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S-wave receiver function studies in African sedimentary basins

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UNIVERSITY OF THE WITWATERSRAND



School of Geosciences

Faculty of Science

S-WAVE RECEIVER FUNCTION STUDIES IN AFRICAN SEDIMENTARY BASINS

by

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Submitted in Partial Fulfillment of the Requirements for the Degree of

Master of Science

Declaration

I, **Helio Filemone Inguane**, declare that this dissertation is my own unaided work. It is being submitted to the Degree of Master of Sciences at the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination to any other University.

.....

(Signature of candidate)

..... day of Year.....

Abstract

Sedimentary basins are the result of prolonged subsidence of the Earth's surface. They occupy 45% of the African surface. Knowledge of their area and depth is important because they often contain mineral, energy and groundwater resources. The transition between crust and mantle (the Moho) is believed to hold important clues to the Earth's evolution and has been the subject of many studies, including P-wave Receiver Function (PRF) studies, to determine the structure and composition of the crust and uppermost mantle in Southern and Eastern Africa.

The PRF method relies on the partial conversion of P-waves produced by teleseismic earthquakes to S-wave waves at the Moho. The travel-time delay between the direct P-wave and the Ps phase is used to deduce the thickness and average velocity of the crust. However, the PRF technique fails in regions where there is strong intracrustal layering (such as sedimentary basins), because the reverberations produced by the layers arrive at the seismometer simultaneously with the Ps phase. Here the S-wave Receiver Function (SRF) method works better, as the Sp phase arrives before any reverberations produced by intracrustal layering. In this study I have used the SRF method to investigate crustal structure beneath sedimentary basins in Southern and Eastern Africa.

The aim of this research was to constrain the crustal thickness and shear wave velocity in seven sedimentary basins in Eastern and Southern Africa using S-wave Receiver Functions (SRFs). Teleseismic earthquakes with magnitude ≥ 5.5 and 60 to 82 degree epicentral distance were used to generate the SRFs using data acquired by seismic stations that were deployed between 2007 and 2013 in three rift basins (Lake Albert, Lake Edward and Rukwa) and four pull-apart basins (Mandawa, Mozambique, Rovuma and Ruvu).

A moveout correction was made to align the SRFs obtained from different earthquakes, enabling them to be stacked to reduce random noise and enhance the signal-to-noise ratio of the S_p phase and the accuracy of the pick of the S_p arrival time. The S_p arrival time uncertainties, typically 0.05 s of time error, were estimated for each station using the bootstrapping method. The surface wave group velocity models for each station (at 10, 15, 20, 25 and 30s periods) were used to constrain the depth-velocity models. The grid search modeling was performed using the DISPER80 package.

The following crustal thicknesses (H) and average crustal shear velocities (V_s) were obtained:

- Lake Albert and Lake Edward rift basins situated within the Mesoproterozoic Ruwenzori orogenic belt: H of 38.8 ± 2.4 km and 33.83 ± 0.9 km, respectively; V_s of 3.72 km/s and 3.73 km/s, respectively;
- Rovuma, Mandawa and Ruvu pull-apart basins within the Neoproterozoic Mozambique mobile belt: H of 32.73 ± 1.8 km, 37.79 ± 2.2 km and 39.63 ± 2.2 km, respectively; V_s of 3.68 km/s, 3.76 km/s and 3.79 km/s, respectively; and
- Phanerozoic Mozambique pull-apart basin: H of 36.9 ± 2.1 km and V_s of 3.7 km/s.

These results were compared with previous studies. The crustal thickness reported in a global review of Proterozoic terrains (Durrheim and Mooney, 1994) ranged between 40-55 km, while Rudnick and Fountain (1995) reported an average thickness of 43 km. For stations located in basins in the Mesoproterozoic Ruwenzori orogenic belt, this study produced estimates of H and V_s of 36.3 ± 2.4 km and 3.7 km/s, respectively. V_s is similar to estimates by Julià et al. (2005) and Tugume et al. (2013) for stations in the same region, while H is a few kilometers thinner. This study obtained H of 32-40 km beneath Neoproterozoic pull-apart Tanzanian coastal basins located within the Mozambique mobile belt (Chatellier and Slevin, 1988), while Tugume et al. (2013) estimated the crust of the adjacent Tanzanian craton to be 39 km thick.

In the Phanerozoic Mozambique basin, this study found H and V_s of 36.9 ± 2.1 km and 3.7 km/s, respectively. Kgaswane et al. (2009), using joint inversion of receiver functions and Rayleigh wave dispersion, estimated the H and V_s for the northeast Limpopo belt (west of the Mozambique basin) to be 40 ± 3 km and 3.7 km/s, respectively.

In general, this study found thinner crust and slower V_s than previous studies. However, it is important to note that the station locations were different. Previous studies analysed P-wave Receiver Functions (PRFs) recorded by stations located in the interior of the continent and near to the Tanzania craton, while this study analysed S-wave Receiver Functions (SRFs) recorded by stations located in rift and coastal sedimentary basins. It is likely that the crust thinned during extension and continental break-up.

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1. INTRODUCTION

The Precambrian is the longest eon in the geological time scale. It is divided into two eras, viz. the Archean (3900 to 2500 Ma) and the Proterozoic (2500 to 570 Ma). The Archean is considered to have been a relatively unstable period: oceanic crust was accreted to cratons or was subducted beneath them; calc-alkaline rocks melted, producing tonalities; and the lower continental crust was metamorphosed and deformed to produce gneisses and granulites (Windley, 1981). The Archean terrains are characterized by relatively thin crust (between 27 and 40 km) and low seismic velocities. The Proterozoic is considered to be a period of stabilization. During this period the Earth's interior cooled and the density of the lithosphere increased, which triggered subduction as well as increases in crustal thickness (Hargraves, 1981).

The crust is the outer shell of the solid Earth and is made of relatively low density rocks (Allen and Allen, 2005). The crust is divided into two different types, viz. oceanic and continental. The continental crust is mostly older than oceanic crust, and is divided into two layers: upper and lower crust. The upper crust can be investigated using direct methods such as borehole drilling to extract rock samples, while the lower crust is investigated through indirect measurements such as refraction seismology. The upper continental crust is typically 20 to 25 km thick, and the density varies from 2500 to 2700 kgm⁻³. The lower continental crust ranges in thickness from 10 to 45 km, with a density that varies between 2800 to 3100 kgm⁻³. The mantle is divided into two regions, the upper and lower mantle. The upper mantle extends to a depth of 680±20 km, while the lower mantle extends to the core-mantle boundary at a depth of approximately 2900 km. The mantle density increases with depth.

The discontinuity between the crust and the mantle is called the Moho discontinuity. The Moho depth varies between regions, as was reported in global studies of the Precambrian by Durrheim and Mooney (1994) and Rudnick and Fountain (1995). There are many regions where the crustal thickness and composition remain unknown. The cause of the increase in velocity at the Moho is still debated but many studies postulated that the abrupt seismic velocity increase beneath the Moho is caused by changes in the rock type (Kearey et al., 2009; Stein and Wyssession, 2003; Cook et al., 2010). In some regions the transition from crust to mantle appears to be relatively sharp, in other areas gradational, and in yet other areas it seems to have a laminated structure with horizontal variations.

Seismological techniques are widely used by geophysicists to investigate the structure of the Earth's interior. The seismic reflection technique is widely used to explore the upper few kilometers of the crust for oil and gas. The seismic refraction technique is often used for deeper studies. In refraction seismology,

widening the distance between source and receiver increases the depth of investigation. Teleseismic events are often used as energy sources to probe as deeply as the Moho discontinuity, the Lithosphere-Asthenosphere Boundary (or LAB), and the boundary between the upper and lower mantle. Another popular technique is the Receiver Function method, which I have used for this study.

Receiver Functions (RFs) are time series that are calculated by deconvolving the vertical component signal from the longitudinal component of teleseismic waveforms. RFs are used to study the crustal and mantle structure beneath the receiver (Ammon, 1991; El Tahir et al., 2013; Lay and Wallace, 1995; McNamara and Owens 1993; Midzi and Ottemöller, 2001). The Receiver Function techniques are divided into two different types and named according to type of the incident wave: a P-wave Receiver Function (PRF) originates when an P-wave crossing a discontinuity is partially converted to a S-wave, the latter phase being denoted Ps; while an S-wave Receiver Function (SRF) originates when an S-wave crossing a discontinuity is partially converted into a P-wave, the latter phase being denoted Sp.

Many studies have attempted to use the PRF method to determine the thickness (H) and average shear wave velocity (V_s) of the Precambrian Mozambique, Ubendian and Ruwenzori belts and the crust beneath various Phanerozoic sedimentary basins in southern and eastern African. These studies are reviewed in Chapter 2 of this dissertation. However, the PRF results were generally poor for stations situated above sedimentary basins filled with unconsolidated Mesozoic–Cenozoic rocks because the intra-crustal layering produces seismic phases that obscure the Ps phase. The main objective of this study is to investigate whether better results can be obtained using the S-wave Receiver Function method (SRF), as the Sp phase does not coincide with the crustal reverberations.

A grid search method is an algorithm that systematically explores the solution space. In this study, the grid search method was used to find the crustal thickness (H) and average shear wave velocity (V_s) that best satisfies the observed SRFs. The selection of velocity-depth models was constrained by comparing their associated dispersion curves with Rayleigh wave group velocity dispersion curves obtained by Raveloson et al. (2015). The crustal thickness and shear wave velocity are presented and compared with the results of previous regional and global studies.

Many studies have been conducted using the S-wave Receiver Function (SRF) technique and produced credible results. Hansen et al. (2009a, 2009b, 2010) investigated the crustal structure of the Transantarctic and Gamburtsev Mountains (East Antarctica) and Kaapvaal craton (South Africa); Kind et al. (2012) the Lithosphere–Asthenosphere boundary (LAB) beneath the North American craton; and Vinnik et al. (2004, 2005) upper mantle discontinuities in Iceland and the Afro-Arabian hotspot; while Wilson et al. (2006) analyzed the suitability of the SRF technique in sedimentary basins using deep focus earthquakes.

2. REVIEW OF THE GEOLOGY OF THE STUDY AREA

The Precambrian terrains in Africa are composed of various cratons and mobile belts. There are various belts in Eastern and Southern Africa, but in this study only three belts are considered, viz. the Mozambique, Ubendian and Ruwenzori belts (Fig. 1).

Allen and Allen (2005) defined sedimentary basins as “*regions of prolonged subsidence on Earth's surface*”. There are various types of sedimentary basins: rift, interior, complex, delta, median, down-warp, pull-apart and subduction basins. Altogether 87 sedimentary basins have been identified in the African continent (Chatellier and Slevin, 1988), occupying 15 million km² or 45% of the surface area (Fig.1).

The seven African sedimentary basins considered in this study can be divided into three groups: the Lake Albert, Lake Edward and Rukwa rift basins (Fig. 1 – red square); the Tanzanian Rovuma, Ruvu, Mandawa pull-apart coastal basins (Fig. 1 – green square); and the Mozambique pull-apart basin (Fig 1 – blue square).

The seismic stations used in this study are widely spread (Fig. 2). According to their geographical location, the seismic stations can be divided into three groups:

- a) Southern stations: MAPH and MSGR (Mozambique basin);
- b) Central stations: INDI and MTWA (Rovuma basin), KIMA and MOHO (Mandawa basin), MANG (Ruvu basin), CHAL (Rufiji Trough) and KYLA (Rukwa basin); and
- c) Northern stations: BUTI (Lake Albert basin) and KATE (Lake Edward basin).

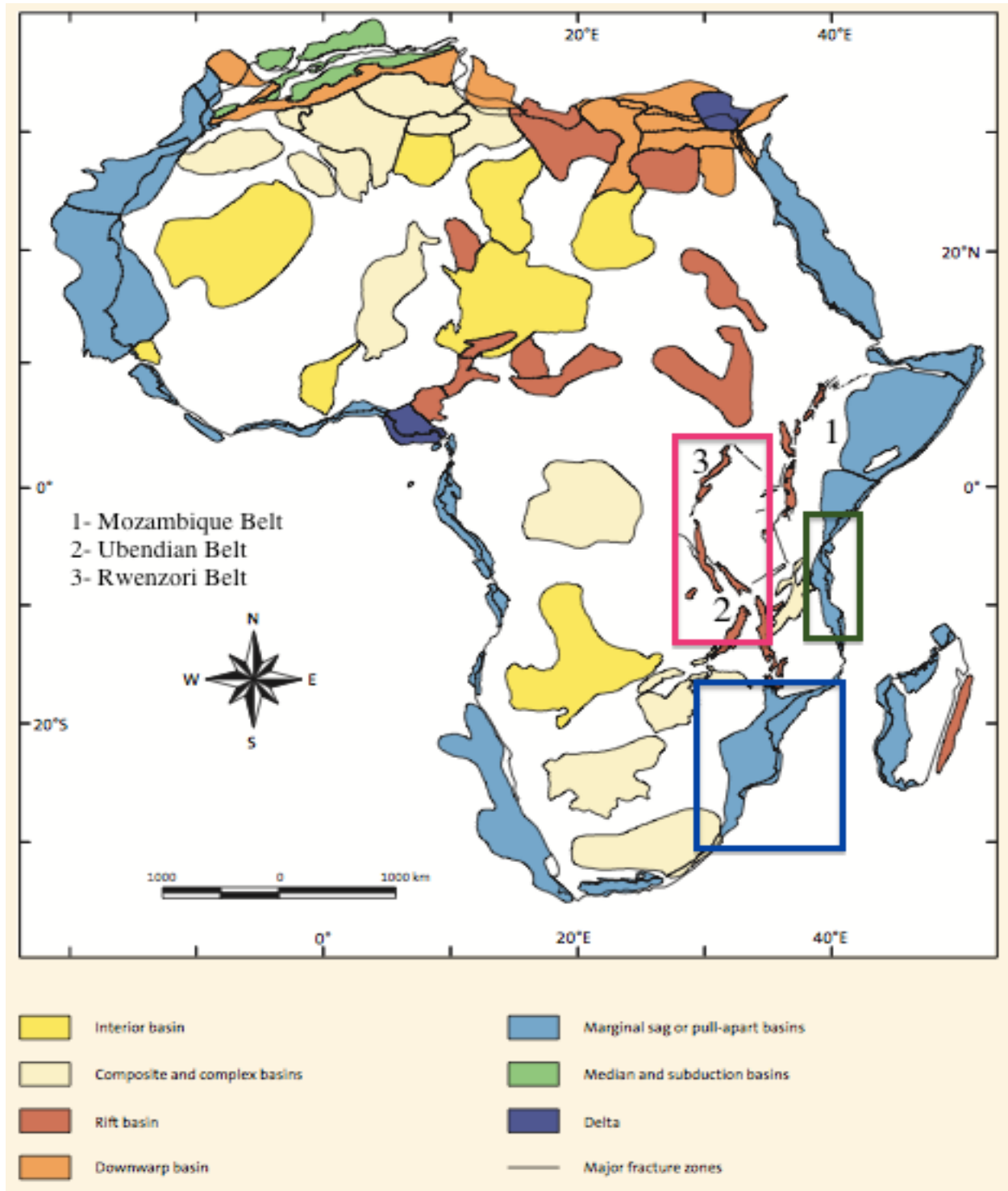


Figure 1 – Types and locations of African mobile belts and sedimentary basins. Blue square, Mozambique pull-apart basin; green square, Tanzanian coastal pull-apart basins; and red square, rift basins (Schlüter, 2008).

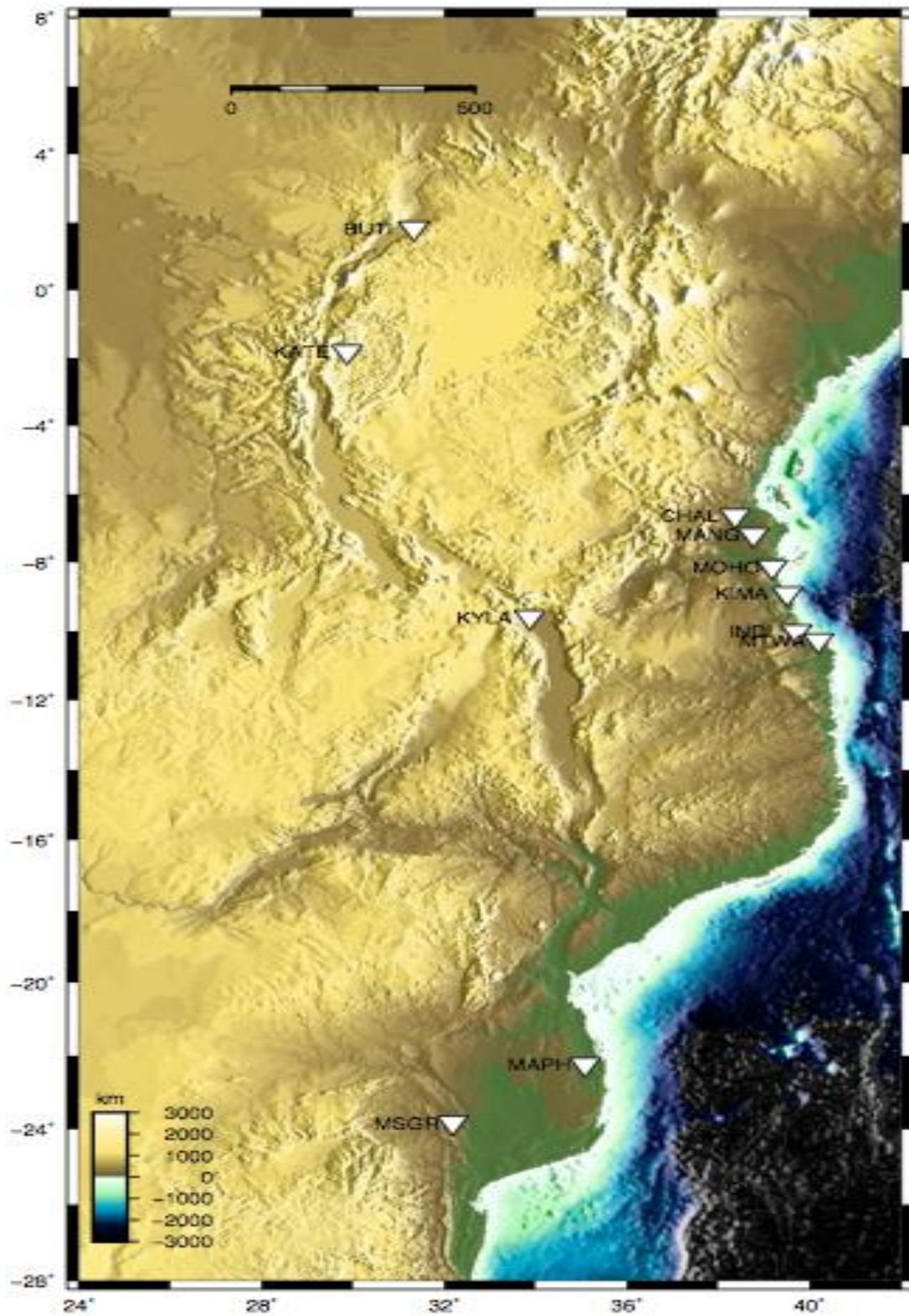


Figure 2 – Map showing the location of the seismic stations deployed in different African sedimentary basins, using topography as background.

2.1. Precambrian Mobile Belts

2.1.1. Mozambique mobile belt

The Mozambique mobile belt is a Precambrian orogenic belt. It stretches from southern Egypt to northern Mozambique. This belt is the result of collisions between east and west Gondwana cratonic elements and is considered to be a Himalayan-type continental collision zone formed between 1200 Ma and 450 Ma by multiple collision events. The southern part of the belt is composed of Neoproterozoic gneisses, while blocks of Archean and Mesoproterozoic continental crust compose the northern part. Granitoids and gneisses dated at 2.7 Ga and xenocrysts aged of 3.0–3.5 Ga have been identified (Dugda et al., 2005; De Waele et al., 2008; Schlüter, 2008; Tugume et al., 2013).

The crustal thickness of Mozambique Mobile Belt was estimated by Dugda et al. (2009) to be between 36 and 42 km with a Poisson's ratio of 0.26 to 0.27. Tugume et al. (2012) estimated 0.24 to 0.28 for the Poisson's ratio, confirming previous studies conducted by Julià et al. (2005 and references within) and obtained similar crustal thicknesses as Dugda et al. (2009). Kachingwe (2014) reported average crustal thickness of 37.3 km and a Poisson's ratio of 0.26.

2.1.2. Ruwenzori Belt

The Ruwenzori Belt is a Mesoproterozoic terrain composed of volcanic sediments and 1.85 Ga granitoids. This belt is bordered by the Tanzania Craton (southeast), Kibaran Belt (south) and the western branch of East African Rift Valley (Begg et al., 2009). The geophysical studies conducted in this terrain by Tugume et al. (2012) estimated a crustal thickness of 37 km and Poisson's ratio of 0.25.

2.1.3. Ubendian Belt

The Paleoproterozoic Ubendian belt is composed of volcanic rocks, amphibolites, gneisses and metasediments formed during two orogenic events. The metamorphism occurred during the first orogenic event between 2.025 Ga and 1.97 Ga. The second orogenic event occurred between 1.94 Ga and 1.86 Ga. Granitoid intrusion and rock deformation overprints granulite and eclogite facies rocks (De Waele et al., 2008; Fernandez-Alonso et al., 2001; Schlüter, 2008; Tugume et al., 2012). The geophysical studies conducted in this terrain by Tugume et al. (2012) reveals a crustal thickness of 42 km and 0.26 for Poisson's ratio.

2.1.4. Seismic structure of Precambrian crust

There is no significant difference in the thickness and composition of the lower crust of the Archean Tanzania craton and the Proterozoic mobile belts surrounding (Juliá et al., 2005). Shear velocities >3.9 km/s in the lowermost crust in the Mozambique and Ubendian belts indicate the presence of a substantial proportion of mafic rocks such as amphibolites, mafic granulites and mafic gneisses. This has been interpreted to indicate that crustal evolutionary processes in the Archean and Proterozoic were similar (Rudnick and Gao, 2003; Juliá et al. 2005). However, the lowermost crustal layer beneath the Mozambique and Ubendian belts is thinner and has lower average shear wave velocities than many other Proterozoic terrains (Durrheim and Mooney, 1994).

2.2. Phanerozoic Basins

2.2.1. Mozambique Basin

The Mozambique pull-apart basin is a Phanerozoic basin characterized by large coastal plains covered by Cenozoic sediments lying above Mesozoic grabens and horsts (Nairn et al., 1991). This basin covers approximately $185,000 \text{ km}^2$ onshore, while another $100,000 \text{ km}^2$ is offshore. The maximum sedimentary thickness reported is 10 km in the Zambezi river delta (Rutten et al., 2008). The Mozambique basin strata reveal the presence of Karoo-age basalts, erupted as a consequence of Mesozoic rifting, in different sites on the Mozambique coastal plains. Tertiary and upper Cretaceous sediments overlie this basaltic basement (Fonseca et al., 2014; Salman and Abdula, 1995; Simpson et al., 1979; Watts, 2001).

The Mozambique Coastal Plain (MCP) is a Mesozoic/Cenozoic terrain situated in the northern part of Mozambique basin and extends for 1200 km of the 2470 km Mozambican coast line (Fig. 3). The strata reveal Mesozoic to Cenozoic sediments overlying Karoo lavas. The lavas were extruded from the Lebombo fracture, located southwest of the Mozambique basin. The sediments were deposited afterwards. The seismic studies conducted in the Mozambique basin revealed the presence of natural gas beneath the Mozambique Coastal Plains. The reserves in the Pande and Temane natural gas blocks were estimated at 3-trillion cubic feet.

2.2.2. Rovuma Basin

The Rovuma pull-apart basin is a Phanerozoic basin and geographically shared between Mozambique (20 %) and Tanzania (80 %) (Fig. 4). This basin covers $17,000 \text{ km}^2$ onshore and

12,500 km² offshore. The oldest sedimentary unit is the Makonda group, located in the northern part of the Rovuma basin, and consists of sandstone and conglomerate. The southern part of the basin has marine characteristics from Neocomian to Albian-Aptian age (Nairn et al., 1991). The Rovuma basin strata reveal the presence of Cenozoic sediments as a result of regional uplift, faulting and the influence of the East African Rift Valley. The sediment thickness reported is 10 km (Key et al., 2008). Seismic studies were conducted in the Rovuma basin for gas and oil prospection both offshore and onshore. On the Mozambican side of the Rovuma basin a natural gas block with 85-trillion cubic feet of reserves was discovered.

2.2.3. Mandawa Basin

The Mandawa pull-apart basin is Karoo-aged basin located along the Tanzanian coast (Fig. 4). Sediments were deposited from the Triassic to the Early Jurassic (241 – 183 Ma). This basin is the major salt basin in Southern Africa, with extension to the southeast towards the Indian Ocean. The Mandawa basin strata reveal the presence of Permo–Triassic and early Jurassic clays, halite (salt) evaporites and Paleogene sediments (Hudson and Nicholas, 2014). The sediment thickness estimated is around 4.5 km. Oil and gas companies conducted geophysical studies in 1952 and afterwards deep wells were drilled.

2.2.4. Ruvu Basin

The Mesozoic Ruvu pull-apart basin is located in northern Tanzania and covers 15,000 km² (Fig. 4). This basin trends in a northeasterly direction and is bounded by the Tanga fault zone (Kingston, 1988). In the Ruvu basin, Permo–Triassic Karoo outcrops are found in the northeast part; while in the southeast Neogene rocks crop out. The Ruvu marine strata reveal the presence of Mesozoic sediments, Tertiary conglomerates, Aptian/Albian sandstones and limestones, and Miocene deltaic sandstone clasts. The sediment thickness estimated is around 3 km. The crustal thickness estimated was 34 and 41 km (Young, 2012).

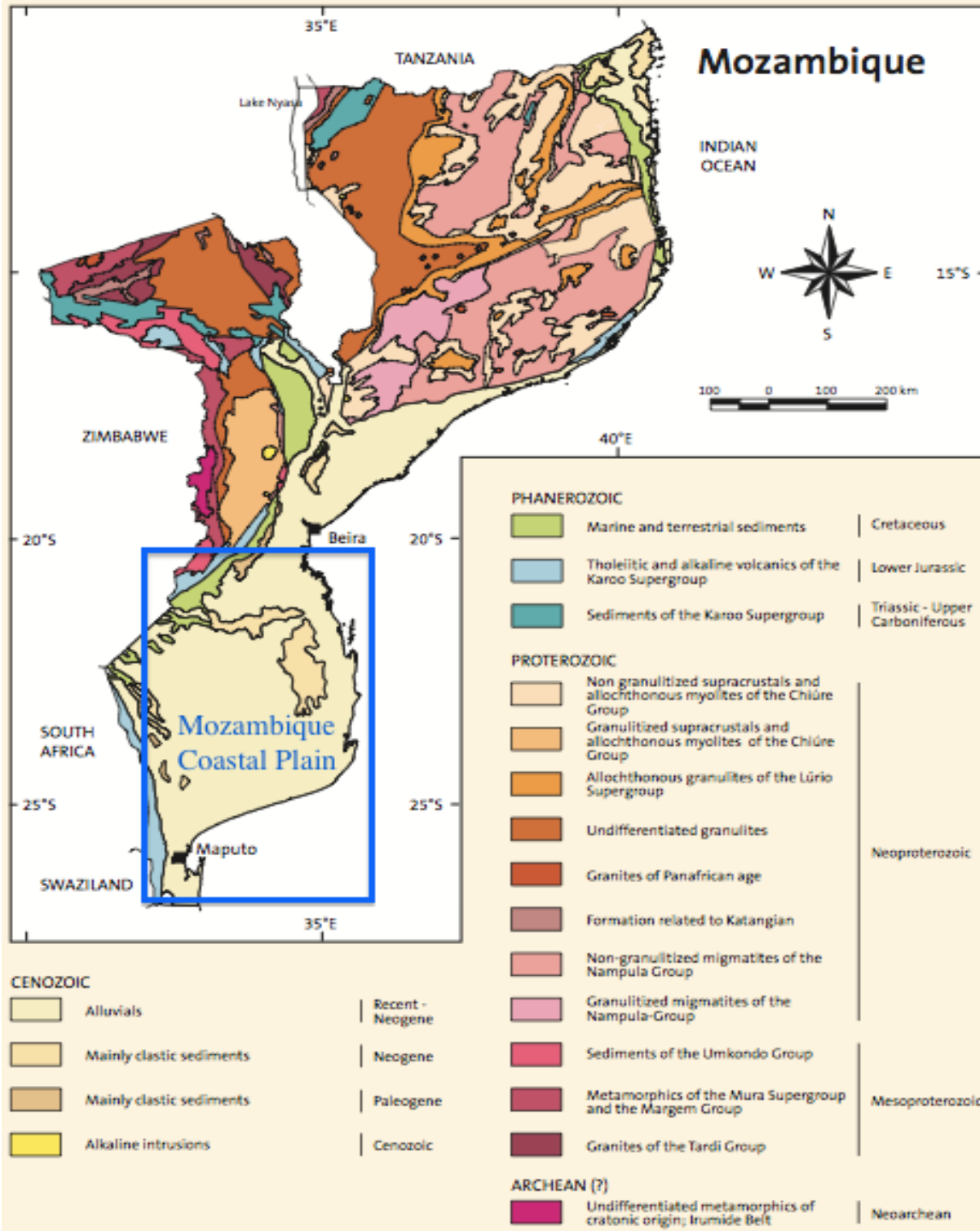


Figure 3 – The geologic map of Mozambique. Blue square area: The Mozambique Coastal Plains (MCP), northern Mozambique basin, which is considered to be the major coastal terrain in southern Africa (Schlüter, 2008).

2.2.5. Rukwa Basin

The Rukwa rift basin is located southeast of the Paleoproterozoic terrain (the Ubendian belt) and along the west branch of the East African Rift System. This sedimentary basin is located southwest of Tanzania between Lake Niassa and Lake Tanganyika and is bounded by two faults: the Kanda fault (southwest) and the Lupa fault (northeast). The Rukwa basin is approximately 300 km long and 50 km wide (Fig. 4). The Rukwa strata reveal the presence of uplifted Paleoproterozoic metamorphic rocks from the Ubendian belt (Fernandez-Alonso et al., 2001; Roberts et al., 2010). The sediment thickness is around 11 km. Geophysical studies (gravity and reflection seismic) were conducted and boreholes drilled for rock sample collection (Kim et al., 2009; Roberts et al., 2010).

2.2.6. Lake Albert Basin

The Lake Albert intra-continental rift basin is located in Uganda, in the western branch of the East African Rift Valley, and was formed during the Miocene (Fig. 5). This basin is 80 to 160 km long and 30 to 80 km wide. It stretches from Lake Edward (southern part) to Uganda and Sudan (northern part). The strata reveal the presence of Cambrian metamorphic rocks and Cretaceous sedimentary rocks, overlain by Neogene sediments. The Lake Albert basin is considered to be a half-graben basin (Karp et al., 2012). The sediment thickness is over 4 km (Abeinomugisha and Obita, 2011). The first well for exploration purposes was drilled in 1938. Afterwards aeromagnetic and gravity surveys were conducted to acquire data over the Albertine Graben in order to update the geophysical database (Abeinomugisha and Obita, 2011).

2.2.7. Lake Edward Basin

The Lake Edward basin is a rift basin composed of a half-graben filled by sediments. This basin is located between the Democratic Republic of Congo and Uganda (Fig. 5). The basin is 80–160 km long and 30–80 km wide. The strata reveal the presence of Proterozoic medium- and high-grade metamorphic rocks, Pleistocene fluvial and lacustrine sediments and Cenozoic volcanoclastics. The Cenozoic volcanoclastic deposits are sourced from the Toro Ankole and Virunga volcanic provinces (McGlue et al., 2006). Gravity modelling indicated the average sediment thickness to be 5 km (McGlue et al., 2006).

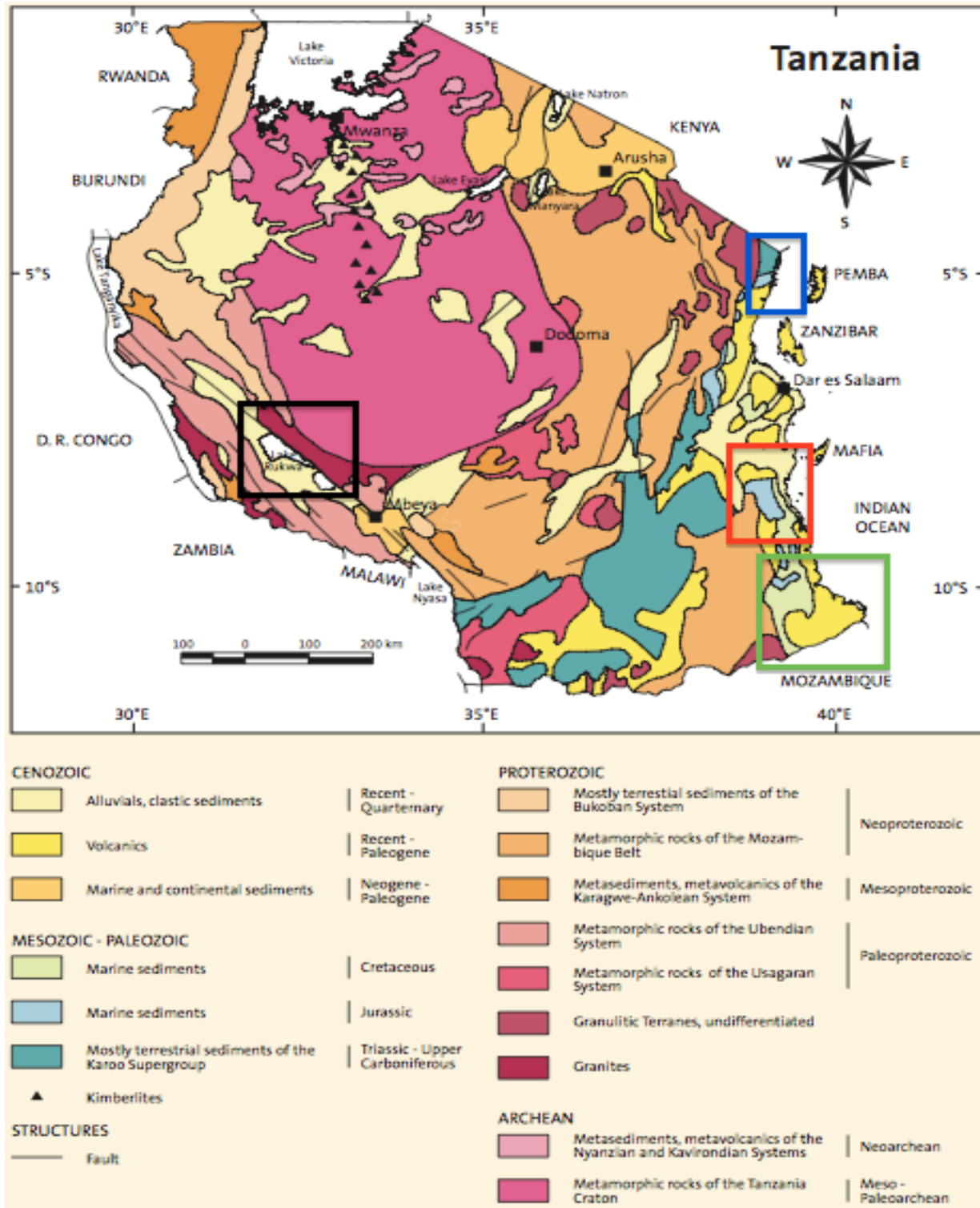


Figure 4 – Tanzanian sedimentary coastal basins and rift basins. Black square – Rukwa rift basin; green square – Rovuma basin; red square – Mandawa basin, and blue square –Ruvu basin (Schlüter, 2008).

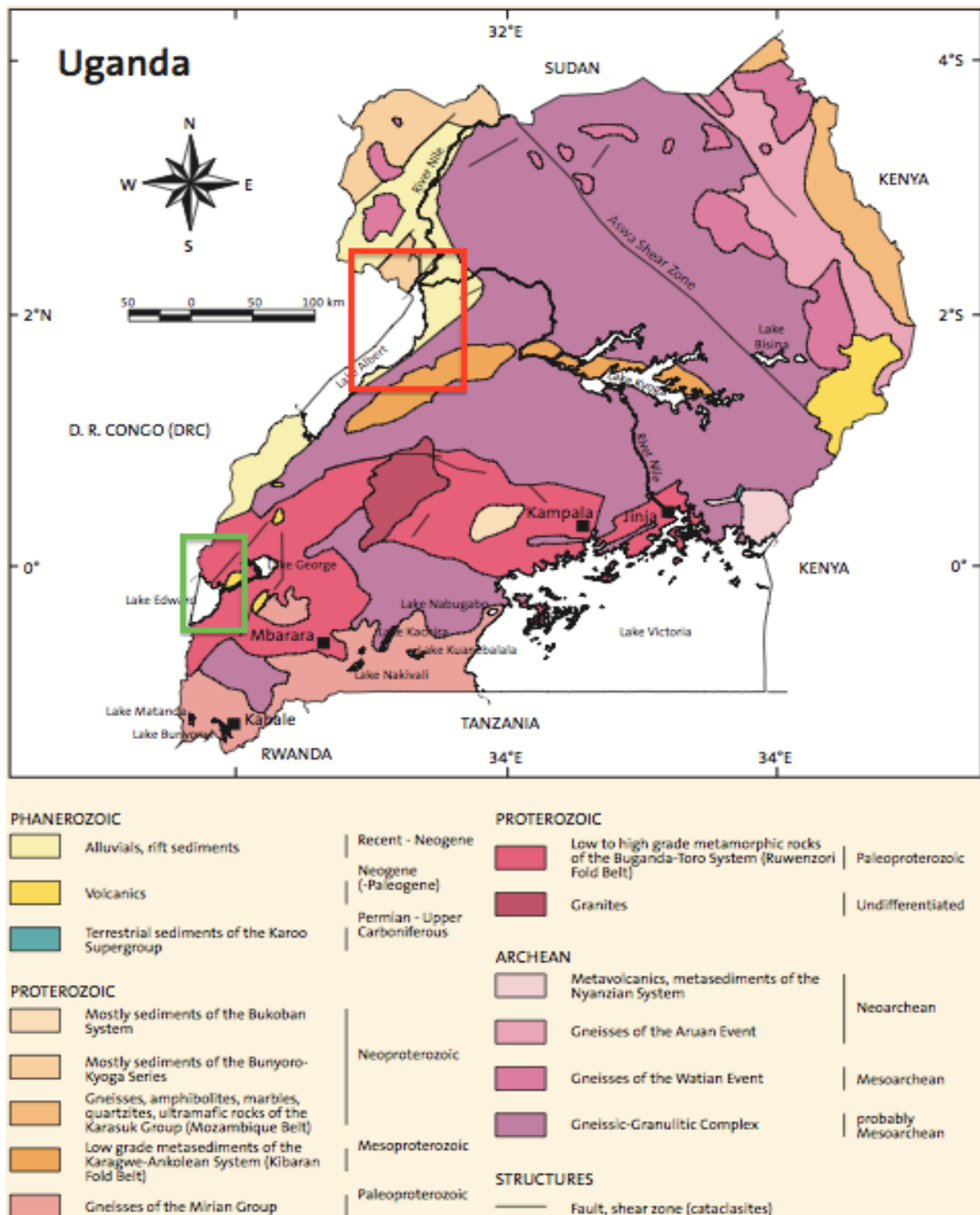


Figure 5 – Uganda geology map. The area delimited by the red square represents the Lake Albert basin, and the green square the Lake Edward basin (Schlüter, 2008).

3. DATA AND METHODOLOGY

3.1. Data

The data used in this research project were collected using temporary seismic stations deployed in southern and eastern Africa during the MACOMO (Madagascar, Comoros and Mozambique) project and the AfricaArray East Africa Broadband Seismic Experiment (AAEASE).

The MACOMO project team deployed seismic stations in Mozambique, Comoros and Madagascar. I was a member of the team that deployed and serviced the stations. Data from two stations in southern Mozambique were used for this study, namely Massingir (MSGR) and Mapinhane (MAPH), labeled as part of the XV network. These seismic stations were deployed between August 2011 and August 2013, and were equipped with broadband portable Nanometrics Trillium Compact sensors and RefTek 24-bit data loggers.

In the AAEASE project, seismic stations were deployed in eastern Africa. Data recorded by nine seismic stations from two different networks (ZP and YH) deployed in Tanzania and Uganda were used in this study. From the ZP network, the Butiaba (BUTI) and Katwe (KATE) stations, both in Uganda, and Kyela (KYLA) in Tanzania were used. These seismic stations were deployed in different periods between August 2007 and August 2010. From the YH network, Chalinze (CHAL), Kilwa (KIMA), Lindi (INDI), Manero Mango (MANG), Mohoro (MOHO) and Mtwara (MTWA) seismic stations, deployed in Tanzania between February 2010 and July 2011, were used. Both the ZP and YH network were equipped with portable broadband sensors (Streckeisen STS-2G1, Guralp CMG-3T and Guralp CMG-40T), 24-bit RefTek data loggers, and GPS (Global Position System) antennas. More detailed information for each station (location, deployed instruments and period of deployment) is provided in Appendix A.

3.2. Receiver Functions (RFs)

Receiver Functions are time series. These time series result from deconvolution of the vertical component signal from the longitudinal component of teleseismic waveforms, and are used to study the crustal and mantle structure beneath the receiver (Ammon, 1991; El Tahir et al., 2013; Lay and Wallace, 1995; McNamara and Owens 1993; Midzi and Ottemöller, 2001).

3.2.1. Converted phases and receiver function types

When an incident teleseismic body wave passes through a discontinuity formed by two layers with different physical properties, the wave will be reflected, refracted and converted into P or S phases. The receiver function technique attempts to isolate the converted phases (Ps and Sp) generated from teleseismic earthquakes by many discontinuities within the Earth such as the Moho.

The incident P-wave passing through a discontinuity creates a converted S-wave (P-wave to S-wave or Ps), which is refracted towards the normal according to Snell's law. It follows the direct P-wave (the precursor). Similarly the incident S-wave creates a converted P-wave (S-wave to P-wave or Sp) when passing through the discontinuity. It is faster and will become a precursor of the direct S-wave. The Sp ray is refracted away from the normal to the interface and thus has a longer ray path than the Ps ray and has a conversion point that is further from the station (Fig. 6).

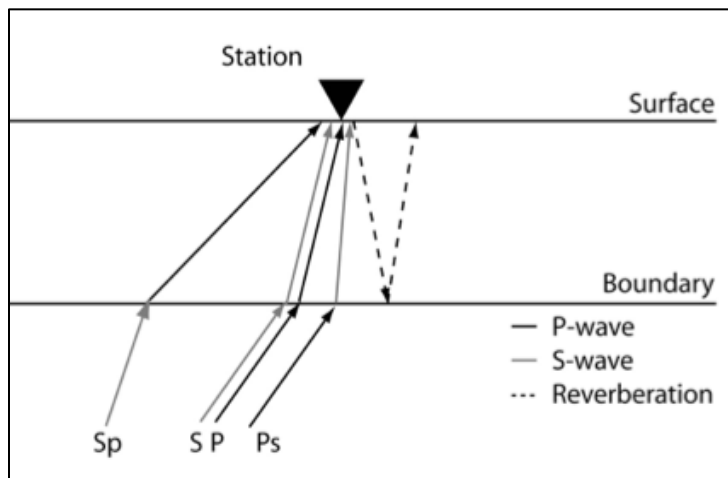


Figure 6 – Illustration showing the ray paths of direct P and S phases, converted Ps and Sp phases and reverberations passing through the boundary on their way to the receiver (Hansen et al., 2009b).

The Receiver Function techniques are divided into two different types and named according to type of the incident wave:

- a) PRF (P-wave Receiver Function) – when an P-wave crossing a discontinuity generates a converted phase, P-wave to S-wave or Ps; and

- b) SRF (S-wave Receiver Function) – when an S-wave crossing a discontinuity generates a converted phase, S-wave to P-wave or Sp.

Both Receiver Function techniques are widely used and considered powerful techniques for investigating crustal structure (Hansen et al., 2009a, 2009b; Kumar et al., 2006; Lin et al., 2011; Vinnik et al., 2004; Young, 2012).

3.2.2. Receiver Function technique

Consider a P-wave as an incident wave passing through an interface h (Fig. 7 – Left). The receiver records the direct phase wave (Pp), the converted phase wave (Ps) and the multiple phases (PpPhs, PsShs and PsPhs+PpShs). The idealized waveform (Fig. 7 – Right) clearly shows the sequence of phase arrivals recorded by the receiver.

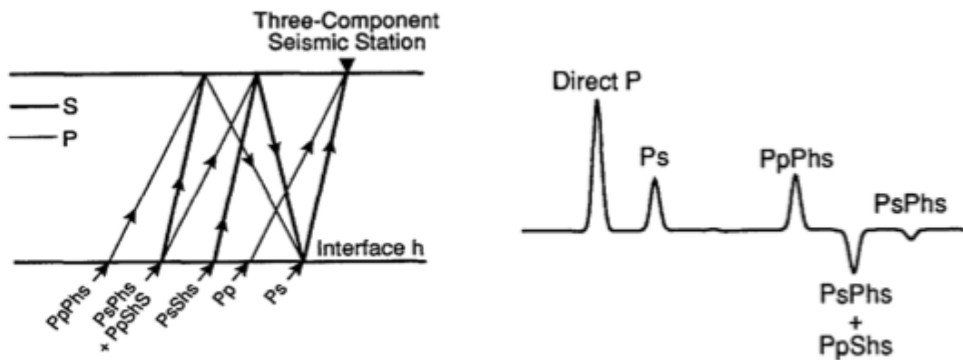


Figure 7 – (Left) Diagram showing direct phase (Pp), converted phase (Ps) and multiple phase ray paths passing through interface h and arriving at the three-component seismic station (receiver) and (Right) the correspondent waveform of phases arriving at receiver (Ammon, 1991).

In the P-wave Receiver Function, the precursor phase is the direct P-wave followed by the converted phase (Ps) and multiples in the seismogram coda. The converted phase (Ps) is not a precursor because it behaves as an S-wave and is slower than the P-wave.

3.2.3. Receiver Function generation

In the Receiver Function technique, teleseismic body waveforms are processed and modeled to extract the crustal and mantle structure information underneath the receiver. This information

can be contaminated and it is necessary to clean the signal by removing source and mantle path effects to obtain the Receiver Function.

The three-component seismic trace is modeled as a convolution

$$s(t) = s_e(t) * e(t) * i(t) * h(t) \dots\dots\dots (1)$$

where

$s(t)$ is the three component seismic trace;

$s_e(t)$ is the earthquake source time function;

$e(t)$ is the impulse response of the layered earth structure;

$i(t)$ is the impulse response of the recording instrument; and

$h(t)$ is the receiver function.

The variables $s(t)$ and $h(t)$ are vectors (Langston, 1979; Hansen and Dueker, 2009 and Ammon, 1991).

The ray path of an incoming P-wave one-dimensional velocity structure is shown in Fig. 7. The source time function is removed by deconvolving the vertical $z(t)$ component from the radial $r(t)$ or transversal $t(t)$ component of motion. Thus:

$$z(t) = \sum_{k=0}^n z'_k s_e(t - t_k) \dots\dots\dots (2)$$

$$r(t) = \sum_{k=0}^n r'_k s_e(t - t_k) \dots\dots\dots (3)$$

$$t(t) = \sum_{k=0}^n t'_k s_e(t - t_k) \dots\dots\dots (4)$$

Where

k is the number of phases. If $k = 0$, the arrival represents the direct P-wave, and $k=1$ is the Ps arrival.

$s_e(t - t_k)$ is the earthquake time source function.

t_k is the arrival time of k -th arrival.

z'_k , r'_k and t'_k are the vertical, radial and transversal amplitude of the normalized k -th arrival.

Seismic waveforms contain all the information recorded along the path, from source to receiver. To obtain the true structure response beneath the receiver is necessary rewrite the equations (2), (3) and (4):

$$z(t) = s_e(t) * i(t) * e_z(t) \quad \dots\dots\dots (5)$$

$$r(t) = s_e(t) * i(t) * e_r(t) \quad \dots\dots\dots (6)$$

$$t(t) = s_e(t) * i(t) * e_t(t) \quad \dots\dots\dots (7)$$

where

* indicates the convolution operator

$s_e(t)$ is the earthquake source time function;

$i(t)$ is the seismometer (instrument) response, and

$e_z(t)$, $e_r(t)$ and $e_t(t)$ represent the earth response for vertical, radial and transversal components.

From theoretical calculations for typical crustal structures it was concluded that crustal reverberations and converted phases appear predominantly on the vertical component of motion, even if a high-contrast interface is allowed to dip (Langston, 1979). Langston (1979) suggested that the vertical component of motion $z(t)$ behaves like a Dirac delta function, hence.

$$e_z(t) \sim \delta(t) \quad \dots\dots\dots (8)$$

$$z(t) \simeq s(t) * i(t) \quad \dots\dots\dots (9)$$

Equation (5) is combined with conditions (8) and (9) and equations (6) and (7) are rewritten in the frequency domain using the Fourier Transform (Langston, 1977 and 1979; Ammon, 1991; Langston and Hammer, 2001), remembering that multiplication in the frequency domain is the equivalent operation to convolution in the time domain:

$$Z(\omega) = S(\omega)I(\omega) \quad \dots\dots\dots (10)$$

$$R(\omega) = S(\omega)I(\omega)E_r(\omega) \dots\dots\dots (11)$$

$$T(\omega) = S(\omega)I(\omega)E_t(\omega) \dots\dots\dots (12)$$

where

ω is the angular frequency, $\omega = 2\pi f$;

$Z(\omega)$, $R(\omega)$ and $T(\omega)$ are the Fourier Transforms of the vertical, radial and transverse components of motion, $z(t)$, $r(t)$ and $t(t)$;

$I(\omega)$ is the Fourier transform of the instrument response $i(t)$; and

$E_z(\omega)$, $E_r(\omega)$ and $E_t(\omega)$ represent the Fourier Transform for the vertical, radial and transverse components of earth motion $e_z(t)$, $e_r(t)$ and $e_t(t)$, respectively;

$S(\omega)$ is the Fourier transform of the source time function $s(t)$.

The source equalization is defined by:

$$H(\omega) = \frac{R(\omega)}{Z(\omega)} = \frac{S(\omega)I(\omega)E_r(\omega)}{S(\omega)I(\omega)} \dots\dots\dots (13)$$

where $H(\omega)$ is the ratio of the Fourier transform spectra of the radial $R(\omega)$ and the vertical response $Z(\omega)$ (Ammon, 1991). The Receiver Function $h(t)$ is obtained by calculating the inverse Fourier transform of $H(\omega)$.

The deconvolution is unstable because the signal is band-limited and is contaminated by random noise. A Gaussian filter is introduced to limit the final frequency by removing the high frequency noise, while a zero phase low-pass filter helps to remove side lobes on the data (Langston, 1979).

$$H(\omega) = \frac{R(\omega)Z^*(\omega)}{\phi(\omega)} G(\omega) \dots\dots\dots (14)$$

The Receiver Functions with high frequency are sensitive to small-scale heterogeneities, while lower frequency Receiver Functions help to image large-scale structures (Tang et al., 2016). The Gaussian filter lowers the frequency content of the signal, decreasing the frequencies caused by scattering of small-scale heterogeneities (Lodge and Helffrich, 2009).

$$G(\omega) = \xi e^{\frac{-\omega^2}{4\alpha^2}} \dots\dots\dots (15)$$

The Gaussian filter is normalized to unit amplitude in the time domain by the constant ξ and α is the width of the Gaussian filter (Ammon, 1991). The function $\phi(\omega)$ is the autocorrelation of $R(\omega)$ with any spectral troughs filled to a level (known as the “water level”) depending on the parameter c (Langston, 1979). To perform suitable deconvolution, the water level parameter is allowed to vary to find a minimum value that can give a stable deconvolution. Ammon (1991) stated that the cost of stabilizing the deconvolution in an approximation to the ideal source equalization and filling the spectral holes is a change in the amplitude of the Receiver Function. In the original study by Langston (1979), the Gaussian filter and parameter c were set to the following values 0.07 and 0.01, respectively.

The function $\phi(\omega)$ is given by

$$\phi(\omega) = \max\{Z(\omega)Z^*(\omega), c * \max[Z(\omega)Z^*(\omega)]\} \dots\dots\dots (15)$$

where

$Z^*(\omega)$ is the complex conjugate of radial Fourier transform $Z(\omega)$ and c is the water level parameter.

3.3. S-wave Receiver Function technique

In the PRF, the converted phase P_s arrives after the direct P-phase and before multiple phases, while in the SRF the converted phase S_p is a precursor phase followed by the direct S-phase and multiple phases in S waveform coda (Hansen et al., 2007; Kind et al., 2012; Kumar et al., 2006 and Wilson et al., 2006). Due to the velocity differences between P (faster) and S (slower) waves, the S_p -phase arrives first because it travels as a P-wave through the crust, while the direct S-wave arrives later because its slower velocity.

In the PRF technique, the multiple phases arising from an intracrustal discontinuity arrive a few seconds after the arrival of the direct P phase, about the same time as the converted P_s waveform. These multiples (or reverberations) are caused by the layered crustal structure underneath the receiver. The multiple phases can mask the converted phase discontinuities

arising from the Moho (crustal–mantle discontinuity) and Lithosphere–Asthenosphere Boundary (Vinnik et al, 2004, 2005; Wilson et al., 2006). According to Zheng et al. (2005), thick sedimentary structures produce reverberations that deteriorate the crustal and upper mantle image. Therefore, the PRF results on sedimentary terrains are often unreliable due to difficulties in clearly identifying converted phases produced by discontinuities.

The S-wave Receiver Function (SRF) technique solves the reverberation problem by isolating the Sp (S- to P-wave) conversion phase in the S waveform. The SRF has less contamination from first-order multiples, and arrivals produced by conversions at discontinuities can be seen with less difficulty. In the final SRF waveform, the direct S phase is aligned with the zero time offset. The multiples are imaged on the negative side of the zero time offset, while converted phase are imaged on the positive side of zero offset. However, the S-waves are characterized by lower frequencies than P-waves, resulting in lower spatial resolution (Yuan et al., 2006). This is one of the weaknesses of the S-wave Receiver Function method. Furthermore, the SRF epicentral distance has a more limited range than the PRF. The SRFs contain the conversion at the Moho, Lithosphere–Asthenosphere boundary, and the 410 and 660 km discontinuities, but these conversions are not imaged at all epicentral distances. The distances vary for each discontinuity. The Moho is detected between 55 – 95 degrees; 55 – 80 degrees for the Lithosphere–Asthenosphere boundary; 80 – 95 degrees for the 410 km discontinuity; and 95 – 100 degrees for the 660 km discontinuity (Yuan et al., 2006).

To generate an SRF, the teleseismic events are selected according to pre-defined parameters (magnitude and epicentral distance), the recorded waveforms are rotated from the Z–N–E to the Z–R–T (or L–R–T) coordinate system using each event’s back azimuth. The (R) Radial component of motion of each event is deconvolved by the (Z) vertical component in the time domain. The Receiver Function amplitude and time axis are reversed to make the SRF resemble a PRF (Hansen et al., 2010 and references within; Kind et al., 2012; Tugume et al., 2013; Wilson et al., 2006).

3.3.1. Event selection

The teleseismic events selected for SRF analysis have magnitude ≥ 5.5 and epicentral distances between 60 degrees and 82 degrees (Appendix B). The teleseismic events travel long distances. To avoid records with attenuated and weak signal amplitudes, I selected only teleseismic events

with magnitudes ≥ 5.5 . The epicentral distance range of 60 to 82 degrees was selected to avoid interference between S and SKS phase arrival times and phase triplication. Rays closer than 60 degrees can be triplicated. Triplication occurs when three rays from the same source but with different paths and travel times emerge at the same epicentral distance. Beyond 82 degrees there is a contamination of the SV component waveform caused by the dual presence of S- and SKS-phases at same epicentral distance (Fig. 8). For the SRF technique using the Sp converted phase, epicentral distances between 60 and 82 degrees are suitable, while for the SKSp conversion a wider range is needed, i.e. above 82 degrees.

The data from the XV, YH, and ZP networks used to generate SRFs in this study were requested from IRIS using software Jweed 4.1. Only SRFs with a clear converted phase discontinuity arrival time are used in further processing steps. In Appendix B, a summary of all selected event parameters (date, location, event time, depth and magnitude) is presented. Using event information from Appendix B, the epicentral map (Fig. 9) was plotted using only the events with good S-wave receiver functions.

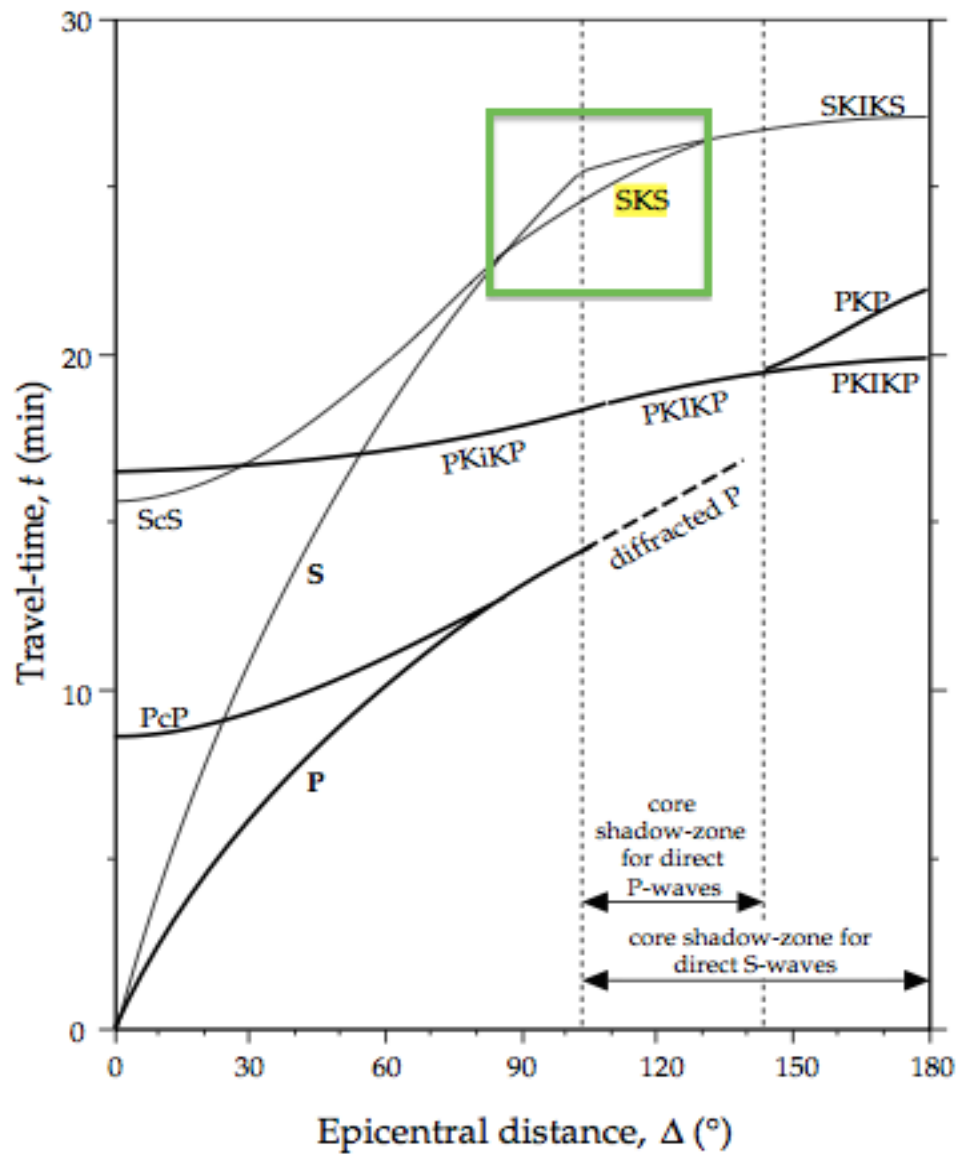


Figure 8 – Travel time curves for epicentral distances from 0 to 180 degrees, showing the P and S wave shadow zones and the interference region between the S and SKS phases in the green square (Lowrie, 2007).

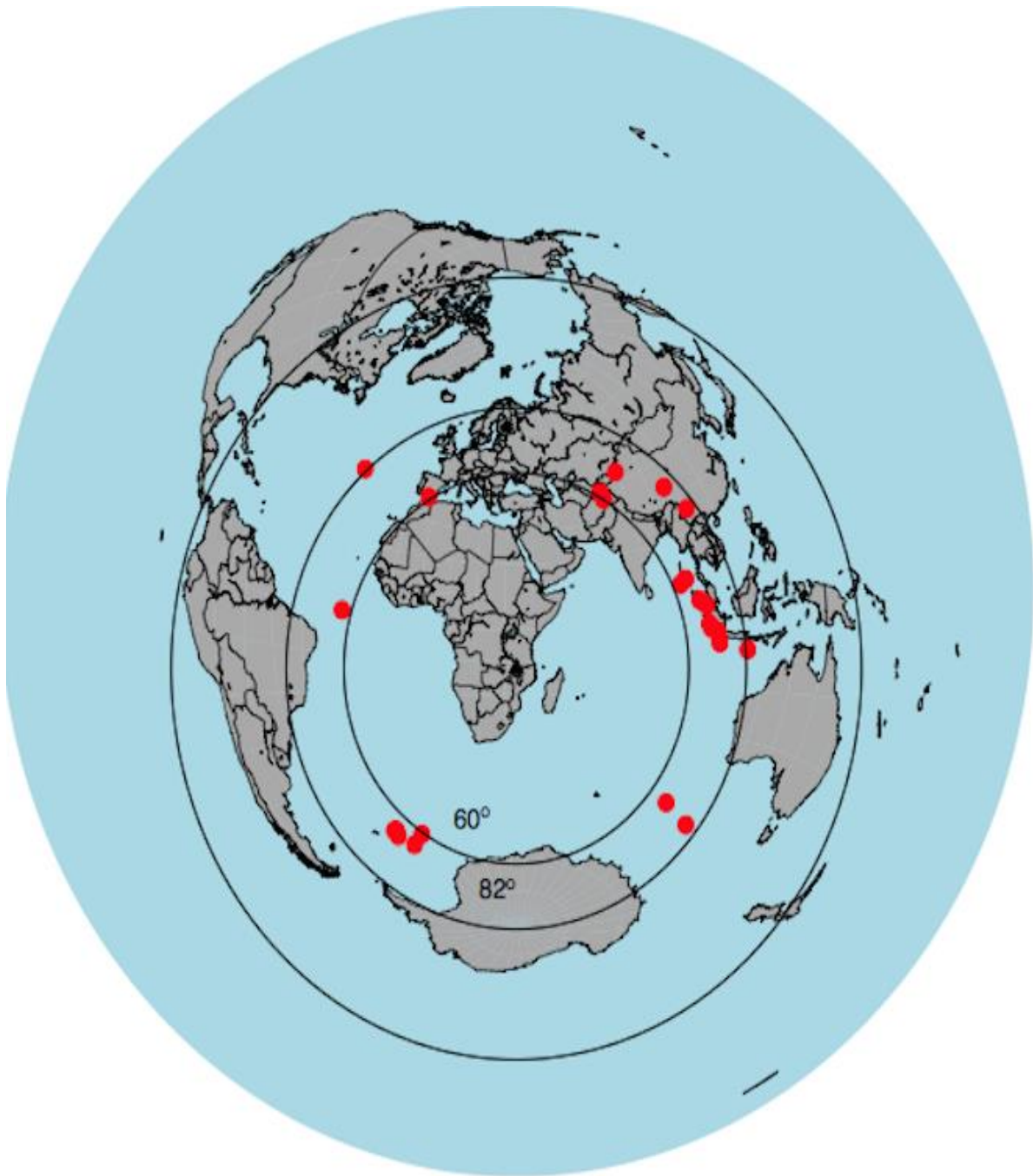


Figure 9 – Map showing location of events used to generate SRFs and centered on the middle of seismic stations used to collect data at 60 to 82 degrees of epicentral distance.

3.3.2. Isolation of the converted phase

Typically a seismic station records three components of ground motion: vertical (Z), East-West (E), and North-South (N) (Fig. 10). The seismograms contain information about the earthquake source, the path through the mantle and crust, as well as the response of the instrument.

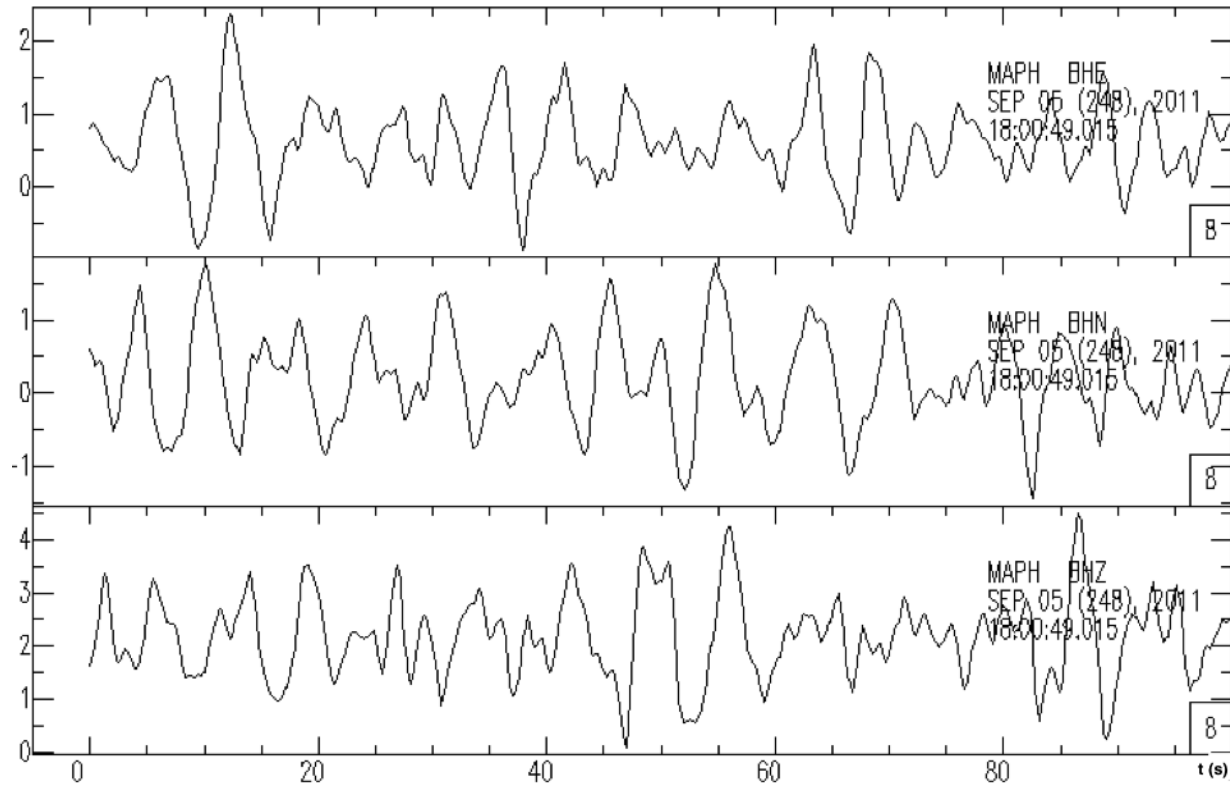


Figure 10 – Example of 3-component seismograms at MAPH seismic station. BHE, BHN and BHZ refer to the East-West, North-South and vertical component, respectively.

The receiver function technique attempts to isolate the converted phases. Seismograms must be rotated to enhance the converted phases. These rotations depend on the incident angle of the incoming wave toward the receiver (Fig. 11).

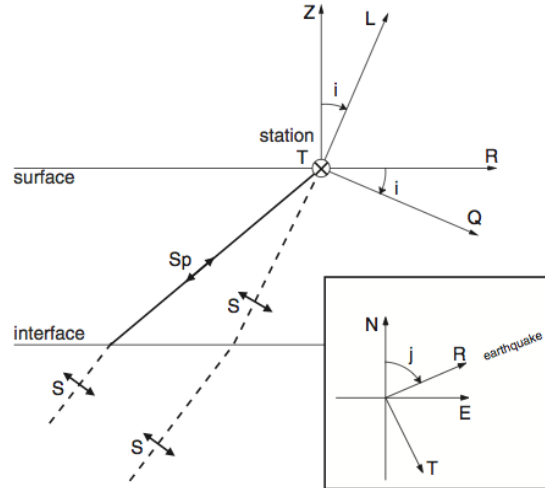


Figure 11 – Cartoon showing the rotated components using incident angle (i) and ground motion of the incoming converted S-waves ray (continuous line) and direct S-wave ray that is free of conversion (dashed line). Inside the black square, the back azimuth angle projection (j) and R points towards the earthquake epicenter (Kind et al., 2012).

Before coordinate system rotation, the seismograms are in the Z (vertical) – N (North–South) – E (East–West) coordinate system. The P-wave energy is dominant in the vertical component of the seismogram, while the S-wave dominates the horizontal component. Aiming to enhance the converted signal amplitude, the rotation is executed using the back azimuth. The back azimuth can be calculated if the hypocenter and receiver coordinates are known (Soudoudi, 2005). The ZRT (or P–SV–SH) coordinate system rotation is executed when R is computed using the theoretical back azimuth (Kind et al., 2012).

The SRF waveform has weak Moho and LAB conversion (Fig. 12 – a). To enhance these conversions coordinate system rotations are carried out. Two coordinate system rotations are carried out: The first rotation (to ZRT) is carried out to enhance the signal from the crust–mantle discontinuity (the Moho), while the second rotation (to LQT) is carried out to enhance the signal from the Lithosphere–Asthenosphere discontinuity. In this section, the arrival times are presented with negative values because the coordinate rotation is executed before the SRF is deconvolved and the amplitude and time reversed. This is contrary to the final results where the arrival times are positive due to reverse of time and amplitude.

To enhance the signal, rotations are executed and coordinate systems are changed.

- i. As a result of the first rotation, the R and T components contain energy from N and E components and enhance the signal amplitude of phases arriving in the first 10s, which is the range where the expected Moho conversion arrival time should fall (Fig. 12–a). The expected Sp Moho conversion arrival time is between – 3 and – 6s before the Ss-phase.
- ii. As a result of the second rotation, the L component contains P–wave energy and the Q component contains SV energy. This rotation smooths the amplitude signal in the first 10s and enhances the converted phase signal amplitude between –10 and –20 s, which is a range of expected arrival time for Lithosphere–Asthenosphere discontinuity converted phase (Fig. 12–b).

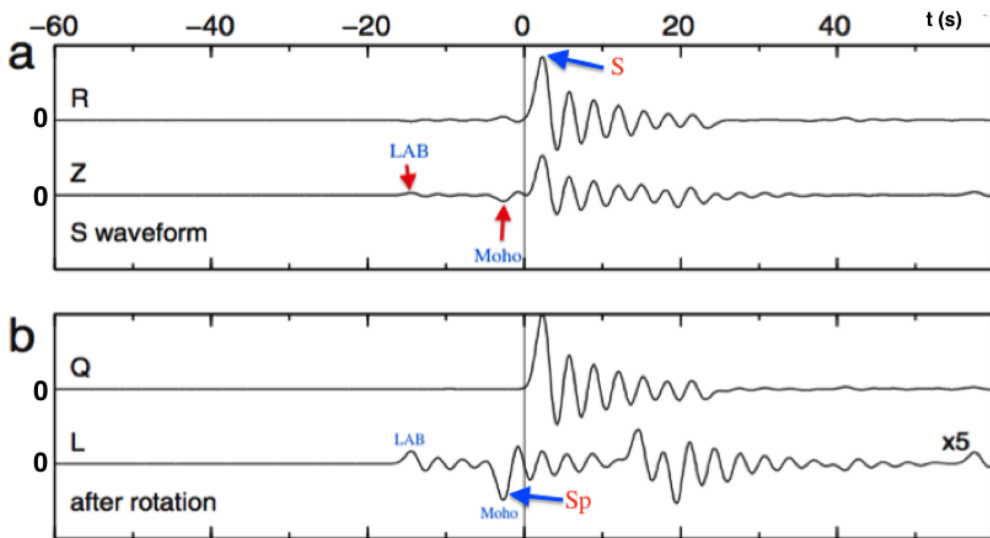


Figure 12 – S-wave waveforms: (a) Result of waveform first rotation from ZNE to ZRT coordinate system while (b) result of second rotation from ZRT to LQT coordinate system (Kind et al., 2012).

After the first rotation of the 3-component seismogram, from ZNE to ZRT coordinate system, the converted waves are polarized and appear on the Z and R components, while the T component contains little energy. The polarization of the body waves (i.e. preferential direction of the

motion of the seismic wave passage) causes the Ps wave energy to appear on the R (radial) and T (transversal) components, and the Sp wave energy to appear on the Z component.

The seismograms were rotated only once because the first rotation enhances the crust–mantle discontinuity, which is the focus in this research.

The S-waveform contains a wealth of information about the Earth's interior underneath a particular receiver. The information is a combination of different contributions, such as the source function, the instrument response, the upper mantle response and the response of the site in the vicinity of the receiver (Kind et al., 2012; Kumar et al., 2006; Midzi and Ottemöller, 2001). The receiver functions must carry only the crustal and upper mantle response. Therefore, the S-waveform is deconvolved to remove effects of the source, the far-field path and instrument, leaving only the direct S and the Sp phase conversion.

3.3.3. Deconvolution

Deconvolution is an inverse filter used to remove the effects of other signal processing filters or the instrument response. It restores the wave shape into the initial form as before the filtering process.

Deconvolution can be executed in either the frequency or the time domain. Deconvolution in time domain is more common than in the frequency domain because it is not easy to find a stable result of division between vertical (Z) and radial (R) components in the frequency domain (Kind et al., 2012). Ligorria and Ammon (1999) introduced an iterative time domain deconvolution approach. This approach constrains the spectral shape at longer periods and improves the extraction of information from the original signal.

The S waveforms are cut 10s after and 100s before the direct S-phase arrival. Kind et al. (2012) stated this window should be used because deconvolution in the time domain needs a certain length for stability reasons.

In SRF processing, the radial component (R) is deconvolved from the correspondent vertical component (Z) in the time domain using 500 iterations (Ligorria and Ammon, 1999). The frequency content of the SRF is controlled using the Gaussian weight parameter. The Gaussian

weight (a) used was $a = 1.0$. According to Hansen et al. (2009a,b), stable and reliable results are found using $a = 1$.

For the PRF, the radial and the transversal receiver functions are generated because the S phases are polarized in SH and SV components. The quality of the PRF is evaluated using the least mean squares minimization to measure the misfit between the original and generated waveform. The misfit is the percentage of recovery of original radial component when compared with generated radial receiver function. It is measured using the difference between the radial component seismogram and the convolution of the vertical component seismogram with the radial receiver function (Tokam et al., 2010; Tugume et al., 2012).

The same procedure is used for SRFs. The least mean squares minimization measures the misfit between original and predicted waveform. Different to the PRF technique, where the radial and transverse receiver functions are generated, in the SRF technique the vertical receiver function is generated. The misfit is measured using the difference between the vertical component of the seismogram and the convolution of the radial component seismogram with the vertical receiver function. A misfit criterion is used to select receiver functions. If the fit of any receiver function is $< 75\%$, then this receiver function is discarded. The selected SRFs were used in further analysis.

The source function and instrument response are removed using deconvolution, leaving only the response of crustal and upper mantle structure in the SRF signal. As a result of deconvolution, the SRF waveforms amplitudes from different sources are equalized so that it appears that all SRFs have same source. The converted phase S_p usually has a higher frequency than the direct S phase. Chopra et al. (2010) suggested that the frequency difference is due to variations of attenuation effects between P- and S-waves in unconsolidated basins.

The arrival time of the direct S-phase signal during the rotation is taken as the zero offset. The resultant SRF has the converted phases on the negative side of zero time offset, while the multiples phases are on the positive. In order to make a SRF resemble PRF, the time and amplitude axis are reversed. As a result of this axis reversal, the converted phases produced by the discontinuities (Moho and Lithosphere–Asthenosphere Boundary) appear on the positive side of the zero time offset while the multiples on the negative side (Fig. 13 – center).

The PRF has converted phases and multiples on the same side of time axis and it is not easy to identify the converted phase of LAB, while the SRF is free from contamination by multiples and the S-phase is at zero offset (Fig 13).

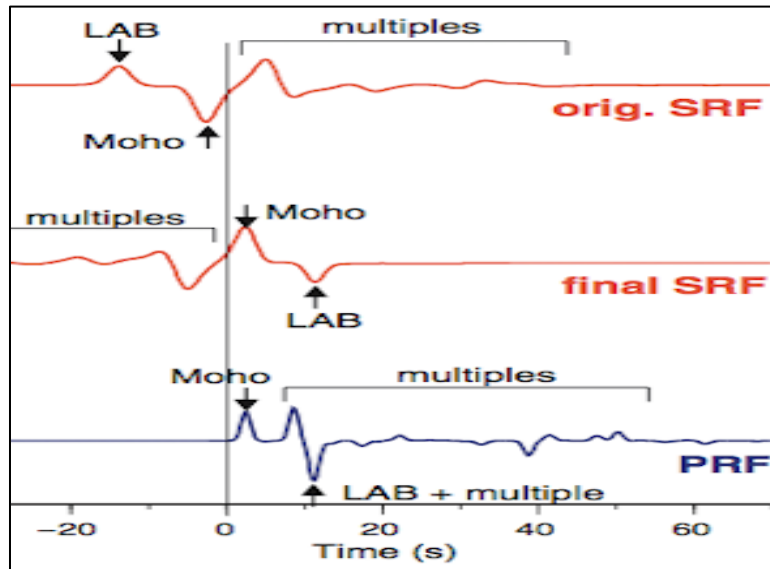


Figure 13 – Image showing the original SRF with the LAB and Moho converted phases on the negative side of the zero time offset (top). To make the SRF resemble a PRF (bottom), the time and amplitude are reversed (center) (Kind et al., 2012).

SRFs were generated for 132 teleseismic earthquakes recorded at Butiana (BUTI) station. In Fig. 14, nine SRFs are displayed showing examples of good and bad SRFs. The good SRFs (5, 6, 7, 8, 9) have the first arrival (direct S-wave) close to the zero offset (5, 9) or at the zero offset (6, 7, 8); the second arrival, the Moho conversion Sp, amplitude is smaller than the first arrival and time window range is between 3 and 6 seconds. In the bad SRFs (1, 4), the first arrival (direct S-wave) has a lower amplitude than the second arrival, the Moho conversion Sp phase. SRFs (1, 3) have the Moho conversion Sp with negative amplitude. SRF (2) has the first arrival far from the zero offset, while SRF (4) does not show a clear first arrival (direct S-wave).

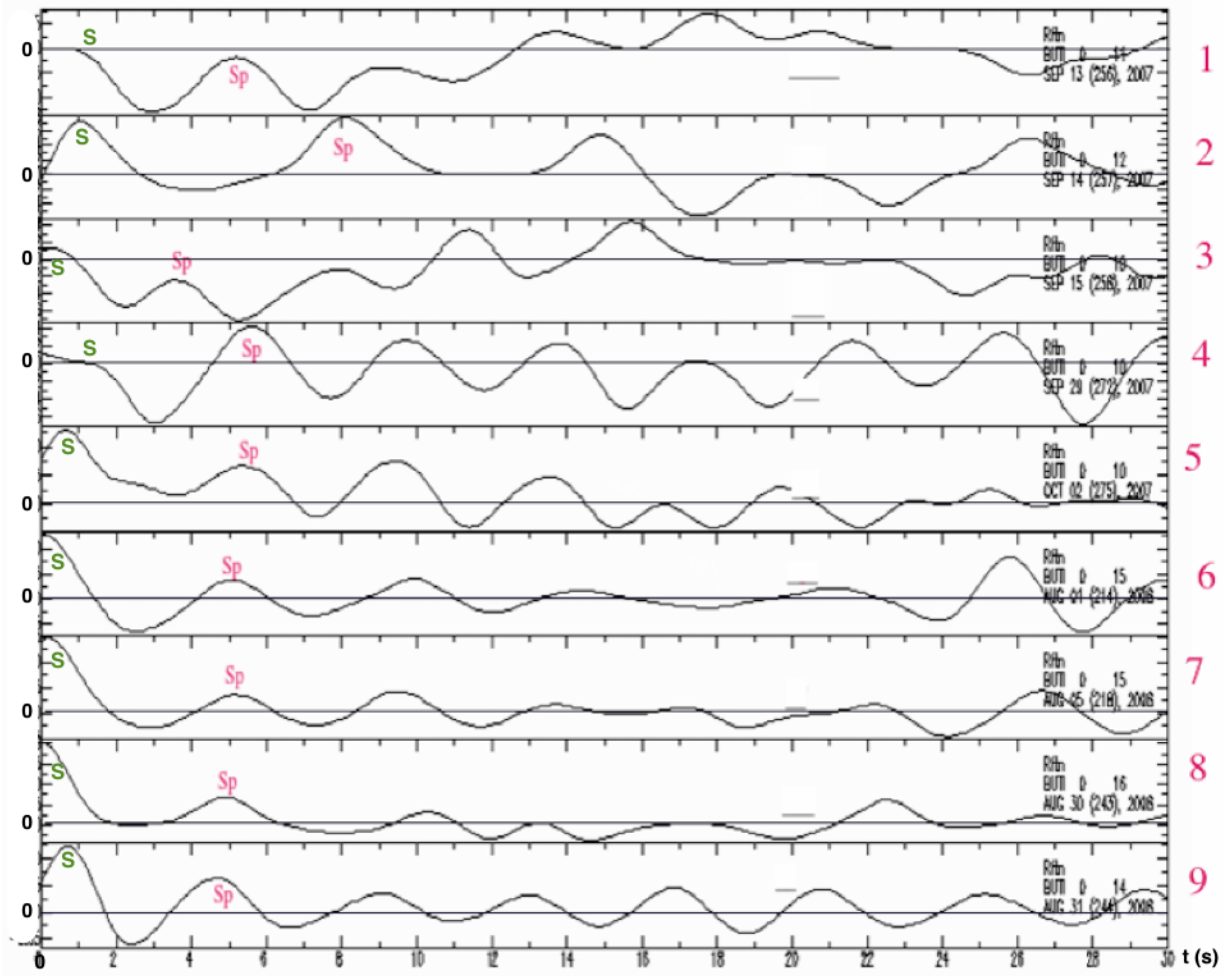


Figure 14 – Sample of SRFs generated from BUTI station data.

In the SRF, the positive amplitudes indicate the velocity increases with depth, while the negative amplitudes mean that velocity decreases with depth. The Moho conversion Sp is expected to have a positive amplitude due to the abrupt increase of velocity at the Moho discontinuity (Fig. 15).

Several good SRFs were generated using data recorded in the Lake Albert sedimentary basin (Fig. 15). In the SRF, the first spike close to the zero time offset is the direct S phase. The second spike represents the Sp Moho conversion. The arrival time range of the Sp Moho conversion is between 3 and 6 s. The time arrival is less than 4 s for thin crustal structures and above 5 s for thicker crustal structures.

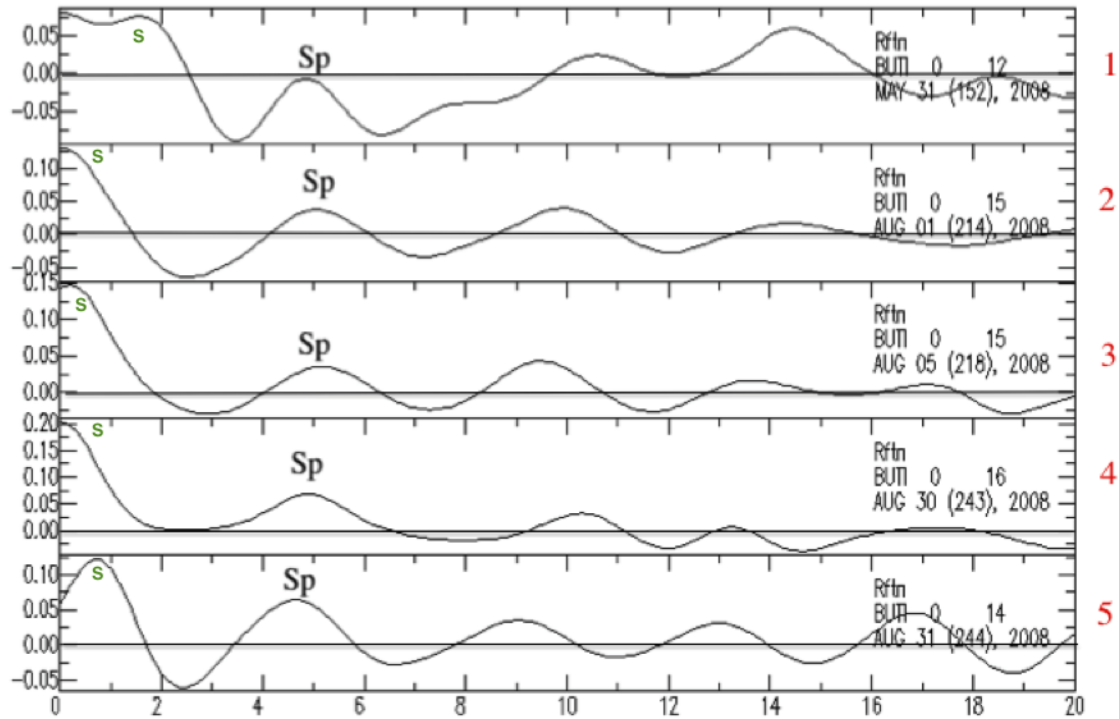


Figure 15 – SRF generated from BUTI station data using Gaussian weight $a=1$. In the five SRFs displayed, the Moho conversion is clearly seen.

3.4. Moveout correction and stacking

In the SRF waveform, the converted phases have a low amplitude and weak signal. Stacking the SRFs obtained from different events can increase the signal-to-noise ratio. However, a moveout correction that aligns all traces must first be made.

3.4.1. Moveout correction

The conversions from different discontinuities and epicentral distances travel with different energy and traces are not aligned due to velocity variation (Fig. 16). To align all traces for each discontinuity, a moveout correction is applied.

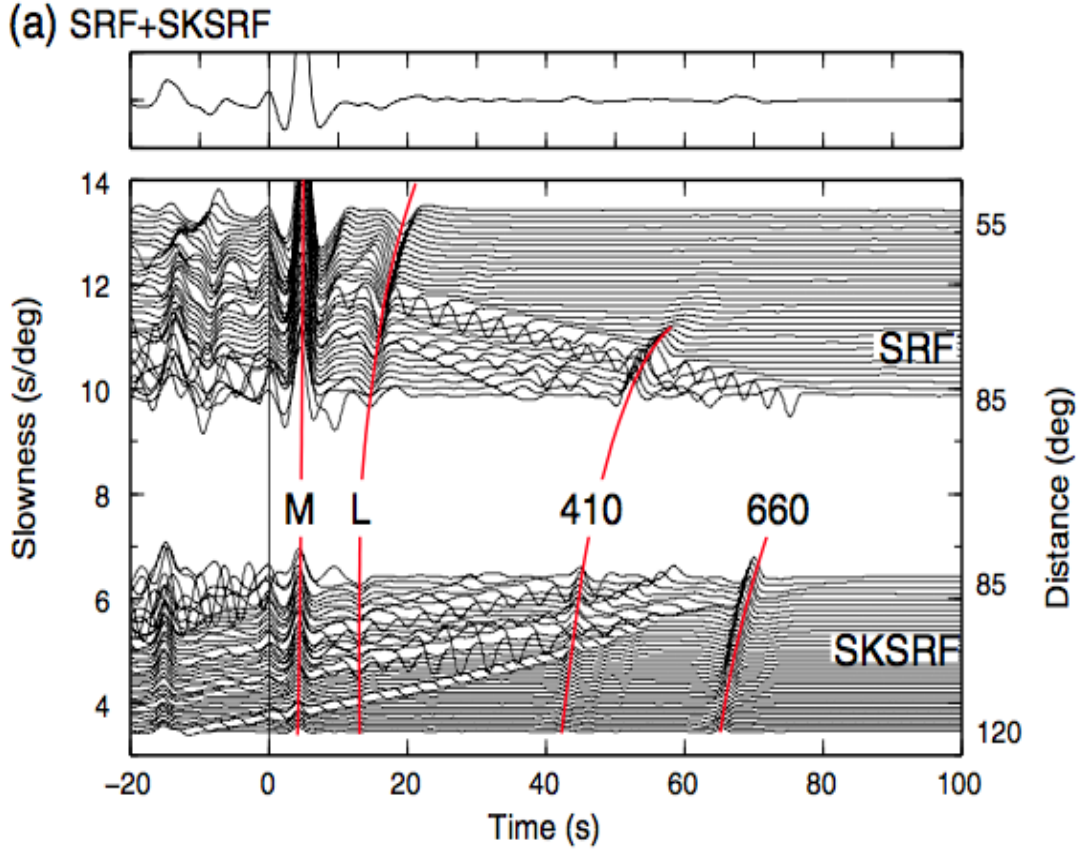


Figure 16 – Receiver function computed for S- and SKS-phase before a moveout correction, hence arrivals are not aligned (Kind et al., 2012).

The moveout is the difference between the travel time arrivals from two offsets separated by certain distance (Kearey et al., 2002; Kind et al., 2012). The moveout correction aligns the traces by taking the differences in the ray paths into account.

In SRF analysis, the arrival time delay considered is between the direct S-wave and Sp phase recorded by the receiver.

$$t_{s_p} - t_s = h \left\{ (v_p^{-2} - s^2)^{1/2} - (v_s^{-2} - s^2)^{1/2} \right\}$$

where

t_{s_p} is the Sp conversion arrival time;

t_s is the direct S-wave arrival time;

the h is the thickness; and

s is the ray parameter.

All traces are displayed as time functions because arrival times of different traces vary as a function of offsets. When all traces are displayed, the direct S-wave, the converted phases and multiples are not aligned due to variations in offset (Stein and Wysession, 2013). The trace is time shifted using the calculated moveout correction. The arrivals now appear at same time for all offsets (Fig. 17).

(b) SRF+SKSRF after moveout correction

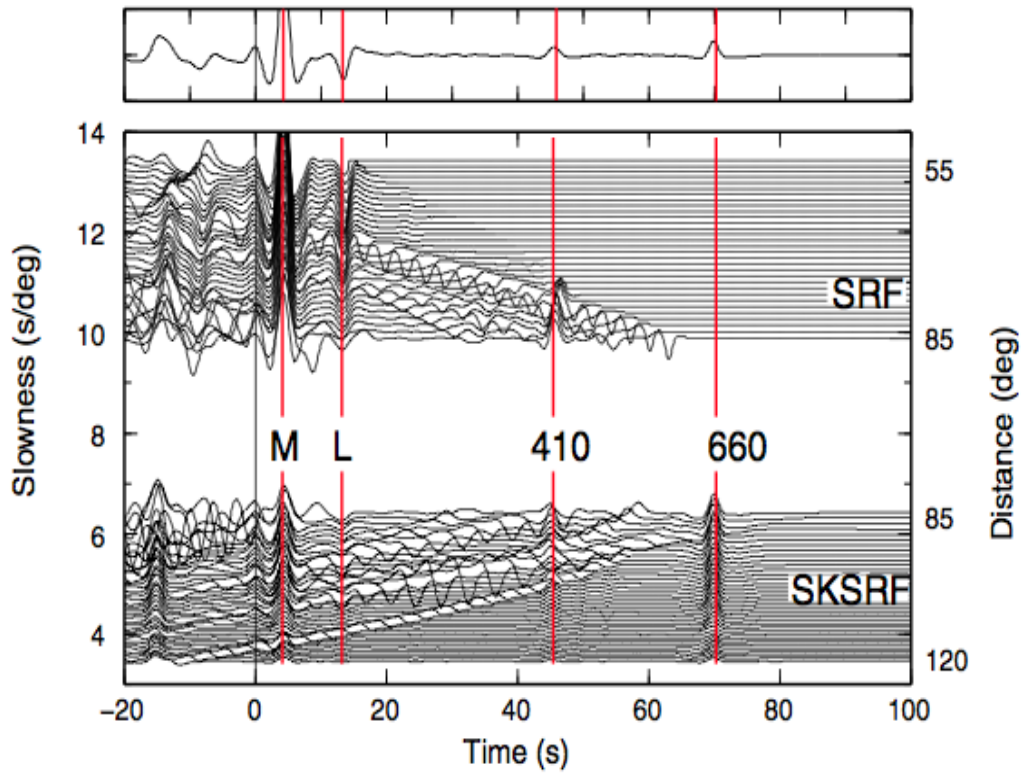


Figure 17 – After moveout correction, the arrivals from each discontinuity are approximately parallel to the zero time offset (Kind et al., 2012).

In moveout correction processing it is assumed that the epicentral distances are known. The moveout changes according to the depth of the discontinuity being investigated e.g. the Moho moveout is less than 6 s, while it is 8 s to 10 s for depths of 440 km and 660 km, respectively (Kind et al., 2012). We used a theoretical slowness of 6.4 s/degree to execute the moveout correction. As result of moveout correction all traces, whether from different or identical sources, are lined up and the converted phases appear in the same time window.

3.4.2. Stacking

Stacking is an important processing step and is a method that sums traces to produce one trace. It reduces random noise and enhances the signal-to-noise ratio (Lay and Wallace, 1995; Mohsen et al., 2006; Hansen et al., 2009a,b). It enhances the conversions due to constructive summation of the Moho conversion because the pre-stack Moho conversion signal is usually small. Bock and Kind (1991) confirmed that stacking many traces enhances the appearance of converted phases.

All traces per station were displayed using Adobe Illustrator software and randomly colored to pick the traces more easily (Fig. 18). All traces with a clear Sp Moho conversion arrival time were picked and the others were discarded using the following criteria:

- i. Traces with either late or early arrivals. The expected range of Moho time arrival is between 3 and 6 s. Arrival times out of this range were discarded;
- ii. Traces with higher amplitude not close to zero offset were discarded because the Moho conversion can fall away from expected arrival time range; and
- iii. Traces with too high or too small Moho conversion amplitude, as this indicates an unrealistic velocity contrast.

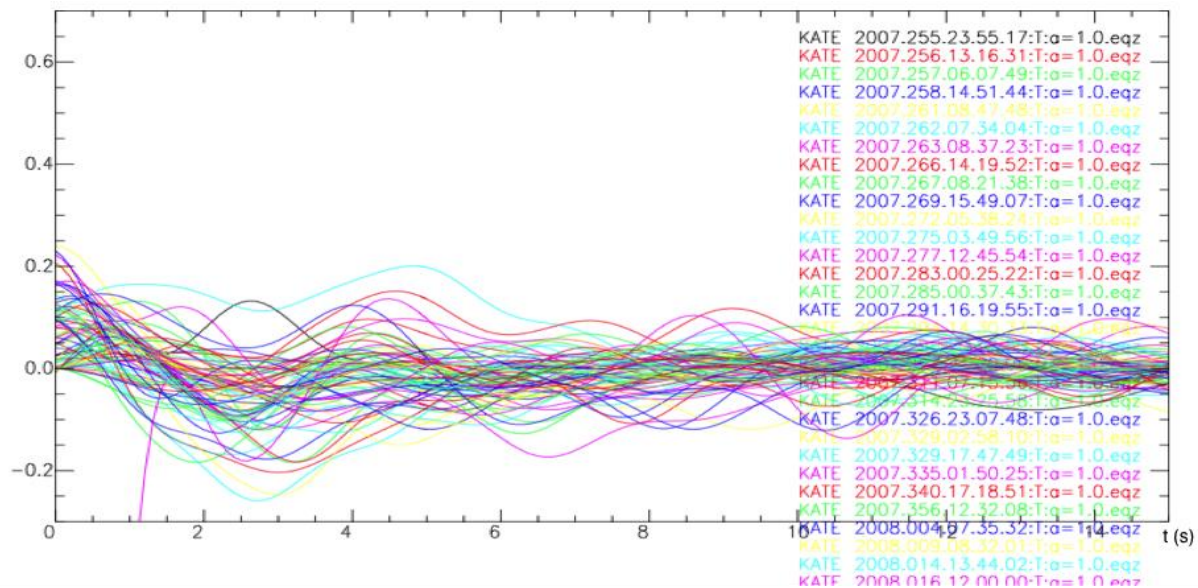


Figure 18 – Plot of generated SRFs from KATE station data, colored for easy manual picking of converted phases and event identification.

Only SRFs that display a clear Moho conversion within the expected arrival time range were selected for further analyses (Fig. 19). The selected SRFs are displayed in different colors and have different amplitude. It is clearly seen that the Sp Moho conversion arrival time range between 3 s and 6 s (Fig. 19 – left). We summed the amplitudes of all Sp conversions (Moho conversion) and displayed the result as a single trace (Fig. 19 – right). We then extracted the time arrival per station using the method of Owens et al. (2000). The stacks were computed using stacking codes in MatLab and the Sp conversion arrival time was extracted. In Table 1, the detailed Moho conversion arrival time for each station is presented, and plots of stacking results are shown in Appendix C. From the resultant trace, it is easy to estimate the exact Sp conversion arrival time, but not the range. The arrival time was then used in a grid search for Moho depth and crustal velocity.

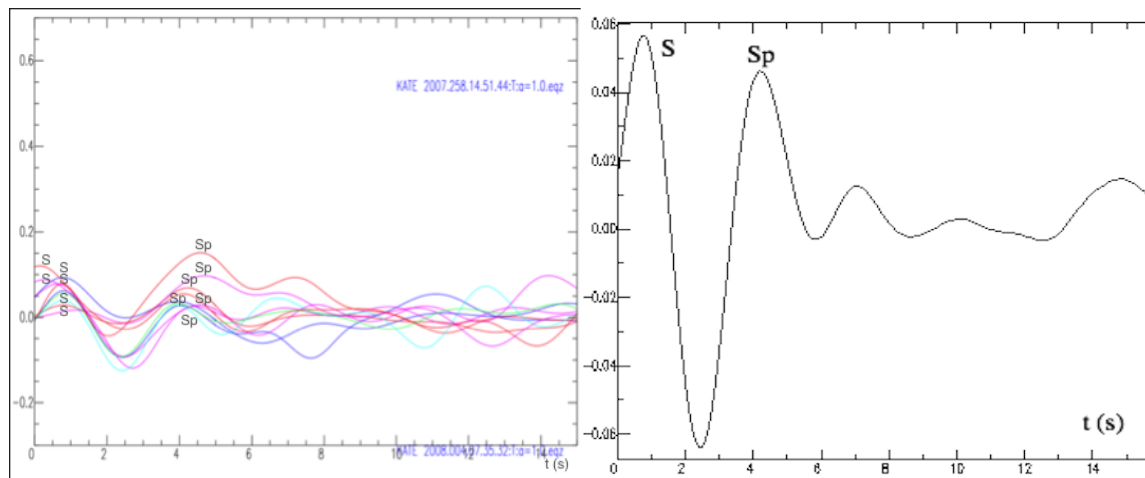


Figure 19 – Stacking of SRF signals recorded at the KATE station Left: traces that satisfied the selection criteria. Right: stacked signal showing an improvement in the signal-to-noise ratio of the Sp phase and the arrival time of the Sp conversion (ca. 4.2 s).

H-k stacking is frequently used in Receiver Function studies. The H-k stacking approach constructs a stacking surface by adding receiver function amplitudes along converted travel-time curves for a range of crustal models that are assumed to consist of a single layer over a half-space (Tang et al., 2016). However, we did not use H-k stacking because our model consisted of two layers over a half-space so that the sedimentary layer could be taken into account. The model parameters were systematically varied to search for the optimum solution.

Table 1 – Summary of data used to estimate Moho arrival time.

Station Name	Number of events recorded	Number of SRFs used for stacking	Moho time arrival (s)
BUTI	132	8	5.2
CHAL	33	1	5.1
INDI	32	5	4.0
KATE	132	10	4.2
KIMA	69	5	5.1
KYLA	104	3	4.0
MANG	32	3	5.4
MAPH	105	6	4.0
MOHO	32	4	5.2
MTWA	59	4	4.5
MSGR	107	3	4.9

3.5. Bootstrapping

The measurement of statistical confidence intervals has become one of the methods to assess data quality. Bootstrapping is a confidence interval measurement method that is widely used for different scientific purposes. Sacchi (1998) defined the bootstrap method as “*a computer intensive technique for assigning measures of accuracy to statistical estimates*”. In geophysics, the bootstrapping technique is used to estimate different error parameters such as magnitude uncertainties, earthquake depths and velocity errors (Sacchi, 1998).

In bootstrapping, the original data set is randomly resampled producing new data sets. The resampled data sets have same characteristics as the original data set (Efron and Tibishirabi, 1991). These new datasets are used to estimate the errors bars, standard deviation or confidence intervals for a chosen parameter without making assumptions about the statistical distributions. The error bars represent the measurement of the variations or dispersion of data using the average or mean. The number of resamples created from bootstrapping distribution is equal to the sample distribution. The inference using bootstrap distribution is equivalent to inferences about the original data set. This procedure is possible because the bootstrap standard error represents the standard deviation of the original data set.

In general, a peak in the data set is identified and used as reference to the degree of uncertainty by comparing the amplitude of the peak obtained from the various resampled data sets. The reference we used was the peak associated with the conversion arrival time (Fig. 20).

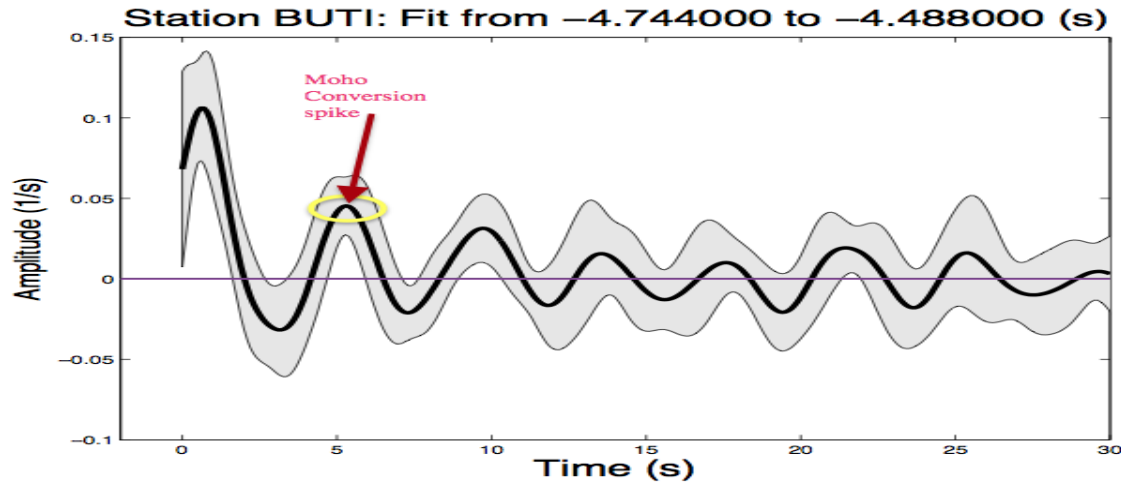


Figure 20 – The bootstrap result plotted using BUTI seismic station showing the spike or conversion time arrival point used as reference point to measure the bootstrap error.

The bootstrapping method was used for all stations to compute the bootstrapping distribution from all selected and stacked SRFs. The SRFs were randomly resampled using the Efron and Tibishirabi (1991) method with 200 bootstrap iterations. Efron and Tibishirabi (1991) discuss the selection of number bootstrap iterations and considered 200 bootstrap experiments to be adequate. The Sp Moho conversion was used as the reference peak.

The bootstrap processing follows the steps below:

- i. All the SRFs for a particular station are resampled to create a new SRF data set;
The SRF set is denoted $X = x_1, x_2, x_3, \dots, x_N$.
 x_i , represent the SRFs and N is the size of the SRF population.
- ii. The resampled SRF set is denoted $H = h_1, h_2, h_3, \dots, h_N$.
The resampled data set H is used in further analysis (Efron and Tibshirani, 1991). The resampled data is the bootstrap population. The bootstrap population size is N , and this resampled data is used instead of original data. The resampled data has the same characteristics as the original data.
- iii. The resampled SRF data set is picked using a random number generator to select the elements of the H data set;
- iv. From the resampled data, sub data sets are created. The sub data sets (Y) result from randomly picking and allow repetition of a resampled SRF in the same data set. The resampling is repeated 200 times to create two hundred different data sets;

$$Y = y_1, y_2, y_3, y_4, y_5, y_6, \dots, y_{198}, y_{199}, y_{200}.$$

- v. The distribution mean (M) and standard deviation (S) are estimated using results produced by all the sub data sets. The inference of standard deviation or parameters estimated using sub data sets will represent the parameter of the original data.

$$M = \frac{\sum y}{N} \text{ and } S^2 = \frac{\sum (y-M)^2}{N-1}$$

- vi. The time standard deviation (2σ) results are displayed as shadow regions (Fig. 21).

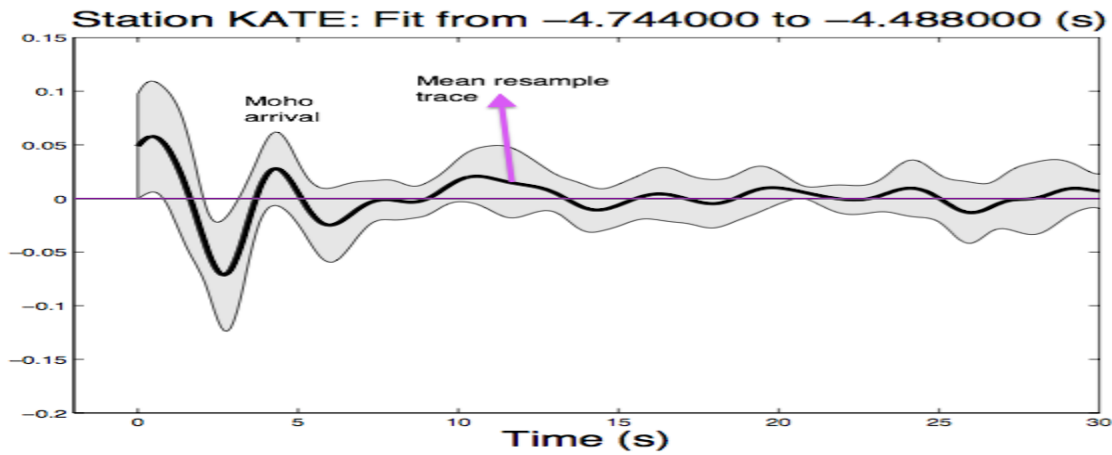


Figure 21 – The bootstrap result plotted using KATE seismic station data. The mean resample trace represents the mean value distribution, grey area covered by bootstrapping sample distribution, and Moho time arrival used as reference point to extract the error bars.

Example of bootstrapping processing using KATE station data:

- i. Original data set X , composed by 10 S-wave receiver functions.

$$X = x_1, x_2, x_3, x_4, x_5, x_6, x_7, x_8, x_9, x_{10}.$$

- ii. From original data set X , create a bootstrap data set H .

$$H = h_1, h_2, h_3, h_4, h_5, h_6, h_7, h_8, h_9, h_{10}.$$

- iii. From bootstrap data set H , create sub data sets. The sub data sets (Y) must have the same size (population), and be randomly picked allowing repetition on picking.

$$Y = y_1, y_2, y_3, y_4, y_5, y_6, \dots, y_{198}, y_{199}, y_{200}.$$

Example of randomly created sub data sets

$$y_1 = h_1, h_2, h_3, h_4, h_5, h_6, h_7, h_8, h_9, h_{10}.$$

$$y_2 = h_{10}, h_2, h_3, h_4, h_5, h_6, h_6, h_8, h_9, h_1.$$

$$y_3 = h_{10}, h_2, h_7, h_4, h_5, h_6, h_6, h_8, h_9, h_1.$$

..

$$y_{200} = h_1, h_2, h_1, h_4, h_5, h_6, h_6, h_8, h_9, h_1.$$

The distribution mean and standard deviation are calculated and displayed in plots (Fig. 22).

Only stations with coherent results of standard deviation (2σ) of bootstrapping distribution were approved. Most stations yielded a bootstrap time error of less than 0.05 s. Stations with inconsistent results were discarded. The inconsistency was caused by unclear time error standard deviation estimation at Moho spike arrival time (Fig. 22). The error was used in further processing steps to estimate uncertainties in estimates of the crustal thickness and shear velocity.

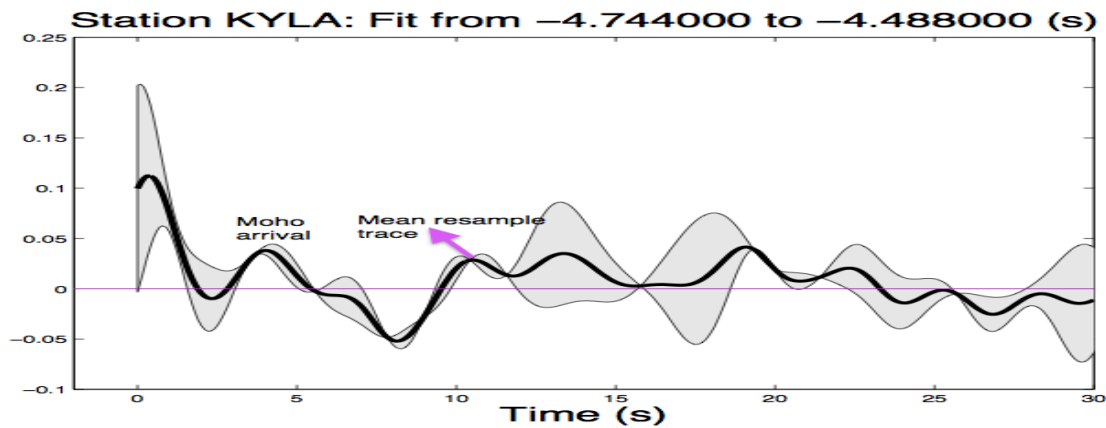


Figure 22 – The bootstrap result plotted using KYLA seismic station data showing inconsistent error bars.

3.6. Crustal thickness and shear wave velocity estimation

The crustal thickness and shear wave velocity was estimated using the Grid Search method. To estimate depth, the Sp arrival time was modeled along with the Rayleigh group velocity for each station. The dispersion curves were used to determine the average shear velocity, while the SRFs were used to determine the traveltime between the discontinuity and the surface.

3.6.1. Dispersion curves

Kearey et al. (2009) defined surface waves as *seismic waves whose travel paths are restricted to the vicinity of a free surface, such as Earth's surface*. There are two types of surface waves: *Rayleigh waves*, which exhibit a combination of P- and SV- particle motion interaction that describes a retrograde ellipse; and the *Love wave*, which exhibits horizontally-polarised (SH) particle motion (Kearey et al., 2002, 2009) (Fig. 23).

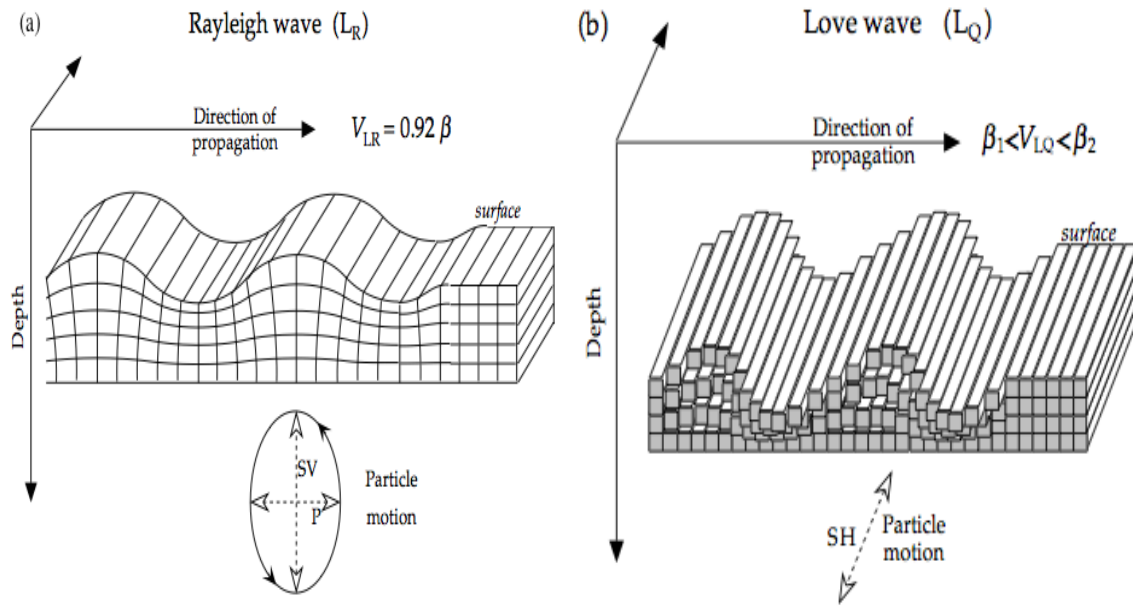


Figure 23 – Cartoon showing particle motion and propagation direction of surface waves (a) Rayleigh waves and (b) Love waves (Lowrie, 2007).

Lowrie (2007) defined dispersion as *dependence of velocity on wavelength*. Surface waves that propagate in a layered medium are dispersive. In addition to providing information about the velocity of the upper mantle and lower crust, dispersion can cause changes in waveforms. As a consequence of the dispersion, waves with different periods travel at different velocities. The velocity variation is caused by wavelength; the longer wavelength waves that travel faster and penetrate deeper than short wavelength waves (Lowrie, 2007; Stein and Wyssession, 2003). A plot of a group or phase velocity as function of frequency is called a *dispersion curve*. On a seismogram, it is clearly seen that surface waves with longer periods and higher amplitudes arrive after the body waves (Fig. 24).

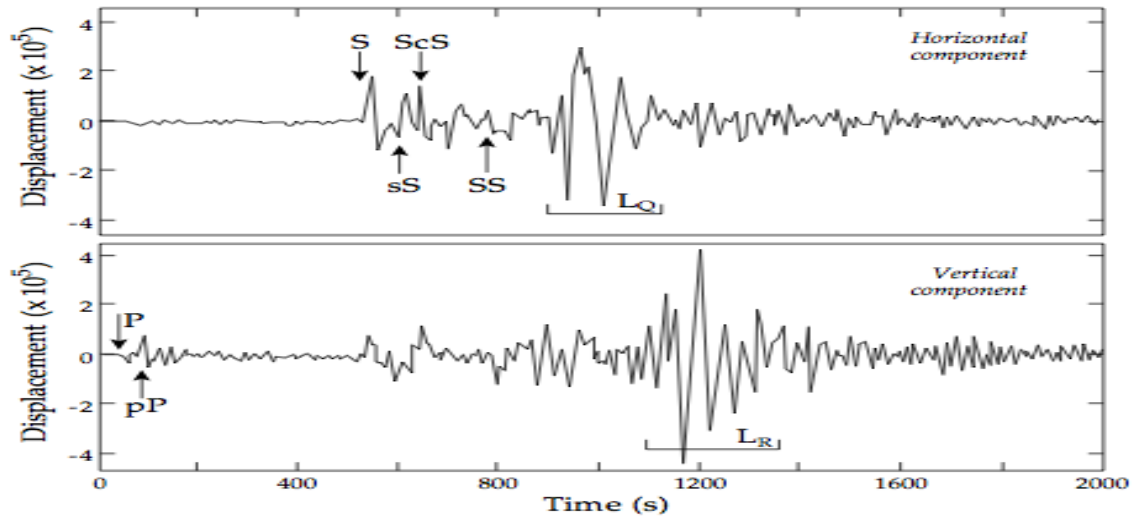


Figure 24 – Broadband seismogram of an earthquake in Peru recorded at Harvard. The Rayleigh wave (LR) recorded by the vertical component has higher amplitude than body wave phases P and pP, while the horizontal component Love wave (LQ) has a higher amplitude than body wave phases S, sS, ScS and SS (Lowrie, 2007).

An example of group velocity dispersion curves for station MSGR is shown in Fig. 25. A detailed list of periods and corresponding velocities for each station is presented at Appendix D.

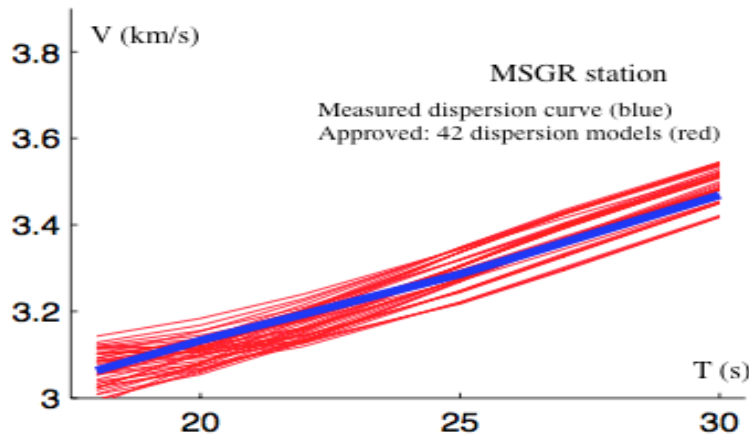


Figure 25 – A dispersion curve for MSGR station modeled using group velocity and periods. The observed dispersion curve plotted from dispersion data is shown in blue, and dispersion curves produced by models that fitted the observed dispersion curve within an acceptable range are shown in red.

The group velocities for each station at periods of 10, 15, 20, 25 and 30 seconds (Appendix E) were determined from Rayleigh waves and extracted from ambient noise by Raveloson et al. (2015). Raveloson et al. (2015) used the same dispersion curves in studies of the Congo basin.

The steps used to generate dispersion curves were:

- i. Definition of the range of parameters for the grid search. The parameters used to compute velocity–depth models are crustal density, crust and mantle velocity, Moho depth (varying from 25-50 km), shear velocity (varying from 3.4-3.9 km/s), and the Poisson’s ratio for sediments, upper mantle and crust. The parameters are shown in Table 2.
- ii. Generation of synthetic dispersion models for various velocity–depth models using the DISPER80 package (Saito, 1988). DISPER80 is a FORTRAN subroutine package used to estimate normal modes of Earth models. The Earth models are considered as non-rotating perfectly elastic transversally isotropic and laterally homogeneous (Saito, 1988).
 - a) The velocity-depth models consisted of three layers: two crustal layers and an uppermost mantle layer. The sedimentary layer thickness was allowed to vary from 0.01 to 1 km. The Moho depth window was set between 25 and 50 km, the minimum Moho depth is 25 km and the maximum Moho depth is 50 km. A 1 km increment is used for crustal structure sampling.
 - b) The shear velocity window was set between 3.4 and 3.9 km/s. The minimum crust velocity is 3.4 km/s and the maximum crust velocity is 3.9 km/s. The models were created using the minimum velocity 3.4 km/s as a starting point for sampling and using 0.05 km/s increment for the crustal shear velocity.
 - c) Starting from minimum velocity (minimum crust velocity = 3.4 km/s), DISPER80 created a velocity–depth models sampling all Moho depths between 25 and 50 km. Then the intermediate loop is closed.
 - d) The minimum crust velocity (3.4 km/s) is incremented 0.05 km/s, and the sampling is executed as previously explained at c).
 - e) The procedures explained at c) and d) are executed until the maximum velocity crust, 3.9 km/s, is achieved. Then the main loop is closed. More details of the scripts used are shown in Appendix F.

- iii. The synthetic dispersion models are compared to the measured group velocity models. The group velocity dispersion produced by each model is compared with the measured group velocity for all periods. If the values fall out of the velocity range for each period, it is considered a bad model and therefore discarded. The models that fall within the bootstrap time error (0.05 s) are approved and used in further analysis. The selection of synthetic models that fit the measured dispersion curves is made using the grid search method. Approved dispersion models are used to estimate the mean shear velocity. The dispersion curves for each station are shown in Appendix F.
- iv. The stacked S_p -S arrival time and Rayleigh group velocities are modeled to estimate the Moho depth for all approved models.

Table 2 – Parameters used when SRFs and dispersion velocities were modeled using the grid search method.

Parameter	Value (units)
Sediment density	2.7 g/cm ³
Sediment velocity	3.2 km/s
Sediment Poisson's ratio	0.25
Sediment thickness	0.01 – 1 km
Minimum Crustal velocity	3.4 km/s
Maximum Crustal velocity	3.9 km/s
Crustal Poisson's ratio	0.25
Crustal velocity increment	0.05 km/s
Crustal thickness increment	1 km
Minimum Moho depth	25 km
Maximum Moho depth	50 km
Upper Mantle velocity	4.5 km/s
Upper Mantle Poisson's ratio	0.28

3.6.2. The Grid Search Method

Grid Search is a method which considers all the possible solutions to seek those with minimum error and drawn from same size of grid in model space (Menke, 2012; Havskov and Ottemöller, 2010). This method is very sensitive to small depth and velocity uncertainties when solutions are modeled. The Moho conversion arrival time is modeled with the Rayleigh group velocity to estimate the crustal thickness due to sensitivity of both parameters to S-wave velocity. In grid search modeling, I used the Rayleigh group velocity dispersion curves; the SRFs and the parameters listed in Table 2. The grid search modeling follows the procedure described below:

- i. A two layer model was used, a crustal layer (including the sedimentary layer) and an upper mantle layer;
- ii. Generate synthetic velocity–depth models and select the models that fit the measured dispersion curve (as detailed in previous section);
- iii. The grid search is set to accept the dispersion curves models within the margin of the bootstrap error, 0.05 s, when compared with the measured dispersion curve;
- iv. The predicted model Sp Moho conversion arrival time is compared to the Moho conversion time from the real data. The observed conversion time is estimated using the stacking method;
- v. If the compared model fits both the dispersion curve and SRF within the error, it is accepted; if not, it is discarded;
- vi. The grid search results are presented in ranges and bin sizes on histograms. For crustal shear velocity, the histogram bin size is 0.05 km/s and ranges between 3.5 and 4.0 km/s (Fig. 26). For crustal thickness, the histogram bin size is 2 km and ranges between 30 and 40 km (Fig. 27).
- vii. Finally, to estimate the final results from values presented in histograms (Appendix G), the following procedure was used
 - a) For the mean values of crustal thickness and crustal shear wave velocity:

$$M = \frac{\sum X}{N}$$

where M is the mean value of the population, X are individual values from the population, and N is the population size.

b) For the uncertainty:

$$s^2 = \frac{\sum (X - M)^2}{N - 1}$$

where S is the population standard deviation;

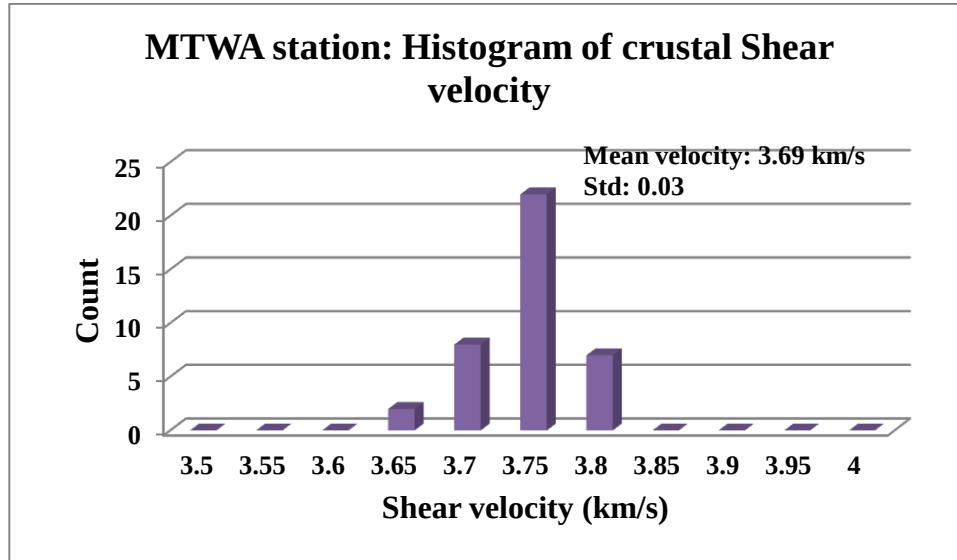


Figure 26 – Plot of crustal shear velocity histogram for MTWA station for a crustal shear velocity range from 3.5 to 4 km/s and 0.05 km/s bin size.

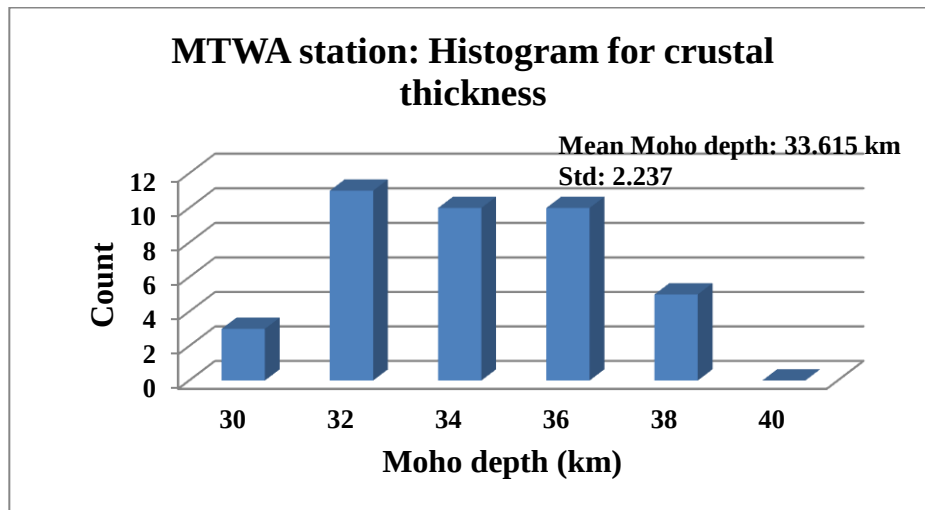


Figure 27 – Plot of crustal thickness histogram for MTWA station for a crustal thickness range from 30 to 40 km and 2.0 km bin size.

4. RESULTS

In this study I used teleseismic events recorded by eleven seismic stations deployed between 2007 and 2013 in seven African Mesozoic/Cenozoic sedimentary basins within Precambrian Belts. The S-wave receiver functions (SRFs) were generated from recorded data and the Grid Search method was used to estimate crustal thickness and shear wave velocity. The summary of all results is presented in Table 3.

Teleseismic events with magnitude ≥ 5.5 and epicentral distances 60 to 82 degrees recorded by eleven stations were used in this study. The number of teleseismic events recorded by five stations (BUTI, KATE, KYLA, MAPH and MSGR) was ≥ 100 , two stations (KIMA and MTWA) had ≥ 50 and four stations (CHAL, INDI, MANG and MOHO) were < 50 (see Table 1).

A total of 837 SRFs were generated from the teleseismic events. However, only 50 SRFs passed the quality control criteria used for manual picking of the Moho arrival time. The selected SRFs were used in further analysis steps, viz. migration in depth, stacking, bootstrapping and grid search. The BUTI and KATE stations have the highest number of selected SRFs (8 and 10 respectively) while only one SRF from the CHAL station was used. Because of the high level of uncertainty and the unrealistic result of 6.1s for the Moho arrival time, the CHAL station was discarded from the further processing. The results of the stacking step are shown in Appendix C.

The results of the bootstrapping error analysis are shown in Appendix D. The error estimated for eight stations was 0.05 s. From the bootstrap plot for KYLA and MAPH station, it was not possible to estimate the error bars. At MAPH station the amplitude of the Moho conversion was too low, and at KYLA station the error estimated was 0.01s, considered as an unreliable result, so both were discarded.

In the grid search, the crustal velocity-depth models for different depths were first modeled using group velocity dispersion curves. The results are shown in Appendix G. The grid search results estimated for average crustal shear wave velocity range between 3.67 and 3.79 km/s with 0.03 km/s of uncertainty. The stations with fastest shear wave velocity are BUTI and MANG; while INDI, MSGR and MTWA stations have slower shear velocities. The results of crustal structure thickness were between 31.86 and 39.63 km. The stations with thinner crust are MTWA, INDI and KATE; while stations with thicker crust are MANG and BUTI. The plot of crustal thickness

and shear wave velocity results are in range shown in histograms (see Appendix H). The crustal thickness histograms bin size is 2 km and the range is 30-44 km, while the average crustal shear velocity histogram bin size is 0.05 km/s and the range is 3.5-4.0 km/s.

The BUTI station is the seismic station with the highest standard deviation, while KATE and INDI have less uncertainty (standard deviation). MAPH, KYLA and CHAL are the stations that recorded the noisiest signals, with the result that 3 or fewer SRFs were used.

The summary of all results is presented in Table 3. Using these results, the seismic stations can be divided into three groups. The first group of stations produced good results (BUTI, INDI, KATE, KIMA and MSGR); the second group produced acceptable results (MANG, MOHO and MTWA); while the third group (CHAL, KYLA and MAPH) failed to meet different quality control criteria and thus no results were obtained.

Table 3 – Summary of data used to model crustal thickness

Belt	Basin	Station name	Seismic events recorded	SRFs used	Moho Arrival time (s)	Dispersion curves models fitted	Average velocity (km/s $\pm$ std)	Crustal Thickness (km $\pm$ std)
Mozambique	Rovuma	INDI	32	5	4.0	15	3.67 $\pm$ 0.03	31.86 $\pm$ 1.0
		MTWA	59	4	4.5	39	3.69 $\pm$ 0.03	33.61 $\pm$ 2.2
	Mandawa	KIMA	69	5	5.1	41	3.75 $\pm$ 0.03	37.43 $\pm$ 2.2
		MOHO	32	4	5.2	42	3.76 $\pm$ 0.03	38.16 $\pm$ 2.1
	Ruvu	MANG	32	4	5.4	30	3.79 $\pm$ 0.03	39.63 $\pm$ 2.2
	Rufiji	CHAL	33	1	-	-	-	-
Ubendian	Rukwa	KYLA	104	3	4.0	-	-	-
Ruwenzori	Lake Albert	BUTI	132	8	5.2	45	3.72 $\pm$ 0.04	38.8 $\pm$ 2.3
	Lake Edward	KATE	132	10	4.2	12	3.73 $\pm$ 0.03	33.83 $\pm$ 0.9
	Mozambique	MSGR	107	3	4.9	42	3.7 $\pm$ 0.03	36.9 $\pm$ 2.1
		MAPH	105	3	4.0	-	-	-

5. DISCUSSION OF RESULTS

In summary, I compared the estimated results in this research with the results of previous studies in the same terrains (see Table 4). The regional results were estimated by Tugume et al. (2012, 2013), Last et al. (1997), Julià et al. (2005), Dugda et al. (2005), Young et al. (2012), Kachingwe (2014) and Kgaswane et al. (2009); while global model results were estimated by Durrheim and Mooney (1994) and Rudnick and Fountain (1995). The results for three terrains: Mozambique Mobile Belt, Ubendian Belt and Mozambique basin are analyzed in terms of consistency.

Table 4 – Summary of crustal structure and shear wave velocity results for Mozambique mobile belt, Ubendian orogenic belt, Mozambique basin and global models for Proterozoic terrains.

Age	Terrain	Crustal thickness (km)	Shear wave velocity (km/s)	Source
Archean	Limpopo Belt	40 ± 3	3.7	Kgaswane et al., 2009
Proterozoic	Global Models	43.6 ± 4.6		Rudnick and Fountain, 1995
		40 – 55		Durrheim and Mooney, 1994
Mesoproterozoic	Ubendian Orogenic Belt	43 ± 4	3.9	Tugume et al., 2013
		42		Tugume et al., 2012
		40 – 45	3.74	Last et al., 1997
		40 ± 2	4.0	Julià et al., 2005
		36.3 ± 2.4	3.7 ± 0.01	This study
Neoproterozoic	Mozambique Mobile Belt	38 – 43	3.9	Tugume et al., 2013
		36 – 39	3.74	Last et al., 1997
		37 – 42		Tugume et al., 2012
		37 – 42		Dugda et al., 2005
		37 – 42		Kachingwe, 2014
		38 ± 2	4.0	Julià et al., 2005
		34 – 41		Young, 2012
		36.14 ± 2.6	3.72 ± 0.04	This study
Phanerozoic	Mozambique Basin	36.9 ± 2.1	3.7	This study

5.1. Uncertainties associated with the S-wave Receiver Functions method

The SRF method uses shorter epicentral distances (60 to 82 degrees) than the PRF method (30 and 90 degrees). Because of the small epicentral distance range the events with magnitude greater than 5.5 relatively few. The SRFs are analyzed and the conversion phase arrival is manually picked. Amplitudes of SRFs can vary greatly because of anisotropy, lateral variations in structures beneath the station, and unsuccessful rotation into the great-circle-path (Tang et al., 2016). In the case of the BUTI and KATE stations, the SRFs have large amplitudes, probably because the stations are above rift basins. The CHAL station recorded events with weak signal, probably owing to scattering beneath the seismic station.

The receiver functions with higher frequency are sensitive to small-scale heterogeneities, while lower frequency receiver functions help image larger-scale structures. The grid search technique uses a low-density layer with lower impedance. The lower density reduces the amplitudes of the multiples. The main difference between the gradient and the density solutions is that the gradient diminishes all amplitudes, whereas the density change principally affects the amplitudes of the multiples (Lodge et al., 2009).

5.2. Crustal structure

5.2.1. Mozambique Mobile Belt

Various studies of the crustal structure of the Mozambique Mobile Belt have been conducted using receiver functions, although different approaches were used.

- Last et al. (1997) modeled the S-wave phase velocity and receiver functions separately and estimated 3.74 km/s for crustal structure shear velocity and 36-39 km for crustal thickness;
- Julià et al. (2005) jointly inverted receiver functions and Rayleigh wave dispersion curves and estimated the crustal thickness crustal shear velocity to be 36-40 km and 3.9-4.1 km/s, respectively, and concluded that the Moho is gradational;

- Dugda et al. (2005), Tugume et al. (2012) and Kachingwe (2014) calculated PRFs using 1.0 and 2.5 Gaussian filters, analyzed the PRFSs with H–k stacking, and estimated the crustal thickness to be 37-42 km;
- Young (2012) used the H-k stacking method to analyze PRFs generated from data recorded in Tanzanian coastal basins. The crustal shear wave velocity estimated is 3.8 km/s and crustal thickness is 34-41 km; and
- Tugume et al. (2013) used joint inversion of PRFs and Rayleigh wave dispersion to estimate crustal thickness of 38-43 km.

In this study the crustal thickness and shear wave velocity were estimated to be 36.1 ± 2.6 km and 3.72 km/s, respectively (Table 4). I found thinner crust and slower shear wave velocity than previous studies. The estimated crustal shear wave velocity in this study is 0.1 to 0.2 km/s slower than estimated by Julià et al. (2005) and Tugume et al. (2013). This might be explained by differences in method and station locations: previous studies used P-wave receiver functions while I used S-wave receiver functions; previous studies used stations in the interior and near to the Tanzania craton, while I used stations in coastal sedimentary basins.

The results of crustal thickness uncertainties accepted using Receiver Function method are ± 3 km (Tugume et al., 2013). If the results of the uncertainties are $> \pm 3$ km, the Moho depth is considered gradational (or not well resolved). Young (2012) found uncertainties for the Moho depth in Tanzanian coastal basins. These uncertainties are ± 5 km, ± 6.5 km and ± 7.5 km for Ruvu, Mandawa, Rovuma basin, respectively.

The global model crustal structure thickness for Proterozoic terrains estimated by Durrheim and Mooney (1994) ranges between 40-55 km, while Rudnick and Fountain (1995) estimated 43 km. The result estimated in this study is ~ 3 km thinner than the global model estimated range.

The pull-apart Tanzania coastal basins, within the Mozambique Mobile Belt, are located between the oceanic crust and the Tanzanian craton (Chatellier and Slevin, 1988). I obtained crustal thicknesses of 32-40 km. Tugume et al. (2013) estimated the crust of the adjacent Tanzanian craton to be 39 km thick.

5.2.2. Ubendian Orogenic Belt

Various studies of the crustal structure of the Ubendian Orogenic Belt were conducted using receiver functions.

- Last et al. (1997) modeling shear phase velocity and receiver functions separately and estimated 3.74 km/s for shear velocity and 40–45 km for the crustal thickness;
- Julià et al. (2005) used joint inversion and Rayleigh wave dispersion curves, and estimated crustal thickness to be 38–42 km;
- Tugume et al. (2012) estimate the same value as Julià et al. (2005);
- Tugume et al. (2013) estimated 3.7 km/s for shear wave velocity and 43 ± 4 km for crustal thickness.

This study estimated a crustal thickness of 36.3 ± 2.4 km and shear wave velocity of 3.7 km/s (see Table 4). The crustal shear velocity estimated in this study is the same as previously estimated by Julià et al. (2005) and Tugume et al. (2013), while the crustal thickness estimated in this study is thinner than previously determined.

The global model crustal structure thickness for Proterozoic terrains, Durrheim and Mooney (1994) estimated ranges between 40–55 km, while Rudnick and Fountain (1995) estimated 43 km. The result estimated in this study is ~ 3 km thinner than these global studies.

5.2.3. Mozambique Basin

For the Mozambique Basin, I estimated crustal thickness and shear wave velocity of 36.9 ± 2.1 km and 3.7 km/s, respectively. Kgaswane et al. (2009), using joint inversion of receiver functions and Rayleigh wave dispersion, estimated 40 ± 3 km for crustal thickness and 3.7 km/s for shear velocity for the northeast Limpopo Belt, which is the west of the Mozambique basin.

The MSGR station is located in the western Mozambique coastal plains within the Mozambique basin. Published geological maps strata reveal post-Karoo deposits, Upper Jurassic to Cenozoic sediments, overlying eroded Karoo rocks. The pull-apart basins are located between thin oceanic crust and thick continental crust (Chatellier and Slevin, 1988), while the Mozambique basin is located between the oceanic crust and the Limpopo belt. This study estimates crustal shear velocity that is similar to Kgaswane et al. (2009).

5.2.4. Global studies

Durrheim and Mooney (1994) stated that the high layer velocity found at the base of crust ($V_p > 7.0$ km/s and $V_s > 4.0$ km/s) is likely composed of rocks with basaltic average composition. The average shear wave velocity estimated in this study is between 3.70 – 3.72 km/s, indicating that the Precambrian terrains considered in this study do not have a thick basaltic lower crust. The seismic velocity differences in the Precambrian crust obtained in this study compared with others regions in the Earth show that the formation processes of these regions are likely different, even though they have same age (Rudnick and Fountain, 1995).

5.3. Eastern and Southern African uplift

In the second part of the discussion, I compare crustal thickness results estimated in this study with elevation for each station to analyze if there is a first order correlation between crustal thickness and the topographic uplift observed in Eastern and Southern Africa terrains. The causes of uplift Eastern and Southern African are debated. The topography present in Africa varies. The highest regions are inland and elevation decreases towards the coastline. I use the Airy model to analyze the principal cause of topography uplift in terrains considered in this studied.

The Airy hypothesis assumes that surface elevation is compensated by thickness variation underneath the uplifted area (Kearey et al. 2009). The Airy relation is:

$$h = r \frac{(\rho_m - \rho_c)}{\rho_c},$$

Where

ρ_m (kg/m³) – density of the uppermost mantle layer;

ρ_c (kg/m³) – density of the crustal layer;

r (m) – thickness of the root;

h (m) – elevation topography.

Consider the initial conditions at with $\rho_m = 3,300$ kg/m³ and $\rho_c = 3,000$ kg/m³. To support an increased elevation of 1,000 m, the crust must be thickened by 10,000 m.

The results in Table 5 were used to examine the relationship between crust thickness and elevation (Fig. 28). The correlation is poor, hence we can conclude that simple Airy-type isostasy does not apply. The type of isostatic compensation is named overcompensated isostatic. This type of isostatic compensation is due to erosion of the mountains and consequent reduction of elevation.

Crustal thickness is a common cause of uplift during orogenesis (Kearey et al., 2009). I conclude that crustal thickening is not the first-order cause of uplifted terrains in Southern and Eastern Africa

The Southern and Eastern African plateau uplift is a result of dynamic uplift processes combined with isostatic compensation resulting from heating of the thermal lithosphere occurred in Mesozoic (Nyblade and Robinson, 1994 and references within; Nyblade and Sleep, 2003). The main reason for the thermal perturbations in the lithosphere above the convective region in the asthenosphere is the presence of African Superwell. Nyblade and Robinson (1994) and Pasyanos and Nyblade (2007) discussed the uplift in the Southern and Eastern African plateau exhaustively, and concluded that is caused by the African Superwell, which is located underneath the Southern and Eastern African plateau. The heat flow in Southern Africa plateau is about 9mWm^{-2} higher than global averages from similar terrains, indicating that the heat source is deep seated and not crustal (Nyblade and Robinson, 1994).

Table 5 – The crustal structure thickness versus elevation for each stations deployed in the Mozambique mobile belt, Mozambique basin and Ubendian belt.

Station	BUTI	INDI	KATE	KIMA	MANG	MOHO	MSGR	MTWA
Crustal thickness (km)	38.8	31.86	33.83	37.43	39.63	38.16	36.9	33.61
Elevation (m)	623	14	950	7	366	4	135	32

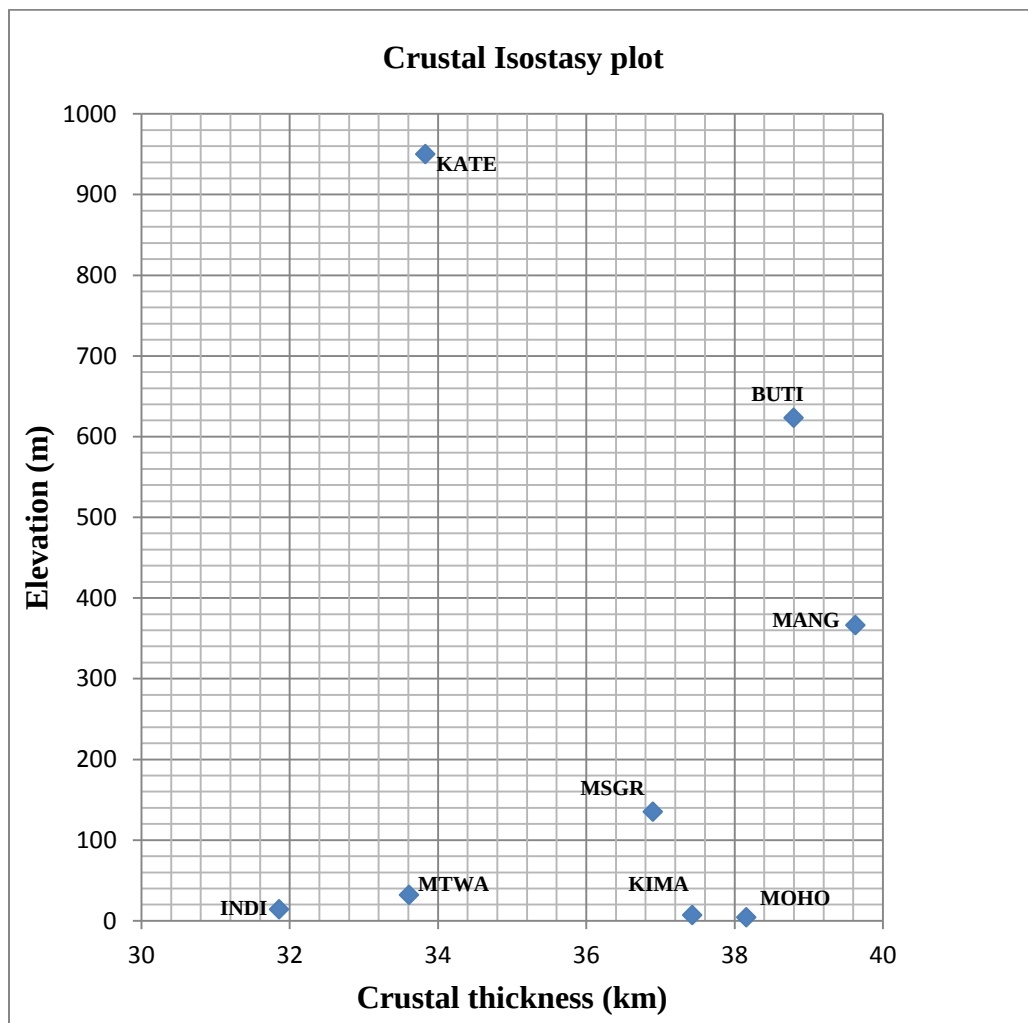


Figure 28 – Crustal thickness versus elevation for each station where Moho depths were determined using SRFs.

6. CONCLUSION

Shear wave Receiver Functions generated from teleseismic events and modeled with Rayleigh group velocity dispersion curves using the Grid Search method were used to study the crustal thickness and shear wave velocities for seven Phanerozoic sedimentary basins located within three African Proterozoic Orogenic Belts.

The SRF method was shown to be suitable in sedimentary terrains due to the early arrival of the Sp Moho conversion, free from contamination by crustal reverberations, enabling the time of the Sp Moho arrival to be accurately measured. This is not the case for the P-wave Receiver Function Method.

During the study, I used data from the temporary seismic stations deployed for less than two years and epicentral distances between the 60 and 82 degrees. The number of usable events recorded by stations in coastal sedimentary basin was fewer than the number recorded by seismic stations in rift basins. The seismic signal received in the coastal sedimentary basins (Tanzanian coastal basin and Mozambique Coastal Plain) suffered scattering because of the unconsolidated geology.

The crustal thickness in late Precambrian and Phanerozoic terrains estimated using SRFs was found to be slightly thinner than previous studies conducted in the same terrains and global average models. The SRF method indicated a sharp Moho, in contrast with previous results using the PRF method that indicated a gradational Moho. The results of crustal shear velocity are similar or slower than previously obtained in the same terrains.

I conclude that crustal thickening is not the first-order cause of uplifted terrains in southern and eastern Africa.

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8. APPENDICES

Appendix A. Station location, period of deployment and instrument deployed.

Station Location (Station Code)	Net work	Location				Period of deployment	Equipment
		Lat. (°)	Long. (°)	Elev. (m)	Basin, Belt		
Butiaba (BUTI)	ZP	1.82	31.33	623	Lake Albert basin, Rwenzori Belt	2007.08.29 to 2008.11.26	I
Katwe (KATE)	ZP	-0.14	29.87	950	Lake Edward basin, Rwenzori Belt	2007.09.02 to 2008.12.01	II
Keyla (KYLA)	ZP	-9.60	33.87	501	Rukwa basin, Ubendian Belt	2008.12.19 to 2010.08.05	II
Chalinze (CHAL)	YH	-6.64	38.37	219	Ruvu basin, Mozambique Mobile Belt	2010.02.21 to 2011.07.13	I*
Manero Mango (MANG)	YH	-7.20	38.77	366	Ruvu basin, Mozambique Mobile Belt	2010.02.20 to 2011.07.05	I
Mohoro (MOHO)	YH	-8.14	39.18	4	Rufiji Trough, Mozambique Mobile Belt	2010.02.12 to 2011.07.15	I*
Kilwa (KIMA)	YH	-8.92	39.51	7	Mandawa basin, Mozambique Mobile Belt	2010.02.16 to 2011.07.16	I*
Lindi (INDI)	YH	-10.02	39.71	14	Rovuma basin, Mozambique Mobile Belt	2010.02.13 to 2011.07.16	I*
Mtwara (MTWA)	YH	-10.28	40.19	32	Rovuma basin, Mozambique Mobile Belt	2010.02.14 to 2011.07.17	I
Massingir (MSGR)	XV	-23.83	32.18	135	Mozambique basin	2011.08.29 to 2013.08.30	III
Mapinhane (MAPH)	XV	-22.25	35.08	42	Mozambique basin	2011.08.11 to 2013.08.31	III

a) Instruments deployed in seismic station: All stations used 24 – bit Reftek 130 Data logger;

I – Broadband Seismometer Guralp CMP 3T (* is applied for CMP 40T);

II – Broadband Seismometer Streckeisen STS – 2;

III – Broadband Nanometrics Trillium Compact;

b) Pictures of instruments deployed in the seismic stations.



Picture 1. An example of complete seismic station deployed in sedimentary basin.

- 1 – GPS – Global Position System;
- 2 – Breakout Box;
- 3 – Battery Box;
- 4 – Power Box;
- 5 – Data logger; and
- 6 – Seismic Box.



Picture 2. Broadband seismometer: Streckeisen STS – 2 G1.



Picture 3. Broadband seismometer: Guralp CMG – 40T.



Picture 4. Broadband seismometer: Nanometrics Trillium Compact



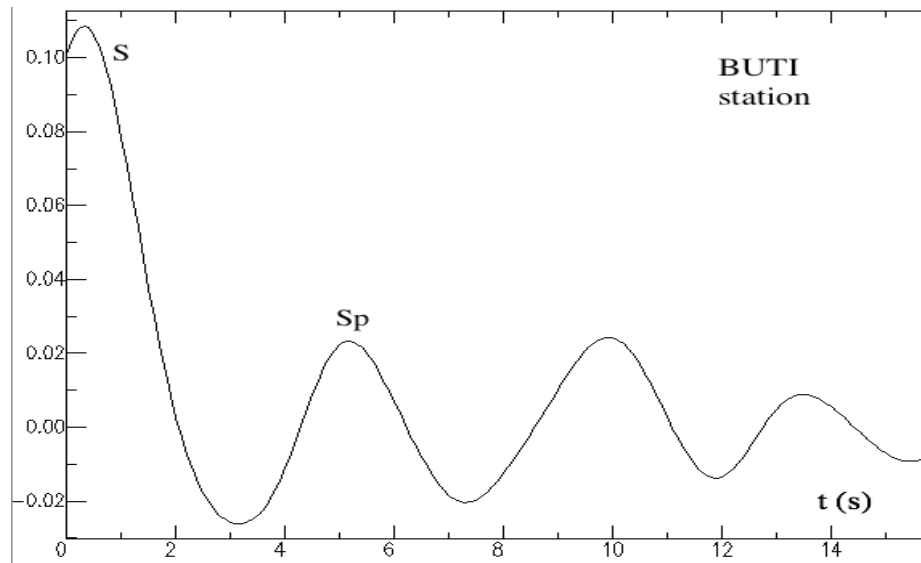
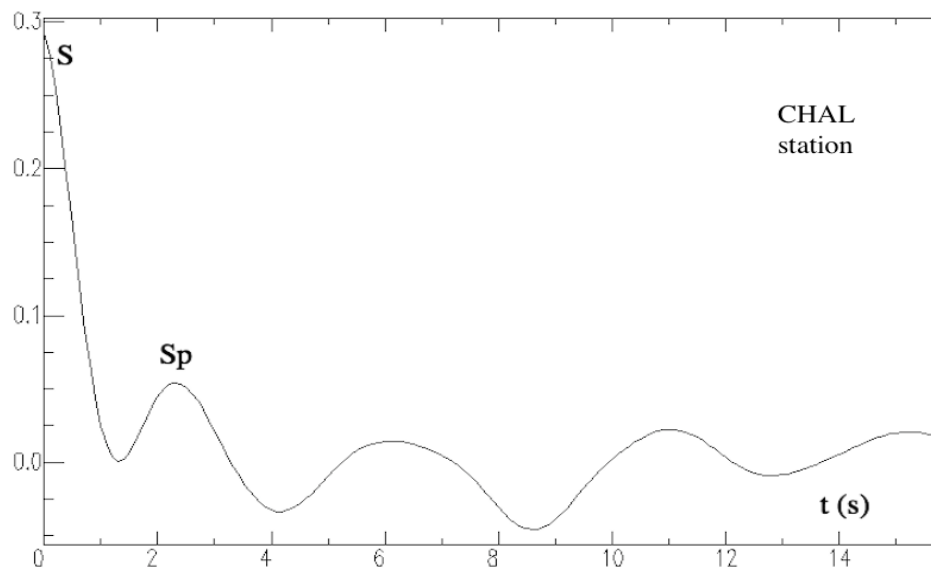
Picture 5. Data logger 24 – bit Reftek 130

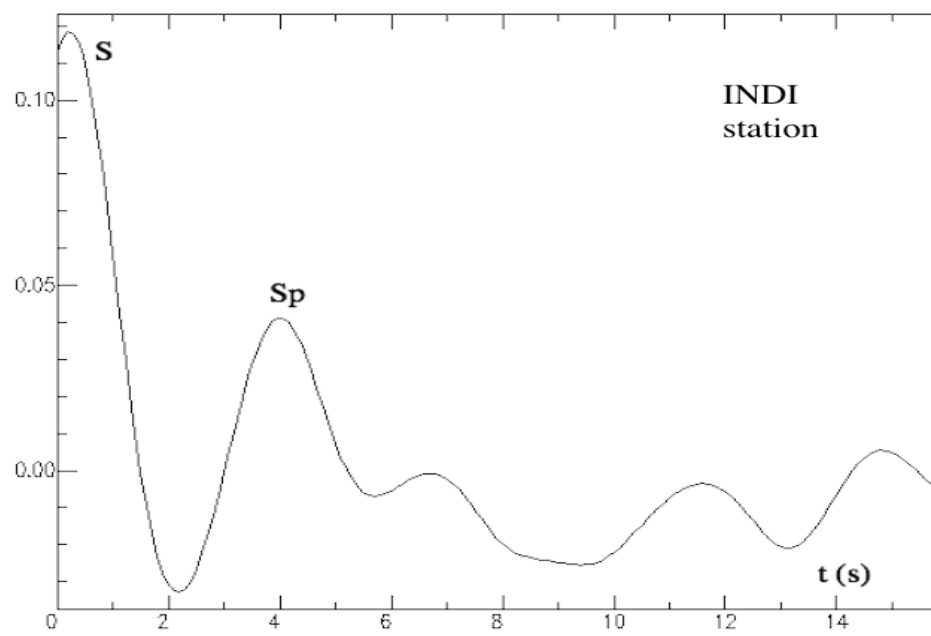
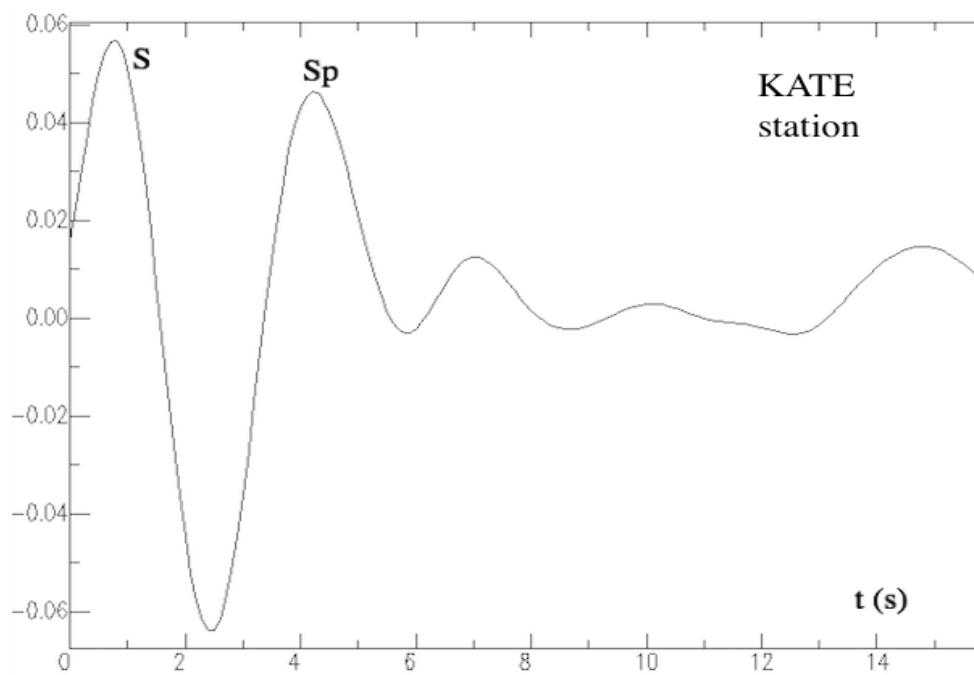
Appendix B. *List of events used to generate SRF.*

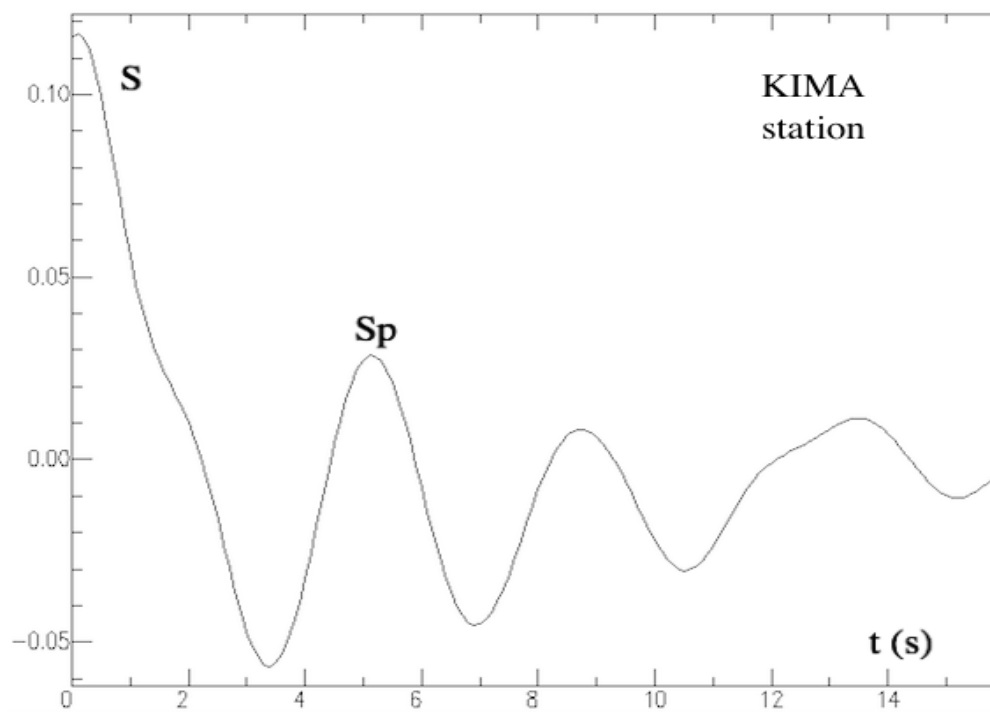
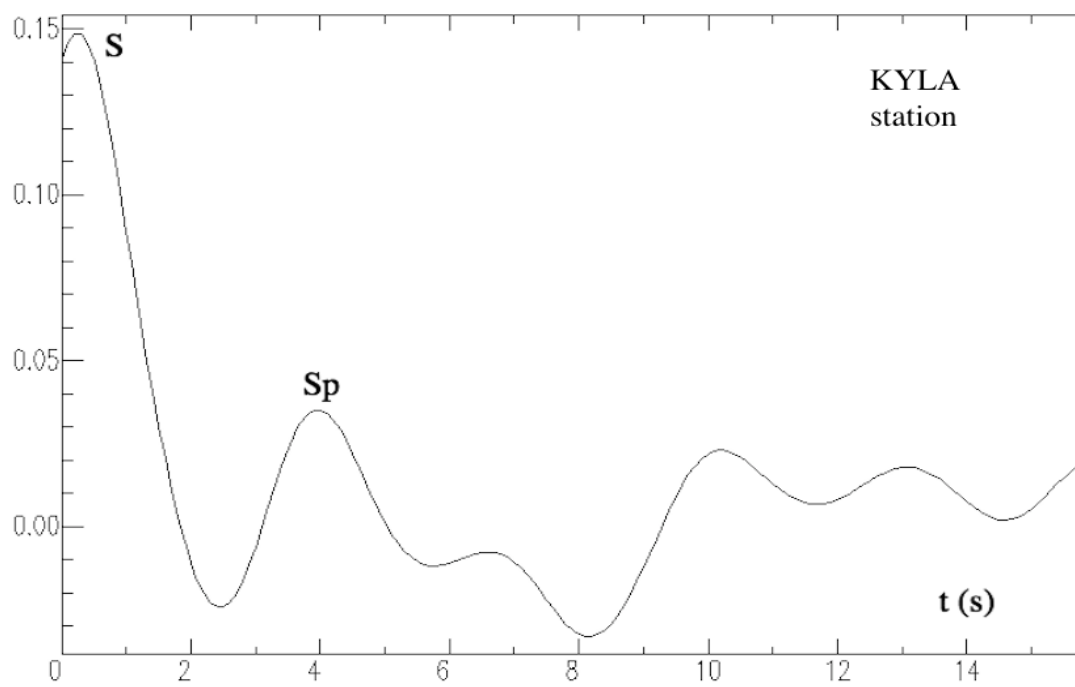
Date	Time	Latitude	Longitude	Depth (km)	Magnitude
2007-09-15	14:45:32	- 2.85	101.10	65.5	5.5 Mw
2008-01-04	07:29:19	- 2.79	100.91	40.6	6.0 Mw
2008-01-22	17:14:58	1.03	97.45	40.6	6.2 Mw
2008-02-12	01:29:41	- 2.98	101.15	48.7	5.5 Mw
2008-05-12	06:27:59	31.06	103.37	7.6	7.9 Mw
2008-05-31	04:37:54	- 41.25	80.48	0.5	6.4 Mw
2008-07-01	01:54:40	- 58.25	- 21.80	3.4	5.6 Mw
2008-07-14	04:44:52	2.20	96.53	33.6	5.6 Mw
2008-07-23	19:54:44	32.82	105.54	5.9	5.5 Mw
2008-07-30	20:15:12	-59.71	-27.88	136.2	5.5 Mw
2008-08-01	08:32:43	32.03	104.72	11	5.7 Mw
2008-08-05	09:49:17	32.75	105.49	6	6.0 Mw
2008-08-26	03:07:29	-6.338	104.46	21.7	5.7 Mw
2008-08-27	01:35:32	51.60	104.15	16	6.3 Mw
2008-08-30	08:30:53	26.24	101.88	11	6.0 Mw
2008-08-30	08:30:53	26.24	101.88	11	6.0 Mw
2008-08-31	08:31:10	26.23	101.97	10	5.6 Mw
2008-10-08	14:07:15	29.76	90.33	2.4	5.5 Mw
2009-07-09	11:19:17	25.71	101.09	13.9	5.7 Mw
2009-11-10	02:48:46	8.02	91.95	22.7	6.0 Mw
2009-12-17	14:55:38	42.33	30.52	15.1	5.5 Mw
2010-03-11	06:22:18	- 57.28	- 27.93	307.5	5.6 Mw
2010-04-11	22:08:11	37.00	-3.47	619.6	6.3 Mw
2010-04-25	21:09:47	-55.70	-27.91	26	5.6 Mw
2010-04-27	17:17:14	-50.02	114.18	12.7	5.6 Mw
2010-06-03	05:35:44	33.29	96.19	18.6	5.5 Mw
2010-08-13	07:58:49	36.88	-32.90	12.8	5.5 Mw

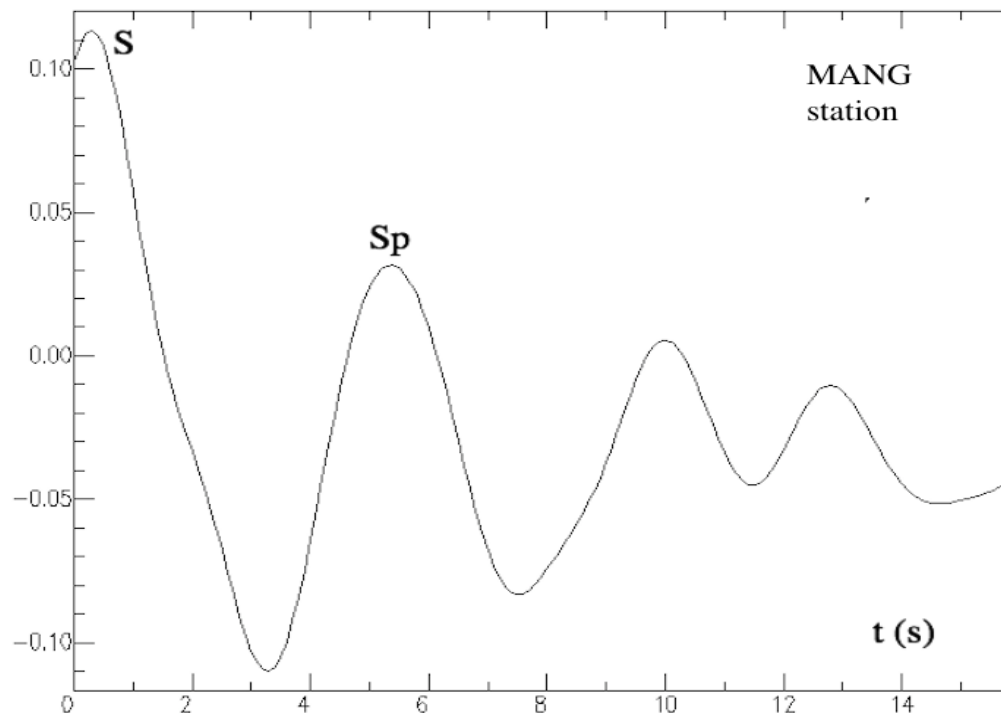
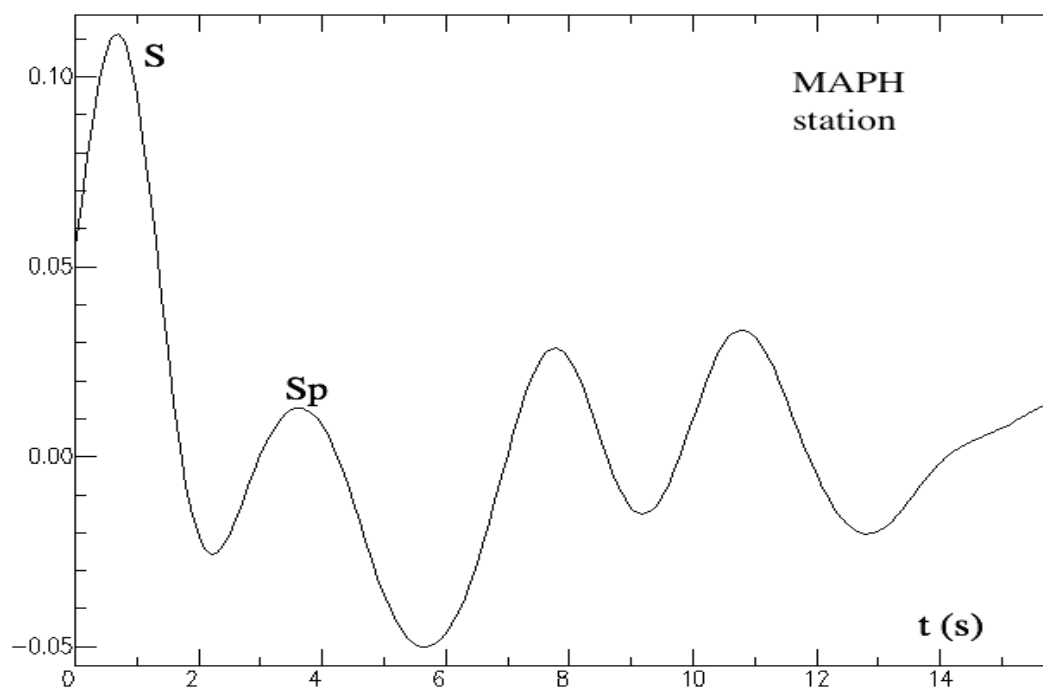
2010-08-29	06:37:48	-55.87	-27.09	31.7	5.5 Mw
2010-10-03	04:30:15	-47.31	100.26	10	5.5 Mw
2010-11-19	21:55:15	1.18	100.09	215.8	5.6 Mw
2010-12-07	04:27:22	-58.02	-7.61	10	5.9 Mw
2010-12-23	16:28:14	-8.06	119.97	202.9	5.5 Mw
2011-01-02	15:19:31	-4.56	101.32	25.7	5.8 Mw
2011-01-17	19:20:56	-5.10	102.60	39.1	6.0 Mw
2011-02-06	11:21:17	-8.25	104.11	15	5.7 Mw
2011-02-06	11:21:17	-8.25	104.11	15	5.7 Mw
2011-02-28	11:04:13	-58.91	-16.28	10	5.6 Mw
2011-05-28	17:07:41	-5.70	103.49	32	5.7 Mw
2011-07-08	05:53:03	0.96	-26.41	10	5.6 Mw
2011-07-11	08:53:58	-4.06	121.69	35	5.5 Mw
2011-07-15	13:26:02	-60.76	-23.52	10	6.0 Mw
2011-09-05	17:55:11	2.96	97.89	91	6.7 Mw
2011-11-07	11:59:31	36.50	71.10	212.1	5.7 Mw
2012-04-24	14:57:10	8.86	93.94	14.1	5.6 Mw
2012-04-25	07:42:23	9.01	93.94	9	5.9 Mw
2012-05-12	23:28:43	38.61	70.35	10	5.7 Mw
2012-07-12	14:00:34	36.52	70.90	198	5.8 Mw
2012-09-14	04:51:47	-3.319	100.59	19	6.3 Mw
2012-09-15	16:32:21	-10.73	113.86	10	5.7 Mw
2013-01-28	16:38:53	42.60	79.70	15	6.2 Mw

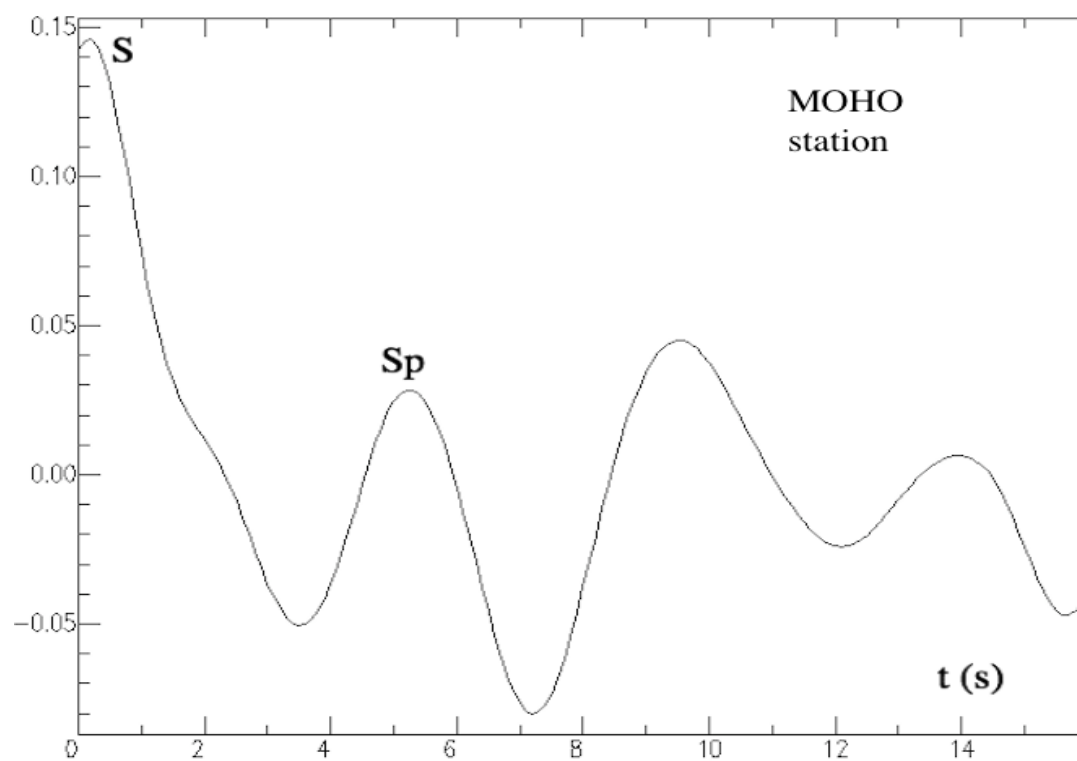
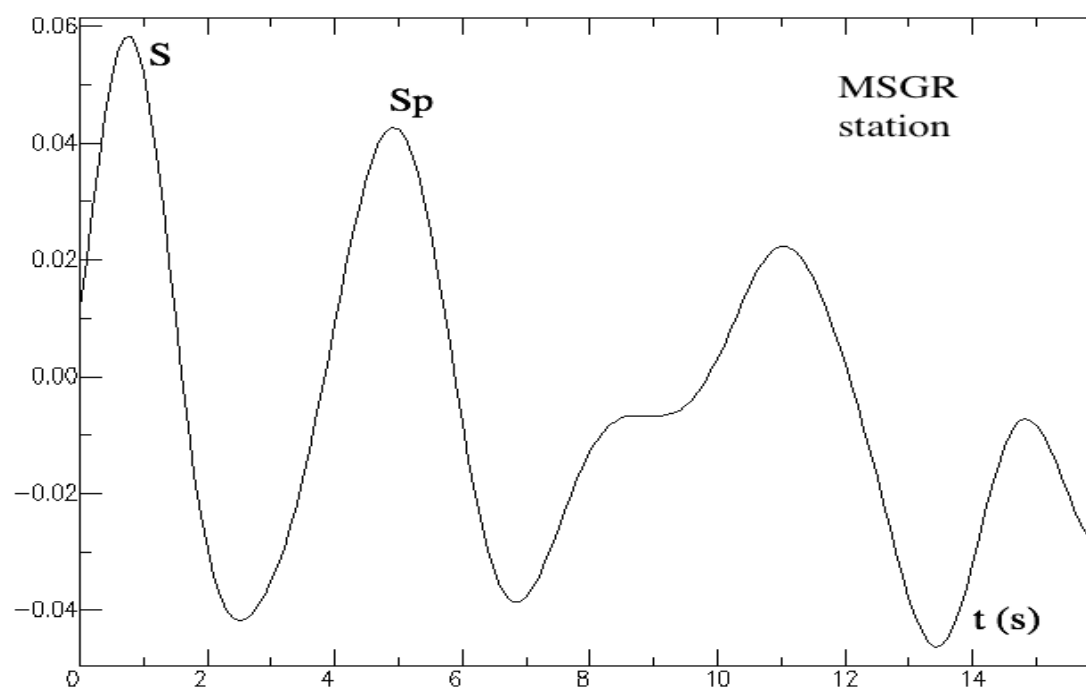
Appendix C. *Plots of stacking results for each station.*

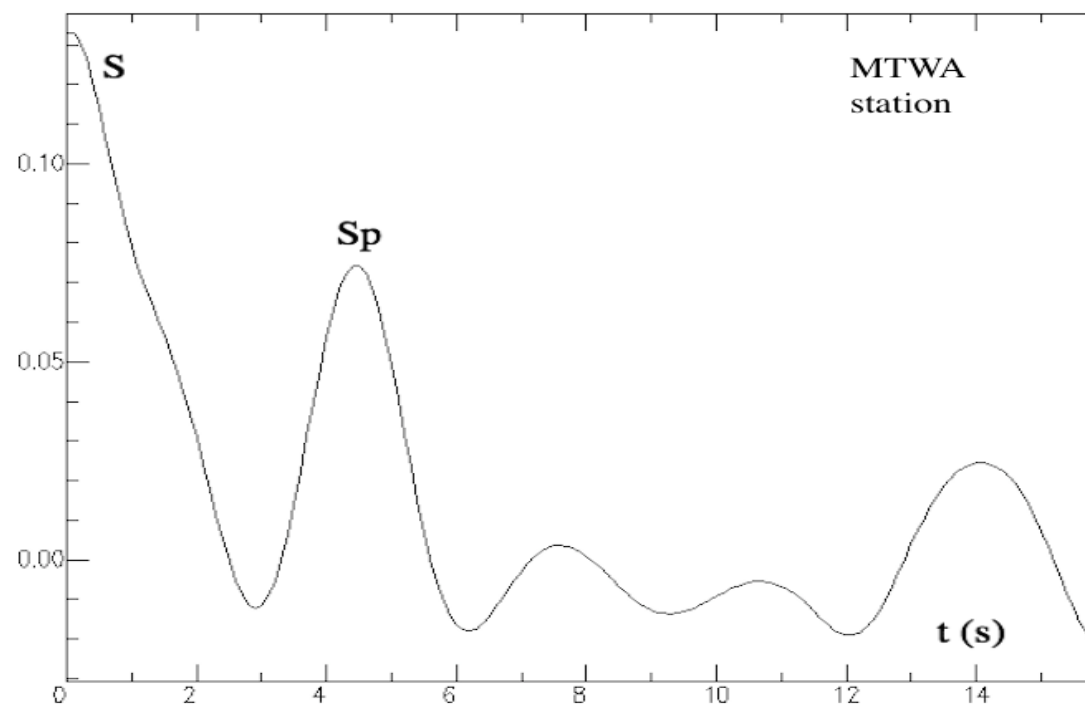




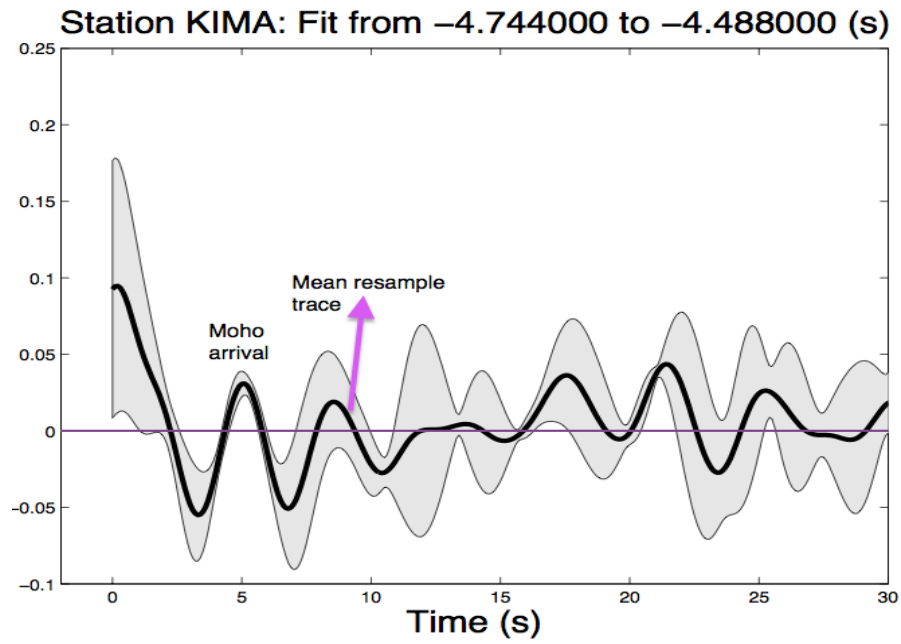
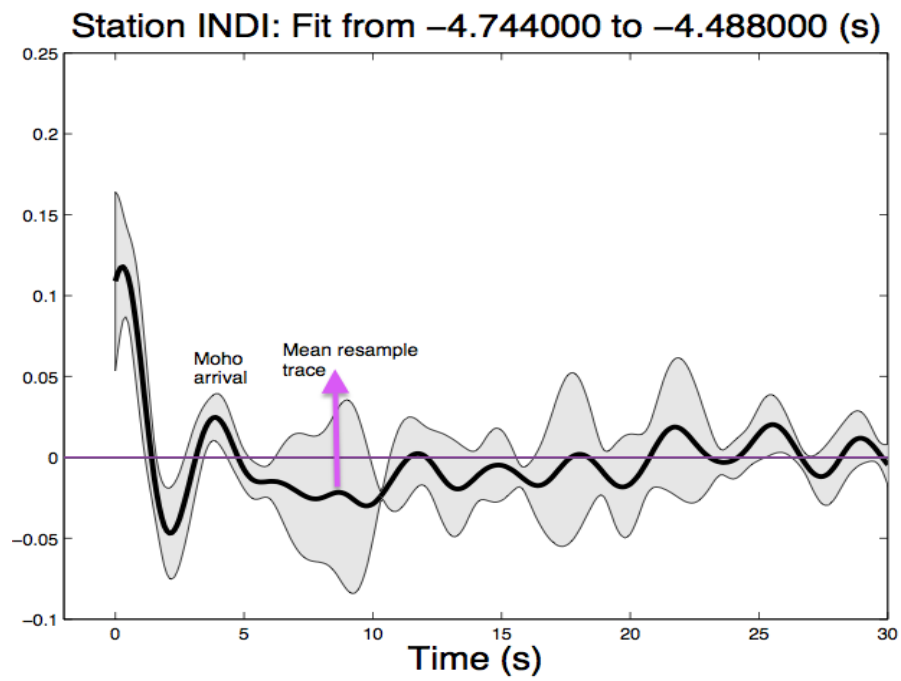


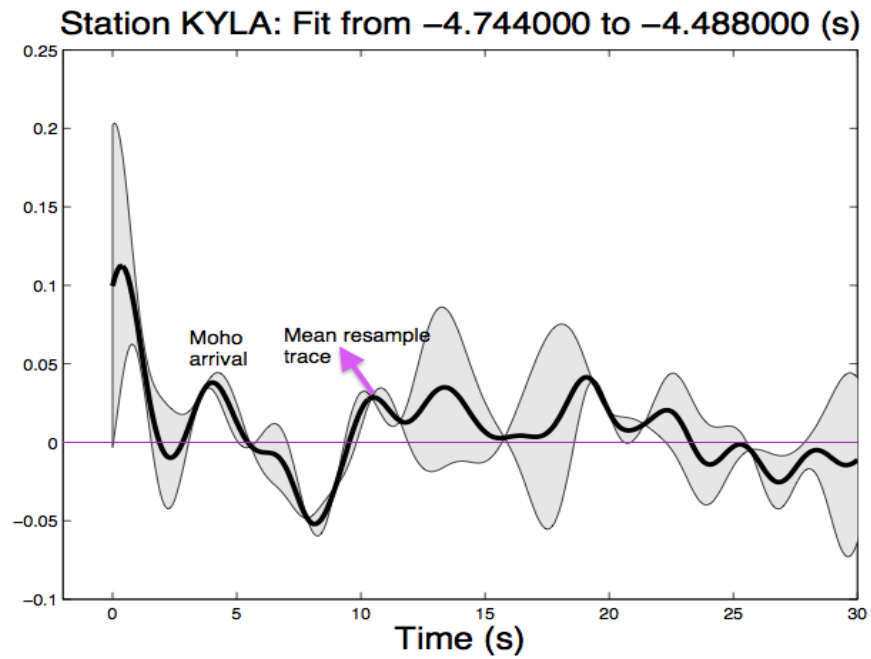
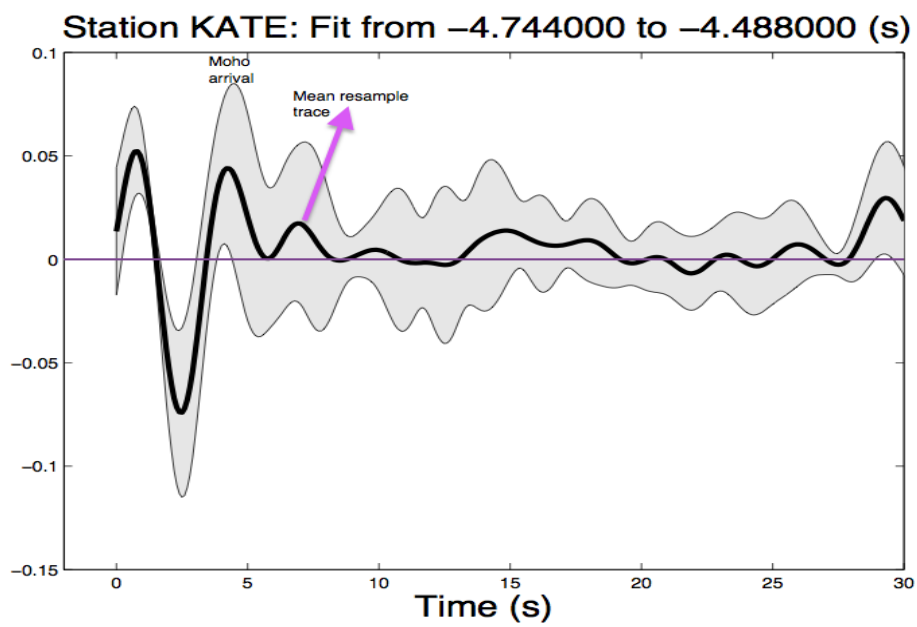


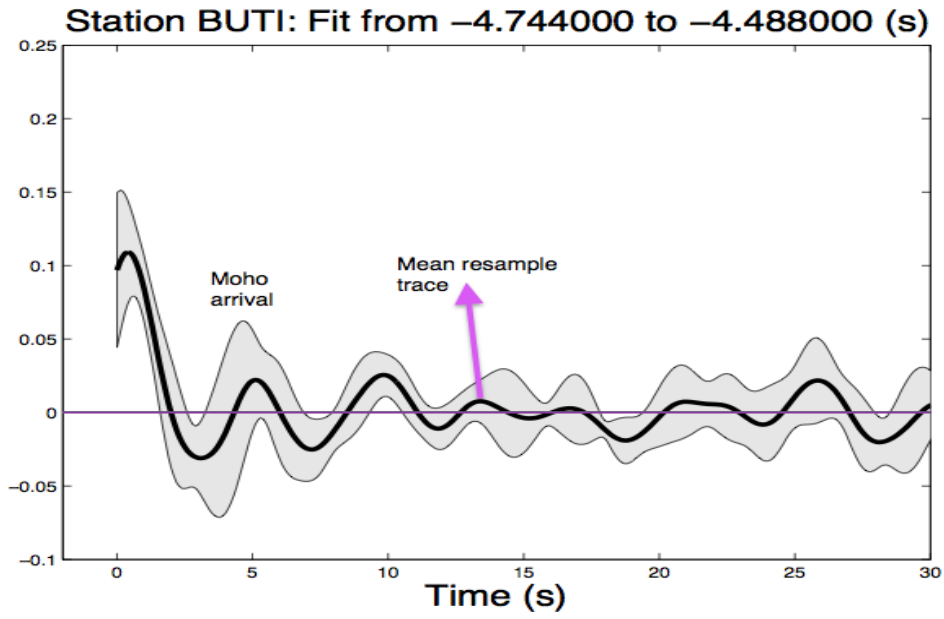
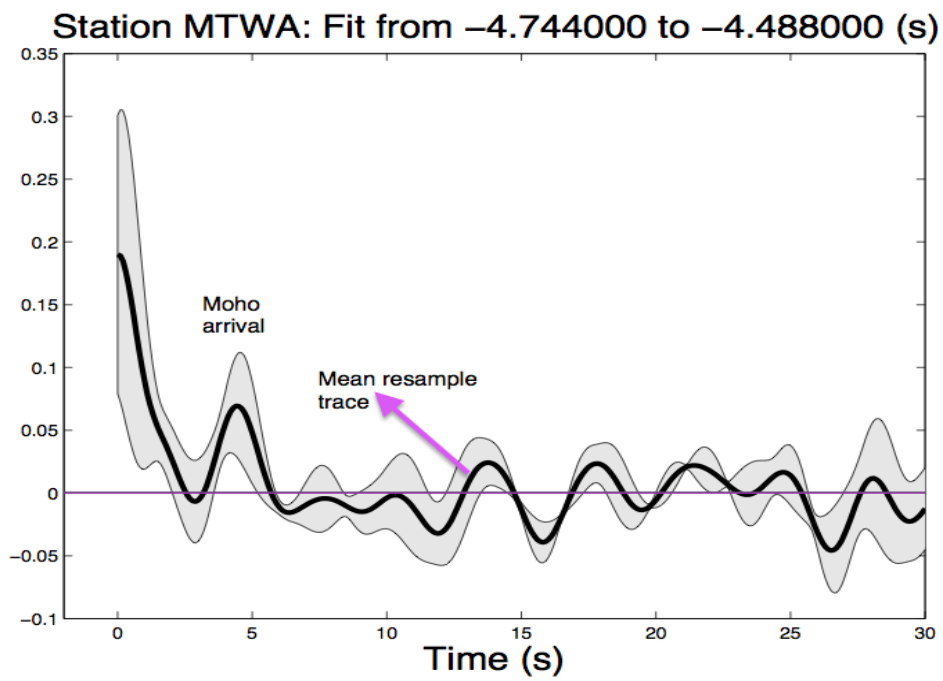




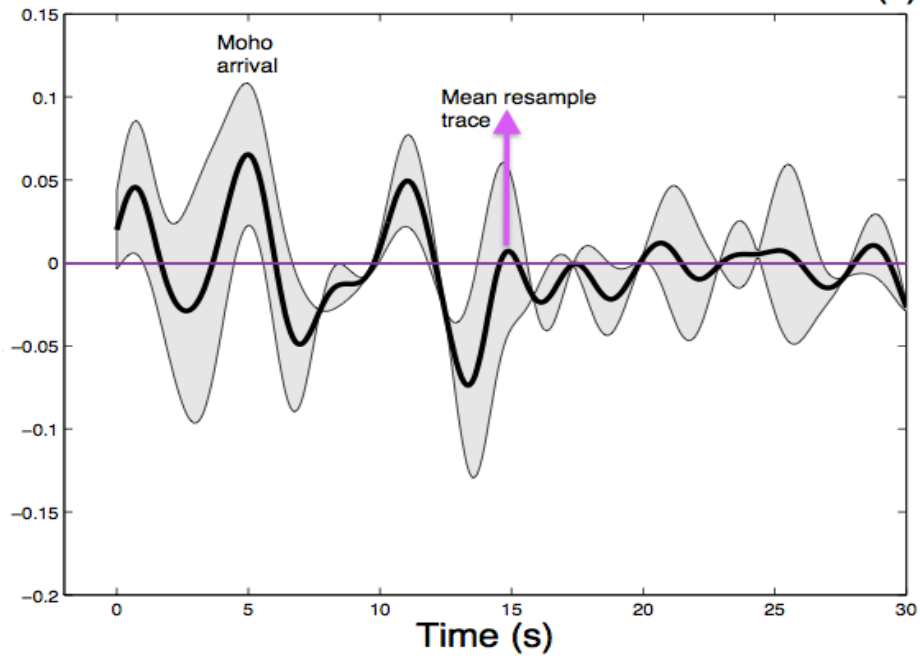
Appendix D. *Plots of bootstrap result per station.*



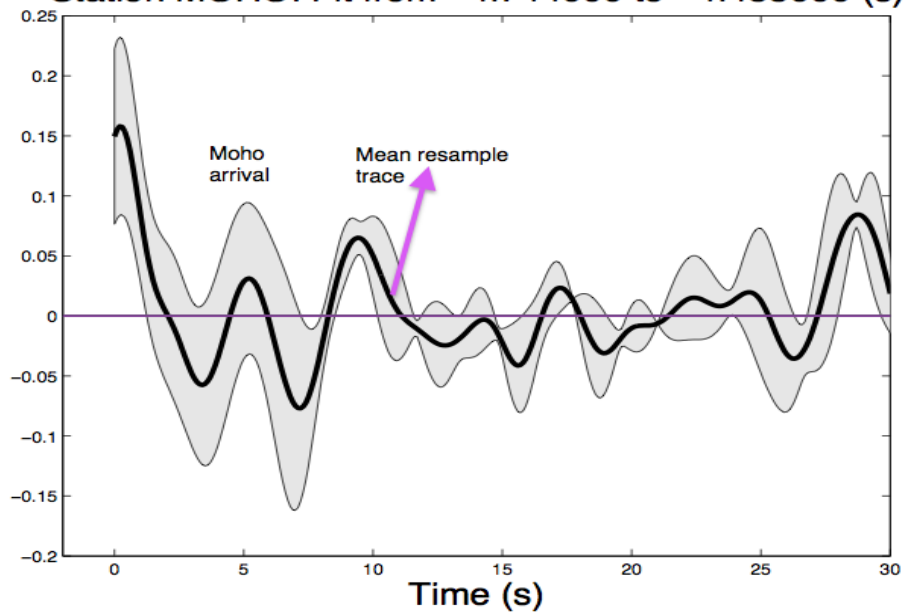


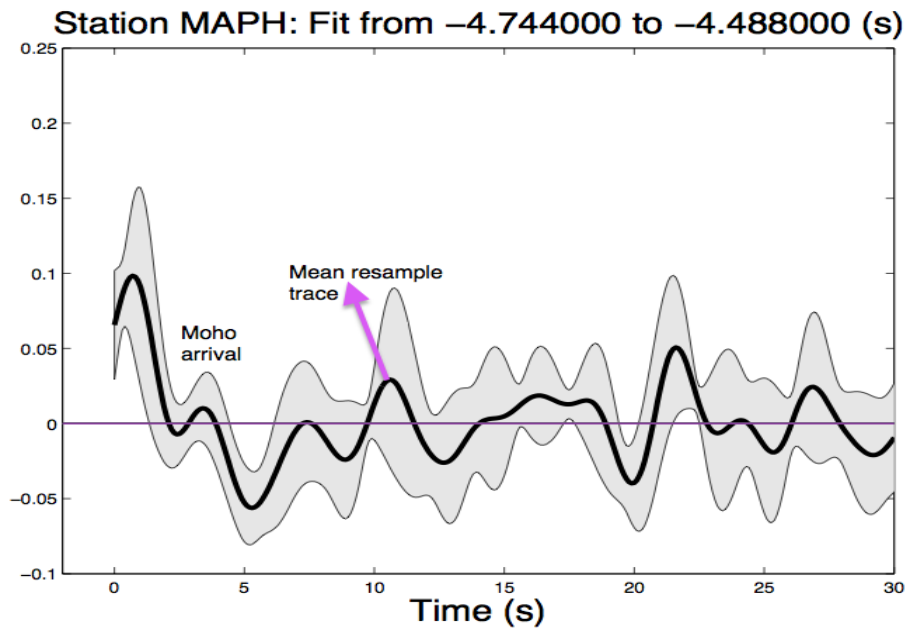
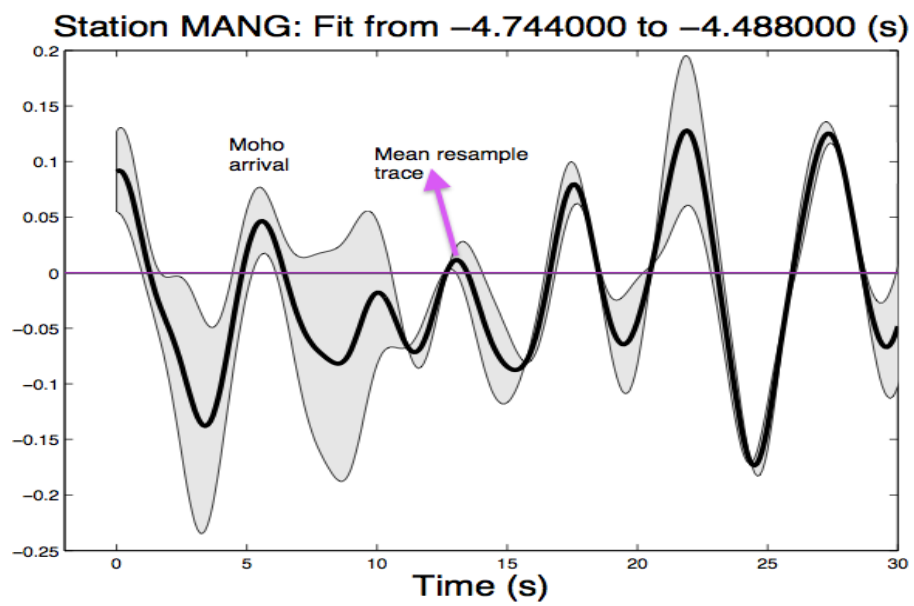


Station MSGR: Fit from -4.744000 to -4.488000 (s)



Station MOHO: Fit from -4.744000 to -4.488000 (s)





Appendix E. Group velocity dispersion curves data with parameters as period, Rayleigh velocity and dispersion errors for each station.

	Dispersion curve data		
Station Name	Period	Rayleigh velocity (km/s)	Error
BUTI	10	2.94	0.08
	15	2.94	0.08
	20	3.12	0.02
	25	3.29	0.08
	30	3.47	0.03
CHAL	10	2.94	0.06
	15	2.94	0.98
	20	3.2	0.08
	25	3.38	0.02
	30	3.54	0.02
INDI	10	2.9	0.02
	15	2.89	0.08
	20	3.18	0.02
	25	3.38	0.03
	30	3.53	0.09
KATE	10	3.12	0.02
	15	3.13	0.04
	20	3.2	0.02
	25	3.34	0.08
	30	3.49	0.02

KIMA	10	2.91	0.09
	15	2.91	0.09
	20	3.2	0.02
	25	3.39	0.04
	30	3.53	0.05
KYLA	10	2.96	0.02
	15	2.96	0.02
	20	3.07	0.04
	25	3.24	0.04
	30	3.39	0.05
MANG	10	2.9	0.04
	15	2.9	0.03
	20	3.22	0.02
	25	3.41	0.02
	30	3.57	0.02
MAPH	10	2.9	0.09
	15	2.9	0.02
	20	3.16	0.03
	25	3.36	0.05
	30	3.49	0.05
MOHO	10	2.93	0.09
	15	2.93	0.02
	20	3.21	0.04
	25	3.39	0.02
	30	3.53	0.04

MSGR	10	2.97	0.02
	15	2.95	0.02
	20	3.13	0.04
	25	3.28	0.02
	30	3.46	0.04
MTWA	10	2.87	0.04
	15	2.86	0.03
	20	3.18	0.02
	25	3.4	0.04
	30	3.54	0.05

Appendix F. *Grid Search scripts used to generate velocity – depth models and estimate shear wave velocity and crustal thickness*

```
#!/usr/bin/env perl -w

# Create output file that lists names of models generated
open(OUT,">MODELS");

# Define dispersion data file and error associated with
# that velocity dispersion data
$dspData = "BUT1.disp";
$error = 0.05;

# Setup search limits/increments for different variables
$crustVel = 3.2;      # define fixed values (Vs of ice and upper mantle and the ice thickness)
$mantUpVel = 4.5;
$SedimentCrst = 0.01 – 1 km;
$SedimentRho = 2.7;
$crstPR = 0.25;      # define additional fixed values (ice density and poisson's ratio)
$mantPR = 0.28;      # for all layers
$SedimentPR = 0.25;

$crustAvgVelMin = 3.4; # min/max search values and increment for crustal velocity (km/s)
$crustAvgVelMax = 3.9;
$crustAvgVelInc = 0.05;
$crstLowCutoff = 4.22;
$MohoDepMin = 25;     # min/max search values and increment for Moho depth (km)
$MohoDepMax = 50;
$MohoInc = 1;

# make starting values equal to the minimum search limits
$crustAvgVel = $crustAvgVelMin;
$MohoDep = $MohoDepMin;

$crustUpVel = $crustAvgVelMin;

# Increment through values to make models
while ($crustUpVel <= $crustAvgVelMax) {
    system("echo  \");
    system("echo Running models with $crustUpVel Upper Crust velocity");
    $crustAvgVel = $crustAvgVelMin;

    while ($crustAvgVel <= $crustAvgVelMax) {

        # to do approach similar to Last et al. 1997, the upper and lower crustal Vs are
```

```

# averaged by slowness to equal the average Vs of the crust
# $crustUpVel = $crustAvgVelMin;
$crustLowVel = 1 / (((1 / $crustAvgVel) * 2) - (1 / $crustUpVel));

if ($crustLowVel <= $crustLowCutoff && $crustLowVel >= $crustUpVel) {
    system ("echo Lower Crust Vs $crustLowVel and Upper Crust Vs $crustUpVel");
    # we still have o.k. lower crust velocity - proceed

    while ($MohoDep <= $MohoDepMax) {

        # also, to do approach similar to Last et al. 1997, will make upper and lower crust the
        same thickness
        $crustUpThick = (($MohoDep - $SedimentCrst) / 2) + $SedimentCrst;

        # setup name of velocity model to reflect these values
        # Modified by Cristo
        $LowCrst = substr $crustLowVel, 0, 5;
        $name = "velMod_CU_" . $crustUpVel . "_CL_" . $LowCrst . "_CAv_" . $crustAvgVel .
        "_M_" . $MohoDep . "_";
        print OUT ($name, "\n");

        # Calculate appropriate density values for associate velocities/poisson's ratios
        $crustUpVp = $crustUpVel * ((2 * ($crstPR - 1) / (2 * $crstPR - 1)) ** 0.5);
        $crustLowVp = $crustLowVel * ((2 * ($crstPR - 1) / (2 * $crstPR - 1)) ** 0.5);
        $mantVp = $mantUpVel * ((2 * ($mantPR - 1) / (2 * $mantPR - 1)) ** 0.5);
        $crustUpRho = $crustUpVp * 0.32 + 0.77;
        $crustLowRho = $crustLowVp * 0.32 + 0.77;
        $mantRho = $mantVp * 0.32 + 0.77;

        # make corresponding velocity model
        system ("makeVelMod_Sediment.csh $SedimentVel $crustUpVel $crustLowVel
        $mantUpVel $SedimentCrst $crustUpThick $MohoDep $iceRho $crustUpRho $crustLowRho
        $mantRho $icePR $crstPR $mantPR $name");

        # increment Moho Depth
        $MohoDep = $MohoDep + $MohoInc;
    }

}

# reset $MohoDep to min value and increment $crustAvgVel
$MohoDep = $MohoDepMin;
$crustAvgVel = $crustAvgVel + $crustAvgVelInc;

}
$crustUpVel = $crustUpVel + $crustAvgVelInc;

```

```

}
close(OUT);

# Now, open up the list of models and for each one, generate their predicted velocity disperion;
# for the AGAP modeling, we will be using phase velocities
open(TABLEA,"MODELS");
open(OUT2,">XY_FILES");
@tablea = <TABLEA>;

system ("echo Making synthetic dispersion curves \.\.\\.");
for ($i = 0; $i < @tablea; $i++) {
    chomp ($tablea[$i]);
    ($model) = (split /\s+/, $tablea[$i])[0];
    # temporary rename of model to run dispersion code
    system ("cp $model testModel");
    system ("dosrf testModel < list.periods > /dev/null");
    $scrName = $model . ".xy";
    print OUT2 ($scrName, "\n");
    system ("sac2xy cray1.sac tmp > /dev/null");
    system ("awk '{print \$2 }' < tmp > tmp2");
    system ("paste list.periods tmp2 > $scrName");
    system ("/bin/rm testModel *.sac");
}
close(TABLEA);
close(OUT2);
system ("echo Done making synthetic dispersion curves");

# Now, for each of those generated models, see if the predicted values all fall within the error
# for the disperion data.
open(TABLEB,"XY_FILES");
@tableb = <TABLEB>;

for ($j = 0; $j < @tableb; $j++) {
    chomp ($tableb[$j]);
    ($xyFile) = (split /\s+/, $tableb[$j])[0];
    # Modified by Cristo
    #($n1,$n2,$n3) = (split /\_/, $xyFile)[0,1,2];
    #($end) = substr($n3,0,6);
    # $velName = $n1 . "\_" . $n2 . "\_" . $end;
    ($velName) = (split /\.xy/, $xyFile);

    open(TABLEC,$xyFile);
    open(TABLED,$dspData);
    @tablec = <TABLEC>;
    @tabled = <TABLED>;
    $OOB = 0;

```

```

for ($k = 0; $k < @tablec; $k++) {
    chomp ($tablec[$k]);
    chomp ($tabled[$k]);
    ($sper,$phsVel) = (split /\s+/, $tablec[$k])[0,1];
    ($sperD,$phsVelD) = (split /\s+/, $tabled[$k])[0,1];

    # make sure looking at same period value and compare
    # phase velocity values
    if ($sper eq $sperD) {
        $phsVelLimUp = $phsVelD + $error;
        $phsVelLimLow = $phsVelD - $error;

        if ($phsVel <= $phsVelLimUp && $phsVel >= $phsVelLimLow) {
            # value falls within specified phase velocity range
        } else {
            # value has fallen outside of range, so we want to
            # flag this as a bad model
            $OOB = 1;
        }
    }
}

# check to see if had any out-of-bounds values
if ($OOB eq 0) {
    # all datapoints fall within acceptable range
    print "Model ", $xyFile, " is acceptable\n";
} else {
    # some data fell outside of acceptable range
    system ("rm $xyFile $velName");
}
close(TABLED);

}
close(TABLEB);

unlink("MODELS");
unlink("XY_FILES");
system ("echo Done with the dispersion curve matching, moving to SRF timing script");

```

Script interpretation

1. Variables definition

dispData = Station.disp; => station name vary according to the station data used.

error = 0.05s; => bootstrapping estimated error for all stations.

Sedmtvel = 3.2 km/s; => sediments velocity.

MantleUpvel = 4.5 km/s; => upper mantle velocity.

SedmCrust = 0.1 – 1 km; => sediments thickness given by range because real thickness was unknown.

SedmRho = 2.7 g/m³; => sediments density.

CrustPR = 0.25; => crustal Poisson's ration.

MantlePR = 0.28; => mantle Poisson's ration.

SedmPR = 0.25; => sediments Poisson's ration.

CrustAvgVelMin = 3.4 km/s; minimum crustal velocity expected to be consider in a good model.

CrustAvgVelMax = 3.9 km/s; maximum crustal velocity expected to be consider in a good model.

CrustAvgVelInc = 0.05; velocity increment used to create an intermediate loop.

CrustLowCutoff = 4.22 km/s;

MohoDepMin = 25 km; minimum crustal thickness expected in model sampling for Moho depth.

MohoDepMax = 70km; minimum crustal thickness expected in model sampling for Moho depth.

MohoInc = 1 km; Moho depth increment used in crustal thickness sampling.

2. Initial conditions

CrustAvgVel = CrustVelMin; use a 3.4 km/s as a starting point for velocity

MohoDep = MohoDepMin; use a 25 km as starting depth.

CrustUpVel = CrustAvgVelMin; the loops have fixed velocity but sample different depths, starting from 25 to 50 km.

3. Processing

While (CrustUpVel ≤ CrustAvgVelMax)

{ CrustAvgVel = CrustAvg VelMin. , make the starting values equal to the minimum search limits

While (CrustUpVel ≤ CrustAvgVelMax) to do similar approach as Last et al. (1997), the upper and lower crustal V_s are averaged by slowness to equal the average of the crust

$$\{ \text{CrustLowVel} = \frac{1}{\left(\frac{1}{\text{CrustAvgVel}}\right)^2 - \left(\frac{1}{\text{CrustUpVel}}\right)}$$

If (CrustLowVel ≤ CrustLowCutoff and CrstLowVel ≥ CrstUpVel)

While

{ (MohoDep ≤ MohoDepMax) , to do similar approach as Last et al. (1997), will make the upper and lower crust have same thickness

$$\text{CrustUp} = \frac{\text{MohoDep} - \text{SedmntCrust}}{2} + \text{SedmntCrust}$$

CrustUpVp = CrustUpVel * $\sqrt{\frac{2 * (\text{CrustPR} - 1)}{2 * \text{CrustPR} - 1}}$, Calculate appropriate desnity values for associate velocities/Poisson's ratios

$$\text{CrustLowVp} = \text{CrustLowVel} * \sqrt{\frac{2 * (\text{CrustPR} - 1)}{2 * \text{Crustpr} - 1}}$$

$$\text{MantleVp} = \text{MantleUpVel} * \sqrt{\frac{2 * (\text{MantlePR} - 1)}{2 * \text{MantlePR} - 1}}$$

$$\text{CrustUpRho} = \text{CrustUpVp} * 0.32 + 0.77$$

$$\text{CrustLowRho} = \text{CrustLowVp} * 0.32 + 0.77$$

$\text{MntlRho} = \text{Mntl} * 0.32 + 0.77$; make corresponding velocity models. More detailed in script **b)**

$\text{MohoDep} = \text{MohoDep} + \text{MohoInc}$, the intermediate loop finishes here. All velocity – depth models were created for velocity starting point $\text{CrustAvgVel} = \text{CrustAvgVelMin}$, $\text{CrustAvgVel} = 3.4$ km/s.

$\text{MohoDep} = \text{MohoDepMin}$, The Moho depth is reset to start a new loop where:

$$\text{CrustAvgVel} = \text{CrstAvgVel} + \text{CrustAvgVelInc}$$

$\text{CrustUpVel} = \text{CrstUpVel} + \text{CrustUpVelInc}$, When the CrustAvgVelMax (3.9 km/s) is achieved starting from CrustAvgVelMin (3.4 km/s) using a 0.05 km/s velocity increment.

Close out; all intermediate loops were executed and velocity – depth models created.

a) Open a list of models and generate predicted velocity dispersion

```
For (i = 0; i < table; i ++)  
open ( Table A, model)  
open (out, xy_files)  
dosrf (components dispersion vel for a given period)  
sac2xy (get xy value from sac files)  
close
```

b) Test if dispersion models fall within bootstrap error

open (Table B, xy_files) , for each generated models, see if the predicted values all fall within the bootstrap error

for (j = 0; j < table; j ++)

open (Table C, xy_files)

open (Table D, dispData)

OOB = 0 , initialize flag

for (k = 0; k < table; k++)

if (per = perD

phsVelLimUp = pshVelD + error

phsVelLimLow = pshVelD – error

if ($\text{phsVel} \leq \text{phsVelLimUp}$ and $\text{phsVel} \geq \text{phsVelLow}$) , value falls within specified phase velocity range

else OOB = 1, value fallen outside of the range, flag this as a bad model

if OOB = 0, value fallen inside of the range, flag as a good model and all data points are acceptable.

close (tabled) , For all accepted models, create a table of velocity with correspondent Moho depths and close de processing.

Script 2 – matchSRF

```
#!/usr/bin/env perl -w
```

```
# This script goes through the models that fit the dispersion data (within error) and determines  
# the associated, predicted Sp-S time. Used to find ones with closest timing to the stacked  
# SRF for the examined station
```

```
# NOTE: user must specify list of events so that average ray parameter / slowness for the stack  
# can be determined AND the timing of Moho conversion on the stack must be provided for  
# comparison.  
# Additionally - an error is specified (which brackets how closely the timing must be matched by  
# synthetics to be considered a match to the data. I chose this error based on the spread of  
# timings in the Moho conversion peak from the different events used in the stack.
```

```
$evtList = $ARGV[0]; # Name of the SRF list (specify when script runs)  
$SRF = STATSRLF; # Timing of Moho conversion  
$error = STATERRSRF; # Time error
```

```
$MTlimitU = $SRF + $error;  
$MTlimitD = $SRF - $error;
```

```
# First determine the average ray parameter to use for calculations  
system ("getValues.csh $evtList");  
open(TABLEA,"tempOut");  
@tablea = <TABLEA>;  
$pTotal = 0;  
$pPTotal = 0;  
$count = 0;  
print "\n";  
for ($i = 0; $i < @tablea; $i++) {  
    chomp ($tablea[$i]);  
    ($depth,$distance) = (split /\s+/, $tablea[$i])[3,4];
```

```
# adjust some values so not in scientific notation  
$distanceM = substr($distance,0,8);  
$distanceE = substr($distance,11,1);  
$dist = $distanceM * (10**$distanceE);
```

```
$depM = substr($depth,0,8);  
$depE = substr($depth,11,1);  
$dep = ($depM * (10**$depE)) / 1000;
```

```
# Now, use that distance and depth to figure out the corresponding slowness  
system ("taup_time -mod iasp91 -h $dep -deg $dist -ph S > slowOut");  
open(TABLET,"slowOut");  
@tablet = <TABLET>;
```

```

chomp ($tablet[5]);
($rayParam) = (split /\s+/, $tablet[5])[5];
close(TABLET);
unlink("slowOut");

# Now, use that distance and depth to figure out the corresponding slowness
system ("taup_time -mod iasp91 -h $dep -deg $dist -ph P > slowOutP");
open(TABLETP, "slowOutP");
@tabletP = <TABLETP>;
chomp ($tabletP[5]);
($rayParamP) = (split /\s+/, $tabletP[5])[5];
close(TABLETP);
unlink("slowOutP");
unlink("tempOut");

# Convert rayParam in s/deg to slowness in s/km
$slow = $rayParam / 111;
$slowP = $rayParamP / 111;

# add this value to running total to compute average
$spTotal = $spTotal + $slow;
$spPTotal = $spPTotal + $slowP;
$count = $count + 1;
}
# Compute average slowness/ray parameter
$spAvg = $spTotal / $count;
$spPAvg = $spPTotal / $count;
print "Average S ray parameter: ", $spAvg, "\n";
print "Average P ray parameter: ", $spPAvg, "\n";
print "Moho Time on SRF: ", $SRF, "\n";

# Get list of models that fit dispersion data (within error)
system ("ls *.xy > XY_LIST");

open(TABLEB, "XY_LIST");
@tableb = <TABLEB>;

for ($j = 0; $j < @tableb; $j++) {
    chomp ($tableb[$j]);
    ($xyFile) = (split /\s+/, $tableb[$j])[0];
    # Modified by Cristo.
    #($n1, $n2, $n3) = (split /\./, $xyFile)[0,1,2];
    #($send) = substr($n3, 0, 6);
    # $velName = $n1 . "_" . $n2 . "_" . $send;
    ($velName) = (split /\./, $xyFile);

```

```

# open each model file to get velocity info
open(TABLEC,$velName);
@tablec = <TABLEC>;
chomp ($tablec[1]);
($IceVelP,$IceVelS,$IceThck) = (split /\s+/, $tablec[1])[2,3,5];
chomp ($tablec[2]);
($CrstUpVelP,$CrstUpVelS,$CrstUpThck) = (split /\s+/, $tablec[2])[2,3,5];
chomp ($tablec[3]);
($CrstLowVelP,$CrstLowVelS,$CrstLowThck) = (split /\s+/, $tablec[3])[2,3,5];
close(TABLEC);

# Use P,S velocities to compute vertical slownesses for each layer
$IceSlowP = ((1 / ($IceVelP ** 2)) - ($pPAvg ** 2)) ** 0.5;
$IceSlowS = ((1 / ($IceVelS ** 2)) - ($pAvg ** 2)) ** 0.5;
$CrstUpSlowP = ((1 / ($CrstUpVelP ** 2)) - ($pPAvg ** 2)) ** 0.5;
$CrstUpSlowS = ((1 / ($CrstUpVelS ** 2)) - ($pAvg ** 2)) ** 0.5;
$CrstLowSlowP = ((1 / ($CrstLowVelP ** 2)) - ($pPAvg ** 2)) ** 0.5;
$CrstLowSlowS = ((1 / ($CrstLowVelS ** 2)) - ($pAvg ** 2)) ** 0.5;

# Use slownesses and layer thicknesses to compute layer travel times
$IceTime = ($IceSlowS - $IceSlowP) * $IceThck;
$CrstUpTime = ($CrstUpSlowS - $CrstUpSlowP) * $CrstUpThck;
$CrstLowTime = ($CrstLowSlowS - $CrstLowSlowP) * $CrstLowThck;

# And sum up to find the total, predicted time for the Moho for this model
$time = $IceTime + $CrstUpTime + $CrstLowTime;
#print $velName,": ", $time, "\n";

# Now compare this predicted time to the Moho conversion time on the real data
# (within specified error)
if ( ($time <= $MTlimitU) & ($time >= $MTlimitD) ) {
    # this model fits both dispersion and SRF within error so
    # we'll keep this model's files
    #print "Model ", $velName, " fits both dispersion and SRF: ", $time, "\n";
} else {
    # while this model fits dispersion, it doesn't fit the SRF
    # so we will remove associated files
    system ("/bin/rm $xyFile $velName");
}
}
close(TABLEB);
unlink("XY_LIST");

```

Script description

This script goes through all models that fit in the dispersion data and determines the associated, predicted Sp – S arrival time. Used to find ones with closest timing to the stacked SRF for examined station

eventList = ARGV [0]; input while running the program

srf = statsrf, timing of the Moho conversion arrival

error = staterror , time error estimated using bootstrap method

MTlimitU = srf + error

MTlimitD = srf – error

get.values.csh take ray parameters values, from SAC header, to use for calculations system

table a = TABLE A (output)

pTotal = 0

pPTotal = 0

count = 0

for (i = 0, i < table a; i ++)

Calculate ray parameter using Tau P (S)

Calculate ray parameter using Tau P (P)

Saved as slow S and slow P respectively

Convert ray parameter into slowness, using distance and depth to estimate the correspondent slowness

Slow = ray param / 111 , convert ray parameter in s/deg to slowness in s/km

Slow P = ray param / 111

pTotal = pTotal + slow

pPTotal = pPTotal + slow P

Count = count ++

pAvg = pTotal / count , compute average S ray parameter

pPAvg = pPtotal/ count , compute average P ray parameter

Open (Table B, xy_List) , Get dispersion models that fall within the bootstrap error

Table b = Table B

For (j = 0; j < Table B; j ++)

Open (table C, VelName) open each model file to get velocity information

a) Use P, S velocities to compute vertical slownesses for each layer

$$SedimntSlowP = \sqrt{\frac{1}{(sedimntVelP)^2} - (pPAvg)^2}$$

$$SedimntSlowS = \sqrt{\frac{1}{(sedimntVelS)^2} - (pAvg)^2}$$

$$CrustUpSlowP = \sqrt{\frac{1}{(CrustUpVelP)^2} - (pPAvg)^2}$$

$$CrustUpSlowS = \sqrt{\frac{1}{(CrustUpVelS)^2} - (pAvg)^2}$$

$$CrustLowSlowP = \sqrt{\frac{1}{(CrustLowVelP)^2} - (pPAvg)^2}$$

$$CrustLowSlowS = \sqrt{\frac{1}{(CrustLowVelS)^2} - (pAvg)^2}$$

b) Use slownesses and layer thickness to compute layer travel times

$$\text{sedimentTime} = (\text{sedimntSlowS} - \text{sedimntSlowP}) * \text{sedimntThck}$$

$$\text{crustUpTime} = (\text{CrustUpSlowS} - \text{CrustUpSlowP}) * \text{crustUpThck}$$

$$\text{crustLowTime} = (\text{CrustLowSlowS} - \text{CrustLowSlowP}) * \text{crustLowThck}$$

c) sum partial times to compute total time

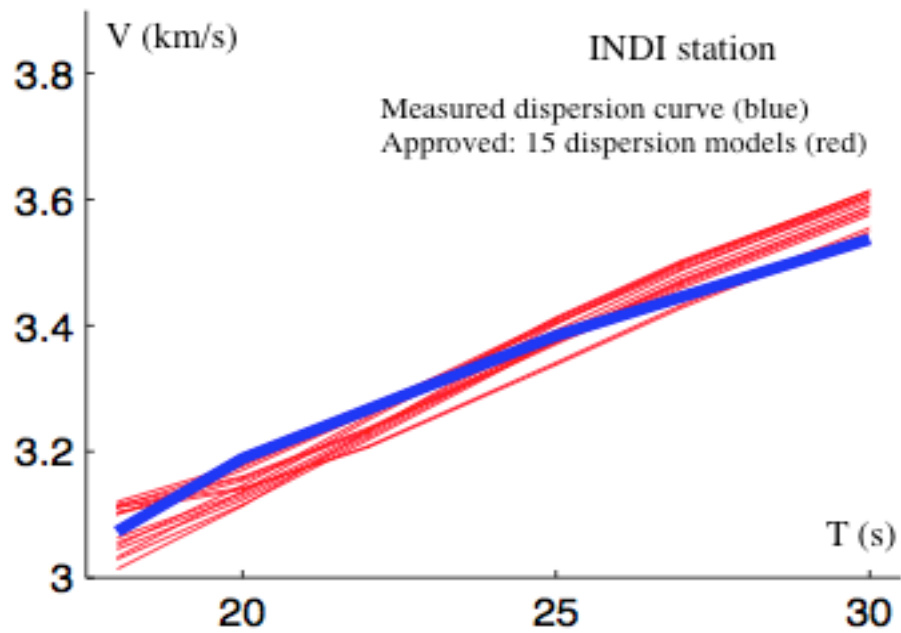
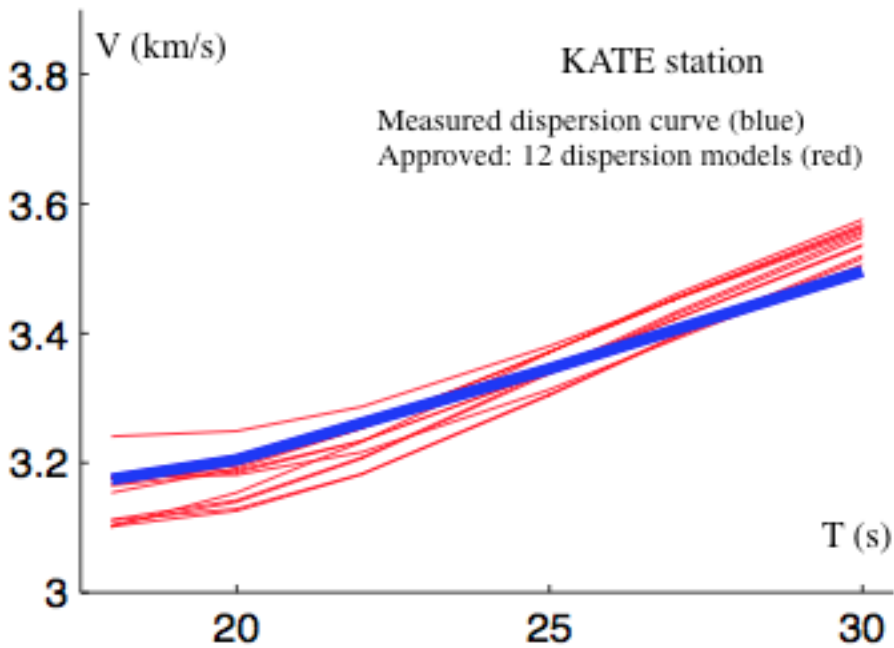
$$\text{time} = \text{SedmntTime} + \text{CrustUpTime} + \text{CrustLowTime}$$

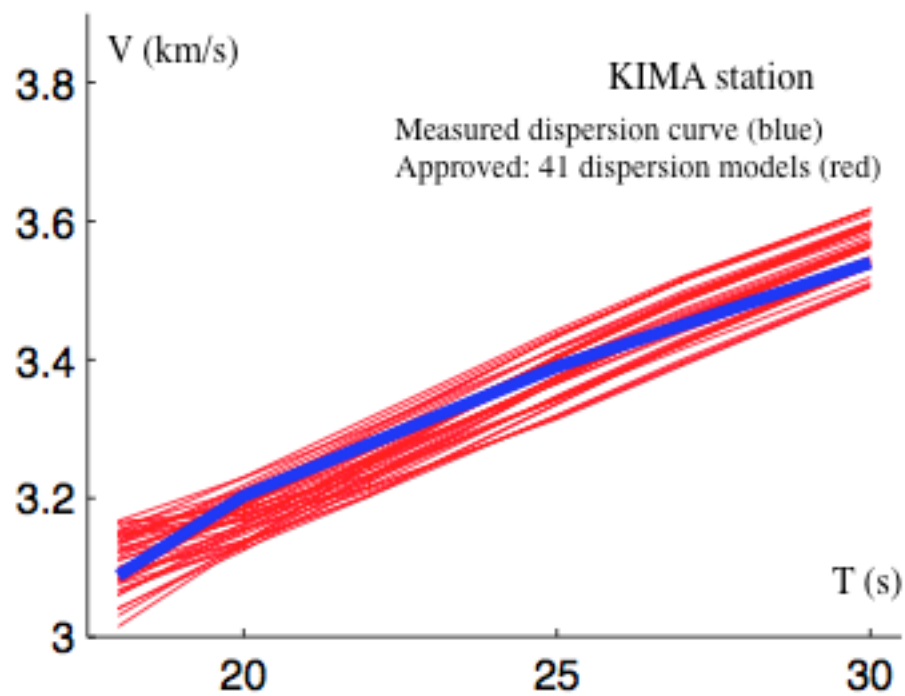
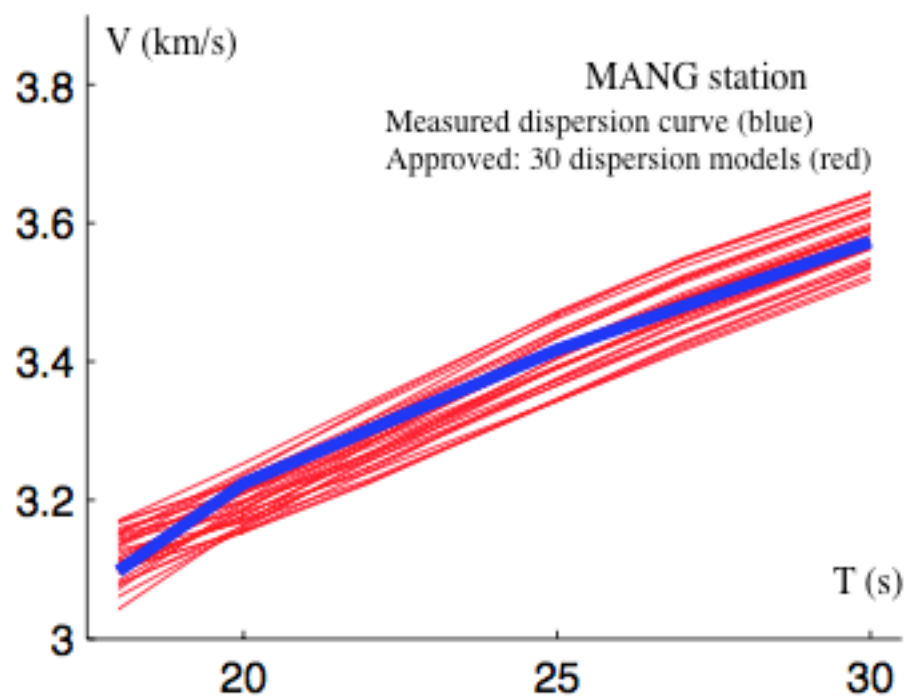
d) Compare observed and calculated Moho conversion arrival time

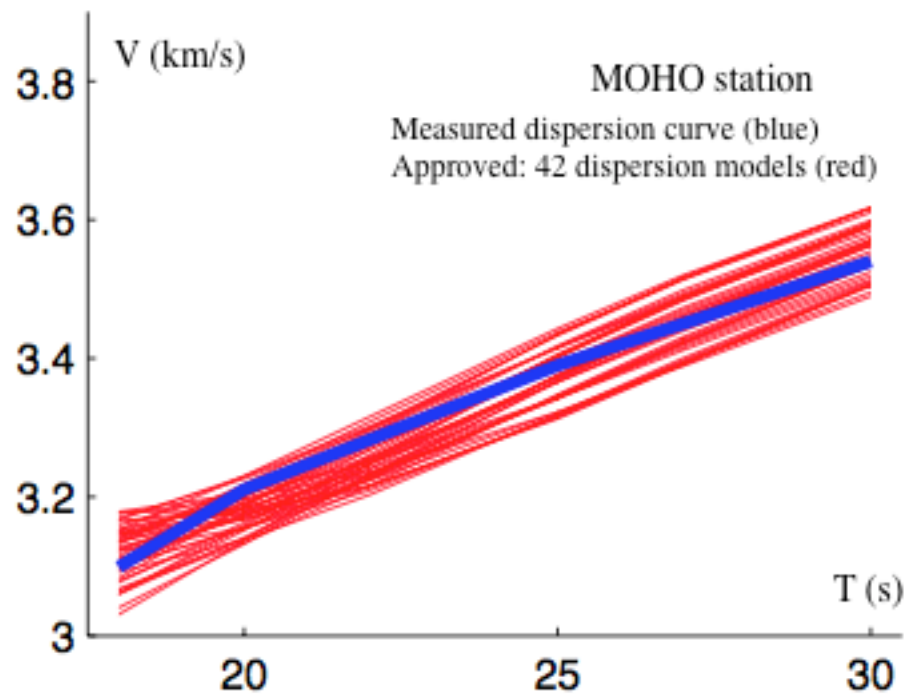
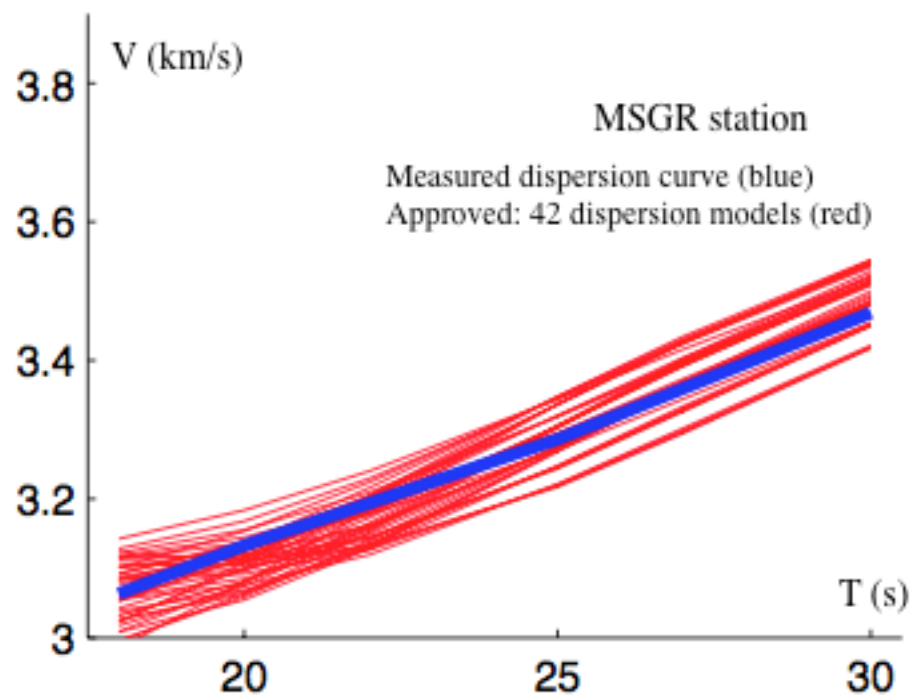
If $(\text{MTlimitD} \leq \text{time} \leq \text{MTLimitU})$

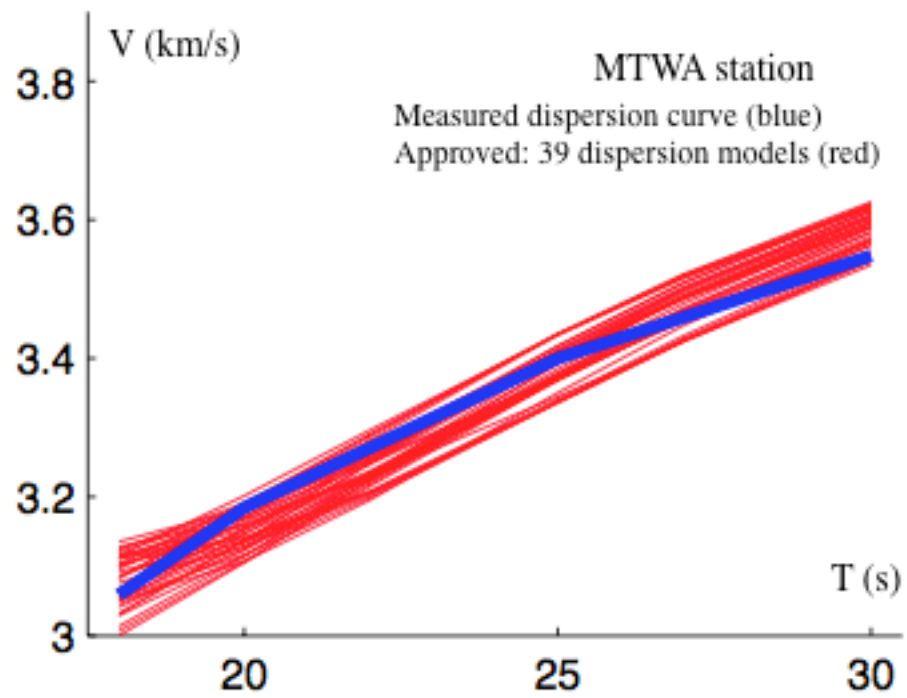
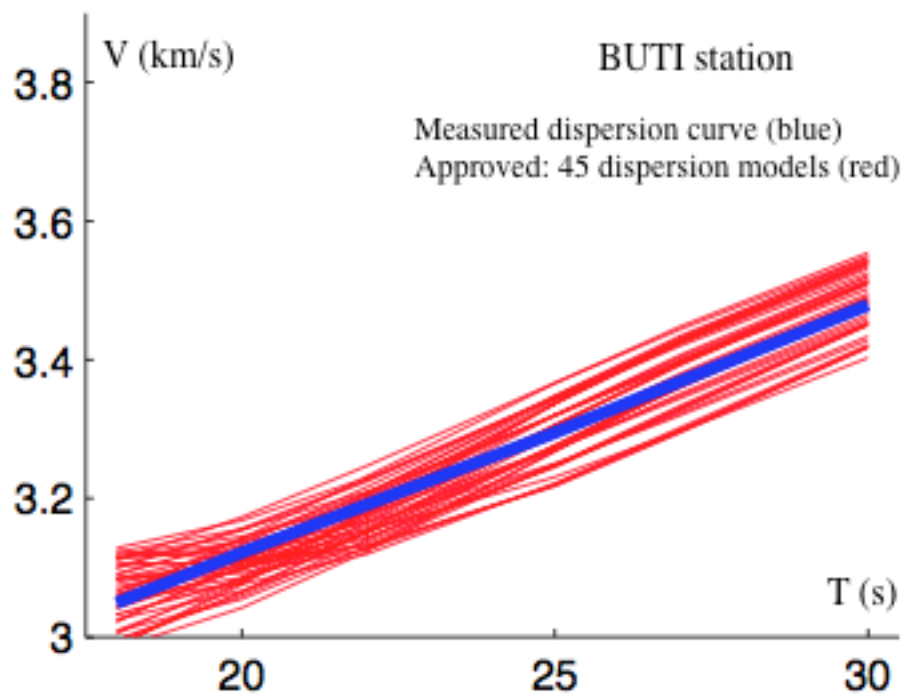
if model is good , those that fall within the error are kept and those that fall outside the error are deleted.

Appendix G. *Plots of dispersion curves, measured dispersion curve (blue) and dispersion curves modeled using 200 iterations (red).*



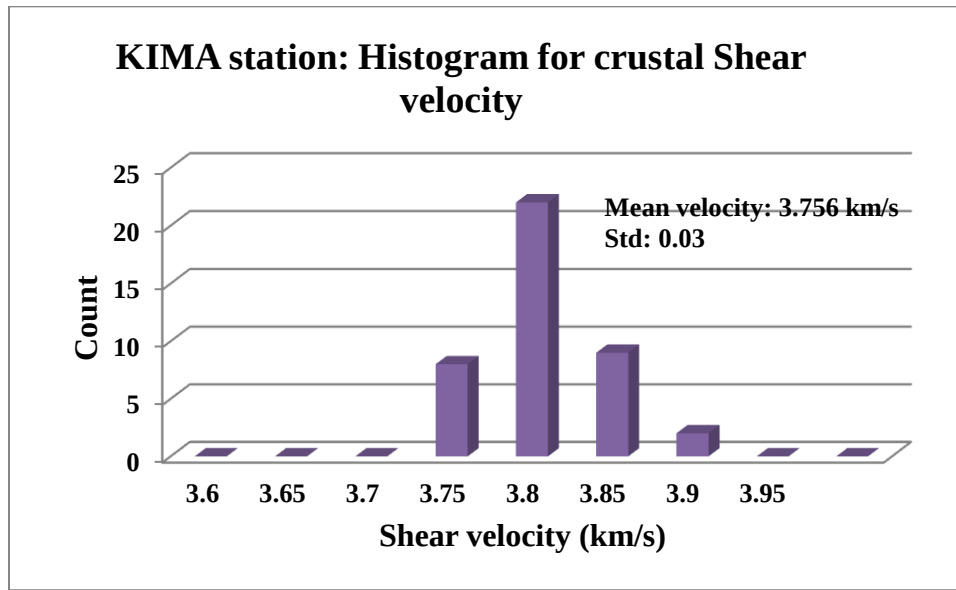
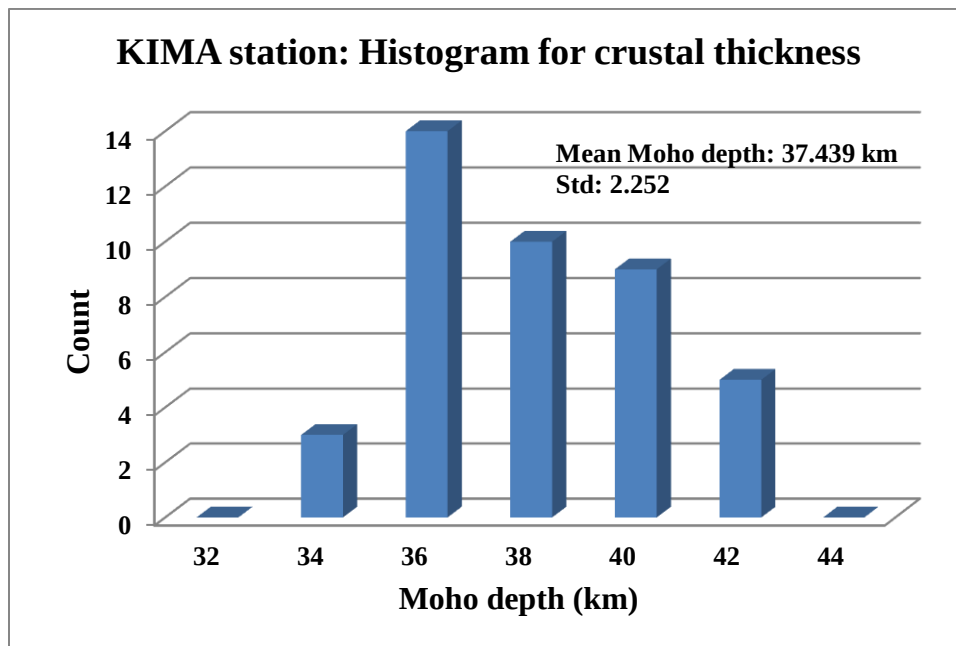




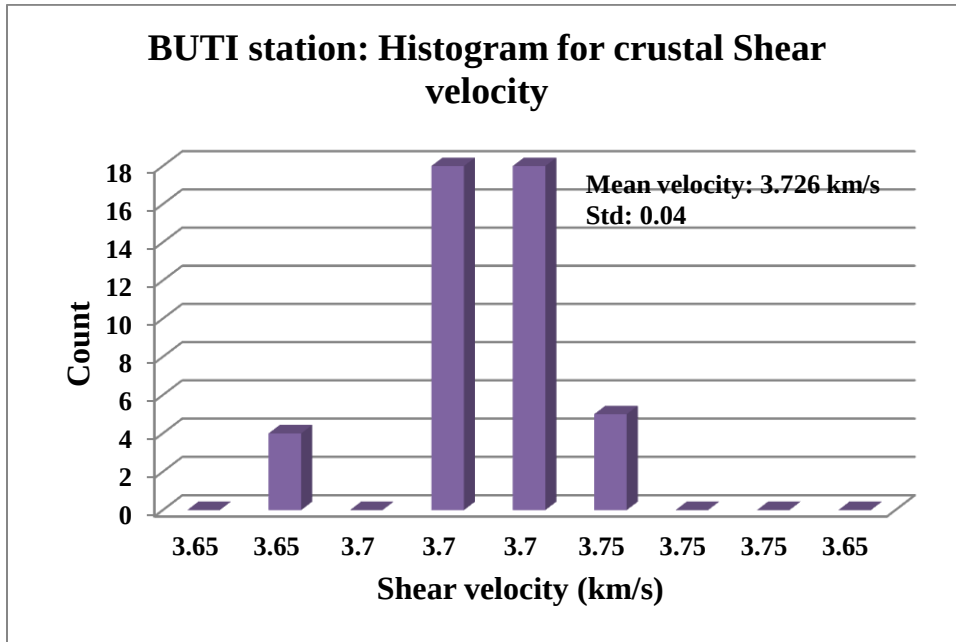
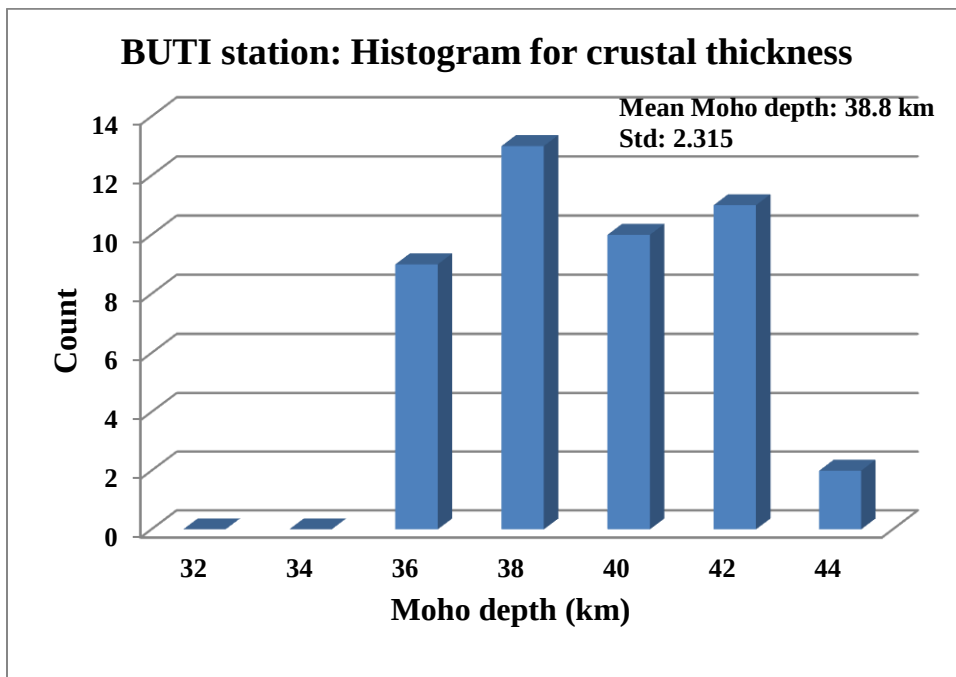


Appendix H. Histograms for eight stations (among all eleven stations) those were found results for Moho depth (bin size 2km) and crustal shear velocity (bin size 0.05 km/s).

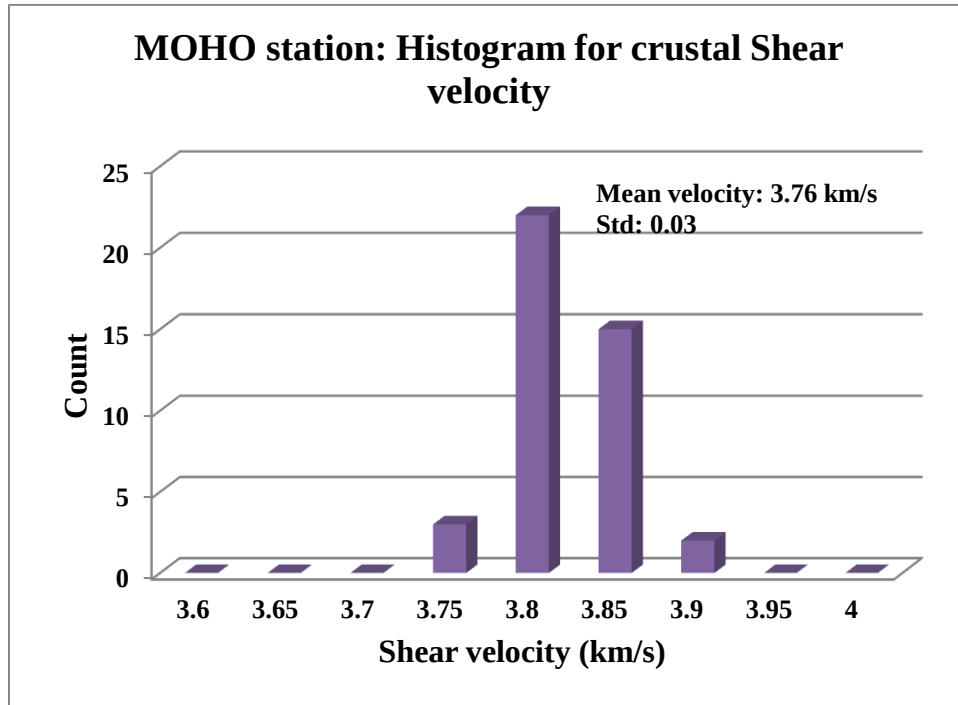
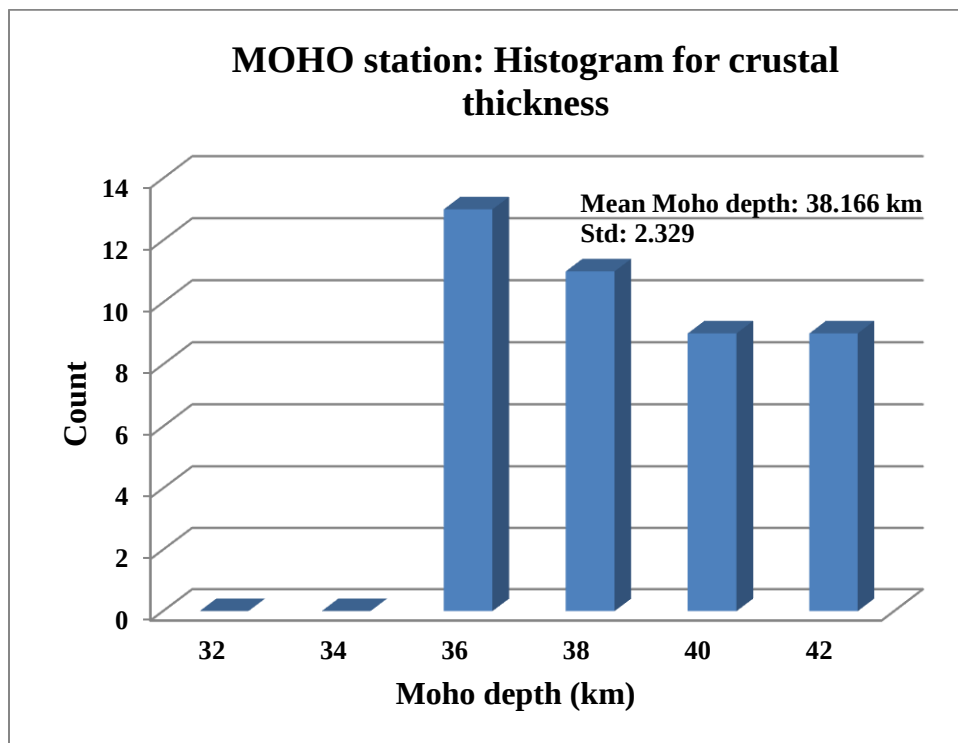
1. KIMA station



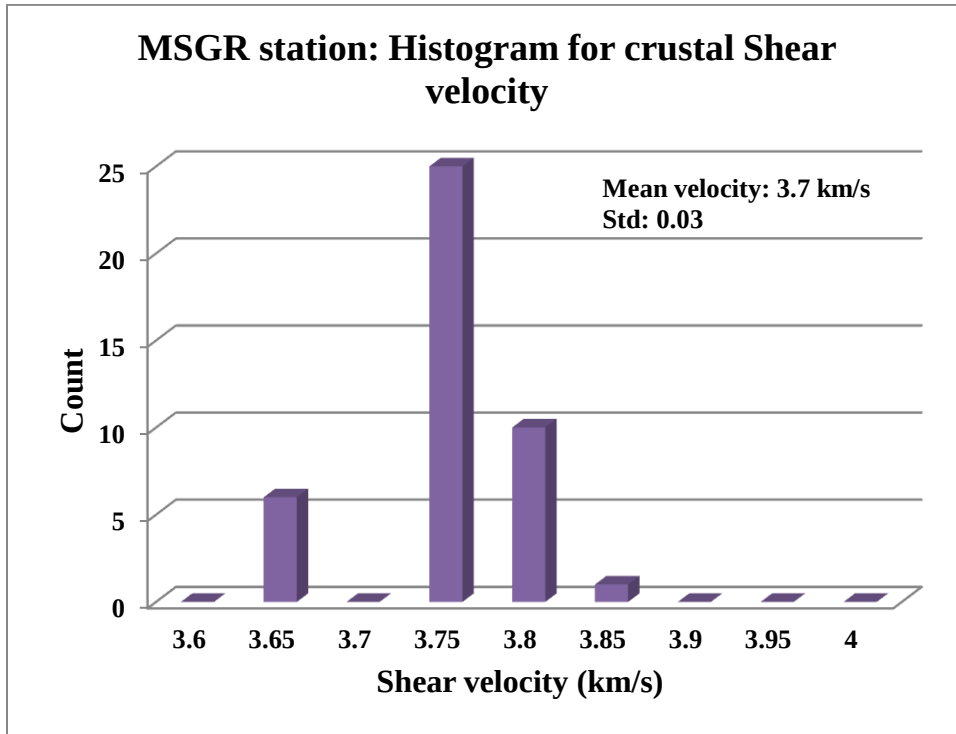
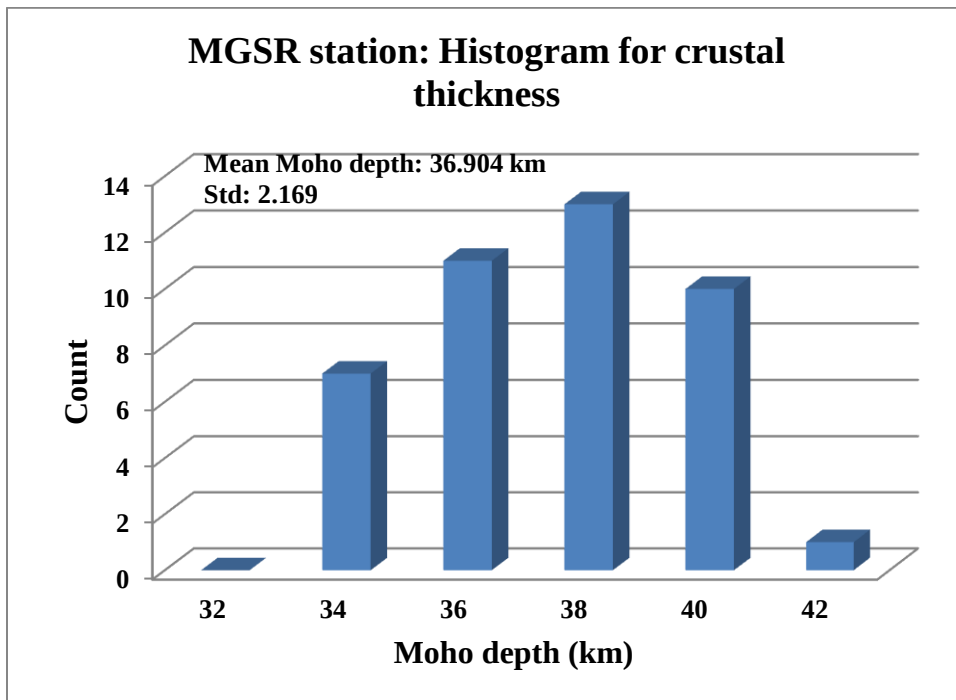
2. BUTI station



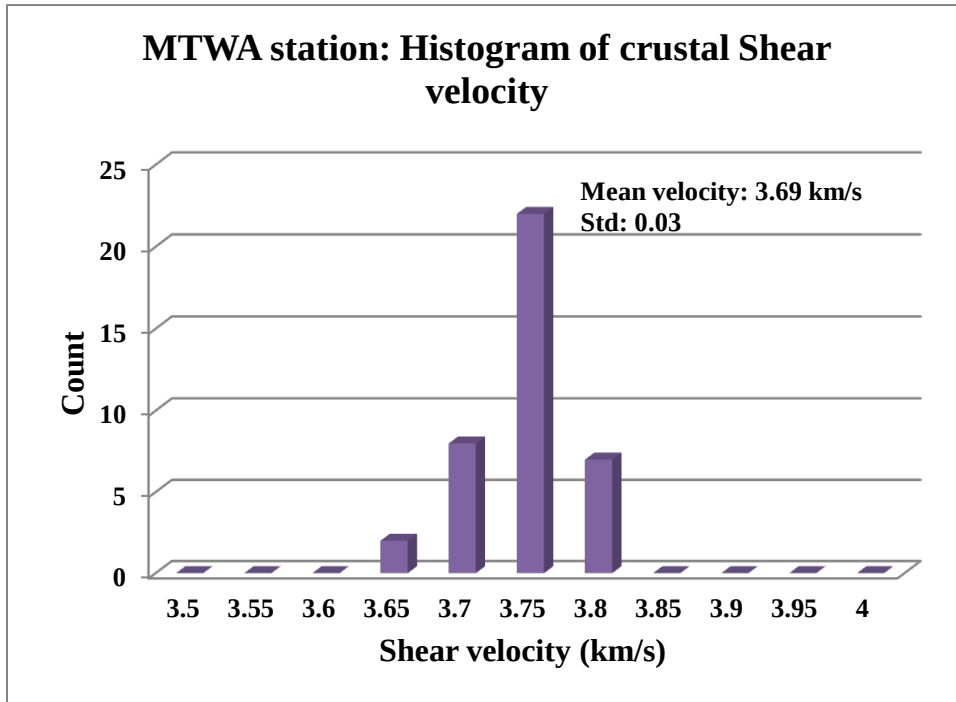
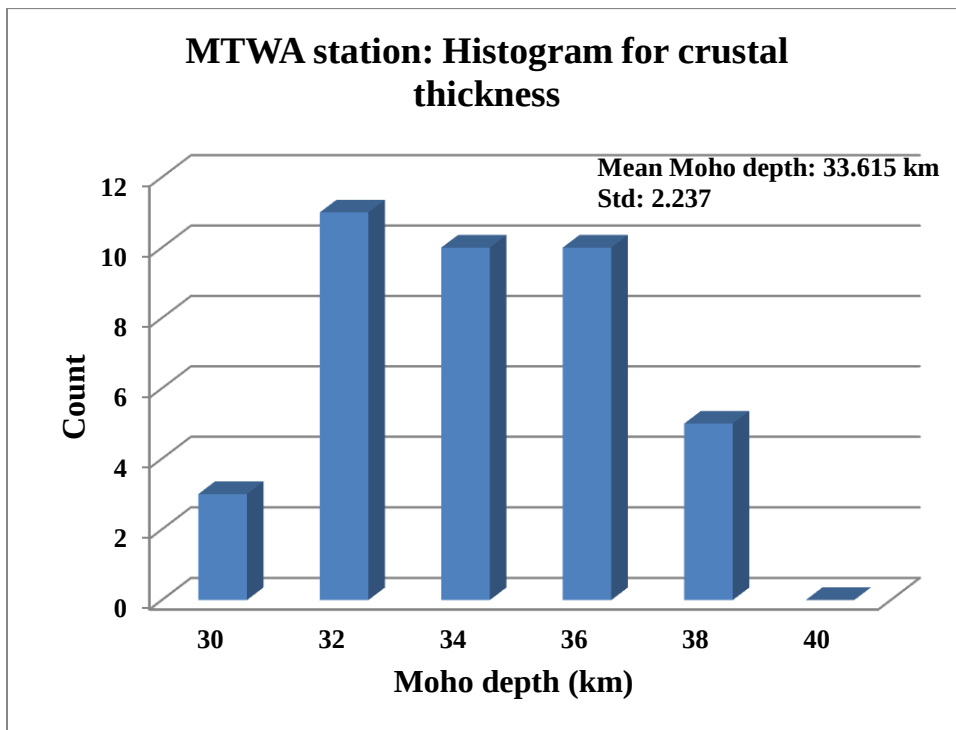
3. MOHO station



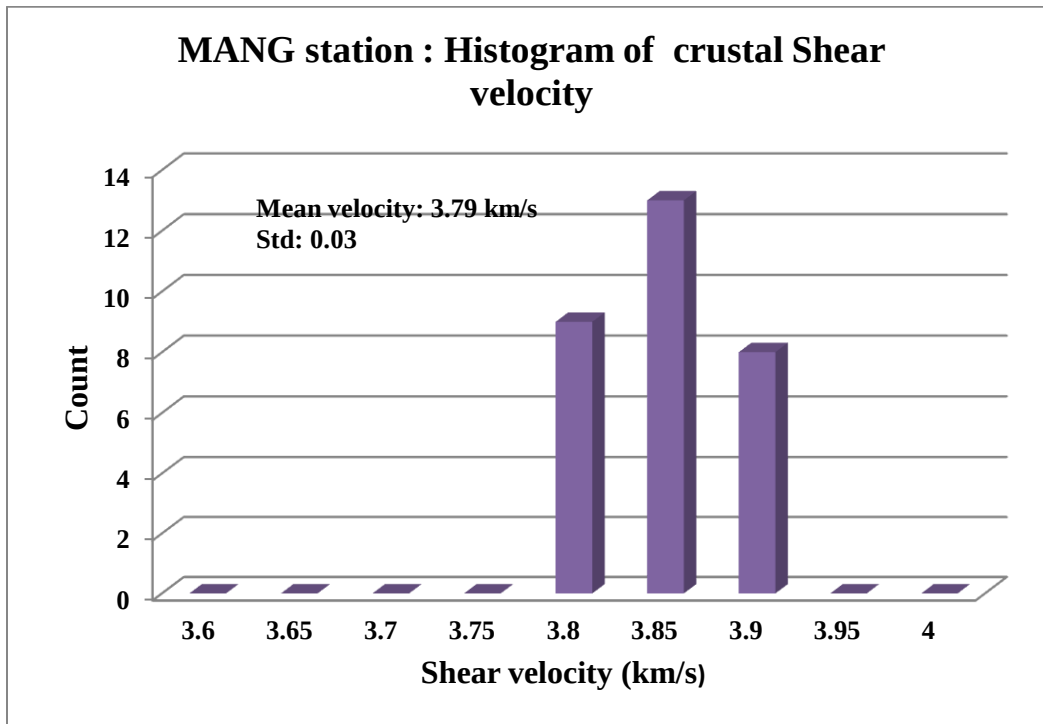
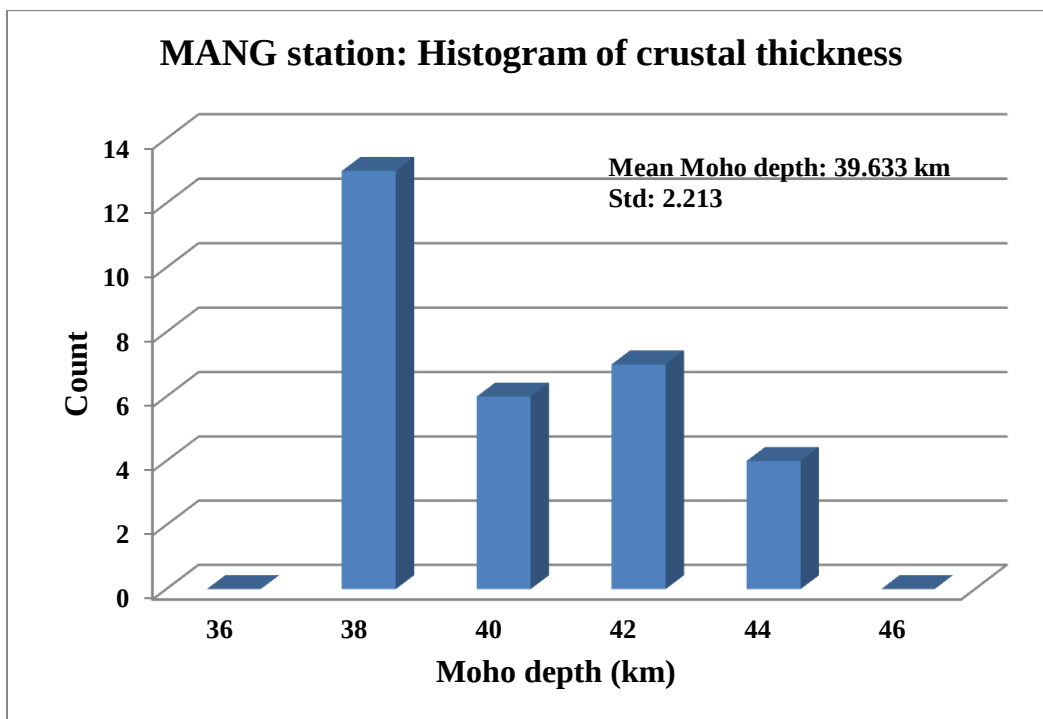
4. MSGR station



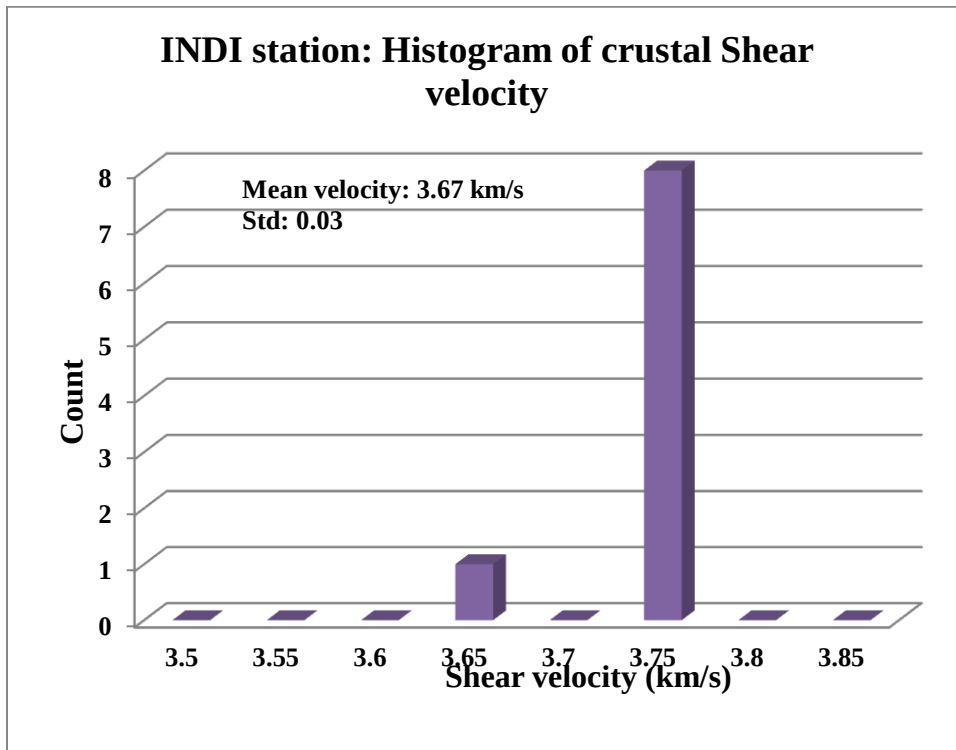
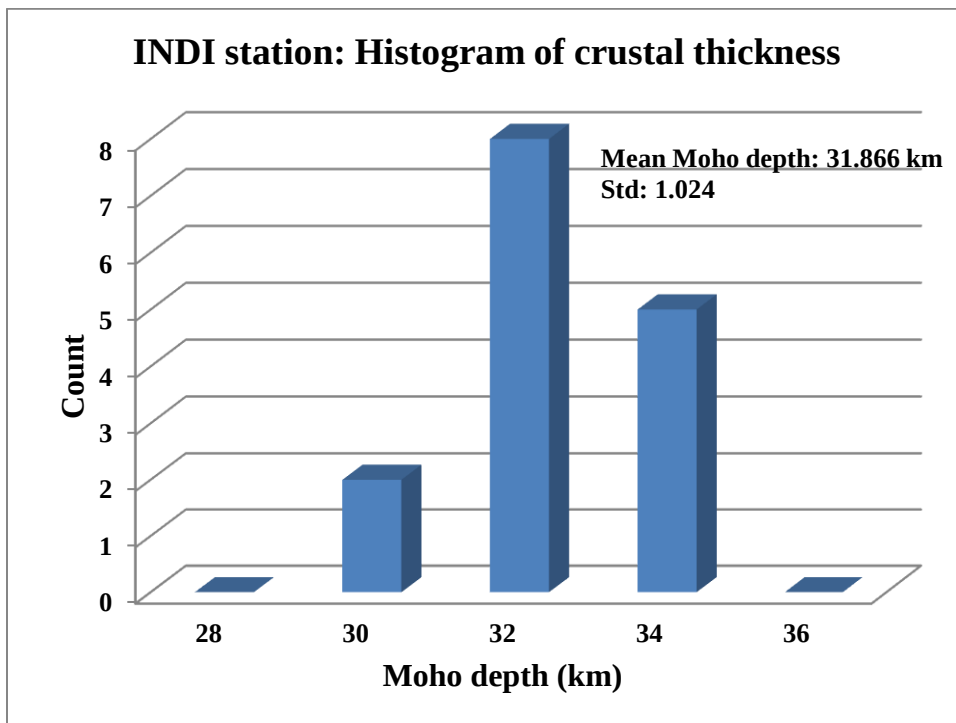
5. MTWA station



6. MANG station



7. INDI station



8. KATE station

