excluded segments with amplitudes > 10 times the standard deviation of the data in the processed time window. For each station pair, the 4-hour time segments were cross-correlated in the time domain and stacked over 3-month periods. Then, for each station pair, we applied automatic frequency-time analysis on each of the 3-monthly stacks (using signal to noise ratio of 10) to obtain the group and phase velocity dispersion curves (Figure 6.2). Finally, for each station pair, the 3-monthly dispersion curves were averaged to obtain the final curve with its standard deviation. The group velocity dispersion curves between 30-35 s showed large variations over the 3-monthly dispersion curves that resulted in large uncertainty. Therefore, we decided to exclude this period interval from further analysis. The final group (3-30 s) and phase (3-35 s) dispersion curves were used to derive their dispersion maps as will be described next.

The interstation dispersion data (Figure 6.3-A,D) were inverted using the method of Barmin et al. (2001) to determine group (3-30 s) and phase (3-35 s) velocity maps for a $1^\circ \times 1^\circ$ grid. The method depends on constructing tomographic maps by minimizing a misfit function that considers three terms: 1) data misfit, 2) model smoothness, and 3) the perturbation from the reference model. The minimization of each term is controlled by weighting factor that governs its contribution to the misfit function. We varied the weights of these terms systematically until we obtained acceptable results considering the resulting data misfit, the resolution, and the model perturbations (Figures 6.3 and 6.4). For further details about the method the reader is referred to Barmin et al. (2001) and Levshin et al. (2005).

Figure 6.3 Group and phase velocity inversion results for a period of 15 s. Shown are the measured interstation path velocities (left), the lateral resolution (middle) and the final dispersion maps (right).

6.3.2 Helmholtz Tomography

We used the Automated Surface Wave Phase Velocity Measuring System (ASWMS) that was developed by Jin and Gaherty (2015) to estimate phase velocities for periods between 30 and 100 seconds. ASWMS is applied to the vertical components of teleseismic earthquake records at nearby stations to calculate the multi-channel cross-correlations of the fundamental mode Rayleigh wave. A window function is applied to the cross-correlation waveforms to check the coherency of the fundamental mode Rayleigh wave signal. The coherent signals are then subjected narrow band-filters around target frequencies using low and high cut-offs of $\pm 10\%$ of the central period. The initial phase delay is estimated by minimizing the misfit between the coherent band-filtered waveforms and a predicted five-parameters wavelet. The initial phase delays are inverted to obtain initial phase velocity maps using the Eikonal equation. The (frequency-dependent) amplitudes of the (frequency-dependent) wavelets provide a good approximation of the amplitude of the power spectrum. Finally, these amplitude measurements are used to correct the phase velocity maps for the effects of multipathing using the Helmholtz equation. For more details the reader is referred to Jin and Gaherty (2015).

We used earthquake records between June 2012 to May 2017, with magnitude $M_w > 6$, and distance larger than 30 degrees from the center

6. Crustal and upper mantle shear wave velocity structure of Botswana

Figure 6.4 Group and phase velocity maps obtained from ambient noise tomography.

of the study area. We used all earthquakes recorded by more than 10 stations. We finally used 469 earthquakes to perform earthquake Helmholtz tomography (Figure 5). We estimated the phase velocity maps at periods between 30 and 35 s with one second intervals, and between 35 and 100 s with five second intervals. A selection of the resulting phase velocity maps are shown in Figure 6.6.

Figure 6.5 The earthquake distribution (left) and the azimuthal distribution of the events (right).

Figure 6.6 The phase velocity maps from Helmholtz tomography.

6.3.3 Combinning ambient noise and Helmholtz dispersion maps

At this stage, from ambient noise, we have group and phase velocity maps between 3-35 s. On the other hand, we have phase velocity maps from Helmholtz tomography between 30 and 100 s. The phase velocity dispersion maps in the overlapping range between 30-35 need to be merged in order to provide a smooth transition from the ambient noise to the Helmholtz tomography maps. The merging process was carried out using a weighting scheme that guarantees a gradual transition from the ambient noise measurements to the Helmholtz tomography using the following formula:

$$C = \begin{cases} C_{an} & T < 30 \text{ s} \\ \frac{(W_{an} * C_{an}) + (W_{eq} * C_{eq})}{W_{an} + W_{eq}} & 30 \text{ s} \leq T \leq 35 \text{ s} \\ C_{eq} & T > 35 \text{ s} \end{cases} \quad (6.1)$$

with

C : the resulting phase velocity map of the merging process at period T .

C_{an} : the ambient noise phase velocity map at period T .

6. Crustal and upper mantle shear wave velocity structure of Botswana

- W_{an} : the weights of the ambient noise phase velocity maps.
(for periods increasing from 30 to 35 s the weights decrease from 5 to 1).
 W_{eq} : the weights of the Helmholtz tomography phase velocity maps.
(for periods increasing from 30 to 35 s the weights increase from 1 to 5).

The previous formula states that the ambient noise phase velocity maps were used for periods below 30 s and the Helmholtz phase velocity maps were used for the periods higher than 35 s. For the periods between 30 and 35 s, relative weights were used (decreasing from 5 to 1 for ambient noise, and increasing from 1 to 5 for Helmholtz tomography) to guarantee a smooth transition between the ambient noise and the Helmholtz tomography phase velocity data. This scheme was preferred over other merging schemes because it prevented creating abrupt changes in the dispersion curves that might lead to artifacts at a later inversion stage.

6.3.4 Joint Inversion of Group and Phase Velocity Dispersion Curves

The group and phase velocity maps provide the local variations of the group and phase velocity at each grid point. Group and phase velocity curves at each grid point were jointly inverted to produce the local 1D shear wave velocity structure at that location. The inversion of all grid points results in a 3D model of the crust and upper mantle. The inversion was carried out using iterative damped least squares optimisation based on the Levenberg-Marquardt algorithm (Moré, 1978) similar to Greve et al. (2014). We used an earlier version of the Mineos software for the forward calculation of the fundamental mode Rayleigh wave group and phase velocities (Woodhouse and Doornbos, 1988). The 1-D inversion approach aimed to minimise an objective function f (depending on measured data $\mathbf{d}$ and 1-D subsurface model $\mathbf{m}$) that consists of three misfit terms: the data misfit, the vertical smoothness, and the norm of the perturbation from the reference model following the equation:

$$\min(f(\mathbf{d}, \mathbf{m})) = \| (G(\mathbf{m}) - \mathbf{d}) + \alpha (\text{Model smoothness}) + \beta (\text{Model perturbation}) \|_2^2 \quad (6.2)$$

where the first term represents data misfit. $G(\mathbf{m})$ is the calculated dispersion data for model $\mathbf{m}$, and $\mathbf{d}$ is the measured dispersion curve. The model smoothness term is calculated using the first difference of the 1-D model following Julià et al. (2000), and α is a parameter controlling the contribution of the model smoothness term to the objective function. The model perturbation term is calculated using the difference between the 1-D reference subsurface model and the estimated model, and β is a parameter controlling the contribution of the norm of the perturbation from the reference model to the objective function.

Figure 6.7 The 1-D inversion of the group and phase velocity dispersion curves. (A) shows the inversion of the average group and phase dispersion curves to derive the average V_s velocity structure in Botswana using the modified AK135 model as a starting model. The subplot in (A) shows the reference, measured, and inverted group and phase dispersion curves. (B) shows the inversion of the group and phase velocity dispersion curves at grid point 25° East and -23° South. The subplot in figure (B) shows the location of the grid point on the map of Botswana. (C) shows the sensitivity of the phase velocity with depth using the 1D the average V_s velocity structure in Botswana that is plotted in (A). (D) shows the sensitivity of the group velocity with depth using the 1D the average V_s velocity structure in Botswana (A).

We inverted the group (3-30 s) and phase (3-100 s) dispersion curves to estimate the 3D subsurface structure of Botswana. We first inverted the average group and phase velocity curves to estimate the average 1-D shear-wave velocity structure in Botswana. We used an interpolated version of AK135 as a reference model as used by Kgaswane et al. (2009) in South Africa. Then, the average 1-D subsurface model was used as a reference model to invert the group and phase velocity curves at all grid points in Botswana. The inversion was carried for the shear-wave velocity only. For the crustal part with $V_s < 4.1$, the P-wave velocity and density were coupled to shear-wave velocity using the empirical relation found by Brocher (2005). For the deeper part of the model with $V_s > 4.1$, the P-wave velocity and density were kept in fixed ratios to shear-wave velocity.

The 1-D models were discretized from the surface until the mantle discontinuity at 410 km depth. We discretized the 1-D models using 2.5, 5, and 10 km depth steps for the crust (to 35 km, which is the average crustal thickness in Africa), the uppermost mantle (to 100 km), and the upper mantle (410 km), respectively. The increasing depth steps were chosen to mimic the decrease in short scale length sensitivity of the long period Rayleigh wave dispersion measurements (Figure 6.7-C, D). The

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inversion of the average group and phase velocity curves is presented in Figure 6.7-A. The inversion of a single grid point is presented in Figure 6.7-B.

6.4 Results and Discussion

6.4.1 Group and Phase Velocity

The group and phase velocity maps provide insight into the variations of the shear-wave velocity structure. The group (3-35 s) and phase (3-25 s) velocity dispersion maps are approximately sensitive to the crustal structure. The longer period phase velocity maps (25-100) are more sensitive to the upper mantle (Figure 7C-D).

The short period group (3-35 s) and phase (3-25 s) velocity data show a correlation with well known tectonic features. The Okavango Rift Zone, as well as the Passarge and Nosop basins have a clear low shear velocity signature on the short period dispersion maps that distinguish them from the surrounding tectonic blocks (Figure 6.4: 5-30 s group and 5-15 s phase velocity maps). On the intermediate and long period phase velocity maps (25-100 s), a low velocity zone in central Botswana at 30 s is connected to a low velocity zone in the northeast at 40 s and longer periods (Figure 6.6). At 100 s phase velocity, a clear high velocity anomaly at the northwestern corner can be linked to the deep root of the Congo Craton.

6.4.2 Crust and upper mantle shear-wave velocity structure

6.4.2.1 The crust

The 3D shear wave velocity structure shows two low velocity zones that are related to the deep sedimentary basins in Botswana (Figure 6.8A-B). In the southwest, the Nosop basin has a low shear wave velocity that extends to 15 km depth. Cross section A-A' shows the Nosop sedimentary basin as a deep low velocity root that continues to 20 km depth in the central part of the section. The Passarge Basin in central Botswana has low shear-wave velocities in sections C-C' and B-B' that extend further to the southeast until the border of the Kalahari Craton. A zone of relatively high velocities is present beneath the Nosop Basin (section A-A'). It can be linked to a deep cratonic root as will be discussed later. On the other hand, the Passarge Basin shows a connection, at its southeastern border, to a low velocity zone that extends into the mantle (section B-B').

The Okavango Rift Zone in sections C-C' and D-D' is identified as a crustal low velocity anomaly that is connected to the mantle as shown in section D-D'. The low crustal velocity is compatible with a high V_p/V_s ratio found by recent receiver function studies (Yu et al., 2015c; Fadel

et al., Under review), and a shallow Curie depth of 8-15 km from 3D inversion of magnetic data (Leseane et al., 2015). Our results and the previous observations support the hypothesis of ascending hot fluids in the crust from the mantle through weak zones in the lower crust (Leseane et al., 2015; Yu et al., 2015c; Fadel et al., Under review). These hot ascending fluids can be the main cause for the noticeable earthquake activity at the Okavango Rift Zone (Figure 6.1-C).

Sections B-B' and C-C' run through the location of the 6.5 M_w Botswana earthquake. The earthquake occurred within a low velocity zone that is directly connected to the strongest low velocity zone in the mantle that is found beneath northern Botswana (section B-B'). Section C-C' shows that the low velocity zone at the earthquake location seems to have a connection, through a narrow weak zone, to deeper low velocity anomalies beneath northwestern and central Botswana. The profile C-C' runs from the Okavango Rift Zone through the Passarge Basin and ends at the border of the Kaapvaal Craton. It crosses the seismicity of the Okavango Rift Zone and appears to align with the pattern of the high magnitude earthquakes (4.5-6.5 M_w) extending from the Okavango Rift Zone to the border of the Kaapvaal Craton (Figure 6.1-C). Moreover, it also agrees with the low velocity segments that rise from the upper mantle to the crust in section C-C' (moving from northwest to southeast). The contact with low-velocity mantle material, suggestive of elevated temperature or partial melting, through a lower crustal weak zone may affect the crustal properties and may explain the earthquake activity in this zone.

6.4.2.2 The upper mantle

The uppermost mantle structure of Botswana is dominated by the high shear-wave velocities of the cratons and low shear wave velocities in central and northern Botswana that extend to the borders of the cratons.

The Kaapvaal Craton is characterized by high shear wave velocities in the southeast. Section A-A' shows the deep cratonic root of the Kaapvaal Craton that extends to 200 km depth with a positive anomaly of more than 2% that increases in magnitude with depth. This is in agreement with body wave tomographic images from James et al. (2001), Fouch et al. (2004) and Youssof et al. (2015) that showed that the deep keel of the Kaapvaal Craton extends to more than 250 km depth. Moreover, the Zimbabwe Craton also shows up as a high velocity anomaly with a deep root, as can be seen in the 180 km depth slice. Furthermore, it seems that the deep root of the Zimbabwe Craton has a higher shear wave-velocity than the Kaapvaal Craton, potentially reflecting a slightly different composition or temperature.

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Figure 6.8 The 3D inversion results.

At the western part of section A-A', a positive shear-wave velocity anomaly is found ($\sim 3\%$) approximately between 50-90 km depth. This is suggestive of a deep cratonic root for the Rehoboth Province as suggested by various studies (Wright and Hall, 1990; Begg et al., 2009). A recent magnetotelluric study by Muller et al. (2009) showed that the Rehoboth Province has a similar lithospheric conductivity as the western block of the Kaapvaal Craton. A major part of the Rehoboth province was formed during the Proterozoic (2.2-1.9 Ga) around an Archean nucleus (Van Schijndel et al., 2011). The high shear wave velocities between 50-90 km may be related to remnant oceanic lithosphere, because it has been suggested, based on a petrological analysis, that the Rehoboth Province is formed in an oceanic environment (van Schijndel et al., 2014).

Between the Kaapvaal Craton in the east and the deep cratonic root of the Rehoboth Province in the west, a zone of low shear wave velo-

cities appears. It can be related to the Kheis Belt and the Okwa Block. Petrological analysis of rock samples from Okwa Basement Complex by Mapeo et al. (2006) have shown that the Okwa block was emplaced during a narrow time span at 2056 ± 3 Ma. The timing corresponds with the intrusion of the Bushveld Complex in the Kaapvaal Craton in South Africa at ~ 2.05 Ga and suggests a link between the two events. Recent P- and S-wave body wave tomography results by Youssof et al. (2015) have shown that the Bushveld Complex is a dominant feature of the South African upper mantle, showing up as a low V_p anomaly and non-connected low S-wave velocity lobes. Given the evidence from the petrological analysis and the arguments of the non-connected lobes of the S-wave tomography, we interpret this low shear-wave velocity zone as another lobe linked to the Bushveld Intrusion that may be connected at deeper depths. Future body-wave tomography may provide additional evidence for this interpretation.

The northeastern tip of Botswana has the lowest shear-wave velocity of the 3-D mantle model. This low-velocity anomaly (section B-B') extends further to the Okavango Rift Zone (section D-D') and to the border of the Kalahari Craton (section C-C'). The 6.5 M_w Botswana earthquake may be related to this anomaly because it occurred within a low-velocity zone that is directly connected to this negative velocity anomaly (section B-B'). The low velocity zone is suggestive of upwelling from the mantle, which may explain the extensional mechanism of the earthquake and its large focal depth (29 km, USGS). The strong negative mantle velocity anomaly at the northern tip of Botswana may be the deeper extension of the East African Rift System since it is located at the extension of the southwestern branch of the EARS with Lake Kariba as its last surface expression. The negative mantle anomaly extends from the east to west beneath the Okavango Rift Zone in northern Botswana (section D-D'). Moreover, section C-C' shows that it also extends from the north to the south until the border of the Kaapvaal Craton.

The western part of section D-D' shows two high velocity zones at ~ 50 -100 km and ~ 160 -200 km depth. Their location agrees with the location of the high V_p anomaly from teleseismic body wave tomography of SAFARI data by Yu et al. (2017), who interpreted the anomaly as a deep root of the Congo Craton. Our observations suggest that the two high shear-wave velocity anomalies belong to the Congo Craton, and that the low-velocity anomaly in between is linked to the EARS. This low velocity anomaly of the EARS appears to erode the margin of the Congo Craton in a similar way as the Kaapvaal Craton: section B-B' shows that a fingerlike low-velocity anomaly seems to erode the edge of the craton at the center of the section.

Finally, our results show a low velocity anomaly (~ 30 -50 km) under the Okavango Rift Zone in the central part of section D-D'. The low velocity anomaly appears to be connected to a deeper upper mantle low-

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velocity anomaly at the western end of section D-D'. This may explain the earthquake activity of the Okavango Rift Zone, which has been attributed to ascending fluids from the mantle through lower crustal weak zones (e.g., Leseane et al., 2015; Yu et al., 2015c; Fadel et al., Under review).

6.5 Conclusions

We presented the first 3D shear wave velocity model of Botswana using data from 51 seismological stations in Botswana. We used two complementary methods, ambient noise and Helmholtz tomography, to derive Rayleigh wave group (3-30 s) and phase velocity (3-100 s) maps of Botswana. The group and phase velocity data were jointly inverted to determine a 3D shear wave velocity model beneath Botswana.

Our results show that the upper mantle low velocity expression of the EARS extends from the northern tip of Botswana until the border of the Congo Craton in the west and the Kaapvaal Craton in the south. This low velocity anomaly seems to erode the margins of the cratons by finger-like low-velocity bodies.

The upper mantle low velocity anomaly of the EARS seems to affect the crust through weak low velocity zones at the Okavango Rift Zone and the border of the Kaapvaal Craton. This extension of the EARS until the border of the Kaapvaal Craton may explain the occurrence of the 6.5 M_w Botswana on the 3rd of April 2017 with its deep crustal hypocenter and its extensional mechanism.

Our results further confirm the presence of deep roots for the Kaapvaal and Zimbabwe cratons with high velocity anomalies extending to more than 200 km depth, and we also observe cratonic roots for the Rehoboth Province and the Congo Craton.

6.6 Acknowledgments

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tomography codes are freely distributed through University of Colorado Boulder website. The Helmholtz tomography code is freely distributed on GitHub through Lamont Doherty Earth Observatory, Columbia University. Maps and figures were produced using GMT and Python.

3D object-oriented image analysis in 3D geophysical modelling

7

Abstract

Geophysical data are the main source of information about the subsurface. Geophysical techniques are, however, highly non-unique in determining specific physical parameters and boundaries of subsurface objects. To obtain actual physical information an inversion process is often applied, in which measurements at or above the Earth surface are inverted into a 2- or 3D subsurface spatial distribution of the physical property. Interpreting these models into structural objects, related to physical processes, requires *a priori* knowledge and expert analysis which is susceptible to subjective choices and is therefore often non-repeatable. In this research we implemented a recently introduced object-based approach to interpret the 3D inversion results of a single geophysical technique using the available *a priori* information and the physical and geometrical characteristics of the interpreted objects. The introduced methodology is semi-automatic and repeatable, and allows the extraction of subsurface structures using 3D object-oriented image analysis (3D) in an objective knowledge-based classification scheme. The approach allows for a semi-objective setting of thresholds that can be tested and, if necessary, changed in a very fast and efficient way. These changes require only changing the thresholds used in a so-called ruleset which is composed of algorithms that extract objects from a 3D data cube. The approach is tested on a synthetic model which is based on *a priori* knowledge on objects present in the study area (Tanzania). Object characteristics and thresholds were well defined in a 3D histogram of velocity vs. depth, and objects were fully retrieved. The real model results showed how 3D OOA can deal with realistic 3D subsurface conditions in which the boundaries become fuzzy, the object extensions become unclear, and the model characteristics vary with depth due to the different physical conditions. As expected, the 3D histogram of the real data was substantially more complex. Still, the 3D OOA-derived objects

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were extracted based on their velocity and their depth location. Spatially defined boundaries, based on physical variations, can improve the modeling with spatially dependent parameter information. With 3D OOA the non-uniqueness on the location of objects and their physical properties can be potentially significantly reduced.

7.1 Introduction

Geophysical data are the main source of information about the subsurface. To obtain actual physical information an inversion process is applied, in which measurements at or above the Earth surface are inverted into a 2- or 3D subsurface spatial distribution of the physical property related to these measurements (e.g., seismic wave subsurface distribution in case of seismic measurements). However, the resulting 3D models need further interpretation to describe the subsurface lithological and structural components or to define the subsurface anomalous target e.g., mineral ore, fluid intrusion or specific target object.

It is well known that a single geophysical technique cannot determine the lithology, and that the usage of multiple geophysical techniques is more reliable due to the overlap between the different lithological units within the same range of values of the physical parameters, e.g., limestone and granite have the same electrical conductivity values but differ in magnetic susceptibilities (Bedrosian, 2007). A lot of work has been done on combining two or more geophysical techniques to define an explicit lithological interpretation of the studied areas, e.g., Bosch (1999), Bedrosian et al. (2007), Guillen et al. (2008), Infante et al. (2010), and Bauer et al. (2012). The joint structural and lithological interpretation of different geophysical techniques was done based on pixel-by-pixel analysis for the correlated patterns in the resulting 3D models, and assigning trends to distinct lithological units (Gallardo and Meju, 2011). However, these approaches were focusing on defining specific ranges of physical parameters, e.g., velocity and resistivity, for each classified subsurface object. They did not take in consideration other parameters such as depth or geometry, which can help in the classification process.

For a single geophysical technique, e.g., seismic reflection/refraction or tomography, the interpretation and the classification have been done without a standardized methodology and were often done based on contouring or expert interpretation of the 3D seismic velocity data cube and *a priori* information about the studied area (e.g., Ebbing et al. (2001), Li (2010) and Sanchez et al. (2011)). However, this can be sensitive to subjective choices that can often make the process unrepeatable. The mentioned approaches, either for single or joint interpretation techniques, were mainly voxel-based (the 3D nomenclature of pixel). However recent object-based analysis techniques have proven a superior performance over the pixel-based ones. However, to our knowledge, they were

never applied in 3D geophysical interpretation.

In this research we implement a recently introduced object-based approach to interpret the 3D inversion results of a single geophysical technique using the available *a priori* information and the physical and geometrical characteristics of the interpreted objects. The introduced methodology is semi-automatic and allows the extraction of subsurface structures using 3D object-oriented image analysis (3D OOA) in an objective knowledge-based classification scheme. 3D OOA is an advanced image analysis technique that has originated in the biomedical field (Schönmeyer et al., 2006). To our knowledge, it is applied here for the first time to 3D seismological data to identify 3D objects. The study area in which we apply this methodology is the Tanzania Craton and surroundings. The method is first tested on a synthetic model that simulates the study area. Subsequently, it is used to extract 3D subsurface objects from a real 3D seismic tomographic model of the study area. However, it should be kept in mind that this work mainly targets the subsurface structures or objects which may not be the same as lithological classification, which needs more physical parameters to be done.

7.2 Study area

The study area is the central part of the East African Rift System (Figure 7.1 (A)) for which Adams et al. (2012) created a 3D shear wave seismic tomography model. It is a voxel-based model that ranges from a depth of 50 km down to 400 km at 10 km increments. The subsurface model can be divided into two main parts (Adams et al., 2012). The upper part of the model ($< \sim 200$ km) is characterized by the high velocities of the Tanzanian Craton and the surroundings basement complex, and low velocities of the surrounding tectonic elements, such as rift branches, Proterozoic mobile belts and Cenozoic volcanism. The lower part is dominated by low velocities related to a plume head, or part of the deep mantle super plume in western Africa (see Adams et al. (2012) for more details about the 3D shear wave tomography model).

7.3 Methodology

7.3.1 Synthetic model creation

The synthetic model is a 3D voxel velocity model that was designed in IGMAS+ (3D Interactive Gravity and Magnetic Application System (Götze and Lahmeyer, 1988; Schmidt et al., 2010)) based on the interpretation of the tomography model of the study area provided by Adams et al. (2012). The vertical and lateral dimensions are close to reality, and also the velocity variations of the different objects are physically realistic. It contains four main objects: craton (C), rift (R), plume (P) and three horizontal lay-

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Figure 7.1 (A) Elevation map of the study area with inset map showing the location of the study area within Africa. (B) The shear wave tomography model of the study area. Adapted from Adams et al. (2012).

Figure 7.2 The composition of the synthetic model. (A) five vertical sections that compose the synthetic models; (B) the detailed explanation of the objects vertical section number 3, the model elements are: Rift (R), craton (C), layer (L), boundaries between rift and layers (RL), boundaries between rifts and craton (RC), boundaries between plume and upper objects (PB), and plume (P) (the colors of the section do not describe the velocity of the objects). (C) the detailed explanation of vertical section number 3 in the grey scale of the 3D image stack.

ers (L1, L2, and L3) (Figure 7.2). The craton and rifts were equally divided into three zones (C1, C2, C3, and R1, R2, and R3, respectively) to simulate increased velocity with depth (Figure 7.2). Vertical gradual boundaries (RL and RC) were implemented to describe the transition from rifts to surroundings (either layers or craton, respectively), and a horizontal gradual diffuse boundary (PB) was defined between the plume and objects above. The model was gridded into depth slices using Ordinary Kriging, and subsequently converted into a 3D image stack using XML code to be used for 3D OOA in eCognition, the most widely used commercial OOA software package (Blaschke (2010); www.ecognition.com).

7.3.2 Real model preparation

The S-wave seismic tomography model (Figure 7.1 (B)) was converted into a 3D image stack to be used in eCognition. It included 36 slices ranging from a depth of 50 km to 400 km with a 10 km depth interval and 0.5 degree horizontal spacing in X and Y directions. The slices were gridded using a grayscale color scheme with a linear color distribution, as commonly used in the biomedical field (e.g., Schönmeier et al. (2006)). Then they were converted into a 3D image stack in the same way as the synthetic model.

7.3.3 3D object-oriented analysis

In general terms, object-oriented analysis (OOA) is a 2-step procedure, comprising image segmentation (creation of contiguous segments or objects that are relatively homogeneous in terms of adjacent pixel or point values), and their subsequent classification. The principal advantage over traditional pixel-based methods is that the generated segments are associated with attributes (e.g., spectral, textural, geometric, contextual, or physical), which are employed in the object classification. OOA is thus a form of knowledge-driven analysis that mimics the approaches of cognitive visual image analysis (Martha et al., 2011; Kerle and de Leeuw, 2009).

Released in 2000, eCognition was the first commercial OOA software, and has become the most frequently used tool, with some 50-55% of all published scientific studies on OOA using it (Blaschke, 2010). The segmentation/classification process in eCognition is implemented through rulesets. This allows complex processing schemes to be developed that can comprise a number of different segmentation and classification routines, supported by additional object refinement steps. The rule-set is semi-automatic since the used segmentation and classification algorithms are already developed and existing in eCognition package. The user needs only to define the optimum thresholds to define the target segmentation scheme or to extract the target object. 3D OOA is a recent extension of these concepts, allowing the processing of data cubes such as created from laser scanning data (LiDAR) or tomographic analysis. The origins of this extension lie in the biomedical field, such as the processing of 3D X-Ray computed tomography images (CT; Schönmeier et al. (2006)). More recently the technique has also been tested in the remote sensing community through an example of 3D OOA processing of LiDAR point cloud data (Willhauck, 2012).

Also in the 3D domain the process starts with the segmentation of an image stack into relatively homogenous segments that, together with their associated attributes, are used in the subsequent classification process. The main challenge facing OOA is the determination of the thresholds for both segmentation and classification steps, and the se-

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Table 7.1 An overview of applied segmentation and classification algorithms adapted from eCognition Developer (2011a,b)

<i>Algorithm</i>	<i>Description</i>
<i>Multi-threshold segmentation:</i>	It splits the images into objects and classifies them based on pre-defined thresholds that can be user defined or can be adapted with the combination of other automatic approaches.
<i>Classification:</i>	The classification algorithm evaluates the likelihood that an image object belongs to a specific class based on a class description that can be a threshold or membership function for one or a combination of the different object features. Those can be spectral, textural, geometrical or contextual features. The algorithm allows also the use of fuzzy logic conditions to describe such functions.
<i>Assign Class:</i>	The algorithm determines whether an object belongs to a class or not based on a single threshold condition.
<i>Pixel-based object resizing:</i>	It allows a user defined object to grow, shrink or be coated in a user defined direction (x and/or y and/or z), or a user defined object class, or within specific pixel values. The coating option allows the "coating" pixels on objects to be classified into another class, which is different from the growing option that allows the object itself to grow in the surrounding pixels.
<i>Merge region:</i>	The algorithm merges adjacent segments belonging to the same class into one segment.
<i>Convert image objects:</i>	It connects the 2D segments into 3D segments.
<i>Export view:</i>	It exports the classification results with their georeference information into one of different raster formats, e.g., jpeg or tif.

lection of suitable segment attributes. They usually are based on user interpretation (knowledge-driven), or on statistical data analysis techniques, such as Random Forests (Stumpf and Kerle, 2011). In the present work, the thresholds necessary for defining the 3D objects are based on analysis of the 3D histogram of velocity vs depth of the 3D subsurface model, supported by *a priori* geological knowledge about the subsurface objects, available from previous studies. The inferred thresholds guided the segmentation and subsequent classification of the generated segments through different algorithms as summarized in Table 7.1.

Figure 7.3 3D histogram of the synthetic model explaining the process of the object extraction based on the object velocity values and their depths. Rift (R), Craton (C), layer (L) boundaries between rift and layers (RL), boundaries between rift and craton (RC), boundaries between plume and upper objects (PB), and plume (P).

7.4 Results

7.4.1 3D OOA of the synthetic model

The 3D histogram of the synthetic model (Figure 7.3) describes the distribution of velocity values with depth. The separability of different objects was used to define the main characteristics of the objects in the 3D image stack (Figure 7.2). The 4 main features of the synthetic model, the plume (P), craton (C), rift (R) and horizontal layering (L1, L2, and L3), can be identified in the 3D histogram as distinct objects. The gradual boundaries between the main features (RL, RC, and RB) (Figure 7.2) result in small features that are partly overlapping with the main features.

The cratonic parts (C1, C2, and C3) are distinguishable as three blocks with higher velocity values (4.575, 4.475, and 4.375 km/sec, respectively) in the intermediate to shallow depth level. The plume (P) is visible on the opposite side of the histogram at the deep levels, but has a relatively low velocity. In between the rift (R1, R2, and R3) and the horizontal layers (L1, L2, and L3) are positioned. The rift, on average, has low velocity values of 3.975, 4.075, and 4.175 km/sec for the R1, R2 and R3 depth levels, respectively. The horizontal layers are located at the same depth intervals as the rifts, but have intermediate velocity values of 4.175 km/sec, 4.275 km/sec, and 4.375 km/sec for layers 1, 2, and 3, respectively.

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The complexity in the interpretation of the 3D histogram comes from the gradual boundaries (RL, RC, and PB) that have intermediate velocity values in between the 4 main features. The boundary between two objects was divided into 3 parts that have gradually changing velocity values. For example, RL1 (Figure 7.2 (B)) was divided into 3 parts RL1-1, RL1-2, and RL1-3 with velocity values of 4.025, 4.075, and 4.125 km/sec, respectively. They describe the transition between the first level of the rift (R1: 3.975 km/sec) and layer 1 (L1: 4.175 km/sec) (see the 3D histogram in Figure 7.3). Some of these boundaries (Figure 7.3) have a distinctly different velocity (e.g., RL1-1), some overlap with the velocity of other units and are only distinguishable in depth (e.g., C2 and RC3-3, and also RL1-2 and RC1-1), and some are overlapping in both velocity values and depth ranges (e.g., L1 and RC1-2, as well as L3, RC3-2, and PB3). Layers 1, 2, and 3 share the same velocity values and depth ranges with the boundaries between the craton and rift (RC1-2, RC2-2, and RC3-3, respectively). And although the rift boundaries at the different levels (RL1-1 to RL3-3 and RC1-1 to RC3-3) have a shallow depth (50-220 km), and the plume boundaries (PB1-PB3) have a deep location (190-295 km), there was a zone (190-220 km) in which both boundaries were mixed in depth but could be separated by velocity (Figure 7.2 (B) and (C)).

The ruleset is based on these distinctions and often use a combination of velocity values and depth. The object extraction process was done in 5 steps.

In step 1 segments were created and classified based on velocity values through the *multi-threshold segmentation* algorithm (Table 7.1). Eleven thresholds were used to create the segments and classify them into twelve velocity classes based on the histograms of the 3D image stack (Figure 7.3).

Step 2 classified the plume and the craton based on their velocity and depth location, using the *assign class* algorithm (Table 7.1). The plume was classified using a velocity of 4.075 km/sec and a depth of >225 km (Figure 7.4 A). The craton was classified using velocity values of 4.575 km/sec for C3, and 4.375 km/sec at a depth of ≤ 125 km for C1, and 4.475 km/sec at a depth of ≤ 170 km for C2 (Figure 7.4 (B)).

In the next step the rifts (R1-R3) were classified. For R1 and R2 the same approach was followed as for step 2. They were classified based on their velocity values and depth (3.975 km/sec for R1, and 4.075 km/sec with a depth range of 120-200 km for R2), also using the *assign class* algorithm. Only the depth range was used to classify R2 and thereby prevented classifying other objects (P, RL1-2, and RC1-1) with the same velocity value, but at different depths (Figure 7.3). To classify R3 a different approach was followed (*pixel-based object resizing*) (Table 7.1). This was necessary due to the presence of PB1 with the same velocity and within the same depth range. The rift (R2) was allowed to grow vertically

(-Z-direction only) and the growing was constrained to within the 4.175 km/sec velocity class that was related to R3 of the rift (Figure 7.4 (I), (J), and (C)).

The fourth step focused on the surrounding horizontal layers (L1-L3). L1 was classified based on its velocity of 4.175 km/sec and its depth of ≤ 125 km. This classification included segments that belonged to the boundaries between the craton and the rift (RC1-2) (Figure 7.3). This mis-classification was corrected using the *pixel-based object resizing* algorithm with the coating option in the X direction. This was done by allowing the coating on the neighborhood class of 4.275 km/sec, which was classified as the original boundary class, as shown in Figure 7.4 (K), (L), (M), and (D). L2 and L3 were classified in a similar way, using the *pixel-based object resizing* algorithm with the coating option, but in the -Z direction. The segments below L1 were classified into L2 (Figure 7.4 (N), and (E)), and the pixels below L2 were classified into L3 (Figure 7.4 (O), (P), and (F)).

In the last step the boundaries were classified. The plume boundaries (PB1-PB3) have velocity values of 4.175, 4.275, and 4.375 km/sec, respectively. PB1 and PB2 with 4.175 and 4.275 km/sec velocity, respectively, were classified using the *pixel-based object resizing* algorithm with coating only in the +Z direction (Figure 7.4 Q). This way pixels coating the plume, and with velocity values equal to 4.175 and 4.275 km/sec, were classified into the plume boundaries class (Figure 7.4 (R)). The remaining part of the plume boundary of 4.375 km/sec was mixed with the rift boundary. To separate this remaining part of the plume boundaries first the other rift boundaries needed to be classified. The largest part of the rift boundaries (velocity values of 4.025, 4.075, 4.125, 4.175, 4.225, 4.275, and 4.325 km/sec) was classified using the *assign class* algorithm. Also, the segments of the 4.375 km/sec class with a shallow location (< 190 km) were classified into rift boundaries (Figure 7.4 (R)). The remaining parts of the plume boundaries in the mixed boundaries zone were classified using the *pixel-based object resizing* algorithm with a growing option in the +Z direction, and limiting the growing process to 2 pixels only to prevent including upper pixels that belonged to the rift boundaries (Figure 7.4 (R)). Everything remaining afterwards with a velocity of 4.375 km/sec was assigned to the rift boundaries (Figure 7.4 (S)). The complete 3D rift boundaries and plume boundary can be seen in Figure 7.4 (G) and (H), respectively.

Figure 7.4 3D OOA of the synthetic model. Figures (A:H) show the 3D OOA extraction process with the plume P (A), craton C (B), rifts R (C), layers L1, L2, and L3 (D, E, and F, respectively, and displayed with 50% transparency), the rift boundary RB (G), and the plume boundaries PB (H). (I) and (J) (section 3 in Figure 7.2) show the vertical growing process of the rift into the 4.175 km/sec class. The white circles show the rift and plume boundaries that share the same velocity value and depth range. The white arrows indicate the growing process in -Z direction. (K), (L), and (M) (depth slice 60 km) show the process of correcting the mis-classification in the rift boundaries during the classification of layer 1. (L) highlights the mis-classified segment and the white arrows indicate the coating process in the X direction. (M) shows the corrected segment indicated by the white arrow and surrounded by white line. (N), (O), and (P) (section 3) illustrate the classification of layers 2 and 3. Figure (N) shows layer 1 and the white arrow indicates the coating process in -Z direction. The white arrows in (O) indicate the coating process of layer 2 in -Z direction to define layer 3. (N) layer 3 after the coating process. (Q), (R), and (S) (section 3) explain the process of classifying the rift and plume boundaries. Figure (Q) shows a white arrow that indicates the coating process of the plume in +Z direction in order to define the plume boundaries. (R) plume and rift boundaries except 4.375 km/sec class in blue color, while the white arrow indicates the 2 step growing process of the plume boundaries in order to define the plume boundaries in the 4.375 km/sec class. (S) the final classification result.

The final model mimics the input model perfectly, using a semi-automatic approach in which we only needed number of thresholds to be fed in the predefined segmentation/classification algorithms of eCognition. However, the interpretation was accomplished in an approach similar to cognitive visual image analysis. The 4 main classes were accurately resolved. The gradual boundaries were the most challenging part in the classification process, due to their overlap in velocity and depth with the 4 main features. In reality boundaries are fuzzy. The information on them, and possibly also on the main features, might not be known *a priori*. However, this additional challenge can be satisfactorily solved as shown for the 3D seismological model in the next section.

7.4.2 3D OOA of real seismological model

The approach for the 3D seismological model is similar to the one followed for the synthetic model. Due to different boundary conditions and the available *a priori* information, however, the steps were done in a different order and with different thresholds. Thresholds were again based on the analysis of the 3D histogram of the seismological model supported by visual interpretation and the *a priori* knowledge from the interpretation of the model of Adams et al. (2012) (Figure 7.5). The main features and their boundaries were defined and used to set the corresponding thresholds.

A low velocity zone was defined on the histogram by its velocity (<4.292 km/sec) and location (depth $>\sim 150$ km). The zone was split into an outer (LV) (4.18-4.29 km/sec) and inner (ILV) (<4.18 km/sec) part based on the presence of two separate low velocity clusters around 4.25 km/sec and 4.13 km/sec, respectively, as shown by the white circles in Figure 7.5. The rifts showed a characteristic low velocity in the shallow zone (Figure 7.5). They were defined by a depth and velocity ranging from 4.18-4.47 km/sec at 50 km depth, to 4.28-4.32 km/sec at 140 km depth. This low velocity zone was differentiated from other objects in the shallow part of the model by its diffuse character, a zone marked by low counts (frequency) in a broad range of depths and velocities. The craton and high velocity zone in the shallow part (SHV) have a high velocity (4.47-4.74 km/sec) and a depth of $>\sim 190$ km. The zone surrounding these objects was defined as a fuzzy high velocity boundary (BSHV) and a transition between the high velocity objects (Figure 7.5). A similar fuzzy boundary was identified between the deep low velocity zone and the rifts, and is indicated in Figure 7.5 as a boundary shallow low velocity (BSLV). This boundary was too fuzzy to set a specific velocity range, thus the threshold to define these two boundaries zones was determined by taking the intermediate velocity between the high and low velocity objects (4.37 km/sec). Another deep high velocity zone (DHV) exists below the low velocity zone (LV and ILV). It has the same velocity characteristics in the 3D histogram (velocity >4.4 km/sec) as the shallow high velocity objects, but is located at a depth $>\sim 340$ km. The boundaries between

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Figure 7.5 3D histogram explaining the process of the object extraction based on the objects' velocity values and their depth. Rifts (R), craton (C), shallow high velocity (SHV), boundary shallow high velocity (BSHV), boundary shallow low velocity (BSLV), low velocity zone (LV), inner low velocity zone (ILV), boundary deep low velocity (BDLV), and deep high velocity (DHV). The white circles indicate the two anomaly clusters of the low velocity zone (LV), and (ILV) and the white curved arrows indicate the decay of these anomaly clusters toward the boundary deep low velocity zone (BDLV).

this object and the deep low velocity object (BDLV) were determined based on the decay of the two low anomaly clusters (Figure 7.5). The final threshold was 4.29 km/sec.

The final ruleset consists here of 6 steps. Due to the increased fuzziness different steps were taken to define the objects compared to the synthetic example. The maximum velocity in the model was 4.74 km/sec, and the minimum 4.05 km/sec. Adams et al. (2012) used 10 classes (4.15-4.65 km/sec with 0.05 km/sec interval) to analyze the model.

In the first step, the image stack was classified into 26 velocity classes (4.05-4.75 km/sec with ~ 0.03 km/sec interval), to increase the smoothness of objects only without adding any artifacts to the original data.

Step 2 classified the deep high velocity object (DHV) and its boundary (BDLV), using the *assign class* algorithm. The thresholds for the deep high velocity zone were a velocity of >4.37 km/sec and a depth of >340 km. The boundary was classified using a velocity range of 4.29-4.37 km/sec and a depth >300 km (Figure 7.6 (A) and (B)).

The low velocity zone was classified in step 3 based on its low velocity (4.05-4.29 km/sec) and deep location (depth >150 km). The inner part (ILV) was first defined based on a velocity of < 4.18 km/sec (Figure 7.6 (C)). The outer part (LV) was composed of the remainder with a velocity of 4.18-4.29 km/sec (Figure 7.6 (D)).

In the fourth step, the rift (R) was classified based on its low velocity and shallow depth. Both depth and velocity were defined in ranges to cover the fuzzy characteristics of this object (4.18-4.47 km/sec at 50 km depth, to 4.28-4.32 km/sec at 140 km depth). A sequence of *assign class* algorithm steps was applied to define the velocity range at each depth (Figure 7.6 F).

Step 5 focused on classifying the craton (C) and the high velocity zones (SHV) in the shallow part of the model (50-200 km and 4.47-4.74 km/sec) into two objects. The first is the craton located within the rift branches. It was defined using a small part of the depth range (50-80 km) with the highest velocity in the model that was characteristic for the craton only (4.67-4.74 km/sec). It was then allowed to grow in the X, Y and Z direction until it filled the craton's central main part. The remaining high velocities were assigned to the high velocity zone surrounding the rift. The extracted craton and the shallow high velocity zone can be seen in Figure 7.6 (H).

The final step classified the high (BSHV) and low velocity (BSLV) boundaries using the *assign class* algorithm based on their velocity ranges 4.29-4.37 km/sec for the low velocity boundary (Figure 7.6 (E)), and 4.37-4.47 km/sec for the high velocity boundary (Figure 7.6 (G)).

The resulting model (Figure 7.6 (H)) showed the possibility to resolve objects that are known to be present in the region based on *a priori* knowledge, which were as much as possible translated into objective criteria. The uncertainty of the extracted objects is related to the uncertainty of the shear wave seismic tomography model, which ranged from 25 to 50 km for the depths between 50-250 km and 75 to 100 km for deeper depths. Difficulties in the classification process were mainly confined to boundary zones, similar to what was observed for the synthetic model. The real conditions were more challenging to resolve than in the synthetic case, since the boundaries did not have sharp and clear characteristics in the 3D histogram and were visually not easily identifiable. Therefore, the boundaries of the objects were defined in separate classes (Figure 7.6 (B), (E), and (G)). These separate classes allowed a degree of freedom to describe the fuzzy characteristic of the objects and helped in identifying the body of the main objects.

An issue to be aware of for future use was that eCognition showed the model as a mirror image around the Y (North-South) axis of the model. However, this was only a visualization issue since the exported sections

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Figure 7.6 3D OOA of the real model. (A:H) show the 3D OOA extraction process where (A) illustrates the deep high velocity object (DHV), (B) the boundary deep low velocity (BDLV), (C) the inner low velocity zone (ILV), (D) the outer part of the low velocity zone (LV), (E) the boundaries of the shallow low velocity (BSLV), (F) the rifts (R), (G) the boundaries of the high velocity objects (BSHV), and (H) the craton (C) and the shallow high velocity objects (SHV).

and depth slices had a correct orientation and geo-referenced position.

7.5 Discussion and Conclusions

3D OOA demonstrated the possibility to extract meaningful objects from 3D geophysical data in a semi-automatic way, in which the user does not need to interpret objects manually. The approach allows for a semi-objective setting of thresholds that can be tested and, if necessary, changed in a very fast and efficient way. These changes require only changing the thresholds used in the ruleset or by defining additional algorithms to be included in the ruleset for the extraction process.

The results of the synthetic model analysis showed the capability of 3D OOA to retrieve 3D objects from geophysical data. The extraction process of the main objects (craton, rift, plume, and the three horizontal layers (Figure 7.4)) required a classification based on the velocity and depth values. However, the combination of dividing the objects into zones and their gradual boundaries added complexity to the objects configurations and to the object retrieving process, since there are different parts in different objects that have the same velocity values. This made an object extraction process for these objects based on the velocity values and depths alone impossible. Moreover, it showed the need for the *pixel-based object re-sizing* algorithm, and specifically the growing and coating options, in order to define the objects. With this approach it was possible to define objects in conditions where the object boundaries were not well-defined and fuzzy, such as the mixed zone of the rift and plume boundaries. The synthetic model analysis was driven by the *a priori* knowledge on the objects present in the model. Also the object characteristics and thresholds were well defined in the 3D histogram. As expected, the 3D histogram of the real data was substantially more complex. Furthermore, for the real scenarios no knowledge of the actual object densities and depths may be available.

The real model results showed how 3D OOA can deal with realistic 3D subsurface conditions in which the boundaries become fuzzy, the object extensions become unclear, and the model characteristics vary with depth due to the different physical conditions going from shallow to large depths in the subsurface. The 3D OOA-derived objects, the rifts, the craton, the low velocity zone and the deep high velocity zone (Figure 7.6), were extracted based on their velocity and their depth location. However, most of the objects boundaries have strong fuzzy characteristics and it was not straightforward to define objective measures for accurate definition of these boundaries. This was solved by using separate classes to define the objects boundaries which give more confidence to the major classified objects. However, the process to define fuzzy boundaries in a seismological model needs further research to reduce non-uniqueness in the definition of these boundaries. The smooth transition between

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objects and their surroundings is often too fuzzy to be mapped with confidence. However, eCognition is fundamentally also based on the concept of fuzzy classification. This means that segments marked by value transitions, and that may thus belong to more than one class, can also be classified as such. Moreover, a *multiresolution segmentation* algorithm (Baatz et al. (2005); Baatz (2000)) can be used instead of the *multi-threshold segmentation* algorithm since it can deal with complex structures which have different scales. However, the selection of the optimum scale parameter is a challenging task, even when using the Estimate Scale Parameter (ESP) tool (Drăguț et al., 2010) which requires further extension to allow the processing of 3D data. The same is true for another approach for objective segmentation parameter identification, the Plateau Objective Function by (Martha et al., 2011).

The main geophysical application for 3D OOA is any 3D inversion and/or interpretation process. With the boundaries defined, both spatially as well as in depth, an inversion process can be more focused on retrieving actual properties and reducing the degrees of freedom for changing the shape of objects. This might be particularly useful for the inversion of potential field data where lateral variations are often reasonably well modeled but the depth resolution is very poor, thereby leaving a large degree of uncertainty on the final model in both object boundaries and property characteristics. A case study implementation for reducing non-uniqueness in satellite gravity inversions is shown by Fadel et al. (2015). This also might be particularly relevant for recent satellite gravity studies which use seismic tomography models (for a review see van der Meijde et al. (2013)). A clear example where this might be of added value is the crustal thickness modeling in Africa (Tugume et al., 2013) and South America (Van der Meijde et al., 2013). In these studies uniform densities were assumed throughout the continent, due to lack of objective and spatially homogeneous information. Spatially defined boundaries, based on density variations, can improve the modeling with spatially dependent density information. With 3D OOA the uncertainty on objects and physical properties can be potentially significantly reduced.

The eCognition rulesets developed in this research will be made available on our website www.itc.nl/ooa-group.

7.6 Acknowledgments

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3D OBIA in 3D geophysical modelling: East African Rift System case study

Abstract

Non-uniqueness of satellite gravity interpretation has traditionally been reduced by using *a priori* information from seismic tomography models. This reduction in the non-uniqueness has been based on velocity-density conversion formulas or user interpretation of the 3D subsurface structures (objects) based on the seismic tomography models and then forward modelling these objects. However, this form of object-based approach has been done without a standardized methodology on how to extract the subsurface structures from the 3D models. In this research, a 3D object-oriented image analysis (3D OOA) approach was implemented to extract the 3D subsurface structures from geophysical data. The approach was applied on a 3D shear wave seismic tomography model of the central part of the East African Rift System. Subsequently, the extracted 3D objects from the tomography model were reconstructed in the 3D interactive modelling environment IGMAS+, and their density contrast values were calculated using an object-based inversion technique to calculate the forward signal of the objects and compare it with the measured satellite gravity. Thus, a new object-based approach was implemented to interpret and extract the 3D subsurface objects from 3D geophysical data. We also introduce a new approach to constrain the interpretation of the satellite gravity measurements that can be applied using any 3D geophysical model.

This chapter is based on:

I. Fadel, M. van der Meijde, N. Kerle, and N. Lauritsen. 3D object-oriented image analysis in 3D geophysical modelling: Analysing the central part of the East African Rift System. *International Journal of Applied Earth Observation and Geoinformation*, 35, Part A:44-53, 2015.

8.1 Introduction

Inverting gravity data has traditionally been one of the main geophysical challenges. Because of its monopole potential field, and resulting measured bulk property, retrieving accurate object information is often a matter of trade-offs and best guesses. Using *a priori* geological and/or seismological data is a way of reducing this non-uniqueness of gravity inversions.

There are two main generic approaches have been followed in the past. The first is the conversion of the seismic velocities into densities using empirical standard velocity-density relationships (Birch, 1961; Gardner et al., 1974; Christensen and Mooney, 1995; Godfrey et al., 1997; Brocher, 2005) or including temperature, pressure, and mineral composition effects (Ravat et al., 1999; Korenaga et al., 2001; Cammarano et al., 2011). Calculated densities were then used to produce the estimated gravity signal and compare it with the measured gravity, or to constrain any further inversion process (Bezada and Zelt, 2011). This approach usually has to deal with uncertainty in the seismic data and the velocity-density conversion relationships that result from the dependency on and uncertainty in physical parameters such as temperature, pressure and petrology (Bezada and Zelt, 2011; Cammarano et al., 2011). The second approach is based on the interpretation of the seismic models to derive objects (Götze and Lahmeyer, 1988) such as layers, plate boundaries, subduction zones, domes etc. Subsequently the densities of these objects are calculated using one of the velocity-density relationships or through a trial and error approach. After that, the forward signal associated with these objects is compared with the measured gravity (e.g., Sanchez et al., 2011). A following step can be applying inversion approaches to reduce the misfit between the calculated and the measured gravity (e.g., Ebbing et al., 2001). This gravity modelling object-based approach has been done without a standardized methodology on how to extract the subsurface structures from 3D subsurface data, and has traditionally mainly been based on the user experience or pre-existing data (e.g., Ebbing et al., 2001; Sanchez et al., 2011).

In this research we report on a case study from the central part of the East African Rift System and how an advanced 3D object-oriented Image Analysis (3D OOA) approach can provide more objective definitions of object boundaries for gravity modelling and inversion. 3D OOA will be used to extract 3D objects from a tomography voxel model. Then, they will be reconstructed in the 3D object-based interactive gravity modelling environment IGMAS+. Subsequently their density contrast values will be calculated using an object-based inversion. Finally, the gravity response of the final model will be compared with the satellite gravity signal.

8.2 Dataset

8.2.1 3D Tomography Model

The study area is the central part of the East African Rift System. Adams et al. (2012) developed a 3D shear wave seismic tomography model of the area, which is voxel-based and can be divided into two main sections. The first is the crustal part of the model, which extends from 0 to 40 km with 1 km increments. It represents the crustal structure of the model down to a flat Moho at 40 km depth. The second is the upper mantle part of the model ranges from a depth of 40 km down to 400 km at 10 km increments. This part of the model can be divided into two main zones, shallow and deep, based on their relative position in the model. The shallow zone of the upper mantle part (from 40 km down to ~200 km) is characterized by high velocities of the Tanzanian Craton and Ugandan basement complex, and low velocities of the surrounding tectonic settings such as rift branches, the Proterozoic mobile belts and Cenozoic volcanism. The deep zone of the upper mantle part (from ~200 km down to 400 km) is characterized by dominant low velocities related to a head plume or part of the deep mantle super plume in western Africa (Adams et al., 2012).

8.2.2 Satellite Gravity Data

The gravity data used in this study are GOCE only models (see van der Meijde et al. (2013) for details) and the high resolution EIGEN-6C (Christopher et al., 2011), a combined gravity field model of LAGEOS/GRACE, GOCE and DTU10 up to a maximum degree and order of 1420. The free air gravity data were obtained from the International Centre for Global Earth Models (ICGEM) web site (<http://icgem.gfz-potsdam.de/ICGEM/>) with a resolution of 0.1 degree (Figure 8.1A). The effects of the nearby terrain were removed using ETOPO1 topography data (Figure 8.1B), also downloaded through the ICGEM website with similar resolution (0.1 degree) to the models. The effect of the terrain correction on the satellite gravity measurements was small with a maximum value of $3.16 \times 10^{-5} \text{ m/s}^2$ (Figure 8.1C), in line with findings by Mishra et al. (2012). Then, the data were Bouguer corrected, removing the effect due to the mass excess or deficiency as a result of high or low elevation relative to the mean sea level. The Bouguer correction also used ETOPO1 topography data with the same spatial resolution (0.1 degree) as the satellite gravity data and a reference density of 2.67 t/m^3 (Figure 8.1D).

8.3 3D OOA

Object oriented image analysis (OOA), also defined as object based image analysis (OBIA) or geographic object based image analysis (GEOBIA), is an image analysis technique that is based on the analysis of segments,

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Figure 8.1 The correction of the satellite gravity signal. (A) The free air EIGEN-6c data. (B) ETOPO1 topography map. (C) The terrain correction map using ETOPO1 topography and reference density 2.67 t/m^3 . (D) The Bouguer anomaly map after the Bouguer and terrain corrections.

or objects, instead of pixels. In general, OOA is a 2-step procedure starting with image segmentation that is followed by the classification of these segments. The segmentation step comprises of dividing image pixels into contiguous segments or objects, based on pre-defined knowledge based or homogeneity criteria, such as color or intensity. The subsequent classification step uses the segments associated attributes which can be spectral, geometric, textural, contextual, and also geological and physical, to differentiate between the different objects present in the images. Consequently, OOA mimics cognitive visual image analysis techniques through a form of knowledge-driven analysis (Martha et al., 2011). Recently, 3D OOA was developed in the biomedical field to extract 3D objects from image stacks (Schönmeyer et al., 2006). The first application to 3D synthetic and seismological models was recently carried out (Fadel et al., 2014). That study specifically applied the approach in

eCognition software for deriving objects in a 3D seismic tomography model.

eCognition was the first OOA tool when introduced in 2000 and since then about 50-55% of all OOA published scientific studies have used it (Blaschke, 2010). Then the name was changed into Definiens in 2007. After that the package was split into two software packages; Definiens software that is related to biomedical applications, and eCognition that is mainly applied for geospatial applications. eCognition is built on the concept of rulesets that define the object extraction process in which a complex processing scheme of segmentation/classification algorithms can be implemented (Fadel et al., 2014). The ruleset is a semiautomatic since it is built on segmentation and classification algorithms, which are originally developed in the eCognition package, that are used to segment or to classify the 3D objects from the image stack. The user only needs to define the thresholds that are suitable to define the target segmentation scheme or the target object to be extracted.

3D OOA was applied on the 3D tomographic model in 2 stages. First, the ruleset of Fadel et al. (2014) was used to extract the objects for the upper mantle part of the model between >40-400 km (Figure 8.2 Upper Mantle Part). The upper mantle objects can be separated in two zones based on their relative depth in the upper mantle part of the model:

- 1- The shallow zone (>40-~200 km) consists of high velocity objects (craton (C) and shallow high velocity object (SHV)), low velocity object (rifts (R)), and boundary zones surrounding both high and low velocity objects (boundary shallow high velocity (BSHV) and a boundary shallow low velocity (BSLV), respectively).
- 2- The deep zone (~200-400 km) is dominated by a low velocity object (LV) with its inner part (ILV) followed by a high velocity object at the deepest part of the model (deep high velocity (DHV)) with a boundary between it and the upper low velocity object (boundary deep low velocity (BDLV)).

Second, a new ruleset was developed for the crustal part of the model from the surface down to a depth of 40 km. The new crustal part of the model (Aubrey Adams, personal communication, April 29, 2013) was voxel based with 1 km depth interval.

The new ruleset for this part was based on visual interpretation and analysis of the 3D histogram, summarizing the 3D distribution of density values with depth (Figure 8.3 B), similar to Fadel et al. (2014). The tomographic model had a hard boundary of 2 discrete 1 km layers above and below the Moho (Adams et al., 2012), with a major velocity jump over these two layers. The ruleset made use of this hard boundary to distinguish the crustal part from the upper mantle part. The new ruleset consisted of 2 main steps (see Fadel et al. (2015) supplementary data for more details).

Figure 8.2 3D OOA results of the model. The crustal part (0-40 km) shows the OOA results of the five units (U1-5). The upper mantle part (>40-400) shows the OOA results of the different upper mantle objects: craton (C), shallow high velocity (SHV), rift (R), boundary shallow high velocity (BSHV), boundary shallow low velocity (BSLV), low velocity (LV), inner low velocity (ILV), boundary deep low velocity (BDLV), and deep high velocity (DHV). The reader should be aware that the objects have a reverse orientation in the East-West direction (compare them with Figure 8.4) which is a visualization defect of 3D OOA results (adapted from Fadel et al. (2014)).

In the first step *multi-threshold segmentation* was used to separate the crustal from the upper mantle part of the model. This algorithm splits the image into segments and classifies them based on pre-defined thresholds. 40 km depth was used as a threshold to separate the crustal from the upper mantle part (Figure 8.3A). The subsequent focus was on the crustal part of the model (0-40 km). A series of *assign class* algorithms was applied to split it into five objects using thresholds that were defined based on the visual interpretation supported by the analysis of the 3D histogram (Figure 8.3B). These 5 units (Figure 8.2 Crustal Part) do not necessarily have a geo/physical meaning in terms of their velocities, due to the sharp boundary with large velocity contrast at the fixed layer at 40 km depth in the tomography model. However, they highlight the relative velocity changes in this zone. Unit 1 was characterized by very low velocity values up to 3.2 km/sec, and was surrounded by unit 2 which represents the boundary for unit 1 and has a velocity range from 3.2 km/sec up to 3.4 km/sec. Unit 3, with a velocity range of 3.4-3.55 km/sec, was characterized by low variation in the density distribution as can be seen in the 3D histogram. Unit 4, with a velocity range of 3.55-4.1 km/sec, surrounded the final unit 5 that was characterized by velocities higher than 4.1 km/sec thereby representing the highest velocity in the crustal part. 3D OOA behaved as any simple isosurface due to the lack of *a priori* information about the crustal part; in addition to the limited ability of seismic tomography techniques in such shallow depths where other approaches are more preferable such as receiver functions (Liu and Gu, 2012).

8.4 Object-Based Modelling

8.4.1 Objects Reconstruction

The objects extracted in the OOA part were re-constructed in IGMAS+ (3D Interactive Gravity and Magnetic Application System; Götze and Lahmeyer, 1988; Schmidt et al., 2010), which facilitates interactive forward modelling of the subsurface in an object-oriented environment. The results from OOA were exported as classified cross sections that were used to constrain the shape of the objects through the reconstruction process. 27 sections were created in IGMAS+ with 50 km average distance between them to fit the original 3D tomography model that had a horizontal resolution of 0.5 degree. Then, OOA classified sections were imported in IGMAS+ and co-located with the created IGMAS+ sections to start re-constructing the objects. The upper mantle part of the model (>40-400 km depth) was constructed first (see Figure 8.4 Upper Mantle Part and Fadel et al. (2014) for details). Adams et al. (2012) used a 1 km thick layer above and below the 40 km depth to constrain Moho depths and allow for large variations in the velocity between the crustal and the upper mantle part of the model. Hence, a 1 km horizontal layer above and below a depth of 40 km (2 km thick layer between depths 39-41 km)

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Figure 8.3 3D histograms of the model. (A) The histogram is the full 3D histogram of the model from 0-400 km depth. The black arrow at 40 km depth indicates the sharp discontinuity in the model because of the Moho. (B) The histogram shows the crustal part (0-40 km) of the model, while the black-white lines indicate the threshold velocities for the different units.

Figure 8.4 The reconstruction process of the extracted objects in IGMAS+. (FM) The full model image shows the imported sections resulted from OOA and the measured gravity on the top of them. The other images show the reconstruction process of each object: the five units of the crustal part (U1-5), craton (C), shallow high velocity (SHV), rift (R), boundary shallow high velocity (BSHV), boundary shallow low velocity (BSLV), low velocity (LV), inner low velocity (ILV), boundary deep low velocity (BDLV), and deep high velocity (DHV).

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was constructed to represent the Moho as in the original model. This is in line with actual observations on the Moho which often indicate it is a layer with a certain thickness and not a discrete boundary. Subsequently, the crustal part of the model was constructed above (Figure 8.4 Crustal Part).

The tomographic model did not have a realistic Moho topography. Therefore, another model was constructed in which the crustal and the upper mantle parts remained the same; however, the fixed Moho was replaced with a gravity based crustal thickness map of Africa (Tugume et al., 2013). The 2 km thick flat layer representing top and bottom of the Moho transition zone was modified to describe the undulation of the new Moho. Now a realistic Moho topography was included, representing spatial density variations due to Moho undulations. The embedded Moho model showed a large variation in crustal thicknesses that ranged from ~28km at coastal regions to ~42 km in the central part of the study area (Figure 8.5). The alternative option to use crustal thickness from

Figure 8.5 The crustal thickness map of the study area adapted from Tugume et al. (2013).

CRUST2.0, a crustal thickness map that was created by interpolating the crustal velocity estimates from active-source profiling techniques and geologically estimated terrain ages (Bassin, 2000), was not considered because of the debate of using CRUST2.0 in data poor environments (see discussion in Van der Meijde et al., 2013).

8.4.2 Object-Based Inversion

After the definition of the two scenario models based on the results of the 3D OOA, density values for each object were needed to calculate the gravity signal and compare it with the measured satellite signal. Schmidt et al. (2011) introduced a linear object-based inversion approach in IGMAS+, based on the minimum mean square error (MMSE) algorithm introduced by Haas (2008) based on the work of Sæther (1997).

The interactive object-based modelling approach aims to minimize the misfit between the *calculated* and the *measured* gravity signal up to approximation *error*:

$$\text{calculated} = \text{measured} - \text{error} = \sum_{i=1}^n \rho_i P_i \quad (8.1)$$

Where n is the total number of bodies, ρ_i is the constant density of body i , and P_i represents the polyhedron gravity effect of body i (for density 1).

Equation 8.1 simplifies the subsurface structure into a small number of objects. Each object has a constant density value (ρ_i) that is the common value of all voxels within this object i . The calculated gravity effect of object i will be the product of its density ρ_i multiplied by the polyhedron gravity effect in case of a unit density value (e.g., 1 t/m^3 or 1 kg/m^3 or any other unit). The total *calculated* gravity signal will be the summation of the resulting polyhedron effects for the total number of objects n . The object-based inversion approach aims to find the optimal density values ρ_i using MMSE estimation instead of the traditional interactive trial and error approaches. The estimated optimized density values ρ_i , directly related to the defined objects produce a *calculated* gravity signal with a minimum misfit (*error*) to the *measured* signal. It is also possible to add constraints to the inversion process in case of the presence of *a priori* knowledge about the density values of some objects in the model. In that case, the objects with *a priori* known density values are provided in the inversion algorithm as constants while the inversion algorithm is only applied to the objects with unknown density values, as shown in equations 8.2 and 8.3.

$$\text{calculated} = \sum_{ip} \rho_{ip} P_{ip} + \sum_{np} \rho_{np} P_{np} \quad (8.2)$$

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Hence by substitution using equation 8.1,

$$measured - error - \sum_{np} \rho_{np} P_{np} = \sum_{ip} \rho_{ip} P_{ip} \quad (8.3)$$

Where index ip indicates the objects that will be inverted, and np indicates the objects that will not.

A point of attention is the setting of the background density value. The Bouguer anomaly is the result of the density contrast between the background densities of the Earth and the existing densities of the different objects in the subsurface. IGMAS+ allows one density background value to calculate density anomalies and, hence, gravity anomalies relative to it. However, for models with large vertical and spatial density contrasts down to 400 km the background density values changes with depth. Therefore, the use of one value as a density background is not ideal. In order to deal with this limitation, relative object density contrasts were used with respect to a background density of 0. The results from the inversion process are thus density contrast values between objects in the model and the background densities at their equivalent depths. For example, if we assumed that the density background for 0-10 km is 2.67 t/m³, the absolute density value of unit 1 is 2.453 t/m³ (U1 lies in a depth range 0-10 km with a density contrast of -0.217, as shown in Table 8.2).

The measured gravity signal was filtered to remove short wavelengths in order to fit the resolution of the input data (i.e., the seismological model). The smallest object in the model is approximately 100 km. This required the application of a low-pass cosine roll-off filter on the measured gravity signal (Figure 8.6(D)). This removed the high frequency content, needed to be comparable with the calculated gravity signal based on the seismological model. The filtered gravity signal was used to estimate the optimized density contrast values of the different objects, and to evaluate the degree of correlation with the calculated gravity signal.

Parameters	Flat Moho Model			
	Upper mantle part	Upper mantle part with Moho	Complete model	Complete model without Moho
Correlation	0.69	0.71	0.72	0.70
Standard Deviation * 10 ⁻⁵ m/s ⁻²	25	24	14	25
	Undulated Moho Model			
	Upper mantle part	Upper Mantle part with Moho	Complete model	Complete model without Moho
Correlation	0.69	0.71	0.95	0.89
Standard Deviation * 10 ⁻⁵ m/s ⁻²	25	24	11	16

Table 8.1 The inversion results for the different parts of the two models.

Table 8.1 shows the results of the inversion process for the different parts of the two models, with both the flat and the undulated Moho, compared to EIGEN-6c. In both cases, the calculated gravity signal from the

upper mantle part showed 69% correlation with the measured satellite gravity. In case of the flat Moho model, the added Moho discontinuity layer and the crustal part only had a slight effect on improving the results (maximum of 3% improvement). In case of the complete model, however, using the realistic Moho surface based on the crustal thickness map of Africa (Tugume et al., 2013), the results showed a major improvement up to a 95% fit. The undulating Moho discontinuity clearly has a strong influence on the final fit, which is to be expected for a large-scale strong anomaly that also has a significant contribution to the gravity signal among all modeled objects. Another reason for the increased fit is that the applied Moho topography is also derived from gravity data. Therefore, it is expected to give a high correlation and consequently contribute strongly. This strong contribution is reflected in the, unrealistically, high relative density contrast value 5.157 t/m^3 of the Moho layer that results from the inversion, as shown in Table 8.2. This is also partly because the Moho is modeled as a thin layer (2 km thick) and that all the uncertainties in the strong negative crustal density values (U1 -0.217 t/m^3 , U2 -0.244 t/m^3 , U3 -0.278 t/m^3 , U4 -0.312 t/m^3 , and U5 -0.373 t/m^3 Table 8.2) seem to accumulate in the high positive density contrast value of the Moho.

Object	Eigen-6c (t/m^3)	Eigen-6c2 (t/m^3)	GOCO03S (t/m^3)	GO-CONS-GCF-2-DIR-R4 (t/m^3)	GO-CONS-GCF-2-TIM-R4 (t/m^3)
Unit 1 (U1)	-0.217	-0.215	-0.211	-0.211	-0.211
Unit 2 (U2)	-0.244	-0.243	-0.24	-0.241	-0.241
Unit 3 (U3)	-0.278	-0.277	-0.275	-0.275	-0.275
Unit 4 (U4)	-0.312	-0.312	-0.308	-0.309	-0.308
Unit 5 (U5)	-0.373	-0.374	-0.371	-0.372	-0.371
Craton (C)	0.008	0.008	0.007	0.007	0.007
Shallow High Velocity (SHV)	0.011	0.011	0.01	0.01	0.009
Boundary Shallow High Velocity (BSHV)	0.005	0.005	0.004	0.004	0.004
Rift (R)	-0.002	-0.002	-0.003	-0.003	-0.003
Boundary Shallow Low Velocity (BSLV)	-0.015	-0.015	-0.014	-0.014	-0.014
Low Velocity Zone (LV)	-0.052	-0.053	-0.053	-0.053	-0.053
Inner Low Velocity (ILV)	-0.002	-0.002	-0.003	-0.003	-0.003
Boundary Deep Low Velocity (BDLV)	-0.067	-0.067	-0.062	-0.06	-0.058
Deep High Velocity (DHV)	0.172	0.173	0.171	0.172	0.17
MOHO	5.157	5.154	5.132	5.137	5.139
Reference	0	0	0	0	0
Correlation	0.948	0.949	0.948	0.948	0.947
Standard deviation	11.021	11	11.002	11.062	11.089

Table 8.2 The density contrast (t/m^3) estimated from the inversion of the complete model with the undulated Moho using different gravity models.

Table 8.2 (second column) shows the density contrast values for each individual object in the model resulting from the object-based inversion using EIGEN-6c. The values for the crustal part (U1-U5) of the model, rifts (R), the low velocity zone (LV), and its inner part (ILV) showed negative density contrast. For the rifts (R), the low velocity zone (LV), and the inner low velocity zone (ILV), the values are matching with the low velocity values from the tomography model. However, the inner low velocity (ILV) object showed less negative density contrast than the low velocity (LV) object, which contradicts the seismic tomography model, since the inner low velocity (ILV) had the lowest velocity. The craton (C), the shallow high velocity (SHV), and the deep high velocity zone (DHV) objects, showed positive density contrast values that agreed with the high velocity in the

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tomographic model. However, the craton (C) had a higher velocity in the tomographic model than the shallow high velocity (SHV) object. This was contrary to the density contrast, which was lower for the craton. This contradiction can be explained by the large uncertainty in the velocity values that is less than 0.2 km/s at depths <250 km, and not more than 0.3 km/s at greater depths (Adams et al., 2012). The boundaries between Shallow High Velocity (BSHV), Shallow Low Velocity (BSLV), and Deep Low Velocity (BDLV) objects showed realistic transition values. The final maps of the calculated gravity signal for the upper mantle part and the complete model with the undulated Moho are shown in Figure 8.6.

8.4.3 Testing Different Gravity Models

The previous modelling was done using a combined EIGEN-6c model constructed from satellite data and other gravity data sources. This raised the question of how much the different content, resolution and quality of different gravity models influence the final model parameters. For this purpose, different combined and satellite data only models were compared. EIGEN-6c, EIGEN-6c2 (GOCE, GRACE, LAGEOS and surface data), GOCO03S (GOCE, GRACE, CHAMP, SLR), GO-CONS-GCF-2-DIR-R4 (GOCE, GRACE, and LAGEOS), and GO-CONS-GCF-2-TIM-R4 (GOCE only) were used in this test (for more information on the GOCE related models see van der Meijde et al. (2013)). The models were downloaded with the same resolution and subjected to the same processing (terrain correction, Bouguer correction and 100 km cosine roll-off filter). The results for the different gravity models are summarized in Table 8.2.

The results showed that EIGEN-6c2, the most recent and highest resolution model, gives the best results. This was expected since the model has been shown to have a higher accuracy than any other gravity model (Förste et al., 2013). However, the differences in the models (Figure 8.7) were not significant. In addition, the correlations are all high and comparable, and standard deviation results vary little (Table 8.2). When compared to the standard deviation that is obtained for the best model (see Table 8.1) one can conclude that choices made in the creation of the model are of much greater consequence than the variability between models (at 100 km resolution).

8.5 Discussion

It is difficult to quantify how much 3D OOA has improved the modelling of subsurface objects. An expert modeler might have come to a similar result, or maybe even better. The main advantage is that the boundaries are based on physical rules and knowledge, and that they are objective and repeatable. An additional advantage was that the top and bottom of the objects were determined, something that traditionally has been diffi-

Figure 8.6 The calculated gravity signal based on the inversion process. (A) The 100 km filtered gravity signal. (B) The calculated signal from the upper mantle part of the model (>40-400 km) with 69% correlation with (A). (C) The calculated gravity signal using the full model with 95% correlation with (A). (D) The difference between the measured (A) and the calculated (C) signal. (E) The histogram of difference map (D).

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Figure 8.7 Maps show the difference between combined satellite and surface data gravity model *EIGEN-6c2* that showed the best fit in the inversion results (Table 8.2) and the different gravity models. (A) shows the difference between *EIGEN-6c2* and the combined satellite and surface data gravity model *EIGEN-6c*. (B) shows the difference between *EIGEN-6c2* and satellite only gravity model *GOCO03S*. (C) shows the difference between *EIGEN-6c2* and satellite only gravity model (*GO-CONS-GCF-2-DIR-R4*). (D) shows the difference between *EIGEN-6c2* and satellite only gravity model (*GO-CONS-GCF-2-TIM-R4*).

cult due to strong smearing and/or smoothing in the vertical direction.

In this example the boundaries of the objects were fixed, and the inversion only influenced the internal physical properties of the objects. This is, of course, a limitation and can be given more degrees of freedom in future attempts. The method we developed is ideal for incorporating *a priori* information in the identification of objects. We only used geological knowledge on the main geological and tectonic features in the area based on Adams et al. (2012), and crustal thickness from Tugume et al. (2013). More *a priori* information can potentially be incorporated, such as velocity-density relationships, point observations from, for example, receiver function analysis, and petrochemical knowledge on the physical properties of the various defined objects. This way physical properties can be further constrained. In our experiments we observed that for some object parameters had non-realistic density contrast values (e.g., the Moho layer). By adding ranges within which parameters may vary, the result is likely to be more realistic in terms of retrieved physical properties.

Because of the poorly resolved crustal part of the seismological model (Adams et al., 2012), there is no strictly defined structural integrity between the crustal and the upper mantle part of the model. The sharp Moho boundary defined in the tomographic model is not physically realistic in all places, and replacement by another crustal thickness model creates an uncertainty in the physical continuation of the physical properties, thereby adding more uncertainty on structural integrity.

The 31% misfit value between the calculated and measured gravity, based on the upper mantle part of the model, can be explained due to the absence of the crustal part in the model. The dramatic improvement in the correlation and standard deviation after adding the Moho are doubtful when considering the unrealistically high density contrast of the Moho that had a high influence on the measured signal. Therefore, in order to prove the necessity to add a realistic Moho surface, the inversion was carried out excluding the Moho layer (a density contrast=0) from the inversion using *np* index mentioned in Equations 2 and 3. The results of the inversion showed a significant improvement of 89% in the calculated signal (Table 8.1). This underscores the importance of knowledge on the shape of the Moho surface. In addition, it reflects the necessity for density contrast constraints of some objects, since a minor change in one object in the model caused significant changes in the density contrast of all objects. Table 8.2 shows that a negligible change in the measured gravity signal can cause substantial changes in the density contrast values. Another indicator for the high sensitivity of the density contrast values is that the values are not stable with changes in the extent of the measured gravity signal. Some tests were done to remove the edge effect in the calculated gravity signal by reducing the measured gravity extent by 4, 8, and 12% of the total extent of the study

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area. The results of the density contrast for each trial were also different. Moreover, the edge effect was still present in the forward signal and the misfit between the measured and the calculated gravity signal (Figure 8.6C and D) since the forward calculation of IGMAS+ is based on flat Earth assumptions and does not take in to account the spherical shape of the Earth.

The uncertainty in the location of the extracted objects are related to the uncertainty of the seismic tomographic model. Adams et al. (2012) stated that the uncertainties for major features in the model are ranging from 25-50 km for the shallow zone of the upper mantle part of the model (40-250 km) and can be up to 75-100 km for depths > 250 km. The uncertainty of the crustal part of the model (0-40 km) was not discussed in Adams et al. (2012) but was assumed to be the same as for the depth range 40-250 km.

The uncertainties are reflected in the results of the object based inversion. The correlation factor and the standard deviation of the inversion results show that for the complete model, including the undulated Moho density contrasts 5.157 t/m^3 , the correlation was 0.95 and the standard deviation of the misfit was around $11 * 10^{-5} \text{ m/s}^2$ (Figure 8.6-D and E). The misfit can be explained by the inherent uncertainty in the seismic model that also include the object boundaries. The inversion algorithm only allowed change in the density contrasts. That resulted in some objects density contrast to be unrealistic. However, if the algorithm was allowed to change the boundaries location within certain limits that are linked to the uncertainty of the input data, may be both the location of the objects' boundaries and the density contrast values will be improved. This can be done through implementing insight on a priori knowledge about the possible density contrast ranges of the objects that can prevent adding more non-uniqueness to the problem due to the available infinite combination of boundary locations and density contrast values.

8.6 Conclusion

In this research we developed a method to extract 3D objects from a seismic tomography model using 3D OOA. The extracted objects were then reconstructed in an object-based gravity modelling environment (IGMAS+), and the density contrast value for each object calculated using an object-based inversion technique. The calculated gravity signal from the upper mantle part (>40-400 km) of the model produced a 69% correlation with observed gravity. The low fit is mainly due to the absence of the Moho undulation surface, and the crustal part. By adding a flat Moho layer with a 2 km thickness and the crustal objects, based on the tomographic model, the calculated gravity signal only improved by 3%. However, by adding the Moho undulation surface based on the crustal thickness map of Africa by Tugume et al. (2013) and, the intra-crustal

objects from the tomographic model from Adams et al. (2012), the fit of the calculated gravity signal with the observed gravity significantly improved up to 95%. However, the results for the Moho and the crustal part need further verification, and further improvements are expected based on other geophysical data sources, such as receiver functions.

The usage of the different satellite gravity models, either satellite only models or satellite and terrestrial models, only had a minor effect on the inversion results. Uncertainty in the inversion is larger than the variability due to the use of different models (all at 100 km resolution).

Integrating 3D OOA in a 3D object-oriented gravity modelling environment can lead to a significant improvement in converting the voxel-based geophysical models into object-based models based on simple and objective knowledge-based classification rules. In addition, the object-based inversion approach can be improved by using the different available segmentation algorithms, which, for example, allow objects to grow and shrink automatically, and hence allow some flexibility for automatic adjustment of the boundary location, and thus to optimize fitting results.

8.7 Acknowledgments

We thank Dr. Aubreya Adams at Washington University for making the seismic tomography model of the study area available for this study. We would like to thank Prof. Dr. Hans-Jürgen Götze and his research team at Kiel University for providing the academic license for the IGMAS+ software.

Joint inversion of seismic surface wave and GOCE gravity gradients

9.1 Abstract

The GOCE satellite mission has provided the geophysical community with gravity and gravity gradient measurements with an unprecedented accuracy and global homogeneous coverage. However, the full capabilities of the gravity gradient data and their sensitivity to the geometry of the 3D subsurface structures have not been fully exploited yet. One of the main obstacles is the inherent non-uniqueness of the potential field inversion. Therefore, potential field data are usually jointly inverted with other geophysical data. With the emerging large amount of seismological data and the high-quality gravity gradients measurements provided by GOCE, new methodologies are needed to fully exploit the capabilities of both data sets. In this chapter, the gravity gradients at satellite altitude and gravity total field were used in a joint inversion scheme with seismic surface waves. A direct coupling approach between velocity and density was used, based on recent velocity-density relationships. An application using synthetic tests showed that the developed methodology provides promising results, which will need to be tested on real data.

9.2 Introduction

The cooperative inversion concept was introduced by Lines et al. (1988) and was defined as the approach in which we can estimate subsurface models that can fit different geophysical measurements. They divided the concept into two main schemes: sequential and joint inversion. The sequential inversion scheme depends on inverting multiple geophysical measurements separately. The inversion result from one method is used as an input or initial model for the next inversion step(s) of the other geophysical measurements. On the other hand, the joint inversion scheme inverts the multiple geophysical datasets simultaneously based on a coupling relation that relates the different measured physical properties empirically or structurally with each other. This coupling relation conceptually assumes that the rock properties and/or the boundaries

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between the subsurface bodies have a similar or a common response in these datasets. In other words, they have a coupling response which can be addressed empirically or physically. However, both sequential and joint schemes have their own advantages and disadvantages. In general, the main drawback of the joint inversion scheme is that the coupling assumption between the different geophysical methods is not valid for some subsurface environments, while the sequential scheme gives the freedom for the inversion results to vary between the different geophysical inversion steps. On the other hand, the joint inversion approach results in a unique solution, while the sequential inversion has an added freedom in the subsequent inversion step(s) that increases the non-uniqueness of the solution (Gallardo and Meju, 2011).

In this chapter, I will introduce a technique to jointly invert seismic surface wave and gravity observations. Seismic techniques can provide detailed images of seismic velocity variations. However, they have limited sensitivity to density. Recent studies (e.g., Mosca et al. (2012)) have shown that density is required to distinguish between compositional and temperature effects within the Earth, and therefore represents a crucial physical parameter for solid Earth geophysics. The Rayleigh surface wave is well-known to be sensitive to density, albeit only weakly. However, estimating density using surface waves suffers from strong nonlinearity (Takeuchi and Saito, 1972). Recent pioneering work by Blom et al. (2017) has shown that full waveform inversion can theoretically be used to recover the subsurface density structure regardless of the small contribution of the density to the seismic wave field. On the crustal and upper mantle scale, Xing et al. (2016) showed that an improper shallow density structure could lead to errors in shear velocity models at middle and lower crustal depths. Based on the previous findings, it is clear that there is a need for a methodology that can constrain the density structure at crustal and upper mantle depths. Recent studies found novel approaches to combine gravity with seismic data to estimate the subsurface density structure on exploration scale (Moorkamp et al., 2011) and for the crust and the uppermost part of the mantle (Maceira and Ammon, 2009). Moorkamp et al. (2011) jointly inverted seismic body wave, electromagnetic, gravity total field, and gradients. On the other hand, Maceira and Ammon (2009) jointly inverted Rayleigh wave group velocities and gravity anomalies. The gravity gradient data are more sensitive to the subsurface density structure in comparison with the total field gravity data (Zhdanov et al., 2004). Therefore, the excellent global gravity gradient observation needs a new tool to fully exploit their potential. Afonso et al. (2016) discussed that the joint inversion of GOCE gravity gradients and seismic surface wave data is a very promising approach.

The developed methodology that will be presented here is designed to jointly invert gravity and seismic surface Rayleigh wave group and phase velocity measurements. The inversion technique will focus on

using global gravity data derived from the recent GOCE satellite mission. The GOCE satellite was a unique satellite (the GOCE mission ended on 13.11.2013) that was equipped with a three-axes gradiometer and flying at an altitude of 260 km and less. GOCE provided the most detailed measurements of Earth's gravity from space ever by acquiring gravity gradients, i.e., the three-dimensional second derivatives of the gravitational potential. On the other hand, Rayleigh wave group and phase velocity are used as the seismic observations. Rayleigh wave dispersion data at intermediate periods (10-200 s) can be used to explore the crust and upper mantle structure. In the last two decades, the methodology was boosted by the ambient noise technique that enabled estimation of group and phase velocities at relatively short periods that can even be used in exploration applications.

9.3 The data sets

Before going through the details of the joint inversion technique, it is important to describe the different data sets used in it. The gravity data from the GOCE satellite consists of the total field and the second derivatives of the gravity field. The observations are made at two satellite altitude heights at 255 and 225 km (Bouman et al., 2016), and the global grids are made available through internet (<http://goce4interior.dgfi.tum.de>). The gravity total field and gradient grids have been used in several studies to image the crustal and upper mantle structure (Bouman et al., 2015; Fulla et al., 2015; Fadel et al., 2015). The inversion of gravity data is well-known to suffer from non-uniqueness because of the lack of vertical sensitivity, however, in joint inversion, gravity observations have been shown to provide constraints on the density structure (Moorkamp et al., 2011, and references therein). Here, I will use the gravity total field (G_z) and the second vertical gradient (G_{zz}). The full gravity gradient components could be used. However, the G_{zz} component was selected to reduce the computation time, and because it can be directly linked to the subsurface structure, given the geometrical complexities of the other components.

Rayleigh wave group and phase velocity measurements are sensitive to variations in shear wave velocity structure in the Earth. As presented in Chapter 5 and 6, the depth sensitivity depends on the measured periods. Short periods, e.g., 1-30 s, are sensitive to the velocity distribution in the crust. Longer periods, up to 200 s, can provide estimates on the velocity structure of the upper mantle.

9.4 Methodology

The joint inversion technique is designed to minimize the following objective function:

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Figure 9.1 The synthetic model designed to test the joint inversion technique. (1) shows a depth slice at 15 km with a positive and negative shear wave anomaly (+0.46 and -0.44 km/s). The east-west profile A-B goes through -24.5 degrees latitude. The north-south profile C-D profile goes through 27.5 longitude. (2) shows the depth slice at 15 km with a positive and negative density anomaly (+0.24 and -0.16 g/cm³). The east-west profile A-B goes through -24.5 degrees latitude. The north-south profile C-D profile goes through 27.5 longitude.

Figure 9.2 Calculated gravity data for a satellite altitude of 225 km using the 3D model of Figure (9.1) contaminated with normally distributed random noise. (A) shows the total gravity field (Gz) and (B) shows the second vertical gradient (Gzz).

$$\begin{aligned} \phi = & \| [g(m) - d_{obs}]_{sw} * w_{sw} \|_2^2 + \| [g(m) - d_{obs}]_{gv} * w_{gv} \|_2^2 + \\ & \| [\frac{\partial m}{\partial x} + \frac{\partial m}{\partial y} + \frac{\partial m}{\partial z}] * \alpha \|_2^2 + \| [m - m_o] * \beta \|_2^2 \end{aligned} \quad (9.1)$$

Where:

- the first term describes the Rayleigh wave dispersion data misfit where $g(m)$ is the forward solver,
- the second term describes the gravity data misfit term,
- the third term is a model regularization that describes directional smoothness in X, Y, and Z, and
- the last term controls the degree to which the inversion result can vary from the starting model.

I used a 3D solver for the fundamental mode Rayleigh wave group and phase velocities. The 3D Rayleigh wave solver is based on a 1D code in spherical coordinates that is an earlier version of the Mineos software (Woodhouse and Doornbos, 1988). The 1D code iterates over the 2D grid points to calculate the Rayleigh wave group and phase velocities of the 3D input Earth model. The process was parallelized using multi-processors to improve the performance and to reduce the computing time. For the gravity component I used the Tesseroids package (Uieda et al., 2016). Tesseroids is a collection of command-line programs for modeling the gravitational potential, acceleration, and gradient tensor. Tesseroids supports models and computation grids in Cartesian and

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Figure 9.3 Rayleigh wave group and phase velocity calculated for the synthetic model of Figure (9.1) contaminated with normally distributed random noise. (1) shows the group velocity dispersion map at 15 s period. A-B and C-D show the group velocity curves calculated for the same locations as in Figure (9.1). (2) shows the phase dispersion map at 15 s period. A-B and C-D show the phase velocity curves for the same location as in Figure (9.1).

spherical coordinates. The geometric element used in the modeling processes is a spherical prism, also called a tesseroid. Tesseroids also contains programs for modeling using right rectangular prisms, both in Cartesian and spherical coordinates. In this case, the Tesseroid code was used for 3D gravity total field and gradients forward calculations in spherical coordinates.

The developed inversion technique uses the iterative damped least squares approach based on the Levenberg-Marquardt algorithm (Moré, 1978) to minimize the joint inversion misfit function in equation 9.1. The Jacobian matrix of the misfit function was calculated using a central finite difference scheme (Press, 2007). The iterative joint inversion technique iterates until one of the following two criteria are achieved: 1) the number of iterations exceeds a predefined number of iterations, or 2) the misfit function drops below a predefined threshold. The shear wave velocity is the main target for the inversion process. The shear wave velocity was coupled to density and Vp velocity using the relation found by Brocher (2005).

9.5 Synthetic test

The synthetic test has been developed with spatial dimensions that are similar to Botswana since the main idea is to apply the technique on NARS-Botswana and GOCE data in future studies. The model area is 20×20 degrees which covers the area of Botswana (less than 10×10 degrees) adding a 5 degrees extension to all directions to avoid edge effects in gravity modeling as can be seen in the results of Chapter 8. The depth of the model extends to 50 km to cover the crustal and the uppermost mantle. The model was discretized by 5 degree steps in east and west directions and 10 km in the Z-direction to guarantee fast computation to facilitate the testing purpose of this chapter.

Our gravity observations simulate the total field (Gz) and the second vertical derivative (Gzz) measured at a satellite altitude of 225 km. Random noise with a normal distribution, zero mean, and standard deviation of 2% of the maximum amplitude of Gz and Gzz was added to the measured signals to simulate a real case scenario. Rayleigh wave group and phase velocities at periods of 1-50 s were used as the seismic surface wave observations to guarantee dominant sensitivity in depth range of 1-50 km as discussed in Chapter 5 and 6. Random noise with a normal distribution, zero mean, and standard deviation of 0.025 km/s was added to the measured group and phase velocities to simulate the uncertainty level of real observations. Figure (9.1) shows the designed synthetic model. Figures (9.2) and (9.3) show the calculated gravity (Gz and Gzz), and Rayleigh wave dispersion (group and phase velocity) data contaminated with the normally distributed random noise, respectively. Figures (S1) and (S2) of the supplementary material show the clean gravity (Gz

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Figure 9.4 The starting model for the inversion process. (1) shows the shear-wave velocity at 15 km depth. A-B and C-D show the shear-wave velocity profiles similar as in Figure(9.1). (2) shows density at 15 km depth. A-B and C-D are 2D density profiles at the same location of the 2D profiles in the synthetic model Figure(9.1).

and G_{zz}), and Rayleigh wave dispersion (group and phase velocity) data, respectively.

The joint inversion technique was applied on the synthetic data. The calculated group and phase velocities (Figure (9.3)) and the calculated gravity signal (Figure 9.2) were used as the input measured signal. A starting model (Figure 9.4) with a uniform shear wave velocity of 3.5 km/s and 0 density anomaly was used to start the inversion process. The maximum number of iterations was defined as 20 and the termination threshold was defined as 0.01% of the starting misfit to let the inversion minimize the objective function as low as possible within the available number of iterations.

9.6 Results

The inversion results presented in Figure 9.5 show that the developed technique was able to retrieve the 3D subsurface structure, even if the starting model is relatively far from the target model. A visual comparison between the true model in Figure(9.1) and the inversion results show a good recovery of the main features of the model, also when we have added noise to our data. The direct coupling relation enables the recovery of shear-wave and density structure simultaneously. Comparing the retrieved densities in comparison to the true model shows that the inversion was able to retrieve the density structure as well. This is to be expected since the velocity-density relationship was not violated and it shows that the developed technique fundamentally works under the assumptions and environment of the definition of the presented synthetic test. The uncertainties associated with the joint inversion results do usually not provide meaningful bounds on the model parameters. Julià et al. (2000) have shown that uncertainty estimations in least squares inversions do not provide confident boundaries on the model parameters. Therefore, the accepted practice to estimate the uncertainty associated with the inversion process is to carry out the inversion with different weights, smoothing values, and constraints to be able to provide a general overview about the uncertainty of the inverted model and the robust geological features in it (Maceira and Ammon, 2009).

9.7 Discussion and Future Work

The presented technique to jointly invert the gravity and surface wave observation showed promising results when applied on simple synthetic data. This provides a proof of concept for the technique, and opens the door for future applications using real data. However, one of the main concerns regarding the developed technique is the direct-coupling relationship between the velocity and density. The relation of Brocher

9. Joint inversion of seismic surface wave and GOCE gravity gradients

Figure 9.5 The retrieved model from the inversion results. (1) is shear-wave velocity depth slice at 15 km depth. A-B and C-D are 2D shear-wave velocity profiles at the same location of the 2D profiles in the synthetic model Figure(9.1). (2) shows the density at 15 km depth. A-B and C-D are 2D density profiles at the same location of the 2D profiles in the synthetic model Figure(9.1).

(2005) can provide constraints on the crustal structure, however, the relation may be strongly violated in the case of the presence of melt or fluids. Moreover, the relation is not valid for upper mantle conditions. Therefore, one of the main features that should be investigated in the future work is the usage of mineral physics to derive the direct-coupling relation between seismic velocities and density. This can be achieved using mineral physics and thermodynamical modelling based on Gibbs free energy minimization like in the PerpleX software (Connolly, 2005). Another fundamental concern regarding the developed technique is the computing time. The simple example provided in this chapter with $4 \times 4 \times 5$ dimensions took approximately 5 minutes on Dell Precision 3510 laptop with 16 Gb of Ram and using 7 processors. However, the computational demand will exponentially increase with increasing model dimensions. An efficient scheme to compute the Jacobian matrix is required, instead of the currently used central finite difference scheme, to improve the performance of the developed technique. Finally, future applications can potentially include the full gravity gradients from GOCE instead of Gzz component only used in this chapter.

9.8 Acknowledgments

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9.9 Supporting information

Figure S1

Figure 9.6 Gravity calculated data at satellite altitude (225 km) using 3D Earth model in Figure (9.1) without the contribution of the normally distributed random noise. (A) is the total gravity field (G_z) and (B) is the second vertical gradient (G_{zz}).

Figure S2

Figure 9.7 Rayleigh wave group and phase velocity signal using the synthetic model in Figure (9.1) without the contribution of the normally distributed random noise. (1) is the group dispersion map of 15 s period. A-B and C-D are group velocity 2D profiles calculated at the same location of the 2D profiles of the Earth model in Figure (9.1). (2) is the phase dispersion map of 15 s period. A-B and C-D are phase velocity 2D profiles calculated at the same location of the 2D profiles of the Earth model in Figure (9.1).

Through this thesis, I focused on two major components; the first brought insight into the crustal and upper mantle seismic structure of Botswana, and the second investigated two methodological developments to integrate seismological and gravity data.

10.1 The crust and upper mantle structure

The crustal and upper mantle structure of Botswana was found to have three distinctive features: 1) the deep sedimentary basins, 2) the extension of the East African Rift System (EARS) in the lower crust and upper mantle and 3) the cratonic blocks. The findings, based on analysis of the discontinuous- and seismic velocity structure, have been imaged using different techniques. The discontinuous structure was imaged with the receiver function technique, the velocity structure with ambient noise analysis and earthquake based surface waves.

Two dominant sediment basins have been identified, the Passarge and Nosop basins. They both have a remarkable low-velocity signature in the upper crustal velocity structure of Botswana. The receiver functions also show evidence for the presence of thick sediments in these basins. At places the sediment thickness exceeds 10 km. Clearly separated from the sediment controlled low velocity anomalies in the upper crust, we observed low velocity anomalies in the lower crust and upper mantle. A known, and expected feature, was the EARS in northeastern Botswana which appears as very low shear wave velocity combined with a thin crustal thickness. One of the most remarkable findings of this thesis is that these low-velocity expressions extend from the EARS until the borders of the Kalahari and Congo cratons. The extension towards the Congo craton is relatively well known from seismic activity in the area but was not previously imaged. Most striking observation is that this anomaly further extends, through the upper mantle and then coming up again, towards the Kalahari craton, around 500 km southwest of the supposed terminus of the EARS. The low-velocity extension corrodes the margin of the cratons forming low-velocity finger-like structures into the

boundaries of the cratons. The low-velocity 'fingers' probably penetrate the crust at weak structural zones. The first time we observed this the model resulted at some raised eyebrows from the scientific community (and with ourselves too, to be honest). But sensitivity and resolution tests indicated that this feature was not likely to be an artefact. The lack of documented seismic activity was the main argument against our theory of incipient rifting in this area. The noticeable earthquake activities at the Okavango Rift Zone in northern Botswana was well documented but there was no record of such activity in central Botswana. An explanation for this might be found in the fact that this low-velocity anomaly is located beneath the Central Kalahari Game Reserve. With an area of over 50.000 square km it is the 2nd largest game reserve in the world, it covers around 1/10th of Botswana, and is slightly larger than the Netherlands. With the lack of seismometers around and in the park it is very likely that any earthquake activity might not have been detected. On April 3, 2017 a large earthquake struck the area. The combination of the extensional mechanism of the earthquake together the large magnitude (6.5 M_w) provided the support that our hypothesis of rifting in this area was more than likely. Combined with a connection in the upper mantle between the EARS and this region, thin crust and high V_p/V_s ratio, we believe we now have a convincing case that the EARS is actually, at least, 500 km longer than previously thought. There has been some debate in literature about the extend of the Kaapvaal craton, particularly about its extend in western direction underneath southern Botswana. We find that the Kaapvaal and the Zimbabwe cratons have high shear velocities and a deep keels extending to more than 200 km depth. Their boundaries are well defined. The Kaapvaal craton shows a clear boundary in the west, and indications are visible of a mobile belt towards the west. More to the west, towards the border with Namibia, another cratonic root is visible, but only at depth. This might be the elusive Maltahohe craton, visible underneath the deep sands of the Rehoboth province.

10.2 Integrating gravity and seismological observations

In the second part of this thesis, I introduced two different strategies to integrate seismological and gravity observations. The two approaches are fundamentally different from each other. The first approach is an object-based technique. Object based techniques are widely used by the remote sensing and medical communities but this was the first time that it was applied for segmentation of a 3D seismic velocity model for gravity-based inversions. 3D object based segmentation techniques are based on spatial variations of a specific physical parameter or a combination of physical parameters. This makes it intrinsically useful for 3D tomographic models. If combined with a priori geological knowledge

it is possible to divide an upper mantle velocity model into separate segments, where segments can then be clustered based on similarity measures. When used for gravity inversions the biggest advantage is that there are improved definitions of the boundaries both vertically and laterally. In particular, density variations in depth are found to be difficult to detect with traditional inversion techniques.

This approach provides a reproducible and robust extraction process that reduces the subjectivity of the interpretation and inversion of 3D subsurface models. The extracted objects can be later inverted for their density contrast using linear least-square inversion scheme. However, the estimated density contrast values resulting from the inversion scheme are sensitive to minute variations in the 3D model. This sensitivity can be reduced by adding constraints on the density solution space. Density values, or the corresponding density contrasts, should be constrained by boundary values that can be derived from seismic velocities using empirical relations or, even better, mineral physics using thermodynamical modelling based on Gibbs free energy minimization like in the *PerpleX* software (Connolly, 2005).

The second approach does not require a pre-defined 3D seismic velocity model to start with. It will use seismological data jointly with gravity data (total field and gradients). Specific boundary information is now obtained from the satellite gravity gradients. It will be matched to the seismological data and jointly inverted for velocity and density structure. The example shown in this thesis is relatively simple in setup but should be tested with real data. It is a possibility to include, in due time, mineral physics equations in the inversion process. The real world testing and inclusion of additional information is definitely recommended for future work.

10.3 Future Work

In recent years a lot of geophysical data has been acquired in Botswana and is increasingly becoming available through the Botswana geoscience portal as well. Besides the seismological data acquired in this project there are several airborne geophysical datasets, ground-based magnetotelluric measurements available for Botswana. Also the recent development in satellite geophysics, like the GOCE and GRACE satellites used in this research, but also the SWARM mission (measuring the magnetic and electrical conductivity fields) contribute to the data availability to study previously data-poor regions. The large volume of geophysical data that covers Botswana gives ample opportunities to image the crustal and upper mantle structure in unprecedented spatial and depth resolution as we show in this research. The high-quality gravity and magnetic data, the recent MT data from SAMTEX magnetotelluric experiment, and the

emerging data from SAFARI and NARS-Botswana seismological networks need an innovative technique to integrate all of them into a single sub-surface model with different physical attributes e.g., seismic velocities, densities, and electrical conductivity.

As a first step, in order to show the potential of the available of high-quality geophysical data across Botswana, I have inverted the MT data in collaboration with Dr Naser Meqbel from GFZ Potsdam. We used data from 235 MT locations in Botswana that are freely distributed through the SAMTEX project website. We inverted the data using the freely available ModEM 3D MT inversion code to estimate the 3D electrical conductivity structure underneath Botswana. The preliminary results (Figure 10.1) shows interesting features in the area of the earthquake location. The correlation with our seismological models is striking at some locations, but also shows some different features that can't be identified in other datasets. Since this data is looking at a completely different, and largely independent, geophysical parameter, the need for joint interpretation, analysis and inversion is evident.

One of the emerging promising methods is the multi-observables probabilistic inversion technique (LITHMOD3D) developed by the group of Juan Carlos Afonso (presently at Macquarie University, Sydney) and his colleagues. The programme has the potential to invert gravity, seismic, elevation, geoid, MT, and surface heat flow data jointly to derive the thermochemical structure of the lithosphere. The recommended technique links a group of geophysical forward solvers and their corresponding physical properties using thermodynamical modelling. It uses a probabilistic inversion approach that is based on stochastic searches through the entire parameter space (including physical properties, composition, and temperature) to find optimum solutions that provide the best fit to the data. We can thereby retrieve models of the thermochemical structure of the lithosphere (e.g., pressure, temperature, seismic velocities, densities, and electric conductivities). The technique was applied successfully on data from the United States and showed promising results. However, this approach has limitations in case of the absence of a priori information on the compositional and thermal structure. Moreover, such approaches are not well suited for the state-of-the-art geophysical forward solvers due to the high computational demands. Another alternative can be a linearized inversion scheme that includes all the needed forward solvers and sensitivity kernels. We have attempted to make a start with this in the joint inversion approach. Applying this for Botswana is a next step in fully understanding the geodynamics of Botswana and surrounding regions and will be the area of future research.

Figure 10.1 The inversion results of MT observations at 235 locations in Botswana. (A) and (B) show the MT locations and the color indicates the total misfit between the observed and the calculated data after the 3D inversion. (C) shows the location of the 2D profiles D-E and F-G on the 30 km depth slice from the 3D shear wave velocity model presented in chapter 6. D-E is a north-south 2D profile from the 3D electrical conductivity model of Botswana crossing the location of the 6.5 M_w Botswana earthquake. F-G is a east-west 2D section crossing the location of the 6.5 earthquake. The 6.5 M_w is presented in (C), D-E, and F-G as a red star.

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Biography

Islam Fadel was born on 4th of April 1986 in Cairo, Egypt. He obtained a Bachelor's degree in geophysics in August 2007 from the Earth Science Department, Helwan University, Egypt. He has been working as an instructor at Helwan University, Egypt, since September 2007. Moreover, he worked as an instructor at Cairo University, Egypt in the academic year of 2010-2011. On September 2011, he started Erasmus Mundus scholarship to carry out his MSc in Faculty of Geo-information Science and Earth Observations (ITC) with specialization in Earth Resource Exploration. In March 2013, he started his PhD studies at the Faculty of ITC to model the African lithosphere combining GOCE and seismology data.

Author's publication

Journal articles

1. van der Meijde, M., Fadel, I., P. Ditmar, and M. Hamayun (2015), Uncertainties in crustal thickness models for data sparse environments: A review for South America and Africa, *Journal of Geodynamics*, 84,1-18.
2. Fadel, I., van der Meijde, M., Kerle, N. and Lauritsen, N. (2015), 3D object-oriented image analysis in 3D geophysical modelling: Analysing the central part of the East African Rift System. *International Journal of Applied Earth Observation and Geoinformation: JAG*, 35 (2015) Part A pp. 44-53.
3. Fadel, I., Kerle, N. and van der Meijde, M. (2014), 3D object-oriented image analysis of geophysical data. *Geophysical Journal International*, 198 (2014)1 pp. 357-365.

Journal articles under review and in preparation

1. Fadel, I., van der Meijde, M., and Paulssen, H., Crustal structure and dynamics of Botswana. *Journal of Geophysical Research: Solid Earth* (minor revision).
2. Fadel, I., Paulssen, H. and van der Meijde, M., Ambient noise tomography of Botswana. *Geophysical Research Letters* (to be submitted).
3. Fadel, I., Paulssen, H. and van der Meijde, M., Crustal and upper mantle shear wave velocity structure underneath Botswana. *Earth and Planetary Science Letters* (in preparation).

Conference paper

1. Fadel, I., Kerle, N. and van der Meijde, M. (2014), 3D object-oriented image analysis with applications in 3D geophysical modeling. *South-Eastern European journal of earth observation and geomatics*, 3(2014)2S: Special issue on: GEOBIA 2014: Advancements, trends and challenges: 5th Geographic object-based image analysis conference, Thessaloniki, Greece, May 21-24, 2014. pp. 205-208.

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2. Fadel, I., van der Meijde, M., Paulssen, H., and Kwadiba, T., Extending the Rift Zone: evidence from central Botswana. AGU Fall Meeting, 11-15 December 2017, New Orleans, United States of America.
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