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SEISMIC STRUCTURE AND ANISOTROPY IN SOUTHERN  
AFRICA'S LITHOSPHERE: CONSTRAINTS FROM  
BROAD-BAND SURFACE-WAVE DISPERSION

JOANNE M.-C. ADAM

A dissertation presented to the  
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School of Natural Sciences, Geology Department,  
Trinity College Dublin,  
Dublin, Ireland



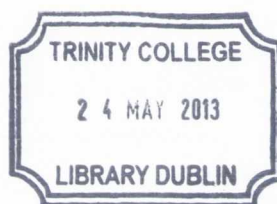
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2013





Thesis 9883

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## DECLARATION

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I hereby acknowledge the authorship of this dissertation and declare that it is my own work except as follows.

Chapter three is a published paper and has been prepared in cooperation with Dr. Sergei Lebedev. I analysed the data, performed the research and wrote the first draft of the manuscript. Dr. Lebedev edited the manuscript. The contents of chapter three have been included in this dissertation without any modification of the published paper.

Most of the results presented in chapter five have been used as a major part of a review paper written by Dr. Lebedev. Dr. Lebedev and myself planned the research. I analysed the data, performed the research and wrote the text of this chapter.

I declare that this dissertation has not been submitted as an exercise for a degree at this or any other university.

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## SUMMARY

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Southern Africa contains some of the oldest crust on Earth, with the Archean Kaapvaal Craton bordered to the North by the Archean-Paleoproterozoic Limpopo belt. To understand the mechanisms of deformation of the Archean craton during and since their formation, I mapped the anisotropic structure of their lithosphere that preserved a record of this deformation.

I measured surface-wave dispersion curves between station pairs across this region and merged them into subsets characterizing four sub-regions within southern Africa. I obtained accurate dispersion curves for each subset in broad period ranges and inverted them for robust profiles of shear velocity and azimuthal anisotropy spanning the depth range from the upper-crust to the asthenosphere.

The mantle-lithosphere beneath the Kaapvaal Craton and Limpopo Belt is characterized by high velocities down to  $\sim 200$  km, confirming observations from previous studies. I showed evidence of a low-velocity zone between 150 and 250 km, weak in the Limpopo Belt and northern Kaapvaal Craton and strong in the southern Kaapvaal Craton. The more pronounced low velocity zone in the southern Kaapvaal Craton probably indicates higher temperature in the asthenosphere and lower lithosphere. This could account, at least in part, for the higher topographic elevation of the southern compared to the northern Kaapvaal Craton.

Measuring the variation of azimuthal anisotropy with depth, I observed its layering with different directions of fast propagation in the upper and lower mantle lithosphere and asthenosphere. I detected an anisotropy boundary at  $\sim 90$  km, which can account for the enigmatic discontinuity at this depth inferred from receiver function studies.

I also resolved the long-standing debate on the origin of the shear-wave splitting in this region. One previous theory postulated that the splitting originated in the lithosphere and was due to the deformation during the Archean collisions between parts of the Kaapvaal Craton and the Zimbabwe Craton. The other theory proposed that the splitting originated in the asthenosphere and was due to the deformation associated with recent plate motion. I show that the fabric in the lower lithosphere of the northern Kaapvaal Craton and Limpopo Belt was parallel to the Archean-Paleoproterozoic sutures, whereas in the southern Kaapvaal Craton it is not. Reconciling the two models, I show that SKS splitting must originate both in the lithosphere and asthenosphere, while the fabric in central and western Kaapvaal



lithosphere is not suture-parallel and probably predates their collision.

Finally, I investigated under what conditions surface waves can resolve the depth of the Moho and presented an efficient strategy for the inversion of surface wave dispersion curves for the Moho-depth. I applied the technique to data from southern Africa and resolved the Moho depth with accuracy.

*It is wrong to think that the task of  
physics is to find out how Nature is.  
Physics concerns what we say about Nature.*

*— Niels Bohr —*

---

## ACKNOWLEDGMENTS

---

I would like to thank my supervisor, Sergei Lebedev, for giving me the opportunity to work with him and accomplish this Ph.D. During the last four years, I always found an open door and open mind to discuss many aspects of my research. I am grateful for his guidance, advice and encouragement.

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## LIST OF EQUATIONS

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## LIST OF ACRONYMS

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|                 |                                                                                          |
|-----------------|------------------------------------------------------------------------------------------|
| AK135           | Earth reference model published by Kennett <i>et al.</i> (1995)                          |
| CHAMP           | CHAllenging Mini-satellite Payload                                                       |
| GEMMA           | GOCE Exploitation for Moho Modeling and Applications                                     |
| GOCE            | Gravity field and steady-state Ocean Circulation Explorer                                |
| GRACE           | Gravity Recovery and Climate Experiment                                                  |
| GSN             | Global Seismographic Network                                                             |
| IRIS            | Incorporated Research Institutions for Seismology                                        |
| JWKB            | Jeffreys-Wentzel-Kramers-Brillouin                                                       |
| K.C.            | Kaapvaal Craton                                                                          |
| LAB             | Lithosphere Asthenosphere Boundary                                                       |
| LVZ             | Low Velocity Zone                                                                        |
| MOHO            | Mohorovičić discontinuity (discontinuity between the crust and the mantle)               |
| P               | Compressional body wave                                                                  |
| P <sub>n</sub>  | P wave that is refracted at the Moho                                                     |
| S               | Shear body wave                                                                          |
| SASE            | South African Seismic Experiment                                                         |
| S <sub>H</sub>  | Horizontally polarized shear body wave                                                   |
| S <sub>V</sub>  | Vertically polarized shear body wave                                                     |
| SKS             | Wave that propagates through the mantle as a S wave<br>and in the outer-core as a P wave |
| STS2            | Streckeisen 2 seismometer                                                                |
| TML             | Thabazimbi-Murchison Lineament                                                           |
| V <sub>P</sub>  | Compressional velocity                                                                   |
| V <sub>S</sub>  | Shear velocity                                                                           |
| V <sub>SH</sub> | Horizontal shear velocity                                                                |
| V <sub>SV</sub> | Vertical shear velocity                                                                  |
| 1-D             | One-dimension                                                                            |

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## LIST OF SYMBOLS

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| SYMBOL    | SI UNIT                          | DENOTATION                                         |
|-----------|----------------------------------|----------------------------------------------------|
| $\rho$    | $\text{kg.m}^{-3}$               | density                                            |
| $\kappa$  | $\text{km.m}^{-1}.\text{s}^{-2}$ | bulk modulus                                       |
| $\mu$     | $\text{kg.s}^{-2}$               | rigidity                                           |
| $V$       | $\text{km.s}^{-1}$               | velocity                                           |
| $V_P$     | $\text{km.s}^{-1}$               | P velocity                                         |
| $V_S$     | $\text{km.s}^{-1}$               | S velocity                                         |
| $V_{SV}$  | $\text{km.s}^{-1}$               | SV velocity                                        |
| $V_{SH}$  | $\text{km.s}^{-1}$               | SV velocity                                        |
| $dV$      | $\text{km.s}^{-1}$               | velocity perturbation                              |
| $C$       | $\text{km.s}^{-1}$               | phase velocity                                     |
| $C_0$     | $\text{km.s}^{-1}$               | isotropic phase velocity                           |
| $dC$      | $\text{km.s}^{-1}$               | phase velocity perturbation                        |
| $U$       | $\text{km.s}^{-1}$               | group velocity                                     |
| $Q$       | $\emptyset$                      | quality factor                                     |
| $T$       | s                                | period                                             |
| $\omega$  | $\text{s}^{-1}$                  | frequency                                          |
| $d\omega$ | $\text{s}^{-1}$                  | frequency perturbation                             |
| $k$       | $\text{m}^{-1}$                  | wave number                                        |
| $dk$      | $\text{m}^{-1}$                  | wave number perturbation                           |
| $x$       | m                                | distance                                           |
| $t$       | s                                | time                                               |
| $w_i$     | $\emptyset$                      | weight on the phase velocity per azimuth i         |
| $n_i$     | $\emptyset$                      | number of phase velocity measurement per azimuth i |
| $N$       | $\emptyset$                      | number of data                                     |
| $F$       | $\emptyset$                      | F-value                                            |

| SYMBOL                                  | SI UNIT            | DENOTATION                      |
|-----------------------------------------|--------------------|---------------------------------|
| $A_1, A_2, A_3$ and $A_4$               | $\text{km.s}^{-1}$ | azimuthal anisotropy parameters |
| $A_{2\varphi}$ and $A_{4\varphi}$       | $\text{km.s}^{-1}$ | amplitude of anisotropy         |
| $\phi_{2\varphi}$ and $\phi_{4\varphi}$ | $^\circ$           | fast-propagation azimuths       |
| $\varphi$                               | $^\circ$           | azimuth                         |
| $\mathbf{C}_d$                          |                    | <i>a priori</i> error matrix    |
| $\mathbf{d}$                            |                    | data matrix                     |
| err                                     |                    | standard deviations             |
| $\mathbf{G}$                            |                    | sensitivity matrix              |
| $\mathbf{m}$                            |                    | model matrix                    |
| $\hat{\mathbf{m}}$                      |                    | estimated model matrix          |
| $\mathbf{R}$                            |                    | resolution matrix               |
| $\mathbb{Z}$                            |                    | set of integers                 |



Part I

INTRODUCTION AND THEORY

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## INTRODUCTION

---

The mechanisms of formation and stabilization of cratons, the ancient remnants of continents, are not well understood. Theories about the formation of continents are debated. One approach to image the deformation undergone by a continent is to study the structures of the lithosphere of the Archean Cratons in southern Africa. The anisotropic fabric within their lithosphere preserves signatures of their Archean formation and evolution.

The Kaapvaal Craton in southern Africa is one rare place in the world that retained Archean rocks and provided numerous xenoliths studies from kimberlite eruptions that allowed to constrain the thermal state as well as the composition of the mantle-lithosphere (Griffin *et al.*, 2003; Bell *et al.*, 2003). The Kaapvaal Craton has formed between 3.7 and 2.7 Ga (De Wit *et al.*, 1992; Eglington and Armstrong, 2004) and comprises four terrains: the Kimberley block (3.0-2.8 Ga), the Pietersburg block (3.0-2.8 Ga), the Witwatersrand block (3.8-3.1 Ga) and the Swaziland block (3.6-3.1 Ga). Results from the geochronologic study of Schmitz *et al.* (2004) showed that the Kimberley block and the Witwatersrand block converged and collided between 2.97 and 2.88 Ga, with probable subduction beneath the Kimberley block. The suture zone is now represented by the magnetic Colesberg lineament (Figure 1.1).

The Limpopo Belt separates the Kaapvaal Craton from the Zimbabwe Craton. It underwent deformation, including significant shear, during the collision between the Kaapvaal Craton and the Zimbabwe Craton (2.9-2.6 Ga) (De Wit *et al.*, 1992) in the Paleo-Proterozoic (~2.0 Ga) (De Wit *et al.*, 1992; Kamber *et al.*, 1995; Holzer *et al.*, 1998; Irvine *et al.*, 2001; Moser *et al.*, 2001; Schmitz *et al.*, 2004; Silver *et al.*, 2004). The Limpopo Belt has a geological structure that differs from the Kaapvaal and Zimbabwe Cratons (Van Reenen *et al.*, 1992; van Reenen *et al.*, 1992). The Limpopo Belt is divided into three main shear zones, trending in the East-West direction. The crust of the Limpopo Belt was thickened during the collision (Holzer *et al.*, 1999). Geophysical studies showed that its present crustal thickness is similar to the crustal thickness in the northern Kaapvaal Craton (Nguuri *et al.* (2001); Nair *et al.* (2006); Kwadiba *et al.* (2003); Kgaswane *et al.* (2009); Yang *et al.* (2008) and Section 5.5.3).

Lithospheric fabrics can preserve signatures of the deformation that may be detected as azimuthal anisotropy (azimuthal dependence of seismic speeds). In southern Africa,

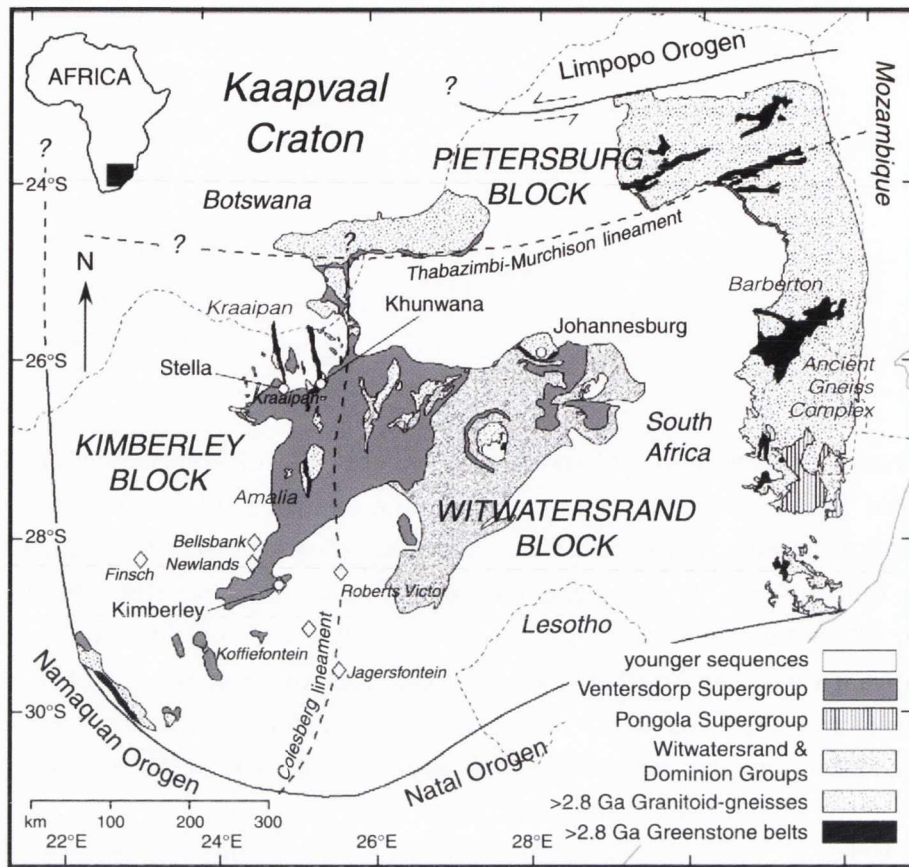


Figure 1.1.: Geographic map of southern Africa with main tectonic units and features labelled. White diamonds represent the location of mantle xenolith (from Schmitz *et al.*, 2004).

previous measurements of azimuthal anisotropy have been debated, with two main end-member theories put forward. One suggests the dominance of anisotropy in the lithosphere (Silver *et al.*, 2004) (due to the deformation from the Archean or Paleo-Proterozoic collision) and the other suggests the dominance of anisotropy in the asthenosphere (Vinnik *et al.*, 1995) (due to deformation associated with the recent plate motion). The main focus of this dissertation is to investigate the azimuthal dependence of the seismic wave velocities in the mantle to describe the deformation of the Kaapvaal Craton and the Limpopo Belt during their Archean and Paleo-Proterozoic evolution. The aim is to resolve the layering of anisotropy with depth but first raise questions regarding the resolvability of anisotropic layers with surface waves.

Chapter 2 presents a general description on the surface waves and the two techniques that are used to measure inter-station surface wave phase velocities. Chapter 3 presents in detail the inter-station phase velocity measurements. The region is divided into four sub-regions and phase velocity azimuthal anisotropy is measured for the four of them. The significance of the anisotropy is carefully examined using statistical tests to establish the robustness of the results.

Chapter 4 first focuses on the resolution of depth-profiles of shear velocity for each sub-region to then further investigate the depth stratification of anisotropy. Phase velocity anisotropic parameters are inverted to obtain accurate depth profiles of shear velocity and azimuthal anisotropy. To ensure the robustness of our results, the model-space is carefully explored and examined in order to determine whether trade-offs between the parameters may limit the resolution of anisotropy.

Finally, Chapter 5 examines the sensitivity of surface-waves to the depth of the Mohorovičić discontinuity and investigate the trade-offs between the crustal and upper-mantle velocity structures with the depth of the Moho. A strategy to resolve the depth of the Moho using surface-wave phase-velocity data is described with application to southern Africa.



## THEORY

Although this thesis focuses primarily on surface waves, it is relevant to define the main seismic-waves generated by an earthquake. In this chapter, I will briefly introduce body waves and focus on a description of the surface waves. I will also present the two techniques used to measure inter-station, surface-wave phase velocities used in the following chapters.

### 2.1 BODY WAVES AND SURFACE WAVES

When an earthquake occurs, it excites two main wave-types: body waves and surface waves. Body waves recorded at teleseismic distances propagate through the deep Earth (Figure 2.1) and are sensitive both to smooth three-dimensional variations in seismic wavespeeds and impedance contrasts (*i.e.* interfaces). Surface waves propagate along the surface of the Earth (Figure 2.1) and are sensitive to lateral heterogeneities within the lithosphere-asthenosphere depth range. They are dispersive and thus present a lot of information within their wave trains.

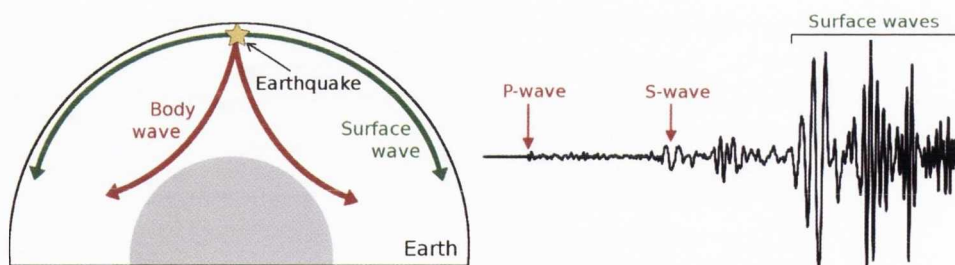


Figure 2.1.: Schematic representation of the different seismic wave types (left) and their recording on a seismogram (right) as they propagate into the Earth. Grey shade schematize the core.



2.1.1.1 *Body waves*

Two types of body waves can be distinguished, each with different particle motions. The first wave is the P-wave (also named compressional or longitudinal wave) for which the direction of oscillation is in the direction of the wave (Figure 2.2). Passing through the Earth, the P-wave generates volumetric change and shear and propagates at the velocity  $V_P$ :

$$V_P = \sqrt{\frac{\kappa + \frac{4}{3}\mu}{\rho}} \quad (2.1)$$

where  $\kappa$  is the bulk modulus,  $\mu$  is the rigidity and  $\rho$  is the density (Aki and Richards, 2002). The P-wave is the wave that travels the fastest and is the first arrival observed in a seismogram (Figure 2.1).

The second type of body wave is the S-wave (also named shear or secondary wave) for which the direction of oscillation is perpendicular to the direction of propagation of the wave if the medium is isotropic (see Figure 2.2). The particle motion is purely shear. The S-wave propagates through the Earth at the velocity  $V_S$  (Aki and Richards, 2002):

$$V_S = \sqrt{\frac{\mu}{\rho}}. \quad (2.2)$$

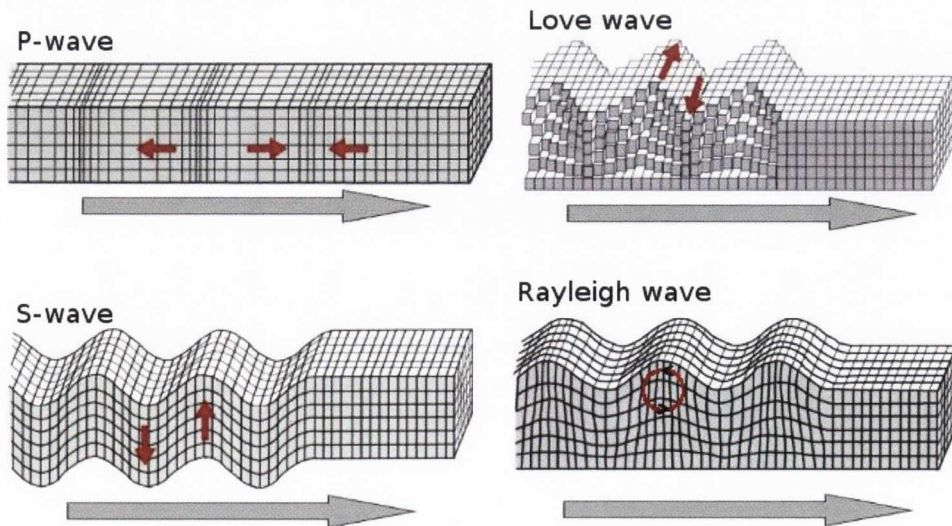


Figure 2.2.: Representation of the particle motion of body waves (P- and S-wave) and surface waves (Love and Rayleigh wave). The wave is propagating from the left to the right. Figure modified after University of Bergen (2005)

S-waves travel slower than the P-wave and then arrive later in a seismogram (Figure 2.1). In an anisotropic medium, the oscillation of the particles can be decomposed into two components: the SV and SH components (respectively for vertical and horizontal direction of particle motion) traveling at different speeds. This discrepancy can be measured and is called the radial anisotropy.

### 2.1.2 Surface waves

Surface waves (Rayleigh and Love waves) propagate in the horizontal direction (along the Earth's surface) and are approximately stationary in the vertical direction. In a layered medium (which is the case of the Earth's crust), interferences of multiples SH waves generate a Love wave. The Love wave is a guided wave at the surface of the Earth and is transversally polarized (see Figure 2.2). The particle motion of a Rayleigh wave is presented in Figure 2.2 and Figure 2.3. It can be considered as the interference of vertically-polarized P and SV waves.

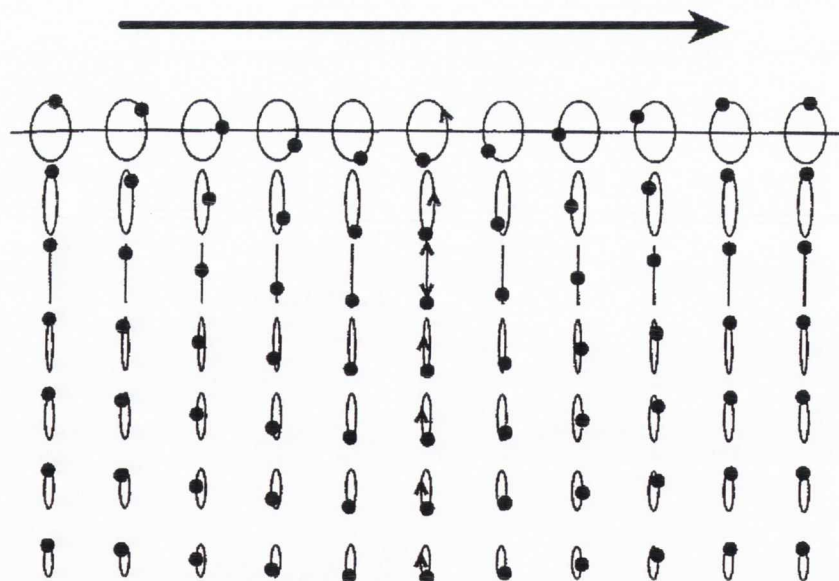


Figure 2.3.: Vertical cross-section of the particle motion for the fundamental mode of Rayleigh wave in an uniform half-space, propagating from left to right. One horizontal wave-length is shown in this picture. Dots are plotted at fixed time point. From Shearer (1999).

As seen in Figure 2.1, surface waves are, for most earthquakes, the most energetic waves, compared to body waves (the exceptions are deep earthquakes, which excite only very long period surface waves).

In the case of a dispersive wave (e.g. Rayleigh and Love surface waves), seismic energy at different frequencies travels at different speeds, creating interferences (constructive and/or destructive). A wave at a single frequency propagates as:

$$A.\cos(\omega t - kx) \quad (2.3)$$

where  $A$  is the amplitude of the wave ;  $\omega$  is the angular frequency (defining the temporal frequency of oscillation:  $\omega = \frac{2\pi}{T}$  where  $T$  is the period) ;  $t$  is the time ;  $k$  is the wave number (defining the spatial frequency of oscillation) and  $x$  is the distance. At each single frequency, the phase speed  $C$  is:

$$C = \frac{\omega}{k} \quad (2.4)$$

The speed at which the energy of the wave travels is the group speed  $U$  often measured as the speed of the envelope of the wave. For a dispersive wave, the equation defining the group velocity is:

$$U = \frac{d\omega}{dk} \quad (2.5)$$

The relation between the group and phase velocities emerges after differentiation of the phase velocity equation:

$$U = \frac{C}{1 - \frac{\omega}{C} \frac{dC}{d\omega}} \quad (2.6)$$

Using Equation 2.6, it is then possible to compute group velocities from phase velocity measurements (or *vice versa*). Figure 2.4 presents phase and group velocity for fundamental mode Rayleigh and Love wave. For more information on group velocities, see Chapter 5 and Appendix C.

Figure 2.6 presents the depth distribution of the sensitivity of Rayleigh and Love waves phase velocities to the S-wave velocity, P-wave velocity and density at five different periods. Rayleigh waves and Love waves present similar sensitivity kernels but with maximum sensitivity at different depths. Short period surface waves are sensitive to shallow structures while long period surface waves are sensitive to deeper structures. At longer periods, surface waves have higher phase velocities (Figure 2.4) and reach greater depths (Nolet, 2008).



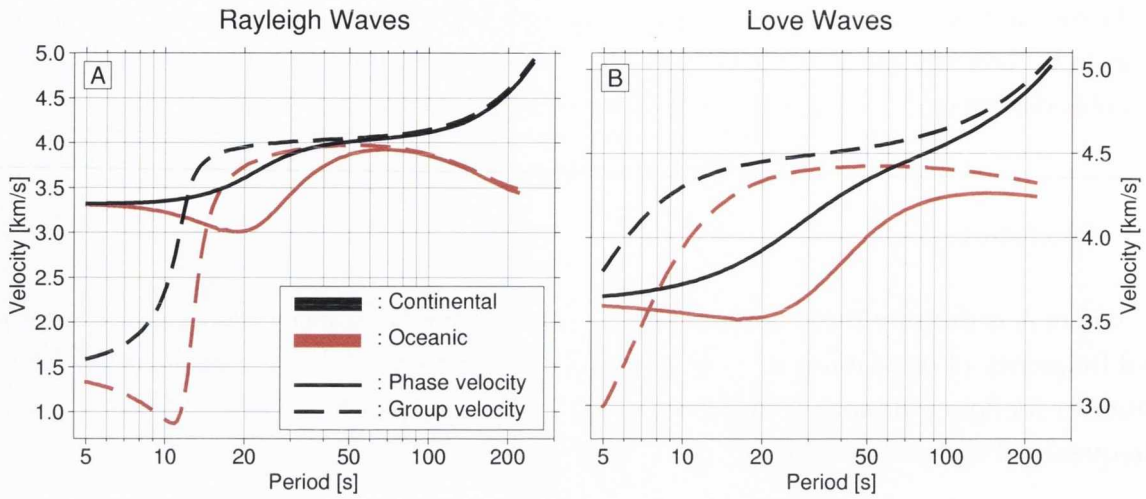


Figure 2.4.: Phase (solid line) and group (dashed line) velocity of fundamental mode Rayleigh (left) and Love (right) waves. The models are for modified reference model AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000).

As surface waves are stationary in the vertical direction, several solutions to the displacement equations may exist, depending on the period of the wave (Aki and Richards, 2002). These solutions are defined by the fundamental mode and harmonics of higher degrees. Figure 2.5 and Figure 2.10 present examples of phase velocity measurements for fundamental and higher mode Rayleigh and Love waves.

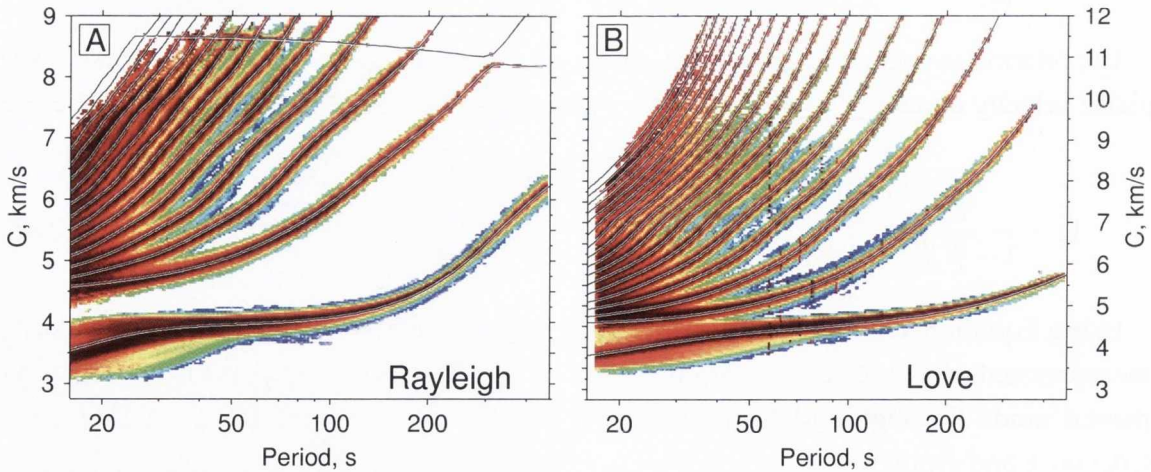


Figure 2.5.: Empirical phase velocity for fundamental 16 first higher-modes Rayleigh (A) and Love (B) waves. Rayleigh and Love phase velocities are computed from waveform fits of approximately 700 000 vertical and 200 000 horizontal respectively broadband seismograms (measurements from Schaeffer and Lebedev, 2012). The white and black lines represents phase velocities for reference model AK135 (Kennett *et al.*, 1995). Figure with courtesy of Andrew Schaeffer.



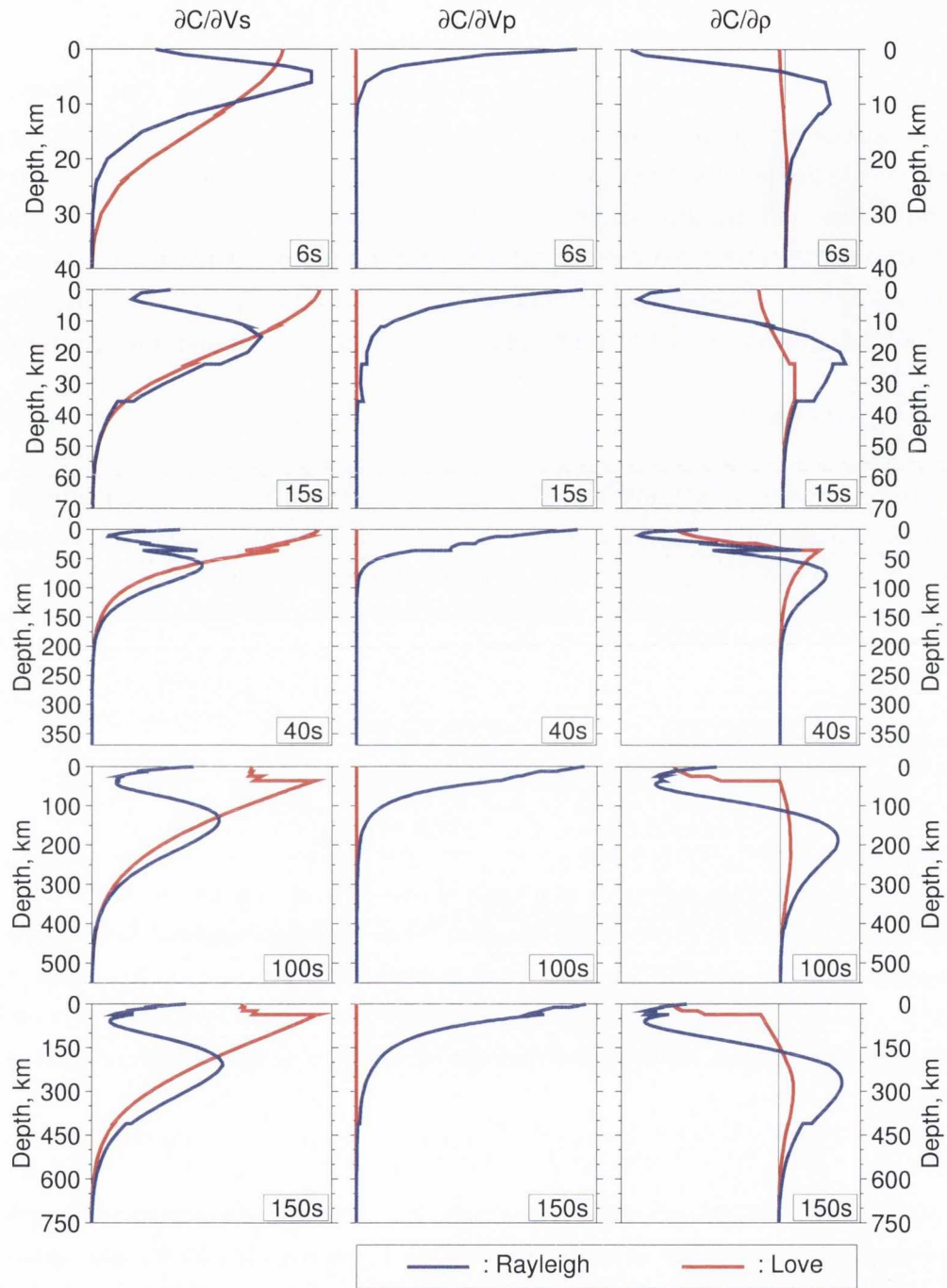


Figure 2.6.: Depth sensitivity of surface waves (at 6, 15, 40, 100 and 150 s). The sensitivity curves are the Fréchet derivatives of the phase velocities of the fundamental-mode Rayleigh and Love waves with respect to S-wave velocities ( $V_s$ ), P-wave velocities ( $V_p$ ) and density ( $\rho$ ). The derivatives were computed for a continental, 1-D Earth model with a 37 km thick crust. Each graph is scaled independently.

## 2.2 SURFACE WAVE DISPERSION CURVES MEASURED USING TWO STATIONS

Surface wave dispersion curves are measured between station-pairs. One advantage of using the two-station method instead of using the event-station method is that the phase (or group) velocities obtained represent a good approximation of the structures below and between the two stations and not between the earthquake and the station. This enables us to constrain the seismic parameters at a smaller scale. An other advantage of using this method is that it is possible to measure phase and group velocities in a broader band of period. Surface waves then sample the Earth in both smaller and greater depths (Figure 2.6).

A combination of two methods is used to measure inter-station phase-velocities: a cross-correlation method and a waveform inversion method. Both approaches are complementary in that they provide measurements in different period bands.

### 2.2.1 Cross-correlation method

Surface wave dispersion curves of Rayleigh (Love) waves can be computed by cross-correlation of the vertical (horizontal) component of the displacement field recorded at two stations that are aligned (or with a backazimuth difference of less than  $7^\circ$ ). A frequency-dependent Gaussian bandpass filter and a frequency-dependent windowing in the time domain is applied to the cross-correlation function to improve the signal-to-noise ratio (Meier *et al.*, 2004).

The surface-wave phase velocities are computed from the calculation of the phase of the spectrogram of the cross-correlation function. However, due to the ambiguity from the arctan function in the calculation of the phase, the phase velocities are selected manually from a set of possible dispersion curves so as to only keep the period bands of the curves relatively close to a reference dispersion curve. Figure 2.8 shows applications of this method by representing the seismograms recorded at each station of the station-pair, the cross-correlation function (and their corresponding spectrograms) and the set of dispersion curves (see Figure 2.7 and Table 2.1 for geographic location and date and time of the earthquakes).

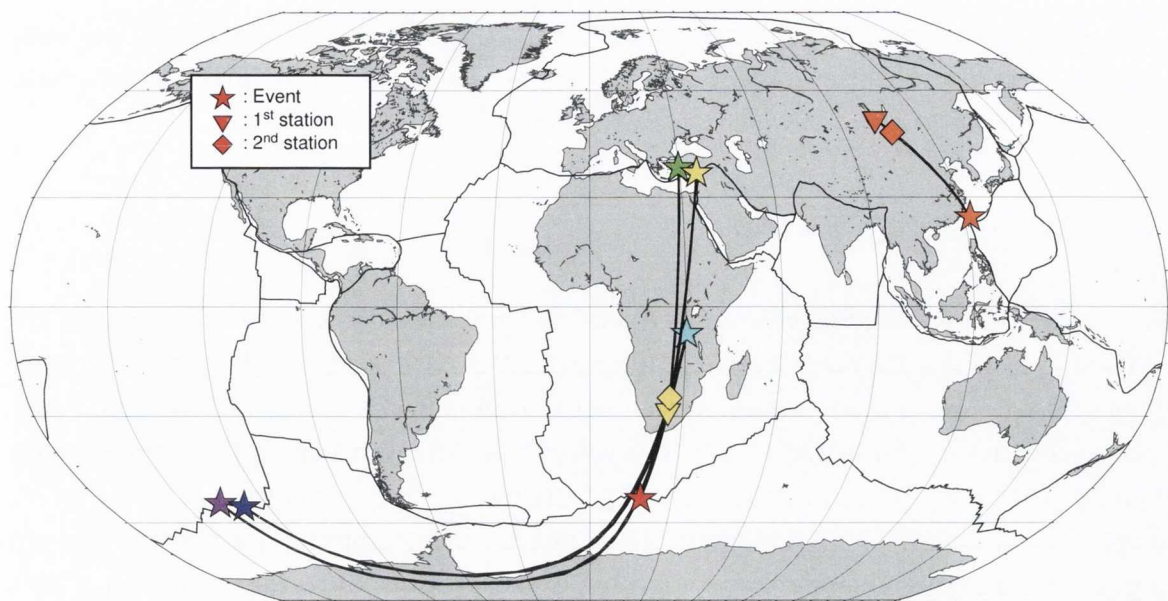


Figure 2.7.: Geographic location of earthquakes (stars) and stations (triangles and diamonds) for which phase velocity measurements are presented as examples. See Table 2.1 for more informations.

Table 2.1.: Properties of earthquakes (date and coordinates) and station-pairs (coordinates) for which phase-velocities measurements are presented as examples. Each pair is represented on the map in Figure 2.7 using the color in the last column. The station BOSA is located at the coordinates (-28.61° ; 25.25°). The station LBTB is located at the coordinated (-25.01° ; 25.59°). The station TLY is located at the coordinates (51.68° ; 103.64°). The station ULN is located at the coordinates (47.86° ; 107.05°)

| Date<br>(dd-mm-yy) | Earthquake<br>time<br>(hh:mm:ss) | Earthquake<br>coordinates<br>(Lat ; Lon) | Station 1<br>name | Station 2<br>name | Color<br>on Figure 2.7 |
|--------------------|----------------------------------|------------------------------------------|-------------------|-------------------|------------------------|
| 01-10-95           | 15:57:23                         | 38.06° ; 29.68°                          | BOSA              | LBTB              | ✕                      |
| 03-09-97           | 06:22:50                         | -55.26° ; -128.98°                       | BOSA              | LBTB              | ✕                      |
| 21-09-97           | 18:13:32                         | -7.290° ; 30.27°                         | BOSA              | LBTB              | ✕                      |
| 04-07-98           | 02:15:51                         | 36.63° ; 35.42°                          | BOSA              | LBTB              | ✕                      |
| 29-08-01           | 03:30:20                         | -52.94° ; 18.44°                         | BOSA              | LBTB              | ✕                      |
| 02-09-01           | 10:06:57                         | -54.31° ; -136.75°                       | BOSA              | LBTB              | ✕                      |
| 13-06-02           | 20:40:28                         | 24.59° ; 121.94°                         | TLY               | ULN               | ✕                      |

Thanks to its elaborate scheme of windowing and weighting in the time and frequency domains, this technique also produces robust and accurate measurements from waves diffracted or scattered at lateral heterogeneities. Indeed, despite the scattering of the wave-form in figure Figure 2.8 (1<sup>st</sup> page, right) it is still possible to measure phase velocities from 0.004 to 0.1Hz (*i.e* 10 to 250 s).



For each station-pair, hundreds of curves (*i.e.* from hundreds of earthquakes) are measured to obtain a robust average dispersion curve that is a good estimate of the phase velocities in the structures between the two stations.

### 2.2.2 Multi-mode inversion method

To measure phase velocities at longer periods, where cross-correlation measurements are difficult due to interferences between long-period surface waves and S waves, I additionally used a multi-mode inversion technique (Automated Multimode Inversion, AMI). Phase velocities are measured from phase velocity perturbations along the source-station path by waveform fitting in time-frequency windows of the seismogram (Lebedev *et al.*, 2005). Phase velocities between a pair of stations are then computed for stations that are aligned on the same great-circle path (within a backazimuth difference of less than  $10^\circ$ ). Waveform fitting is performed in time-frequency windows of the seismograms using the JWKB (Jeffreys-Wentzel-Kramers-Brillouin) approximation (Dahlen and Tromp, 1998). Its elaborate approach for window selection allows to automatically reject portions of the seismograms that can not be modeled accurately under this approximation.

The seismogram is filtered at several frequencies using Gaussian band-pass filters. The minimum and maximum frequencies at which the seismogram is filtered depend on the empirically determined validity of the JWKB approximation. Also, the minimum frequency is the frequency for which 3-4 wavelengths of the fundamental mode is smaller than the distance between the earthquake and the station. If this condition is not fulfilled, the far-field approximation is not valid. The maximum frequency allowed depends on the duration time of the earthquake so as to approximate the earthquake as a point-source.

At each frequency, the time windows within the seismogram are defined by boxcars, including the maxima of the envelope corresponding to the arrivals of the fundamental mode and S and multiple S waves.

In order to avoid fitting waveforms that are affected by scattered energy or strong source-mechanism errors, seismograms recorded at stations for which the source-station azimuth is near a node of the radiation patterns are excluded. Also, to ensure that the seismogram is not dominated by scattered energy, at least three windows must include fundamental modes.

Figure 2.9 presents examples of AMI fits with frequency filters and time-frequency windows for stations BOSA and LBTB in southern Africa (see Figure 2.7 and Table 2.1 for geographic location of both stations and date and time of the earthquakes). Corresponding Rayleigh wave phase velocity measurements are presented in Figure 2.10. In two cases, AMI were able to measure the fundamental and higher-modes (up to mode 7) Rayleigh wave phase velocities.



Thanks to the automation of the procedure, large amount (tens of thousands) of phase velocity measurements along event-station paths can be measured and used to resolve crustal and mantle structure and anisotropy, down to the mantle transition zone.

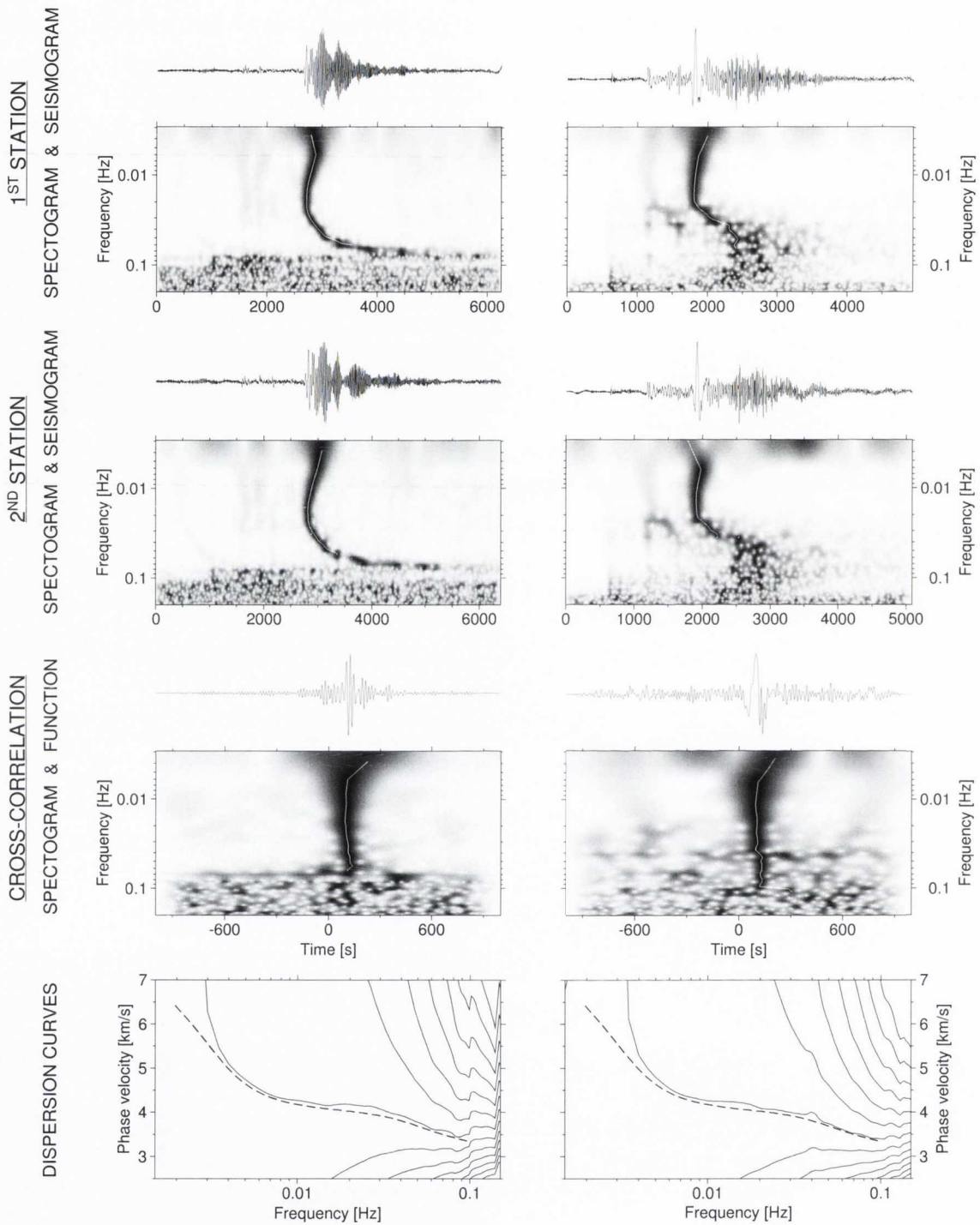


Figure 2.8.: Cross-correlation of teleseismic surface waves. Each column represents an example of phase-velocity measurements at the station-pair BOSA-LBTB using recordings from one earthquake. First/second line: vertical component of the seismogram (and its spectrogram) recorded at the first/second station. White lines are group traveltimes of fundamental mode Rayleigh wave. Third line: cross-correlation function (and its spectrogram) of the two seismograms. High energy in the spectrograms are presented with dark shadings. Bottom line: sets of dispersion curves computed from the phase of the cross-correlation function. The dashed line is a reference dispersion curve. The first column shows data recorded in 2001 on September, 2<sup>nd</sup> (green color on Figure 2.7 and Table 2.1). The second column shows data recorded in 1995 on October, 1<sup>st</sup> (blue color on Figure 2.7 and Table 2.1).

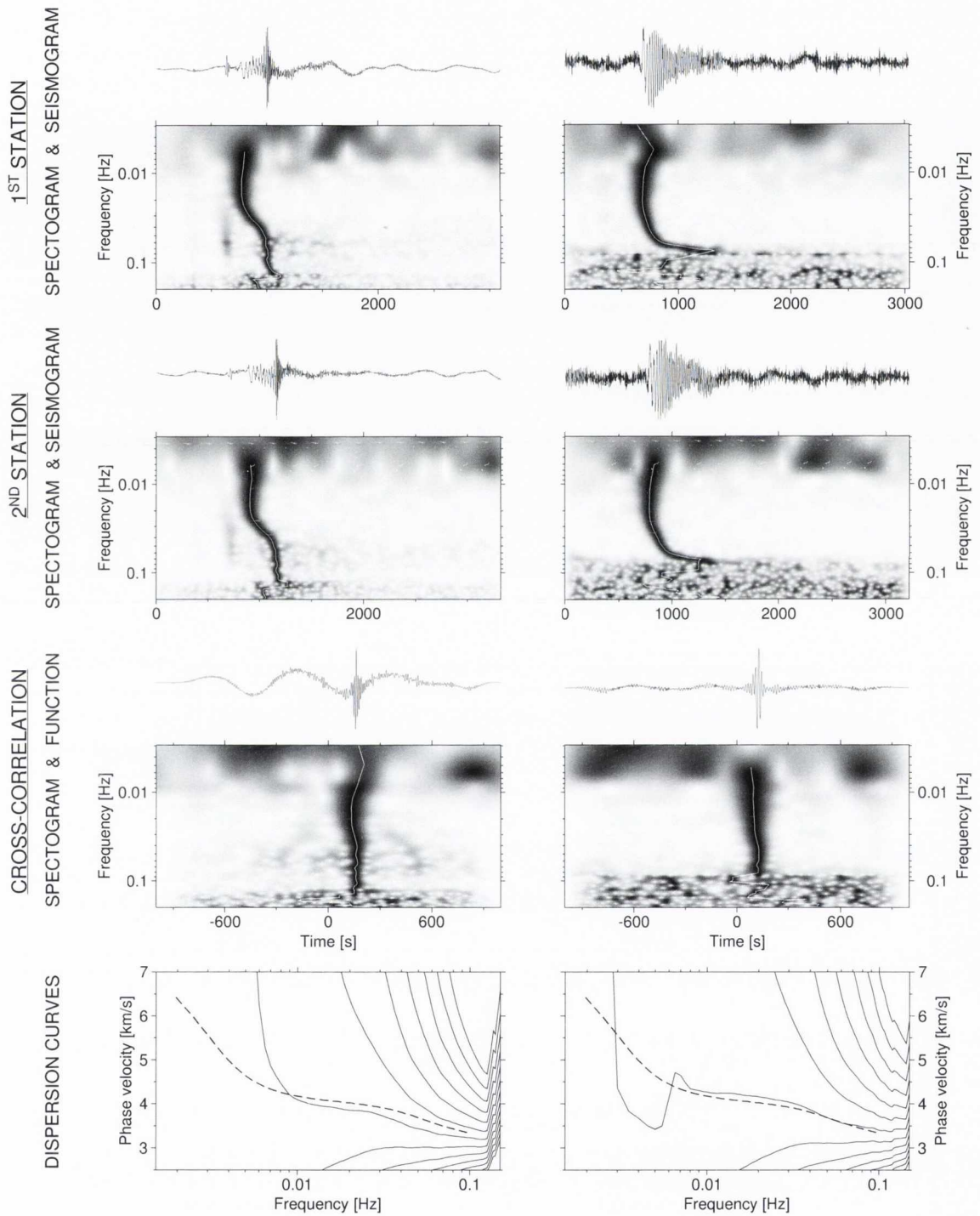


Figure 2.8.: Continued... The first column shows data recorded at the station-pairs TLY-ULN in 2002 on June, 13<sup>th</sup> (orange color on Figure 2.7 and Table 2.1). The second column shows data recorded at the station-pair BOSA-LBTB in 2001 on August, 29<sup>th</sup> (purple color on Figure 2.7 and Table 2.1).

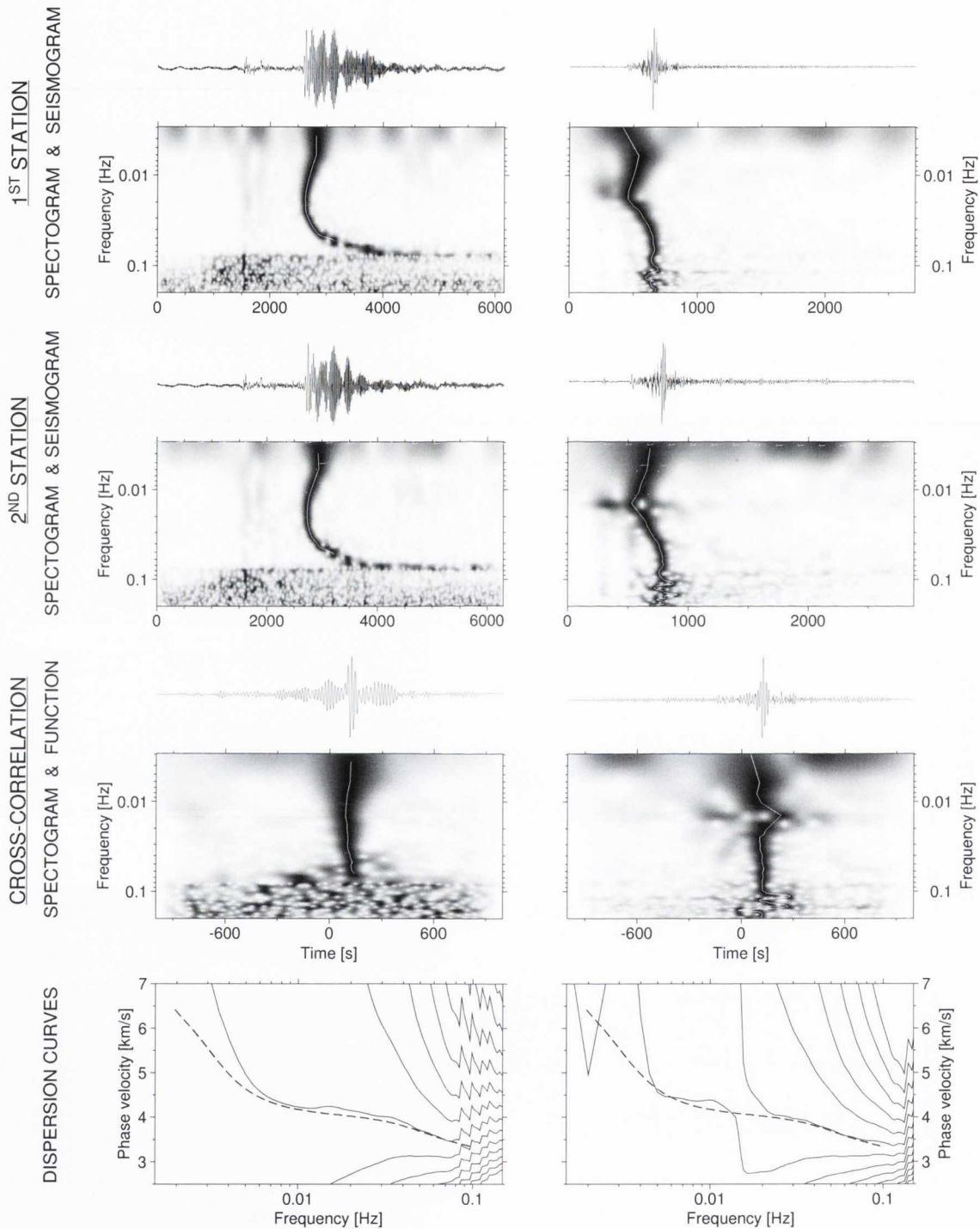


Figure 2.8.: Continued... The first column shows data recorded at the station-pairs BOSA-LBTB in 1997 on September, 3<sup>rd</sup> (cyan color on Figure 2.7 and Table 2.1). The second column shows data recorded at the station-pair BOSA-LBTB in 1997 on September, 21<sup>st</sup> (red color on Figure 2.7 and Table 2.1).



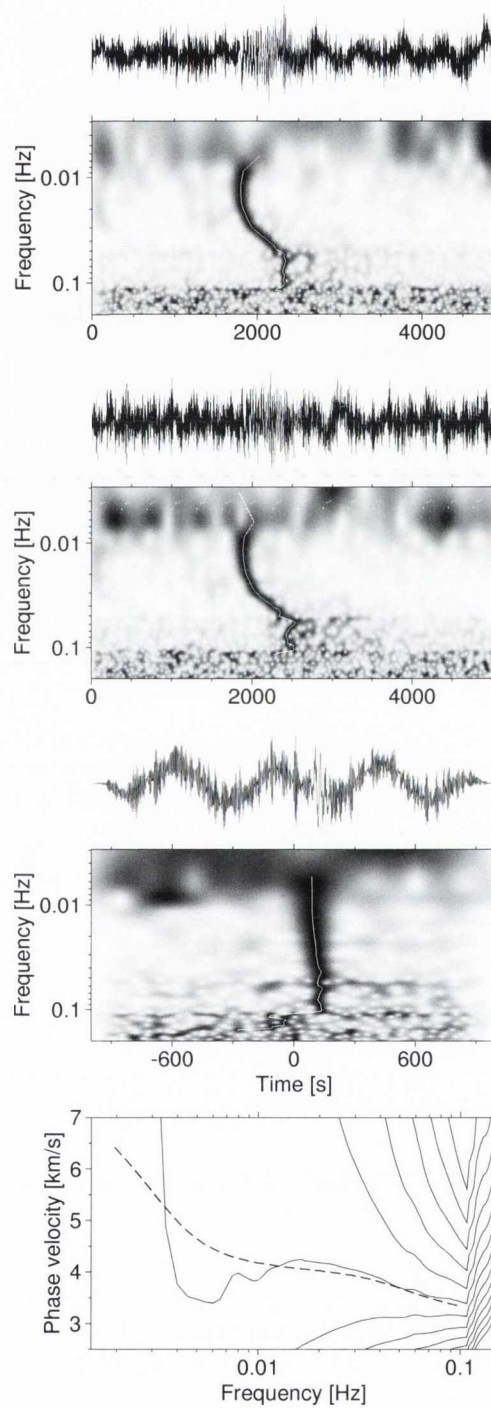


Figure 2.8.: Continued... Data are recorded at the station-pairs BOSA-LBTB in 1998 on July, 4<sup>th</sup> (yellow color on [Figure 2.7](#) and [Table 2.1](#)).

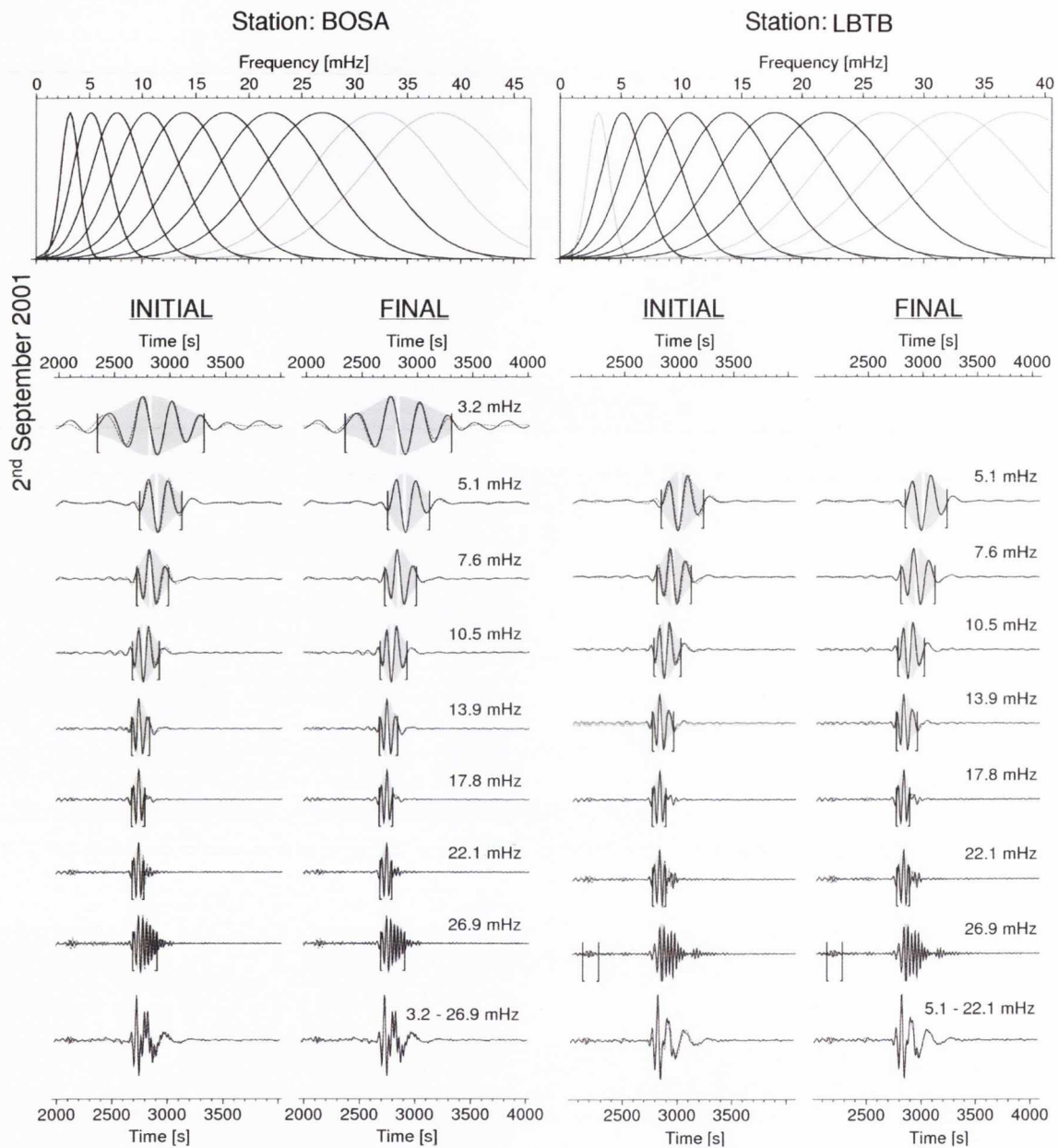


Figure 2.9.: Waveform fitting of data from one earthquake recorded at two stations. Top panel: gaussian band-pass filters. Black lines represent time-frequency windows with successful fits. Bottom panel: filtered waveforms (solid lines) are matched with synthetics (dashed lines) in time-frequency windows (represented by shaded area and demarcated with brackets). The left columns show the initial fit with a reference 1-D Earth model. On the right columns, the Earth model is altered to minimize the misfit between the data and the synthetics. Data have been recorded at stations BOSA (left) and LBTB (right) in 2001 on September 2<sup>nd</sup> (green color on Figure 2.7 and Table 2.1).

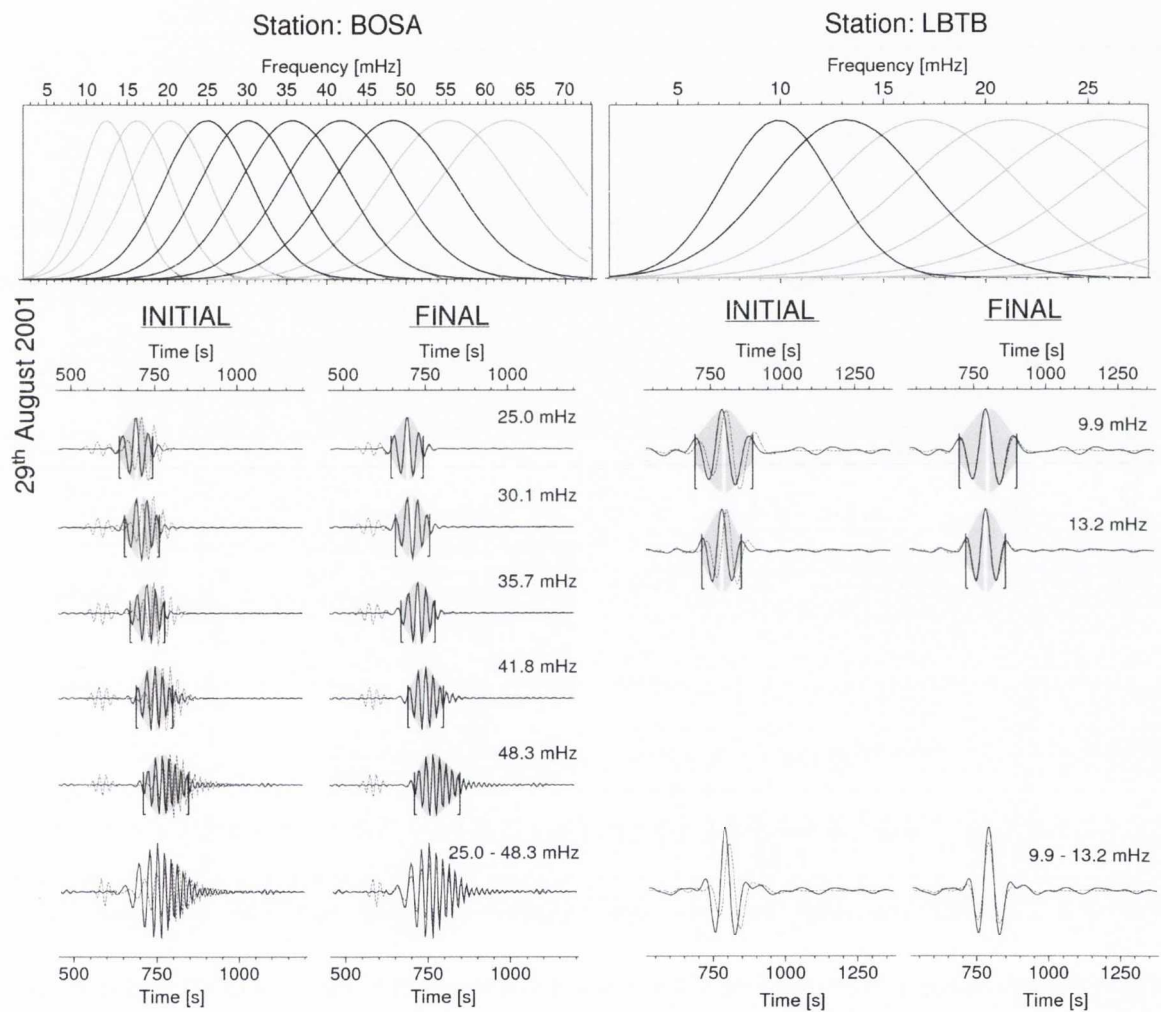


Figure 2.9.: Continued... Data have been recorded at stations BOSA (left) and LBTB (right) in 2001 on August 29<sup>th</sup> (purple color on Figure 2.7 and Table 2.1).

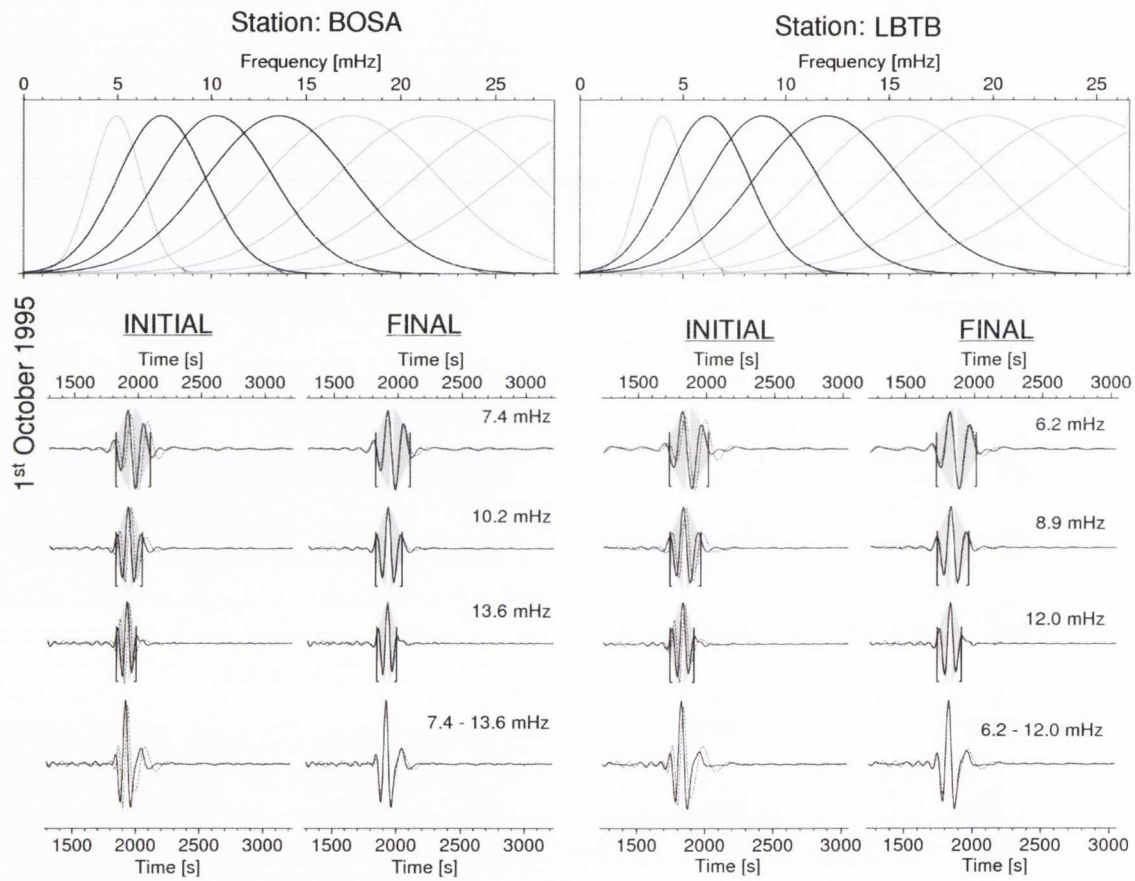


Figure 2.9.: Continued... Data have been recorded at stations BOSA (left) and LBTB (right) in 1995 on October 1<sup>st</sup> (blue color on Figure 2.7 and Table 2.1).



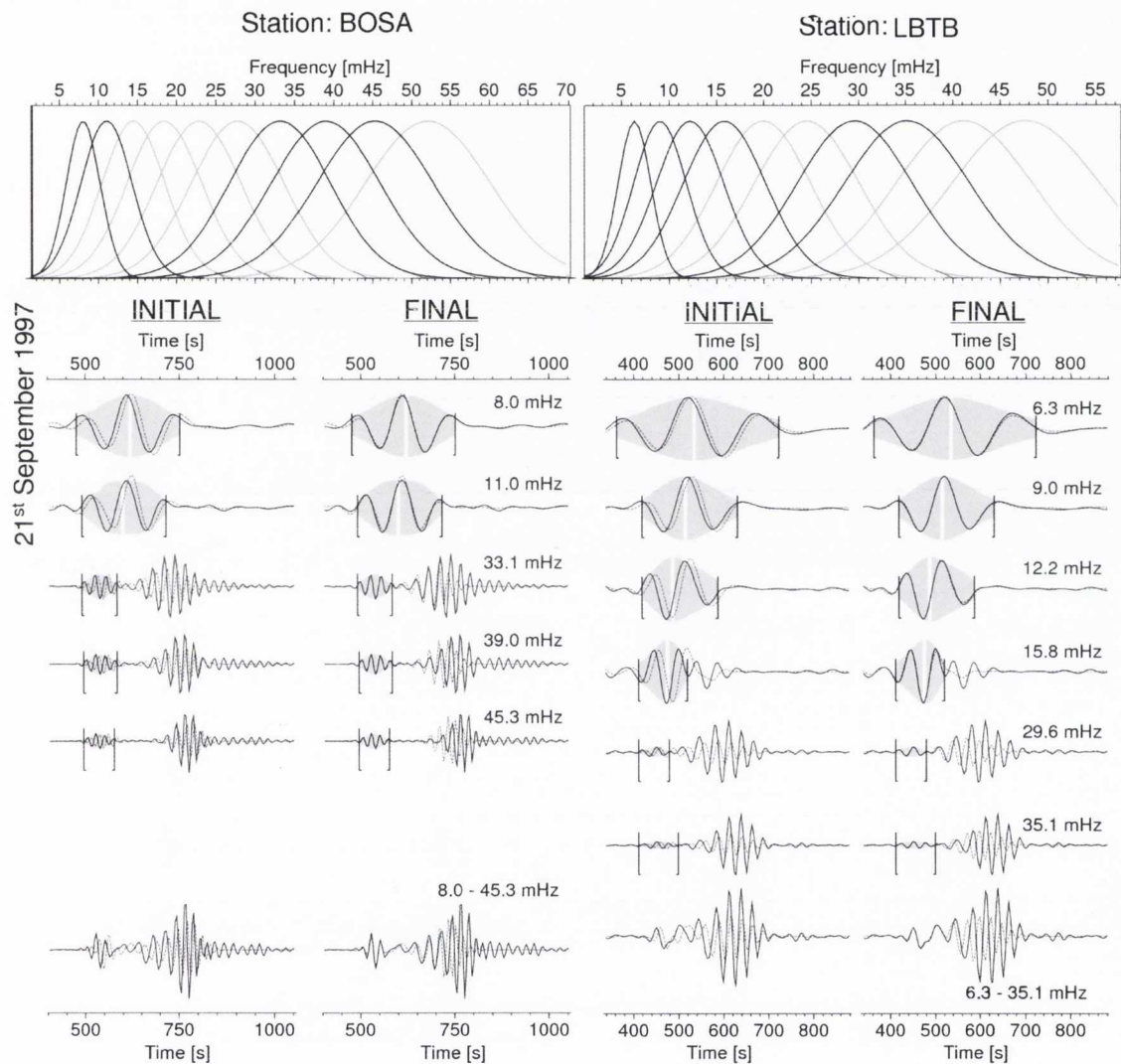


Figure 2.9.: Continued... Data have been recorded at stations BOSA (left) and LBTB (right) in 1997 on September 21<sup>st</sup> (red color on [Figure 2.7](#) and [Table 2.1](#)).

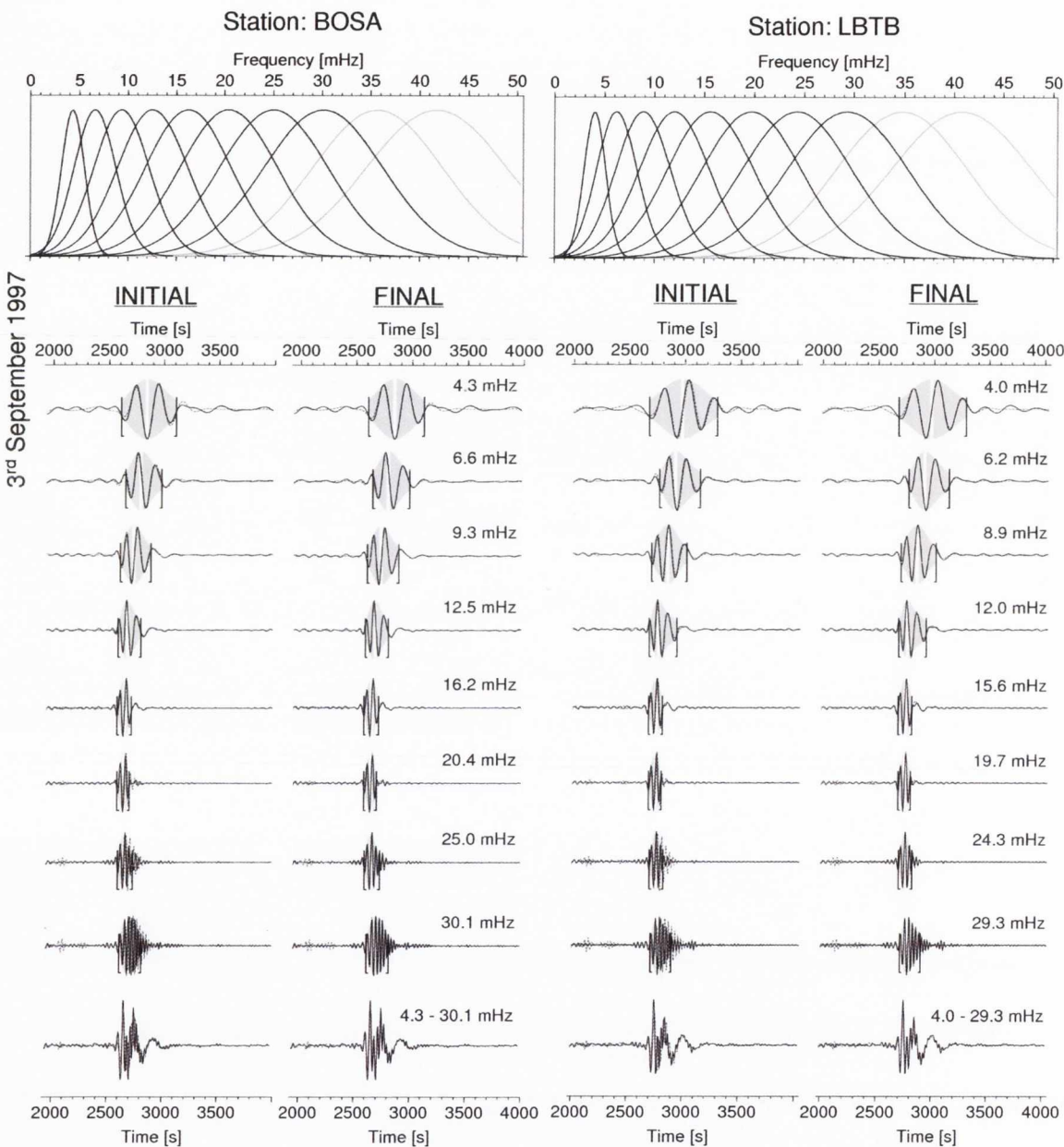


Figure 2.9.: Continued... Data have been recorded at stations BOSA (left) and LBTB (right) in 1997 on September 3<sup>rd</sup> (cyan color on [Figure 2.7](#) and [Table 2.1](#)).

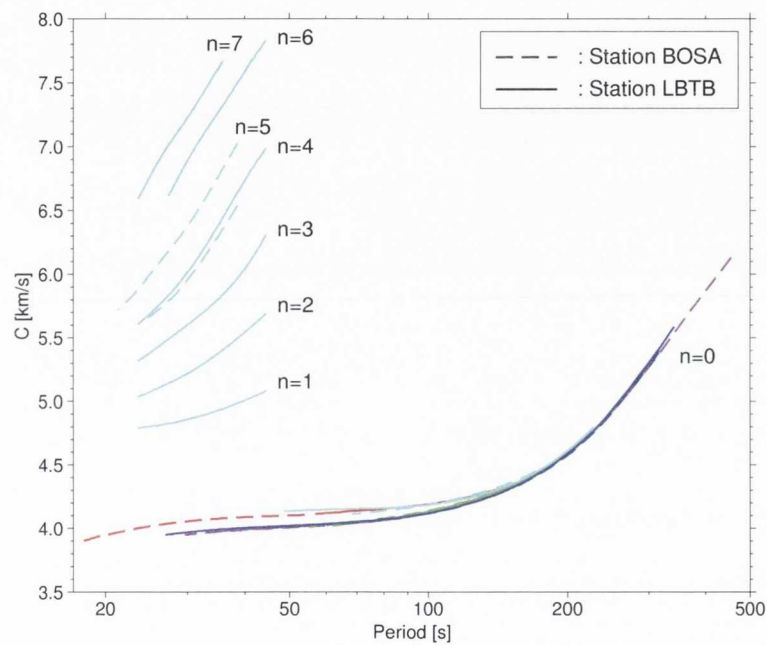


Figure 2.10.: Phase velocity measurements at stations BOSA and LBTB inferred from multi-mode technique. Each color corresponds to a specific earthquake. See [Figure 2.7](#), [Figure 2.9](#) and [Table 2.1](#) for more informations.

Part II

CRUSTAL AND UPPER-MANTLE STRUCTURES IN  
SOUTHERN AFRICA



---

## PHASE VELOCITY AZIMUTHAL ANISOTROPY<sup>1</sup>

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### 3.1 SUMMARY

Seismic anisotropy within the lithosphere of cratons preserves an important record of their ancient assembly. In southern Africa, anisotropy across the Archean Kaapvaal Craton and Limpopo Belt has been detected previously by observations of S K S wave splitting. Because S K S splitting measurements lack vertical resolution, however, the depth distribution of anisotropy has remained uncertain. End-member interpretations invoked the dominance of either anisotropy in the lithosphere (due to the fabric formed by deformation in Archean or Paleoproterozoic orogenies) or that in the asthenosphere (due to the fabric formed by the recent plate motion), each with significant geodynamic implications.

In order to determine the distribution of anisotropy with depth, we measured phase velocities of seismic surface waves between stations of the Southern African Seismic Experiment. We applied two complementary measurement approaches, very-broadband cross-correlation and multimode waveform inversion. Robust, Rayleigh- and Love-wave dispersion curves were derived for four different sub-regions of the Archean southern Africa in a period range from 5 s to 250-400 s (Rayleigh) and 5 s to 100-250 s (Love), depending on the region. Rayleigh-wave anisotropy was determined in each region at periods from 5 s to 150-200 s, sampling from the upper-crust down to the asthenosphere. The Jackknife method was used to estimate uncertainties, and the F-test to verify the statistical significance of anisotropy.

We detected strong anisotropy with a N-S fast-propagation azimuth in the upper-crust of the Limpopo Belt. We attribute it to aligned cracks, formed by the regional, E-W extensional stress associated with the southward propagation of the East African Rift. Our results show that it is possible to estimate regional stress from short-period, surface-wave anisotropy, measured in this study using broadband-array recordings of teleseismic surface waves.

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Rayleigh-wave anisotropy at 70-120 s periods shows that the fabric within the deep mantle lithosphere of the Limpopo Belt and northern Kaapvaal Craton is aligned parallel to the Archean-Paleoproterozoic sutures at block boundaries. This confirms that the fabric within the lithosphere created by pervasive ancient deformation is preserved to this day. Suture-parallel fabric is absent, however, in the deep lithosphere of the western Kaapvaal Craton, suggesting that it was not reworked in the collision with the craton's core, either due to its mechanical strength or because the deformation mechanism was different from those that operated in the North. Anisotropy at periods greater than 120-130 s shows fast directions parallel to the plate motion and indicates shear-wave anisotropy in the asthenosphere.

The depth distribution of anisotropy revealed by surface-wave measurements comprises elements of both end-member models proposed previously: anisotropy in the asthenosphere shows fast-propagation directions parallel to the plate motion; anisotropy in the Limpopo and northern Kaapvaal lithosphere shows fast directions parallel to the Archean-Paleoproterozoic sutures. The distribution of SKS splitting orientations across southern Africa reflects anisotropic fabric both within the lithosphere (dominating the splitting beneath the Limpopo Belt and northern Kaapvaal Craton) and within the asthenosphere (dominating beneath the western Kaapvaal Craton).

### 3.2 INTRODUCTION

The study of cratons (the ancient cores of continents) takes us as far back in time as the Archean Eon, over 2.5 b.y. before present. The rocks within the lithosphere of cratons preserve a record of their Archean formation and evolution. The rocks' anisotropic fabric, in particular, is an indicator of the patterns of deformation during the enigmatic assembly of cratons. Such fabric gives rise to azimuthal seismic-velocity anisotropy (directional dependence of seismic wave speeds) and causes detectable perturbations in the propagation of seismic waves that traverse it within the crust and underlying mantle. The character and the lateral and vertical variability of the fabric can thus be determined from seismic observations, yielding important information on how cratons evolved.

The Archean Kaapvaal Craton in southern Africa (Figure 3.1) is large ( $1.2 \times 10^6 \text{ km}^2$ ) and offers abundant (and well studied) kimberlite nodules from the deep lithosphere. It has also been sampled by a dense array of broadband seismic stations, the Southern African Seismic Experiment (SASE), making it an exceptional natural laboratory for the study of ancient continents (De Wit *et al.*, 1992; Carlson *et al.*, 2000; James *et al.*, 2001; Silver *et al.*, 2001; Schutt and Leshner, 2010). The oldest crustal ages in the western and central parts of the craton are 2.9-3.2 Ga and 3.0-3.6 Ga, respectively, and the bulk of its mantle lithosphere has probably formed prior to 3 Ga (Griffin *et al.*, 2004). The older, eastern part of the Kaapvaal Craton stabilised before 2.8-3.0 Ga; its western and northern parts,



as well as the Limpopo Belt adjacent to it on the North, are thought to have accreted to it in convergent-margin processes between about 2.9 and 2.6 Ga in the late Archean or, for the Limpopo Belt, around 2.0 Ga in the Paleoproterozoic (De Wit *et al.*, 1992; Kamber *et al.*, 1995; Holzer *et al.*, 1998; Irvine *et al.*, 2001; Moser *et al.*, 2001; Schmitz *et al.*, 2004; Silver *et al.*, 2004).

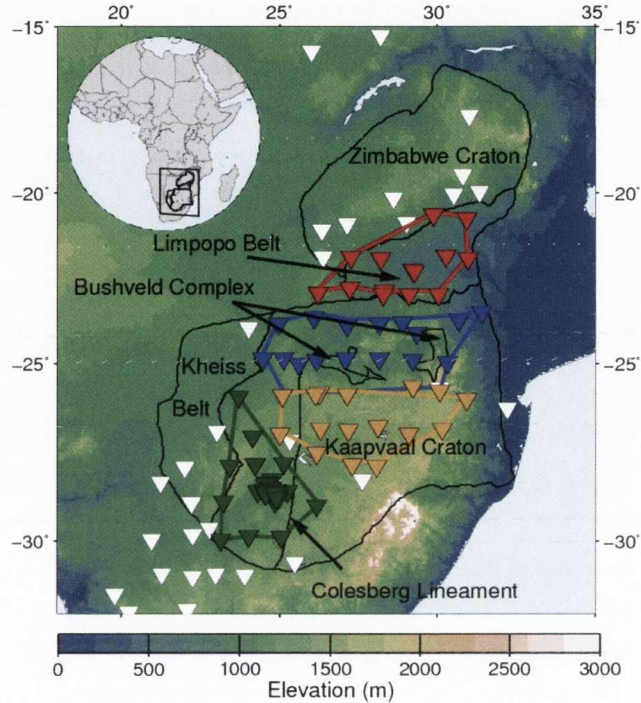


Figure 3.1.: Geographic map of southern Africa, with the main tectonic units labelled. Most of the seismic stations (triangles) were deployed by SASE. The four sub-regions that we identify are indicated with different colours of the station symbols.

As most Archean cratons, the Kaapvaal Craton and the Limpopo Belt are underlain by a thick, cold, stable mantle lithosphere, compositionally buoyant due to its depletion in basaltic components (Jordan, 1975, 1988; Carlson *et al.*, 2000; Begg *et al.*, 2009; Muller *et al.*, 2009; Fulla *et al.*, 2011). Anomalously high S- and P-wave velocities beneath the Kaapvaal Craton and the Limpopo Belt reflect anomalously low temperatures within the mantle beneath them and show that their lithosphere currently extends down to at least 180-200 km depths (e.g. James *et al.*, 2001; Fouch *et al.*, 2004a; Priestley *et al.*, 2006; Chevrot and Zhao, 2007; Wang *et al.*, 2008; Begg *et al.*, 2009; Priestley and Tilmann, 2009; Fishwick, 2010; Adams and Nyblade, 2011; Li, 2011). Mantle xenolith data sample the temporal evolution of the lithosphere since its formation. They show that most of the Kaapvaal Craton and Limpopo Belt mantle root (down to 180-200 km depths) has formed and stabilised in the Archean Eon, and that it has been attached to the overlying crust since that time (Pearson *et al.*, 1995; Carlson *et al.*, 1999, 2000; Griffin *et al.*, 2004).

According to seismic receiver-function and surface-wave studies, the crust beneath the northern Kaapvaal Craton and Limpopo Belt is thicker (40–45 km) than beneath the central and southwestern Kaapvaal Craton (35–40 km) (Nguuri *et al.*, 2001; Stankiewicz *et al.*, 2002; Nair *et al.*, 2006; Yang *et al.*, 2008). The world's largest layered mafic intrusion of the Bushveld Igneous Complex in the northern Kaapvaal Craton developed at around 2.06 Ga (Figure 3.1) and was associated with a modification of both the crust and mantle lithosphere (Good and De Wit, 1997; Carlson *et al.*, 1999; Eglington and Armstrong, 2004). The position of the Bushveld Complex may have been influenced by the Thabazimbi-Murchison Lineament (TML) (Figure 1.1) (Good and De Wit, 1997) which extends across the North of the Kaapvaal craton in the NNE-SSW direction, subdividing it with faults.

A major finding of the SASE deployment of broadband seismic stations was the peculiar distribution of SKS splitting across the Kaapvaal Craton and the Limpopo Belt. The splitting of SKS waves is an indicator of azimuthal anisotropy along their nearly vertical path from the core-mantle boundary to the seismic station. Normally, the splitting measurements are dominated by anisotropy within the lithosphere-asthenosphere system. The anisotropy reflects the crystallographic preferred orientation of olivine (the dominant mineral in the mantle) that develops in response to shear strain (*e.g.* Silver, 1996; Savage, 1999; Becker *et al.*, 2006; Long and Becker, 2010). In southern Africa, the splitting times (delays between the arrivals of the SKS waves polarized in the fast and slow propagation directions) are relatively small, around 0.6 s on average (Silver *et al.*, 2001, 2004). This is about half of the global average, or of the global average over cratons only (Becker *et al.*, 2012). The fast-propagation azimuths, however, displayed a coherent spatial pattern, with NE-SW fast azimuths in the western Kaapvaal Craton rotating to ENE-WSW and E-W azimuths in the northern Kaapvaal and the Limpopo Belt (Section 3.7). Little or no SKS splitting was detected in the central Kaapvaal Craton.

Silver *et al.* (2001, 2004) interpreted these fast-propagation directions as evidence that the SKS splitting occurs in the lithosphere. The important implication was that the splitting indicated fabric within the mantle lithosphere that formed at 2.9 to 2.6 Ga during collisional accretion of the western and northern parts of Kaapvaal Craton (Kimberly and Pietersburg blocks, respectively) and, later, the Limpopo Belt to the craton's core (Witwatersrand block in the central part of the craton today). This interpretation was in contrast to an earlier inference by Vinnik *et al.* (1995), who detected SKS splitting with NNE-SSW fast-propagation directions at stations within the southern Kaapvaal Craton and concluded that the splitting reflected the shear in the mantle associated with the current motion of the African Plate.

Silver *et al.* (2001, 2004) postulated that the SKS splitting originates in the lithosphere because a significant proportion of their measurements (particularly, the E-W ones in northern Kaapvaal and Limpopo) did not match the plate-motion direction. Their measurements in the western Kaapvaal, however, showed NE-SW and NNE-SSW fast direc-



tions, similarly to the data of Vinnik *et al.* (1995), implying an ambiguity in what the SKS splitting may indicate (Archean fabric in the lithosphere or recent shear beneath it?). Fundamentally, the ambiguity is due to the lack of vertical resolution inherent in SKS measurements. In order to determine the vertical distribution of anisotropic fabric beneath southern Africa, measurements of different types are required.

Surface waves can provide the necessary vertical resolution. Global and continent-scale tomographic models derived from surface-wave observations have been successful in mapping the layering of azimuthal anisotropy within continental lithosphere and asthenosphere at large scales (e.g. Simons *et al.*, 2002; Debayle *et al.*, 2005; Yuan and Romanowicz, 2010). Analysis of surface-wave data from dense broadband arrays has been producing increasingly high resolution of distributions of anisotropy, both lateral and vertical (Pedersen, 2006; Yang and Forsyth, 2006; Zhang *et al.*, 2007; Deschamps *et al.*, 2008a; Beghein *et al.*, 2010; Fry *et al.*, 2010; Endrun *et al.*, 2011; Lin *et al.*, 2011).

In southern Africa, a number of surface-wave studies targeted the isotropic shear-speed structure of the lithosphere, all detecting a thick, high-velocity root beneath the Archean cratons (e.g. Priestley, 1999; Freybourger *et al.*, 2001; Saltzer, 2002; Priestley *et al.*, 2006; Li and Burke, 2006; Yang *et al.*, 2008; Fishwick, 2010; Adams and Nyblade, 2011; Fishwick and Bastow, 2011; Li, 2011). Radial anisotropy, determined by joint inversions of Rayleigh and Love wave measurements, was found to be similar in amplitude to the global average (Dziewonski and Anderson, 1981), as for cratons elsewhere (Lebedev *et al.*, 2009), with horizontally polarized shear waves traveling a few percent faster than vertically polarized ones (Freybourger *et al.*, 2001; Saltzer, 2002). Measurements of azimuthal anisotropy, however, have proven difficult. Freybourger *et al.* (2001) searched for but did not detect azimuthal surface-wave anisotropy. Li (2011) inverted Rayleigh-wave measurements for isotropic phase-velocity maps and the whole-region-average phase-velocity anisotropy and found predominantly NE-SW fast-propagation directions in a 40–110 s period range, with amplitudes mostly under 1%. Interestingly, Pn measurements in the central Kaapvaal Craton indicated similar, NNE-SSW fast-propagation directions within the uppermost mantle, although the small dataset available was not sufficient to isolate completely the signal of anisotropy from that of isotropic heterogeneity (Kwadiba *et al.*, 2003). Overall, three-dimensional variations of azimuthal anisotropy beneath southern Africa are still unknown.

In this study, we apply powerful recent implementations of inter-station measurement methods to the complete SASE dataset and measure phase velocities of the Rayleigh and Love fundamental modes in broad period ranges. Analysis of the new measurements reveals the lateral and vertical distribution of azimuthal anisotropy across the Kaapvaal Craton and the Limpopo Belt, from the upper-crust down to the asthenosphere. We reconcile our results qualitatively with published shear-wave splitting measurements, identify

the likely depth ranges of the origin of the splitting, and discuss the implications of the results for southern Africa's lithospheric and asthenospheric dynamics.

### 3.3 PHASE VELOCITIES

#### 3.3.1 *Measurements*

We collected seismograms recorded by two permanent (Geotech KS-54000 borehole seismometers) and 112 temporary (STS2) broad-band stations in southern Africa (Figure 3.1). The temporary stations belonged to the Southern African Seismic Experiment (SASE) which included two deployments: the Anatomy of an Archean Craton (55 broad-band seismic stations at 82 sites) and the Kimberley Telemetered Array (30 temporary stations) (Carlson *et al.*, 1996). We apply a small (-0.688 s) "Quanterra bug" timing correction (Bolton *et al.*, 1998) to the long-period-channel data from the two permanent stations, BOSA and LBTB. All seismograms underwent instrumental response correction to displacement. The horizontal components were rotated to the radial and transverse components.

We use two methods to obtain Rayleigh- and Love-wave dispersion curves for pairs of stations in a broad range of periods. With the first method, we compute the cross-correlation of the displacement field from a source recorded at two stations (Meier *et al.*, 2004) that are aligned with the source (within a back-azimuth window of a  $10^\circ$  half-width). Thanks to its elaborate scheme of windowing and weighting in the time and frequency domains, this method enables measurements of inter-station dispersion curves from short (5-20 s) to long (up to hundreds of seconds) periods, from the phase of the phase of the cross-correlation function. Short period teleseismic surface waves are always diffracted and often show distinct multiple arrivals of the fundamental mode (multi-pathing). These waves would be difficult to model accurately with any techniques and models available. It turns out, however, that if the complex wavefield does not change too much between nearby stations, then the cross-correlation of the diffracted surface waves recorded at these stations can produce robust, accurate phase-velocity measurements (Meier *et al.*, 2004; Endrun *et al.*, 2008; Lebedev *et al.*, 2009).

The second method uses the Automated Multimode Inversion of surface- and S- waveforms (Lebedev *et al.*, 2005) that, for successful waveform fits, can measure average phase velocities between sources and stations. Dispersion curves that average between pairs of stations are then computed from these "source-station" dispersion curves, with the stations and sources, again, approximately on the same great-circle path (Lebedev *et al.*, 2006). This method is complementary to the first one in that it provides more phase-velocity measurements at long periods (over 50 s), at which the fundamental mode often interferes with S and multiple-S arrivals, especially for Love waves, making cross-correlation



measurements difficult. Dispersion curves from both methods and all earthquakes are then combined, and an average curve can be computed for each station pair.

We use transverse and vertical components to measure phase-velocity curves for Love and Rayleigh waves, respectively. Measurements are carefully examined interactively, in order to remove any rough (not smooth) portions of dispersion curves and outliers. Surface-wave diffraction and mode interference may cause errors in phase-velocity measurements performed on seismograms from any single event (e.g. Friederich *et al.*, 2000; Pedersen, 2006; Bodin and Maupin, 2008). Because these effects have a strong frequency dependence, selection of only smooth portions of phase-velocity curves reduces the impact of the errors. The averaging of many measurements, performed on signal from events in different regions, particularly at opposite azimuths from the station-station pair, is another important element of the method and produces robust, accurate phase-velocity measurements (Meier *et al.*, 2004; Lebedev *et al.*, 2006).

### 3.3.2 The four sub-regions

A robust phase-velocity curve for a pair of stations can be computed as an average over tens to hundreds of curves measured using seismograms from single events (Meier *et al.*, 2004; Lebedev *et al.*, 2006). If average dispersion curves are computed for various station combinations within an array, a tomographic inverse problem can be set up and solved for the distributions of the isotropic and anisotropic heterogeneity (e.g. Zhang *et al.*, 2007; Deschamps *et al.*, 2008a; Endrun *et al.*, 2011). Because of the relatively short deployment time of the SASE stations and the remoteness of southern Africa from regions of abundant seismicity, there are not enough “one-event” measurements to compute robust averages for most of the pairs of SASE stations.

We thus devise an alternative approach: we divide southern Africa into sub-regions, each with a relatively homogeneous structure within it, and invert our measurements for sub-region-average, phase-velocity structure and anisotropy. The four regions we identify are the Limpopo Belt and the northern, central and southwestern Kaapvaal Craton (Figure 3.1). The crustal thickness is relatively uniform within each of the sub-regions, according to published receiver-function and surface-wave studies (Nguuri *et al.*, 2001; Stankiewicz *et al.*, 2002; Nair *et al.*, 2006; Yang *et al.*, 2008; Kgaswane *et al.*, 2009). Isotropic phase-velocity heterogeneity at periods sampling the crust and the mantle is also relatively weak within each sub-region (Yang *et al.*, 2008; Li, 2011). Each of the four regions contains numerous two-station pairs (Figure 3.2, top) and is sampled by a very large number of phase-velocity measurements (Figure 3.2, middle and bottom; Table Table 3.1).

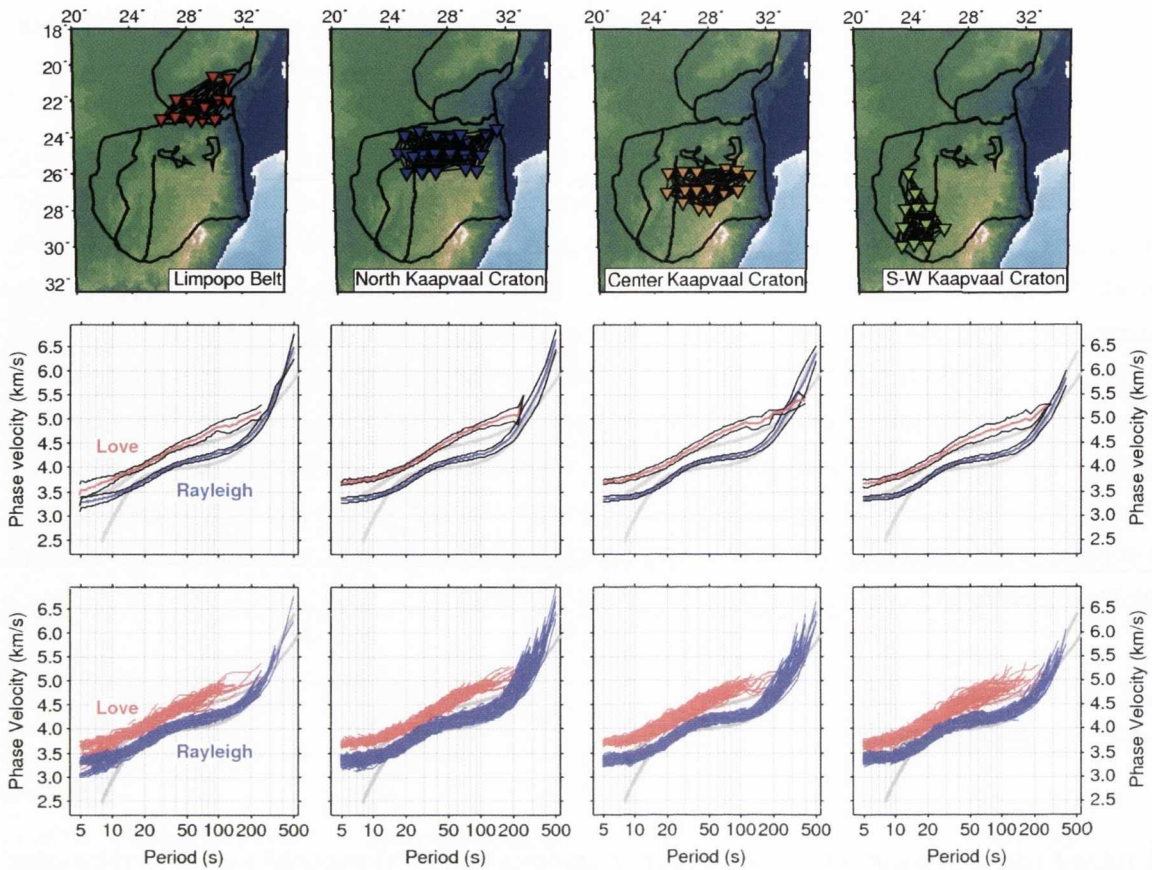


Figure 3.2.: Rayleigh and Love phase velocity measurements. Top: stations and interstation paths within each sub-region. Middle: region-average, phase-velocity curves for Love and Rayleigh waves. Bottom: all dispersion curves, each measured using a pair of stations and a single event. Grey dispersion curves were computed for the reference model AK135 (Kennett *et al.*, 1995).

### 3.3.3 Isotropic-average, Rayleigh and Love phase-velocity curves

Using the numerous phase-velocity measurements within each of the four regions (Table 3.1; Figure 3.2), we obtain accurate isotropic-average dispersion curves in very broad ranges of periods. The period ranges extend from 5 s to 250-400 s for Rayleigh and from 5 s to 100-250 s for Love waves, depending on the region (Figure 3.2, Figure 3.3). Phase-velocity measurements in such period ranges can resolve shear-velocity structure from the upper-crust down to the lower lithosphere and asthenosphere (Figure 3.4).



Table 3.1.: Number of stations, station pairs and Rayleigh and Love dispersion curves for the four sub-regions and the entire region.

|                         | # stations | # station<br>pairs | # Rayleigh<br>dispersion curves | # Love<br>dispersion curves |
|-------------------------|------------|--------------------|---------------------------------|-----------------------------|
| Limpopo Belt            | 16         | 112                | 1065                            | 455                         |
| N Kaapvaal Craton       | 25         | 191                | 2241                            | 500                         |
| Central Kaapvaal Craton | 18         | 113                | 1119                            | 431                         |
| S-W Kaapvaal Craton     | 34         | 190                | 1222                            | 530                         |
| Southern Africa         | 88         | 618                | 6188                            | 2083                        |

In Figure 3.3 we compare the phase-velocity curves from the four regions with each other, with global-reference curves, and with published previous measurements. Both Rayleigh and Love curves from the four regions are well above global average values and are, overall, very similar across the four regions (Figure 3.3A). Zooming in on the differences between the regions, we plot Rayleigh-wave phase-velocity anomalies (Figure 3.3B), computed as differences of observed curves and the curve computed for the reference model AK135 (Kennett *et al.*, 1995), which is close to a global continental average.

Phase velocity curves in Figure 3.3(A and B) are computed by simple averaging over all measurements within the regions. Our measurements, however, are distributed unevenly as a function of azimuth of the inter-station path (specifically, paths with ~N-E back-azimuths align with more events, mainly in S-E Asia, than those at other back-azimuths). In the presence of azimuthal anisotropy, this may result in a bias in the average phase-velocity curves. The bias can be removed by computing both the isotropic values and anisotropy of phase velocities explicitly, as we describe in Section 3.4. In Figure 3.3(C and E) we plot our isotropic-average phase velocity curves with the effect of anisotropy removed; Figure 3.3(D and F) show the anomaly of the isotropic-average curves relative to AK135. We do not plot the isotropic curves up to the very long periods spanned by the simple-average curves because the number of measurements at these periods was insufficient to compute anisotropy accurately. The effect of azimuthal anisotropy, however, is small compared to the phase-velocity anomalies relative to the global average. Therefore, for most purposes, simple-average curves (Figure 3.3A and B) provide useful accuracy up to the very long periods (>200 s).

In all Archean parts of southern Africa sampled by our data, phase velocities are much greater than the global average, both at the short periods that sample the crust and at the long periods that sample primarily the mantle lithosphere, indicating high shear velocities in both layers. The reduction in the anomaly at 15-30 s (Figure 3.3B, D and F) is in part due to the crust being thicker than the 35 km crust in the reference model (Nguuri *et al.*, 2001; Stankiewicz *et al.*, 2002; Nair *et al.*, 2006; Yang *et al.*, 2008; Kgaswane *et al.*, 2009):

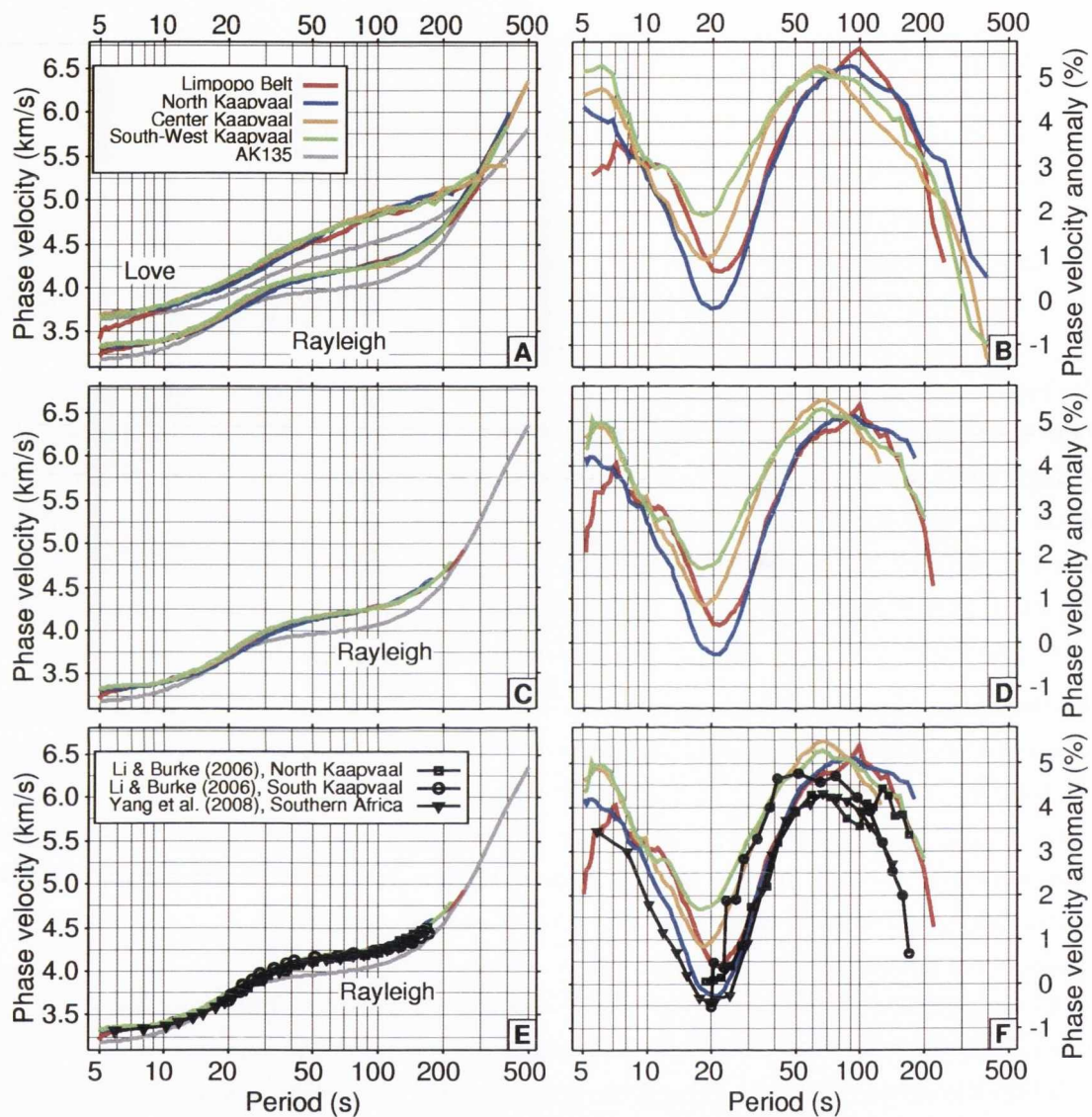


Figure 3.3.: Sub-region-average dispersion curves. A: Rayleigh- and Love-wave dispersion curves for the four sub-regions, computed by simple averaging. B: Same Rayleigh-wave dispersion curves as in A) but presented as the anomaly with respect to AK135 (Kennett *et al.*, 1995), to highlight the differences between the curves. C) Isotropic-average, Rayleigh-wave dispersion curves for the four sub-regions, computed by inversions for azimuthal anisotropy at each period. D) Same curves as in C) but presented as the anomaly with respect to AK135. E) Comparison between our isotropic curves (same as in C) and those from previous studies. F) Same curves as in E) but presented as the anomaly with respect to AK135.



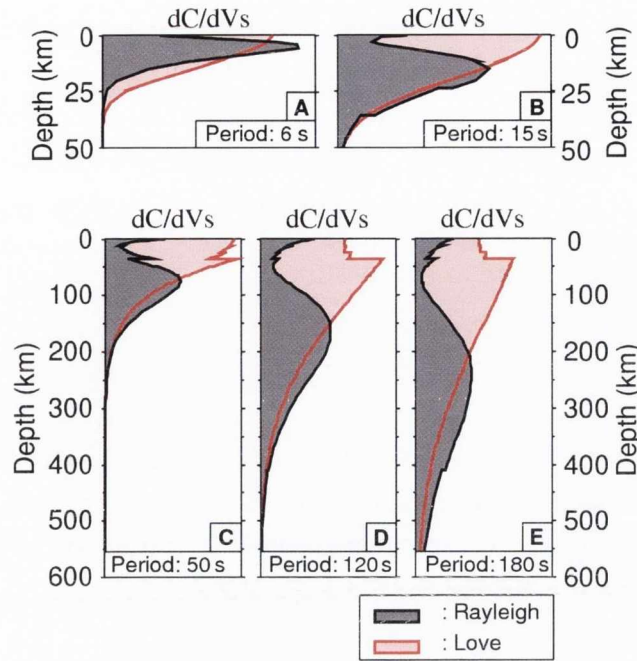


Figure 3.4.: Sensitivity kernels for the fundamental mode Rayleigh (grey) and Love (red) wave phase velocity as a function of depth at five periods (6 s, 15 s, 50 s, 120 s and 180 s). Each graph is scaled independently. The derivatives are computed for a model with a 37 km thick crust.

the northern Kaapvaal Craton and the Limpopo Belt have the thickest crust and show the lowest phase velocities in this period range.

Rayleigh-wave curves published recently by other authors are consistent with our measurements. We compare the published curves and our isotropic-average ones in Figure 3.3 (E and F). The smooth, southern-Africa average dispersion curve of Yang *et al.* (2008) is somewhat slower than all of our curves because it was an average over a larger region around the Kaapvaal Craton, including non-cratonic, southern South Africa. The North-Kaapvaal average curve of Li and Burke (2006) is nearly identical to our northern Kaapvaal and Limpopo curves in most of its 20-180 s period range (apart from its rougher portion with a dip at 70-110 s). The South-Kaapvaal average curve of Li and Burke (2006) is also very similar to our central and southwestern Kaapvaal curves at 25-50 s, although it is 1-2% slower at longer periods. This is probably due to their South-Kaapvaal region extending further South than our sub-regions and sampling the seismically slower southernmost portion of the craton (Yang *et al.*, 2008; Li, 2011).

The Rayleigh-wave curves obtained by different methods are very similar, and the (small) differences between them can be explained by the differences in the boundaries of the regions that the curves average over. This consistency validates the measurements and confirms the accuracy of the currently available Rayleigh-wave data.

The new measurements obtained in this study extend the range of Rayleigh-wave curves to very long periods and also add the broad-band Love-wave curves (Figure 3.3A). The availability of both Rayleigh and Love measurements will be particularly useful for isolating the isotropic-average seismic-velocity structure from the effects of radial anisotropy, so as to relate it to the lithospheric temperature and composition (*e.g.* Khan *et al.*, 2011; Fullen *et al.*, 2012).

### 3.4 INVERSION

Within each sub-region, we measured phase velocities of surface waves travelling at different azimuths. We can use the azimuthal dependence of phase velocities to determine the orientation and amplitude of azimuthal anisotropy, as a function of period, in different parts of the Kaapvaal Craton and in the Limpopo Belt.

#### 3.4.1 Rayleigh wave anisotropy

The azimuthal data coverage is sufficient to constrain azimuthal anisotropy in the period range from 5 s to 150-200 s, depending on the region. The inversion procedure described below is applied independently to data from each sub-region, at every period.

First, we sort phase-velocity measurements into 1-degree bins, according to the station-station azimuth. For example, the North-South direction ( $0^\circ$ ) contains inter-station pairs with azimuths in the range  $[-0.5^\circ; 0.5^\circ]$ . We next use sliding windows in order to smooth the phase-velocity distributions with azimuth. This averages out noise in the measurements and also takes into account the finite station-event back-azimuth window used in the selection of sources for phase-velocity measurements. We choose the half-width of the window of  $15^\circ$ . For example, phase velocity in the North-South direction ( $0^\circ$ ), is an average over all measurements from all inter-station pairs situated at azimuths in the range  $[-15.0^\circ; 15.0^\circ]$ . Such a  $30^\circ$  width window is applied at every  $1^\circ$  step. We have verified that the choice of the width of the window does not influence our results (in the Section 3.10, we test three different window widths and show that the results are robust with respect to the choice of the width). To avoid biases due to larger numbers of measurements at some azimuths compared to others, phase velocity in a window is computed as a weighted average, with weights inversely proportional to the number of measurements at an azimuth. At a period  $T$  and azimuth  $\varphi$ , the averaged phase velocity  $C(T, \varphi)$  is computed as:

$$C(T, \varphi) = \frac{1}{\sum_i w_i(T)} \sum_{i=\varphi-15}^{\varphi+15} w_i(T) \cdot \sum_j C_j(T, i) \quad (3.1)$$



with

$$w_i(T) = \frac{1}{n_i(T)} \quad (3.2)$$

where  $w_i(T)$  is the weight applied at the azimuth  $i$  ( $i \in \mathbb{Z}$ ) and  $n_i(T)$  is the number of measurements at azimuth  $i$ .  $C_j(T, i)$  is the  $j^{\text{th}}$  phase-velocity measurement at the azimuth  $i$  and period  $T$ .

For a weakly anisotropic medium, azimuthal anisotropy of surface waves is described (Smith and Dahlen, 1973) as a harmonic function of the form:

$$C(T, \varphi) = C_0(T) + A_1(T) \cos(2\varphi) + A_2(T) \sin(2\varphi) + A_3(T) \cos(4\varphi) + A_4(T) \sin(4\varphi) \quad (3.3)$$

where  $T$  is the period,  $\varphi$  is the azimuth,  $C$  is the observed phase velocity,  $C_0$  is the isotropic phase velocity and  $A_1, A_2, A_3$  and  $A_4$  are variables that depend on elastic properties of the medium and parameterise anisotropy. At each period, the amplitudes of anisotropy ( $A_{2\varphi}$  and  $A_{4\varphi}$ ) and the fast-propagation azimuths ( $\phi_{2\varphi}$  and  $\phi_{4\varphi}$ ) for the  $2\varphi$  and  $4\varphi$  components of anisotropy, respectively, are calculated as:

$$A_{2\varphi} = \sqrt{A_1^2 + A_2^2} \quad A_{4\varphi} = \sqrt{A_3^2 + A_4^2} \quad (3.4)$$

$$\phi_{2\varphi} = \frac{1}{2} \text{Atan} \left( \frac{A_2}{A_1} \right) \quad \phi_{4\varphi} = \frac{1}{4} \text{Atan} \left( \frac{A_4}{A_3} \right) \quad (3.5)$$

For each period, we solve the Equation 3.3 as  $\mathbf{d} = \mathbf{G} \cdot \mathbf{m}$ , where  $\mathbf{d}$  is the vector that contains the data (phase velocity at each azimuth),  $\mathbf{m}$  is the vector that contains the model  $[C_0, A_1, A_2, A_3, A_4]^T$  and  $\mathbf{G}$  is the sensitivity matrix with the coefficients that link the data and the model. We use the least-squares method with damping to solve this problem. Because our problem is over-determined, we use

$$\mathbf{m} = (\mathbf{G}^T \mathbf{C}_d^{-1} \mathbf{G})^{-1} \mathbf{G}^T \mathbf{C}_d^{-1} \mathbf{d} \quad (3.6)$$

where  $\mathbf{C}_d$  is a diagonal matrix containing the *a priori* errors of the measurements. The *a priori* errors have been estimated as the standard deviation of the measurements, depend-

ing on the number of measurements per azimuth. The error  $\text{err}(T, \varphi)$  at period  $T$  and azimuth  $\varphi$  is

$$\text{err}(T, \varphi) = \sqrt{\frac{\sum_{i=1}^n (C_i(T, [\varphi - 15; \varphi + 15]) - C(T, \varphi))^2}{n}} \quad (3.7)$$

where the summation is over the azimuths within the range defined by the sliding window centred on  $\varphi$ :  $[\varphi - 15; \varphi + 15]$ .  $n$  is the total number of measurements in the window. We also set a minimum error value, equal to the error at the azimuth with the most measurements.

The amplitude of anisotropy we compute is under-estimated somewhat because of the sliding-window averaging (Equation 3.1, Equation 3.2). With a dense azimuthal coverage, the underestimation is well below 10% of the amplitude (as given by the application of a sliding window at an extremum of the cosine function), but the effect can be stronger if the data samples the azimuth range sparsely and unevenly. To estimate how much the amplitudes of anisotropy may be reduced, we apply our algorithm to synthetic data with a known fast direction and amplitude of anisotropy but with azimuthal coverage as in our measured data set. The largest under-estimation amounts to 37% of the amplitude and occurs at very long periods where the azimuthal coverage deteriorates. In the following, we plot and discuss only robust anisotropy results, constrained by data with sufficient azimuthal coverage.

#### 3.4.2 Love-wave anisotropy

Using the same approach as for Rayleigh waves, we also inverted our Love-wave measurements (Figure 3.2) for azimuthal anisotropy. Love-wave phase velocities depend on the horizontally polarized shear velocities, and Love-wave anisotropy should be described by  $4\varphi$  terms only (Trampert and Woodhouse, 2003). Surprisingly, our results showed  $2\varphi$  terms that were larger than  $4\varphi$  terms and gave fast-propagation directions similar to those of Rayleigh waves. This may have been because Love-wave measurements were influenced by the presence of Rayleigh-wave energy on transverse components.

The  $90^\circ$ -periodic azimuthal variations described by the  $4\varphi$  terms (Equation 3.3) are generally more difficult to resolve than the  $180^\circ$ -periodic ( $2\varphi$ ) ones, and any biases due to mode interference, wave front bending, or uncertainties in horizontal component orientations will make the task more difficult yet. Even with the large amounts of Love-wave measurements in each region (Figure 3.2), our dataset does not appear to constrain Love-wave azimuthal anisotropy accurately. We thus focus on Rayleigh-wave anisotropy only.

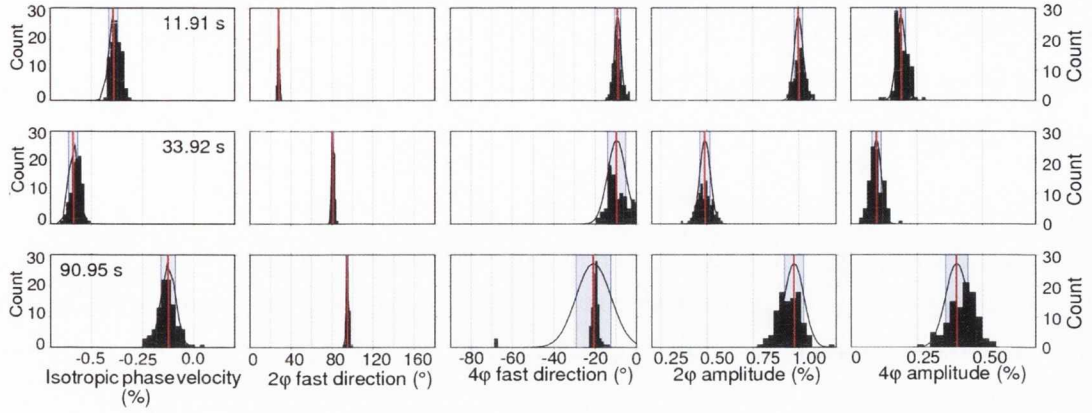


Figure 3.5.: Examples of jackknife distributions of the isotropic and four anisotropic parameters in the Limpopo Belt region at three different periods. Red lines indicate the maximum of the best-fitting Gaussian. Blue areas indicate the half-width of the Gaussian, used to estimate the errors on the parameters.

### 3.5 RESOLUTION AND ERRORS

#### 3.5.1 Jackknife

We estimate the *a posteriori* errors on the parameters of the model  $([C_0, A_1, A_2, A_3, A_4]^T)$  using the jackknife method (Efron and Tibshirani, 1986). With this method, we obtain a first estimate of the lower bound of the error that we can expect on the parameters. The jackknife method provides error estimates by means of repeating the inversion a large number of times (one hundred times in this study), with a dataset from which a randomly selected part of data (30% of the phase-velocity measurements, in this study) has been removed. We remove the random set of data before averaging the data with sliding windows. The distribution of each parameter is expected to follow a Gaussian law, yielding an estimate of the mean and the interval of confidence (half-width of the Gaussian). We use this method to estimate the error on the isotropic-velocity deviation from the reference value and on the fast-propagation azimuths and amplitudes of the  $2\phi$  and  $4\phi$  anisotropy (Equation 3.5). Examples of the results in Figure 3.5 show that, as expected, the values are distributed following Gaussians.

The resulting estimated errors are small for the  $2\phi$  fast propagation azimuth (below  $5^\circ$ ) and are larger for the  $4\phi$  fast azimuth and the amplitudes of anisotropy. The errors increase at long periods because of the decrease in the amount of data.



### 3.5.2 *F-test*

#### 3.5.2.1 *method*

In order to test whether anisotropy is statistically significant (whether  $2\varphi$  or  $4\varphi$  terms are required to explain the data), we perform F-tests. To do so, we first compute the reduced  $\chi^2$  given by the azimuthal anisotropy inversion at every period:

$$\chi^2 = \frac{1}{N - \text{tr}(\mathbf{R})} (\mathbf{d} - \mathbf{G}\hat{\mathbf{m}})^T \mathbf{C}_d^{-1} (\mathbf{d} - \mathbf{G}\hat{\mathbf{m}}) \quad (3.8)$$

where  $N$  is the number of data and  $\mathbf{R}$  is the resolution matrix. The resolution matrix is an identity matrix and its size is equal to the number of unknown (its size is equal to one if we consider only an isotropic inversion, three if we consider an isotropic and the  $2\varphi$  terms and five if we consider a full harmonic function like in the Equation 3.3). The values of  $\chi^2$  are a measure of how well the model fits the data; if the model fits the data well,  $\chi^2$  values tend to be close to 1. A comparison of  $\chi^2$  given by isotropic and anisotropic inversions (with  $2\varphi$  only or both  $2\varphi$  and  $4\varphi$  terms) should show if the data are explained significantly better with anisotropic models, compared to an isotropic one.

We use  $\chi^2$  tests instead of a comparison of variance reductions because we are comparing inversions with different degrees of freedom (an isotropic one has only one parameter and one degree of freedom; anisotropic inversions involve three or five parameters, depending on whether  $2\varphi$  only or both  $2\varphi$  and  $4\varphi$  terms are included).

We perform four series of inversions: first, for isotropic phase-velocity perturbations only; second, for isotropic and  $2\varphi$  anisotropic terms; third, for isotropic and  $4\varphi$  anisotropic terms; and fourth, for isotropic,  $2\varphi$  and  $4\varphi$  terms. F-tests are then performed in order to determine whether the reduction of the  $\chi^2$  between inversions with different degrees of freedom is significant or not (Bevington, 1969).

The ratio of the statistics ( $\chi^2$ ) should follow the Fisher-Snedecor law distribution depending on the number of degrees of freedom in the inversions. For example, if we test whether or not a  $2\varphi$  anisotropic inversion fits the data better than an isotropic inversion, the F-test value will be:

$$F_{1,3} = \frac{\chi_{\text{iso}}^2}{1} / \frac{\chi_{2\varphi}^2}{3} \quad (3.9)$$



and for a test on whether anisotropic model with both  $2\varphi$  and  $4\varphi$  terms fits better than an isotropic model it will be:

$$F_{1,5} = \frac{\chi_{\text{iso}}^2}{1} \bigg/ \frac{\chi_{2\varphi+4\varphi}^2}{5} \quad (3.10)$$

The greater the F-value, the more significant the  $\chi^2$  reduction. In order to gauge statistical significance, we compare the F-values to threshold values, which depend on the degrees of freedom and on the confidence index we choose. We set the confidence index to the commonly used value of 0.05, meaning that the F-value is 5% probable to be an outlier. For a more conservative test, we also set the confidence index to 0.01 (Figure 3.6); the F-value is then only 1% probable to be an outlier (Bevington, 1969). For the test to be significant,  $F_{1,3}$  should be at least 10.13 if a confidence index of 0.05 is assumed and at least 34.12 for a confidence index of 0.01.  $F_{1,5}$  must be at least 6.61 with a confidence index of 0.05 and at least 16.26 with a confidence index of 0.01 (Bevington, 1969).

### 3.5.2.2 F-test results for Rayleigh waves

Figure 3.6 illustrates the anisotropy patterns (A,B) constrained by the inversions of the data from the four regions for the isotropic,  $2\varphi$  and  $4\varphi$  coefficients at different periods, as well as the statistical significance of anisotropy (C-E). The amplitude of anisotropy (Figure 3.6, frames A, B) is given both by the position of the small circles along the y-axis and by the length of the sticks centred on the circles. The fast-propagation direction is given by the orientation of the sticks (as if on a map): vertical means N-S and horizontal means E-W. For the  $4\varphi$  patterns, only one of the two perpendicular fast directions is plotted, for clarity. In Figure 3.6C we compare the misfits given by different inversions, parameterised using a full set as well as different subsets of the isotropic,  $2\varphi$  and  $4\varphi$  coefficients.

The  $4\varphi$  terms are very small almost everywhere, and so are the reductions in misfits that result from adding the  $4\varphi$  terms as inversion parameters (Figure 3.6C). The  $4\varphi$  anisotropy is expected to be weak for long-period Rayleigh waves that sample the olivine-dominated upper mantle (Montagner and Tanimoto, 1991; Trampert and Woodhouse, 2003). An absence of statistically significant  $4\varphi$  anisotropy has been seen in a number of previous regional-array studies (e.g. Deschamps *et al.*, 2008b; Darbyshire and Lebedev, 2009), and our analysis confirms these findings.

In the crust, the  $4\varphi$  terms may play a greater role (Polat *et al.*, 2012). According to our results, however, even in the Limpopo Belt, where the  $4\varphi$  terms are the largest, both their amplitude and statistical significance are well below those of the dominant  $2\varphi$  terms. In the following, we shall thus discuss and interpret only the  $2\varphi$  anisotropy patterns.

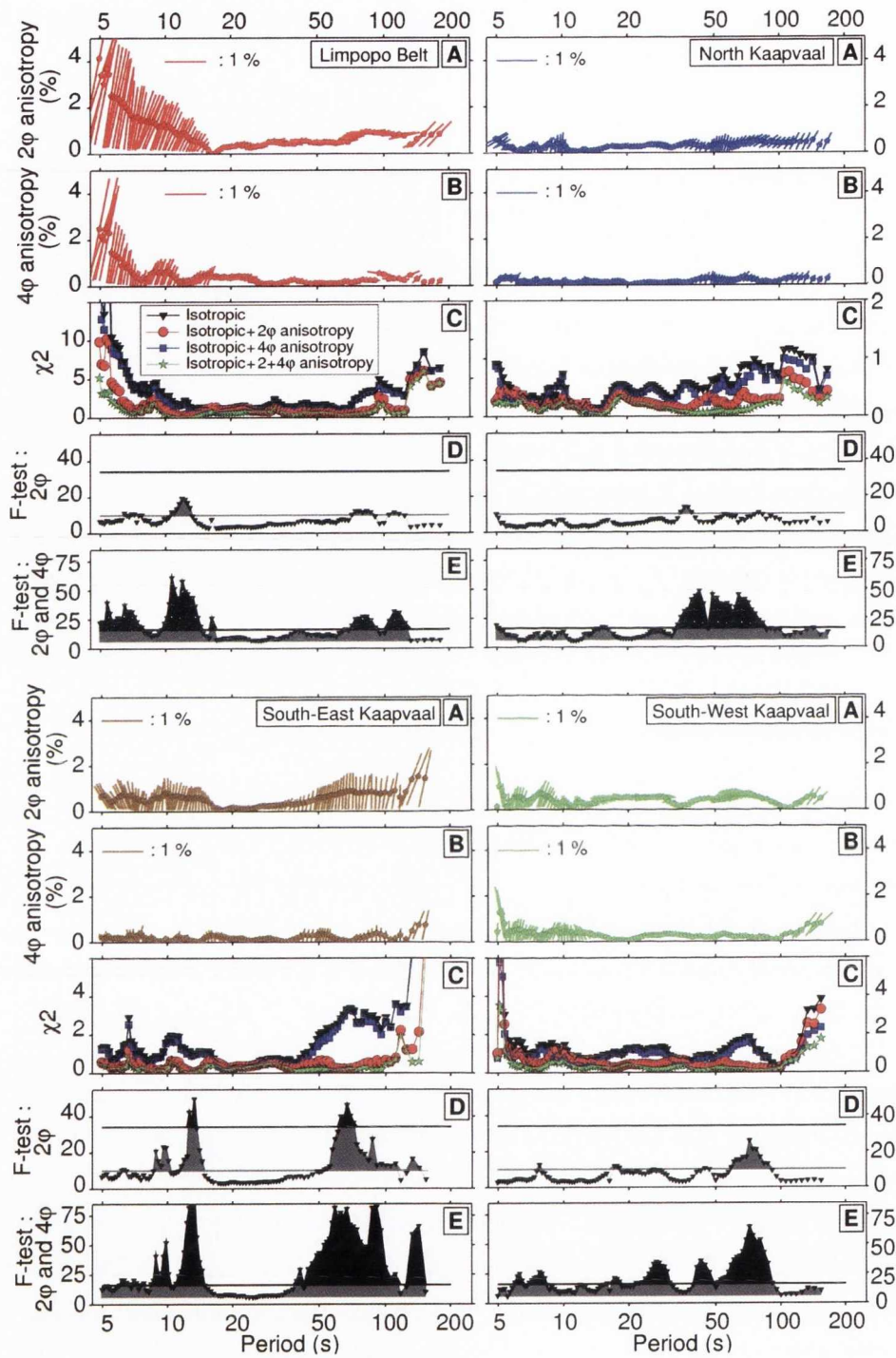


Figure 3.6.: Results of the inversions and F-tests for the four sub-regions. A) Azimuths of fast-propagation directions and amplitudes of anisotropy, the  $2\phi$  terms. The inversion were for the isotropic and both the  $2\phi$  and  $4\phi$  terms. B) Azimuths of fast direction and amplitudes of anisotropy, the  $4\phi$  terms (for clarity, only one of the two perpendicular fast directions is shown). C) Reduced  $\chi^2$  of the inversions considering only the isotropic part (black lines and triangles), the isotropic part and  $2\phi$  anisotropy (red lines and circles), the isotropic part and  $4\phi$  anisotropy (blue lines and squares) and the isotropic part with both 2 and  $4\phi$  anisotropy (green lines and stars). D) F-tests on whether the  $2\phi$  terms are required, compared to an isotropic inversion. E) F-tests on whether the  $2\phi$  and  $4\phi$  terms are required, compared to an isotropic inversion. Shaded areas indicate significance, according to the F-test. Grey: above a confidence index of 0.05, black: above a confidence index of 0.01.



Inversions with both  $2\varphi$  and  $4\varphi$  terms included can produce F-values that are much larger compared to those from inversions without anisotropy (Figure 3.6, Figure 3.15). These F-values are also substantially larger than the F-values from inversions with either  $2\varphi$  or  $4\varphi$  terms only. The addition of the  $4\varphi$  terms by itself, however, produces only a small reduction in the misfits, as illustrated by Figure 3.6C and also by our additional F-tests, presented in Section 3.11 for completeness. The only exception is in the period range 33-83 s for the North Kaapvaal Craton region, where adding  $4\varphi$  anisotropy to the  $2\varphi$  anisotropy improves the fit significantly (Figure 3.15C).

The  $\chi^2$  misfits (Figure 3.6, frames C) show that the introduction of anisotropy reduces misfits substantially in some period ranges. At other periods, an isotropic model is sufficient to explain the data, with the  $\chi^2$  values given by isotropic inversions close to 1.

F-test results (Figure 3.6D and E) indicate whether or not adding  $2\varphi$  (D) or  $2\varphi$  and  $4\varphi$  (E) anisotropy to the models improves the fit to the data, compared to an isotropic model. Northern Kaapvaal Craton does not show significant anisotropy at short periods. Phase-velocity anisotropy in the Limpopo Belt and southern Kaapvaal regions shows different patterns in three distinct period ranges. At periods of 5-17 s and greater than 37 s, these regions are anisotropic. Between 17 and 37 s, however, no phase-velocity anisotropy is required by the data (the F-test values are smaller than the confidence level). Anisotropy in the southwestern Kaapvaal Craton is required at both intermediate and long periods but is less significant at short periods below 17 s.

The F-test appears to be an effective way to identify the period ranges in which the data requires azimuthal anisotropy.

## 3.6 AZIMUTHAL ANISOTROPY

### 3.6.1 *Anisotropy in the four regions*

The results of the inversions for Rayleigh-wave azimuthal anisotropy are displayed in Figure 3.7. At the top, we show examples of data and best-fitting models at four different periods, for each of the four regions. The data (colour dots) are sorted into  $1^\circ$  bins and smoothed using a  $30^\circ$  sliding window (Section 3.4). The best-fitting models with the isotropic and  $2\varphi$  anisotropy parameters are shown with solid lines; models with the isotropic and both  $2\varphi$  and  $4\varphi$  anisotropy parameters are shown with dashed lines. See Appendix A for phase velocity distribution with azimuth at every periods.



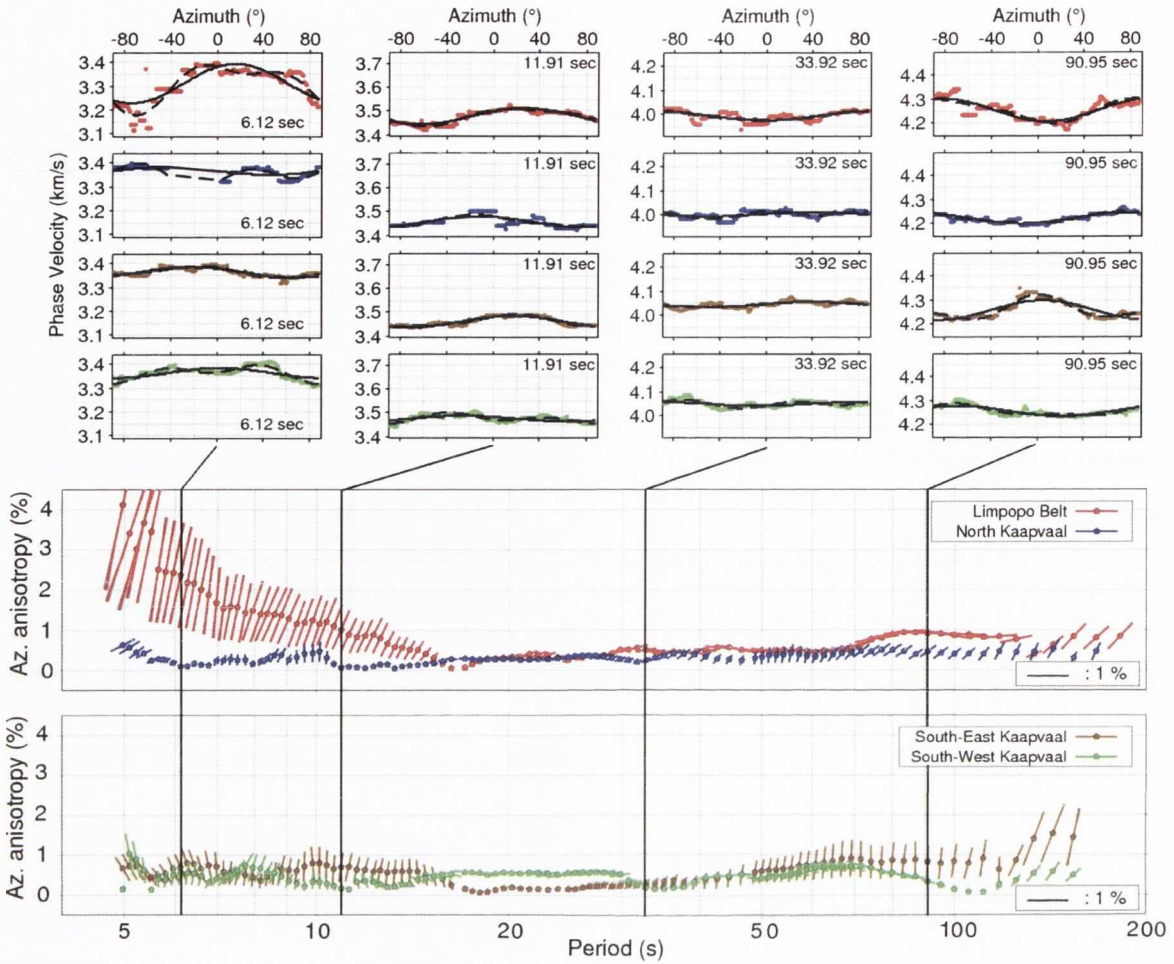


Figure 3.7.: Azimuthal anisotropy in the four sub-regions. Top: Examples of distributions of phase velocities with azimuths at four different periods for each sub-region. Dots are the data, binned and smoothed with a  $30^\circ$  sliding window. Solid black lines: best-fitting models with isotropic and  $2\phi$  terms. Dashed black lines: best-fitting models with isotropic,  $2\phi$  and  $4\phi$  terms (see [Appendix A](#) for phase velocity distribution with azimuth at every periods). Bottom: Azimuths of fast propagation and amplitude of anisotropy as a function of period for each sub-region. The vertical orientation of a stick means a North-South fast-propagation azimuth; a horizontal orientation means an East-West fast-propagation azimuth.

Anisotropy is small in most cases. The examples in [Figure 3.7](#) (top) show that the small amplitude of anisotropy is not an artefact due to noise or inconsistency of measurements but is clearly seen in the data. The small amplitudes are qualitatively consistent with the small SKS splitting times measured in southern Africa ([Fouch \*et al.\*, 2004b](#); [Silver \*et al.\*, 2004](#)).

The two frames at the bottom of [Figure 3.7](#) show the distribution of Rayleigh-wave anisotropy ( $2\phi$ ) with period in the four regions. Accurate measurements of anisotropy must show smooth, gradual changes with period, both in their amplitudes and in their fast-propagation azimuths (unless the amplitude is close to zero). This reflects the grad-

ual changes with period of the shapes and depth spans of the depth sensitivity kernels of Rayleigh-wave phase velocities (Figure 3.4). Large anisotropy differences at adjacent periods cannot be reconciled with realistic Earth structures and indicate errors. In our models, some unrealistic variability, albeit with small anisotropy amplitude, is seen at the shortest periods (5-10 s) in the northern and southwestern Kaapvaal Craton regions. Most of our models show gradual changes with period, as expected.

The largest anisotropy by far is seen at the short (5-10 s) periods in the Limpopo Belt, where the fast-propagation direction is approximately N-S. Rayleigh waves at these periods sample primarily the upper and middle-crust (Figure 3.4). In the southern Kaapvaal Craton, fast directions at periods under 10 s are also around N-S on average, although with a much smaller amplitude. In the northern Kaapvaal Craton, the crust appears to be nearly isotropic.

Intermediate periods of 15-50 s sample the lower-crust and uppermost mantle (Figure 3.4). At these periods, phase-velocity anisotropy is very small (under 0.5%) in three (Limpopo Belt, northern Kaapvaal Craton and central Kaapvaal Craton) of the four regions.

At long periods that sample primarily the lower lithosphere (60-120 s) (Figure 3.4), anisotropy increases to 0.5-1.0%. In all four regions, anisotropy shows nearly constant orientations across most or all of this period range. The fast-propagation directions are dissimilar in different regions: E-W in the Limpopo Belt and southwestern Kaapvaal; NE-SW in northern Kaapvaal; N-S in central Kaapvaal Craton.

At periods longer than 120-130 s, the fast-propagation directions in all regions change towards NNE-SSW. At these periods, Rayleigh-wave sensitivity depth range gradually shifts from mostly-shallower to mostly-deeper than 200 km, increasingly sampling the asthenosphere (Figure 3.4).

### 3.6.2 *Southern-Africa average anisotropy*

The change in anisotropy orientation at periods of 120-130 s is important as it appears to reflect the change of the deformation regime at the lithosphere-asthenosphere boundary. In each of the four regions, however, our inversion results at the longest periods are less robust than at periods below, due to the decrease in the amount of measurements.

We therefore merge the dispersion curves from the four sub-regions and invert the resulting large dataset for southern-Africa-average anisotropy. Additional dispersion curves from longer inter-station paths that cross region boundaries are added (typically, longer paths are likely to yield more long-period measurements). The combined dataset consists of 5313 dispersion curves, from 646 station pairs between 133 stations (Figure 3.8). The dataset is inverted for region-average azimuthal anisotropy (Section 3.4). Figure 3.9 shows fast-propagation azimuths and amplitudes of anisotropy, with examples of data-model fits at three periods.



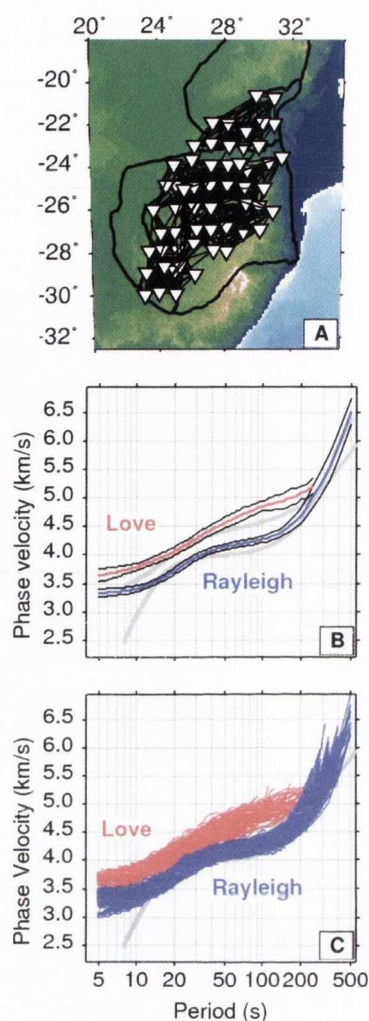


Figure 3.8.: Phase-velocity measurements across the entire region. A) Stations and stations pairs. B) All-region-average dispersion curves for Love and Rayleigh waves. Grey: AK135 (Kennett *et al.*, 1995). C) All dispersion curves, each measured using a pair of stations and a single event.

The N-S fast directions at 5-15 s periods reflect the strong anisotropy in Limpopo Belt and, to a lesser extent, that in the southern Kaapvaal Craton, all with approximately N-S orientation. At periods that sample the lower-crust and mantle lithosphere (15-100 s), the region-average anisotropy is an average over blocks with different fast directions within each (Figure 3.7) and is small (Figure 3.9).

At periods longer than 100-120 s, a robust, region-average, NE-SW direction of fast propagation emerges. This result confirms the approximately NE-SW fast directions yielded by region-by-region inversions, but is more robust, constrained by more data. Anisotropy at the longest periods is likely to be due to fabric in the asthenosphere (Figure 3.4), created by shear associated with the current motion of the African Plate. Flow patterns in



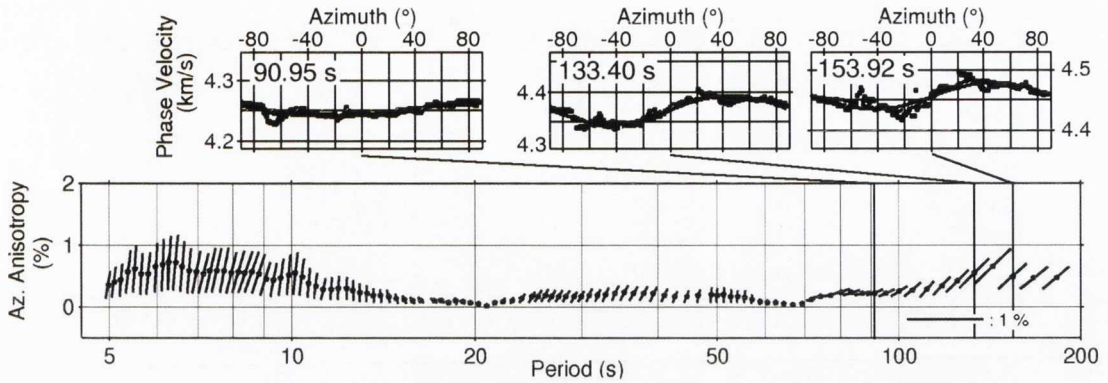


Figure 3.9.: All-region-average azimuthal anisotropy. Top: Example distributions of phase velocities with azimuths at three different periods (the long periods is where all-region average anisotropy is most meaningful). Dots are the data, binned and smoothed using a  $30^\circ$  sliding window. Solid black lines: best-fitting models with isotropic and  $2\varphi$  terms. Dashed black lines: best-fitting models with isotropic,  $2\varphi$  and  $4\varphi$  terms (see [Appendix A](#) for phase velocity distribution with azimuth at every periods). Bottom: the azimuth of fast propagation and amplitude of anisotropy as a function of period. The vertical orientation of a stick means North-South, and horizontal orientation means East-West fast-propagation azimuth.

the asthenosphere can be expected to be smooth, and thus the inversion for a single fast direction beneath the entire southern Africa is meaningful.

### 3.6.3 An alternative inversion approach

To verify our results further, we also invert the data using an alternative approach: tomographic inversions for phase-velocity maps with anisotropy (Lebedev and VanDerHilst, 2008; Deschamps *et al.*, 2008b). This is done for each region separately, using inter-station measurements within that region only and applying very strong lateral smoothing. The smoothing represents lateral averaging over various neighbouring inter-station paths. The inversions thus enhance the averaging that had been done for each station pair over the different one-event measurements available for it ([Section 3.3](#)). The solutions of the tomographic inverse problems (solved separately at different periods) are isotropic-average phase velocities and azimuthal anisotropy within each region. We plot the results of the inversions at a range of periods in [Figure 3.10](#). The inversions were over-parameterised (the sticks that indicate the fast directions and amplitudes of anisotropy are plotted at each grid knot) but also over-smoothed, so that there are almost no variations within each sub-region. We plot anisotropy at all grid knots only to highlight the degree of model smoothness.

The results of the over-smoothed tomography are consistent with the results of our inversions for sub-region-average properties ([Section 3.4](#)). This is to be expected, as the two inversion approaches relate the same data to the same properties of the regions, the

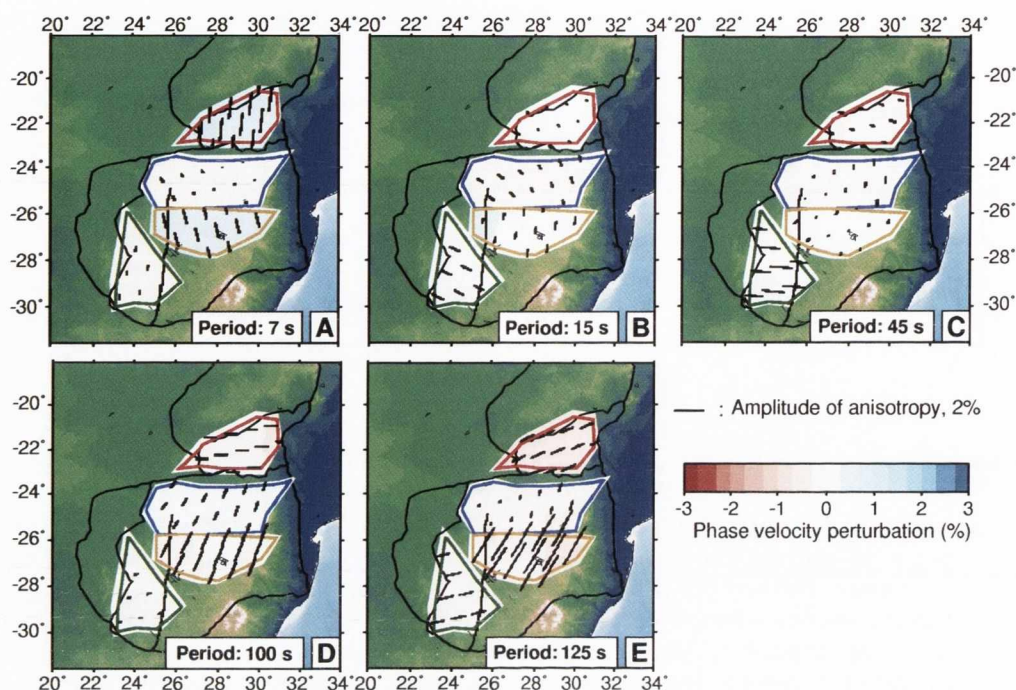


Figure 3.10.: Strongly smoothed phase-velocity tomography at five different periods. Separate inversions are performed for each region, with the results then plotted on the same maps for compactness. Sticks indicate directions of fast propagation. Isotropic phase-velocity anomalies (colour scale) are with respect to the averages over the entire region (Figure 3.8).

region-average anisotropy being of primary interest. The implementation of the two inversions is quite different, however, and the close match between the results confirms their accuracy.

### 3.7 DISCUSSION

The dependence of phase-velocity anisotropy on period indicates the distribution of shear-velocity anisotropy with depth. Although resolving the layering of anisotropy in fine detail would require an investigation of the trade-offs of various anisotropic and isotropic parameters, our phase-velocity results (Figure 3.7, Figure 3.9) display patterns that already reveal clearly the basic elements of the stratification of azimuthal anisotropy beneath southern Africa.

We summarise our inferences in Figure 3.11 and focus on three depth ranges: the upper-crust, lower mantle lithosphere, and asthenosphere. We plot average phase-velocity anisotropy across period ranges that are primarily sensitive to these depth ranges. These averages are meaningful because anisotropy patterns are clear and nearly constant within each of the period ranges, for each of our four regions (Figure 3.7, Figure 3.9). In contrast, phase-velocity anisotropy at periods sampling the lower-crust and upper lithospheric man-



tle (between 10-15 and 50-70 s) is small and more variable (Figure 3.7). This may reflect either the small amplitude or greater depth variability of shear-velocity anisotropy in this depth range.

### 3.7.1 *Upper crust: azimuthal anisotropy, aligned cracks, tectonic stress*

The strong phase-velocity anisotropy at short periods in the Limpopo Belt indicates shear-velocity anisotropy in the upper-crust. This is evident from the gradual decrease in the anisotropy amplitude from 5 s to 10 s period (Figure 3.7), as Rayleigh waves at increasing periods sample progressively deeper crust (Figure 3.4).

The N-S fast-propagation direction (Figure 3.9A) is perpendicular to the E-W sutures between the Limpopo Belt and the Kaapvaal and Zimbabwe Cratons to the South and North of it. It is also perpendicular to the E-W oriented fabric in the deep lithosphere, manifested in anisotropy we see at longer periods. It is thus unlikely to be caused by fabric imprinted into the crust during the Archean-Paleoproterozoic deformation of the Limpopo Belt.

We conclude that the anisotropy is caused by aligned cracks. Cracks in the upper 5 km of the crust can be opened by the current tectonic stress, giving rise to strong anisotropy (e.g. Nur, 1971; Crampin, 1978) with fast-propagation directions parallel to the cracks and perpendicular to the extensional stress component. In eastern southern Africa, the regional stress is characterised by an E-W extension, associated with the South-ward propagation of the East African Rift. This stress field is documented unambiguously by the source mechanisms of numerous earthquakes, showing E-W extension (Figure 3.11A).

The crack-induced anisotropy is the strongest in the Limpopo Belt probably because it is the closest to the tip of the rift. In southwestern and central Kaapvaal Craton, upper-crustal anisotropy is weaker but also with fast directions perpendicular to the regional direction of extension, evidenced by focal mechanisms (e.g. Dziewonski *et al.*, 1994) (Figure 3.11a), and is thus likely to be due to cracks as well.

For the first time, to our knowledge, our results open the possibility of estimation of regional stress patterns from anisotropy measured using teleseismic surface waves. In southern Africa, the stress state is well known from earthquake source mechanisms. Our results suggest that in other regions around the world, where tectonic stress is poorly known, it can be estimated using broadband-array recordings of short-period surface waves, a very common data type.

### 3.7.2 *Anisotropy in deep mantle lithosphere: record of ancient deformation*

Rayleigh waves at 80-120 s sample primarily the lower mantle lithosphere (100-200 km depths) (Figure 3.4) and show an E-W fast-propagation direction beneath the Limpopo



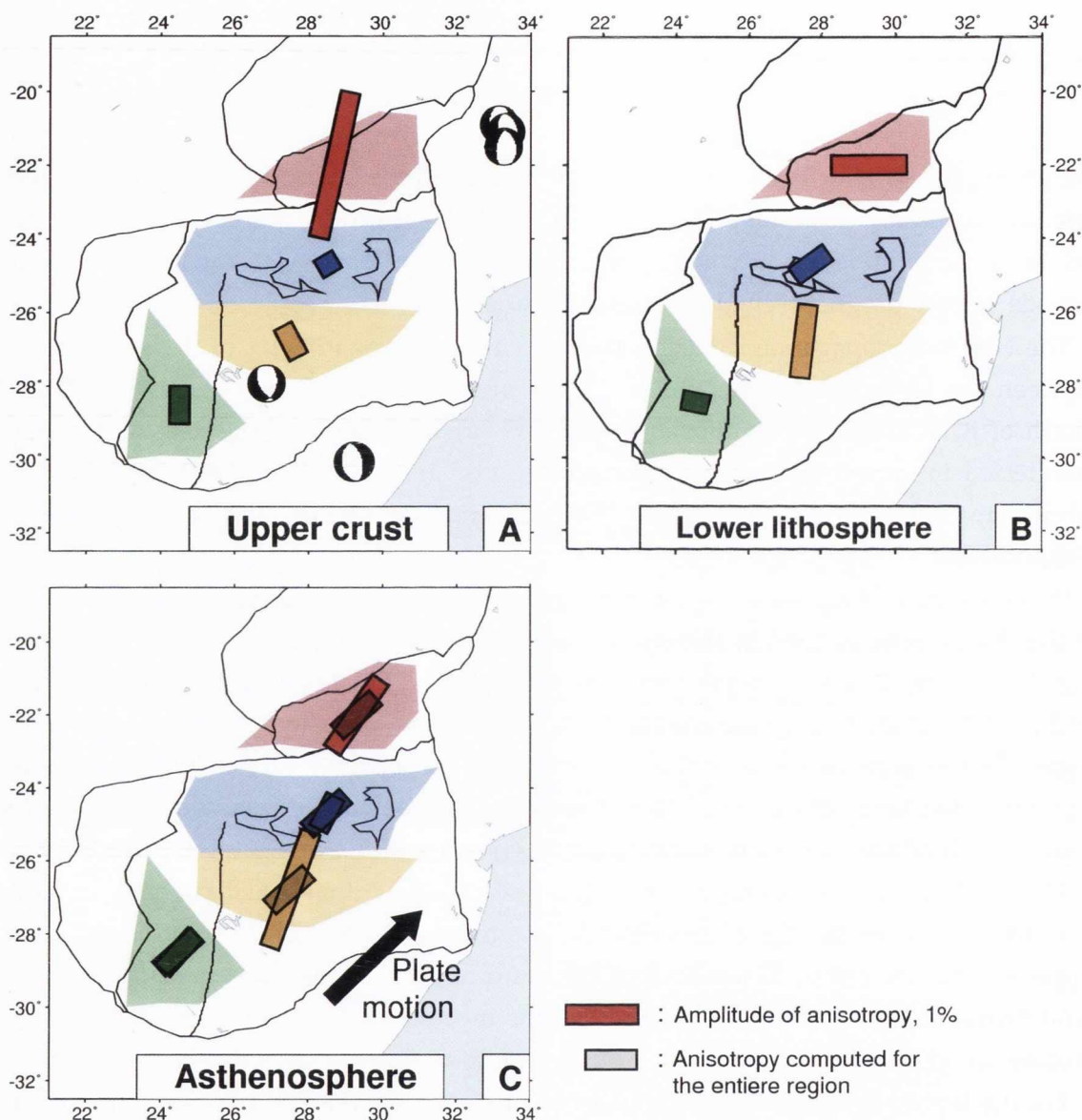


Figure 3.11.: Azimuthal anisotropy of Rayleigh waves sampling primarily the upper crust (A), lower lithosphere (B) and asthenosphere (C). A: Fast-propagation directions of Rayleigh waves at 5-7 s periods. Focal mechanisms of earthquakes are plotted to indicate the regional tectonic stress (*e.g.* Dziewonski *et al.*, 1994). B: Fast-propagation directions at 80-120 s periods, with sensitivity primarily to the deep mantle lithosphere. C: Fast-propagation directions at 130-160 s periods, with sensitivity primarily below 200 km. The plate motion direction is from Gripp and Gordon (1990). Grey bars: fast directions from the inversion for all-region-average anisotropy.

Belt (Figure 3.11B). Beneath the neighbouring northern Kaapvaal Craton, the average fast-propagation direction is similar, ENE-WSW. These fast directions are parallel to the sutures associated with the Archean and Paleoproterozoic orogenies in which what is now the northern Kaapvaal Craton and Limpopo Belt accreted to the Kaapvaal Craton's core (De Wit *et al.*, 1992; Roering *et al.*, 1992; Kamber *et al.*, 1995; Holzer *et al.*, 1998; Silver *et al.*, 2004). The suture of the Limpopo Belt with the Zimbabwe Craton to the North is also late Archean or Paleoproterozoic and also strikes approximately E-W. The anisotropy is likely to indicate suture-parallel fabric which is a record of the ancient deformation of the deep lithosphere.

Western Kaapvaal Craton (the Kimberly Block) also joined the Kaapvaal Craton's core in the late Archean, with the suture between the two now oriented N-S. Western Kaapvaal's lower-lithosphere anisotropy, however, does not show a fast-propagation direction parallel to the suture; it is oriented nearly perpendicular to it. This suggests that the continental collision of the western and central blocks of the Kaapvaal Craton was accommodated by a different mechanism compared to the collisions in the North. The entire lithosphere of the Limpopo and northern Kaapvaal blocks has undergone pervasive deformation, as evidenced by the suture-parallel fabric frozen into it. In contrast, the lithosphere of the western Kaapvaal Craton shows fabric oriented perpendicular to the suture. This may indicate that the Kimberly Block was stronger at the time of the collision, or, alternatively, that it was not subjected to compression or transpression as the northern Kaapvaal and Limpopo blocks were.

### 3.7.3 Anisotropy in the asthenosphere: current flow

At periods longer than 120–130 s (Figure 3.4), Rayleigh waves sample primarily below 200 km depth. At these periods, anisotropy in all sub-regions aligns approximately with the NE-SW motion of the African Plate (Figure 3.11C). The match is confirmed by the inversions of all available data for all-region-average anisotropy, yielding robust, NE-SW fast-propagation directions at long periods (Figure 3.11C, grey bars).

The long-period anisotropy thus indicates fabric in the asthenosphere, created by the shear associated with the current plate motion. Azimuthal anisotropy aligned with the plate motion has been detected previously beneath other cratons, including those in Australia (*e.g.* Debayle *et al.*, 2005) and North America (Snyder and Bruneton, 2007; Deschamps *et al.*, 2008a; Darbyshire and Lebedev, 2009; Yuan and Romanowicz, 2010). In the Kaapvaal Craton, many of the deepest-sampling xenoliths show a sheared texture, which may be evidence of asthenospheric flow fabric (*e.g.* James *et al.*, 2004). Our results confirm that plate motions produce detectable anisotropic fabric within the asthenosphere beneath cratons.



Evans *et al.* (2011) investigated the electrical resistivity structure of the Kaapvaal Craton and reported indications for electrical anisotropy near the lithosphere-asthenosphere boundary, with higher conductivities in a direction parallel to the plate motion. This could be interpreted as evidence for plate-motion parallel flow, although Evans *et al.* (2011) noted that the anisotropy was not quite statistically significant. In an earlier study, Hamilton *et al.* (2006) found highly variable orientations of apparent electrical anisotropy in the crust and lithospheric mantle, which, however, was likely to be due to a combination of factors, including isotropic heterogeneity.

### 3.7.4 *Origin of SKS splitting beneath southern Africa*

Previous SKS splitting measurements in southern Africa (Figure 3.12) prompted two different end-member interpretations, with important implications regarding the deformation of the lithosphere and asthenosphere. According to one interpretation (Vinnik *et al.*, 1995), anisotropy is present in the asthenosphere, reflecting the shear associated with the plate motion, while frozen anisotropy in the ancient lithosphere may also occur but is not dominant. According to the other interpretation (Silver *et al.*, 2004), anisotropy is present in the lithosphere and is weak or absent in the asthenosphere, with the implication that fast-propagation directions inferred from SKS splitting show the orientation of fabric within the lithosphere. This allowed the inference that late-Archean-Paleoproterozoic orogenies in southern Africa were accommodated by pervasive suture-parallel flow in the mantle lithosphere of both the Kimberly and Pietersburg blocks, now in the western and northern parts of the Kaapvaal Craton, respectively, as well as in the lithosphere of the Limpopo Belt to the North. This ancient deformation thus created anisotropic fabric that remains frozen within the deep lithosphere until present.

Our results provide the vertical resolution that the SKS measurements lack. They show (Figure 3.12) that the distribution of anisotropy beneath southern Africa comprises elements of both end-member models. Beneath the northern Kaapvaal Craton and Limpopo Belt, the E-W to ENE-WSW fast propagation directions inferred from SKS splitting match those within the deep mantle lithosphere (Figure 3.12A) and not those within the asthenosphere (Figure 3.12B). This shows that SKS splitting originates within the lithosphere and confirms that the entire lithosphere is likely to have been reworked in the Archean-Paleoproterozoic orogenies (De Wit *et al.*, 1992; Roering *et al.*, 1992; Kamber *et al.*, 1995; Holzer *et al.*, 1998; Silver *et al.*, 2004) undergoing suture-parallel flow, in agreement with the interpretation of Silver *et al.* (2004).

In the Kimberly Block in western Kaapvaal Craton, in contrast, fast-propagation directions inferred from SKS splitting do not match those within the mantle lithosphere (Figure 3.12A) but, instead, match fast-propagation directions within the asthenosphere (Figure 3.12B). SKS splitting in western Kaapvaal is thus likely to originate primarily in



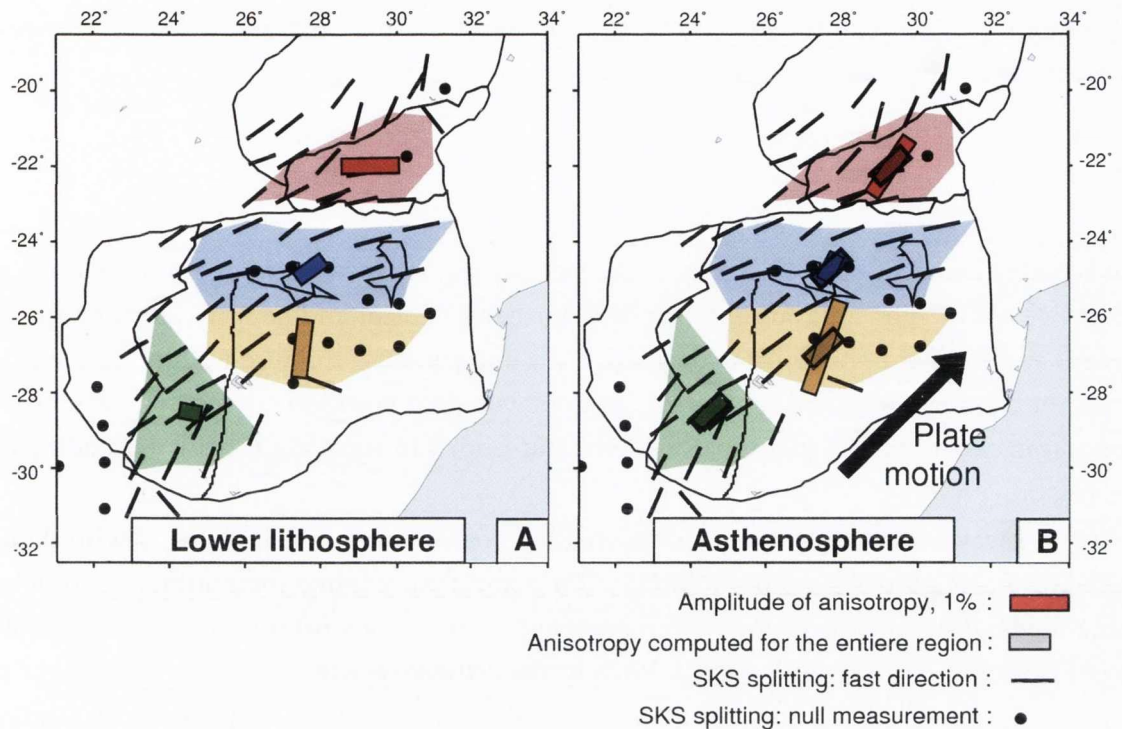


Figure 3.12.: Origin of SKS splitting: a comparison of azimuthal anisotropy in the lithosphere and asthenosphere and the fast directions inferred from SKS splitting. A) Fast-propagation directions at 80-120 s periods, with sensitivity primarily to the deep mantle lithosphere, compared to the fast directions from SKS splitting (black) (Silver *et al.*, 2004). B) Fast-propagation directions at 130-160 s periods, with sensitivity primarily below 200 km (asthenosphere), compared to the fast directions from SKS splitting. The plate motion direction is from Gripp and Gordon (1990).

the asthenosphere and reflects the present plate-motion related flow within it, as suggested by Vinnik *et al.* (1995). The lack of suture-parallel fabric within the lithosphere of the Kimberly block suggests that it was not reworked when it joined the Kaapvaal Craton, unlike the blocks that accreted in the North. This may have been either due to the mechanical strength of this block or because the collision mechanism was different from that in the North.

Small-scale variations in the amplitude of SKS splitting near Kimberley observed by Fouch *et al.* (2004b) are likely to reflect small-scale variations in anisotropy within the lithosphere. Our results represent average anisotropy over relatively large blocks within the Kaapvaal Craton and should be compared to the smooth, regional-scale spatial patterns shown by SKS splitting measurements (Figure 3.11).

Note added in revision: in a just-published study, Vinnik *et al.* (2012) report their joint analysis of P receiver functions and SKS splitting at 23 SASE stations within and around the Limpopo Belt. They conclude that the E-W anisotropic fabric within the belt's lithosphere is underlain by a NE-SW orientated fabric within the asthenosphere. The close

agreement of their results with our results for this part of southern Africa further confirms the occurrence of stratified azimuthal anisotropy.

### 3.8 CONCLUSIONS

We have measured phase velocities of surface waves between pairs of SASE stations in southern Africa. For four sub-regions (the northern, central and southwestern Kaapvaal Craton and the Limpopo Belt), with relatively homogeneous structure within each, we derived robust, Rayleigh- and Love-wave, region-average dispersion curves in a very broad period range: from 5 to 250-400 s (Rayleigh) and from 5 to 100-250 s (Love), depending on the region.

Surface-wave anisotropy indicates the distribution of azimuthal anisotropy with depth in the different parts of southern Africa. Our azimuthal coverage was sufficient to determine Rayleigh-wave anisotropy within each region at periods from 5 to 150-200 s, resolving anisotropy from the upper-crust down to the asthenosphere.

We used the jackknife method to estimate uncertainties and the F-test to determine whether or not anisotropy was required by the data. The F-test results confirmed that  $2\phi$  anisotropy was required between 5 and 17 s and at periods greater than 35 s. No phase-velocity azimuthal anisotropy was required by the data between 17 and 35 s in three out of four regions.  $4\phi$  anisotropy was also inverted for but was found to be small and not statistically significant in almost all cases.

Strong anisotropy measured in the Limpopo Belt at short periods (5-10 s, decreasing with period) is probably due to aligned micro-cracks. The regional E-W extensional stress, associated with the South-ward propagation of the East African Rift, is evident from earthquake source mechanisms and is likely to have opened N-S oriented cracks. This matches the N-S fast Rayleigh-wave propagation directions that we detect both in the Limpopo Belt and, with a smaller amplitude, in the central and western Kaapvaal Craton. Our results show that it is possible to estimate regional stress from surface-wave anisotropy, measured using broadband-array recordings of short-period surface waves.

Fabric within the deep mantle lithosphere of the Limpopo Belt and northern Kaapvaal Craton is aligned E-W to ENE-WSW, parallel to the Archean-Paleoproterozoic sutures where these blocks got attached to the core of the craton and to each other. Our results confirm the earlier inferences made by Silver *et al.* (2004) from their SKS splitting measurements: (i) that the entire lithosphere of these blocks underwent pervasive deformation with suture-parallel flow during the ancient continental collisions, and (ii) that the fabric created by the Archean-Paleoproterozoic deformation is preserved within the lithosphere to this day. Suture-parallel fabric is absent in the deep lithosphere of the western Kaapvaal Craton (Kimberly block). It was not, therefore, reworked in the collision with the central Kaapvaal Craton, unlike the blocks that accreted in the North. This may have been



either due to its mechanical strength or because the collision mechanism was different from those that operated in the North. The depth resolution of the surface-wave data also provides a solution to the debate over the interpretation of SKS splitting measurements: asthenospheric (Vinnik *et al.*, 1995) *vs.* lithospheric (Silver *et al.*, 2004) dominant anisotropy. The distribution of anisotropy beneath southern Africa comprises elements of both models. Anisotropy in the asthenosphere is present and shows fast directions parallel to the plate motion, as argued by Vinnik *et al.* (1995). Anisotropy in the Limpopo and northern Kaapvaal Craton lithosphere shows fast directions parallel to the Archean-Paleoproterozoic sutures, as suggested by Silver *et al.* (2004). The depth intervals where the SKS splitting originates can be determined by comparing the depth distributions of anisotropy and the fast directions given by the splitting measurements. Beneath the Limpopo Belt and northern Kaapvaal Craton, SKS splitting reflects the E-W and ENE-WSW oriented fabric in the lithosphere, as argued by Silver (1996). Beneath the western Kaapvaal Craton, in contrast, SKS splitting reflects the fabric in the asthenosphere, oriented parallel to the SW-NE plate motion, as argued by Vinnik *et al.* (1995).

### 3.9 ACKNOWLEDGMENTS

The authors would like to thank Jeannot Trampert and Balz Kamber for insightful discussions on misfit statistics and on tectonic evolution of the Limpopo Belt, and Derek Schutt, an anonymous reviewer and the Editor, Gabi Laske, for helpful comments on the manuscript. We are grateful to the Kaapvaal Project seismic team for collecting the data. The facilities of the IRIS Data Management System, and specifically the IRIS Data Management Center, were used for access to waveform and metadata used in this study. Figures were generated with Generic Mapping Tools (Wessel and Smith, 1995). This work was funded by the Dublin Institute for Advanced Study and, in part, by Science Foundation Ireland (grants 08/RFP/GEO1704 and 09/RFP/GEO2550).

### 3.10 APPENDIX 1: WIDTH OF THE SLIDING WINDOWS

In this study we smoothed the measurements as a function of azimuth using sliding windows. This reduces the level of noise, by averaging the data at neighbouring azimuths. A window too wide, however, will reduce the amplitude and may bias the phase of the solutions. With a window too narrow, on the other hand, the amount of averaging may not be sufficient to reduce the level of noise.

We used a window half-width of  $15^\circ$  in all inversions above, but here we test other window widths as well. Figure 3.13 (top) shows the data from the Limpopo Belt region at five periods, smoothed using 3 different window widths, and Figure 3.13 (bottom) shows the results of inversions of the data, including both  $2\phi$  and  $4\phi$  anisotropy terms. The



results show that the fast-propagation directions are weakly sensitive to the choice of the width of the sliding window and confirm that  $15^\circ$  window half-width is a reasonable choice.

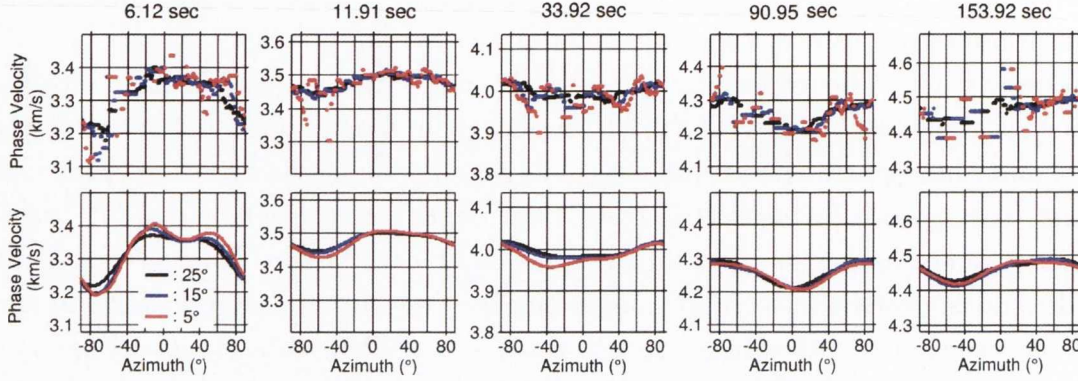


Figure 3.13.: Data (top) and models with  $2\phi$  and  $4\phi$  anisotropy (bottom) for the Limpopo Belt region, at five different periods. Inversions of the data averaged with sliding windows of different widths produce robust results, with weak sensitivity to the window width.

### 3.11 APPENDIX 2: ADDITIONAL F-TESTS

In order to complete the statistical analysis of anisotropy, we present additional tests in Figure 3.15. In contrast to Figure 3.6A(B), where the  $2\phi$  ( $4\phi$ ) anisotropy patterns are obtained from inversions for the isotropic and both  $2\phi$  and  $4\phi$  terms, Figure 3.15A(B) shows  $2\phi$  ( $4\phi$ ) anisotropy given by inversions for only the isotropic and  $2\phi$  ( $4\phi$ ) terms. The two F-tests in Figure 3.15 (C and D) are complementary to the F-tests in Figure 3.6. They test the misfit improvement due to the addition of the  $4\phi$  terms to an isotropic+ $2\phi$  (Figure 3.15 C) and isotropic-only (Figure 3.15D) inversion.

Comparing Figure 3.6A ( $2\phi$  amplitude of anisotropy and azimuth of fast propagation resulting from an inversion for  $2\phi$  and  $4\phi$  anisotropy) and Figure 3.15A ( $2\phi$  anisotropy parameters resulting from an inversion for  $2\phi$  anisotropy only), we find the  $2\phi$  anisotropy parameters to be nearly identical, both in their amplitude and direction of fast propagation. The  $2\phi$  anisotropy in southern Africa, which we interpret and draw conclusions from, is thus robust with respect to whether or not the  $4\phi$  parameters are included into the inversion.

Figure 3.15C presents the F-test results on whether adding  $4\phi$  anisotropy to an isotropic and  $2\phi$  anisotropic inversion improves the fit. Except in the North Kaapvaal region between 33 and 83 s, adding  $4\phi$  anisotropy does not improve the fit significantly.

In the North Kaapvaal Craton (33-83 s periods), the F-test values are greater than the threshold value with a confidence index of 0.05. Figure 3.14 presents the distribution of

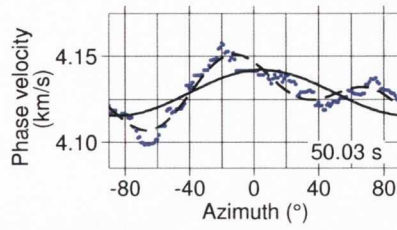


Figure 3.14.: Azimuthal anisotropy at 50.03 s period in the North Kaapvaal Craton region. Blue dots are the data, binned and smoothed with a  $30^\circ$  sliding window. The solid black line is the best-fitting model from an inversion with isotropic and  $2\phi$  anisotropy parameters only. The dashed black line is the best-fitting model from an inversion with isotropic and both  $2\phi$  and  $4\phi$  anisotropy parameters.

phase velocities at a 50.03 s period in the North Kaapvaal Craton region and the best-fitting models given by the inversions with  $2\phi$  anisotropy only and with both  $2\phi$  and  $4\phi$  anisotropy. At this period, the F-test shows that the data requires both  $2\phi$  and  $4\phi$  anisotropy, and the  $90^\circ$  and  $180^\circ$  periodic patterns are clearly seen in the data.

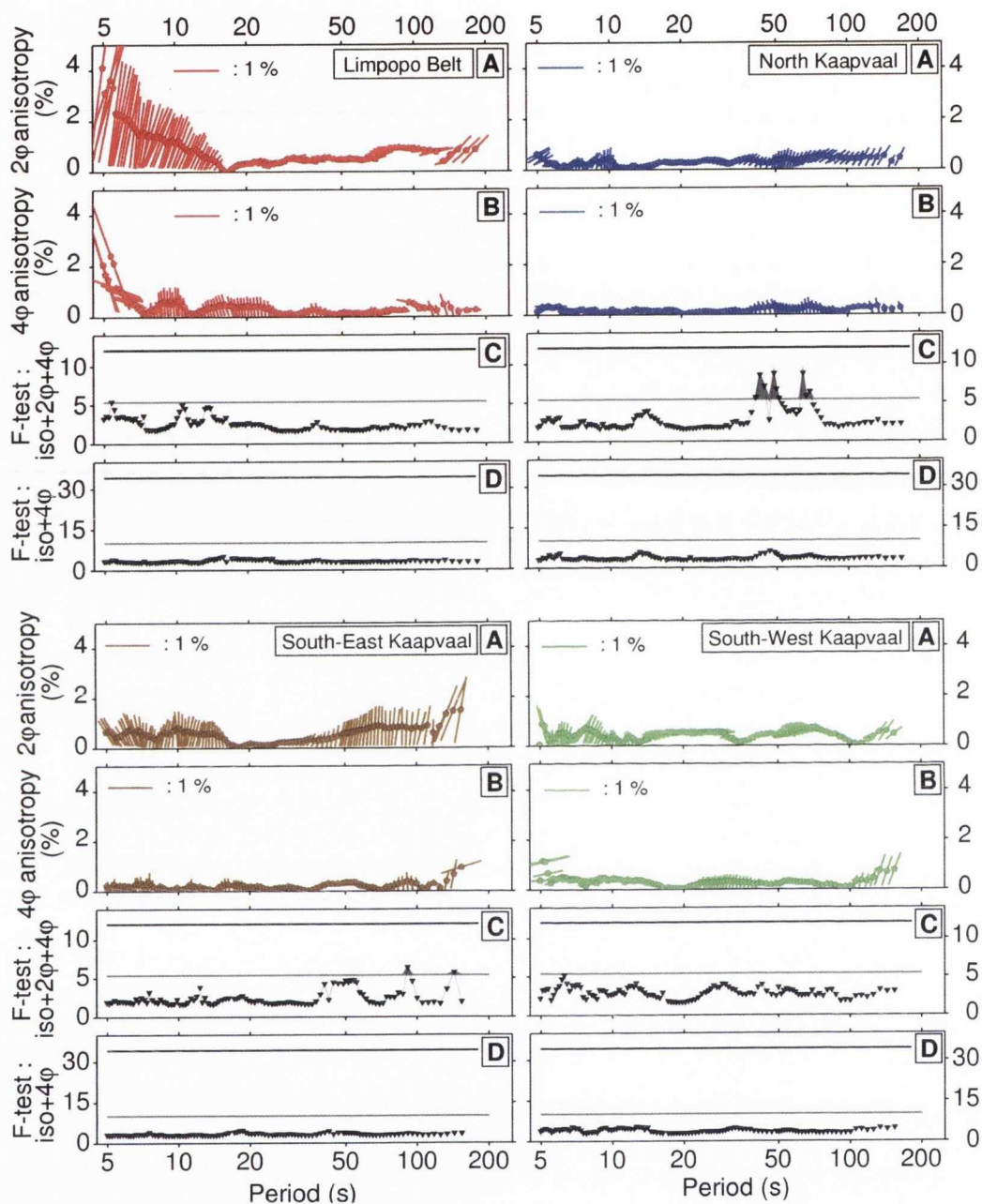


Figure 3.15.: Additional results of the inversions and F-tests for the four sub-regions. A) Azimuths of fast-propagation direction and amplitudes of anisotropy, from inversions for  $2\phi$  anisotropy only. B) Azimuths of fast direction and amplitudes of anisotropy, from inversions for  $4\phi$  anisotropy only (for clarity, only one of the two perpendicular fast directions is shown). C) F-tests on whether an inversion for both  $2\phi$  and  $4\phi$  terms improves the fit, compared to an inversion using only isotropic and  $2\phi$  terms. E) F-tests on whether the addition of the  $4\phi$  terms alone improves the fit, compared to an isotropic inversion. Shaded areas indicate significance, according to the F-test. Grey: above a confidence index of 0.05.



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## AZIMUTHAL ANISOTROPIC SHEAR VELOCITY PROFILES

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In [Chapter 3](#) we show results on azimuthal anisotropy of Rayleigh wave phase velocities, obtained using inter-station measurements. We showed that the distribution of azimuthal anisotropy in southern Africa is complex, with lateral and vertical variations. In order to investigate this further and resolve the depth-distribution of azimuthal anisotropy with more accuracy, we invert Rayleigh isotropic phase velocities and anisotropic parameters to obtain depth-profiles of the vertically polarized shear speed.

In this chapter, we present the inversion technique and investigate the trade-offs between the parameters. We then determine the depth distribution of azimuthal anisotropy in four regions in southern Africa, confirming the results presented in [Chapter 3](#). We resolve azimuthal anisotropy beneath the region with un-precedented accuracy and detect anisotropic boundaries within the mantle-lithosphere.

### 4.1 METHOD

The parameters  $C_0$ ,  $A_1$  and  $A_2$  ([Equation 3.3](#)) are inverted using a non-linear gradient search (Meier *et al.*, 2004; Lebedev *et al.*, 2006; Erduran *et al.*, 2008) to recover depth profiles of the isotropic shear velocity and the  $2\phi$  components of azimuthal anisotropy. The algorithm follows an iterative process, starting with an initial model and improving the parameters to minimize the difference between the model and the data (see [Section 5.3.4](#) for a description on the inversion algorithm). The isotropic starting model is calculated from a modified AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000) in southern Africa ([Figure 5.6](#), [Table 5.2](#)). The depth of the Moho is adjusted in each region with the results presented in [Table 5.4](#). The anisotropic parameters are set to 0. The crust is parameterized with three boxcars and the mantle is parameterized with 12 triangles. The depth of the discontinuities in the crust are also parameters of the inversion. The implementation of the inversion allows the isotropic and anisotropic parameters to be defined with different parameterizations if needed.

Measured and modeled phase velocities are evaluated at azimuths between  $-90^\circ$  and  $75^\circ$  at  $20^\circ$  intervals (*i.e* at  $-90^\circ$ ,  $-70^\circ$ ,  $-50^\circ$ ,  $-30^\circ$ ,  $-10^\circ$ ,  $10^\circ$ ,  $30^\circ$ ,  $50^\circ$  and  $70^\circ$ ). We have

verified that this choice does not influence the results of the inversion by testing different intervals of azimuths (Appendix B) and it resulted that  $20^\circ$  is a good compromise between robustness and efficiency. The vector minimized by the optimization function is defined as the ensemble (*i.e.* for the series of azimuths) of the difference between the modeled and the measured phase velocities, normalized by the standard deviation of the measured phase velocities. Damping coefficients for every parameter are added to the minimization function, as well as smoothing coefficients for the anisotropic parameters.

#### 4.2 RESOLUTION TESTS FOR SHEAR VELOCITY ANISOTROPIC PARAMETERS

The relation between phase velocities and shear velocities is non-linear. Phase velocities can not be explained by an unique shear velocity profile (Figure 2.6). Complex trade-offs between inversion parameters may exist. As shown by Meier *et al.* (2004) and Lebedev *et al.* (2013), the model-space in inversion for isotropic shear-speed profiles is simple, which justifies the use of the gradient-search technique. The gradient-search technique is performed by solving the non-linear problem by adjusting the parameters of the inversion in order to minimize the L2 norm (*i.e.* the sum of the squares of the residuals between the model and the data) and find the parameters that minimize the difference between the model and the data. In the cases presented in this dissertation, 5 iterations are performed to converge to a solution. To invert for the isotropic and anisotropic velocities, we introduce two additional anisotropic parameters. The model-space may, therefore, be more complex. To investigate the complexity of the model-space, we use synthetic dispersion curves and explore the model-space using a grid-search method. By using synthetics, we constrain the quality of the data and focus on the resolvability of anisotropy. In each point of the grid, a gradient-search inversion is performed. The parameters of the corresponding coordinates are fixed and the other parameters are free to vary so as to minimize the deviation between the data and the model. The misfit value is defined as the difference between the model and the data:

$$\text{misfit} = \frac{1}{N} \sqrt{\sum_{i=1}^N \left( \frac{m_i - d_i}{m_i} \right)^2} \quad (4.1)$$

where  $m$  and  $d$  are the ensembles of synthetic and measured phase velocities, respectively, at azimuths between  $-90^\circ$  and  $75^\circ$  at  $20^\circ$  intervals.  $N$  is the number of points and is equal to the product between the number of azimuths (*i.e.* 9) and the number of periods.



#### 4.2.1 Trade-offs between isotropic and anisotropic parameters

We investigate the trade-offs between the isotropic and anisotropic parameters by mapping the two-dimension plane: shear velocity isotropic velocity *vs.* amplitude of shear velocity anisotropy (*i.e.* the norm of the anisotropic parameters) (Figure 4.1). Both the isotropic and anisotropic perturbations in Figure 4.1 are with respect to the reference isotropic shear velocity. The results show that there is no trade-offs between these isotropic and anisotropic parameters. Isotropic structures can not be compensated by azimuthal anisotropy and *vice versa*.

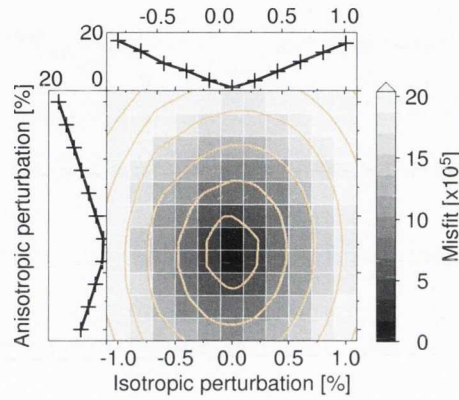


Figure 4.1.: Trade-offs between the shear-wave isotropic velocity and the amplitude of azimuthal anisotropy. Results are shown for synthetic data with AK135 as reference isotropic velocities and 2% anisotropy in the upper-mantle with the azimuth of fast propagation constant during all the test. Isotropic and anisotropic perturbations are for shear velocity in the upper-crust. The grid-spacing of the grid search is represented by the size of the patches. Orange lines show misfit's contours at intervals of  $0.4 \cdot 10^{-5}$  (Equation 4.1). Top (side) frames attached to the main frame represent the minimum misfit values along the y- (x-) direction.

#### 4.2.2 Minimum amplitude of azimuthal anisotropy resolvable

Small amplitudes of anisotropy cannot be accurately resolved. Indeed, as presented in the Figure 4.3, when the amplitude of anisotropy is small, the direction of fast propagation is not resolved. To investigate the threshold amplitude that can be resolved using the inversion technique, we performed synthetic tests and varied the amplitude of anisotropy between 0 and 2% with intervals of 0.1% in the crust and estimated the minimum amplitude at which the azimuth of fast propagation is accurately resolved. In the test presented, the direction of fast propagation is set at  $60^\circ$ . There is no anisotropy in the mantle.

Figure 4.2 presents the azimuths of fast propagation in the upper-crust resulting from inversions for best-fitting model as a function of amplitude of anisotropy for perfect data and for data with noise. We introduced noise to the isotropic-average phase velocities to obtain results that would be comparable with real data. The estimated noise is esti-



imated from real measurements and is described in more detail, as the "complete noise", in Section 5.5 (see also Appendix D). If the data is perfect (*i.e.* no noise), the azimuth of fast propagation is well resolved (Figure 4.2A), whatever the amplitude of anisotropy. However, if the data contains noise, the azimuth of fast propagation can only be resolved accurately if there is at least 0.5% anisotropy (Figure 4.2B). The angle is then accurately resolved with errors of  $\pm 0.5^\circ$ .

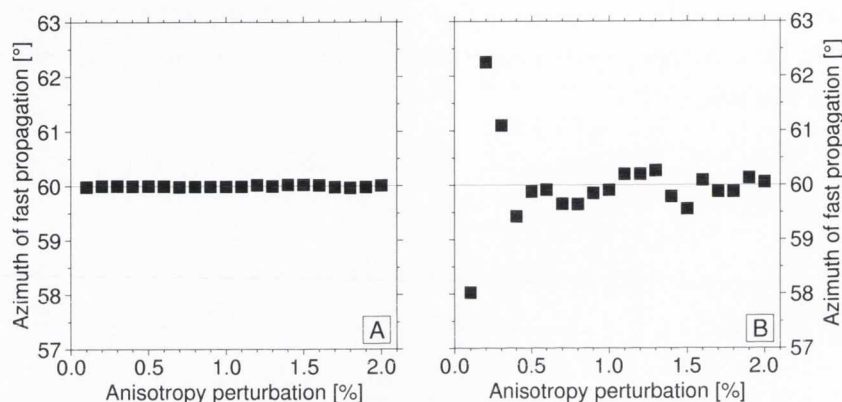


Figure 4.2.: Threshold for minimum amplitude of anisotropy that can be resolved (synthetic tests). Inversion for best-fitting model for azimuthal anisotropy for models with azimuthal anisotropy in the crust. The azimuth of fast propagation is  $60^\circ$  and the amplitude of anisotropy varies and is presented along the x-axis. Results are presented for perfect data (A) and data with noise estimated from real data (B) (see Section 5.5).

#### 4.2.3 Trade-offs between the azimuth of fast propagation and amplitude of anisotropy

We also investigated the trade-offs between the direction of fast propagation and the amplitude of anisotropy to determine whether increases or decreases of amplitude of anisotropy can be compensated by variations of the direction of fast propagation and *vice versa*. We map the two-dimension plane: azimuth of fast propagation *vs.* amplitude of shear velocity anisotropy. Figure 4.3 presents results for synthetic data with AK135 as reference model and 2% anisotropy in the upper-mantle with a fast direction at  $60^\circ$ . The crust is an isotropic medium. Synthetic phase velocities are measured between 4.97 to 250 s.

The model-space seems to be simple and there is only one global minimum. Further investigations of the model-space would then be unnecessary as the best-fitting model resulting from the gradient-search inversion would converge to the solution. There is no trade-offs between the azimuth of fast propagation and the amplitude of anisotropy. Variations in the azimuths of fast propagation can not be compensated by increasing or decreasing amplitude of anisotropy and *vice versa*.

Relatively small misfits are found in an ellipse under the bell-shaped curve. The major axis of the ellipse follows the y-direction (along which the azimuth is constant) and the

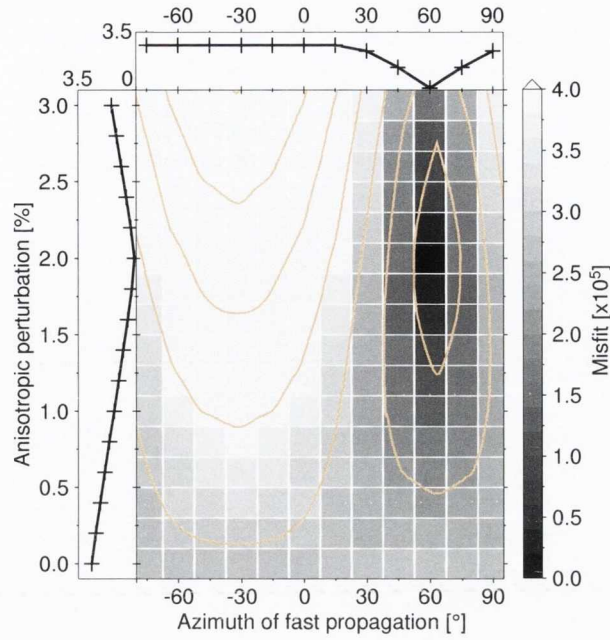


Figure 4.3.: Trade-offs between the shear velocity amplitude of azimuthal anisotropy and azimuth of fast propagation. Results are shown for synthetic data with AK135 (Kennett *et al.*, 1995) as reference isotropic model and 2% anisotropy in upper-mantle with fast direction at  $60^\circ$ . In this grid-search, the anisotropic parameters are fixed in the upper-mantle and free to vary in the other layers. Isotropic parameters are free to vary along every depths. The spacing of the grid-search is represented by the size of the patches. Orange lines show misfit's contours at intervals of  $1.10^{-5}$ . Top (side) frames attached to the main frame represent the minimum misfit values along the y- (x-) direction.

minor axis follows the x-direction (along which the amplitude is constant). The azimuth of fast propagation is better constrained than the amplitude of anisotropy. In this synthetic test (where no noise has been added to the data), the azimuth of fast propagation is perfectly recovered at  $60^\circ$  while the amplitude of anisotropy is not resolved as accurately.

We explored the model-space using real data in the four sub-regions in southern Africa. Figure 4.4, Figure 4.5, Figure 4.6, Figure 4.7 and Figure 4.8 show the two-dimension planes: azimuth of fast propagation *vs.* amplitude of shear velocity anisotropy in six or seven cross-sections corresponding to different depth ranges. The cross-sections represent the model-space for the upper-crust, middle-crust, lower-crust, upper mantle lithosphere, lower mantle-lithosphere and asthenosphere (the asthenosphere may be divided into two layers). The inversions are parameterized with eight or nine boxcars.

The results confirm the observations from synthetic tests. The relatively small misfits are found in an ellipse that is centered under a bell-shaped contour. There is no trade-offs between the azimuth of fast propagation and the amplitude of anisotropy.

If the amplitude of anisotropy is greater than 0.5%, as inferred from the inversion results, then the azimuth is better resolved than the amplitude of anisotropy. In contrast, as in

the case of the upper-lithosphere of the Limpopo Belt (Figure 4.4D) or the upper-crust of the North Kaapvaal Craton (Figure 4.5A), where the amplitude of anisotropy seems to be smaller than 0.5%, the direction of fast propagation is not accurately resolved. This confirms the study in Section 4.2.2 showing that the anisotropy is resolved most accurately if its amplitude is greater than 0.5%.



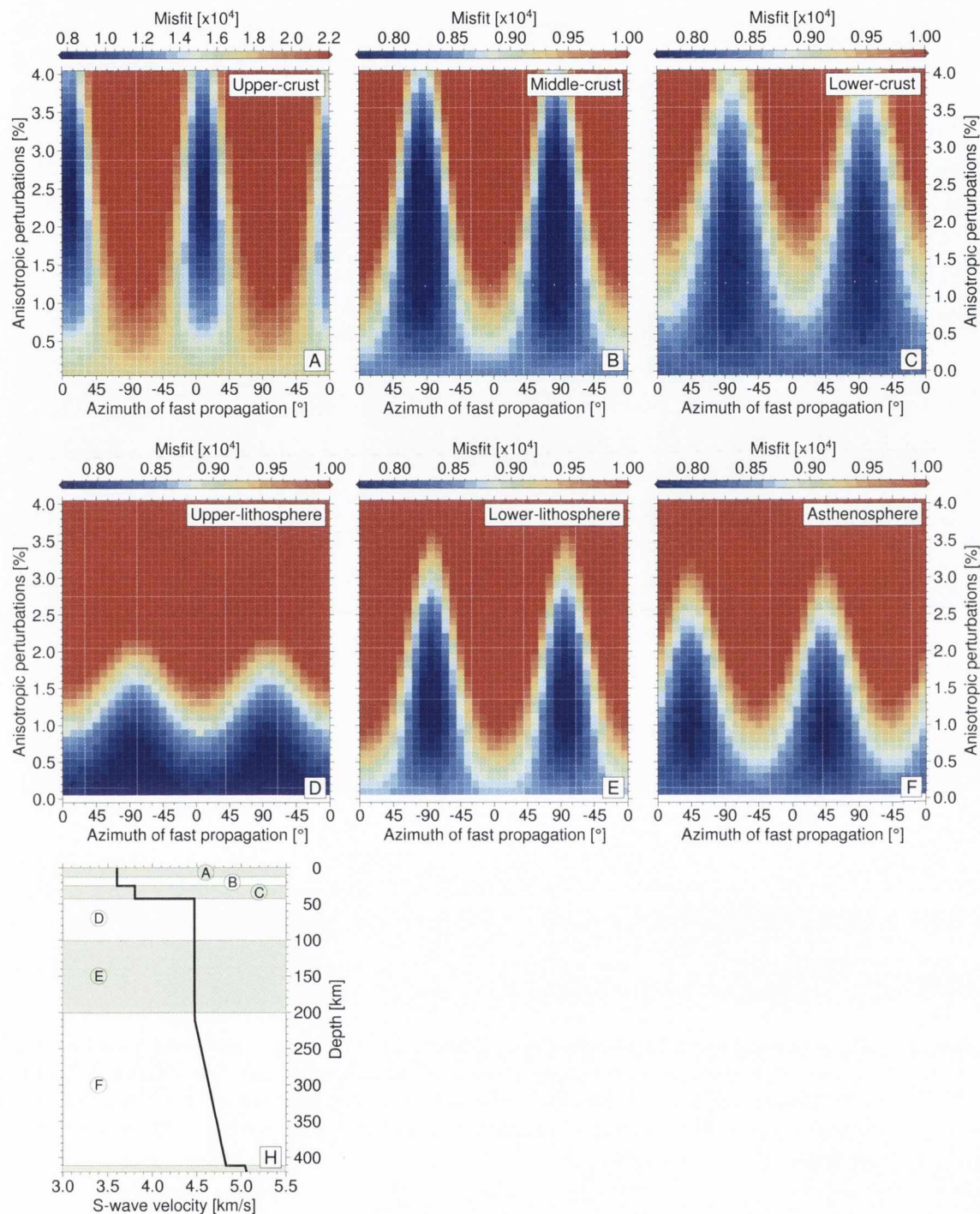


Figure 4.4.: Trade-offs between the shear velocity amplitude of azimuthal anisotropy and azimuth of fast propagation in the upper-crust (A), middle-crust (B), lower-crust (C), upper lithosphere (D), lower lithosphere (E), lithosphere-asthenosphere depth range (F) and asthenosphere (G) of the Limpopo Belt region. The layering is presented in frame G, with AK135 (Kennett *et al.*, 1995) as shear velocity reference.

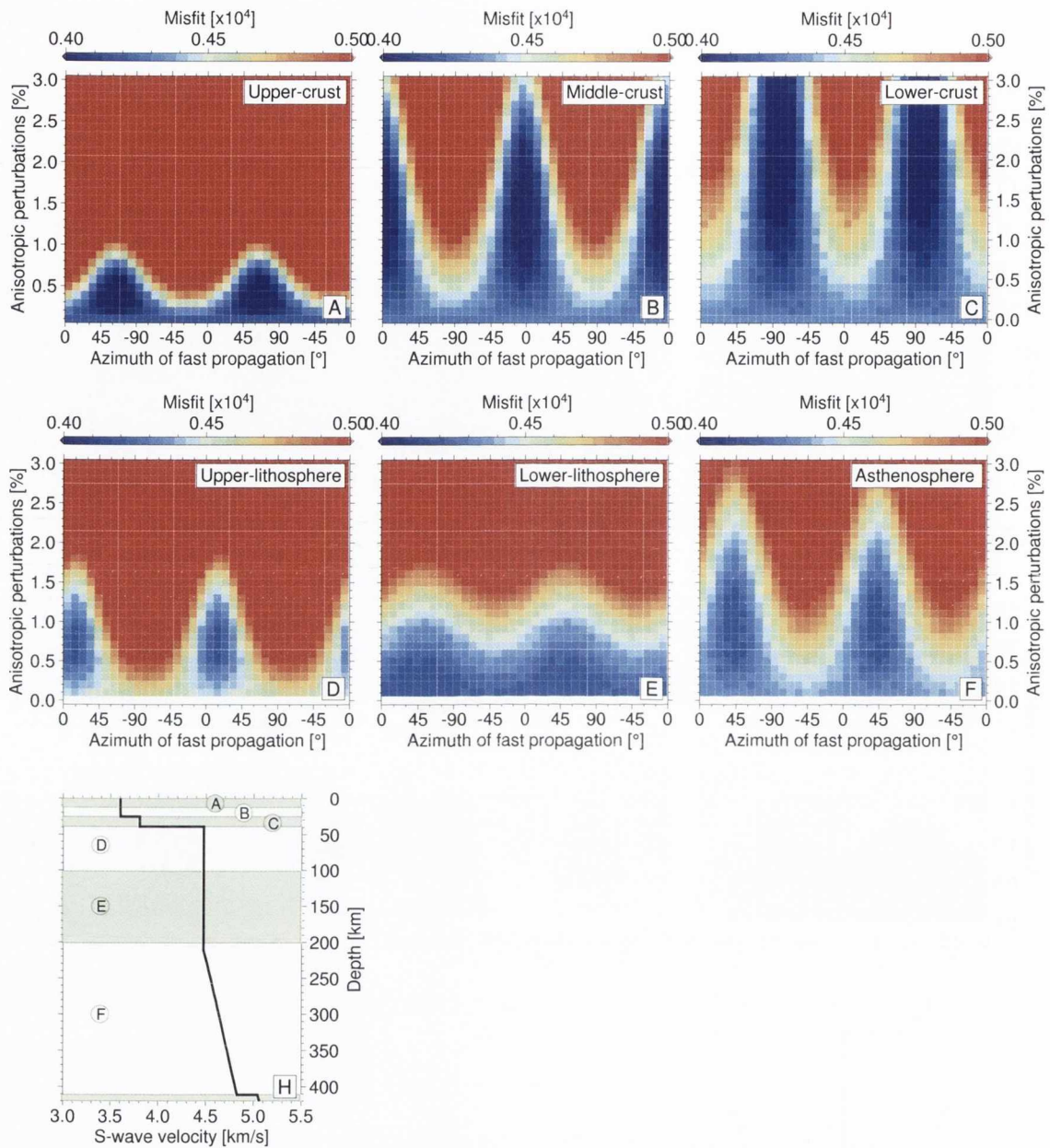


Figure 4.5.: Trade-offs between the shear velocity amplitude of azimuthal anisotropy and azimuth of fast propagation in the upper-crust (A), middle-crust (B), lower-crust (C), upper lithosphere (D), lower lithosphere (E) and asthenosphere (F) of the North Kaapvaal Craton region. The layering is presented in frame G, with AK135 (Kennett *et al.*, 1995) as shear velocity reference.



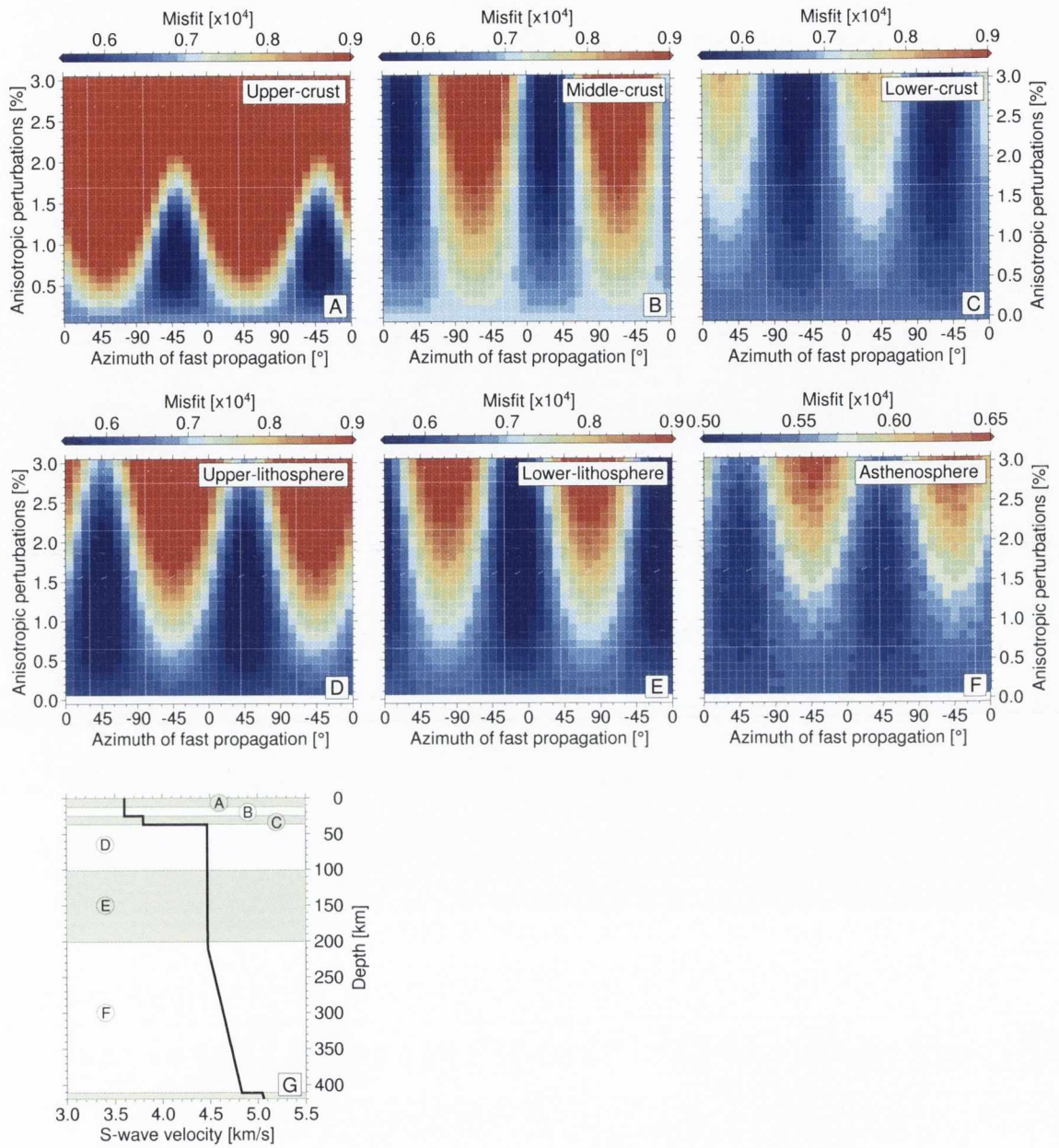


Figure 4.6.: Trade-offs between the shear velocity amplitude of azimuthal anisotropy and azimuth of fast propagation in the upper-crust (A), middle-crust (B), lower-crust (C), upper lithosphere (D), lower lithosphere (E) and lithosphere-asthenosphere depth-range (F) of the South-East Kaapvaal Craton region. The layering is presented in frame G, with AK135 (Kennett *et al.*, 1995) as shear velocity reference.



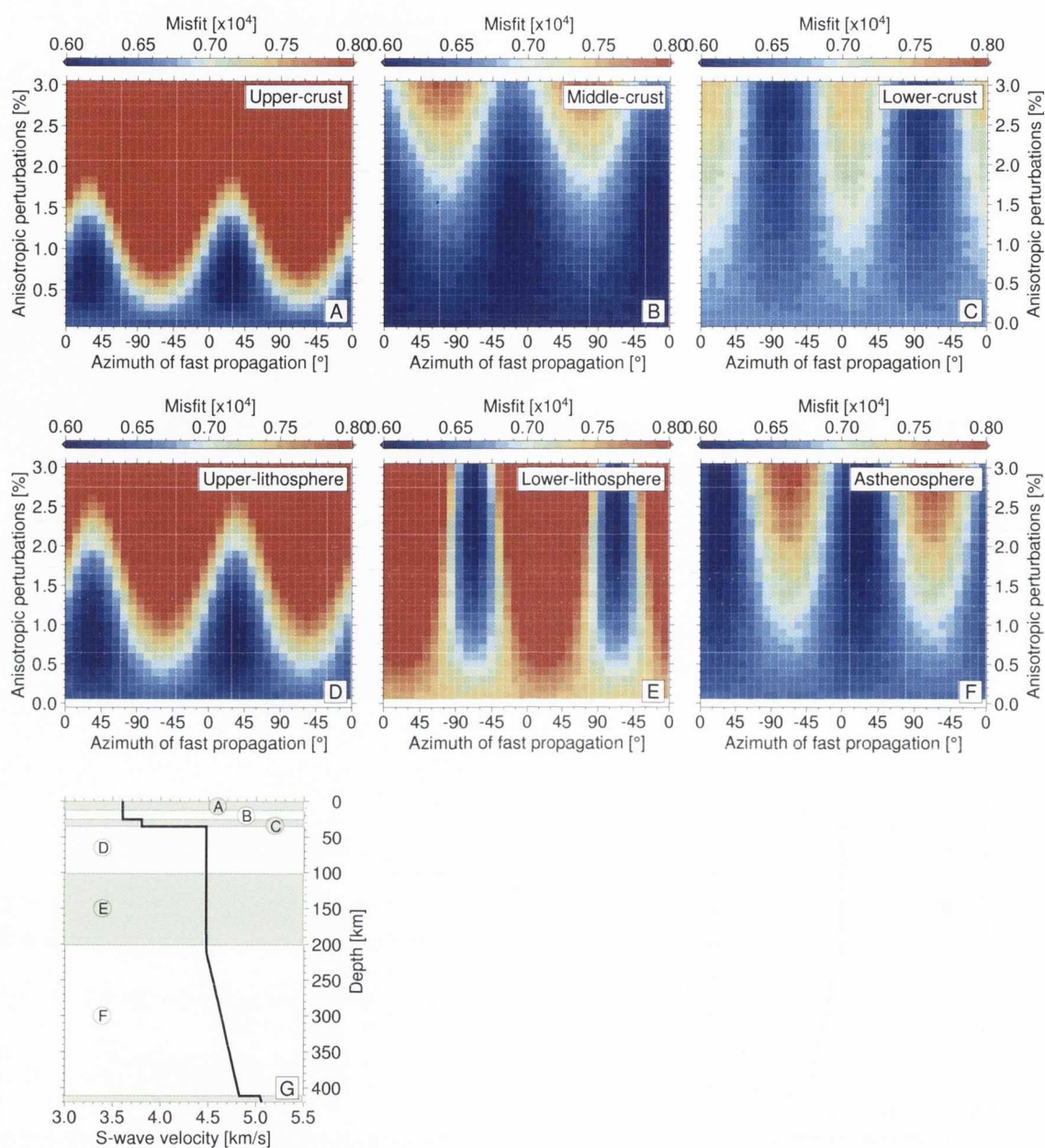


Figure 4.7.: Trade-offs between the shear velocity amplitude of azimuthal anisotropy and azimuth of fast propagation in the upper-crust (A), middle-crust (B), lower-crust (C), upper lithosphere (D), lower lithosphere (E) and asthenosphere (F) of the South-West Kaapvaal Craton region. The layering is presented in frame G, with AK135 (Kennett *et al.*, 1995) as shear velocity reference.

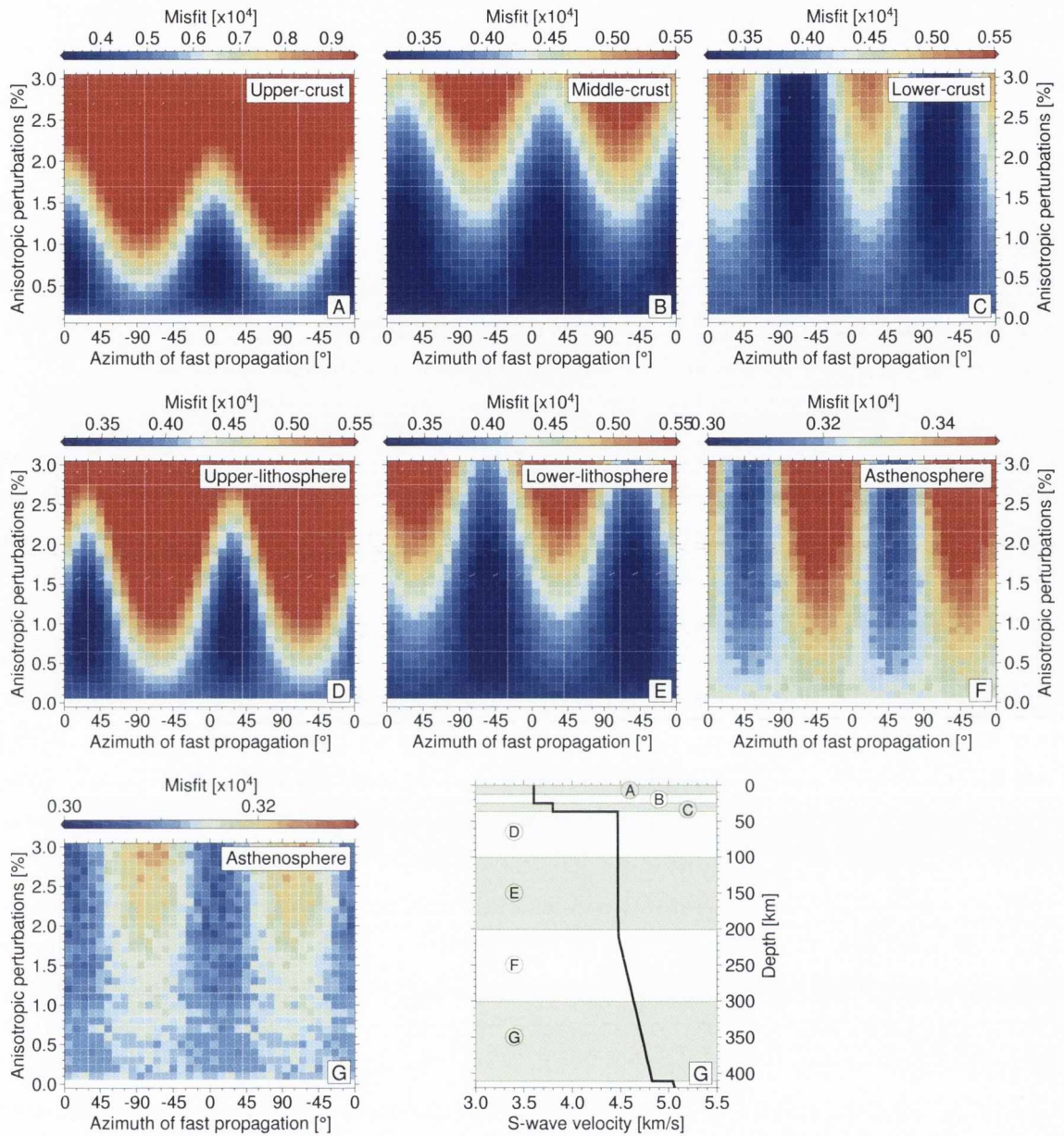


Figure 4.8.: Trade-offs between the shear velocity amplitude of azimuthal anisotropy and azimuth of fast propagation in the upper-crust (A), middle-crust (B), lower-crust (C), upper lithosphere (D), lower lithosphere (E) and asthenosphere (F) of southern Africa. The layering is presented in frame H, with AK135 (Kennett *et al.*, 1995) as shear velocity reference.



### 4.3 SHEAR-VELOCITY PROFILES IN SOUTHERN AFRICA

#### 4.3.1 Introduction

The imaging of the shear velocity of seismic waves enables us to explore the interior of the Earth where the human hand can not reach. Shear velocity anomalies can be interpreted with composition, pressure or temperature changes and contribute important evidence for a better understanding of the dynamics of the crust and the mantle or on the relation between the deep lithosphere and the topography at the surface of the Earth.

Southern Africa is an area of particular interest as it comprises ancient cratons that have remained stable for billions of years. It is therefore important to understand what differentiates the roots of these cratons from the lithosphere of other, younger continents. Several studies imaged the seismic structures of the crust and upper-mantle of the Kaapvaal Craton using different techniques: Rayleigh wave tomography (Li and Burke, 2006; Chevrot and Zhao, 2007; Adams and Nyblade, 2011; Li, 2011), inversion of inter-station surface wave phase velocity measurements (Priestley, 1999; Larson *et al.*, 2006), body wave delay time tomography (James *et al.*, 2001), body wave receiver function (Vinnik *et al.*, 2009), joint inversion of Rayleigh wave phase velocity and ambient noise (Yang *et al.*, 2008) and joint analysis of Rayleigh wave and body wave tomography (Priestley and Tilmann, 2009). The list is not exhaustive but shows the extensive interest in this region.

All studies agreed that the cratonic root extends to at least 180 km and is characterized with high velocities. Geochemical studies of xenoliths (showing the spatial and temporal variations in composition and the thermal state of the lithosphere) from kimberlite eruptions in the South-West part of the Kaapvaal Craton showed that the root of the craton is depleted in iron (Griffin *et al.*, 2003; Bell *et al.*, 2003). The iron-depletion in a cold and thick root would confirm that the craton is mechanically strong and neutrally buoyant (Jordan, 1975), explaining its stability. It is however important to have good constraints on the depth to which the root of the craton extends to confirm that results from these xenoliths studies are indeed related to the root of the cratonic lithosphere (Priestley, 1999).

While all studies agreed that the root of the craton is thick and characterized with fast shear velocities, shear velocity structures below it remains unclear. Qiu *et al.* (1996); Priestley (1999); Li and Burke (2006); Vinnik *et al.* (2009); Li (2011); Adams and Nyblade (2011) show relatively slow shear velocities in a low velocity zone (LVZ) below ~160 km. The presence of such a LVZ has been observed in several cratons around the world [Figure 4.9](#). James *et al.* (2001); Saltzer (2002); Larson *et al.* (2006); Wittlinger and Farra (2007) do not however find any evidence for a LVZ. To understand why studies do not agree on the observation of a LVZ, Li and Burke (2006) suggested that the LVZ is an artifact from radial anisotropy. However, they stated that radial anisotropy by itself is small at the depth of



the LVZ (Freybourger *et al.*, 2001; Saltzer, 2002) and would not be sufficient to account for an apparent low velocity zone. Nevertheless, one should be careful by not interpreting a low velocity zone in  $V_S$  while it is  $V_{SV}$ . Indeed, most published results show shear velocity profiles inferred from Rayleigh wave phase velocity. Here, we also examine  $V_{SV}$  and not  $V_S$ .

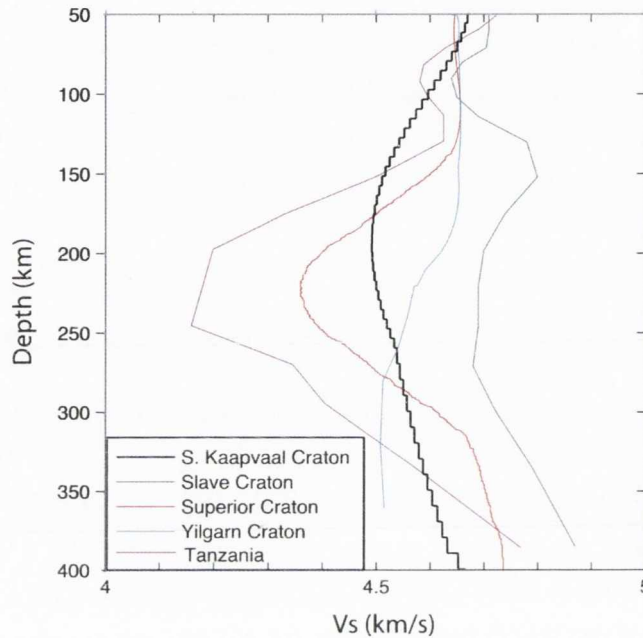


Figure 4.9.: Velocity models in several cratons. Models are shown for the Slave Craton (Chen *et al.*, 2007), the central Superior Craton (Darbyshire *et al.*, 2007), the Tanzania Craton (Weeraratne *et al.*, 2003), an average of models of the Yilgarn Craton from Pedersen *et al.* (2009), and the average velocity model for the southern Kaapvaal Craton. Figure from Adams and Nyblade (2011)

We present in this section the depth distribution of the vertically polarised shear velocity ( $V_{SV}$ ) of three regions in the Kaapvaal Craton and in the adjacent Limpopo Belt, determined by inversion of fundamental mode Rayleigh wave phase velocities. Phase velocities are resulting from inversions for azimuthal anisotropy (Chapter 3), therefore, the results are not biased by azimuthally anisotropic shear velocity structures. The azimuthal distribution of the station-pairs enables us to resolve both azimuthal anisotropy and isotropic phase velocity up to  $\sim 160$  s and at this period, Rayleigh waves are sensitive down to  $\sim 350$  km, we shall plot our results down to this depth.

#### 4.3.2 Method

We invert the isotropic-average Rayleigh wave dispersion curve of each of the four regions using a non-linear gradient search method (Meier *et al.*, 2004; Lebedev *et al.*, 2006; Erduran *et al.*, 2008) (see Section 5.3.4 for a description on the inversion algorithm) for SV velocity

depth profiles. The reference model combined a modified AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000). The depth of the Moho is adjusted with results from Chapter 5.

As the solution of the inversion is non-unique, there is an infinite number of models that would fit to the data. We therefore do not present one best-fitting model for each region, but a range of smooth shear velocity profiles. Each velocity range has been determined after ~65 models resulting from independent inversions. The set of models defining the range have been selected so that the misfit value (that informs about the quality of the misfit ; see Equation 5.2) is no larger than a threshold misfit. This threshold is defined to be the maximum misfit of a set (5-6) of models that would best describe the data.

#### 4.3.3 Results

Figure 4.10 presents the range of the vertically polarized shear speed for each of the four regions. Fits to the data and perturbations between the models and the data are also represented in this figure to illustrate the quality of the fits. The gray bands are determined so that they include the best fitting smooth models. The results show that the four regions are characterized by a high-velocity upper mantle-lithosphere with a maximum velocity of 4.8 km/s at a depth of around 150 km.

A low velocity zone is not required by the data in the Limpopo Belt region and is preferred by the data in the North Kaapvaal Craton. A low velocity zone is required and is stronger in the South Kaapvaal Craton.

#### 4.3.4 Discussion

Our results overall agree with Priestley (1999); Freybourger *et al.* (2001); Li and Burke (2006); Yang *et al.* (2008); Vinnik *et al.* (2009); Li (2011) and Adams and Nyblade (2011) (Figure 4.11) *i.e.* shear velocity profiles show a low-velocity zone with velocity reduction of few percents between 150 and 250 km, below a fast lithospheric mantle.

According to Li and Burke (2006), the low velocity zone is explained by an increase of temperature. They suggested that other candidates for the reduction of velocity would not account for a reduction of a few percent. They argued that petrologic differences observed in xenoliths are insufficient to explain the strong reduction of velocities (James *et al.*, 2004) and that there is no observation of volatile content from xenolith data that would explain a LVZ. A good candidate that would be responsible for a low velocity zone is an increase of iron content at the depth of the LVZ. However, Lee (2003) pointed out that a content increase of iron, calcium and aluminum would only be responsible for a reduction of shear-velocity of up to ~1.5% which do not explain the reduction of

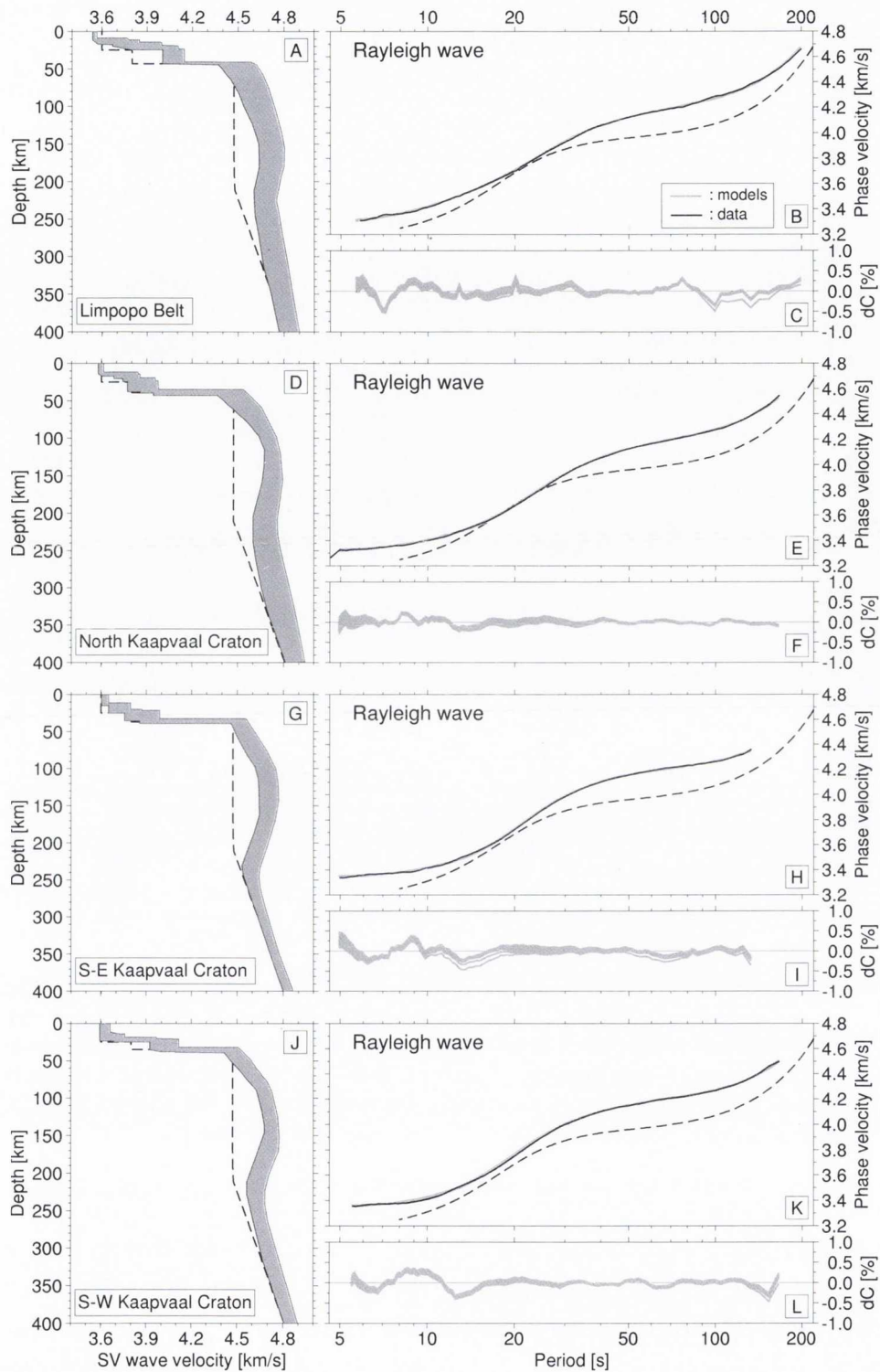


Figure 4.10.: 1-D Shear velocity (vertical component) depth profiles (A,D,G,J) inferred for each of the four sub-region in southern Africa from inversion of Rayleigh wave dispersion curves (B,E,H,K). Dashed lines are reference model AK135 (Kennett *et al.*, 1995), plotted for reference. C,F,I,L: Perturbations of the models (used to generate the shear velocity band) regarding to the data to highlight the quality of the fit.



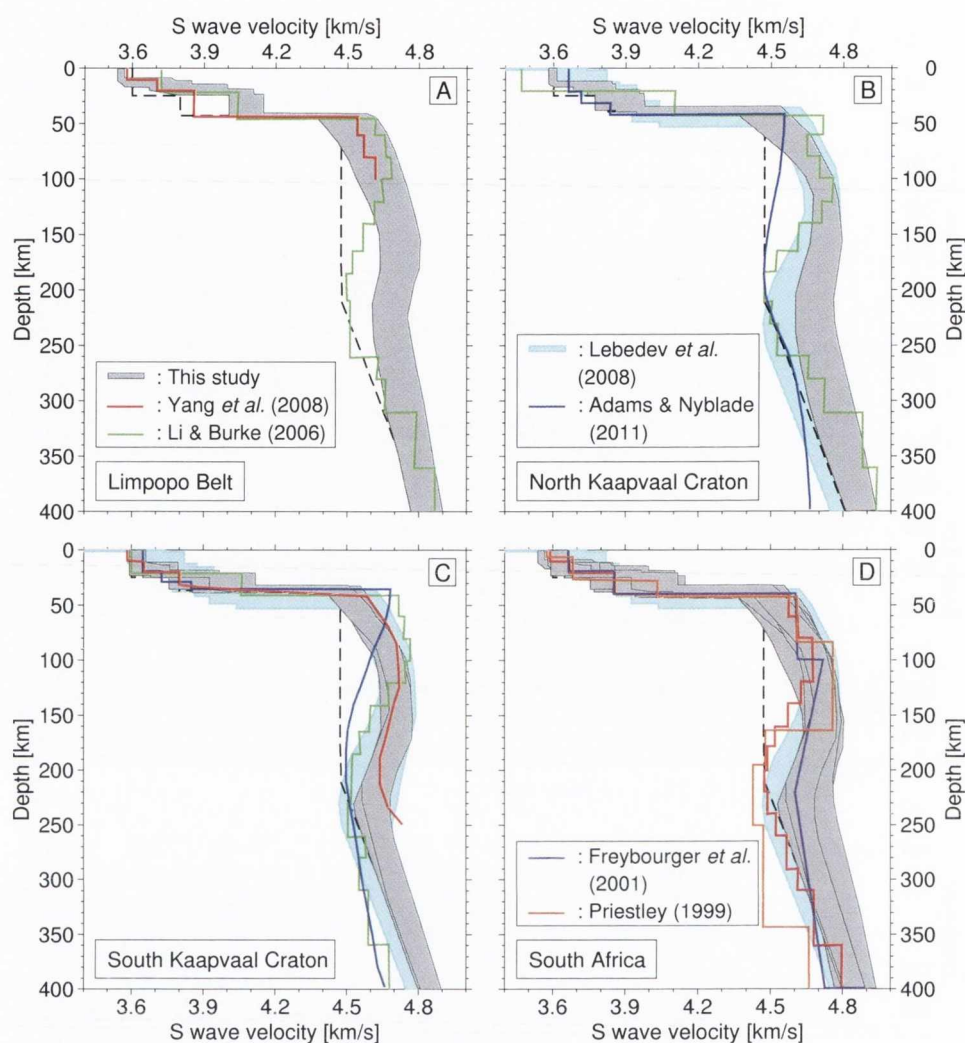


Figure 4.11.: Comparison between shear velocity profiles from this study (gray band) and Yang *et al.* (2008) (red), Li and Burke (2006) (green), Lebedev *et al.* (2008) (blue band), Adams and Nyblade (2011) (dark blue), Freybourger *et al.* (2001) (purple) and Priestley (1999) (orange) in different regions: Limpopo Belt (A), North Kaapvaal Craton (B), South Kaapvaal Craton (C) and average in South Africa (D). The dashed black line represents the reference AK135 (Kennett *et al.*, 1995) for reference.

velocity of few percent as observed. Li and Burke (2006) concluded that an increase of the temperature of  $\sim 220^\circ\text{C}$  would be responsible for the low velocity zone. Relatively higher temperatures in the lithosphere-asthenosphere boundary range, between 150 and 250 km, would increase the buoyancy of the lower-lithosphere and generate high relief. Our results, may thus offer an explanation for the difference of topography between the neighboring central (warmer sub-lithospheric mantle, high topography) and northern Kaapvaal Craton and Limpopo Belt (colder sub-lithospheric mantle, lower topography).

Lithgow-Bertelloni and Silver (1998) and Gurnis *et al.* (2000) proposed that a large-scale upwelling from the core-mantle boundary would be responsible for the high relief in the

entire southern Africa. Our results offer an explanation for the variations in the elevation at a smaller scale, within southern Africa.

#### 4.4 SHEAR-VELOCITY AZIMUTHAL ANISOTROPY IN SOUTHERN AFRICA

##### 4.4.1 Introduction

To further investigate the structures of the lithosphere and upper-asthenosphere of southern Africa, we now focus on the depth distribution of azimuthal anisotropy. Azimuthal variations of the velocity of propagation of a seismic wave informs about the deformation of the medium since its formation.

The origin of the azimuthal anisotropy that generates SKS splitting in southern Africa is debated. One end-member theory (Vinnik *et al.*, 1995) suggested that the splitting occurs in the asthenosphere and is due to flow caused by the current plate motion. Similar interpretations have been proposed in other cratonic regions and tend to validate this theory (Deschamps *et al.*, 2008b; Yuan and Romanowicz, 2010; Debayle and Kennett, 2000). However, Silver *et al.* (2001) and Silver *et al.* (2004) argued that the SKS splitting in southern Africa occurs in the mantle-lithosphere and is related to the deformation during the Archean collisions between what are now part of the Kaapvaal Craton, the Limpopo Belt, and the Zimbabwe Craton. Accurate depth distribution of azimuthal anisotropy would help discriminate between the two end-member theories and also get more detailed evidence on the early dynamics of the cratons.

Receiver function studies (Hansen *et al.*, 2009; Wittlinger and Farra, 2007; Savage and Silver, 2008; Rychert and Shearer, 2009) have found P-to-S and S-to-P conversions at ~100 km beneath southern Africa suggesting shallower than expected lithosphere-asthenosphere boundary (Fishwick, 2010). A converter at a similar depth has been observed under the North American Craton, suggesting, again, a lithosphere-asthenosphere boundary at a shallower depth than expected. Yuan and Romanowicz (2010) argued that this discontinuity would in fact not be related to the LAB, but to a boundary between two layers within the lithosphere, each with a distinct anisotropic signature. In this section, we examine the azimuthal anisotropy in southern Africa and determine whether a boundary at ~100 km between two anisotropic layers could be the phase converter imaged by receiver function studies.

#### 4.4.2 Method

To invert phase velocities and azimuthal anisotropy parameters, we use a similar approach to [Section 4.3.2](#). The reference model for shear velocity is a modified AK135 ([Kennett \*et al.\*, 1995](#)) and crust 2.0 ([Bassin \*et al.\*, 2000](#)). The depth of the Moho is adjusted with results from [Chapter 5](#). The reference model for anisotropy is set to zero at every depth. The anisotropic data are resulting from an inversion which may have introduced errors. Because of this possible noise in the data, we added smoothing parameters to the misfit function.

The solutions are non-unique and an infinite number of models can fit the data equally well. We are therefore searching for a set of models that would fit the data rather than only one model. We follow the same procedure as in [Section 4.3.2](#) and in a first step, search for few best-fitting models and in a second step, search for models that differ from the reference but still fit the data equally well. To select the best-fitting models, we parametrize the inversions with a large number of basis-functions and decrease the smoothing. We therefore have a relatively good estimation of the distribution of anisotropy with depth. However, the data are noisy and we do not expect to fit small scale variations of anisotropy along depth but are seeking for smooth transition of anisotropy. By increasing the smoothing coefficients, we reduce the chance of fitting noise. We perform hundreds of inversions with varying damping and smoothing coefficients to obtain a range of profiles with misfit values no larger than the misfit of the best-fitting profile. We thus obtain a set of smooth profiles that fit the data.

We obtained anisotropic parameters in a broad band of period (up to 200 s) however, the azimuthal coverage of the station-pairs significantly decreased at periods greater than 130 s. Regarding to Rayleigh wave sensitivity to shear velocity ([Figure 2.6](#)), one should be careful while interpreting azimuthal anisotropy at depths greater than 300-350 km.

#### 4.4.3 Results

[Figure 4.12](#), [Figure 4.13](#), [Figure 4.14](#) and [Figure 4.15](#) show 1-D depth profiles of the vertical component of the shear velocity, the azimuth of fast propagation and the amplitude of azimuthal anisotropy. Fits to the data and misfit perturbations ([Equation 5.3](#)) along period are also presented to display the quality of the fits. The gray bands encompass all the best fitting models.



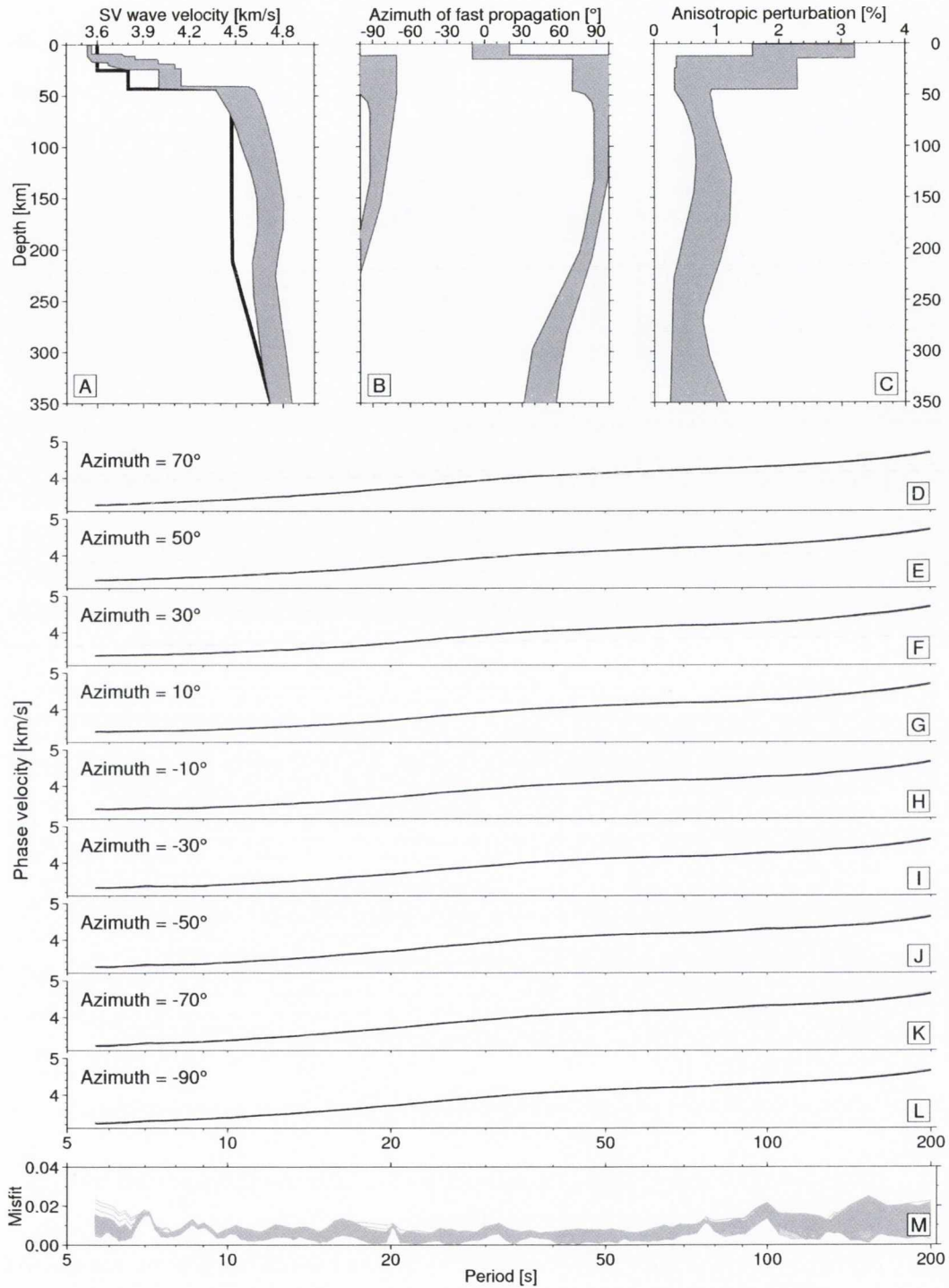


Figure 4.12.: Shear velocity depth-profile ranges (A) and azimuthal anisotropy ranges (B,C) in Limpopo Belt. Thick black line in frame A is modified AK135 (Kennett *et al.*, 1995) for reference. D-L: measures (black) and modelled (gray) dispersion curves. The difference between the curves are smaller than the line thickness. M: Deviation of the models to the data.

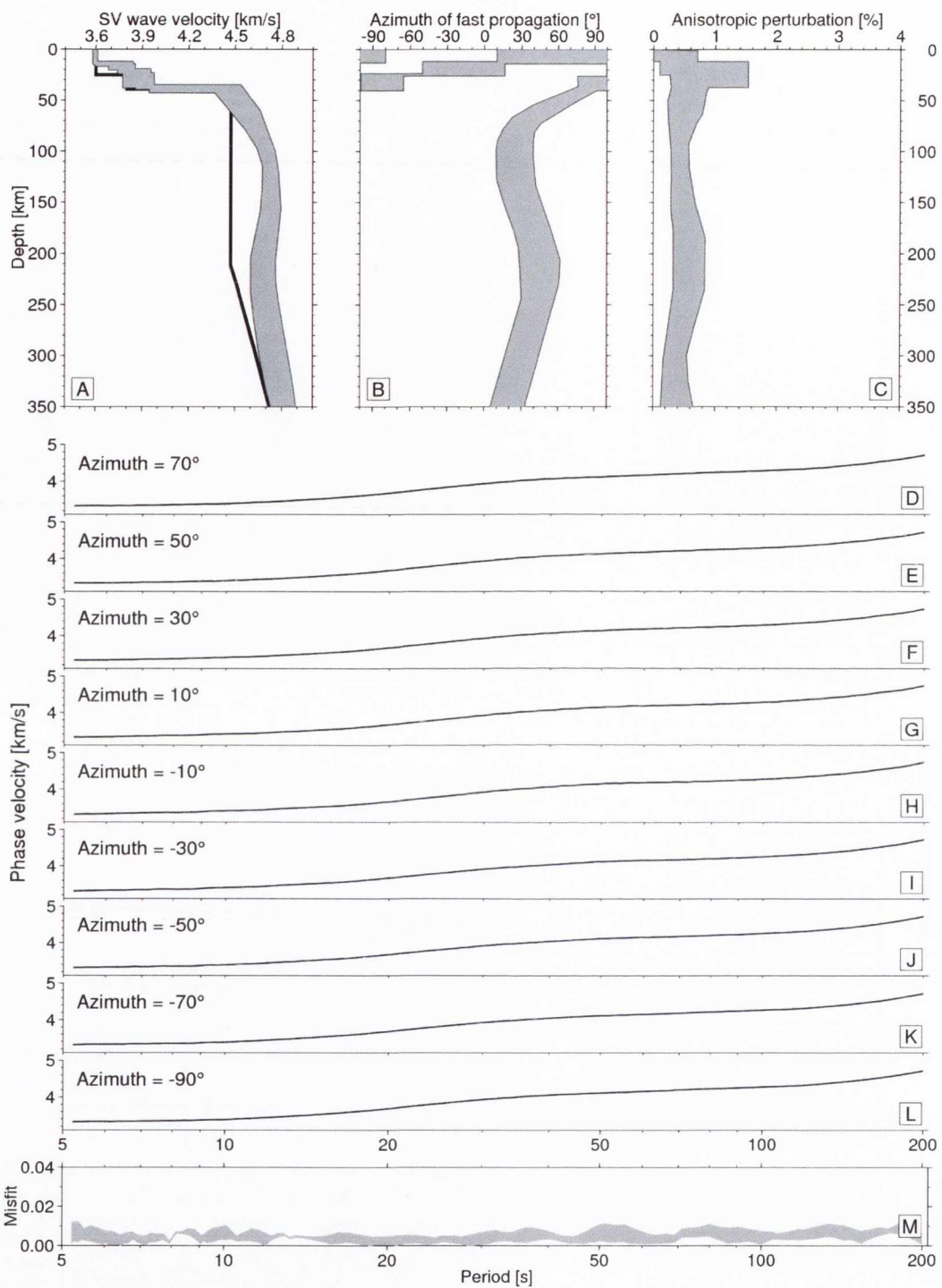


Figure 4.13.: Shear velocity depth-profile ranges (A) and azimuthal anisotropy ranges (B,C) in North Kaapvaal Craton. Thick black line in frame A is modified AK135 (Kennett *et al.*, 1995) for reference. D-L: measures (black) and modelled (gray) dispersion curves. The difference between the curves are smaller than the line thickness. M: Deviation of the models to the data.

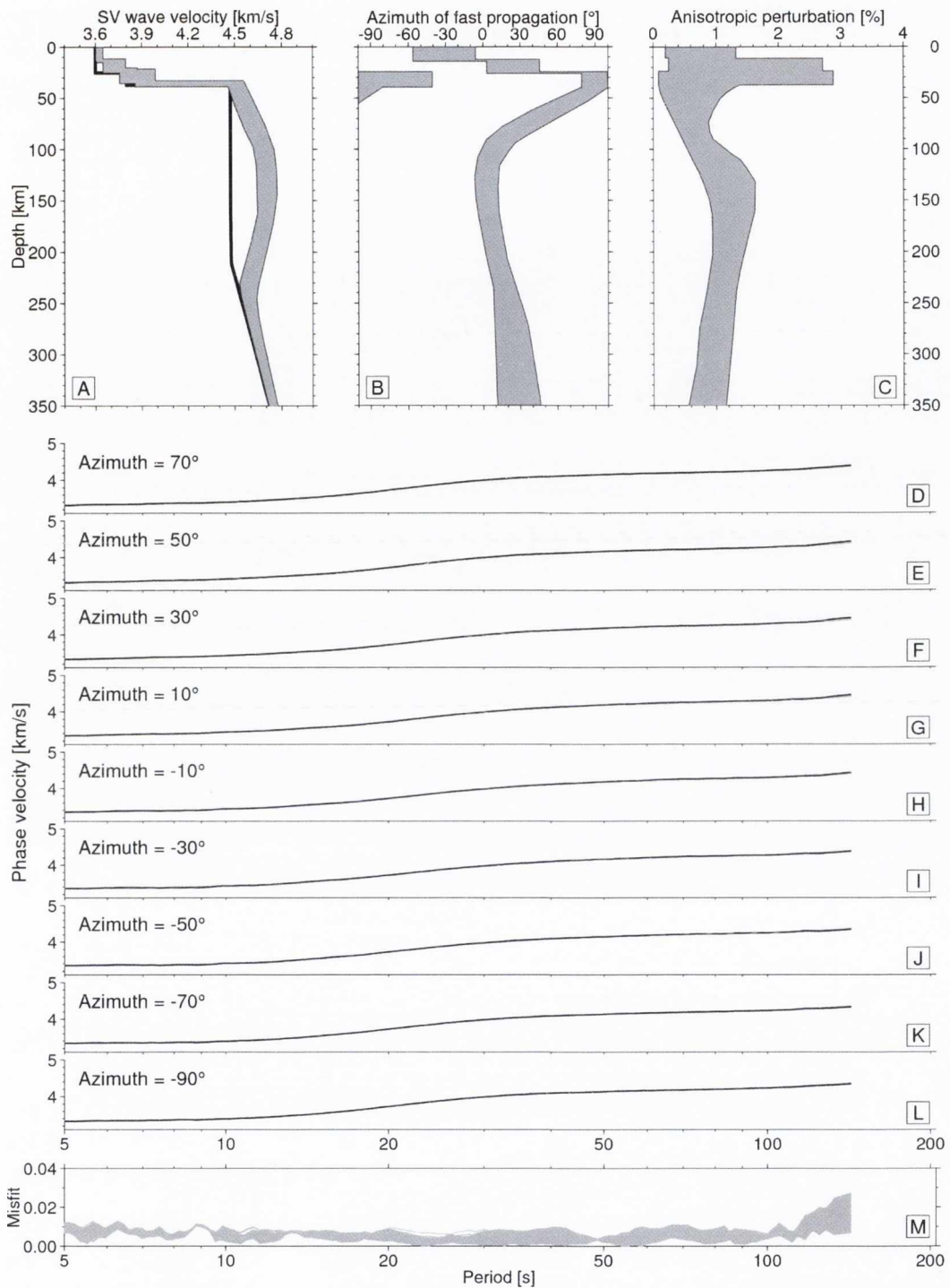


Figure 4.14.: Shear velocity depth-profile ranges (A) and azimuthal anisotropy ranges (B,C) in central Kaapvaal Craton. Thick black line in frame A is modified AK135 (Kennett *et al.*, 1995) for reference. D-L: measures (black) and modelled (gray) dispersion curves. The difference between the curves are smaller than the line thickness. M: Deviation of the models to the data.



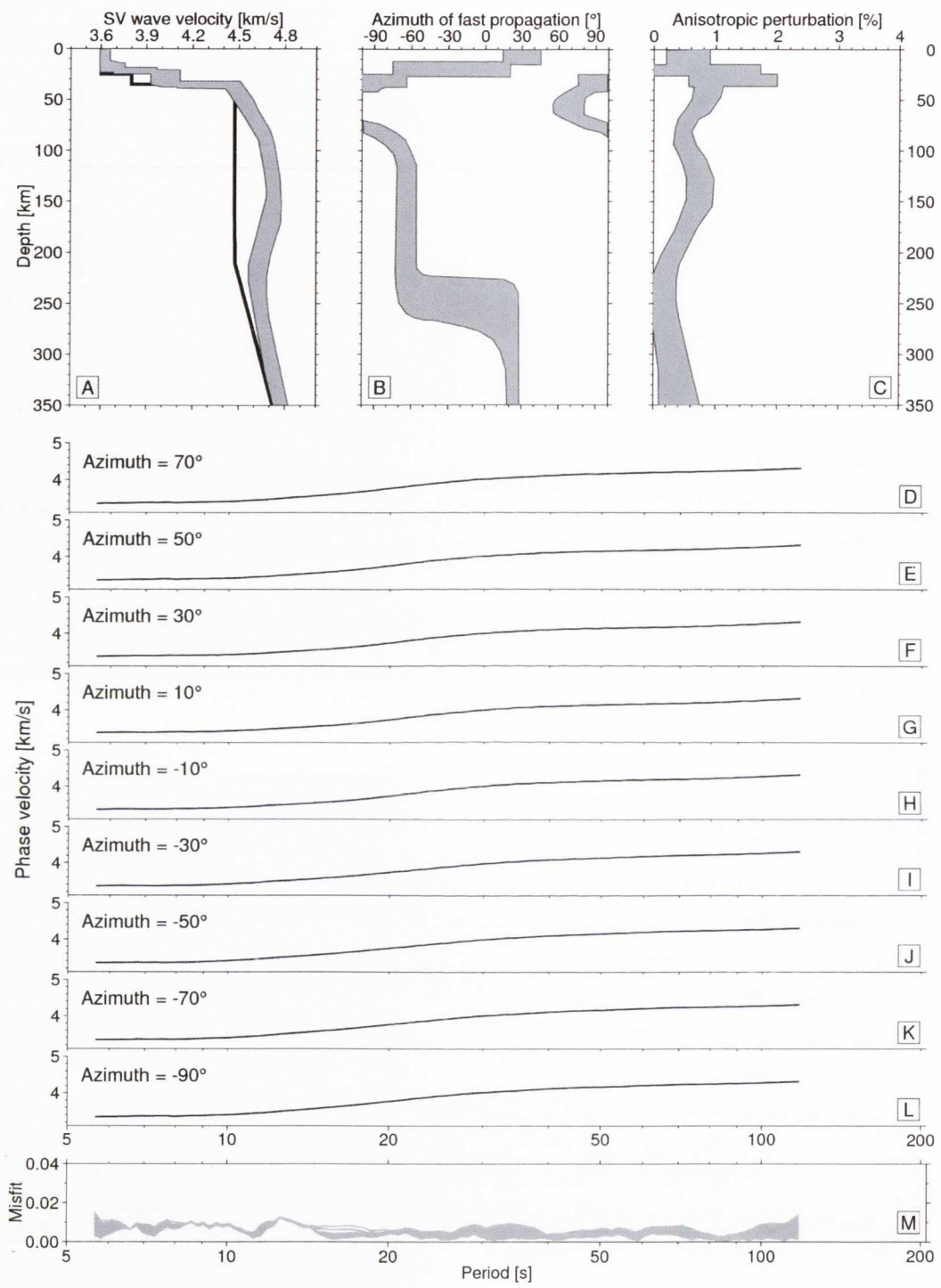


Figure 4.15.: Shear velocity depth-profile ranges (A) and azimuthal anisotropy ranges (B,C) in South-West Kaapvaal Craton. Thick black line in frame A is modified AK135 (Kennett *et al.*, 1995) for reference. D-L: measures (black) and modelled (gray) dispersion curves. The difference between the curves are smaller than the line thickness. M: Deviation of the models to the data.

The results are consistent with the results from the model-space exploration. They provide similar results in their azimuth of fast propagation and amplitude of anisotropy despite the dissimilar, simpler inversion set-up. We are therefore confident with our results and the models presented in this chapter. This observation is expected since, as seen in Section 4.2.3, the model-space is simple and a gradient-search should converge to a best-fitting model. Secondly, the best-fitting models confirm that the azimuth of anisotropy is better resolved than the amplitude.

We observe pronounced transition of fast directions in the three regions of the Kaapvaal Craton between the depth of the Moho and ~100-150 km. An abrupt change of direction is required by the data in the South-West region of the Kaapvaal Craton as smoother profiles would deteriorate the quality of the fit to the data. Figure 4.16 shows a comparison between the best fitting models and a model with smoother transition of fast directions between 40 and 90 km that do not fit the data equally well.

In the North and central Kaapvaal, the data can be explained equally well with a sudden change of direction. Figure 4.17 shows a comparison between the best fitting models and a model with sharp transition of azimuth of fast propagation that fits the data.

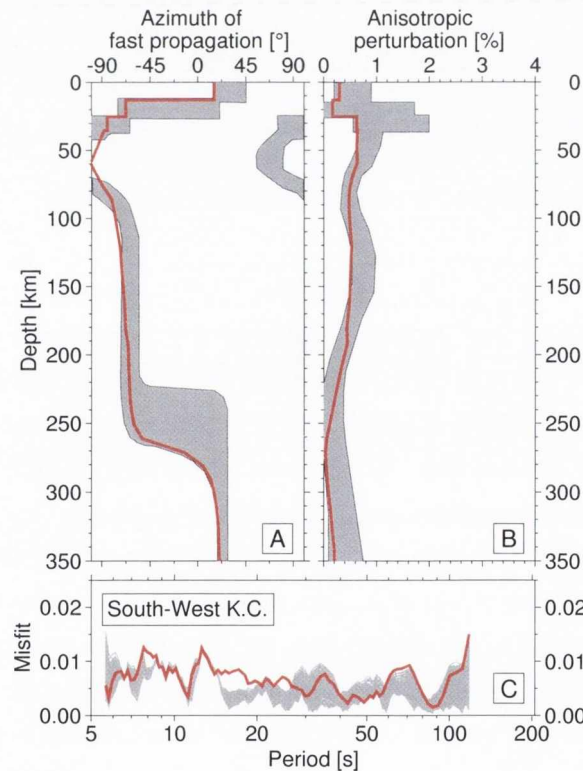


Figure 4.16.: Resolution of fast propagation direction between 40 and 90 km in South-West Kaapvaal Craton. Gray bands show the set of models that best fit the data. Red lines are for a model with no sharp variations of azimuthal anisotropy between 40 and 90 km. C: Deviation of the models to the data.

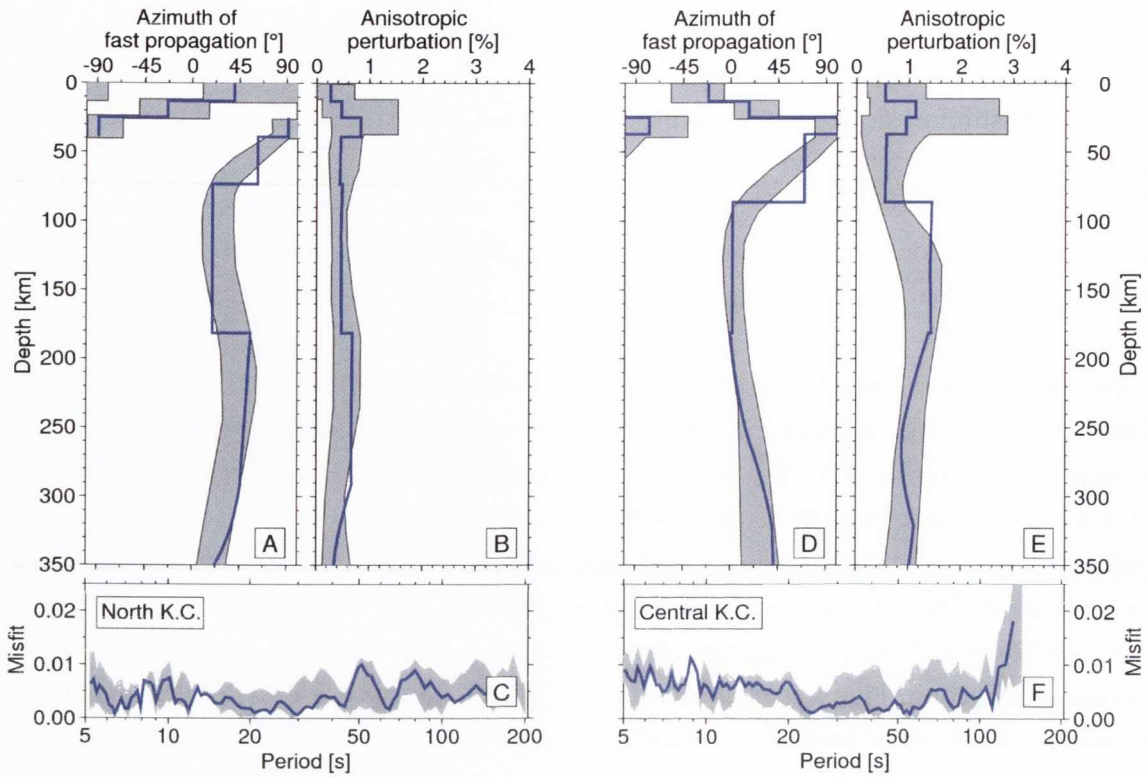


Figure 4.17.: Resolution of the transition of fast propagation directions around 90 km in North (A,B,C) and central (D,E,F) Kaapvaal Craton. Gray bands show the set of models that best fit the data. Blue lines are for a model with sharp variations of azimuthal anisotropy around 90 km. D,F: Deviation of the models to the data.

#### 4.4.4 Discussion

The inversions confirm our initial inferences from the azimuthal anisotropy of Rayleigh wave phase velocity (Section 3.7). They also show what could not be seen from the phase velocities: anisotropy in the mantle-lithosphere is strongly layered, with varying direction of fast propagation with depth. We showed in Section 3.7 that the origin of the SKS splitting was both in the lithosphere and in the asthenosphere. According to Vinnik *et al.* (1995), the SKS splitting originates in the asthenosphere and reflects the plate motion. An other interpretation from Silver *et al.* (2004) suggested that the splitting originates in the mantle lithosphere and is related to the ancient deformation. Our observations of the distribution of anisotropy reconciled with both models (Figure 4.18). Firstly, the SKS splitting in the Limpopo Belt and North Kaapvaal Craton seems to originate in the lower-lithosphere, reflecting the deformation and flow during the Archean collision (2.7 Ga) and subsequent Paleo-Proterozoic deformation (Silver *et al.*, 2004). Secondly, the direction of fast propagation in the asthenosphere (at depth greater than 200–250 km) tend to converge to a similar azimuth in the four regions and align with plate motion ( $\sim 45^\circ$ ) (Figure 4.18), confirming the argument of Vinnik *et al.* (1995). Surface wave studies in Northern Amer-



ica (Deschamps *et al.*, 2008b; Yuan and Romanowicz, 2010) and Australia (Debayle and Kennett, 2000) observed a similar correlation between the azimuth of fast propagation in the asthenosphere and the current plate motion.

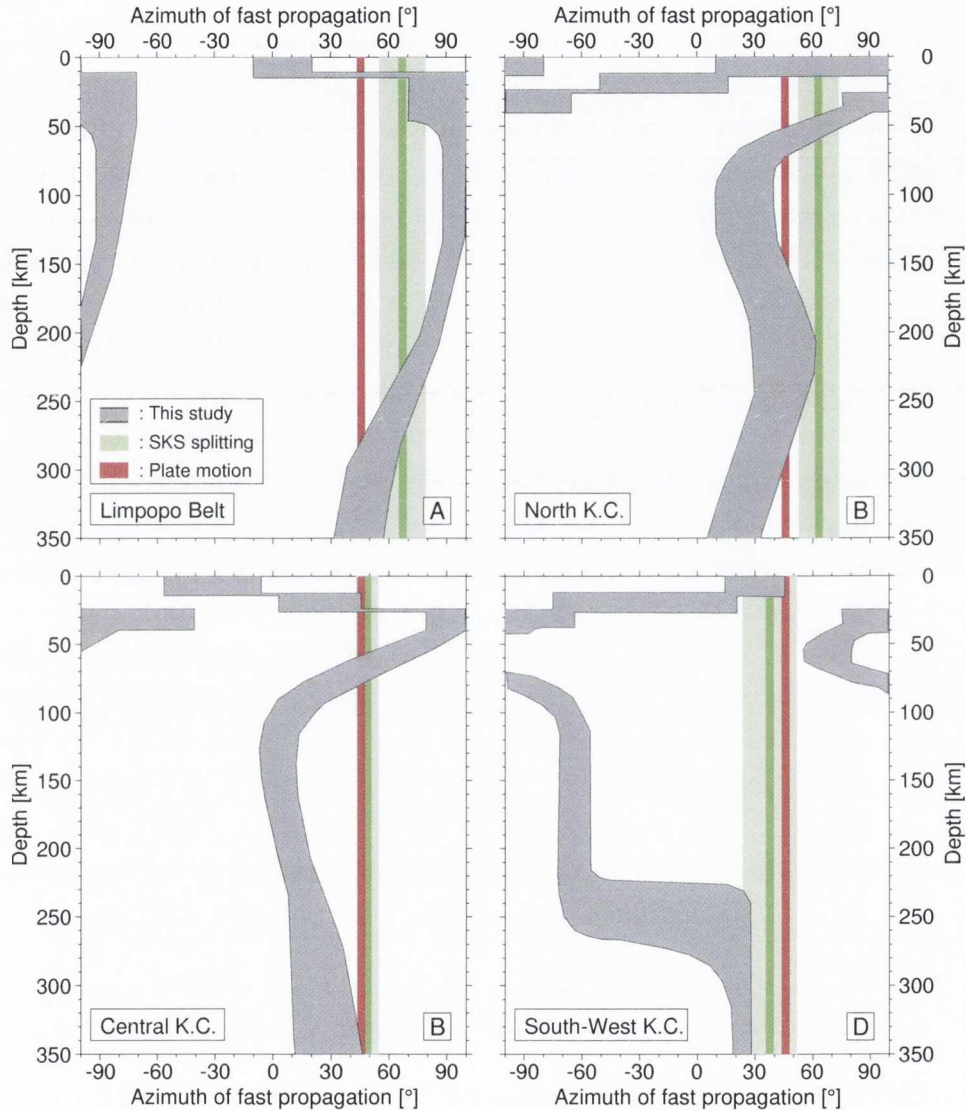


Figure 4.18.: Comparison of depth-profiles of azimuths of fast propagation from this study (gray bands) with plate motion (red) (Gripp and Gordon, 1990) and averaged fast directions inferred from SKS-splitting measurements (Silver *et al.*, 2004) (dark green lines) with their standard deviation (light green band).

The change of direction of fast propagation at  $\sim 90$  km in the Kaapvaal Craton (and particularly in the South-West Kaapvaal Craton) is of particular interest as it may have been observed in receiver function studies, in southern Africa and around the globe (Wittlinger and Farra, 2007; Savage and Silver, 2008; Rychert and Shearer, 2009; Fischer *et al.*, 2010) and by joint inversion of surface wave waveforms and SKS splitting measurements

in North America (Yuan and Romanowicz, 2010) but might not be related to a change of isotropic shear velocity. Indeed, in the case of southern Africa, the evidence of a discontinuity at ~90 km as a change of isotropic velocity is not clear (Figure 4.11). Wittlinger and Farra (2007) argued that this boundary is not imaged by tomographic studies because it was not related to isotropic velocity variations but to anisotropic variations. This boundary would be described by a variation of radial or azimuthal anisotropy within the lithosphere. Rychert and Shearer (2009) compiled P-to-S converted phases from all around the globe and mapped the 90-100 km boundary at a global scale. They interpreted this boundary as the lithosphere-asthenosphere boundary under the oceans and the continents (excluding cratons). Beneath cratons, this boundary could not be related to the LAB as xenolith studies showed that the lithosphere is coupled and stable down to at least 150 km (Fischer *et al.*, 2010, and references therein). Wittlinger and Farra (2007) and Savage and Silver (2008) also argued that this ~90 km boundary is a converter within the lithosphere and is not the LAB.

Using Rayleigh wave phase velocity measurements we did not find evidence of isotropic-averages shear velocity discontinuities at depths around 90 km. We, however, observe an abrupt change of direction of fast propagation and amplitude of anisotropy around the same depth. Indeed, azimuthal anisotropy (Figure 4.16, Figure 4.17) shear velocity profiles show either gradual or sharp transition between two anisotropic layers. We also inverted Rayleigh- and Love-wave dispersion curves for depth profiles of radial anisotropy and found similar results *i.e.* a change of anisotropy more or less sharp at around 90 km (Figure 4.19). Such a transition, if sharp, would generate conversions of body waves that could be recorded in receiver function studies. The transition of azimuthal or radial anisotropy is observed in the three regions of the Archean Kaapvaal Craton. However, the data do not show any evidence for a anisotropic boundary in the Paleo-Proterozoic Limpopo Belt, and could suggest that the mantle-lithosphere of the Limpopo Belt was deformed as an unit.

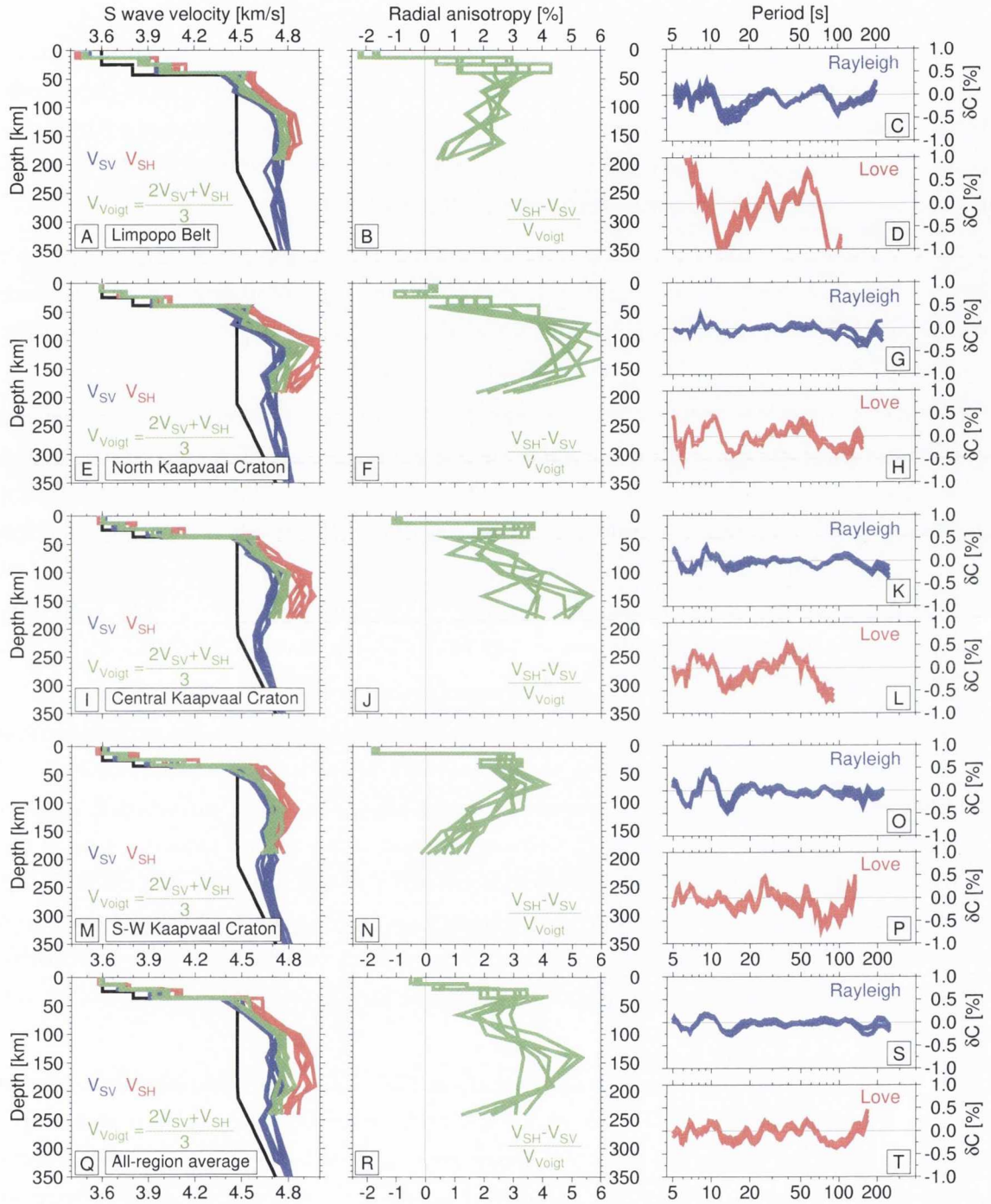


Figure 4.19.: Radially anisotropic shear velocity depth profiles and fits to the data for the four sub-regions in southern Africa and the all-average region. The isotropic Voigt average (green lines in A,E,I,M,Q), horizontally/vertically polarized shear velocities (blue/red lines in A,E,I,M,Q) and radial anisotropy (B,F,J,N,R) are only represented at depths constraints by the data. Right column: deviation between the model and the data (blue: Rayleigh wave, red: Love wave).



## 4.5 CONCLUSION

We inverted measurements of Rayleigh phase velocities and azimuthal anisotropy parameters to obtain depth-profiles of the vertically polarized shear velocity and the depth-distribution of azimuthal anisotropy. Before inverting real data from southern Africa, we focused our study on the inversion technique and its resolution power, in order to establish the robustness of the models obtained with this method.

We found that azimuthal anisotropy can be resolved with accuracy if its signal is greater than the noise in the data. We confirmed this observation by mapping the model-space in inversions of data from southern Africa and showed a clear dependence between the resolution of anisotropy and the amplitude of the noise.

The exploration of the model-space showed that there are no trade-offs between the amplitude of anisotropy and the direction of fast propagation. They have uncorrelated uncertainties and variations of the direction of fast propagation can not compensate for variations of the amplitude of anisotropy (or *vice versa*). The model-space investigation also showed that the *a posteriori* errors on the parameters are small and the azimuth of fast propagation is better resolved than the amplitude. The model-space is simple, with only one global minimum, justifying the use of a gradient search technique to find models that would best fit the data.

We determined depth-profiles of the vertically polarized shear velocity and confirmed results from previous studies: the mantle-lithosphere beneath the Kaapvaal Craton and the Limpopo Belt is characterized by a fast root extending down to approximately 200 km. Below that lid, the velocity seems to decrease, forming a weak low velocity zone in the Limpopo Belt and North Kaapvaal Craton but stronger in the southern part of the Kaapvaal Craton. A temperature differences between the sub-lithospheric mantle of the South Kaapvaal Craton and the North Kaapvaal Craton and Limpopo Belt could be responsible for the south-north decrease of topography at the surface (Lithgow-Bertelloni and Silver, 1998).

We presented robust depth distributions of azimuthal and radial anisotropy and showed that the lithosphere is layered, with distinct directions of fast propagation and amplitudes of anisotropy within the crust and the mantle. The anisotropy in the upper mantle-lithosphere seems to differ from the anisotropy in the lower mantle-lithosphere, suggesting that the upper and lower lithosphere underwent different deformation during their last major deformation episodes and may not have stabilized at the same time. The history of the craton and its neighboring belts seems to be more complex than previously expected.

Previous receiver function studies yielded evidence for a boundary at ~90 km in the Kaapvaal Craton, but also around the globe. This signal may be related to anisotropy.

We confirmed the existence of an intra-lithospheric discontinuity and observed an abrupt change of azimuthal and radial anisotropy at depths around 90-100 km.

Part III

DEPTH RESOLUTION OF THE MOHOROVIČIĆ  
DISCONTINUITY



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## DEPTH RESOLUTION OF THE MOHOROVIČIĆ DISCONTINUITY<sup>1</sup>

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In 1921, Tams (1921) and Angenheister (1921) studied and compared the propagation of surface-waves along oceanic and continental crust. They observed differences in the wave-train between surface-waves propagating in an oceanic or continental crust. Nowadays, the discrepancy of the wave propagation along oceanic and continental crust is well-known ; it is mainly due to the change of the depth of the Moho. Signatures of the crustal thickness are observed in phase-velocity or group-velocity curves of surface-waves as they show a characteristic increase or local minima in velocity due to the Mohorovičić discontinuity (Figure 5.1, Figure 5.2). The periods at which these features in the curves are observed inform about the depth of the discontinuity (Figure 2.6, Figure C.1). The Moho signature in phase or group velocities in the oceanic model is more marked as it correspond to a higher contrast of velocity between the crust and the mantle. Regarding the Tibetan model, the Moho signature is less marked. Despite the large contrast between crustal and mantle velocities, the Moho is deeper than for the oceanic model and because surface waves span large depth ranges, they are less sensitive to velocity contrasts.

In this part of this thesis, we present the sensitivity of surface-waves to the Moho discontinuity by analyzing the effects of the Mohorovičić discontinuity on the shape of the surface-wave dispersion curves. We use the examples of an oceanic crustal model, a typical continental crustal model and a model representing Tibetan crust. Both forward modelling and inversion techniques are used to investigate the sensitivity of the surface-waves. Forward modelling is a deterministic method that determines the dispersion curve corresponding to a particular Earth model. We use this approach to directly observe the influence of a varying model and crustal thickness to the sensitivity of the surface-waves. In contrast, the inversion procedure is using an algorithm which finds a model of the Earth that best fits the surface-wave dispersion-curve, allowing for trade-offs between the parameters. Both methods indicate the period band at which the dispersion curve is the

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<sup>1</sup> The material presented in this chapter forms a part of an article published to the journal *Tectonophysics* as : Lebedev, S., J. Adam & T. Meier. 2013. Mapping the Moho with seismic surface waves: A review, resolution analysis, and recommended inversion strategies. *Tectonophysics*, "Moho" special issue, doi:10.1016/j.tecto.2012.12.030

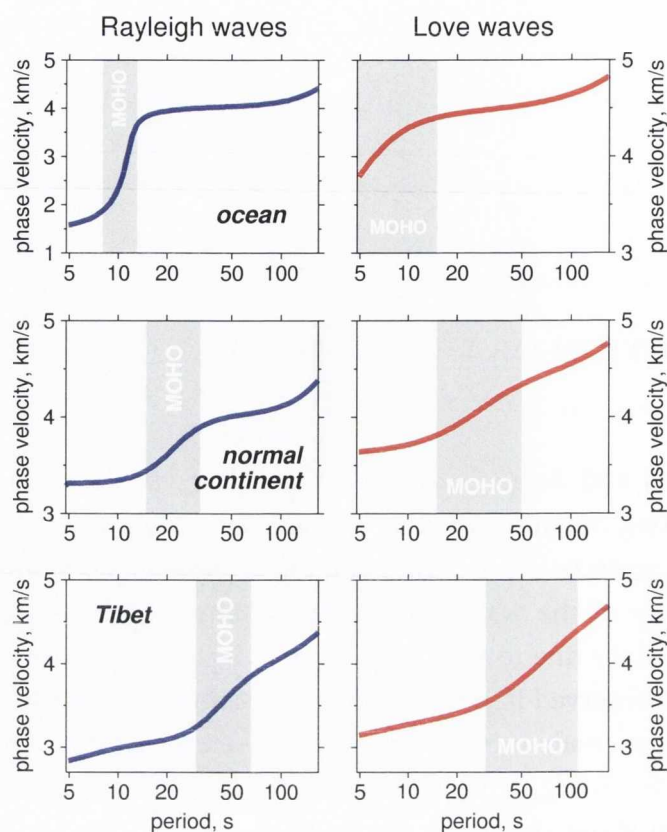


Figure 5.1.: Signature of the Moho in phase-velocity curves of surface-waves. Phase velocities of the fundamental-mode Rayleigh and Love waves were computed for an oceanic with a 5 km water layer and a 6 km thick crust (top), a continental model with a 37 km thick crust (middle), and a model with a 65 km thick crust that fits surface-wave data from NE Tibet Agius and Lebedev (2010). The period ranges with the characteristic phase-velocity increase with period due to the S-velocity increases at the Moho are marked with grey shading.

most sensitive to the discontinuity. Furthermore, the results of the inversion also inform about the impact of the crustal and mantle layers to the resolution of the Moho depth.

We also investigate the trade-offs between the Moho depth and the crustal and mantle shear-velocities and the Moho thickness to analyze the resolution of the Moho. Knowing the Moho-depth resolvability with surface-waves, we then discuss strategies to best estimate the Moho-depth, with applications to real data from southern Africa.

## 5.1 DISCOVERY OF THE MOHOROVIČIĆ DISCONTINUITY

Andrija Mohorovičić was born in Croatia in 1857. After graduating from high-school, he studied mathematics and physics at the University of Prague. He first worked as a meteorologist and then as a seismologist. Between 1891 and 1921, he headed the now called *Andrija Mohorovičić Geophysical Institute* in the faculty of Science at the University of Za-

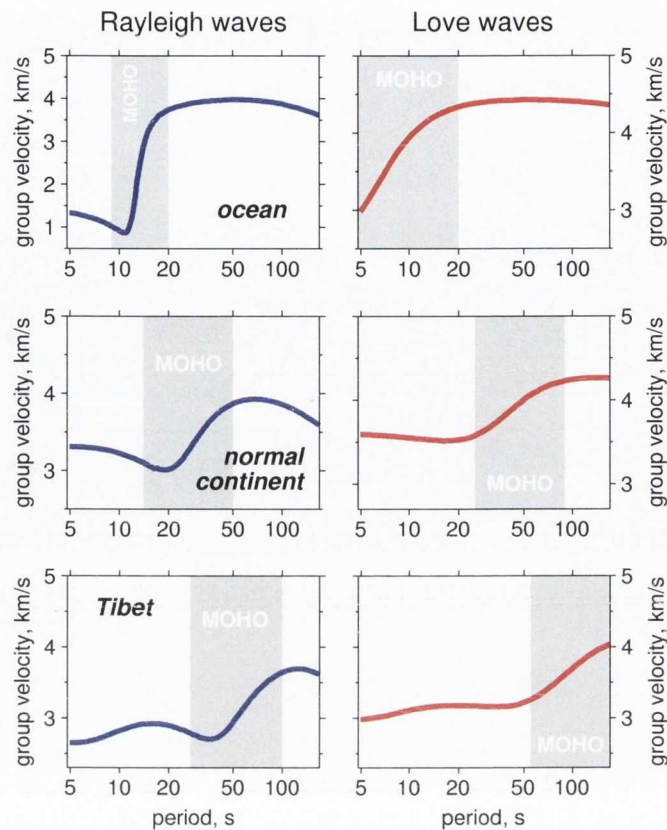


Figure 5.2.: Signature of the Moho in group-velocity curves of surface-waves. Group velocities of the fundamental-mode Rayleigh and Love waves were computed for an oceanic with a 5 km water layer and a 6 km thick crust (top), a continental model with a 37 km thick crust (middle), and a model with a 65 km thick crust that fits surface-wave data from NE Tibet Agius and Lebedev (2010). The period ranges with the characteristic phase-velocity increase with period due to the S-velocity increases at the Moho are marked with grey shading.

greb (Croatia) (Skoko, 1992). He was a pioneer in the study of crustal discontinuities and an important figure in the study of the Earth. In his paper published in 1913 (Mohorovičić, 1913), A. Mohorovičić stated that seismology is “...to investigate the interior of the Earth and to take over where the geologist stops, because the modern seismographs can serve as a binocular for observing even the greatest depths...”. He developed ideas that are now considered as fundamentals and are still relevant today.

After the Kupa Valley, Croatia earthquake of October, 8 1909, A. Mohorovičić collected records of 39 stations from all around Europe (within ~2400 km ; the farthest station being in Moscow). He used the seismographic records to locate the epicenter and also detected two sets of P and S wave arrivals. He interpreted this as evidence for two paths travelled by seismic energy, one near the Earth’s surface and the other refracted by a deeper, higher-velocity medium, and inferred the presence of a velocity discontinuity that is now known to be the discontinuity between the crust and mantle (Mohorovičić, 1910; Skoko, 1992). This was a considerable discovery since at that time the only discontinuities in the



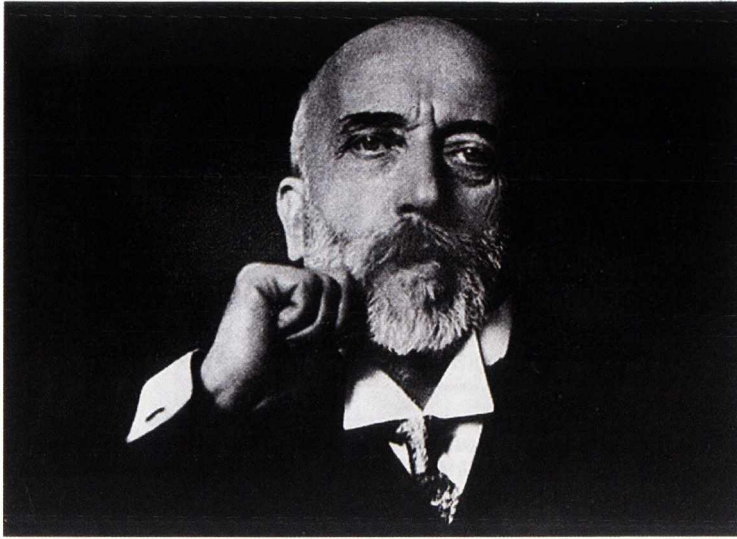


Figure 5.3.: Portrait of Andrija Mohorovičić (Ph. DR Coll. Archives Larbor, 2012).

crust were believed to be very shallow and due to tectonic deformations (Bullen, 2008). The relevance and potency of this discovery was only recognized later, thanks to the paper published in 1912 by Bendorf (Bendorf, 1912). Today, this discontinuity is called the Mohorovičić discontinuity or shortly, the Moho.

## 5.2 WHAT IS THE MOHOROVIČIĆ DISCONTINUITY ?

The Moho discontinuity is the boundary between the Earth crust and the Earth mantle. This discontinuity can be described seismologically and lithologically. From a seismological point of view, the Moho is represented by a jump in the seismic wave velocities. The crust is defined as the outer layers of the Earth with P-wave velocities lower than  $7.6 \text{ km.s}^{-1}$  and S-wave velocities lower than  $4.3 \text{ km.s}^{-1}$  (e.g. Meissner, 1986). The P-wave velocity in the mantle is, normally, over  $8 \text{ km.s}^{-1}$  which creates a contrast of velocity of  $\sim 1.0 - 1.5 \text{ km.s}^{-1}$  (Grad *et al.*, 2009) (Figure 5.5, Figure 5.6, Figure 5.7). This jump of velocity is related to a change of rock compositions, resulting in a contrast in elastic parameters (Grad *et al.*, 2009).

From a lithologically point of view, the Moho discontinuity is described by a jump in density. The density is approximately  $\sim 3.0 \text{ g.cm}^{-3}$  above the Moho discontinuity and  $\sim 3.3 \text{ g.cm}^{-3}$  below it (Talwani *et al.*, 1959; Grad *et al.*, 2009) (Figure 5.5, Figure 5.6, Figure 5.7).

Density, velocity and rheology changes as well as mineralogical and chemical composition changes are related and all define the Moho discontinuity (Burg, 2011; Birch, 1960;

Christensen and Mooney, 1995).

Mapping the Moho discontinuity (by defining its depth and/or lateral variations) is pertinent to understand the geodynamics and geological history of the Earth. The Moho discontinuity is a key element to reconstruct lithospheric processes (extension, collision, etc.), model the dynamics of the crust and the mantle, and model the heat flow (Jiménez-Munt *et al.*, 2008; Tirel *et al.*, 2008).

Several methods exist to map the Moho and seismic ones are particularly effective as they use different types of waves, resolving the Moho at complementary accuracy and resolution. Gravimetry also provides important complementary data on the Moho at a larger scale.

### 5.2.1 Gravimetry

The change of density between the crust and the mantle creates gravimetric anomalies that can be measured at the surface of the Earth (Ranganai *et al.*, 2002; Grad *et al.*, 2009; Reid *et al.*, 2012) or above it *via* satellites (Shin *et al.*, 2007; Sampietro, 2011). Measurements of gravitational effects allow one to reconstruct the depth variation of density and then map the depth of the Moho. Today, the GEMMA (GOCE Exploitation for Moho Modeling and Applications) (Sampietro, 2012) project aims to estimate the depth of the Moho around the world with high-resolution. It uses satellite gravity data from the GOCE (Gravity field and steady-state Ocean Circulation Explorer) mission launched on the 17<sup>th</sup> of March 2009 which measures the static gravity field (ESA - GOCE, 2012). Preliminary results from the GEMMA project show promising high-resolution global models of the depth of the Moho. Furthermore, they detected discrepancies between their models and seismic models suggesting the presence of previously unknown anomalies (Sampietro, 2011). Other satellite missions, including CHAMP (CHAllenging Mini-satellite Payload) (Lühr, 2010) launched on the 15<sup>th</sup> of July 2000 and GRACE (Gravity Recovery and Climate Experiment) (GRACE, 2002) launched on the 17<sup>th</sup> of March 2002, also measured the variations of the gravity field. Shin *et al.* (2007) for example, used the data collected by those satellites to estimate the depth of the Moho beneath Tibet.

### 5.2.2 Controlled-source seismology

Controlled-source seismic exploration uses artificial sources and dense arrays of seismometers to image the crust by measuring the reflections or refractions of seismic body-waves between the source and the station (Prodehl, 2011, and references therein). This technique is relatively expensive and labour-intensive but produces high-resolution profiles of the crust. In Southern Africa, De Wit and Tinker (2004) investigated the crustal



structure and the Moho depth in central Kaapvaal Craton. They used the vibroseis method to generate eight high-resolution deep-seismic reflection profiles over a total of ~500 km and imaged the Moho.

### 5.2.3 *Passive seismology*

In contrast to controlled-source seismology, passive seismology studies records of signals from natural sources (ambient seismic noise, local or teleseismic earthquakes) to image the Earth interiors (Das and Nolet, 1998a; Julia *et al.*, 2000; Zhu and Kanamori, 2000; Bostock *et al.*, 2002; Kind *et al.*, 2002; Stankiewicz *et al.*, 2002; Endrun *et al.*, 2004; Koulakov and Sobolev, 2006; Yang *et al.*, 2008; Nabelek *et al.*, 2009; Rychert and Shearer, 2010; Tkalčić *et al.*, 2012). The large amount of earthquakes and natural sources and available data from stations worldwide enable to either map the crust at a local scale (*e.g.* Durrheim and Green, 1992, studied the crust in Southern Africa using local mine tremors and nearby stations) or map the deep interior at a global scale using teleseismic events (*e.g.* Schaeffer and Lebedev, 2012, used a large amount of surface-wave signals and studied the upper-mantle isotropic and anisotropic structures). Using broad-band seismic networks, Das and Nolet (1998a,b); Meier *et al.* (2007a,b) inverted surface-wave measurements for tomographic models focusing mainly on the Moho discontinuity.

## 5.3 SENSITIVITY AND RESOLUTION OF THE MOHO DISCONTINUITY

### 5.3.1 *Oceanic, continental and Tibetan crust models*

We investigate three 1-D synthetic models of the Earth that correspond to three types of the crust: oceanic (Table 5.1, Figure 5.5), normal continental (Table 5.2, Figure 5.6) and Tibetan (Table 5.3, Figure 5.7). The upper mantle structure of the oceanic and normal-continent are, for simplicity, from the same global reference model. Figure 5.4 shows a geographic map with locations at which the three models are located.

The oceanic crustal model is calculated with a modified AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000) reference models at the coordinates (0°,180°), which corresponds to the mid-Pacific Ocean. The model contains a layer of water and a layer of sediment atop the crystalline crustal layers (Table 5.1, Figure 5.5). The Moho discontinuity is at 11 km depth (below sea level).

The normal continent model is a crustal model of southern Africa's Kaapvaal Craton at coordinates (-28.614°,25.256°), computed with a modified AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000) as reference. It represents typical continental crust and



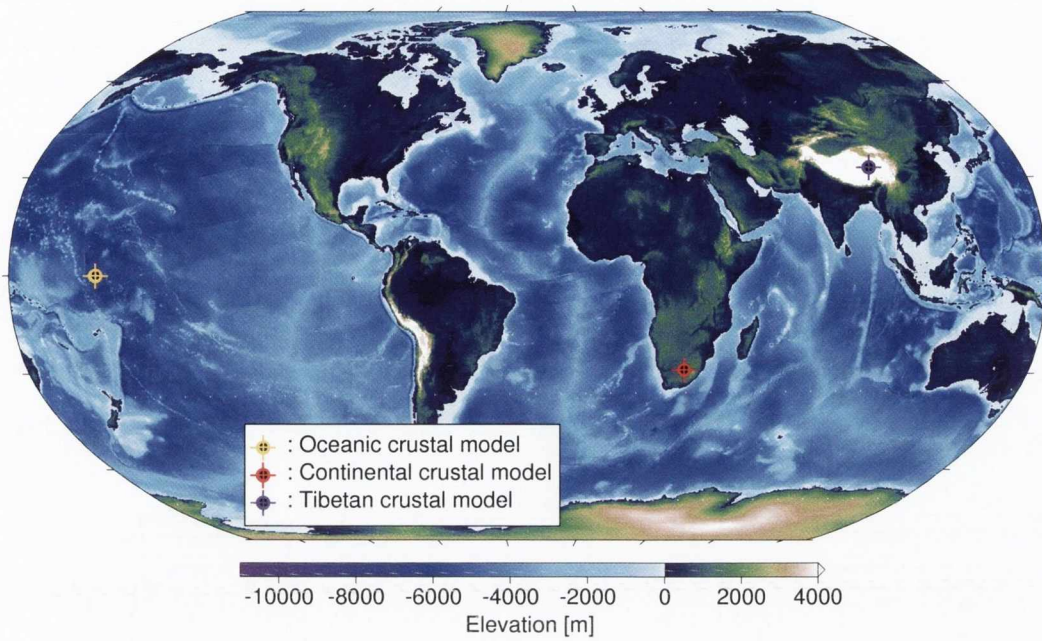


Figure 5.4.: Geographic map and localisation of the crustal models. Yellow circle: oceanic crust. Red circle: normal continental crust. Purple circle: Tibetan crust.

mantle with a Moho depth of 37.0 km (Table 5.2, Figure 5.6).

The Tibetan model is representative of the East-central Qiangtang region in Tibet. This models is the result of an inversion for S-wave ( $V_s$ ) velocity profiles using surface-wave dispersion curves recorded in the region (Agius and Lebedev, 2010). The S-wave velocity model used in this study is the Voigt average representing the average between the vertical ( $V_{sv}$ ) and the horizontal ( $V_{sh}$ ) components of the S-wave velocity (Equation 5.1).

$$V_{\text{Svoigt}} = \frac{2V_{sv} + V_{SH}}{3} \quad (5.1)$$

The P-wave velocity is computed as the product of the voigt average and the  $V_p/V_s$  ratio of this region. The Moho discontinuity is 65 km deep and the middle-crust as well as the upper-mantle comprise a low velocity zone (Table 5.3, Figure 5.7). See Agius and Lebedev (2010) for more information.

Table 5.1.: Earth model of an oceanic crust and upper-mantle calculated with a modified AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000) at coordinates (0°,180°).  $\rho$ : density,  $V_p$ : compressional velocity,  $V_s$ : shear velocity and  $Q$ : quality factor.

| Layer        | Depth range [km] | $\rho$ [kg.m <sup>-3</sup> ] | $V_p$ [km.s <sup>-1</sup> ] | $V_s$ [km.s <sup>-1</sup> ] | $Q$ [Ø] |
|--------------|------------------|------------------------------|-----------------------------|-----------------------------|---------|
| water        | 0.0-4.5          | 1020                         | 1.50                        | 0.00                        | 0.0     |
| sediments    | 4.5-4.6          | 1700                         | 1.80                        | 0.80                        | 50.0    |
| upper-crust  | 4.6-6.2          | 2600                         | 5.00                        | 2.50                        | 300.0   |
| middle-crust | 6.2-8.5          | 2900                         | 6.60                        | 3.65                        | 300.0   |
| lower-crust  | 8.5-11.0         | 3050                         | 7.10                        | 3.90                        | 300.0   |
| upper-mantle | 11.0-...         | 3641                         | 8.02                        | 4.47                        | 394.6   |

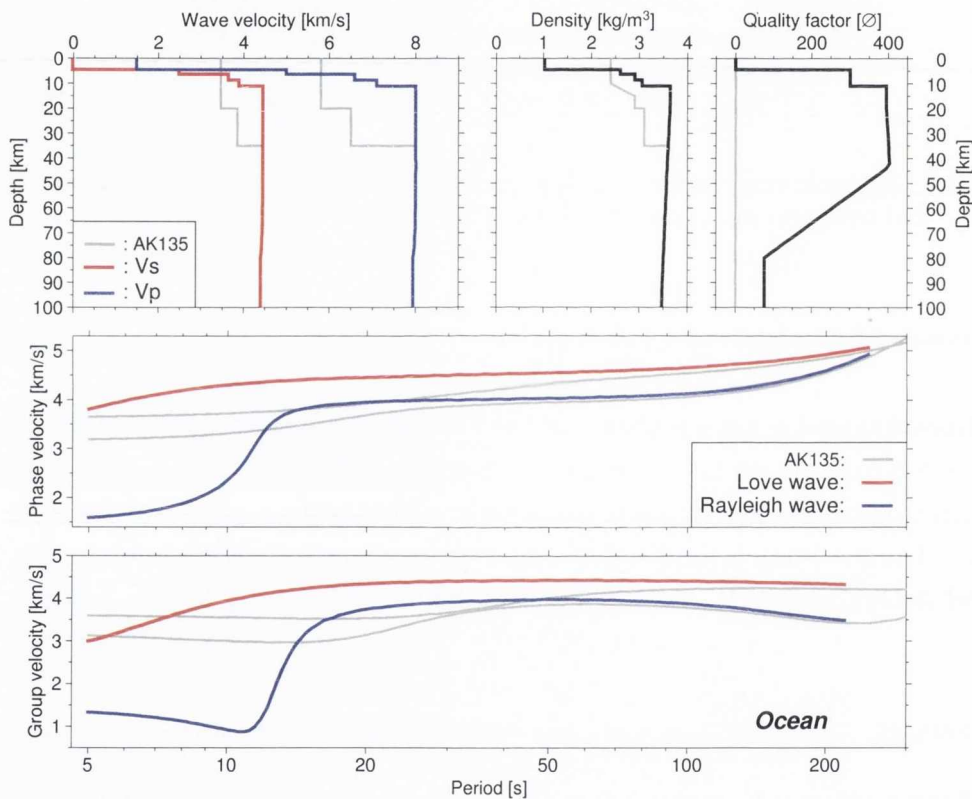


Figure 5.5.: Oceanic Earth model calculated with a modified AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000) at coordinates (0°,180°). Top: shear velocity ( $V_s$ , red), compressional velocity ( $V_p$ , blue), density and quality factor. See Table 5.1 for precise values. Middle: Rayleigh (blue) and Love (red) phase velocities computed for this model (see Section 5.3.3). Gray lines represents the reference AK135. The bottom panel shows the corresponding group velocities (see Appendix C).

Table 5.2.: Earth model of a continental crust and upper-mantle calculated with a modified AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000) at coordinates (-28.614°,25.256°).  $\rho$ : density,  $V_p$ : compressional velocity,  $V_s$ : shear velocity and  $Q$ : quality factor.

| Layer        | Depth range [km] | $\rho$ [kg.m <sup>-3</sup> ] | $V_p$ [km.s <sup>-1</sup> ] | $V_s$ [km.s <sup>-1</sup> ] | $Q$ [Ø] |
|--------------|------------------|------------------------------|-----------------------------|-----------------------------|---------|
| upper-crust  | 0.0-13.0         | 2800                         | 6.20                        | 3.60                        | 300.0   |
| middle-crust | 13.0-25.0        | 2850                         | 6.50                        | 3.60                        | 300.0   |
| lower-crust  | 26.0-37.0        | 2950                         | 6.80                        | 3.80                        | 300.0   |
| upper-mantle | 37.0-...         | 3597                         | 8.21                        | 4.47                        | 401.3   |

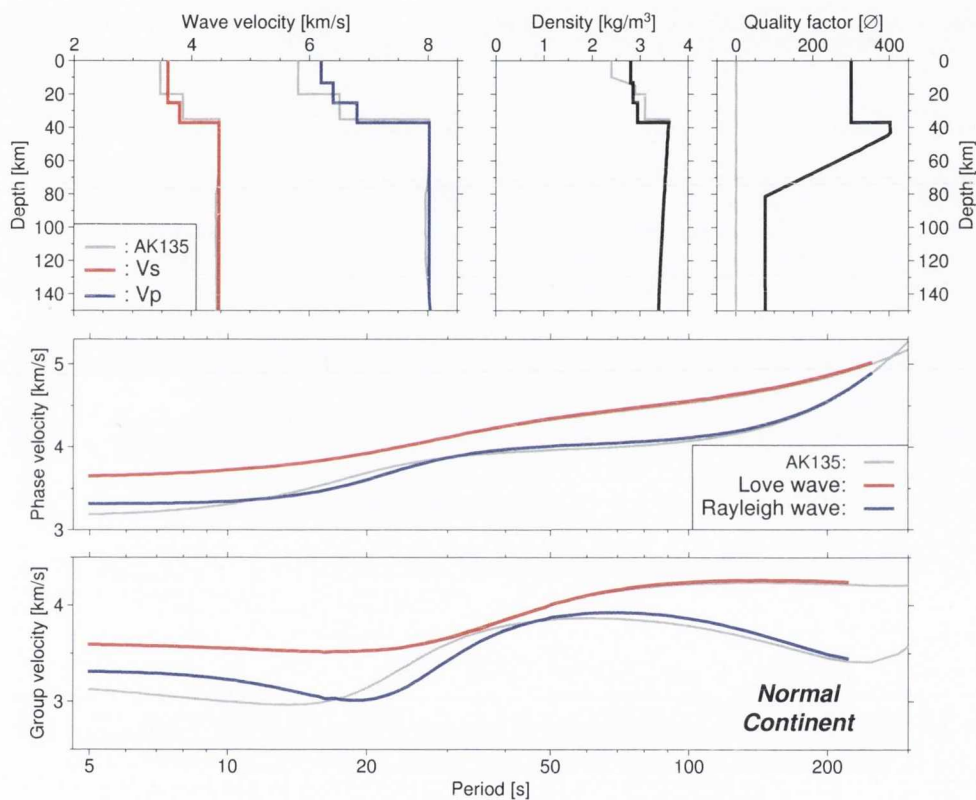


Figure 5.6.: Continental Earth model calculated with a modified AK135 (Kennett *et al.*, 1995) and crust 2.0 (Bassin *et al.*, 2000) at coordinates (-28.614°, 25.256°). Top: shear velocity ( $V_s$ , red), compressional velocity ( $V_p$ , blue), density and quality factor. See Table 5.2 for precise values. Middle: Rayleigh (blue) and Love (red) phase-velocities computed for this model (see Section 5.3.3). Gray lines represents the reference AK135. The bottom panel shows the corresponding group velocities (see Appendix C).



Table 5.3.: Earth model of a Tibetan crust and upper-mantle representative of the East-central Qiangtang region in Tibet (Agius and Lebedev, 2010). $\rho$ : density,  $V_p$ : compressional velocity,  $V_s$ : shear velocity and  $Q$ : quality factor.

| Layer              | Depth range [km] | $\rho$ [kg.m <sup>-3</sup> ] | $V_p$ [km.s <sup>-1</sup> ] | $V_s$ [km.s <sup>-1</sup> ] | $Q$ [Ø] |
|--------------------|------------------|------------------------------|-----------------------------|-----------------------------|---------|
| upper-crust        | 0.0-7.9          | 2700                         | 5.26                        | 2.99                        | 300.0   |
| middle-crust       | 7.9-20.6         | 2700                         | 5.96                        | 3.39                        | 300.0   |
| middle-crust (LVZ) | 20.6-45.1        | 2850                         | 5.89                        | 3.35                        | 300.0   |
| lower-crust        | 45.1-65.1        | 3100                         | 6.33                        | 3.60                        | 300.0   |
| upper-mantle       | 65.1-...         | 3523                         | 8.05                        | 4.49                        | 164.3   |

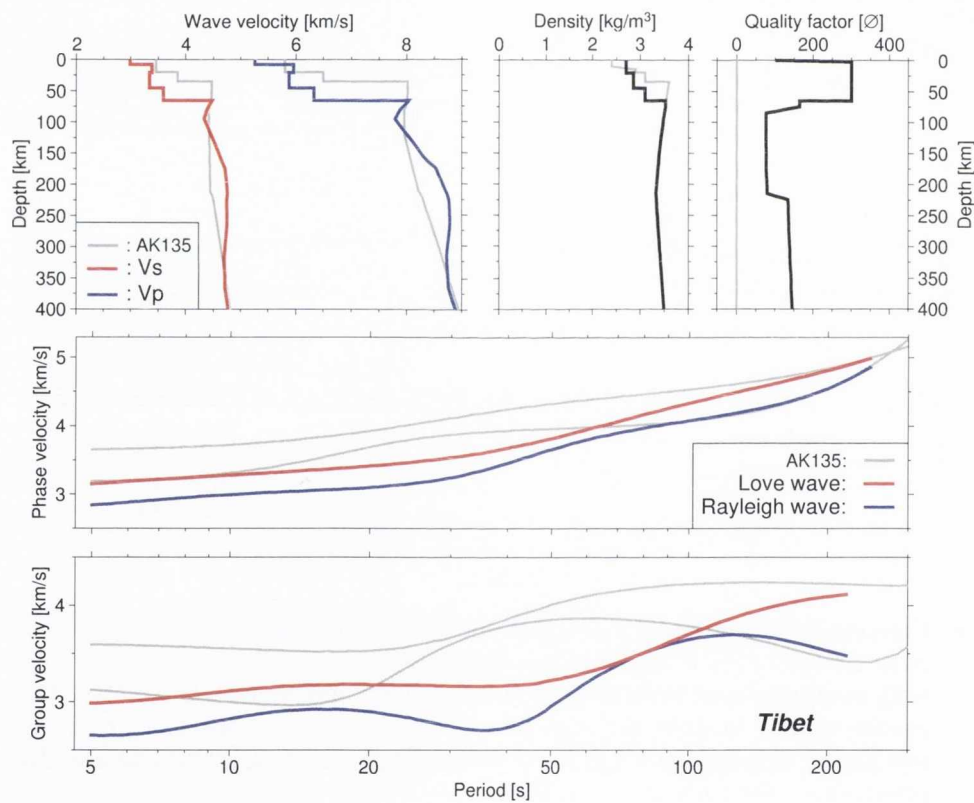


Figure 5.7.: Tibetan Earth model representative of the East-central Qiangtang region in Tibet (Agius and Lebedev, 2010). Top: shear velocity ( $V_s$ , red), compressional velocity ( $V_p$ , blue), density and quality factor. See Table 5.3 for correct values. Bottom: Rayleigh (blue) and Love (red) phase velocities computed for this model (see Section 5.3.3). Gray lines represents the reference AK135. The bottom panel shows the corresponding group velocities (see Appendix C).

### 5.3.2 Generation of Earth models with varying Moho depths

The reference Earth models presented in [Section 5.3.1](#) are used as the reference data (or background models). To generate models with varying Moho depths, the Mohorovičić discontinuity is shifted up and down within a range of approximately  $\pm 15$  km around the actual depth of the Moho. The Moho depth is sampled every kilometer. For each region, the upper-limit of the Moho depth has been adjusted not to interfere with shallower layers. ([Figure 5.8](#)). The parameters ( $\rho$ ,  $V_s$ ,  $V_p$  and  $Q$ ) of the crust and upper-mantle are all adjusted to appropriate crustal or mantle values to reflect the Moho depth changes. Indeed, surface waves are sensitives to depth variations of those parameters ([Figure 2.6](#), [Figure C.1](#)). However, the sensitivity to shear velocity is stronger, we shall then only focus this study on shear velocity variations.

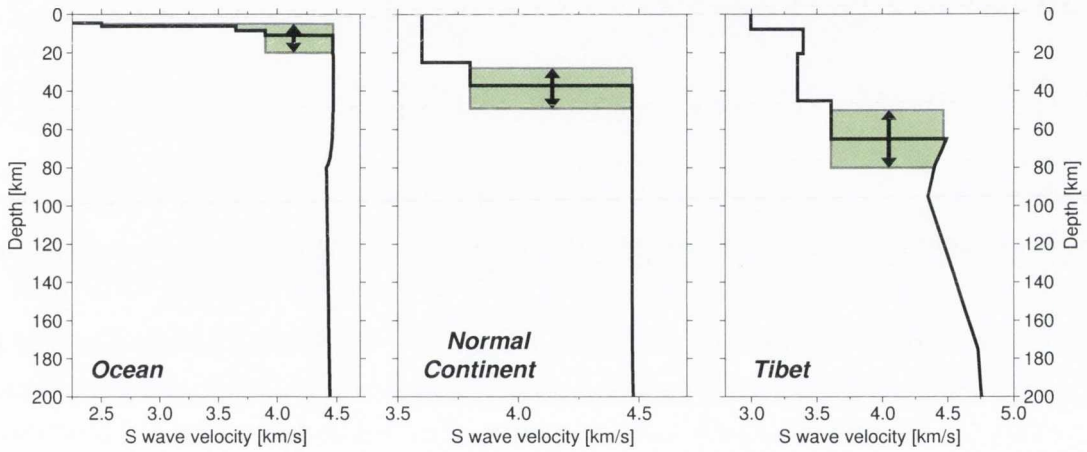


Figure 5.8.: Depth intervals in which the Moho vary for each of the three Earth model (Oceanic, synthetic and Tibet). The depth of the Moho varies between 5 and 20 km for the oceanic crust, 28 and 60 km for the continental crust and 50 and 80 km for the Tibetan crust.

For each region, we obtain a set of models in which the depth of the Moho is the only parameter that vary. To each model of this dataset corresponds a dispersion curve for Rayleigh and Love waves (computed either with forward modeling ([Section 5.3.3](#)) or with an inversion algorithm ([Section 5.3.4](#))). Independently for each wave, the difference between the obtained dispersion curve and the reference dispersion curve informs about the impact of the Moho depth shift. This difference is calculated using two methods:

1. the misfit [ $\emptyset$ ]:

$$\text{misfit} = \frac{1}{N} \sqrt{\sum_T \left( \frac{d - m}{d} \right)^2} \quad (5.2)$$

2. the phase velocity perturbation of the dispersion curve [%]:

$$\text{perturbation} = \frac{m - d}{m} * 100 \quad (5.3)$$

Where  $d$  corresponds to the reference Earth-model,  $m$  to the Earth-model with varying Moho-depth,  $N$  is the number of measurements and  $T$  to the period. Both set of values are important and interesting to explore the sensitivity of surface-waves to the depth of the Moho discontinuity. On the one hand, the misfit measures the deviation between the data and the model. The smaller the misfit, the better the correspondence of the reference and the model. On the other hand, the phase velocity perturbation as a function of period shows period-dependent difference between the data and the model. It shows the period range at which the surface-waves are most sensitive to the change of velocity due to the modification of the depth of the discontinuity.

### 5.3.3 Resolution of the Moho depth using forward modeling

In this section we investigate the sensitivity of surface-waves to the crustal thickness using forward modeling. We directly compare the signatures of the Moho depth in dispersion curves using a set of models that are different only by the depth of the Moho in them.

Theoretical phase-velocities are calculated for an 1-D model using an adaptation of the MINOS code of Guy Masters (Masters; Nolet, 1990). At every single and independent frequency, the fundamental-mode phase velocity is computed for a given Earth-model (Section 5.3.2). During the computation, anisotropy is ignored. Rayleigh- and Love-wave dispersion curves are computed independently.

Group velocities are related to phase velocities and can be computed from them (see Section 2.1). We therefore compute group-velocities as well and investigate their sensitivity to the depth of the Moho. Figure 5.9, Figure 5.10 and Figure 5.11 show the sensitivity of Rayleigh- and Love-waves to the Moho depth in the three tectonic settings: ocean, normal continent and Tibet. The Moho depth range is represented by the black and gray lines in Figure 5.10A and Figure 5.11A.

In all cases, the "V-shape" of the misfit functions (with a null misfit at the reference Moho depth) (Figure 5.9, top) for either Rayleigh- or Love-waves show that the depth of the Moho is confidently resolvable with either phase or group velocities. A variation of only few kilometers in the Moho depth generates detectable ( $>1\%$ ) perturbations of phase- and group-velocities (Figure 5.9 middle- and lower-frames, Figure 5.10D-E, Figure 5.11D-E).



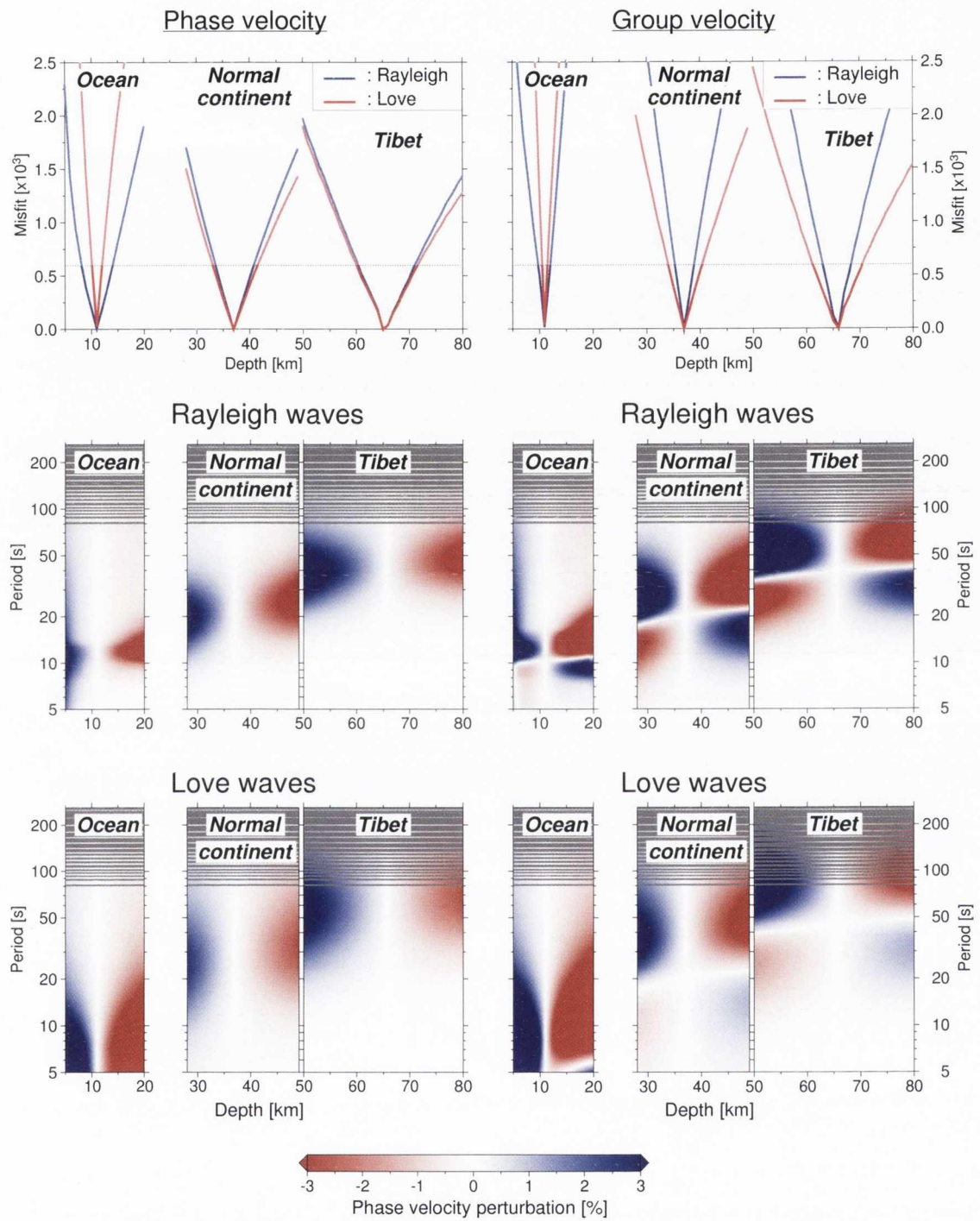


Figure 5.9.: Sensitivity of the phase and group velocities of fundamental-mode surface-waves to the depth of the Moho for each Earth-models (Oceanic, Continental and Tibetan). The reference Moho depth is 11 km for the oceanic model, 37 km for the continental model and 65 km for the tibetan model. The depth of the Moho of each model varies and the difference between the modified model and the reference is computed and represented by misfit values (Equation 5.2): top ; or the phase-velocity perturbation (Equation 5.3) for Rayleigh and Love waves: middle and bottom respectively.

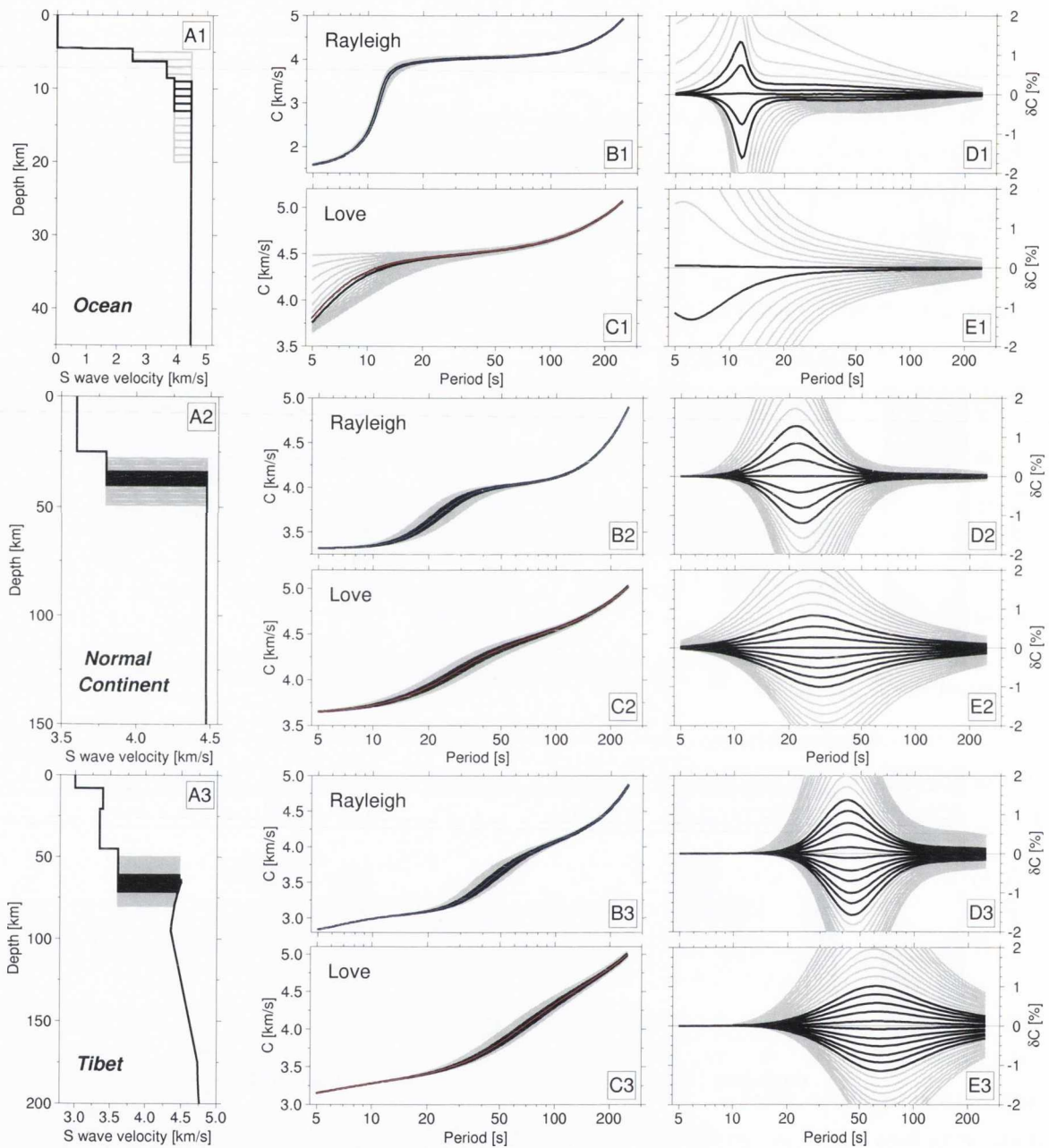


Figure 5.10.: Sensitivity of surface-waves' phase velocity to the Moho depth for each of the three Earth-models (Oceanic:1, Continental:2 and Tibetan:3). The depth of the Moho varies as shown by the grey and black lines representing the 1-D  $V_s$  profiles (A). The corresponding dispersion curves for Rayleigh and Love waves are shown in frames B and C respectively. The phase-velocity perturbations are presented in frames D and E. Models with a misfit value lower than  $6 \cdot 10^{-4}$  are represented with black lines (see Figure 5.9, top frame).

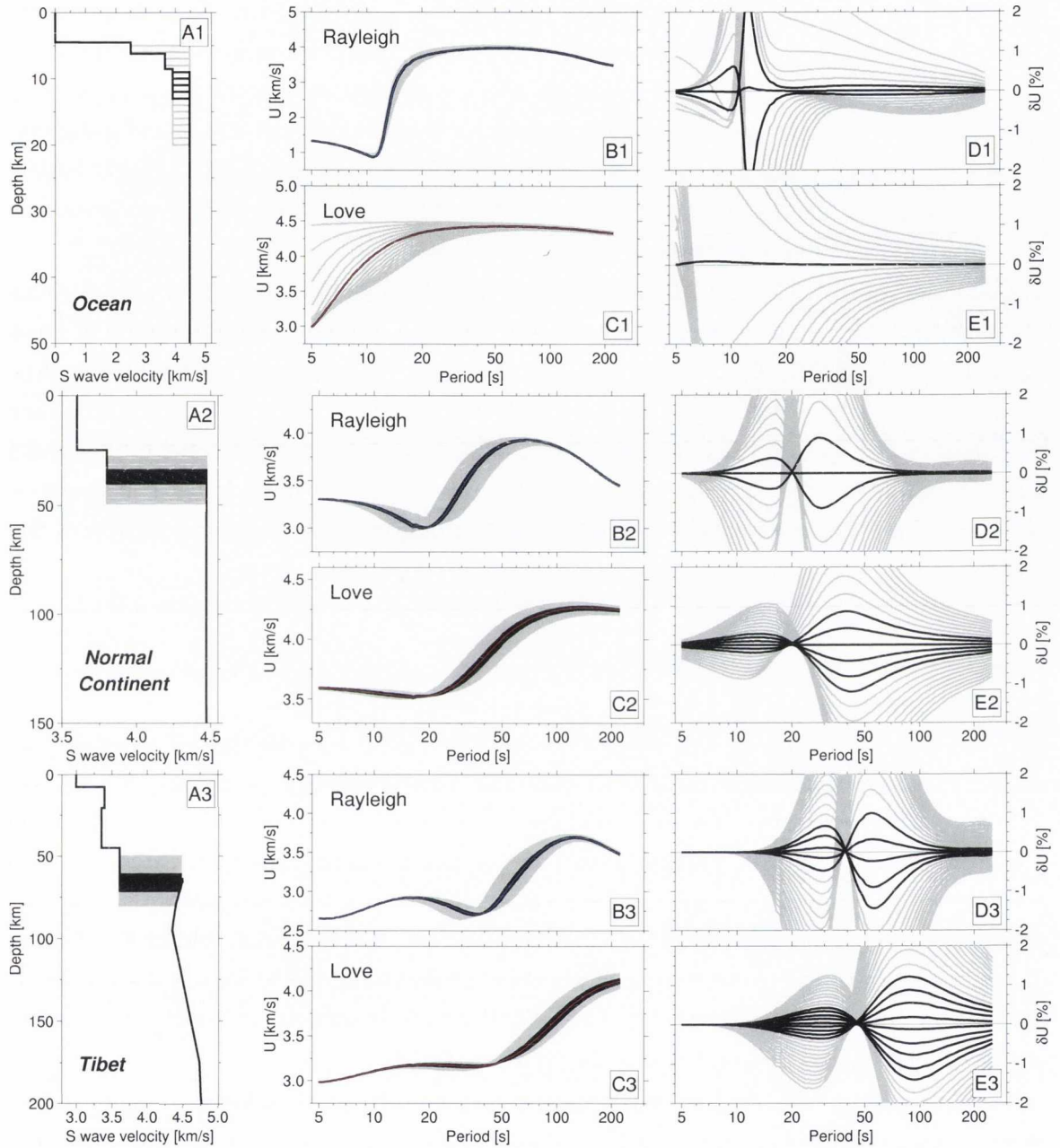


Figure 5.11.: Sensitivity of surface-waves' group velocity to the Moho depth for each of the three Earth-models (Oceanic:1, Continental:2 and Tibetan:3). The depth of the Moho varies as shown by the grey and black lines representing the 1-D Vs profiles (A). The corresponding dispersion curves for Rayleigh and Love waves are shown in frames B and C respectively. The phase-velocity perturbations are presented in frames D and E. Models with a misfit value lower than  $6.10^{-4}$  are represented with black lines (see Figure 5.9, top frame).



The deeper the Moho, the higher the periods at which the surface-wave is the most sensitive to the Moho (Figure 5.9 middle- and lower-frames, Figure 5.10 D-E, Figure 5.11 D-E). Indeed, surface-waves are sensitive to broader depth-ranges with increasing periods (Figure 2.6, Figure C.1). The period-range of sensitivity of the Love-wave is broader compared to the sensitivity of the Rayleigh-wave and is also explained by the depth sensitivity of surface-waves (Figure 2.6, Figure C.1). While, for a similar period, phase and group velocities are sensitive to the same depth range, group velocity sensitivity kernels have a zero and negative sensitivity at high depths. This negative sign of the sensitivity generates an additional node in the phase-velocity perturbations.

Phase and group velocities from both continents show similar sensitivity to the Moho. However, the period-range of the sensitivity of surface-waves to the Tibetan thick crust is broader compared to the period-range of sensitivity to the normal continent. The low seismic velocity above the Moho (see Yang *et al.*, 2012; Agius and Lebedev, 2010) amplifies the signal of the discontinuity. The sensitivity of surface-waves to an oceanic Earth-model differs to some extent from the sensitivity to a continental Earth-models. The presence of water in the oceanic model (Table 5.1) has a strong effect on the propagation of the surface-waves.

#### 5.3.4 Sensitivity and resolution with inversion

Furthermore, an inversion technique is used to not only investigate the sensitivity of surface-waves to the Moho but to also take into account the trade-offs between the parameters. The procedure used is similar to Bartzsch *et al.* (2011). During the inversion procedure, the depth of the Moho is fixed at a range of values in a grid search. At each point of the grid, a separate gradient search is performed and the algorithm adjusts the other parameters (*e.g.* the S-wave velocity in the crust or the mantle, the depth of other discontinuities...) so as to compensate the impact of the change of Moho depth as much as possible and find the Earth-model that best fits the data. Inversions of the phase velocities are performed for each curves with varying Moho depths.

1-D Earth models are obtained from phase velocities with an iterative non-linear gradient-search using a Gauss-Newton algorithm (Meier *et al.*, 2004; Lebedev *et al.*, 2006; Erduran *et al.*, 2008). During the inversion, the S-wave velocity is the only parameters that vary. The P-wave velocity is then adjusted so as to keep the same  $V_p/V_s$  ratio.

The parameters of the model are averaged over basis-functions with box-cars and triangles as described by Nolet (1990). The depth of the discontinuities in the crust (defining the discontinuity between the upper- and middle-crust, the middle- and lower-crust and the lower-crust and upper-mantle *i.e.* the Moho) are also parameters of the inversion and are allowed to vary. In this study, the models have two or three discontinuities and are parameterized with 9 or 10 basis-functions. The crust is parameterized with box-cars and

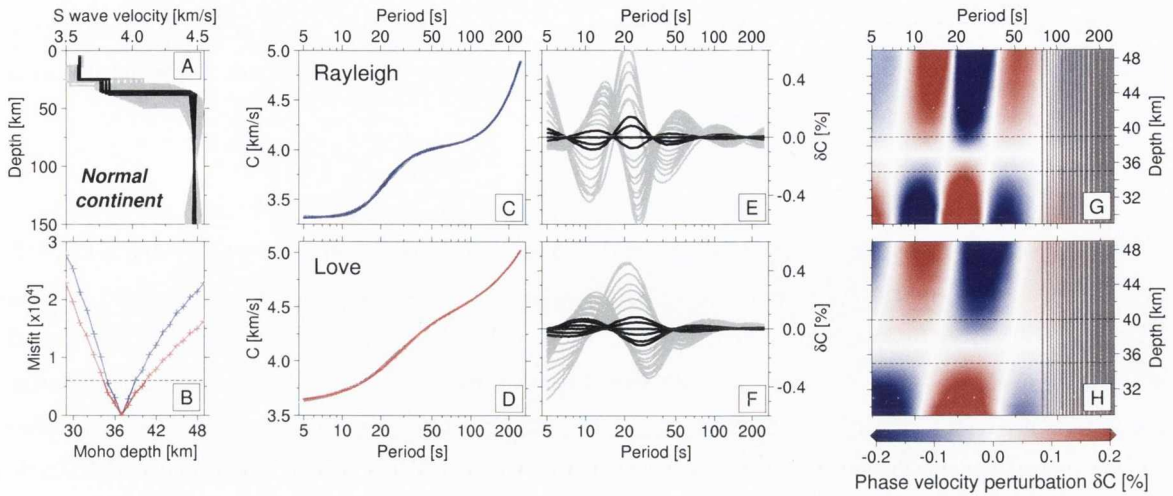


Figure 5.12.: 1-D model-space exploration by inversion of synthetic phase-velocity curves for the continental Earth-model. Inversions of Rayleigh (C,E,G) and Love (D,F,H) waves were performed separately. The model-space was explored by means of a uniform sampling of the Moho-depth axis, with a non-linear, gradient-search inversion at each point. In each of the inversion, the depth of the Moho is fixed but the crustal and mantle structures are allowed to vary to allow trade-offs between the parameters. A: 1-D shear-velocity profiles resulting from each inversion. B: misfit values resulting from the inversions. C (D): Phase-velocity curves. The blue (red) curves are the reference curves for Rayleigh (Love) waves. Behind, in grey or black, are the dispersion curves corresponding to the profiles in A. E (F): Differences between the best-fitting curves and the reference curves for Rayleigh (Love) wave. Grey and black lines are the curves corresponding to the shear-velocity profiles in A. The same information is presented in G (H) using another format. Vertical dashed lines in frames G and H represent the Moho depth-ranges to which the misfit values are smaller than  $6.10^{-3}$  (B).

the mantle is parametrized with triangles. The number of basis-function should be large enough not to affect the results.

The vector minimized by the optimization function is defined by the difference between the theoretical phase-velocity model resulting from the inversion procedure and the data. The misfit is then normalized with the standard deviation or estimated error of the measured phase-velocities (set to  $0.05 \text{ km.s}^{-1}$  for synthetic data). Damping coefficients for each of the parameters are added to the minimisation vector to control the smoothness of the resulting model. The gradient-search converges to the best-fitting model (Erduran *et al.*, 2008) by minimizing the misfit vector.

In this chapter, Rayleigh and Love dispersion curves are inverted independently as its main purpose is to resolve the sensitivity of each surface-wave type to varying Moho depths. Using this algorithm, Rayleigh and Love dispersion curves can, however, be inverted together, to also resolve and take into account radial anisotropy.

We apply the gradient-search method on the three synthetic Earth-models (oceanic, normal continent and Tibet) and vary the depth of the Moho. Similarly to results of the



sensitivity tests of Section 5.3.3 the misfit curves (Figure 5.12B) show the same singular "V-shape" with a minimum value at the correct Moho-depth. The depth of the Moho can thus be resolved with the inversion.

The misfit values (Figure 5.12B) and the amplitude of the phase-velocity perturbations are much smaller (by an order of magnitude) and the dispersion curves (Figure 5.12C-D) are much closer to the reference curve (compared to Figure 5.10B-C). The reason is that the gradient-search compensates incorrect Moho depths with variations in crustal and mantle seismic velocity structures, with the depth of the Moho and the crustal and mantle structure showing strong trade-offs (see Section 5.4.1). The effect of the trade-offs can also be observed in the phase-velocity perturbation curves. When the depth of the Moho is the only parameter that varies, the phase-velocity perturbation show one clear extremum (Figure 5.10D-E). However, when the effect of the Moho-depth perturbation is partially compensated, the phase-velocity perturbations (Figure 5.12 E-F) show oscillations with several extrema as the algorithm compensates for the Moho variations with changes in the shallower and deeper structures to which the surface-wave is sensitive at shorter and longer periods.

Not only can the inversion resolve the depth of the Moho accurately, but it also enables one to investigate the effects of the trade-offs between the parameters of the inversion.

#### 5.4 TRADE-OFFS BETWEEN THE PARAMETERS

During the inversion, changes of the Moho depth trade off with other parameters of the model. In order to map the trade-offs between the parameters, we perform a grid-search across slices of the model-space: lower/middle-crust velocity *vs.* Moho depth ; upper-mantle velocity *vs.* Moho depth ; Moho thickness *vs.* Moho depth. The results of this model-space exploration will help us to better understand the relation between the parameters and identify any possible correlation between them. The model-space is explored using forward modelling with the phase-velocities and group-velocities of a normal continent. Similar results are expected with the oceanic or Tibetan dispersion curves.

##### 5.4.1 *Generation of Earth models*

This section presents the method used to generate the models computed at every point across the two-parameter planes. In the first part, we explore the correlation between the velocity of the S-waves in the middle and lower-crust and the depth of the Moho discontinuity. Note that the velocity of the P-wave is altered depending on the velocity



of the S-wave so that the  $V_p/V_s$  ratio remains constant. In the second part, we explore the correlation between the S-wave velocity of the first 100 kilometers of the upper-mantle and the depth of the Moho discontinuity. In the last part, we examine the resolving power of surface-waves to the thickness of the Moho discontinuity by examining the trade-offs between the Moho thickness and the Moho depth.

#### 5.4.1.1 *Wavespeed variations in the middle and lower-crust*

We investigate the relation between the Moho depth and the velocity in the middle and lower crust using a grid-search method. Earth models are created so the depth of the Moho and the S-wave velocity (and consequently the P-wave velocity) in the middle and lower-crust vary. For each couple of parameters, the misfit (Equation 5.2) is computed and their distribution informs about the trade-offs between the parameters. We first adjust the depth of the Moho as presented in Section 5.3.2. S-wave velocity of the middle and lower-crust is then modified by adding or subtracting the desired velocity variation (Figure 5.13 A-D). The Moho depth varies between 28 and 60 km and the variation of the S-wave velocity in the middle and lower-crust varies between  $-0.25$  and  $0.25 \text{ km.s}^{-1}$ . The P-wave velocity in the middle crust varies then between  $-0.444$  and  $0.444 \text{ km.s}^{-1}$  and between  $-0.447$  and  $0.447 \text{ km.s}^{-1}$  in the lower crust

#### 5.4.1.2 *Wavespeed variation in the upper-mantle*

To investigate the relation between the Moho depth and the velocity of the first 100's kilometers of the upper-mantle, we follow the same procedure as in Section 5.4.1.1. The S-wave velocity in the upper-mantle varies between  $-0.25$  and  $0.25 \text{ km.s}^{-1}$  and corresponds to a variation of P-wave velocity of  $-0.448$  and  $0.448 \text{ km.s}^{-1}$  (Figure 5.13 E, F, G and H).

#### 5.4.1.3 *Thickness of the Moho discontinuity*

To explore the trade-offs between the depth and the thickness of the Moho discontinuity, we define the thickness of the Moho as a linear transition between the parameters of the lower-crust and the parameters of the upper-mantle. We first adjust the depth of the discontinuity at the desired depth (Section 5.3.2). The top of the discontinuity is defined at half of the thickness atop the depth of the Moho discontinuity and the bottom is set at half of the thickness below it.

To avoid extreme transitions between the parameters of the crust and the mantle, the top of the Moho can not be shallower than the top of the lower-crust. We only consider Mohos for which the top of the discontinuity is within the lower-crust (for example, in Figure 5.13 A<sub>3</sub> and B<sub>3</sub>, the thickness of the Moho is restricted as the depth of the discontinuity is shallow and close to the bottom of the middle-crust). In the case of a shallow and/or thin Moho discontinuity, the parameters are adjusted so that the bottom of the Moho has

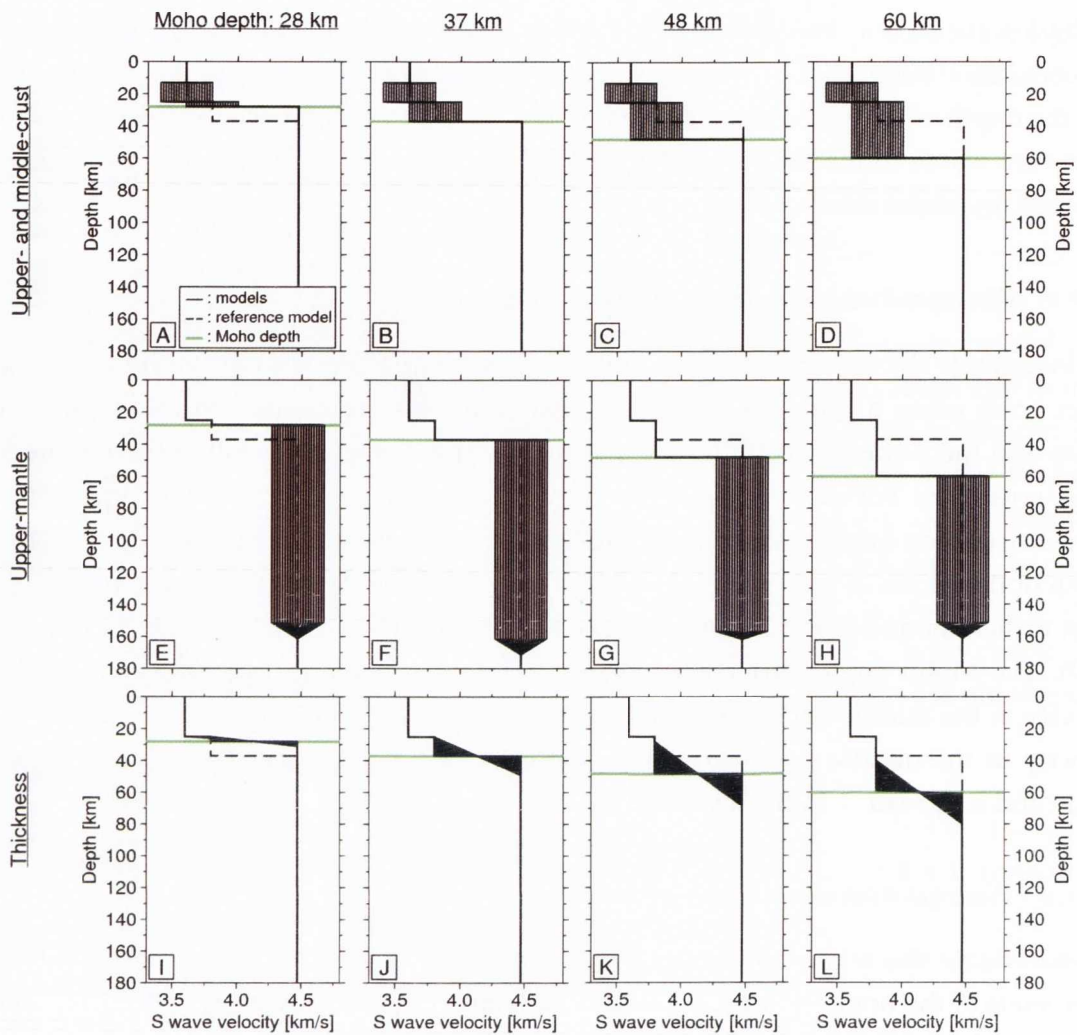


Figure 5.13.: Samples of 1-D synthetic shear-velocity profiles used to explore the trade-offs between the depth of the Moho and the shear-velocity in the upper- and middle-crust (first line), upper-mantle (middle line) and between the depth of the Moho and its thickness (bottom line). Examples at three different depths of the Moho: 29 km (first column), 37 km (second column), 48 km (third column) and 60 km (forth column). The dashed line represents the true synthetic model (unchanged in all the frames). The horizontal green lines represent the depth at which the Moho has been modified. The continuous black lines represent the model with both its Moho depth and shear-velocities altered.

upper-mantle elastic parameters. For example, the [Figure 5.13 A3](#) presents all the models that have a Moho of 29 km deep (solid black lines) and because the bottom of the middle-crust is at 25 km, the Moho can not be thicker than 4 km otherwise, the top of the Moho would be in the middle-crust. In that configuration, the bottom of the Moho discontinuity is shallower than the reference model (horizontal dashed line) but the parameters ( $V_p$ ,  $V_s$ ,  $\rho$  and  $Q$ ) below the Moho are the upper-mantle parameters. With the depths of the top and the bottom of the Moho discontinuity defined, the parameters: S-wave velocity, P-wave velocity, density and quality factors in between are determined with a linear function.



Let  $P_i$  be a parameter at the depth  $i$  between depth  $t$  at the top of the Moho and depth  $b$  at the bottom of the Moho.  $P_i$  is defined by:

$$P_i = P_t + (P_b - P_t) \frac{Z_i - Z_b}{Z_t - Z_b}, \quad (5.4)$$

where  $Z_i$  is the depth of the point,  $Z_t$  is the depth of the top of the Moho discontinuity and  $Z_b$  is the depth at the bottom of the Moho discontinuity.

#### 5.4.2 Results

Trade-offs are expected between the Moho-depth and the shear-wave velocity in the crust and the mantle. Indeed, these shear-velocity perturbations generate phase-velocity and group-velocity perturbations in the same period-range as perturbations due to the Moho-depth variation (Figure 2.6, Figure C.1). Imaged with group velocities, trade-offs between the depth of the Moho and shear velocities in the lower-crust or upper-mantle are weaker than with phase velocities. Despite that trade-offs remain, shear velocity structures above and below the Moho appear to have less effects on the resolution of the Moho-depth.

The trade-offs between the Moho-depth and the shear-speed in the lower-crust and upper-mantle are presented in Figure 5.14, top and middle frames, respectively, for Rayleigh (left column) and Love (right column) waves. They show similar results *i.e.* when estimating the Moho-depth, an increase (decrease) of its depth can be compensated to a large extent by an increase (decrease) of the shear-speed in the lower-crust or the upper-mantle in the model.

An erroneous estimate of the Moho depth can not be compensated by a thicker (or thinner) Moho. Indeed, the bottom line of the Figure 5.14 shows that the trade-offs between the Moho-depth and its thickness are weak. Neither Rayleigh nor Love surface-waves can resolve the thickness of the Moho. Joint analysis with other data (*e.g.* receiver functions, see Endrun *et al.*, 2004; Julia *et al.*, 2000; Lebedev *et al.*, 2002b,a; Shen *et al.*, 2012; Tkalčić *et al.*, 2012) can, however, enable to resolve the depth and the thickness of the Moho discontinuity with higher resolution.

### 5.5 APPLICATION TO REAL DATA - SOUTHERN AFRICA

We have shown that surface-wave phase-velocities can resolve the depth of the Moho. In order to investigate this further, we invert measurements from southern Africa and test the resolution of the depth of the Moho with real data.



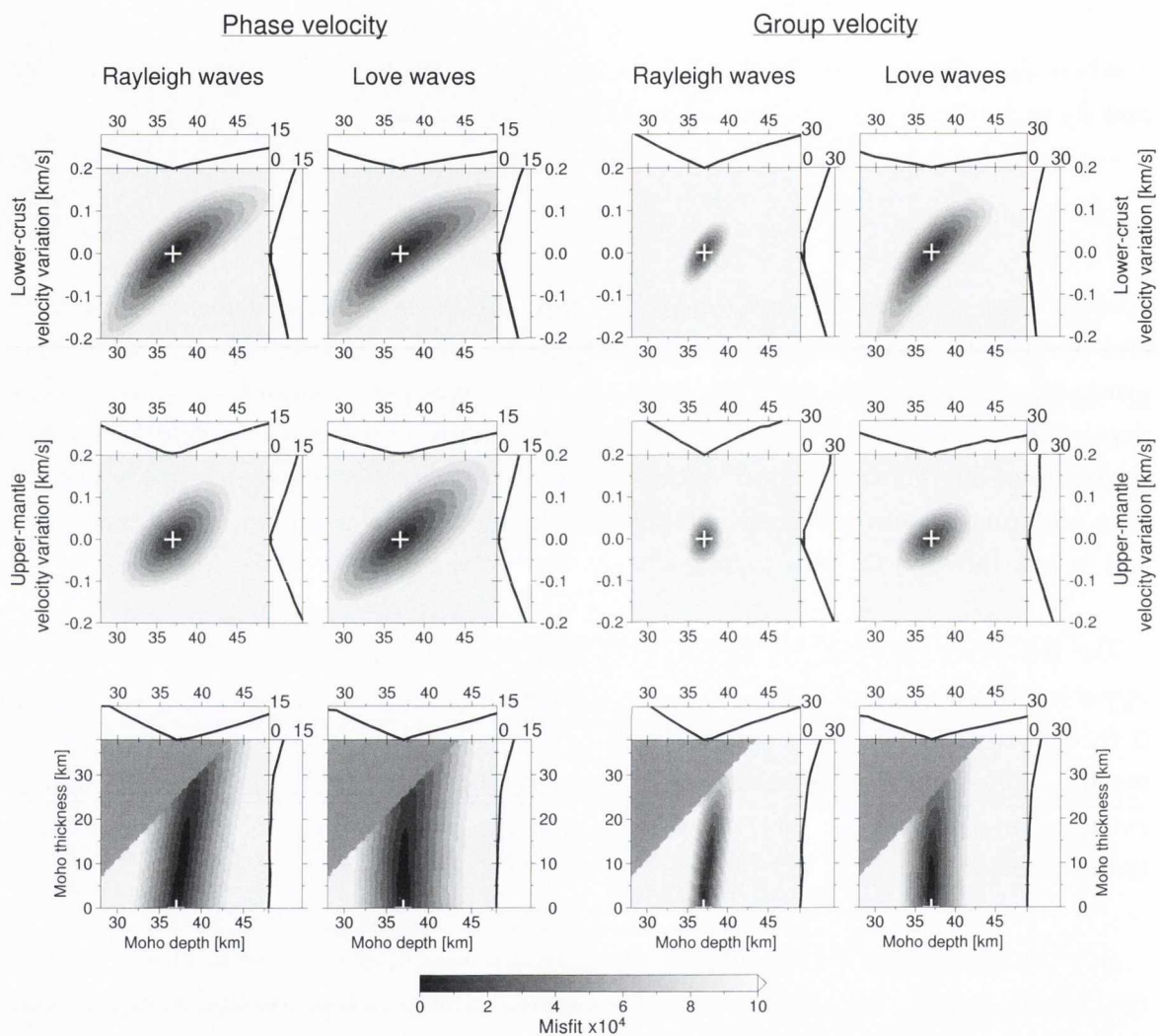


Figure 5.14.: Trade-offs between the Moho depth and other seismic parameters: upper- and lower-crust (top), upper-mantle (middle) and Moho thickness (bottom). Trade-offs are computed for both Rayleigh (left) and Love (right) waves independently. For each couple, the model is computed following [Section 5.4.1](#) and [Figure 5.13](#). The white cross in each frame shows the reference model with a Moho at 37 km depth and thickness of 0 km. Smaller frames attached above (on the right) of each misfit frames represent the minimum misfit-values along the y-(x)-axis.

### 5.5.1 Data acquisition

We use Rayleigh-wave phase-velocity dispersion curves measured in the Limpopo Belt and three regions in the Kaapvaal Craton in Southern Africa (Figure 3.2). Each curve is an average of thousands of inter-station measurements across regions in which the Moho-depth is known to be nearly homogeneous, with variations of only few kilometers (Nguuri *et al.*, 2001; Nair *et al.*, 2006; Yang *et al.*, 2008; Kgaswane *et al.*, 2009). Investigating the average depth of the Moho in such regions is thus meaningful.

In Section 3.3.3, we also present isotropic-average Rayleigh dispersion curves across each region which do not depend on any azimuthal variation of phase-velocities. Both simple-regional average and isotropic-average dispersion curves span very broad period-ranges, are robust and accurate. Both curves should yield similar results regarding the Moho-depth investigation. The following analysis will be applied independently to the simple regional-average and the isotropic-average Rayleigh dispersion curves of each of the four regions.

We also perform further inversion of modified versions of each dispersion curve. The first one is a smoothed version of each dispersion curve, computed to reduce the effect of random noise in the data. To smooth the dispersion curve, we use an over-parameterised gradient-search inversion (presented in Section 5.3.4) and under-damp the parameters. The resulting model fits the data closely, excluding period-dependent random errors that are inconsistent with any plausible Earth-models. The results of the smoothed-data inversions should be consistent with the results of the original dispersion-curve (*e.g.* Deschamps *et al.*, 2008b; Endrun *et al.*, 2011) inversion, although the smoothing operation may, in principle, introduce new biases into the data (see Appendix D for more information).

The second modified version of the dispersion curve is generated by shortening its period-range. Figure 5.9 and Figure 5.10 show that the period-range at which the Rayleigh-wave is sensitive to the Moho is identifiable. We then invert only the period-band of the dispersion curve that is sensitive to the Moho. This enables to remove measurements errors at periods less sensitive to the Moho-depths but could still bring errors during the inversion. This narrow-band dispersion curve has, however, limited sensitivity and an accurate reference model of the crust and the mantle must be used in order to robustly recover the information sought.



### 5.5.2 Results

Previous studies using seismic ambient-noise and teleseismic tomography (Yang *et al.*, 2008), body-wave Pn arrival (Kwadiba *et al.*, 2003), receiver functions (Nguuri *et al.*, 2001; Stankiewicz *et al.*, 2002; James *et al.*, 2003) and joint inversion of receiver function and teleseismic surface-waves (Kgaswane *et al.*, 2009) used measurements from the SASE array and the Kimberley array in southern Africa to *i.a* determine the crustal thickness. Across the region, they found that the depth of the Moho varies between 33 to 50 km. We then expect the Rayleigh-wave to be sensitive to the Moho between ~15 and 35 s (Figure 2.6, Figure C.1) and use these limits as the period-range of the narrow-band Rayleigh-wave dispersion curves.

Figure 5.15, Figure 5.16, Figure 5.17 and Figure 5.18 present the results of both simple regional-average and isotropic regional-average for the three different inversions (original-data, smoothed and narrow-band Rayleigh-wave phase-velocity dispersion curves) from, respectively, the Limpopo Belt, North, South-East and South-West Kaapvaal Craton regions.

Results from Figure 5.15, Figure 5.16, Figure 5.17 and Figure 5.18 show similarities, as expected, with the results of the synthetic inversions for a normal continent (Figure 5.12). The "V-shaped" of the misfit curves are relatively smooth due to noise in the data, but one steep valley with a local minimum can still be clearly identified (at the exception of the misfit curve from the inversion with a smoothed isotropic-average dispersion curves from South-West Kaapvaal Craton region). The frames presenting the phase velocity perturbation as a function of Moho depth and period (Figure 5.15, Figure 5.16, Figure 5.17, Figure 5.18 I,J,K) show the same pattern of period-dependent positive and negative perturbations as seen in the synthetic tests (Figure 5.12 G,H) but with distortions, due to the noise in the measurements.

Smoothing the dispersion curve reduces the misfit values (frames B) and the phase velocity perturbation patterns are more similar to those in synthetic tests (frames F,G,H,I,J,K). Figure 5.19 presents synthetic tests for Moho-depth estimation with synthetic data to which noise has been added (for completeness, see Appendix D). The "high-frequency" noise (Figure 5.19B) is the residual between the Limpopo Belt's simple-average phase velocities and the smooth version of this same curve. Smoothing the dispersion curve does not indeed alter the resolvability of the Moho-depth and does reduce the amplitude of the misfit values. The "complete estimated noise" (Figure 5.19C) is computed as the residual between the Limpopo Belt's simple-average phase velocities and the phase velocities resulting from an inversion that best fits the data with a realistic 1-D Earth model. We would then define this noise as the full noise present in the data. To examine the effect



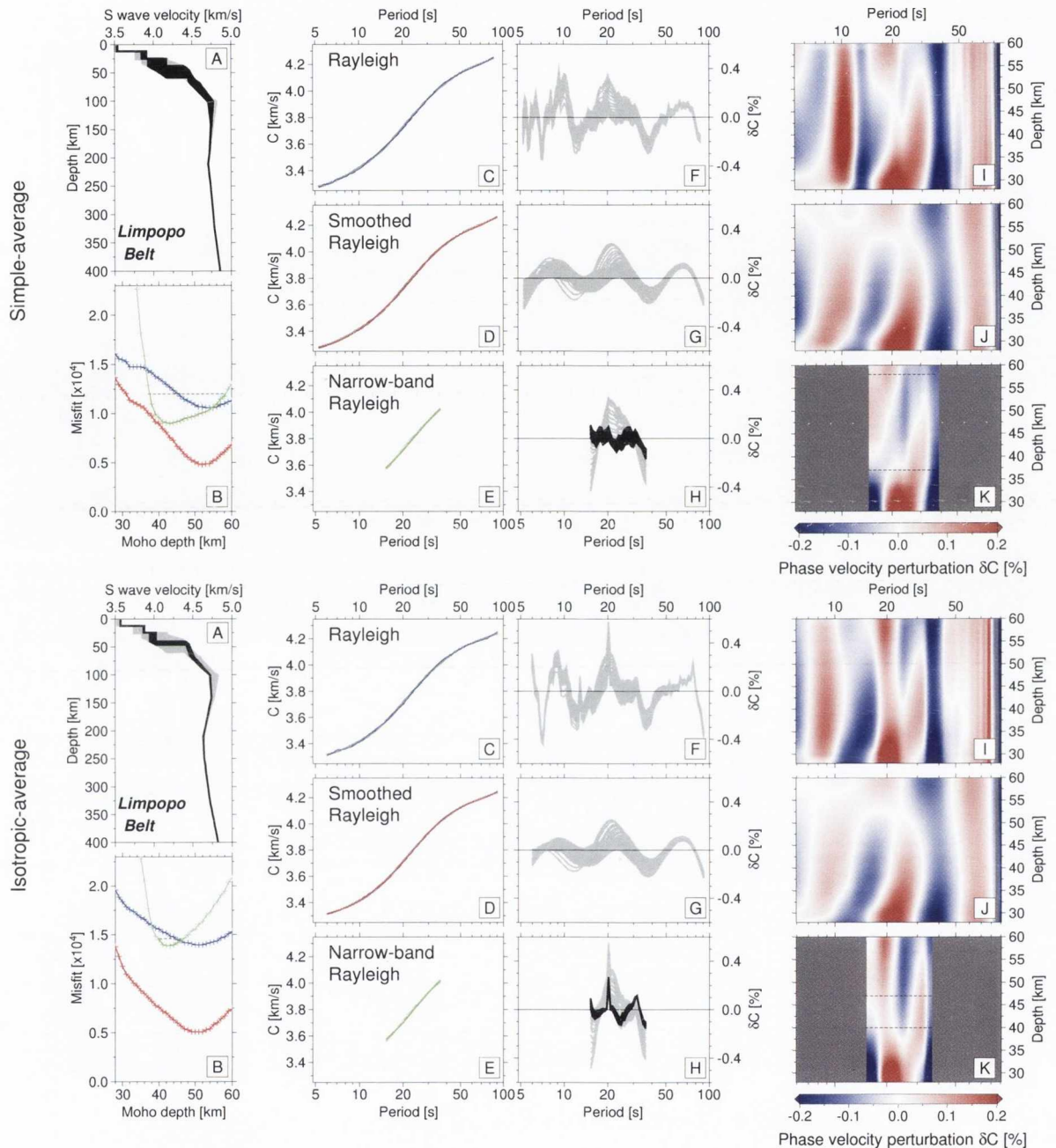


Figure 5.15.: Moho-depth resolution of the Limpopo Belt region using simple-regional average (top) or isotropic-average (bottom) Rayleigh-wave dispersion-curve. See Section 3.3.1 (Section 3.3.3) to see how the curves are measured. A: best-fitting shear-velocity profiles with Moho-depth varying between 28 km and 60 km. B: Misfit values resulting from inversion with the Moho-depth varying. In blue: broad-band dispersion curve (top row: C, F, I); in red: smoothed broad-band dispersion curve (middle-row: D, G, J); in green: narrow-band dispersion curve (bottom row: E, H, K). C, D, E: Rayleigh-wave dispersion curves (colored curves: data; grey and black curves behind the colored curves: dispersion curves corresponding to the models in A). F, G, H: Phase-velocity perturbations between the best-fitting models (colored curves in C, D, E) and the models. The same information is presented in I, J, K using an other format.

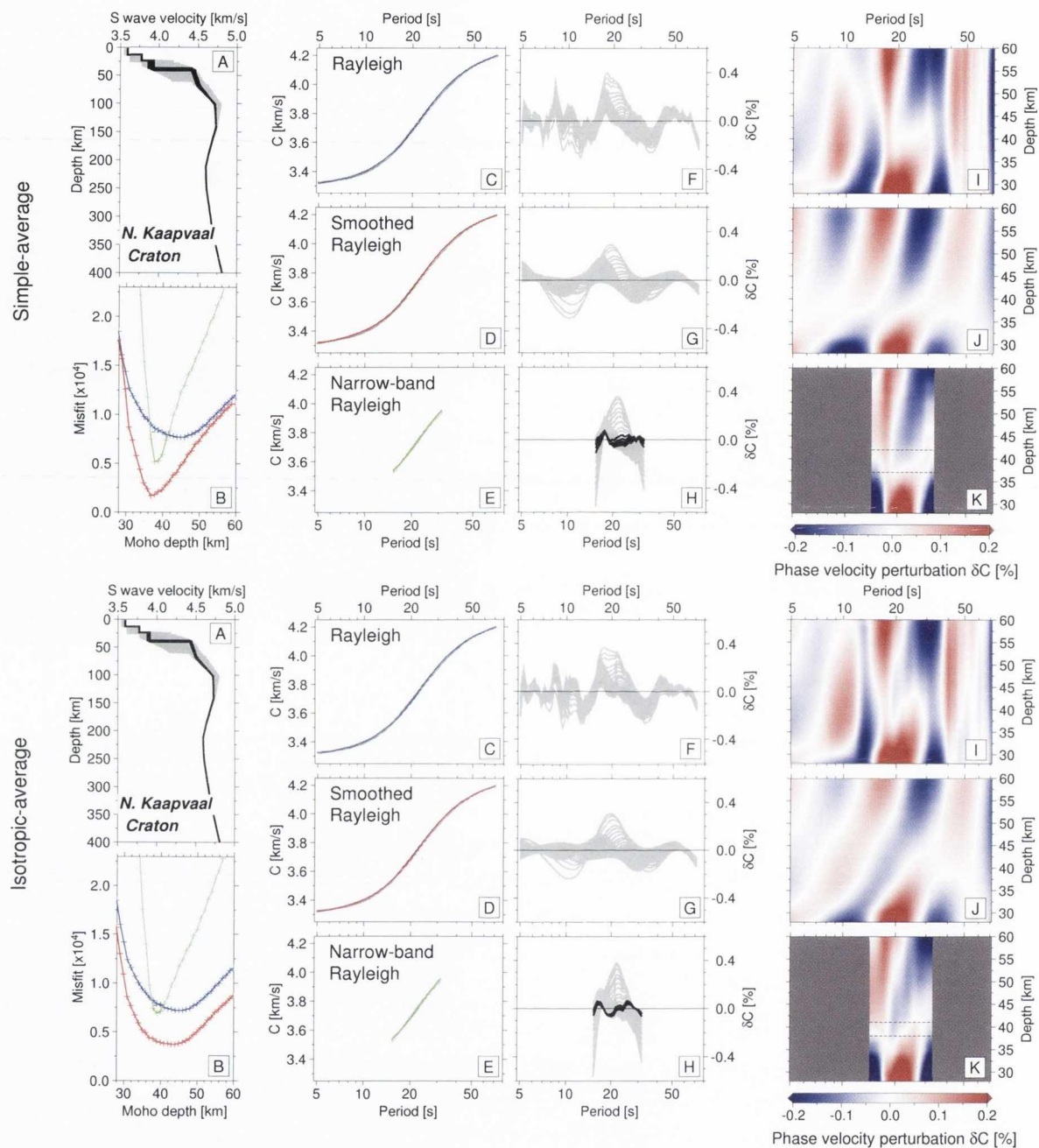


Figure 5.16.: Moho-depth resolution of the North Kaapvaal Craton region using simple-regional average (top) or isotropic-average (bottom) Rayleigh-wave dispersion-curve. See Section 3.3.1 (Section 3.3.3) to see how the curves are measured. A: best-fitting shear-velocity profiles with Moho-depth varying between 28 km and 60 km. B: Misfit values resulting from inversion with the Moho-depth varying. In blue: broad-band dispersion curve (top row: C, F, I); in red: smoothed broad-band dispersion curve (middle-row: D, G, J); in green: narrow-band dispersion curve (bottom row: E, H, K). C, D, E: Rayleigh-wave dispersion curves (colored curves: data; grey and black curves behind the colored curves: dispersion curves corresponding to the models in A). F, G, H: Phase-velocity perturbations between the best-fitting models (colored curves in C, D, E) and the models. The same information is presented in I, J, K using an other format.



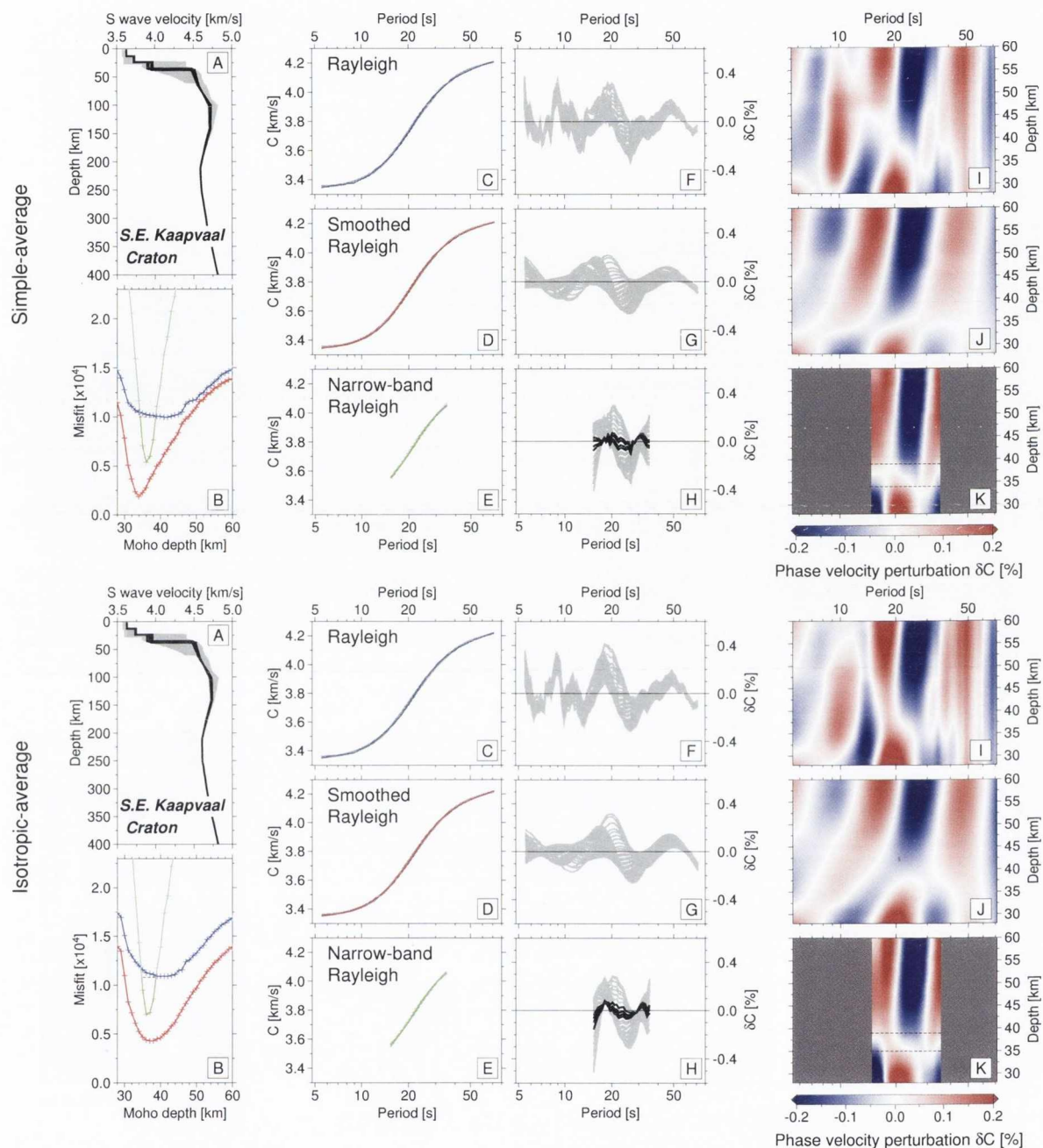


Figure 5.17.: Moho-depth resolution of the S-E Kaapvaal Craton region using simple-regional average (top) or isotropic-average (bottom) Rayleigh-wave dispersion-curve. See Section 3.3.1 (Section 3.3.3) to see how the curves are measured. A: best-fitting shear-velocity profiles with Moho-depth varying between 28 km and 60 km. B: Misfit values resulting from inversion with the Moho-depth varying. In blue: broad-band dispersion curve (top row: C, F, I); in red: smoothed broad-band dispersion curve (middle-row: D, G, J); in green: narrow-band dispersion curve (bottom row: E, H, K). C, D, E: Rayleigh-wave dispersion curves (colored curves: data; grey and black curves behind the colored curves: dispersion curves corresponding to the models in A). F, G, H: Phase-velocity perturbations between the best-fitting models (colored curves in C, D, E) and the models. The same information is presented in I, J, K using an other format.



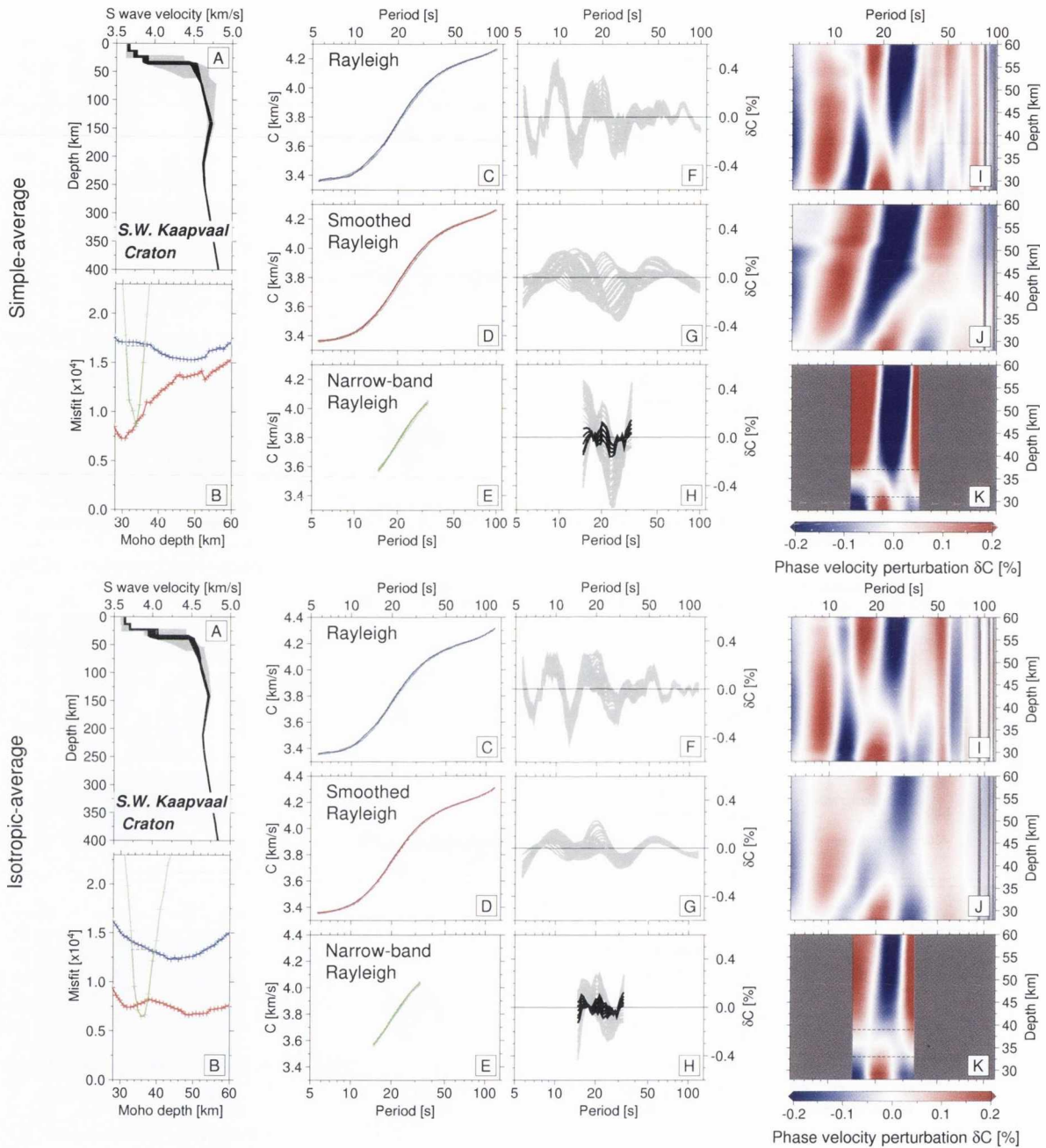


Figure 5.18.: Moho-depth resolution of the S-W Kaapvaal Craton region using simple-regional average (top) or isotropic-average (bottom) Rayleigh-wave dispersion-curve. See Section 3.3.1 (Section 3.3.3) to see how the curves are measured. A: best-fitting shear-velocity profiles with Moho-depth varying between 28 km and 60 km. B: Misfit values resulting from inversion with the Moho-depth varying. In blue: broad-band dispersion curve (top row: C, F, I); in red: smoothed broad-band dispersion curve (middle-row: D, G, J); in green: narrow-band dispersion curve (bottom row: E, H, K). C, D, E: Rayleigh-wave dispersion curves (colored curves: data; grey and black curves behind the colored curves: dispersion curves corresponding to the models in A). F, G, H: Phase-velocity perturbations between the best-fitting models (colored curves in C, D, E) and the models. The same information is presented in I, J, K using an other format.

of smoothing, we evaluate the residual noise after smoothing with a simple "ramp" noise model. The comparison between the results of the curves with "complete estimated" noise and "ramp" noise reveals the effect of the component of the "complete" noise that is similar to the ramp function.

However, even if the smoothing reduces the amplitude of the misfit values, the width of the valley remains unchanged and the Moho-depth resolution is not improved. To improve the resolution of the Moho-depth, we therefore invert a narrow-band dispersion curve sampling the periods most sensitive to the depth of the Moho. Indeed, the misfit curves (Figure 5.15, Figure 5.16, Figure 5.17 and Figure 5.18) show a local minimum with steeper slopes which lead to a better identifiable Moho-depth.

The data of the Limpopo Belt region seems to be very noisy at 40 s and the narrow-band curve do not take this period into consideration. The noise at that period may have affected the results of the inversion. The effects of this noise is noticeable since the Moho-depth would be determined 10 km too deep (Figure 5.19C,G). Similarly, results using smoothed simple-average or isotropic-average dispersion curves from the South-West region of the Kaapvaal Craton (Figure 5.18) are not consistent in this region, with consistency also lacking between the results of inversions of broad-band and narrow-band dispersion curves. The smoothing procedure may have introduced biases in the data. Indeed, the level of broad-band noise in the data seems to be high (Figure 5.18F) (especially at periods lower than 17 s) and may alter the smooth curve. Inversions with the narrow-band curve are less affected by this noise and provide results that are more consistent with previous studies (Nguuri *et al.*, 2001; Nair *et al.*, 2006; Yang *et al.*, 2008; Kgaswane *et al.*, 2009). These tend to confirm that the narrow-band dispersion curves are the most accurate data to resolve the Moho.

Within each region, the results using simple regional-average or isotropic regional-average Rayleigh dispersion curves are mutually consistent. Discrepancies are observed, however, in the results of the narrow-band curves from the Limpopo Belt region. The isotropic curve seems to better resolve the Moho-depth. It has been shown in Section 3.6.1 that up to 4% of azimuthal anisotropy is observed in the upper-crust, and whether or not this is taken into account may have an effect on the resolvability of the Moho.

We shall only discuss the results from the inversions using the narrow-band dispersion curves as they have been showed to be the most accurate. Results as depth-ranges of the Moho in each region (and for either simple-average or isotropic-average dispersion curves) are represented by dashed lines in the frames B and K of the figures Figure 5.15, Figure 5.16, Figure 5.17 and Figure 5.18. They are also summarized in the Table 5.4. The depth-range of the Moho is defined as the Moho-depths at which the misfit values are smaller than a threshold value (represented by the horizontal dashed line in frames A). This threshold corresponds to the misfit value from the inversion for the best fitting model



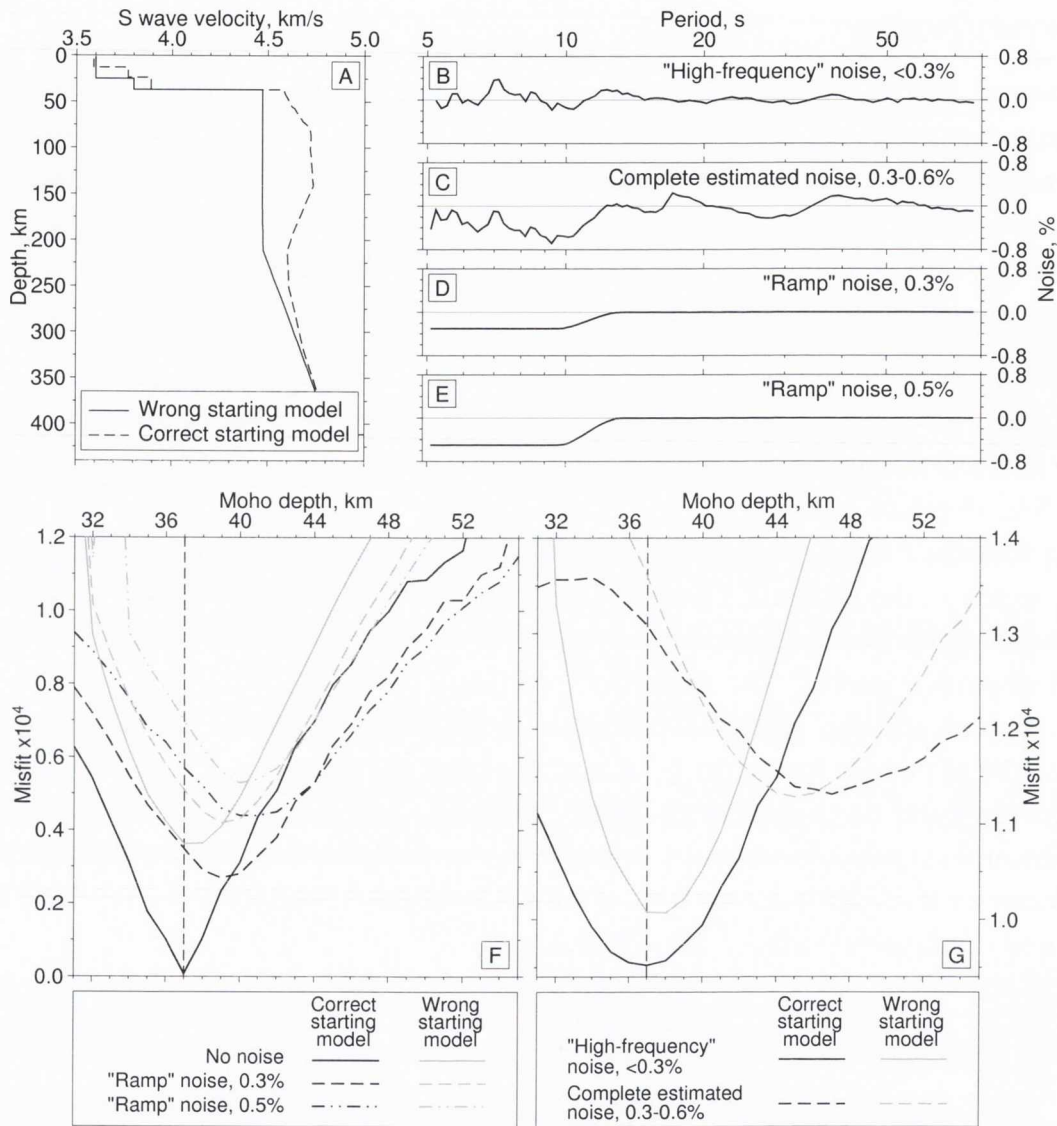


Figure 5.19.: Moho-depth resolution depending on reference Earth-model (true reference model or modified AK135 reference model (Kennett *et al.*, 1995)) or addition of noise. A: Shear velocity of the true reference model and the slightly inexact model. B,C,D,E: shear-velocity noise perturbation for B: "high-frequency" noise *i.e.* residual between the Limpopo Belt's simple-average phase velocities and the smooth version of this same curve; C: complete estimated noise *i.e.* residual between the Limpopo Belt's simple-average phase velocities and the phase velocities resulting from an inversion that best fits the data; D (E): "ramp" noise to simulate a period-dependant noise with maximum amplitude of -0.3 (-0.5) %. F,G: Misfit values. Dashed lines in frames F and G represent the correct Moho-depth.



of the broad-band simple average dispersion curve (frames C). The best-fitting model is found with an inversion with regular parameterisation and damping.

Table 5.4.: Depth-range of the Moho in each of the four sub-regions from simple-average or isotropic-average Rayleigh dispersion curves.

|                       | Simple-average | Isotropic-average |
|-----------------------|----------------|-------------------|
| Limpopo Belt          | 37 – 58 km     | 40 – 47 km        |
| North Kaapvaal Craton | 37 – 42 km     | 38 – 41 km        |
| S-E Kaapvaal Craton   | 34 – 39 km     | 35 – 39 km        |
| S-W Kaapvaal Craton   | 31 – 37 km     | 33 – 39 km        |

### 5.5.3 Discussion

Figure 5.20 presents estimations of the crustal thickness from Nguuri *et al.* (2001) (A), Nair *et al.* (2006) (B), Kwadiba *et al.* (2003) (C), Kgaswane *et al.* (2009) (D), Yang *et al.* (2008) (E) and an interpolated map of the results of this study (F). All maps show similar features with thin crust in the South and thick crust in the North with gradual increase in a North-Eastward direction. All studies overall agree on the Moho depths.

Moho depth inferred from receiver function studies by Nguuri *et al.* (2001), Nair *et al.* (2006) and Stankiewicz *et al.* (2002) (who show similar results to Nguuri *et al.* (2001)) show discrepancies with the results of this study (Figure 5.20 A,B,F). Receiver function measurements have their own uncertainties due to trade-offs between the Moho-depth and the crustal  $V_p/V_s$  ratios. Surface-waves studies have trade-offs between the Moho-depth and the shear velocities in the crust and the mantle. Joint analysis of receiver function and surface-wave data would help to reduce the uncertainties and better constrain the structure of the Moho. Kgaswane *et al.* (2009) inverted receiver-function data and Rayleigh-wave group-velocity (Figure 5.20 E) and present Moho-depth more consistent with the results of this study.

The Limpopo belt region appear to be the more structurally complex region (Figure 5.15). Nguuri *et al.* (2001) and Kgaswane *et al.* (2009) also observed complexity in their data around the Limpopo belt region. Using the receiver-function method, Nguuri *et al.* (2001) observed a complex distribution of the Moho depths within this region and discriminated the Central Zone, South and North Marginal Zones with distinct Moho-depths (see Chapter 1 for a more complete description of the Limpopo belt). Discrepancies between the Central Zone, the South and the North Marginal Zones of the Limpopo belt have

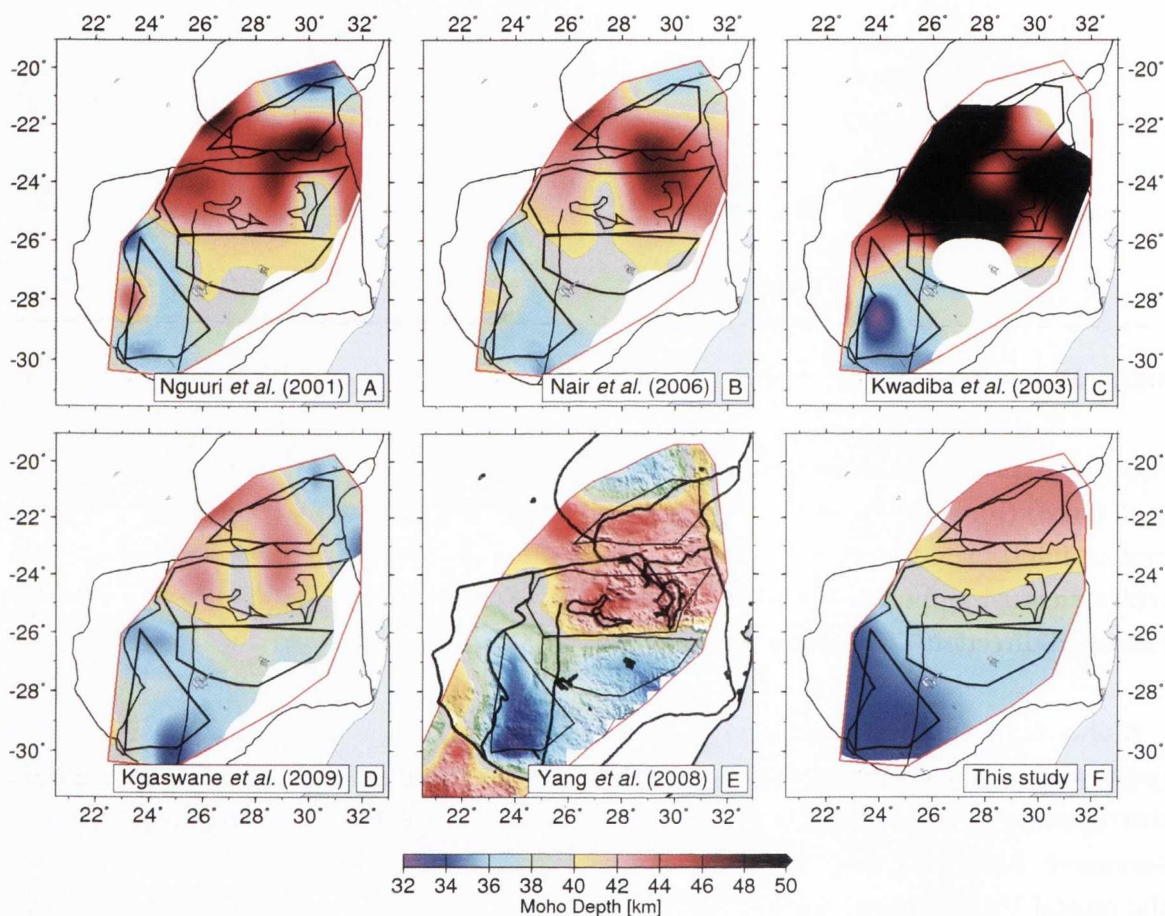


Figure 5.20.: Crustal thickness in southern Africa estimated by previous studies (A-E) and this study (F). A,B: Moho-depth inferred from receiver functions by respectively Nguuri *et al.* (2001) and Nair *et al.* (2006). C: Moho-depth inferred from Pn arrival times by Kwadiba *et al.* (2003). D: Moho-depth estimation from Kgaswane *et al.* (2009) by joint-inversion of receiver-functions and Rayleigh-wave group velocities. E: Moho-depth estimation from Yang *et al.* (2008) who inverted seismic ambient-noise and teleseismic surface wave data (figure modified from original). Black lines mark off tectonic structures and the four sub-regions (see Figure 3.1). Moho thicknesses only inside the red contours are shown for clarity.



also been observed by Kgaswane *et al.* (2009) who jointly inverted Rayleigh-wave group-velocities and receiver-function data. They define the Moho discontinuity as the depth at which the shear velocity is larger than 4.3 km/s. They observed that several stations in the Limpopo Belt show two increases of velocity and exceed 4.3 km/s with lower velocities between them. They did not investigate this further but that may indicate the presence of structural heterogeneities of the Moho in the Limpopo Belt region. Overall, all studies agreed that this region is a thick region with a Moho at a depth around 42–44 km.

To investigate the depth of the Moho discontinuity with surface-wave dispersion curves, we show that it is worthwhile to only invert the period-range of the dispersion curve that is the most sensitive to the depth of the Moho. Indeed, only a limited period-range is sensitive to the Moho and even small errors at short periods (sensitive to the upper layers of the crust) or longer periods (sensitive to the deep mantle lithosphere) can alter the resolvability and bias the estimates of the Moho-depth (Figure 5.19). The best strategy for the inversion of surface-wave data for the Moho with no *a priori* information is, then, to first constrain the crustal and mantle structures with an inversion of the broad-band dispersion curve. The second step is to invert the narrow-band dispersion curve using the results of the first step as reference (or *a priori*) model.

This strategy has been applied to Rayleigh-wave dispersion-curves measurements across southern Africa. Although each curve is an averaged measurement of more than a thousand of curves, the results are promising and this approach could be applied to single station-pairs, so as to increase resolution.

Even though Love-wave shows similar sensitivity to the Moho depth (Figure 5.9 and Figure 5.10), only Rayleigh-wave dispersion curves have been used in this study. Love-wave data were noisier and did not provide robust solutions for the Moho-depth. Joint-inversion of Rayleigh and Love waves have been performed but Love-wave data did not contribute usefully to constrain the Moho-depth.

## 5.6 CONCLUSION

Since its discovery in 1910, the Mohorovičić discontinuity has been the focus of many studies. The depth of the Moho is highly relevant for geodynamic and geological evolution models. Today, the crustal thickness can be resolved at a global or regional scale, depending on the technique used. In this chapter, we have shown how the Moho-depth can be resolved using surface-wave dispersion curves.

Both Rayleigh and Love waves are strongly sensitive to the Moho depth and synthetic tests have shown that a small variation of the depth of the Moho (*i.e.* few kilometers) gen-



erates large and detectable (*i.e.*  $> 1\%$ ) phase velocity perturbations, varying with period. Also, the Moho depth have strong trade-offs with the shear-velocity in the lower-crust and the upper-mantle. A few kilometers of deviation of the depth of the Moho can be compensated almost completely by relatively small variations ( $< 2\%$ ) of the shear velocities in the crust or the upper-mantle. High accuracy of the measurements are then required to resolve the Moho depth with useful precision. Inverting data with a small amount of noise for the Moho is, however, possible. To invert the data for the Moho-depth, the effective strategy proposed here is to first invert the surface-wave dispersion curve alone to determine the shear-velocity depth-profile. In a second step, to only invert for the Moho-depth the period-range of the dispersion curve that is the most sensitive to the presumed Moho-depth using the resulted shear-velocity profile as an *a priori* model (or reference model).

The procedure has been applied to a dataset of several Rayleigh-wave dispersion curves in Southern Africa, each being an average over a region with a relatively homogeneous structure within it. Results over regional-average dispersion curves are very promising and suggest that this procedure will also be applicable datasets sampling a smaller region, increasing the lateral resolution of the Moho-depth variations.

Part IV

CONCLUSION AND BIBLIOGRAPHY

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## CONCLUSIONS

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The principal goal of this Ph.D dissertation was to study the structure and dynamics of ancient continents, with the main focus on the anisotropic seismic structure of the lithosphere and asthenosphere of the Archean Kaapvaal Craton and Limpopo Belt in southern Africa. Mapping the anisotropic structure help us to better understand the mechanisms of deformation of the Archean cratons during and since their formation. In this dissertation, we measured azimuthal and radial anisotropy and its lateral and vertical variations.

We measured a large amount of Rayleigh wave phase velocities in a broad band of periods. We obtained accurate measurements of azimuthal anisotropy and robust depth-dependent profiles of shear velocity azimuthal and radial anisotropy sampling from the upper-crust, down to the asthenosphere.

For the first time to our knowledge, we are able to estimate the regional stress using teleseismic surface-wave measurements. We showed that the azimuthal anisotropy in the upper-crust of the Limpopo Belt is related to aligned micro-cracks, reflecting the southward propagation of the East African Rift. Our results showed that tectonic stress can be estimated from short-period inter-station measurements of surface waves, suggesting that the regional stress could be mapped in regions around the world where it is poorly known but where seismic data are available.

Previous observations of SKS-splitting measurements (Silver *et al.*, 2004) showed lateral variations of azimuthal anisotropy across southern Africa. However, SKS-splitting measurements lack vertical resolution and the origin of the splitting has been subject to debate. On the one hand, Silver *et al.* (2004) argued that the splitting originates in the lithosphere and is related to the Archean collisions between the Kaapvaal Craton, the Zimbabwe Craton and the Limpopo Belt in the North and between the central and western parts of the Kaapvaal Craton in the South. On the other hand, Vinnik *et al.* (1995) argued that the splitting originates in the asthenosphere, reflecting the current plate motion. Using accurate depth-distribution of azimuthal anisotropy, we were able to resolve the debate, reconciling both interpretations. We showed that the fabric in the lower mantle-lithosphere of the Limpopo Belt and the northern part of the Kaapvaal Craton is parallel to



the Archean-Paleoproterozoic sutures. There is, however, no evidence for suture-parallel fabric in the mantle-lithosphere of the south-western Kaapvaal Craton, where the collision mechanism may have been different. In the North, our results thus confirm the interpretation of Silver *et al.* (2004) who suggested that the fast directions, parallel to the Archean-Paleoproterozoic sutures, reflect fabric in the lithosphere. Elsewhere, our results validate the arguments of Vinnik *et al.* (1995) who suggested that the splitting reflects plate motion and originates in the asthenosphere. The direction of fast propagation in the asthenosphere is consistent in the four sub-regions with identified within southern Africa and trend to parallel the plate motion.

A more detailed investigation of the stratification of azimuthal and radial anisotropy beneath the Kaapvaal Craton highlighted an intra-lithospheric boundary (at around 90 km) between layers with different anisotropy. This boundary is not characterized by a variation of isotropic shear velocities. Interestingly, receiver function studies also observed a boundary at ~100 km beneath cratons, in southern Africa and elsewhere, that is not explained with an isotropic shear velocity discontinuity. Our results confirm that this boundary beneath southern Africa could be associated with a change of anisotropy. Importantly, the different orientation of fabric within the upper and lower mantle lithosphere, which probably records the deformation during their last major deformation episodes, suggests that the upper and lower lithosphere may have formed and stabilized at different times.

Our high-quality phase velocity measurements enabled us to measure robust depth-profiles of the vertically polarized shear velocity. The root of the Kaapvaal Craton and the Limpopo Belt extends down to around 200 km and is characterized by fast velocities. We observed a low velocity zone in the four regions, weak in the Limpopo Belt and northern Kaapvaal Craton and stronger in the southern Kaapvaal Craton. Anomalous high relief is observed across the entire region, with the southern part of the Kaapvaal Craton about 1 km higher than its northern part. This could be explained, at least in part, by an increase of the buoyancy in the lower-lithosphere and asthenosphere associated with an increase of temperature of approximately 220°C (Li and Burke, 2006) that could be responsible for the LVZ.

This dissertation also presents results from a research undertaken on the study of the resolvability of the depth of the Moho using surface wave dispersion measurements. We showed that Rayleigh and Love waves are highly sensitive to the depth of the Moho. The depth of this discontinuity can therefore be estimated with accuracy. However, we map several slices of the model-space and showed that the trade-offs between the depth of the Moho and the shear velocity in the lower-crust and the upper mantle-lithosphere are strong. We propose an effective strategy to invert phase-velocity data with measurements errors for the Moho-depth. In a first step we determine the shear velocity structures in the crust and mantle by inversion of the broad-band dispersion curve. In a second step, we only inverted the period-band of the dispersion curve for which the surface wave is

the most sensitive to the presumed Moho depth-range, using the results of the first step as a reference model. Using this technique, we resolved the depth of the Moho beneath southern Africa with good accuracy.

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Part V

APPENDICES



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## DISTRIBUTION OF PHASE VELOCITIES WITH AZIMUTH

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This appendix presents the distribution of Rayleigh- and Love-wave phase velocities with azimuth after weighting and averaging over sliding windows for every periods (between 4.97 and 500.25 s) in the Limpopo Belt region (Figure A.1), North Kaapvaal Craton (Figure A.2), central Kaapvaal Craton (Figure A.3), South-West Kaapvaal Craton (Figure A.4) and all-region-average (Figure A.5). Colored dots are the data. Black solid lines are best-fitting models with isotropic and  $2\varphi$  anisotropic parameters. Black dot lines are best-fitting models with isotropic and 2 and  $4\varphi$  anisotropic parameters.

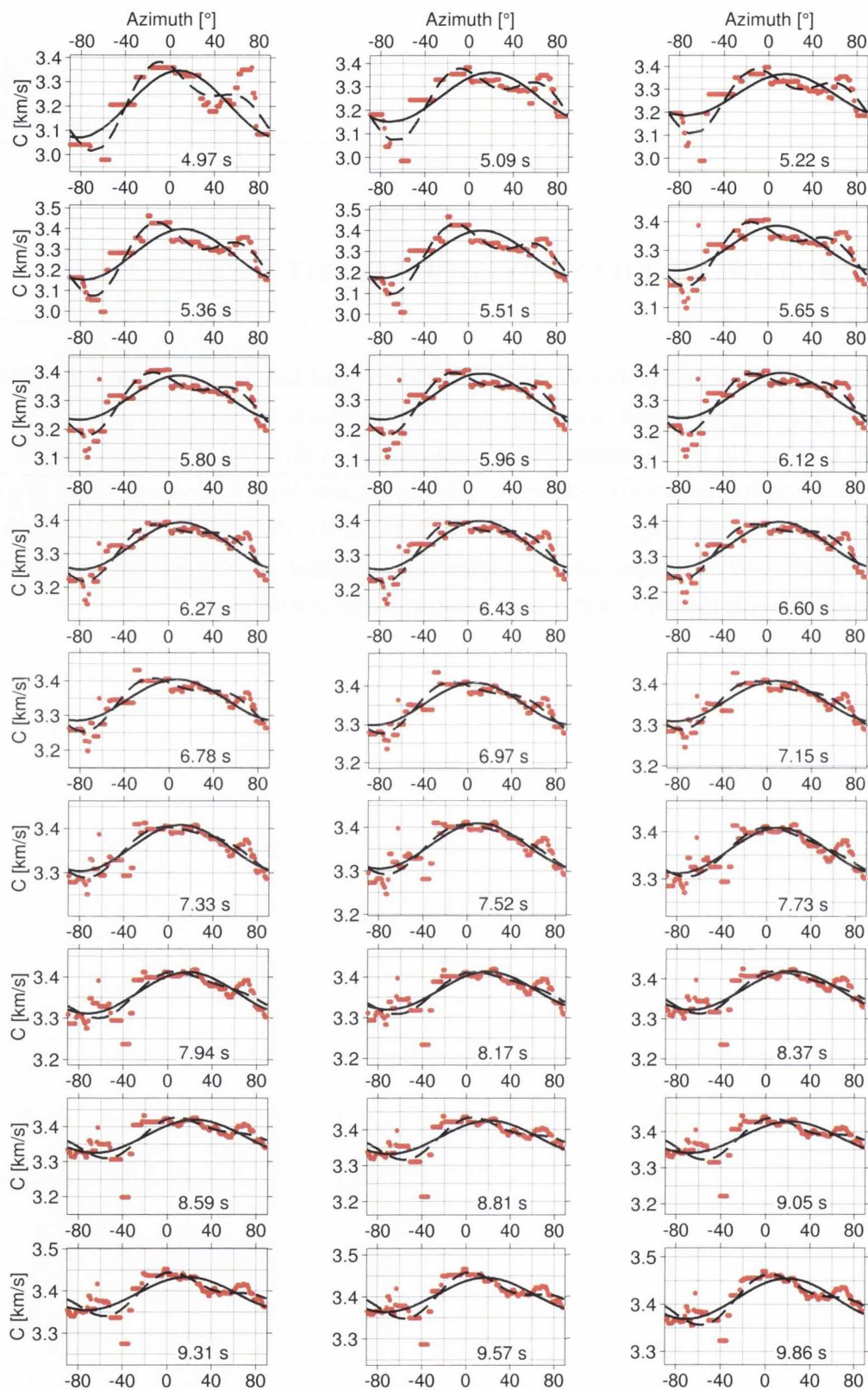
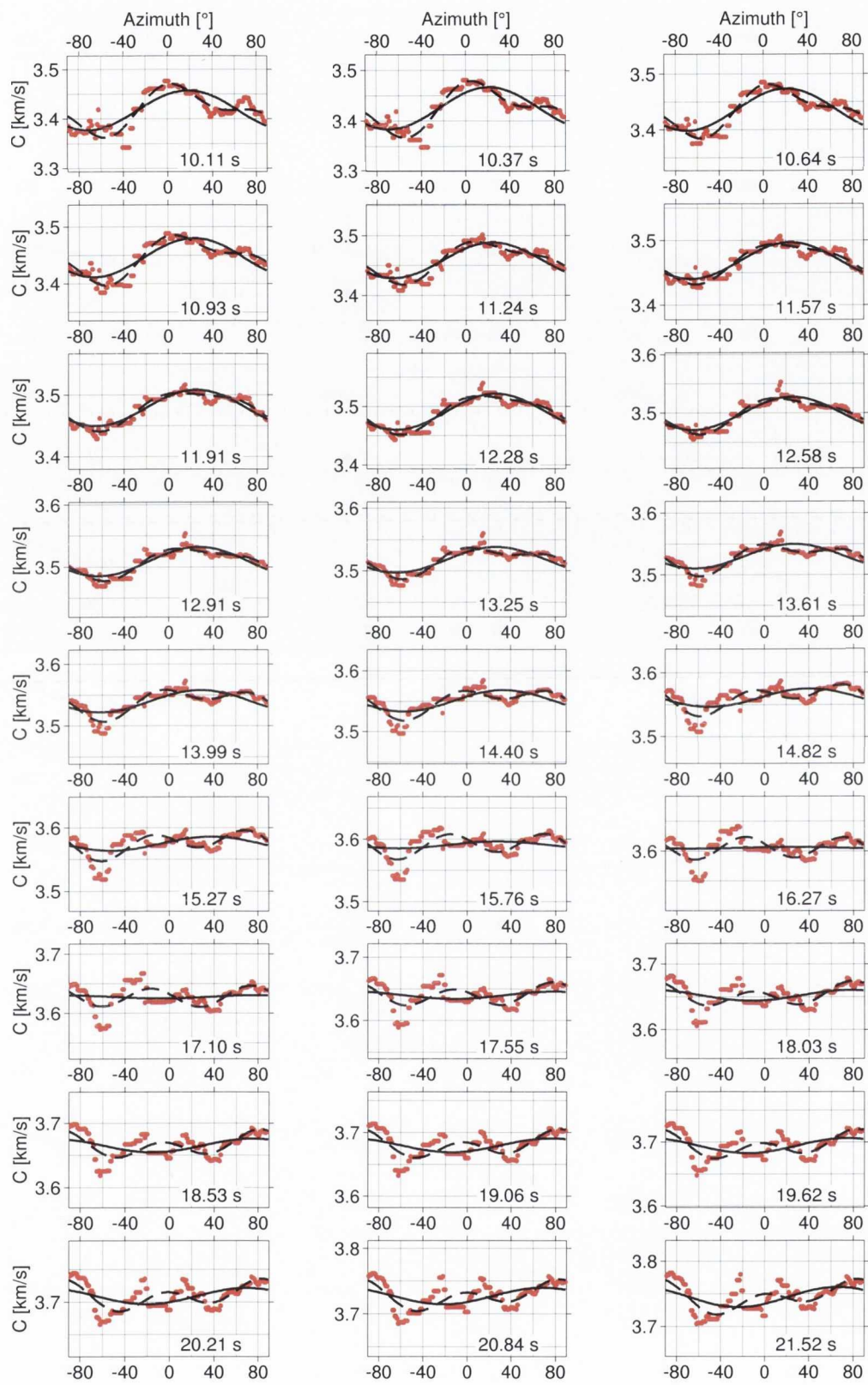
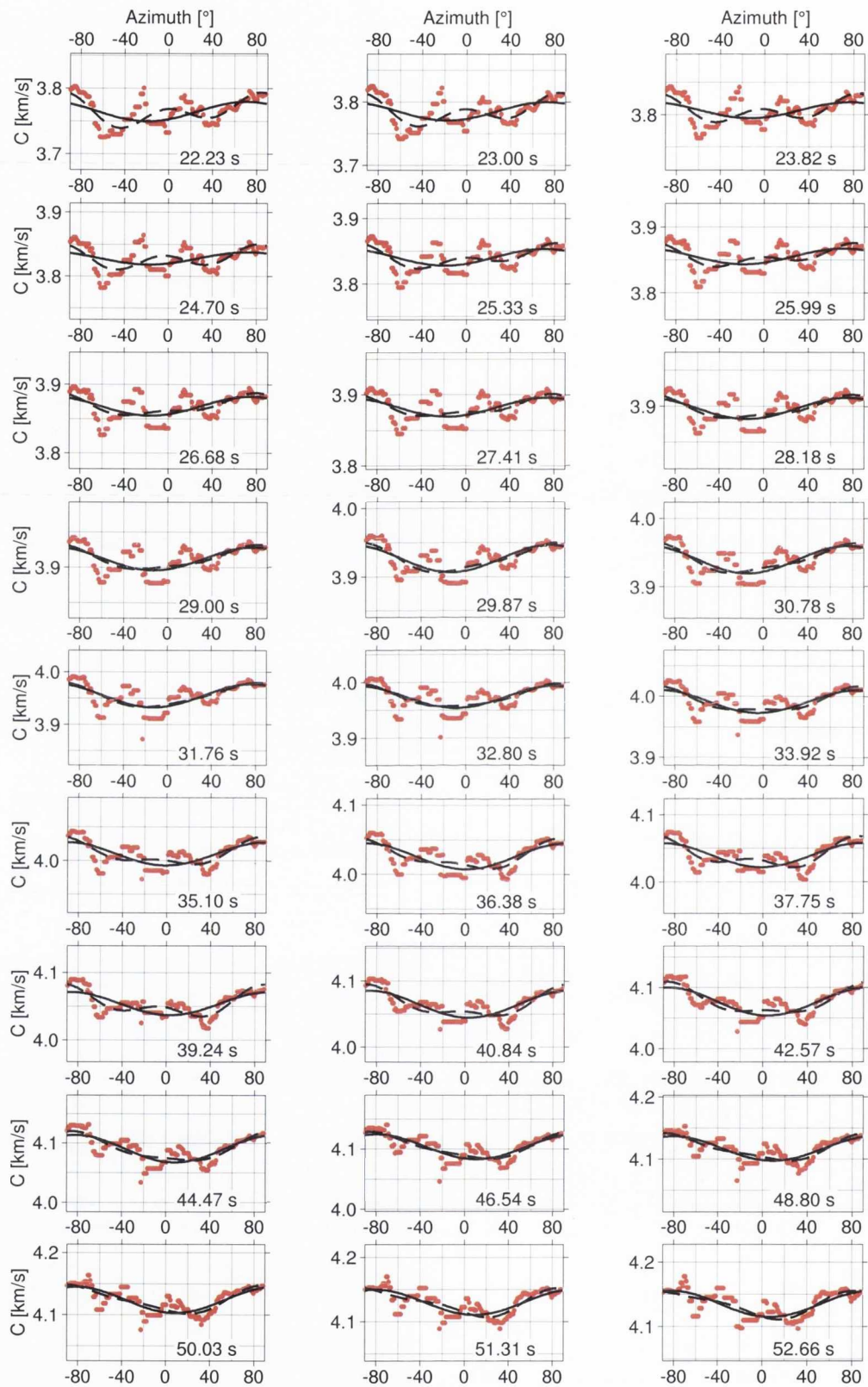


Figure A.1.: Distribution of Rayleigh-wave phase velocities with azimuth - Limpopo Belt (page 1/5)

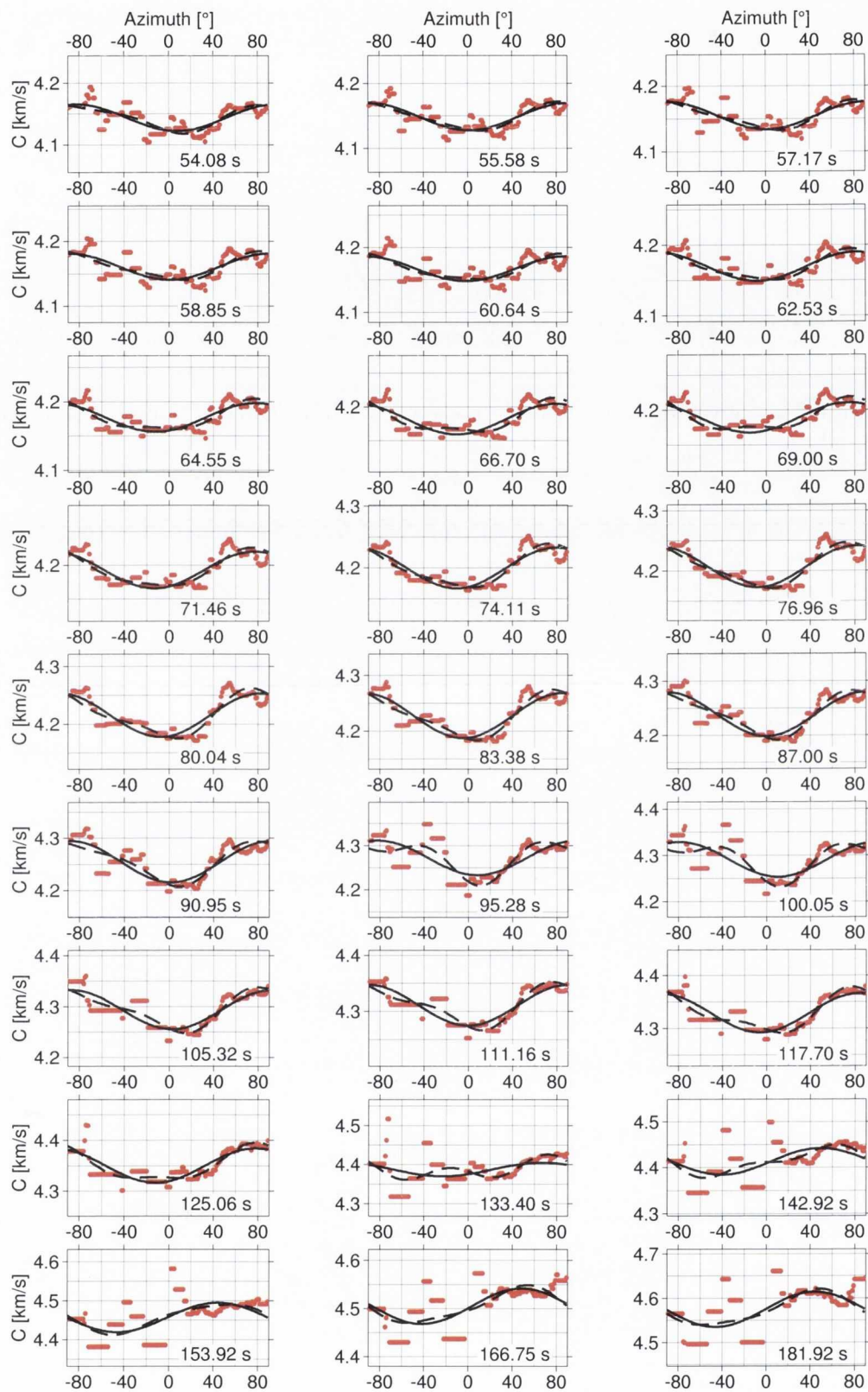


Distribution of Rayleigh-wave phase velocities with azimuth - Limpopo Belt (page 2/5)

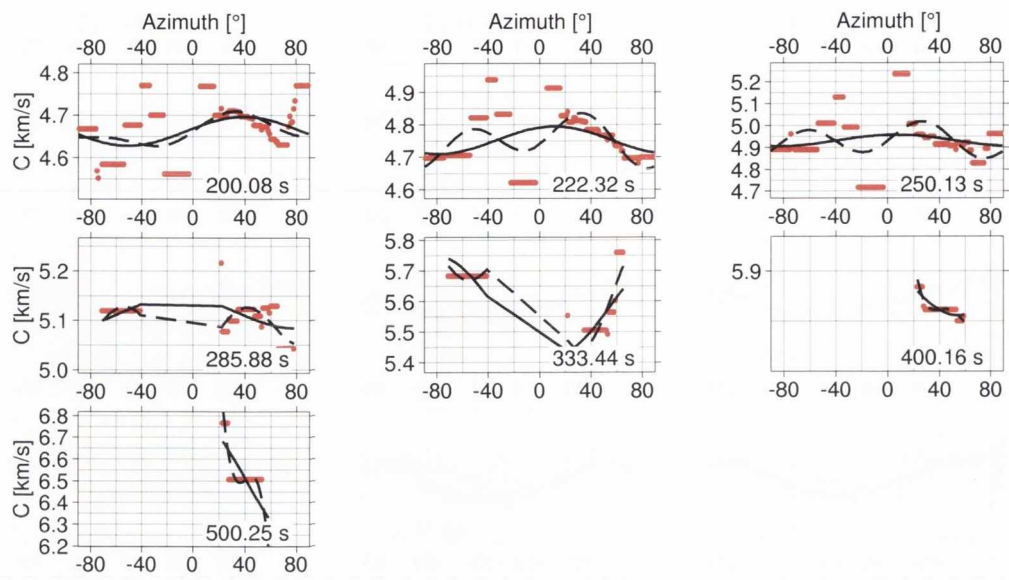




Distribution of Rayleigh-wave phase velocities with azimuth - Limpopo Belt (page 3/5)

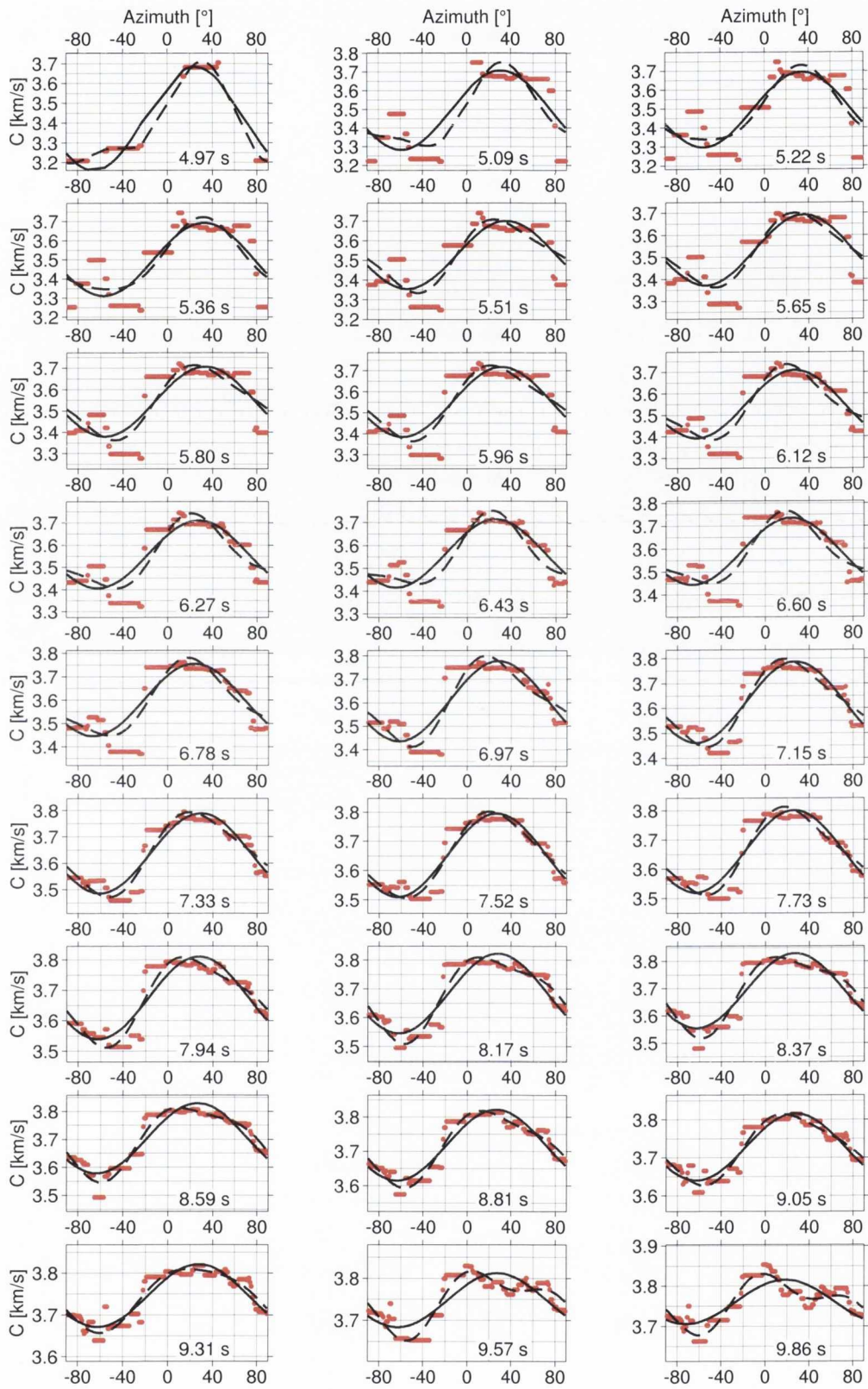


Distribution of Rayleigh-wave phase velocities with azimuth - Limpopo Belt (page 4/5)

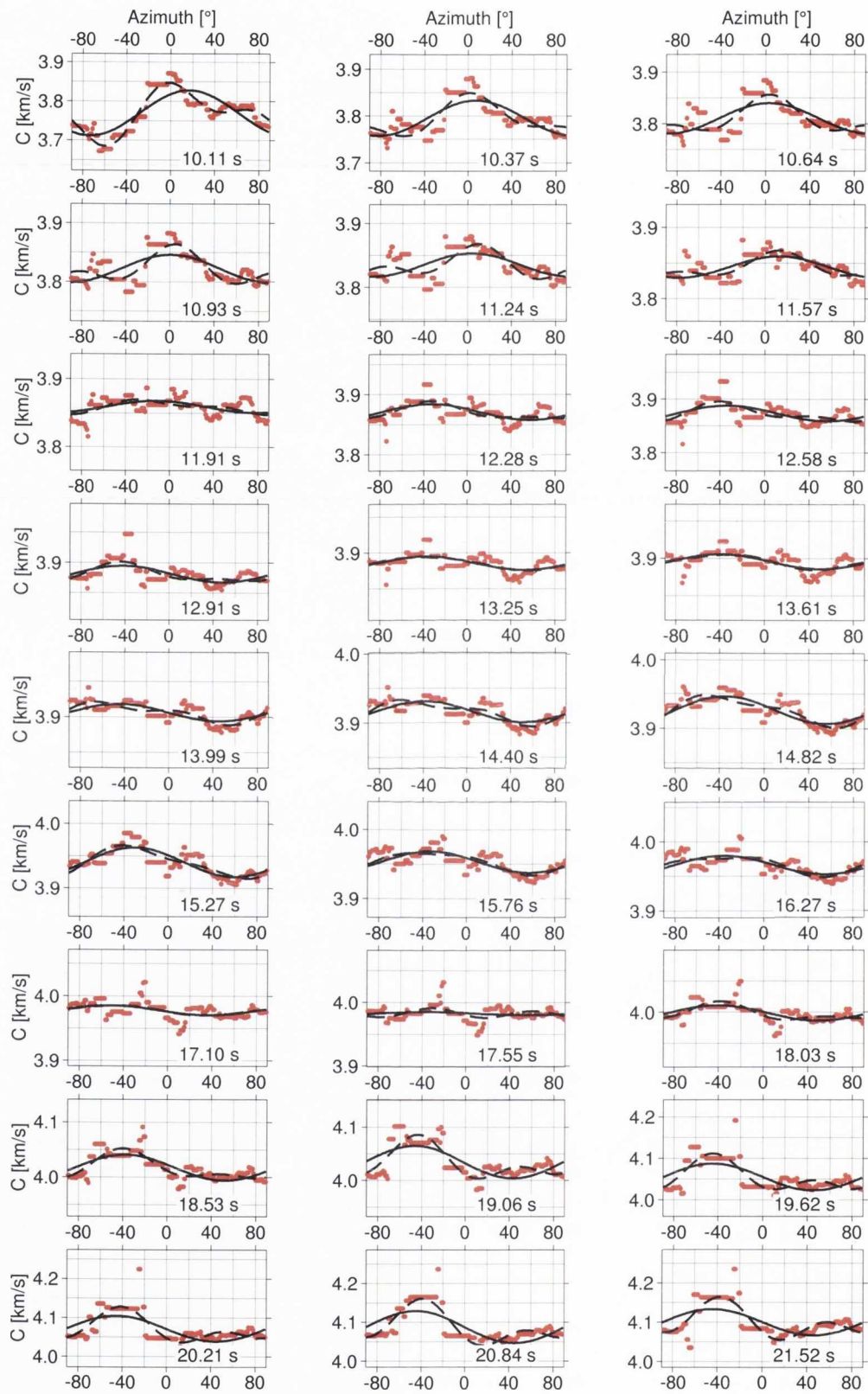


Distribution of Rayleigh-wave phase velocities with azimuth - Limpopo Belt (page 5/5)

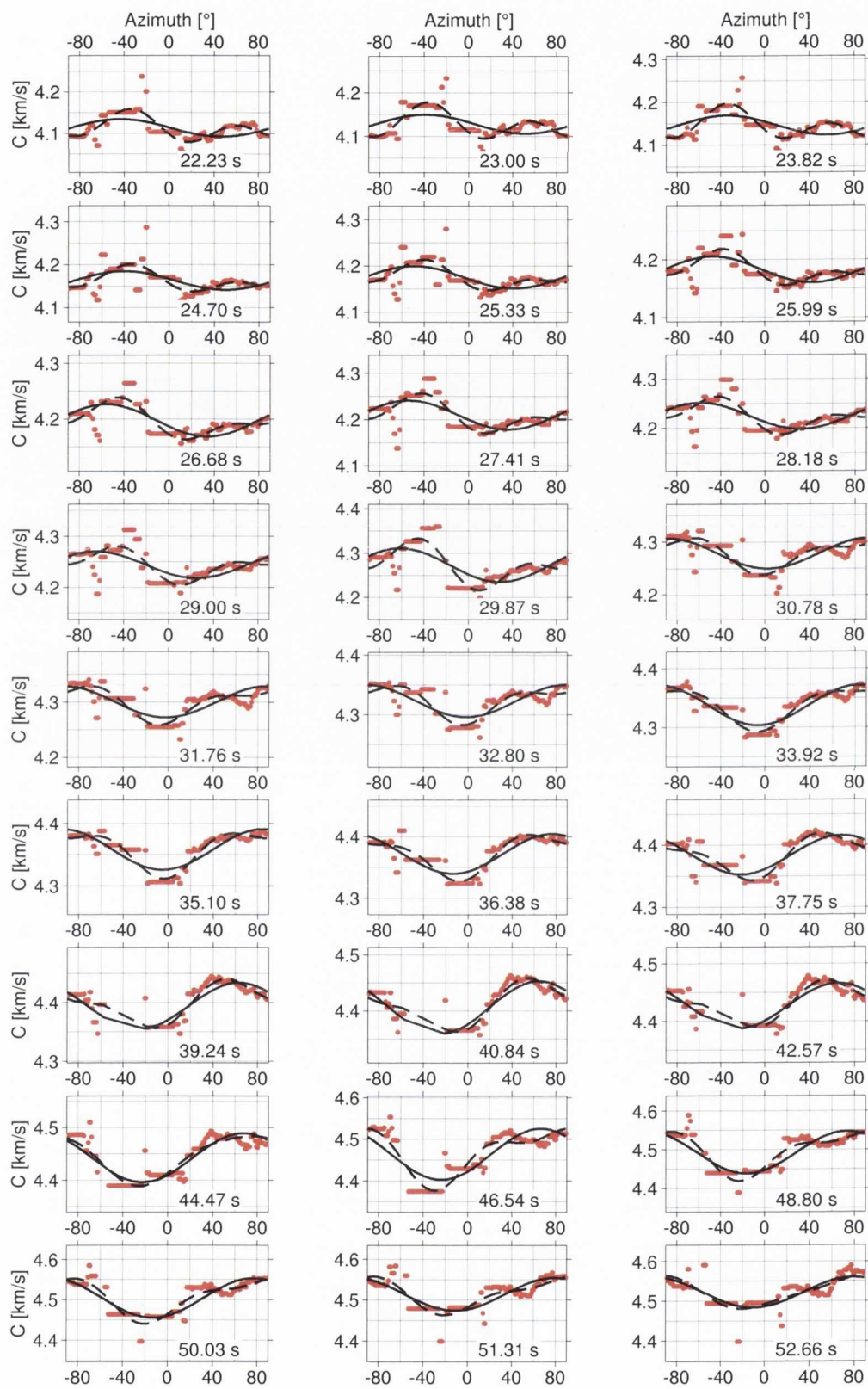




Distribution of Love-wave phase velocities with azimuth - Limpopo Belt (page 1/5)

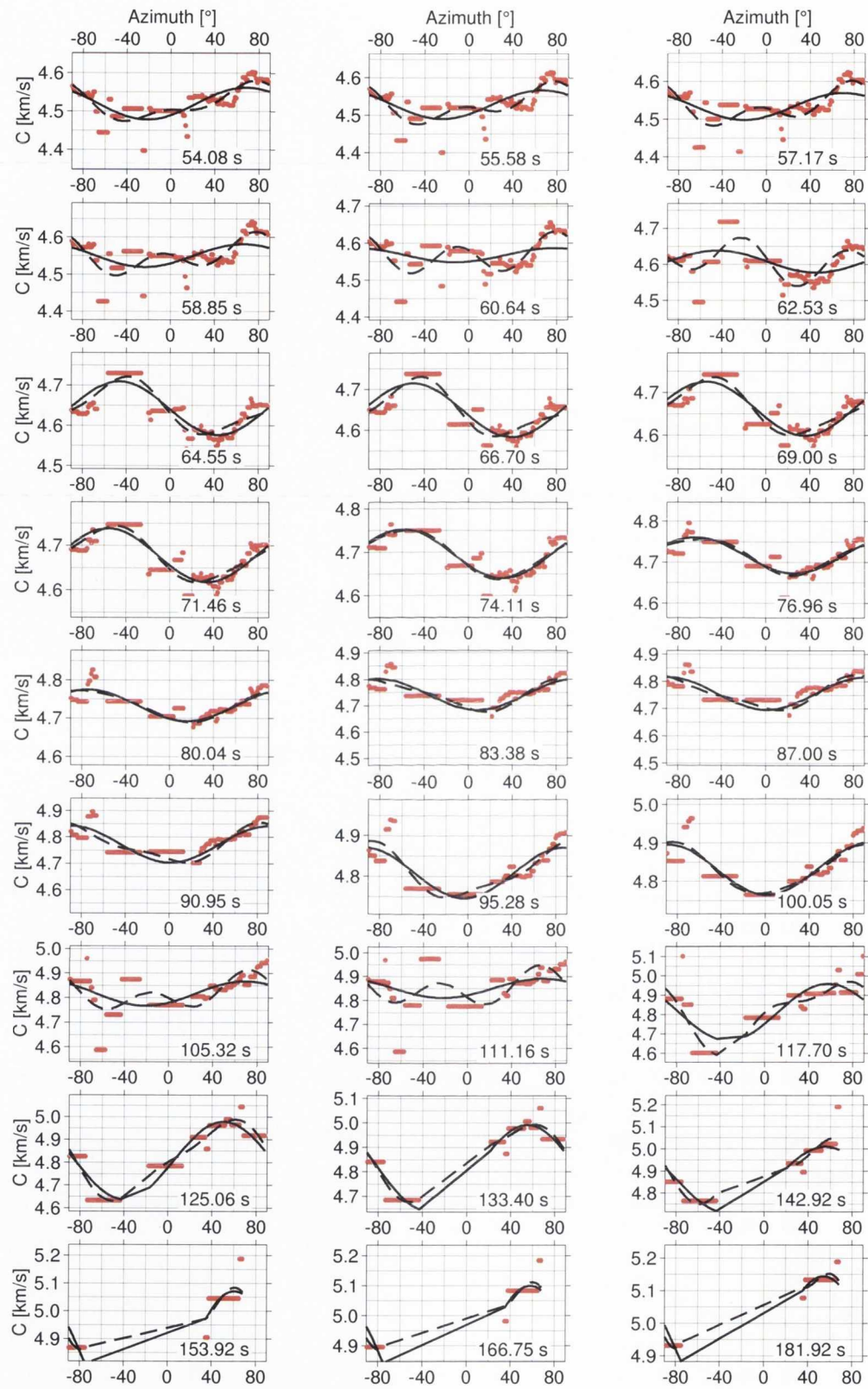


Distribution of Love-wave phase velocities with azimuth - Limpopo Belt (page 2/5)

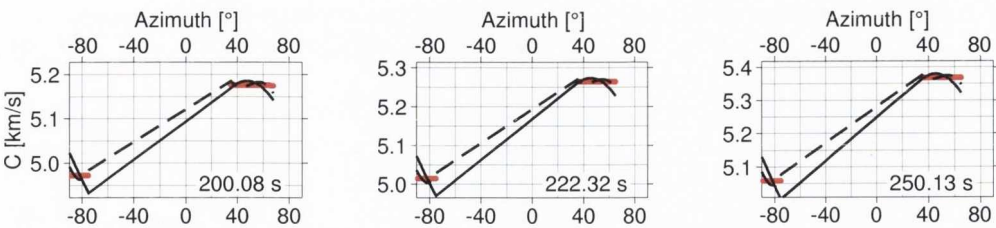


Distribution of Love-wave phase velocities with azimuth - Limpopo Belt (page 3/5)





Distribution of Love-wave phase velocities with azimuth - Limpopo Belt (page 4/5)



Distribution of Love-wave phase velocities with azimuth - Limpopo Belt (page 5/5)

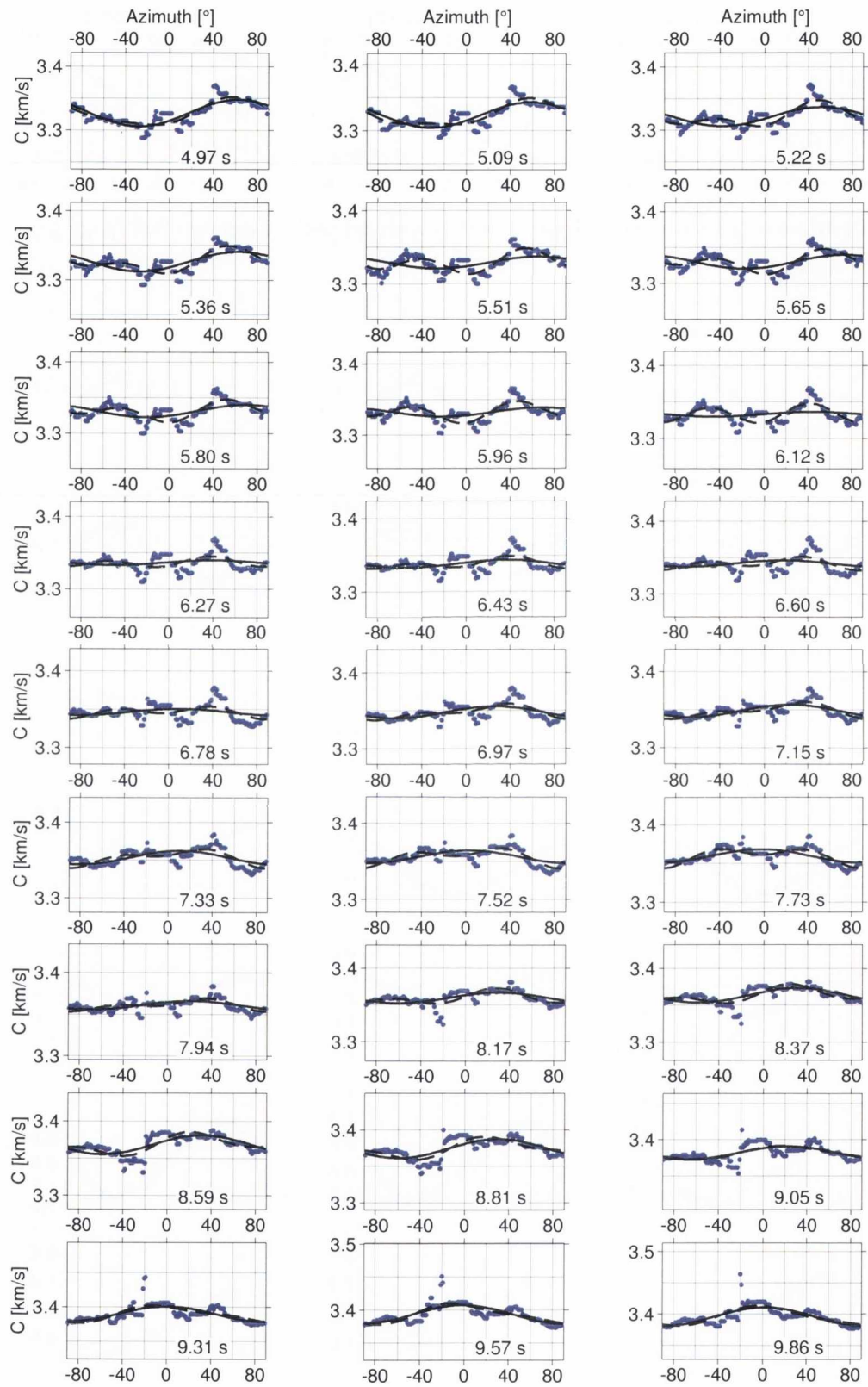
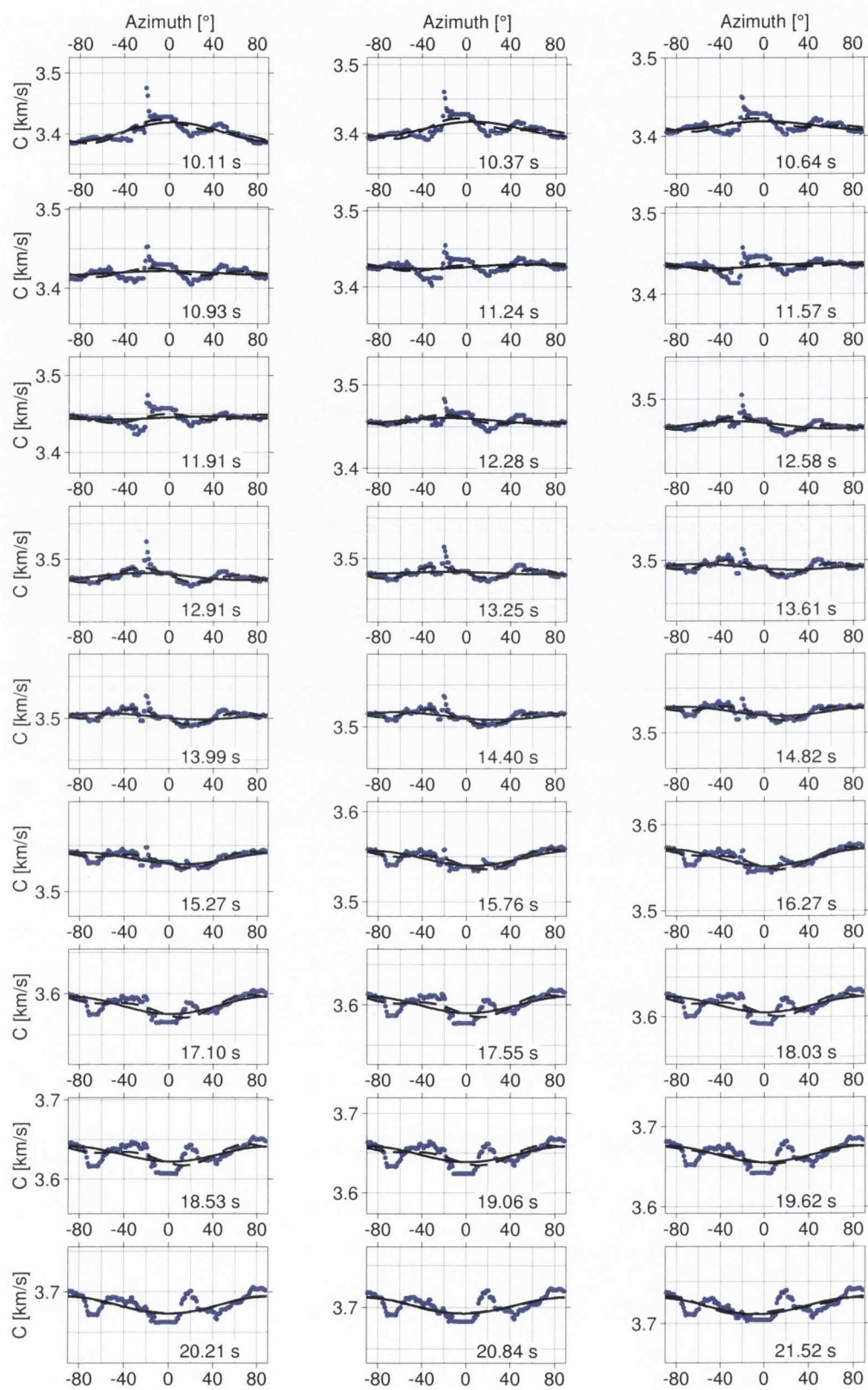
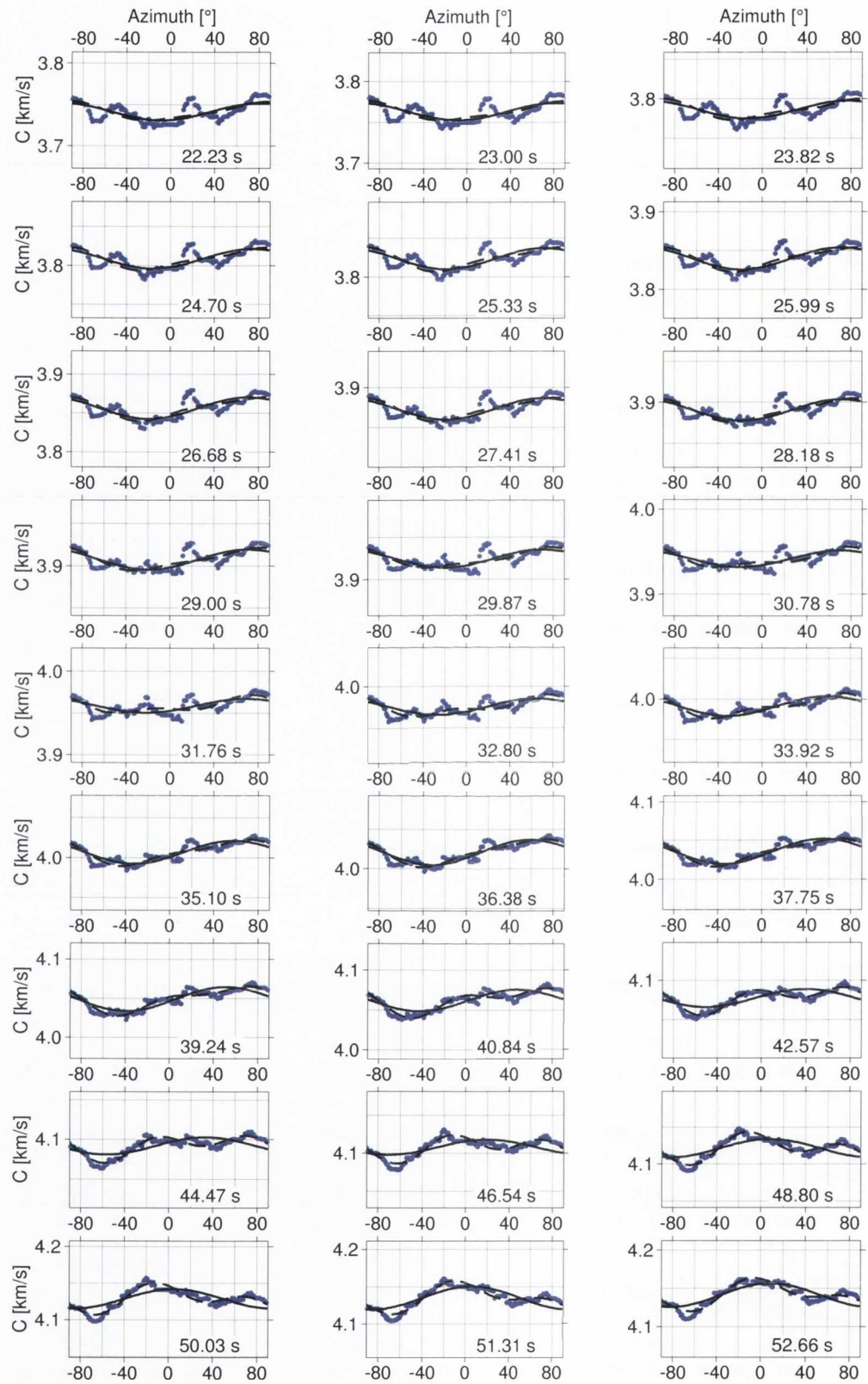


Figure A.2.: Distribution of Rayleigh-wave phase velocities with azimuth - North K.C. (page 1/5)

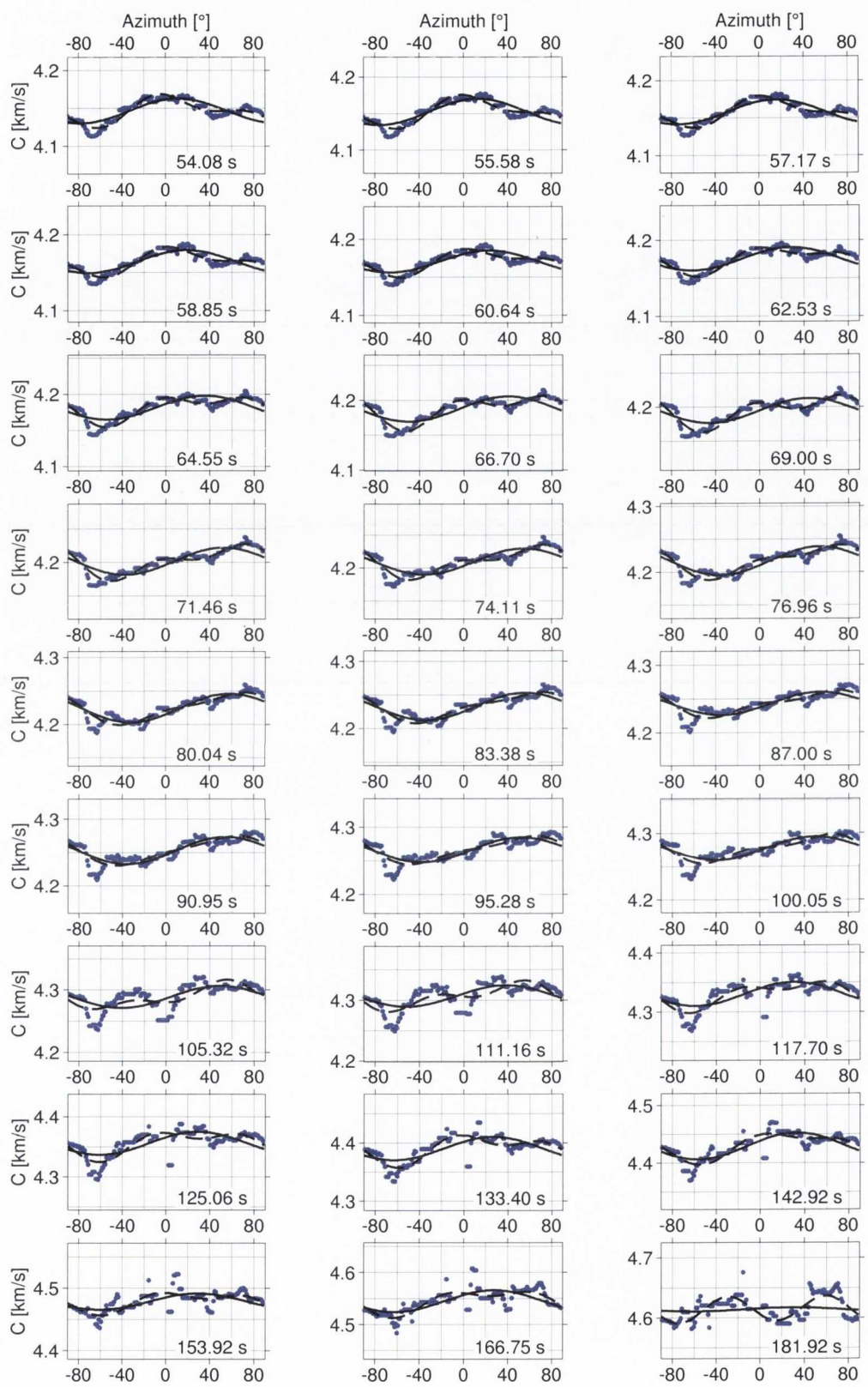




Distribution of Rayleigh-wave phase velocities with azimuth - North K.C. (page 2/5)

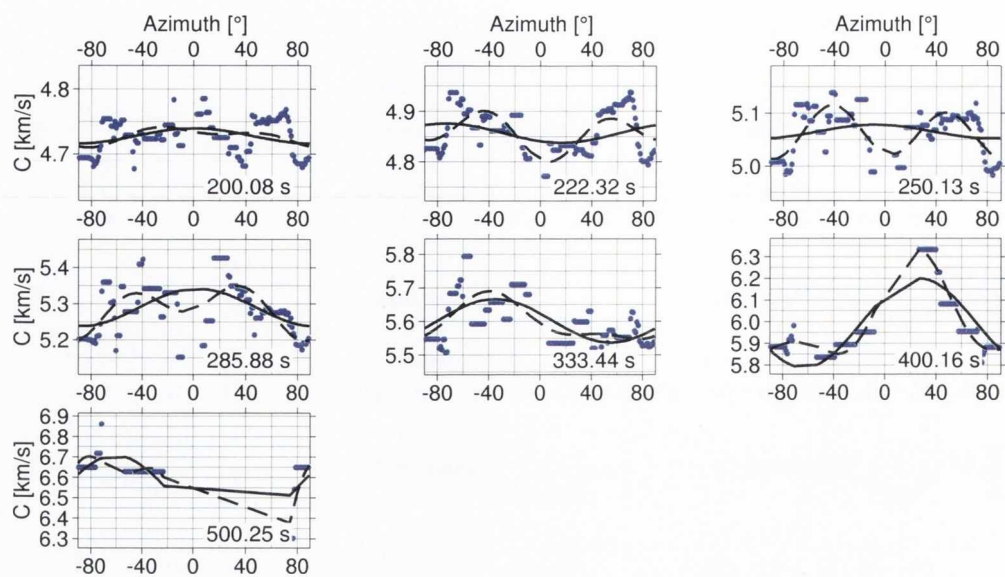


Distribution of Rayleigh-wave phase velocities with azimuth - North K.C. (page 3/5)

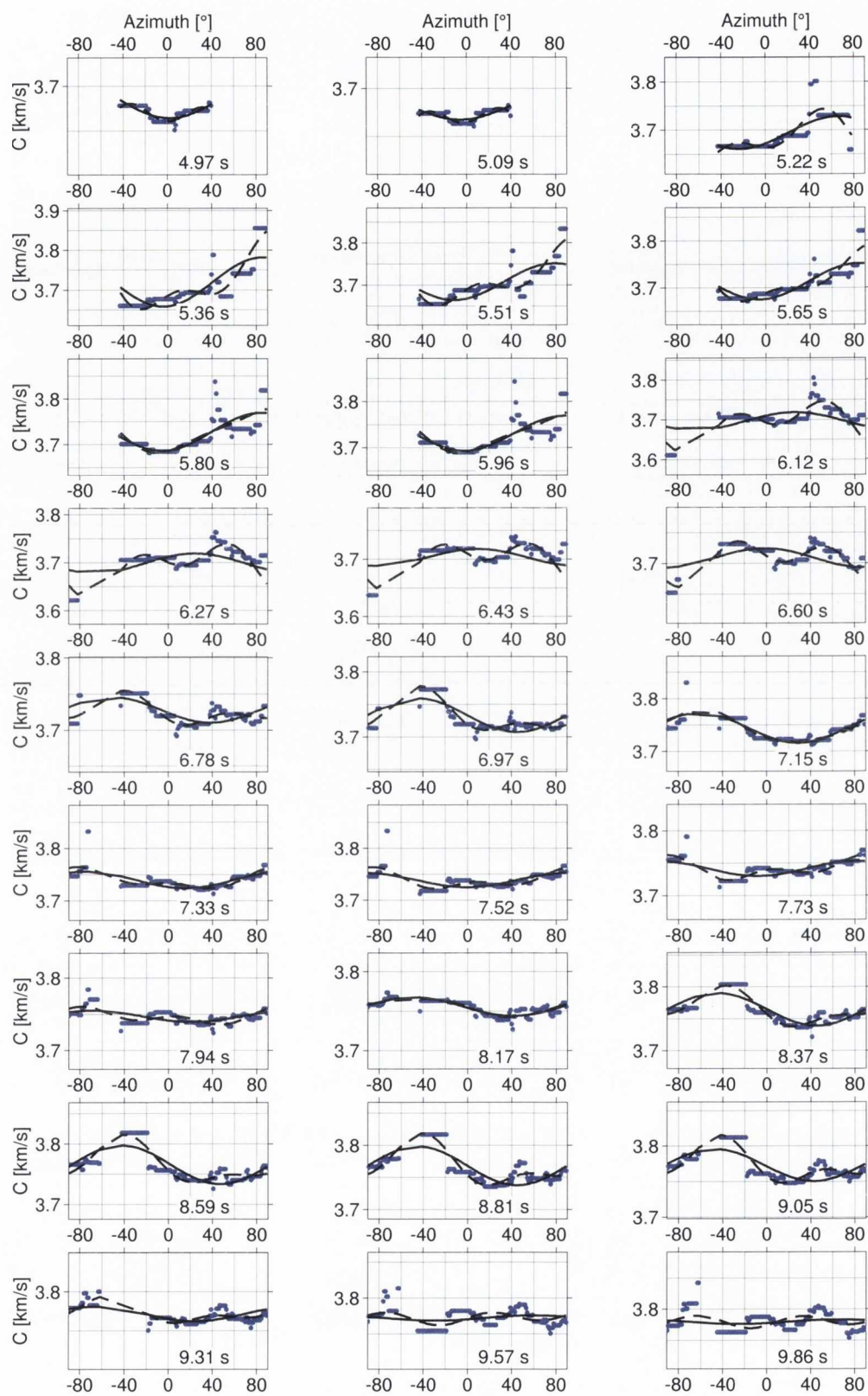


Distribution of Rayleigh-wave phase velocities with azimuth - North K.C. (page 4/5)

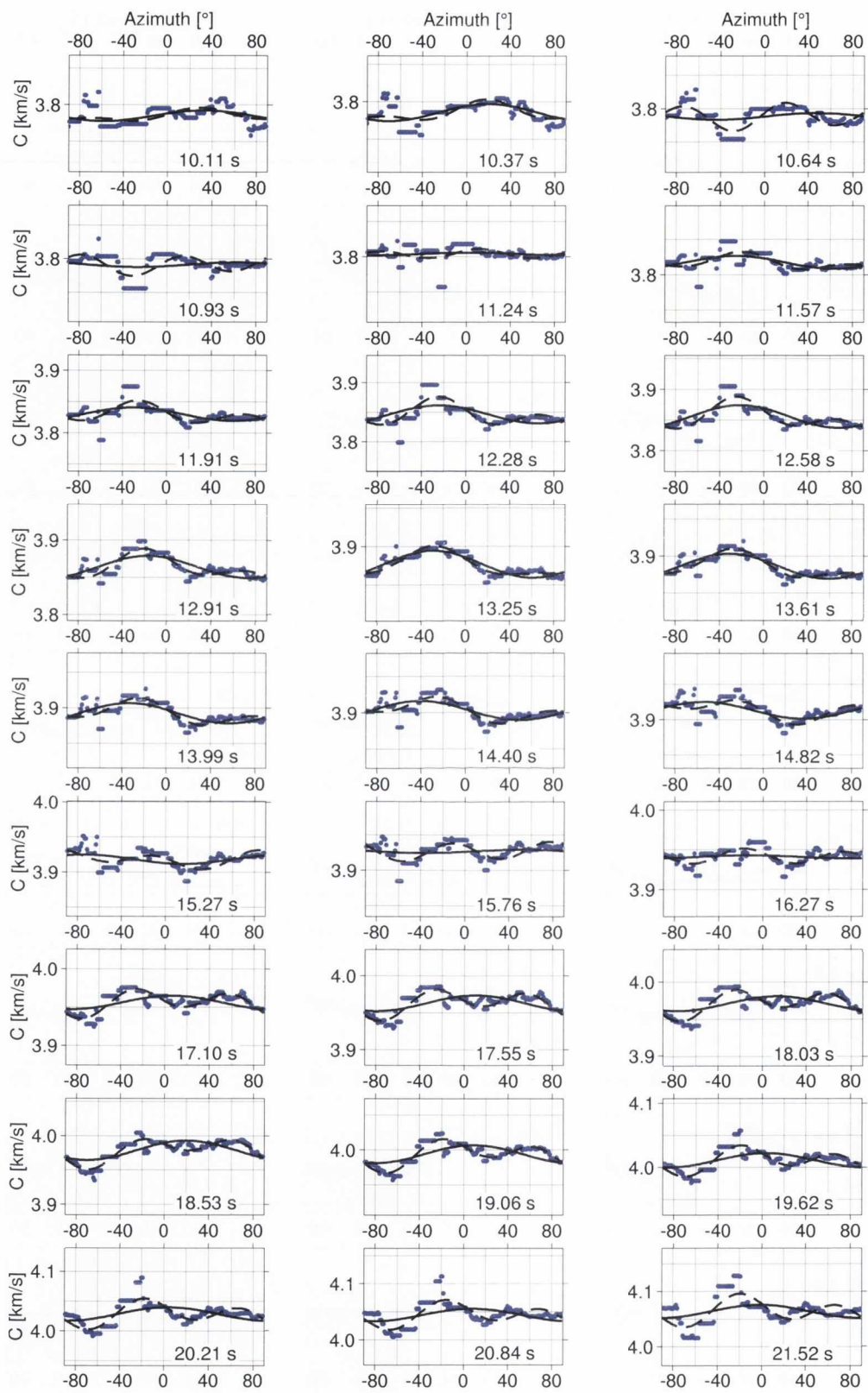




Distribution of Rayleigh-wave phase velocities with azimuth - North K.C. (page 5/5)

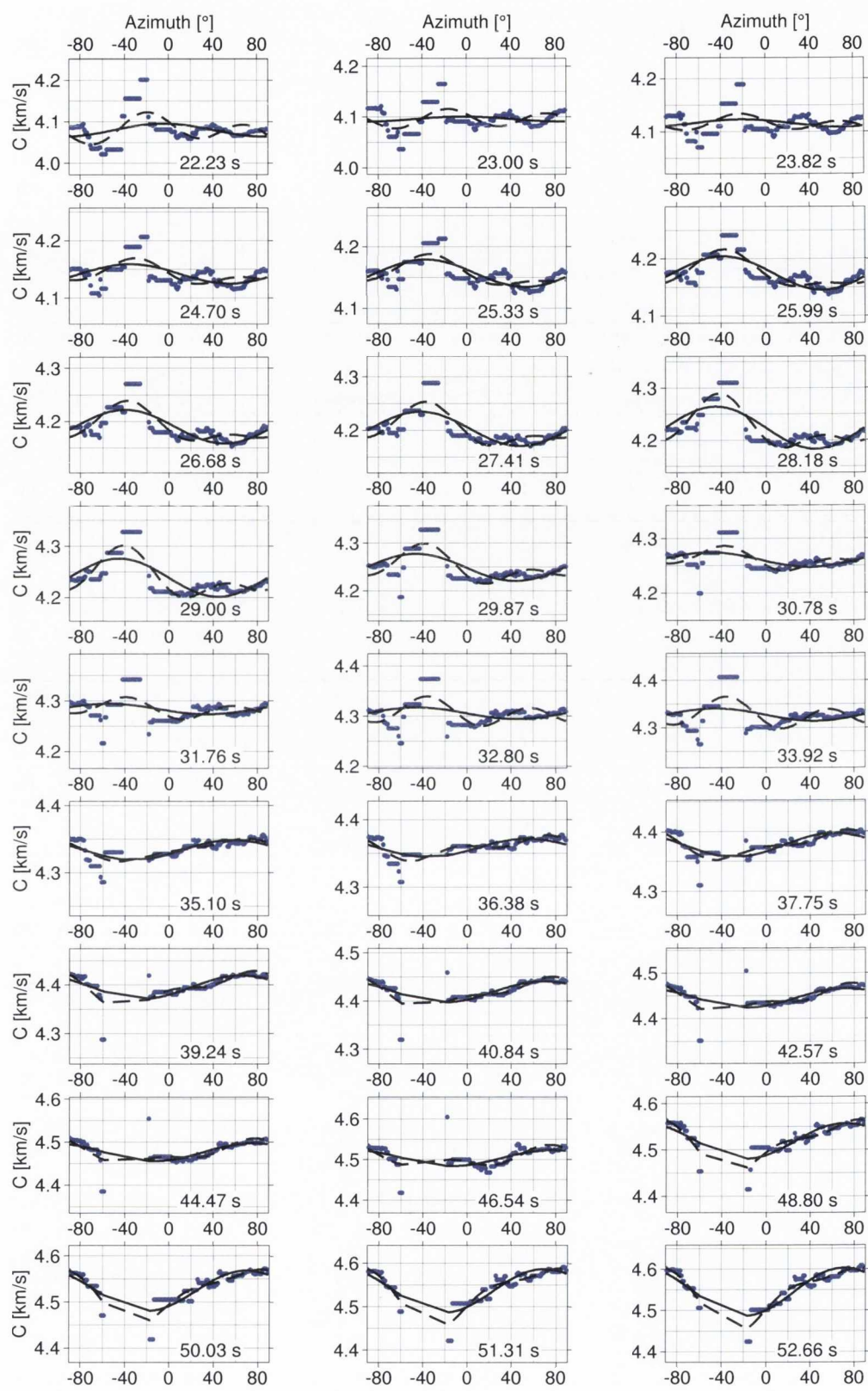


Distribution of Love-wave phase velocities with azimuth - North K.C. (page 1/5)

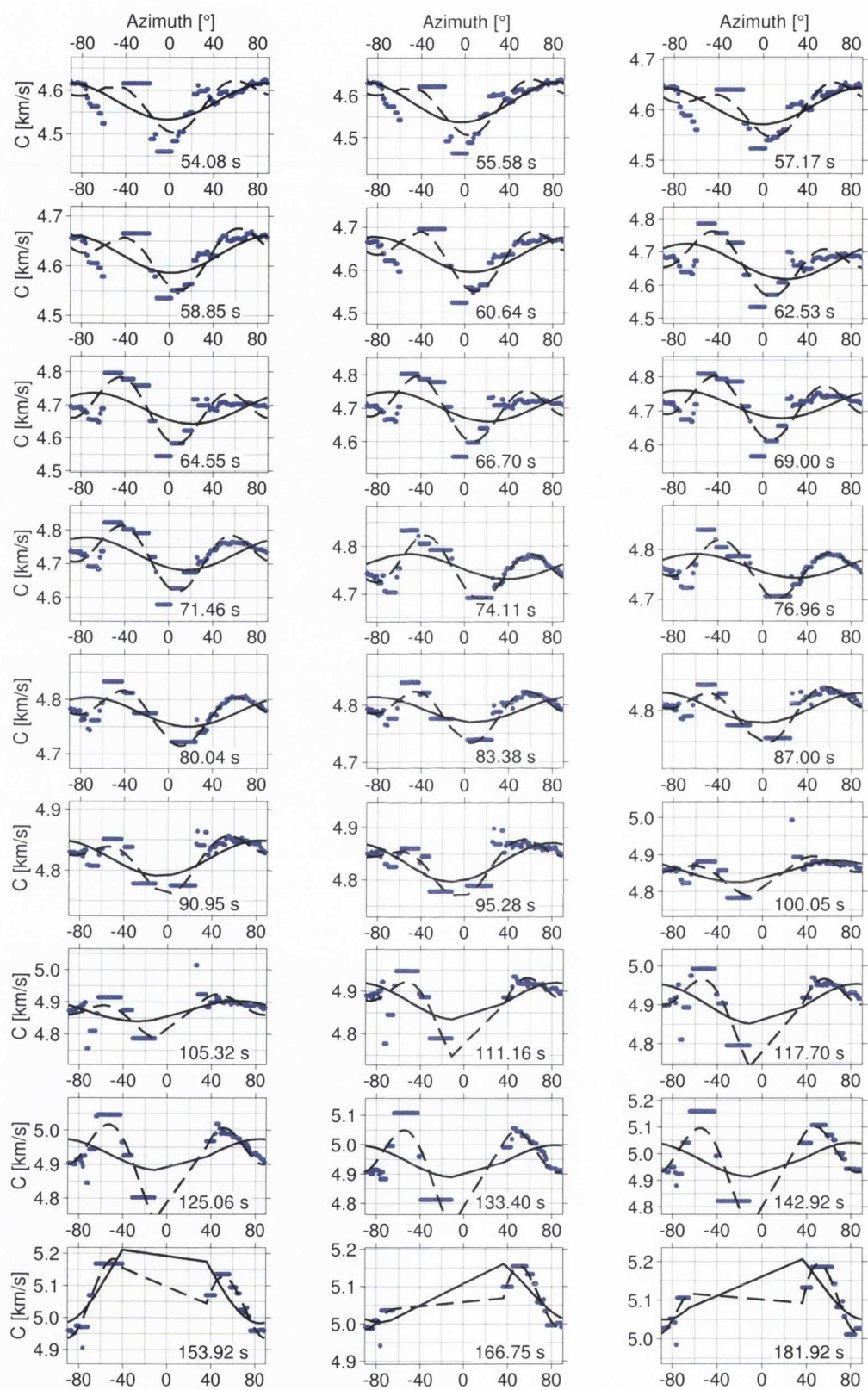


Distribution of Love-wave phase velocities with azimuth - North K.C. (page 2/5)

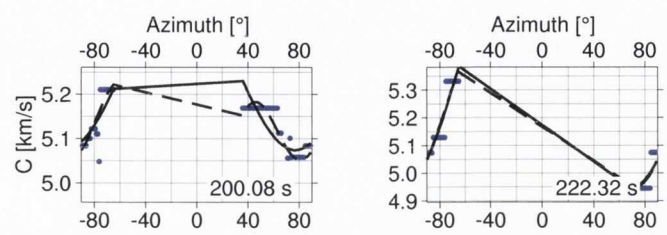




Distribution of Love-wave phase velocities with azimuth - North K.C. (page 3/5)



Distribution of Love-wave phase velocities with azimuth - North K.C. (page 4/5)



Distribution of Love-wave phase velocities with azimuth - North K.C. (page 5/5)



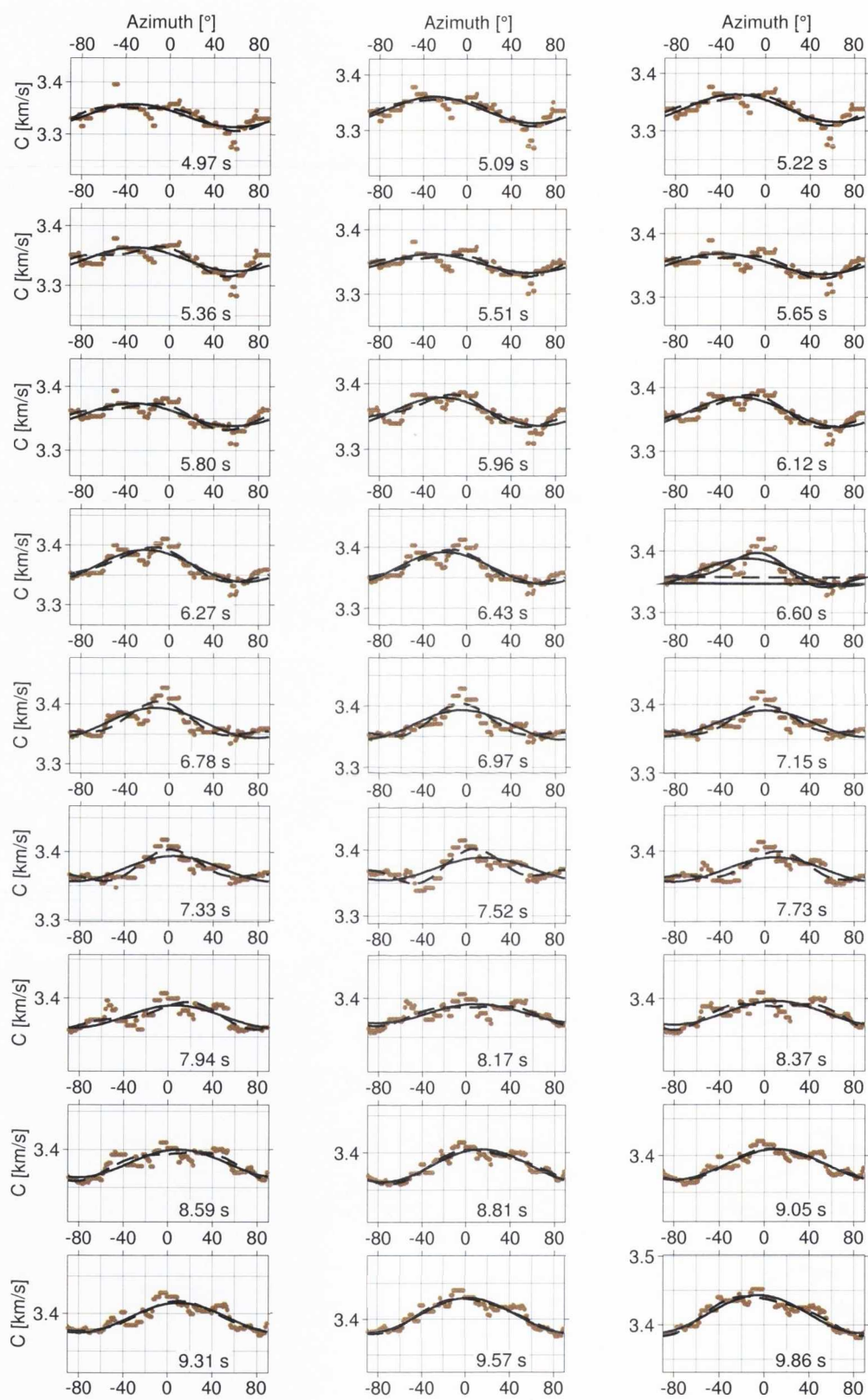
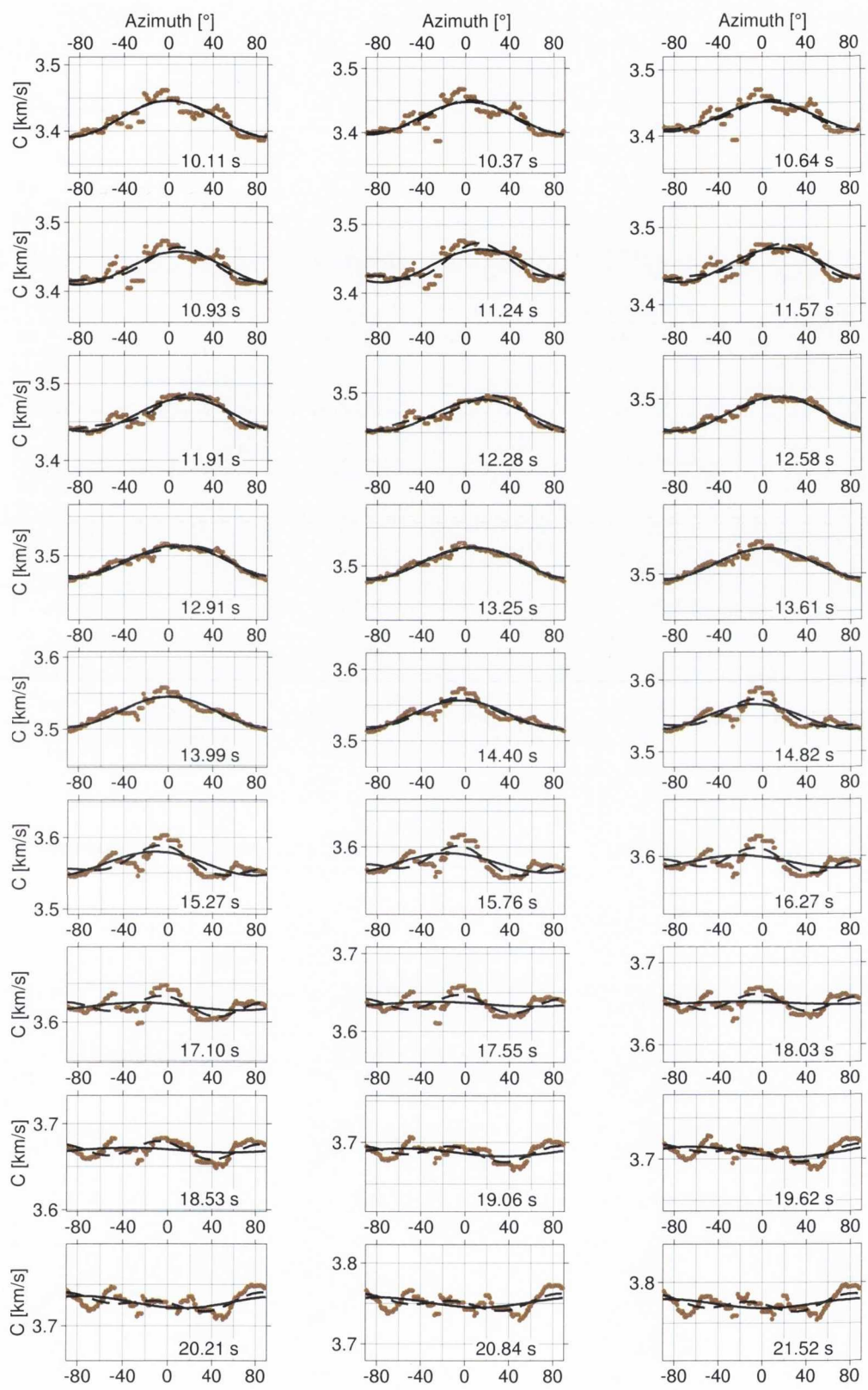
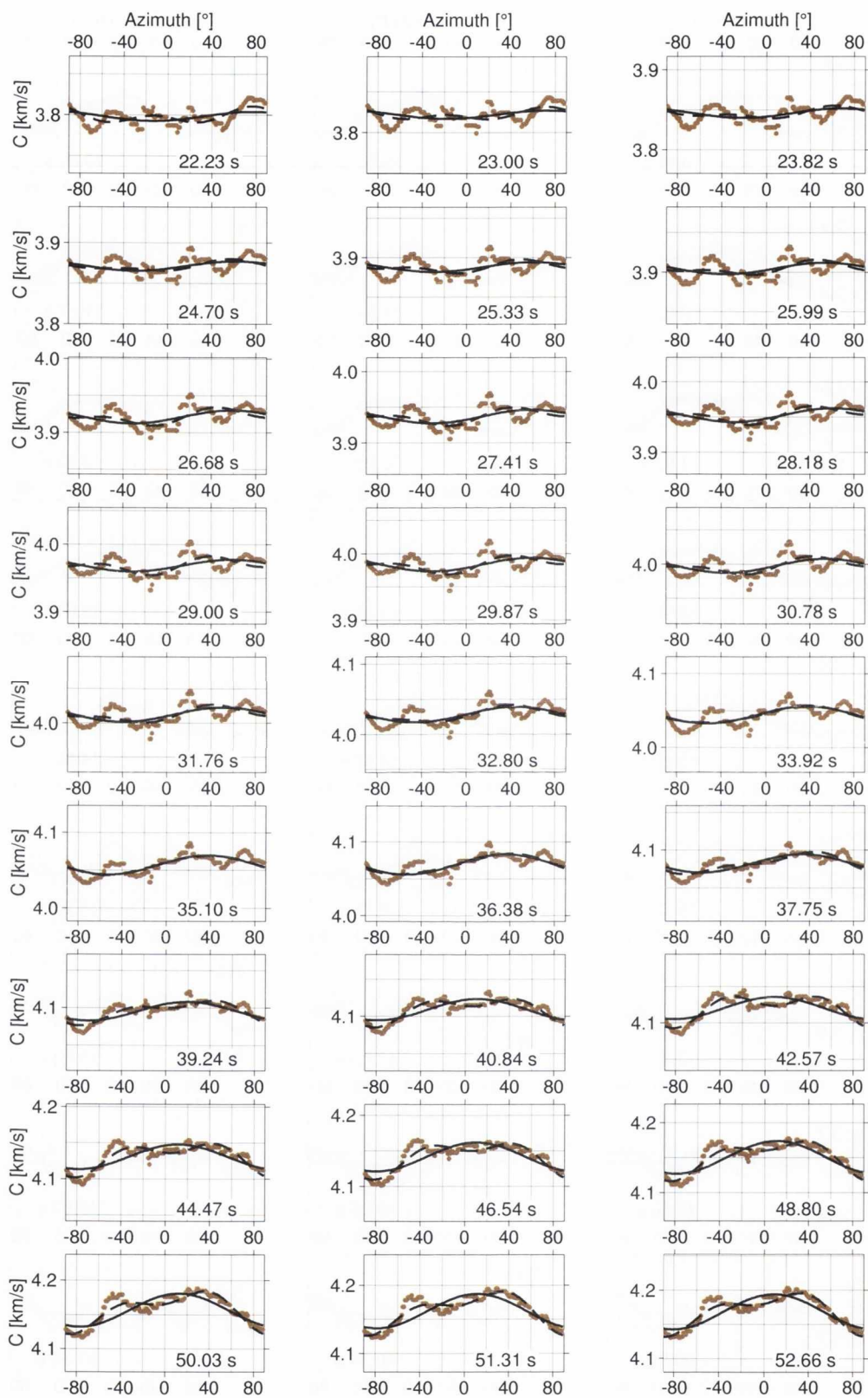


Figure A.3.: Distribution of Rayleigh-wave phase velocities with azimuth - central K.C. (page 1/5)

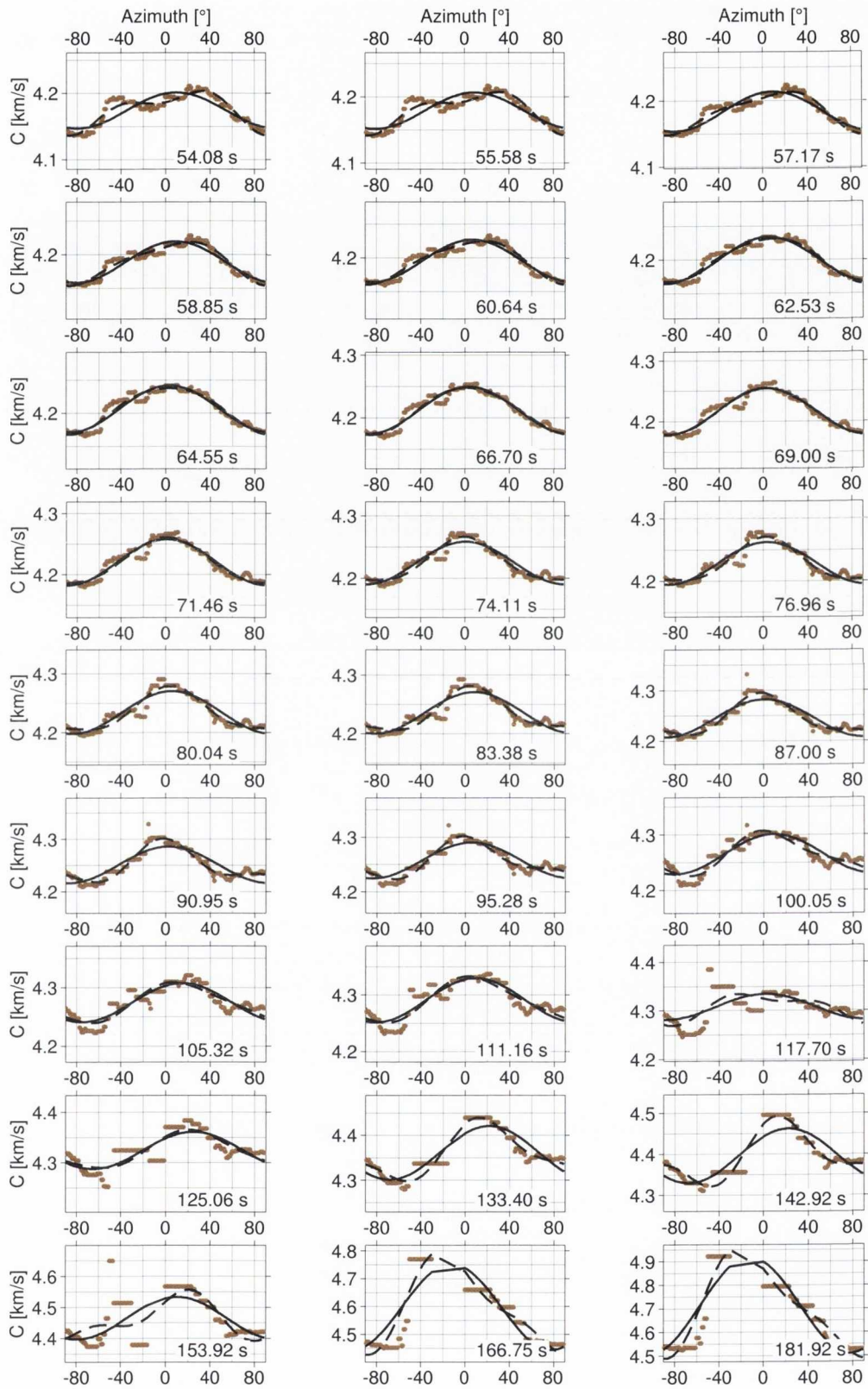


Distribution of Rayleigh-wave phase velocities with azimuth - central K.C. (page 2/5)

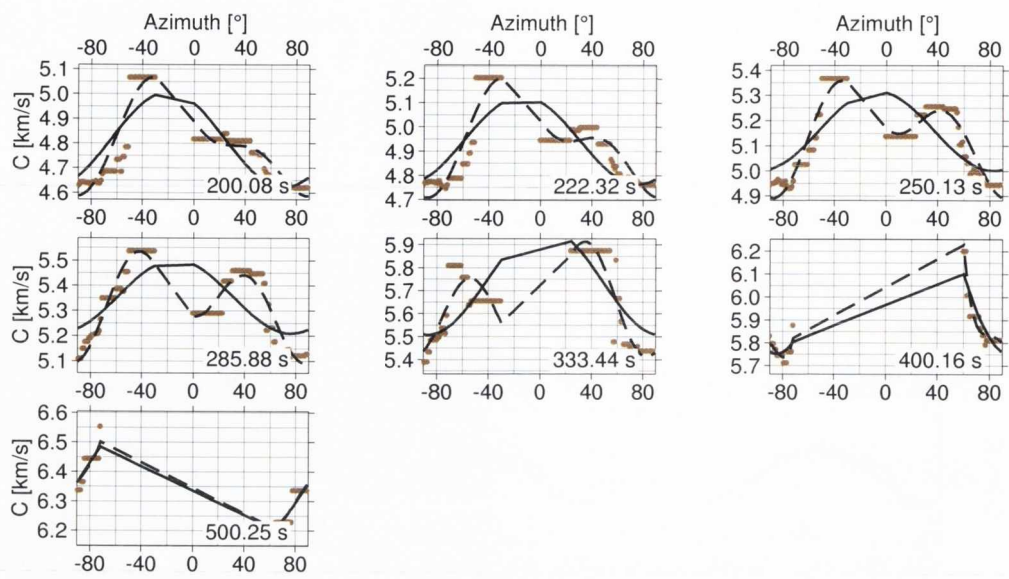


Distribution of Rayleigh-wave phase velocities with azimuth - central K.C. (page 3/5)

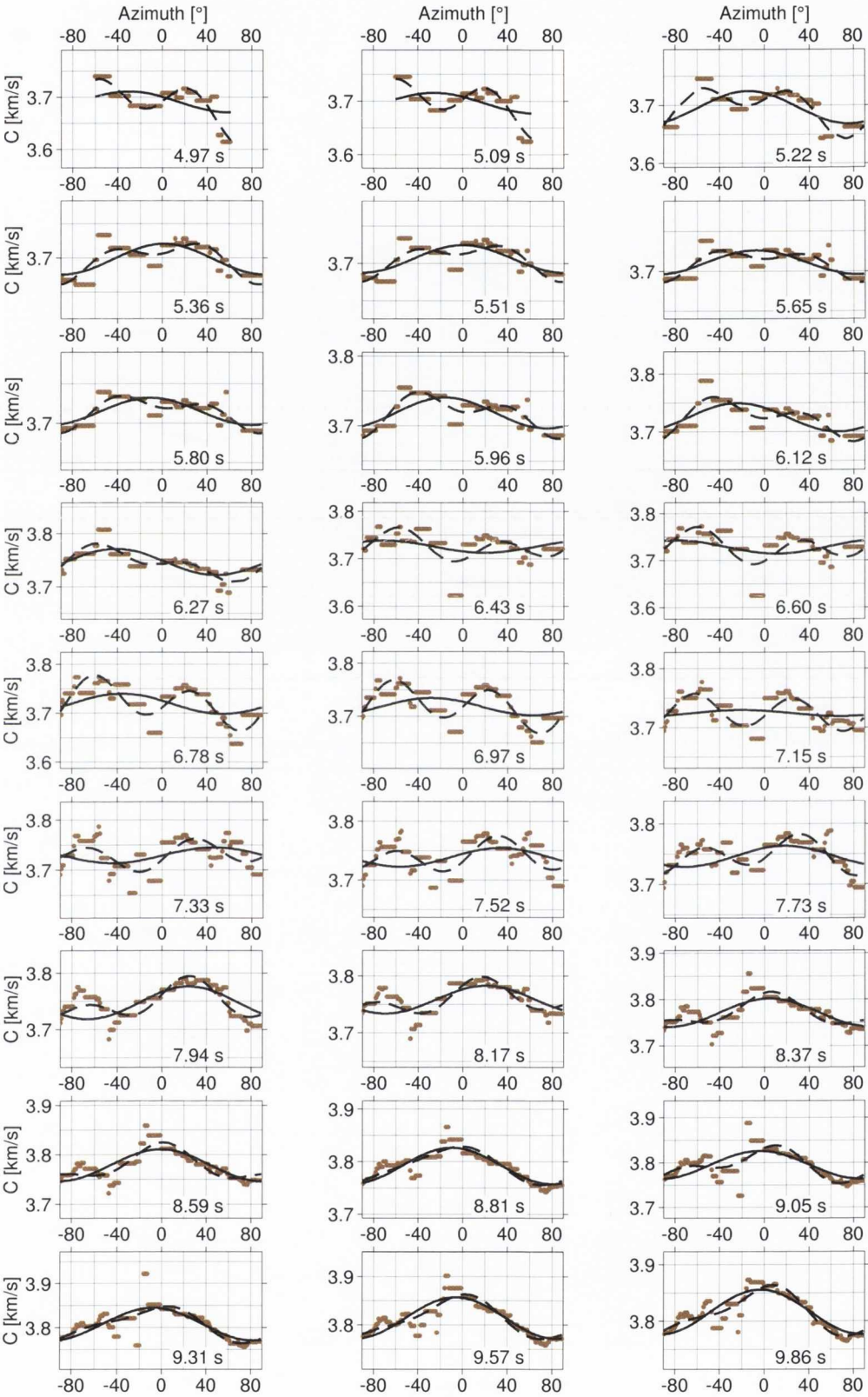




Distribution of Rayleigh-wave phase velocities with azimuth - central K.C. (page 4/5)

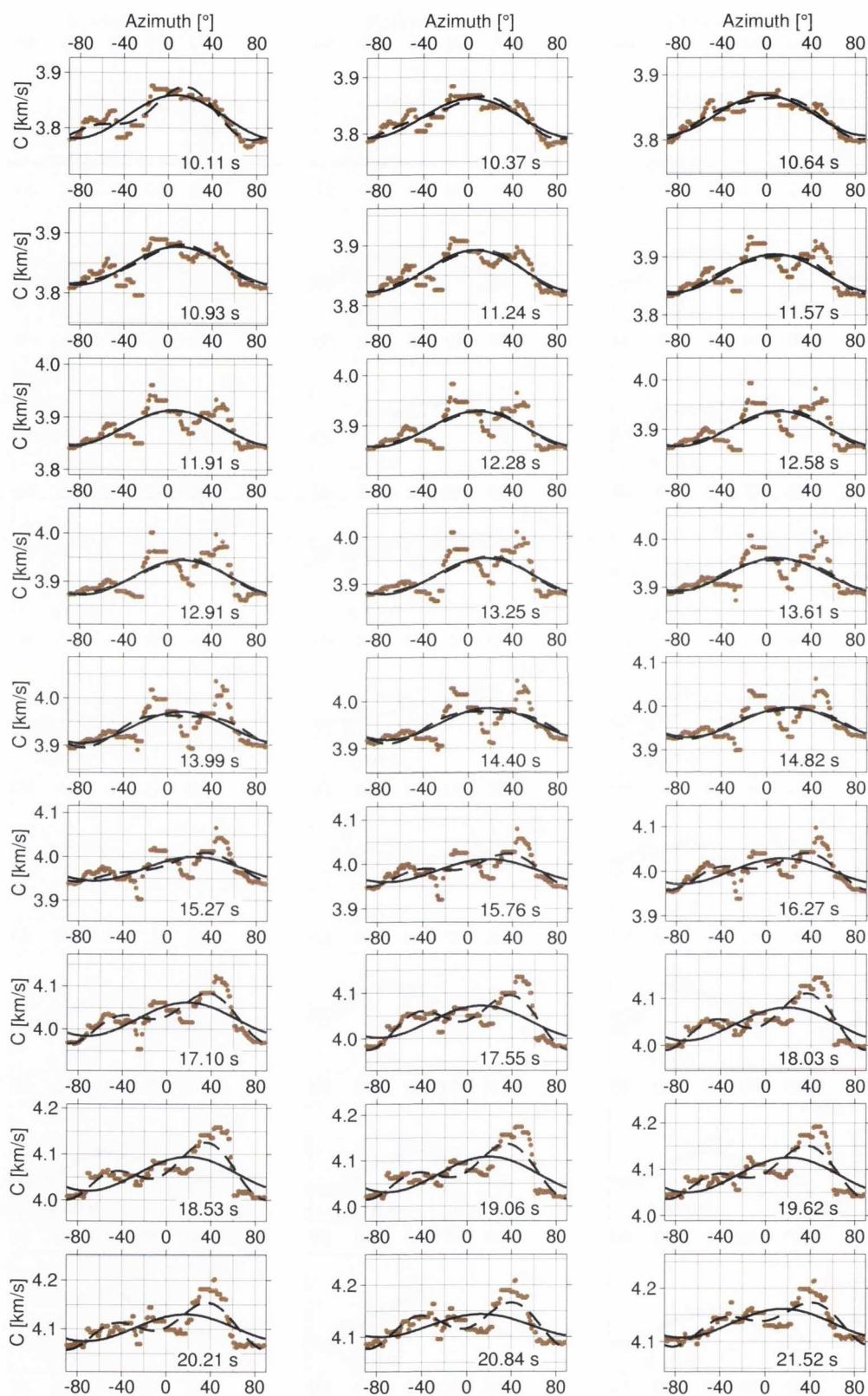


Distribution of Rayleigh-wave phase velocities with azimuth - central K.C. (page 5/5)

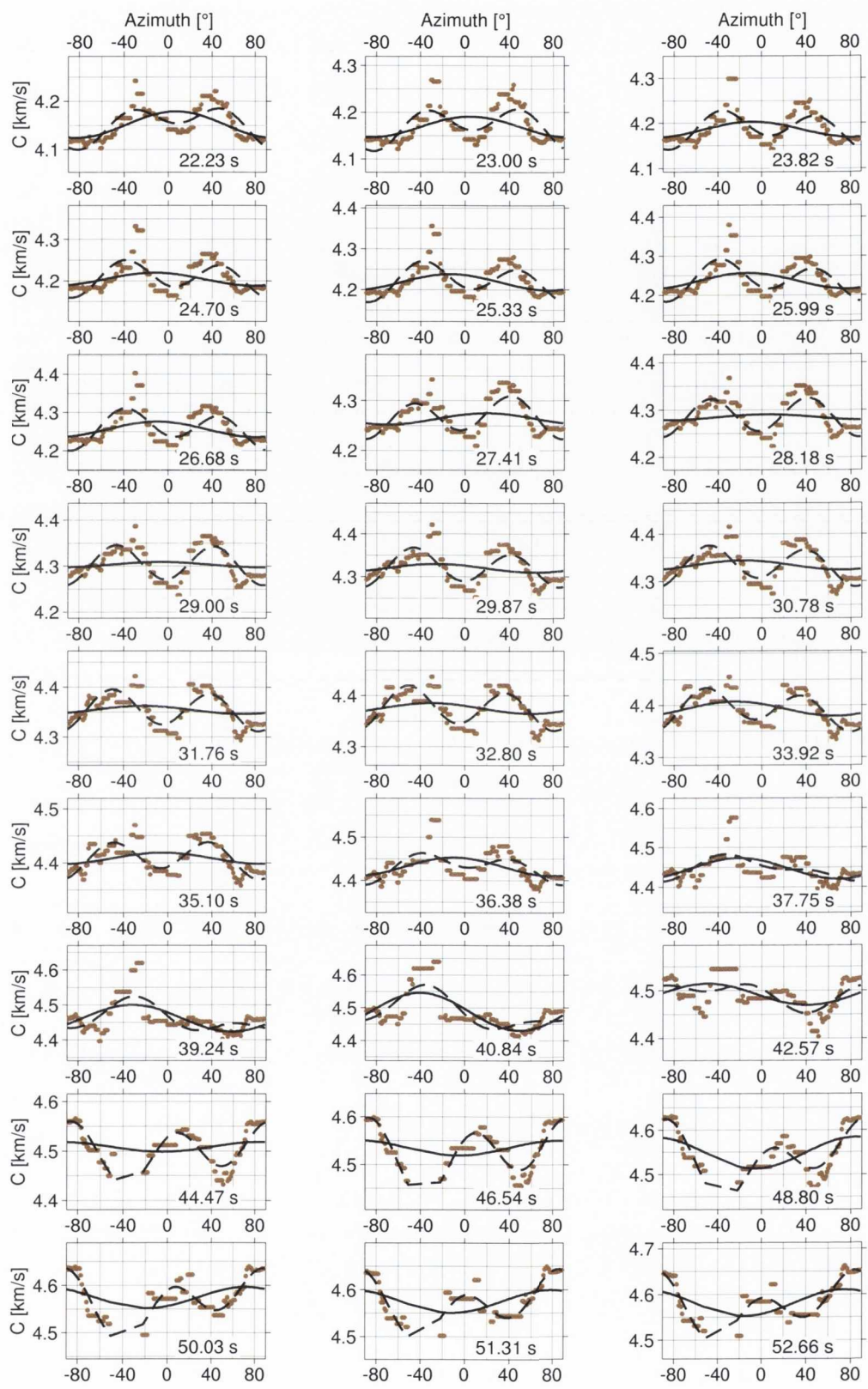


Distribution of Love-wave phase velocities with azimuth - central K.C. (page 1/5)

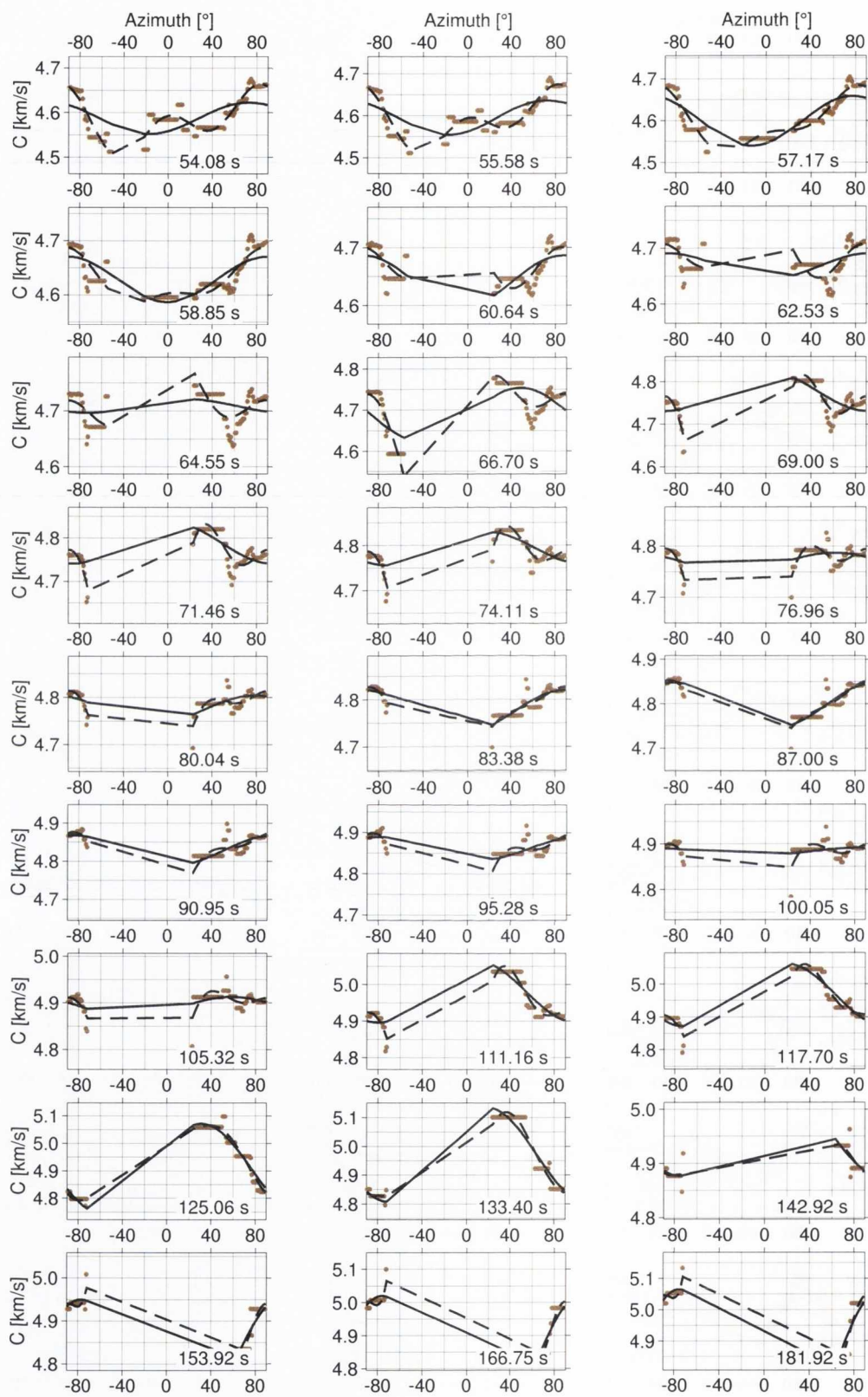




Distribution of Love-wave phase velocities with azimuth - central K.C. (page 2/5)

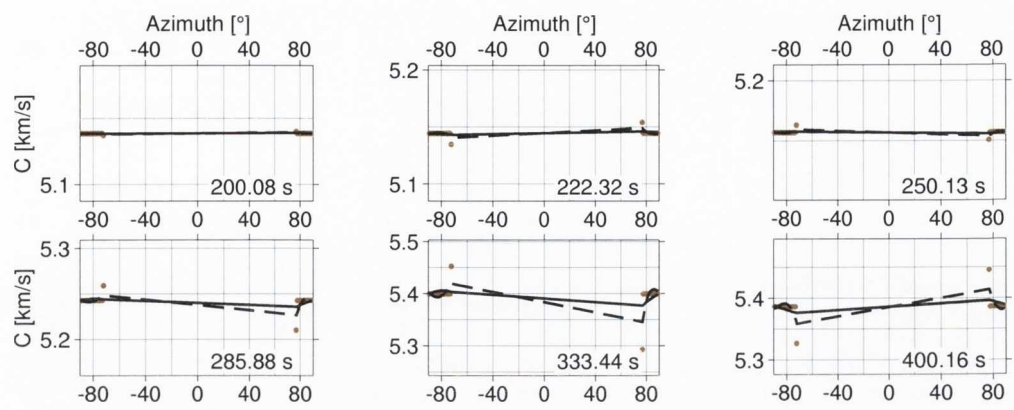


Distribution of Love-wave phase velocities with azimuth - central K.C. (page 3/5)



Distribution of Love-wave phase velocities with azimuth - central K.C. (page 4/5)





Distribution of Love-wave phase velocities with azimuth - central K.C. (page 5/5)

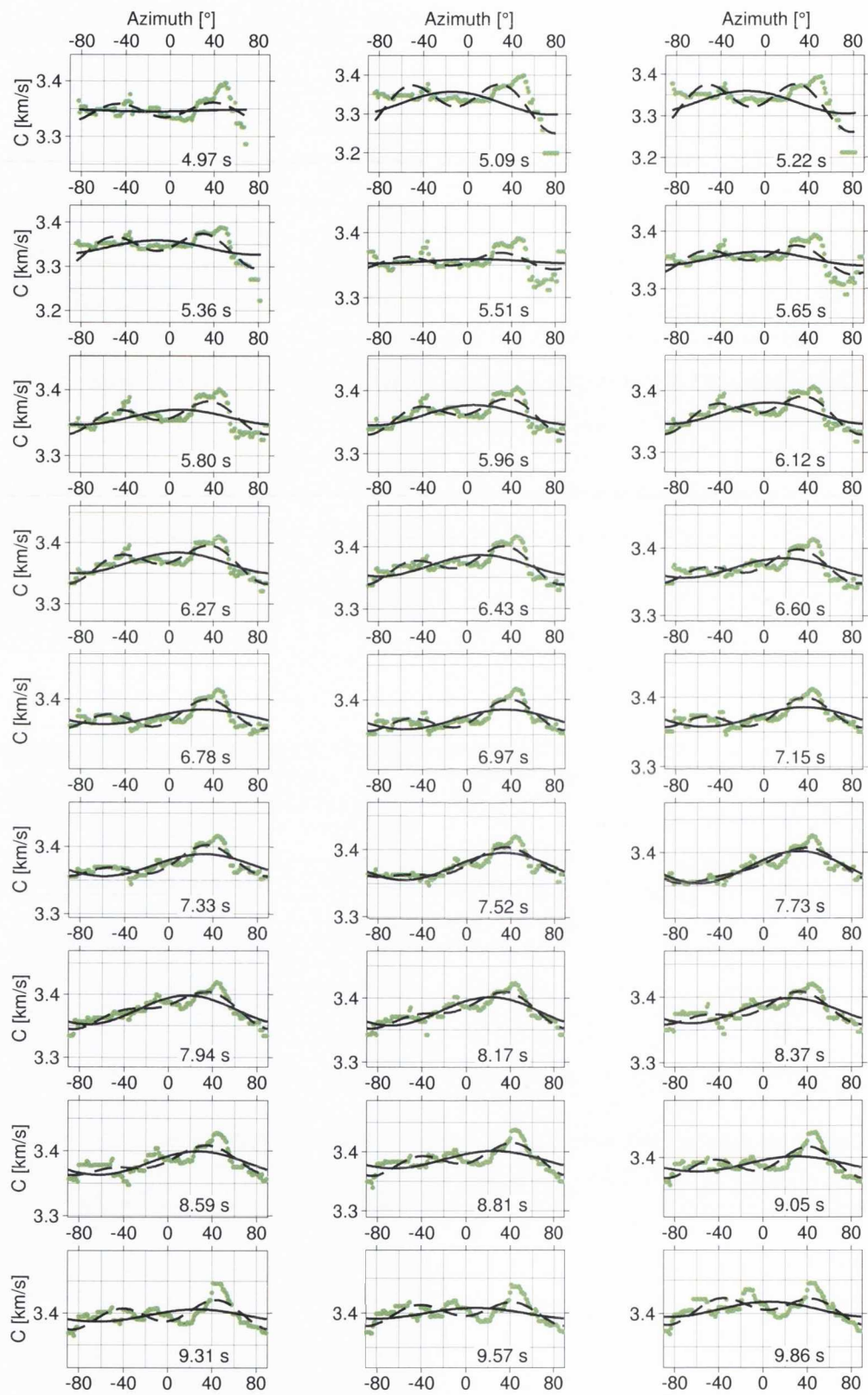
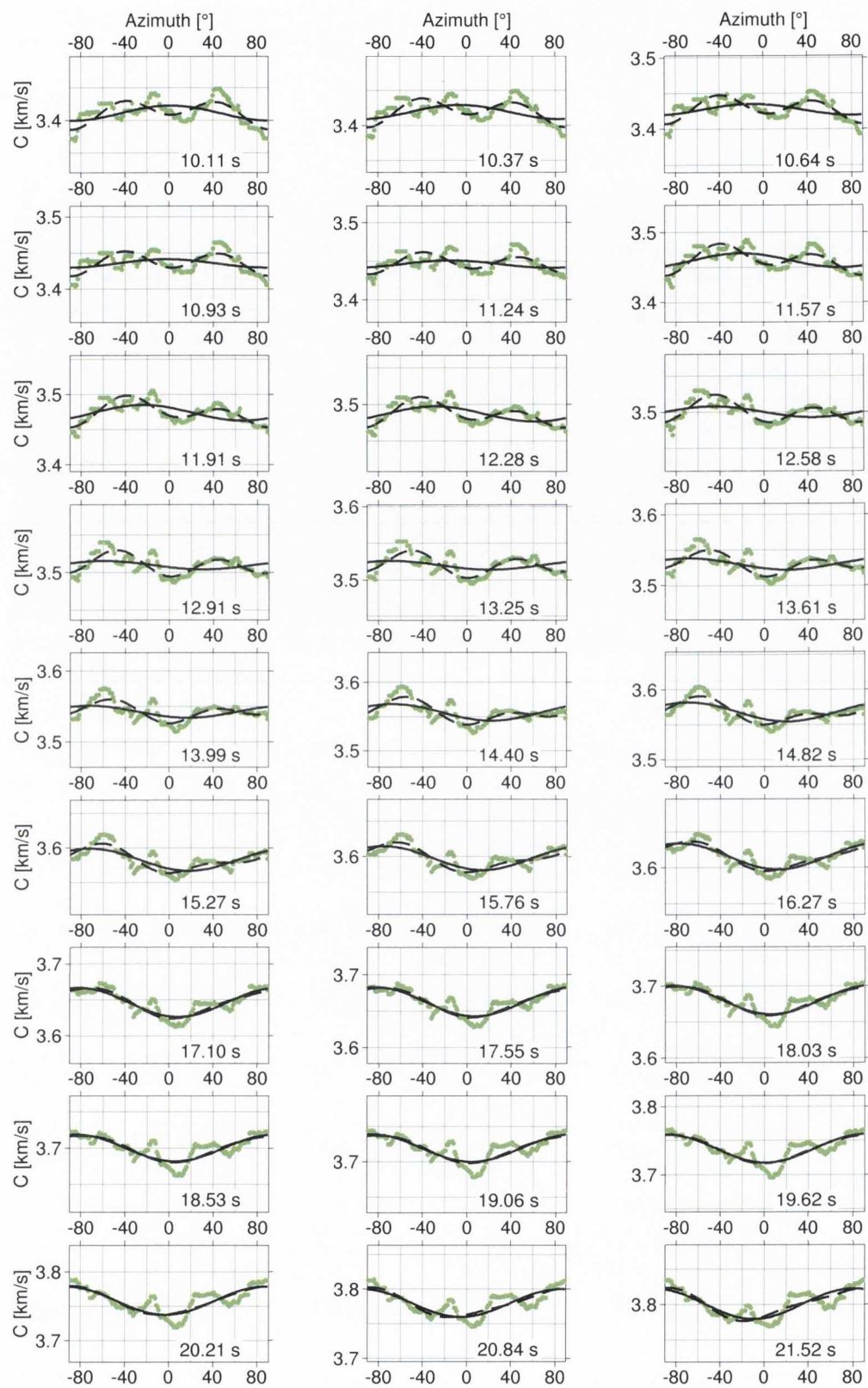
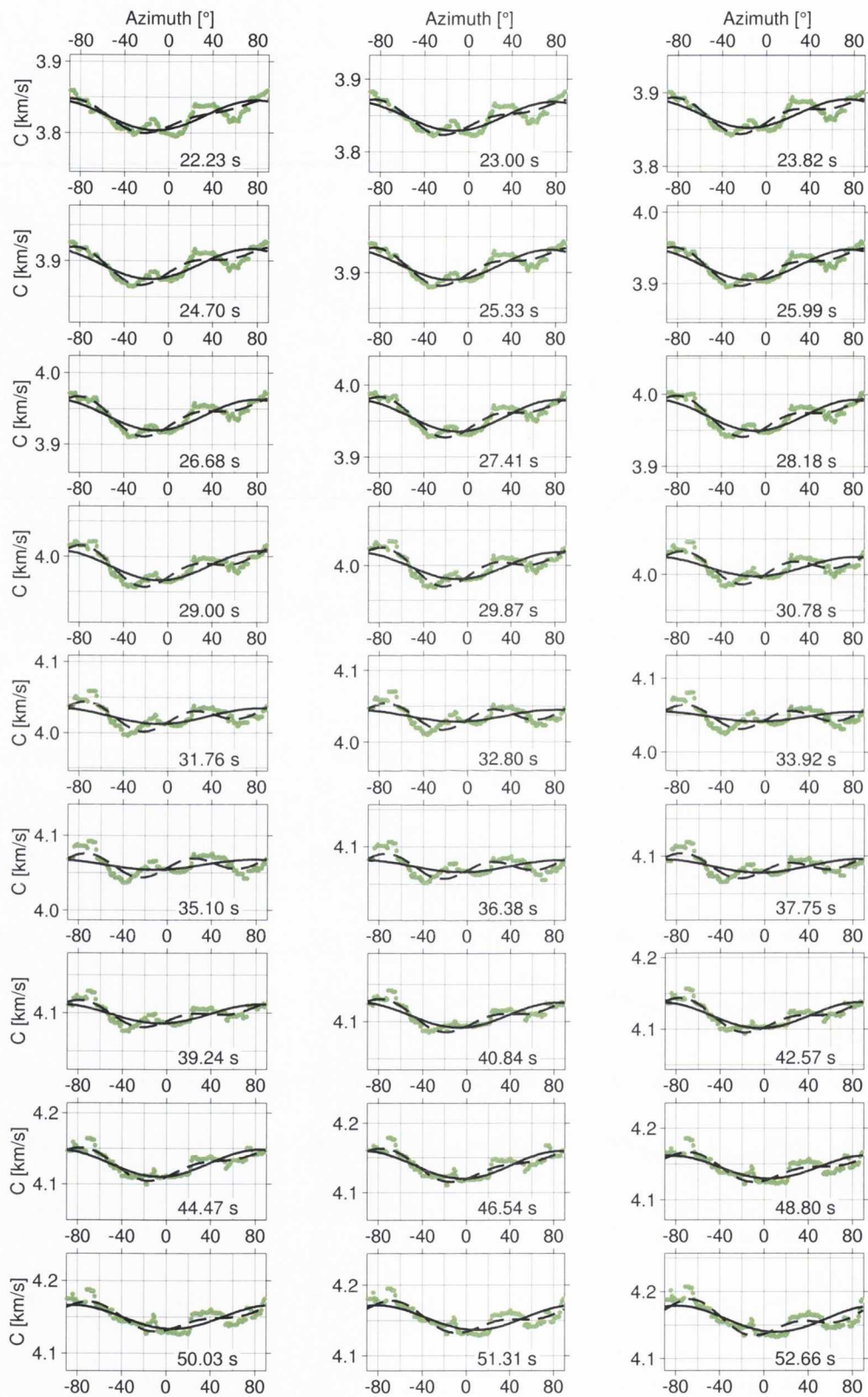


Figure A.4.: Distribution of Rayleigh-wave phase velocities with azimuth - South-West K.C. (page 1/5)

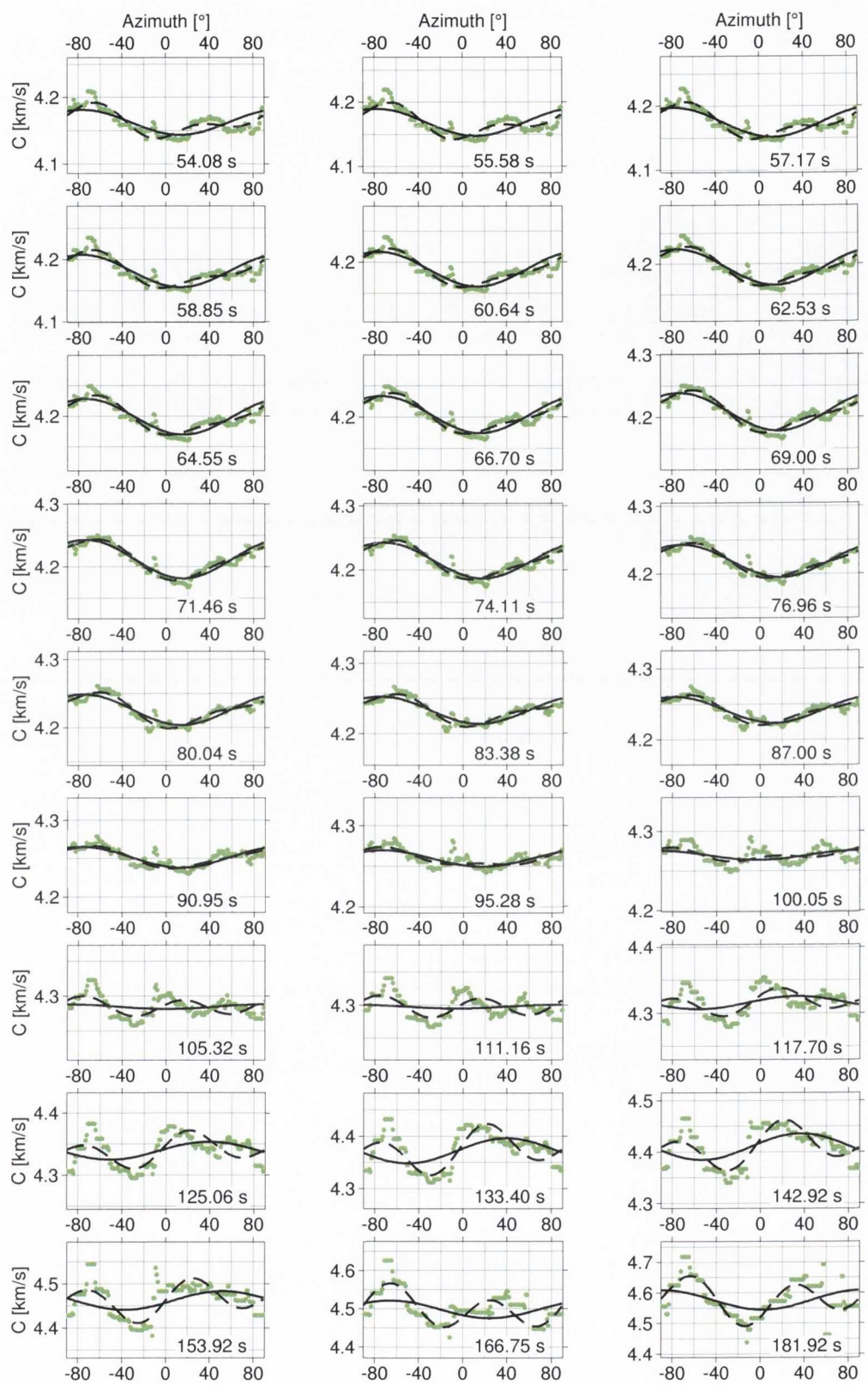


Distribution of Rayleigh-wave phase velocities with azimuth - South-West K.C. (page 2/5)

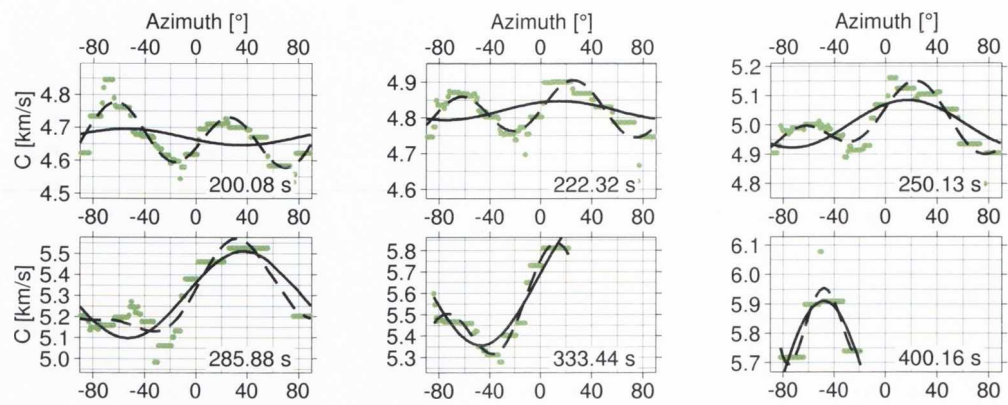




Distribution of Rayleigh-wave phase velocities with azimuth - South-West K.C. (page 3/5)

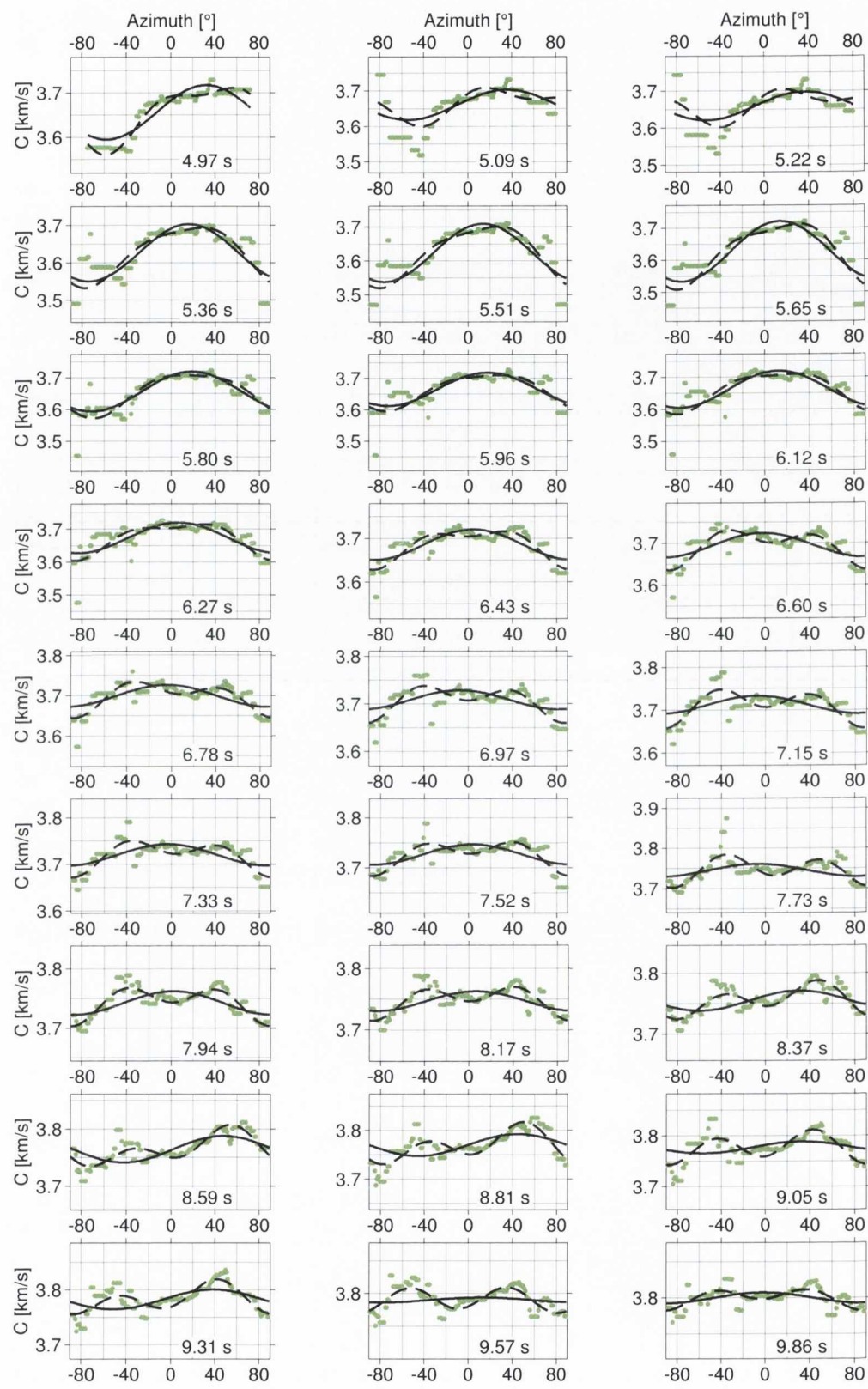


Distribution of Rayleigh-wave phase velocities with azimuth - South-West K.C. (page 4/5)

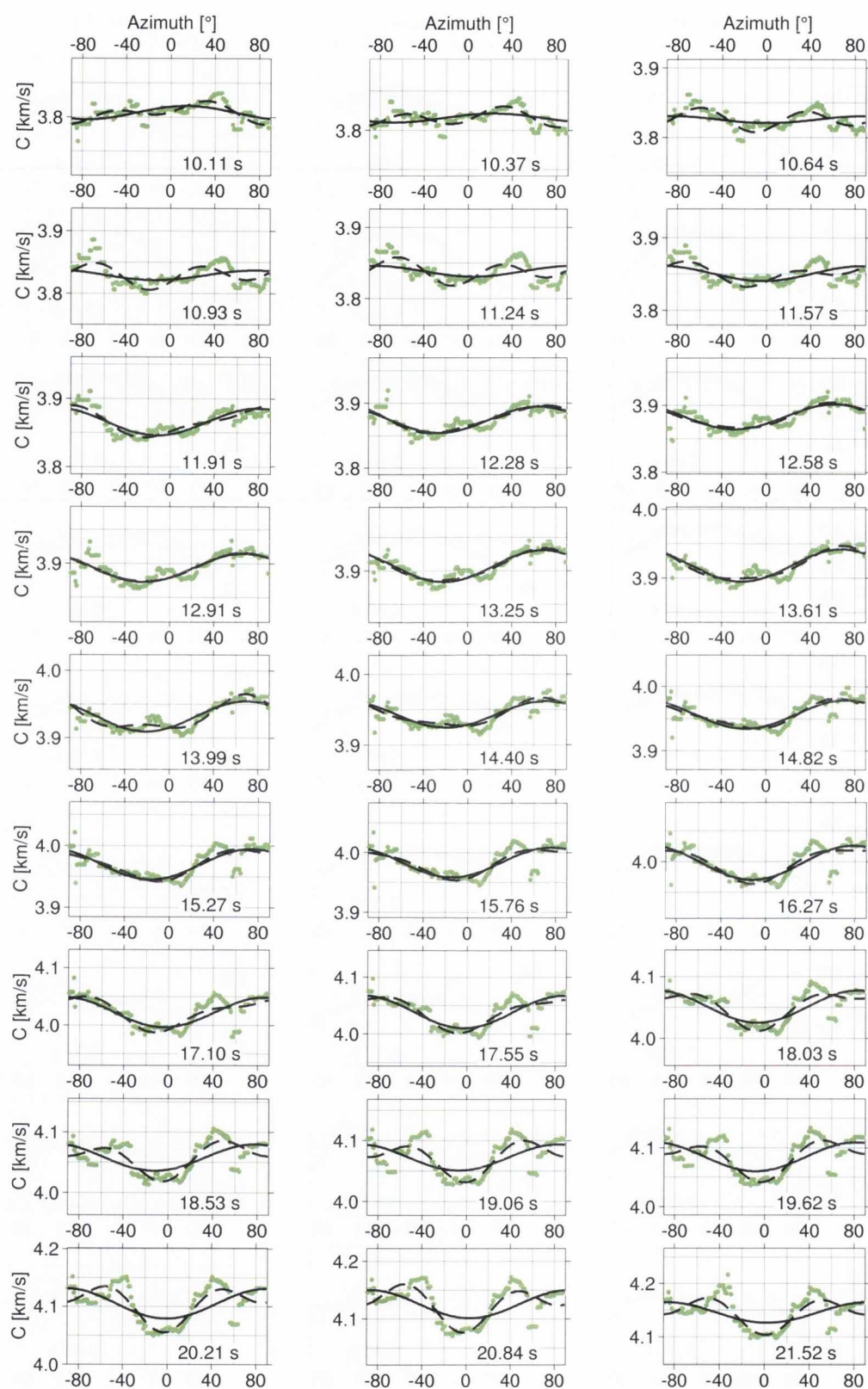


Distribution of Rayleigh-wave phase velocities with azimuth - South-West K.C. (page 5/5)

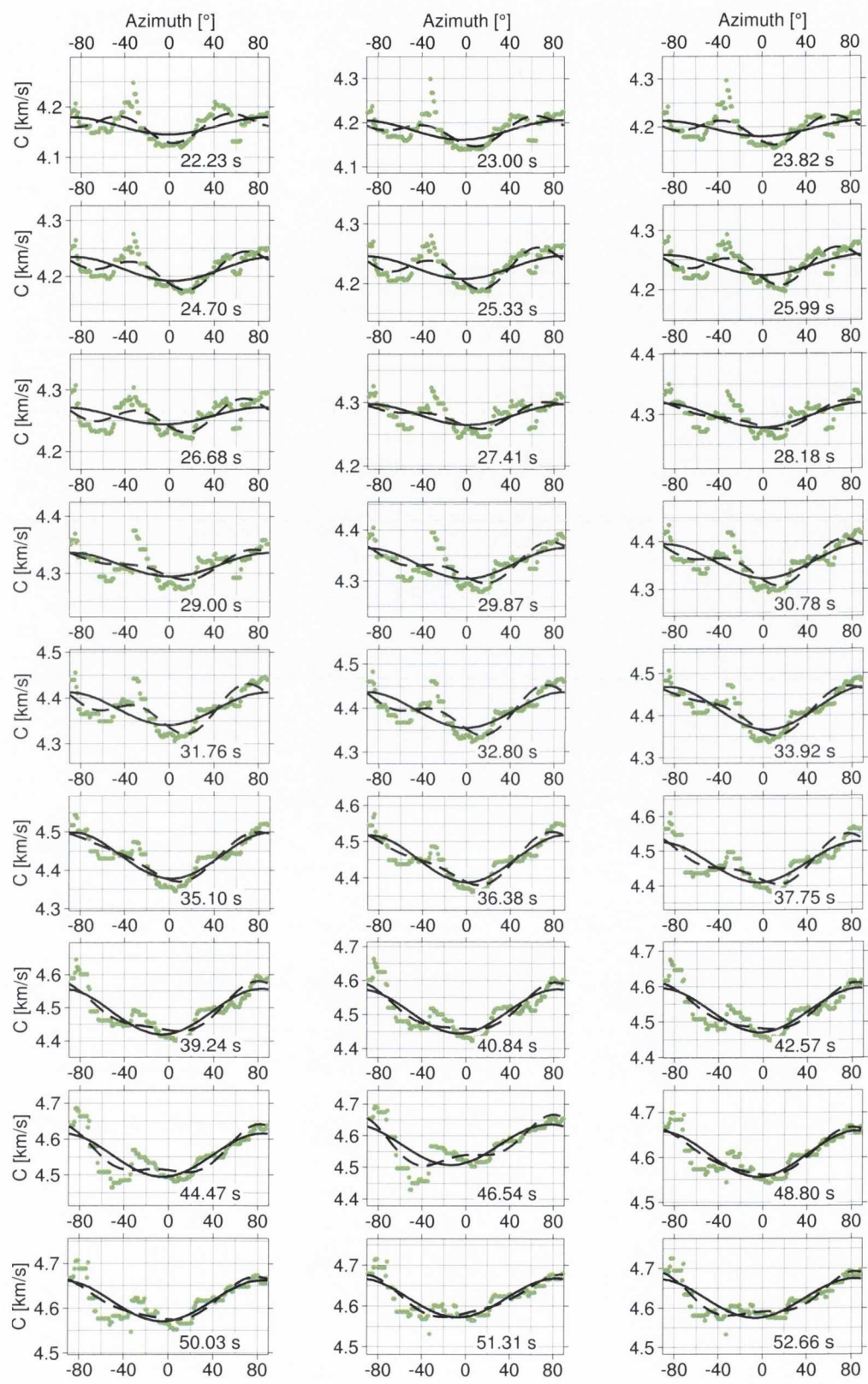




Distribution of Love-wave phase velocities with azimuth - South-West K.C. (page 1 / 5)

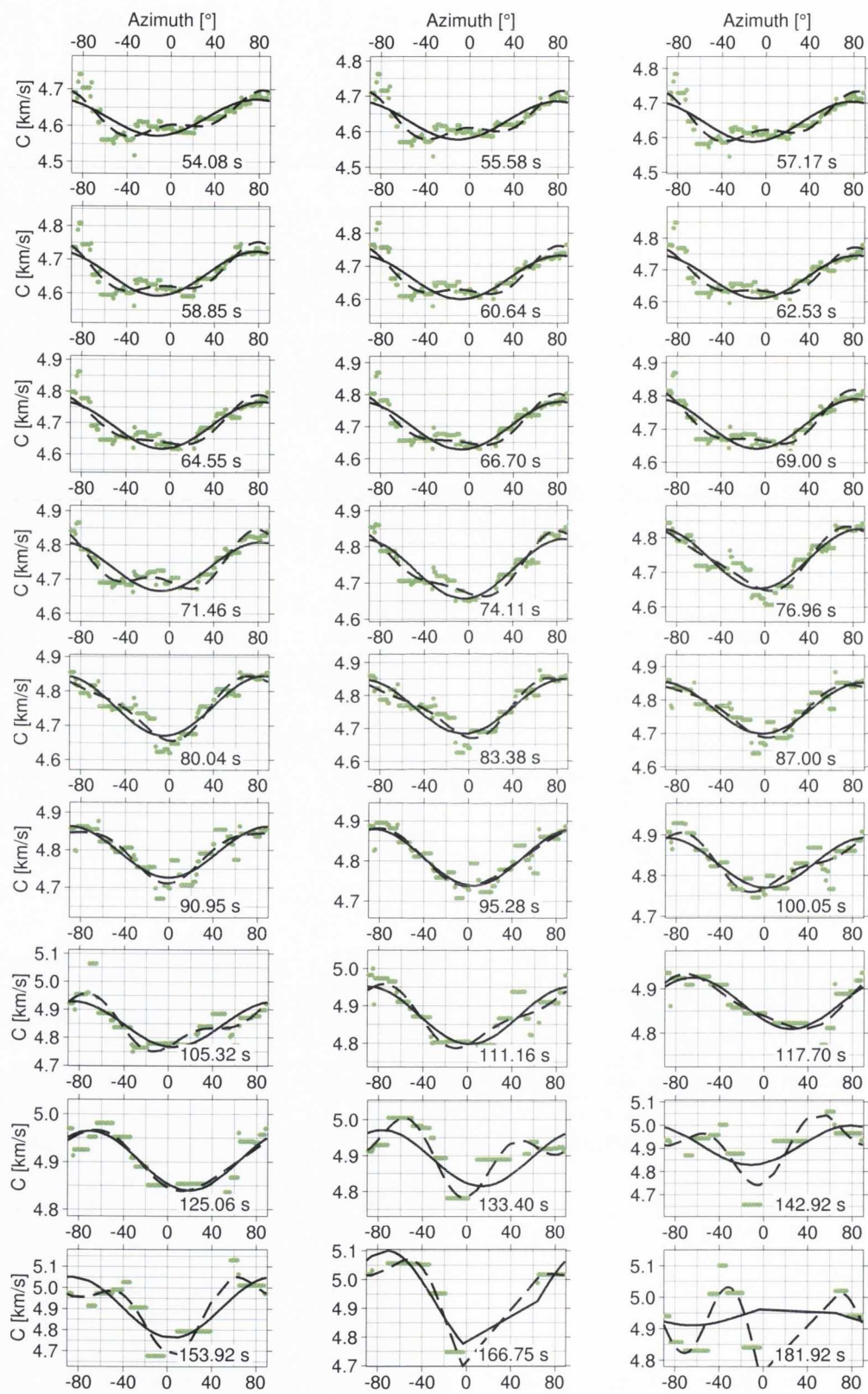


Distribution of Love-wave phase velocities with azimuth - South-West K.C. (page 2/5)

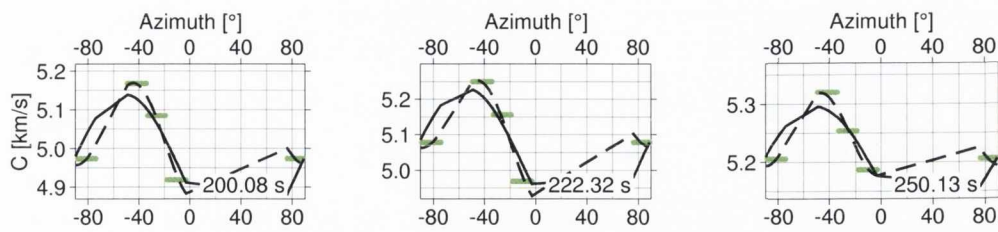


Distribution of Love-wave phase velocities with azimuth - South-West K.C. (page 3/5)





Distribution of Love-wave phase velocities with azimuth - South-West K.C. (page 4/5)



Distribution of Love-wave phase velocities with azimuth - South-West K.C. (page 5/5)

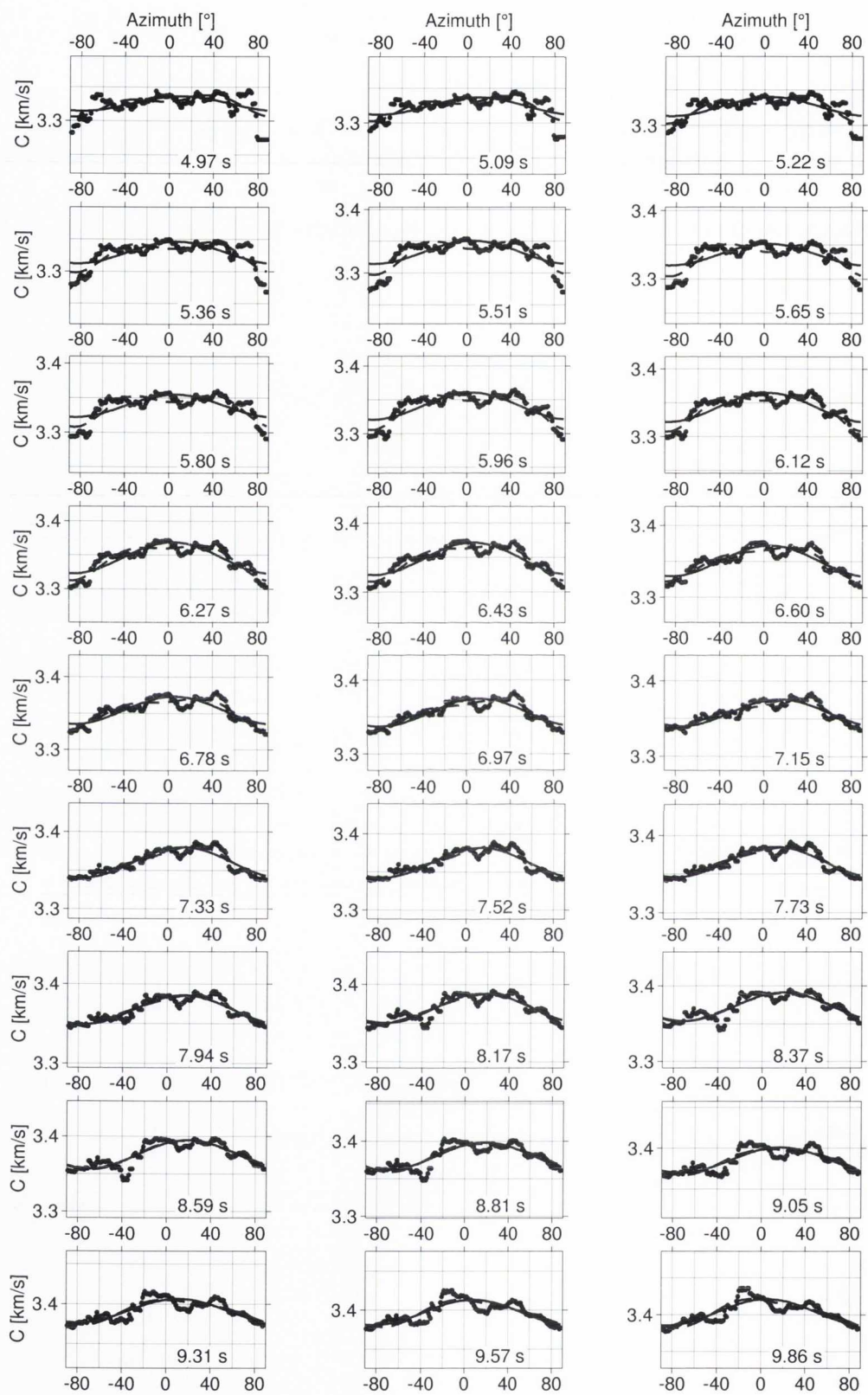
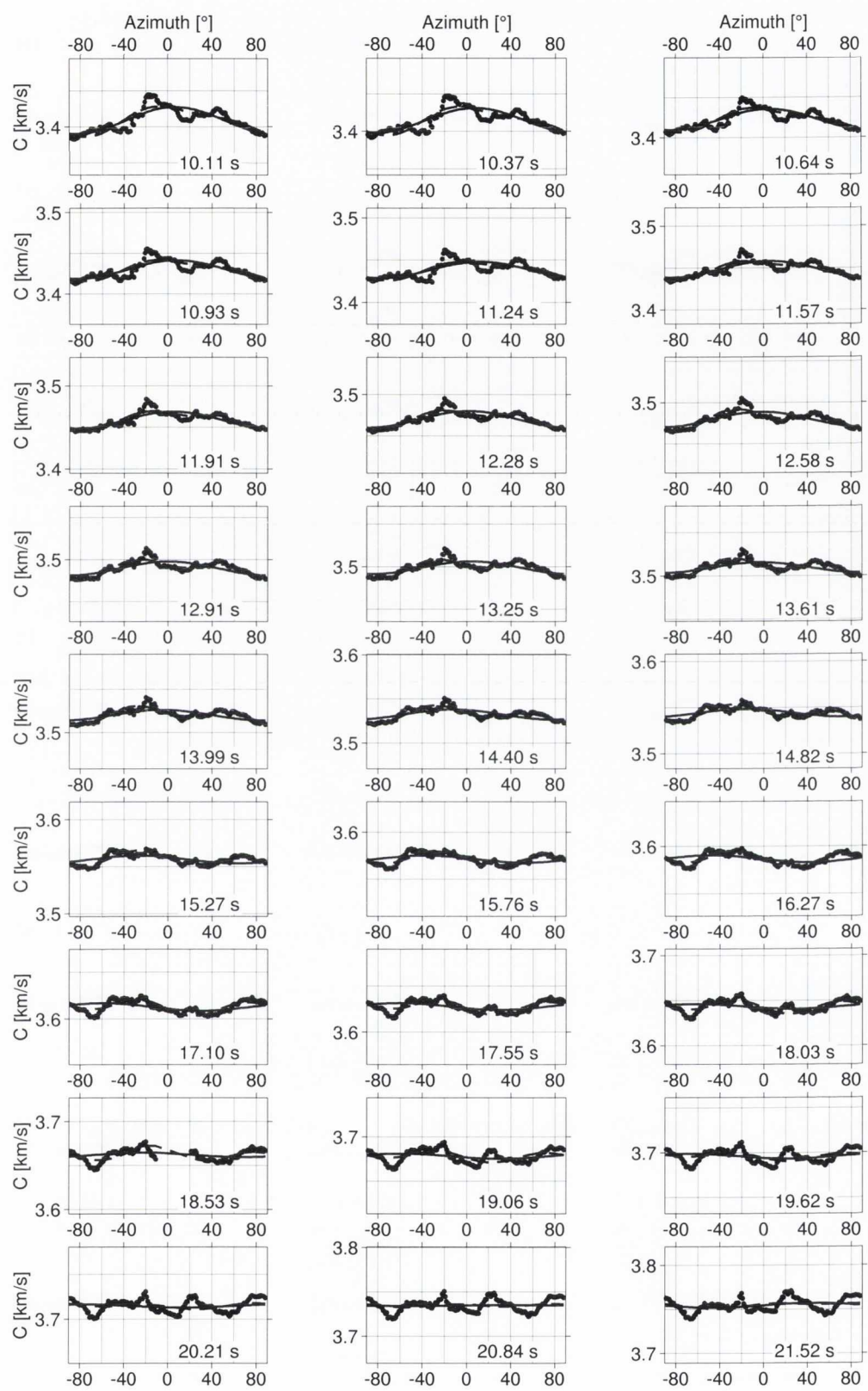
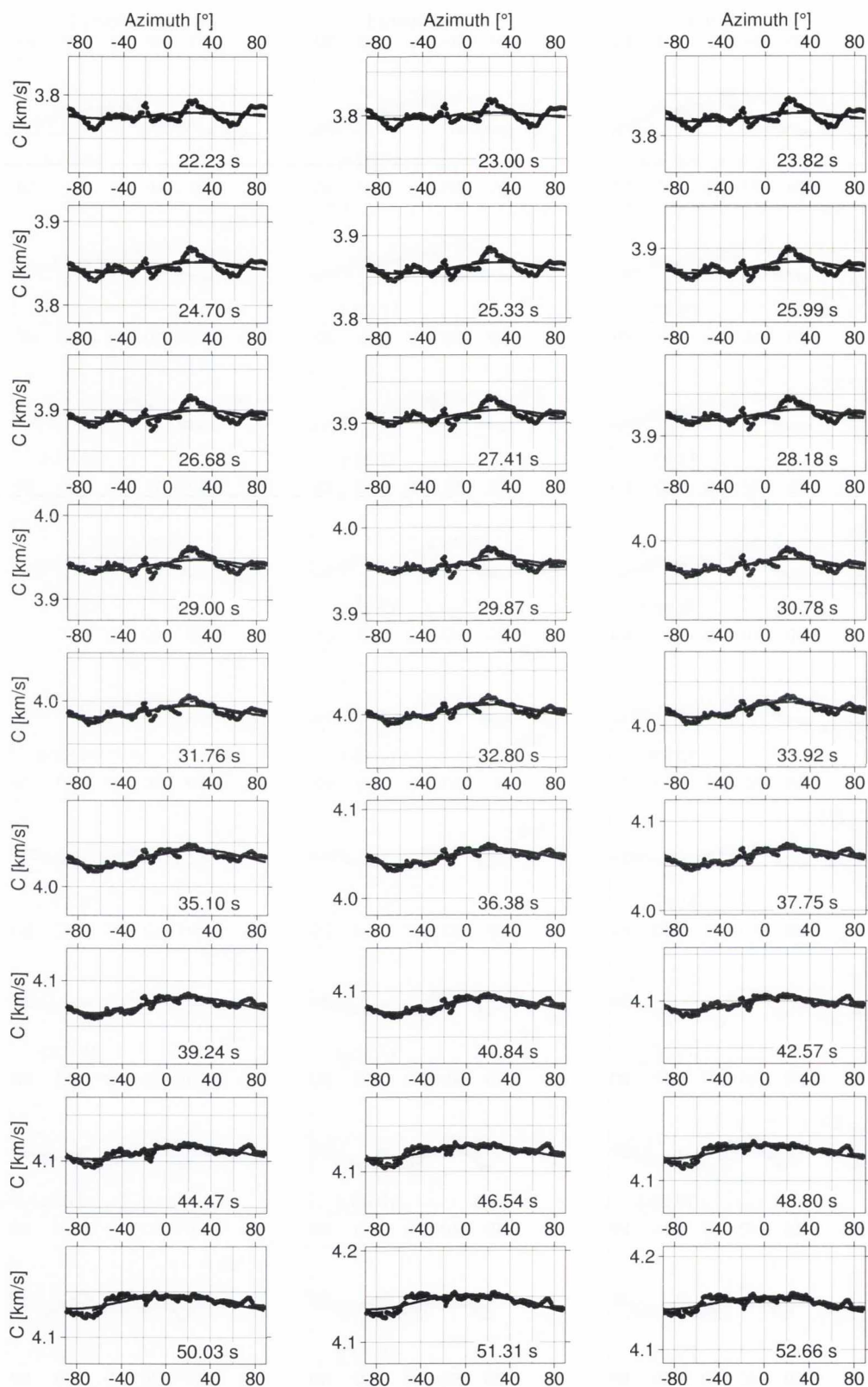


Figure A.5.: Distribution of Rayleigh-wave phase velocities with azimuth - all-region-average (page 1/5)

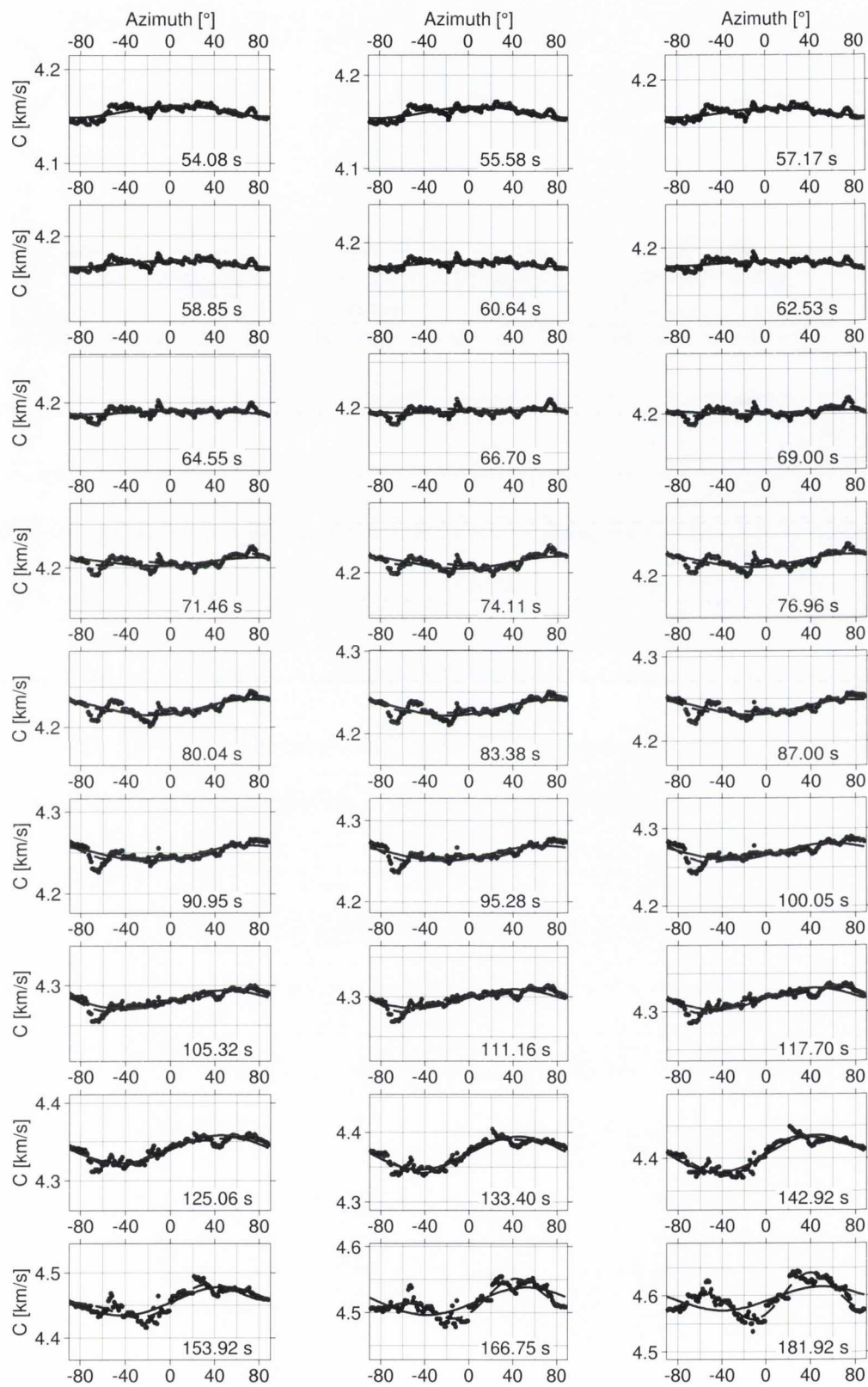




Distribution of Rayleigh-wave phase velocities with azimuth - all-region-average (page 2/5)

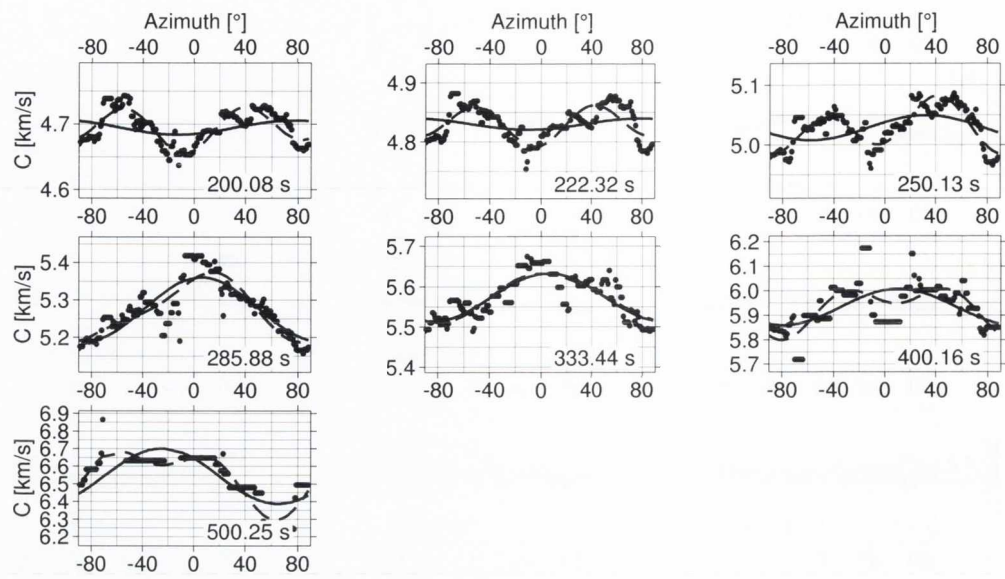


Distribution of Rayleigh-wave phase velocities with azimuth - all-region-average (page 3/5)

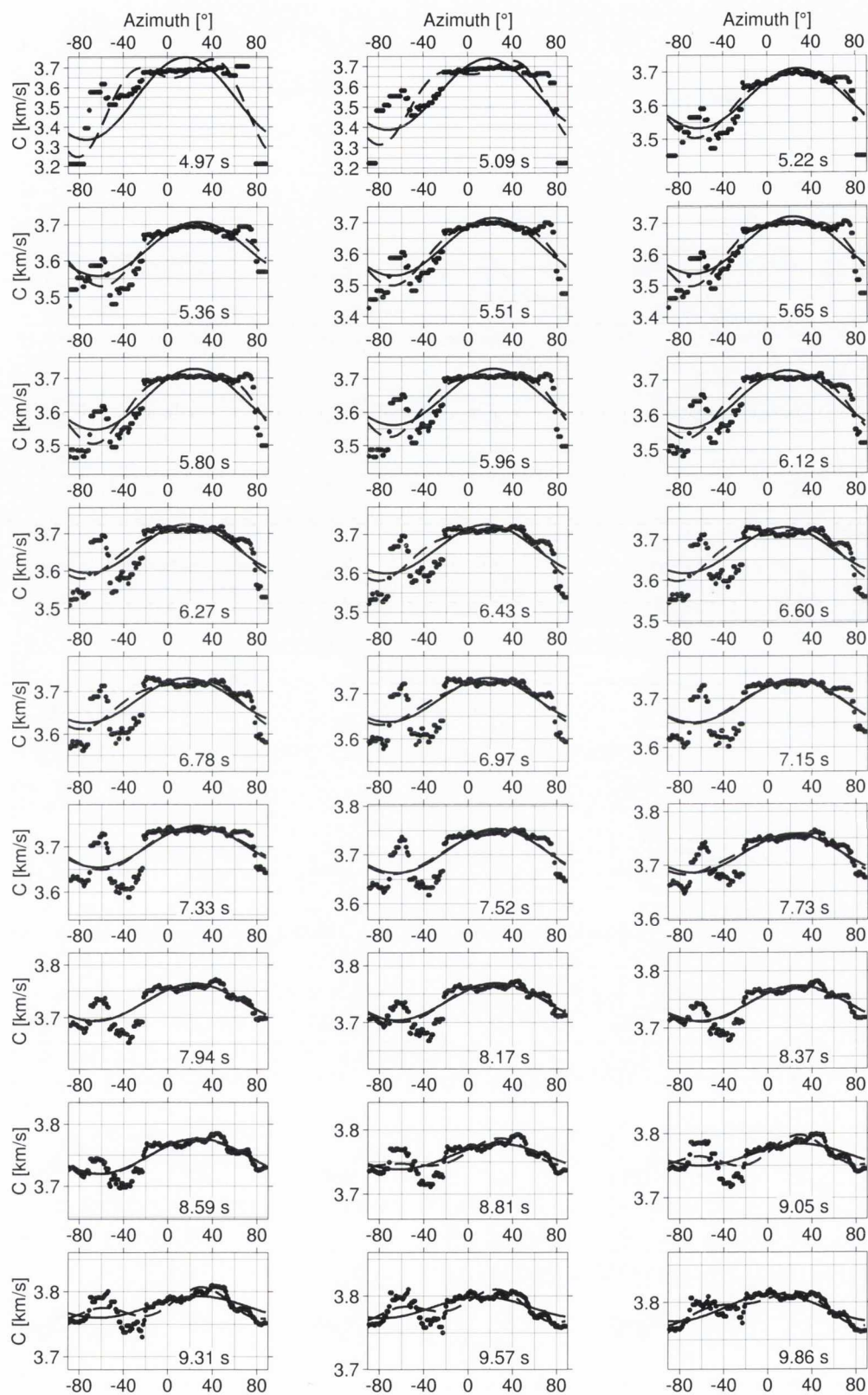


Distribution of Rayleigh-wave phase velocities with azimuth - all-region-average (page 4/5)

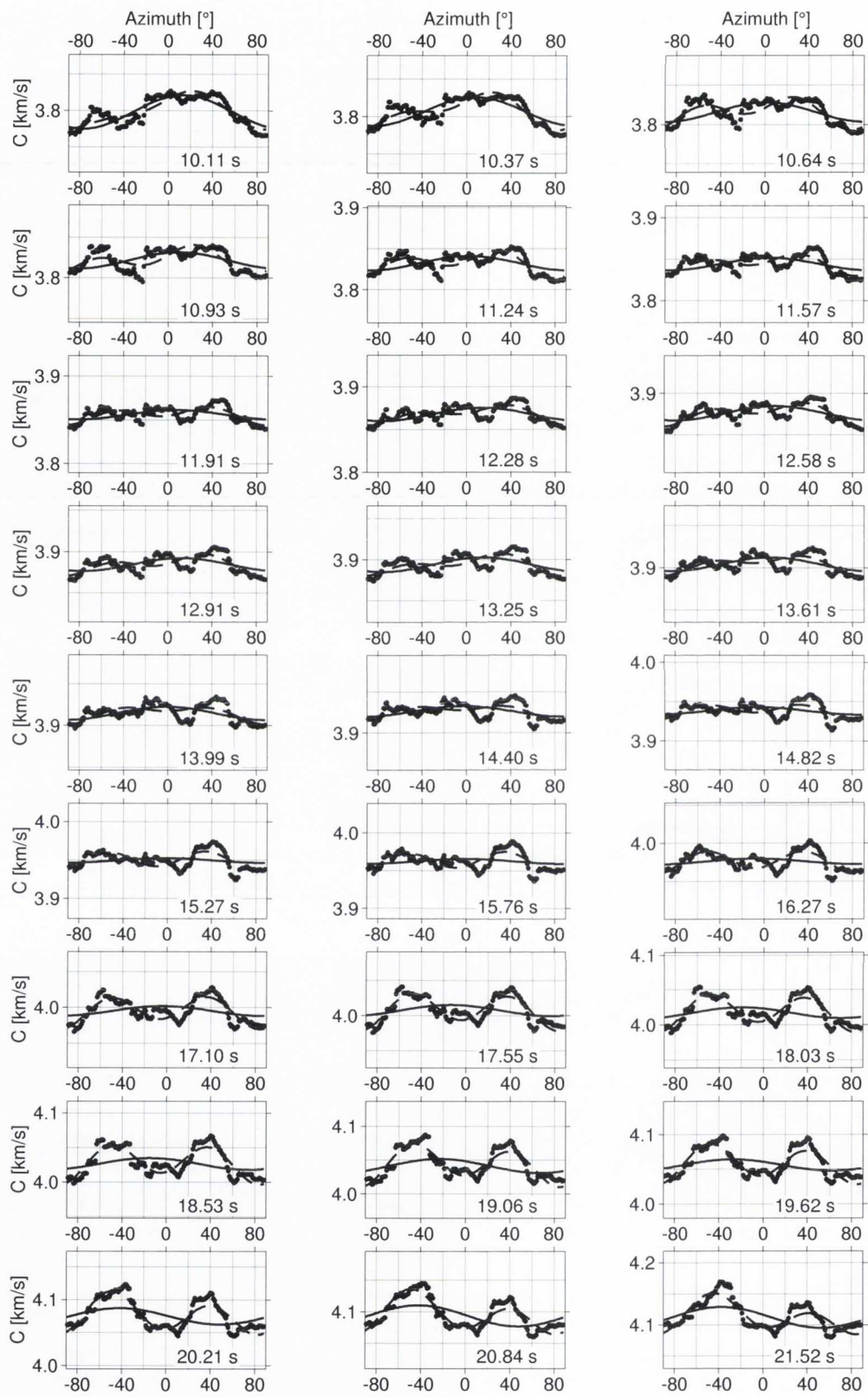




Distribution of Rayleigh-wave phase velocities with azimuth - all-region-average (page 5/5)

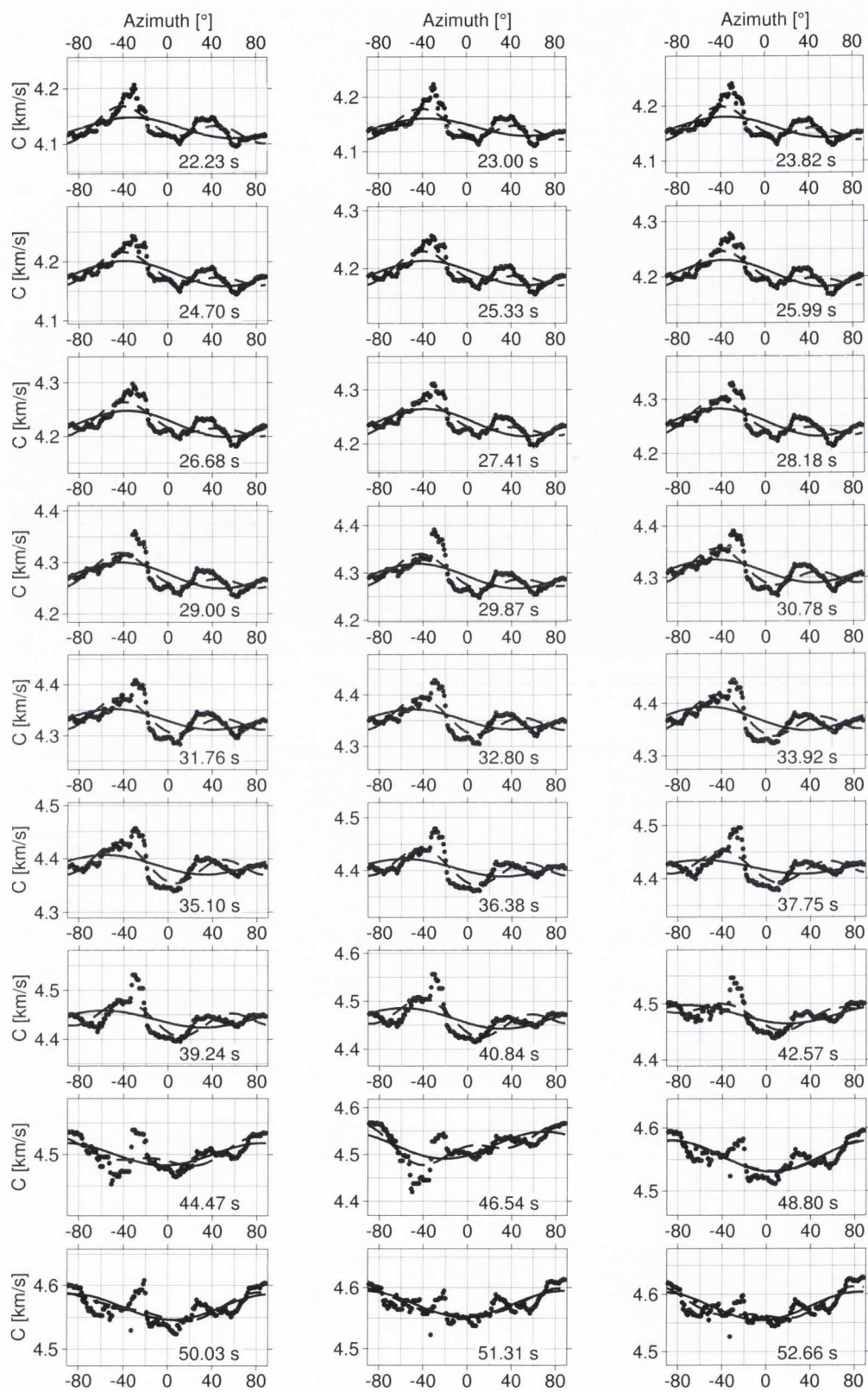


Distribution of Love-wave phase velocities with azimuth - all-region-average (page 1/5)

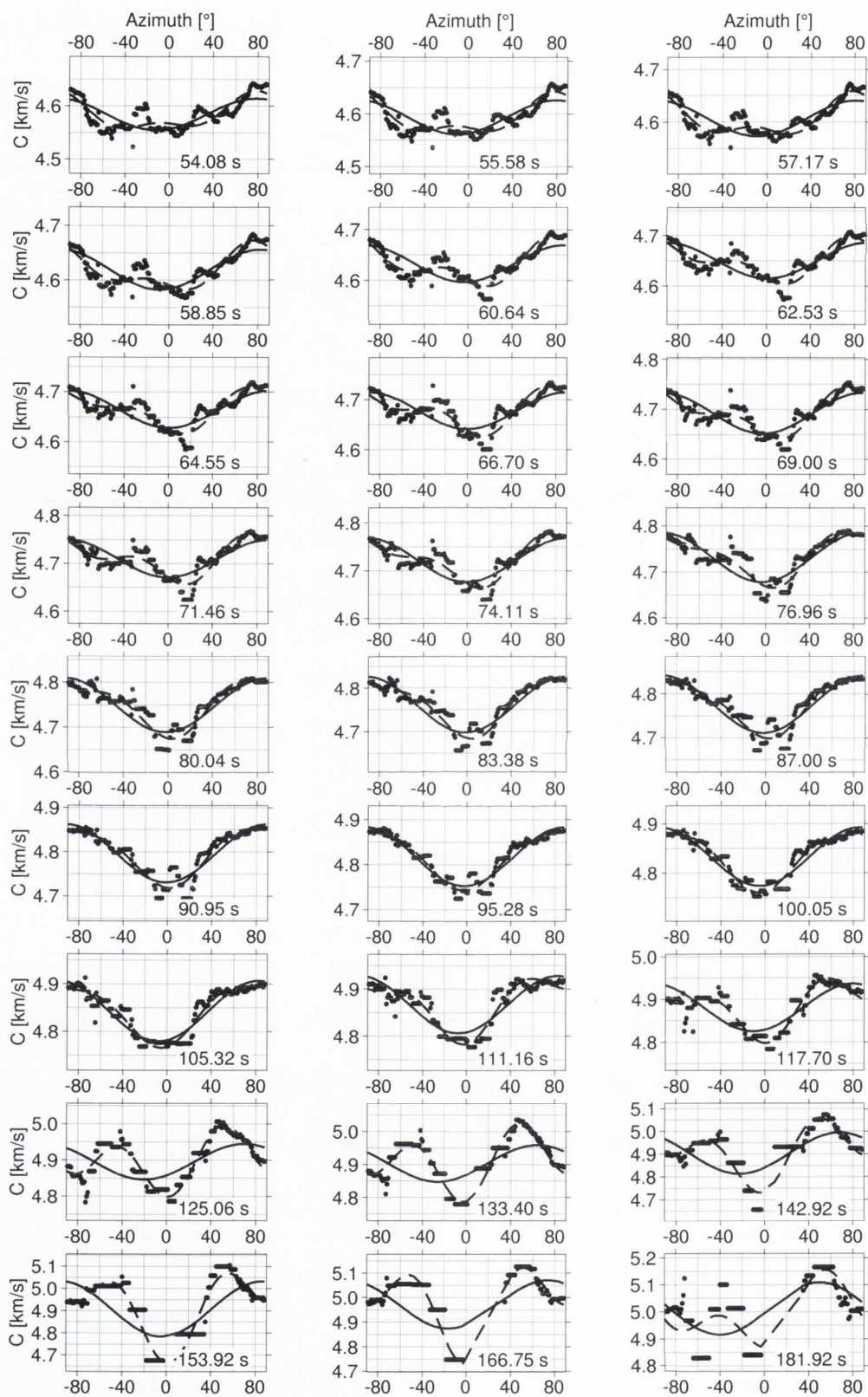


Distribution of Love-wave phase velocities with azimuth - all-region-average (page 2/5)

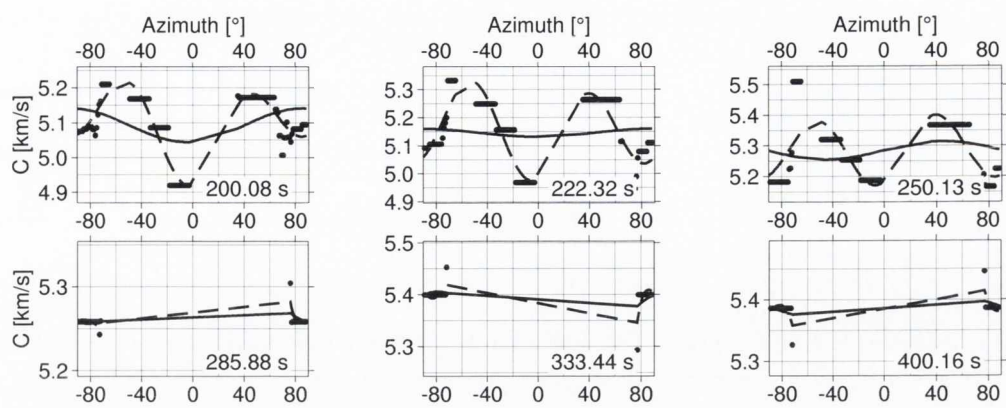




Distribution of Love-wave phase velocities with azimuth - all-region-average (page 3/5)



Distribution of Love-wave phase velocities with azimuth - all-region-average (page 4/5)

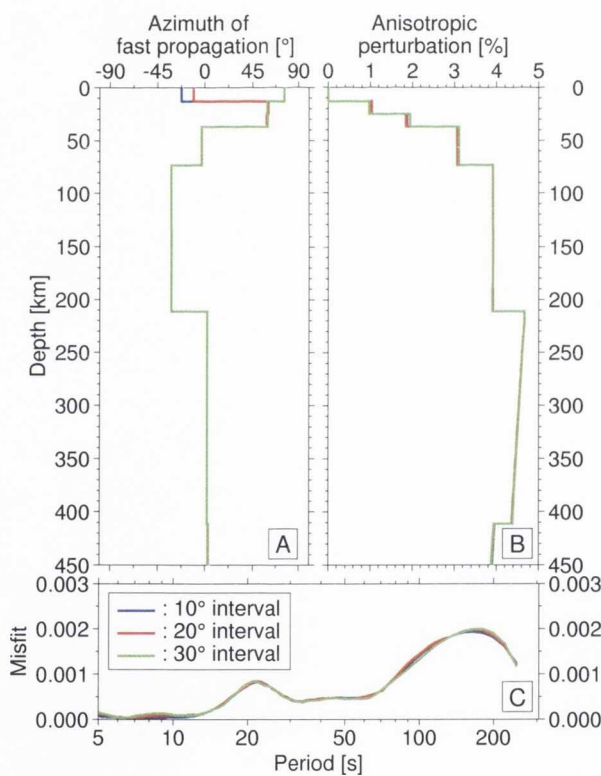


Distribution of Love-wave phase velocities with azimuth - all-region-average (page 5/5)



AZIMUTHAL SAMPLING IN INVERSIONS FOR DEPTH PROFILES

In inversions for depth profiles of shear velocity and azimuthal anisotropy, we computed phase velocities at 20° intervals (*i.e.* at -90°, -70°, -50°, -30°, -10°, 10°, 30°, 50° and 70°). We present in this appendix an exemple of synthetic data fits with three different intervals (10°, 20° and 30°) and show that the result do not depend on this criteria.



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## GROUP-VELOCITY SENSITIVITY KERNELS

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Comparison of the depth sensitivity of phase velocities (Figure 2.6) or group velocities (Figure C.1) with respect to shear velocity, compressional velocity and density shows that both phase and group velocities are sensitive to the same depth-range. Also, a remarkable feature is the alternation of positive and negative sensitivity of group velocities with respect to shear and compressional waves. Furthermore, group velocity sensitivity kernel curves show an additional zero. As the sensitivity kernels of group velocities decreases with depth, its amplitude tends to  $0^-$  (*i.e.* tends to 0 at infinity but remains negative) while the sensitivity kernels of phase velocities tends to  $0^+$  (*i.e.* tends to 0 at infinity but remains positive).

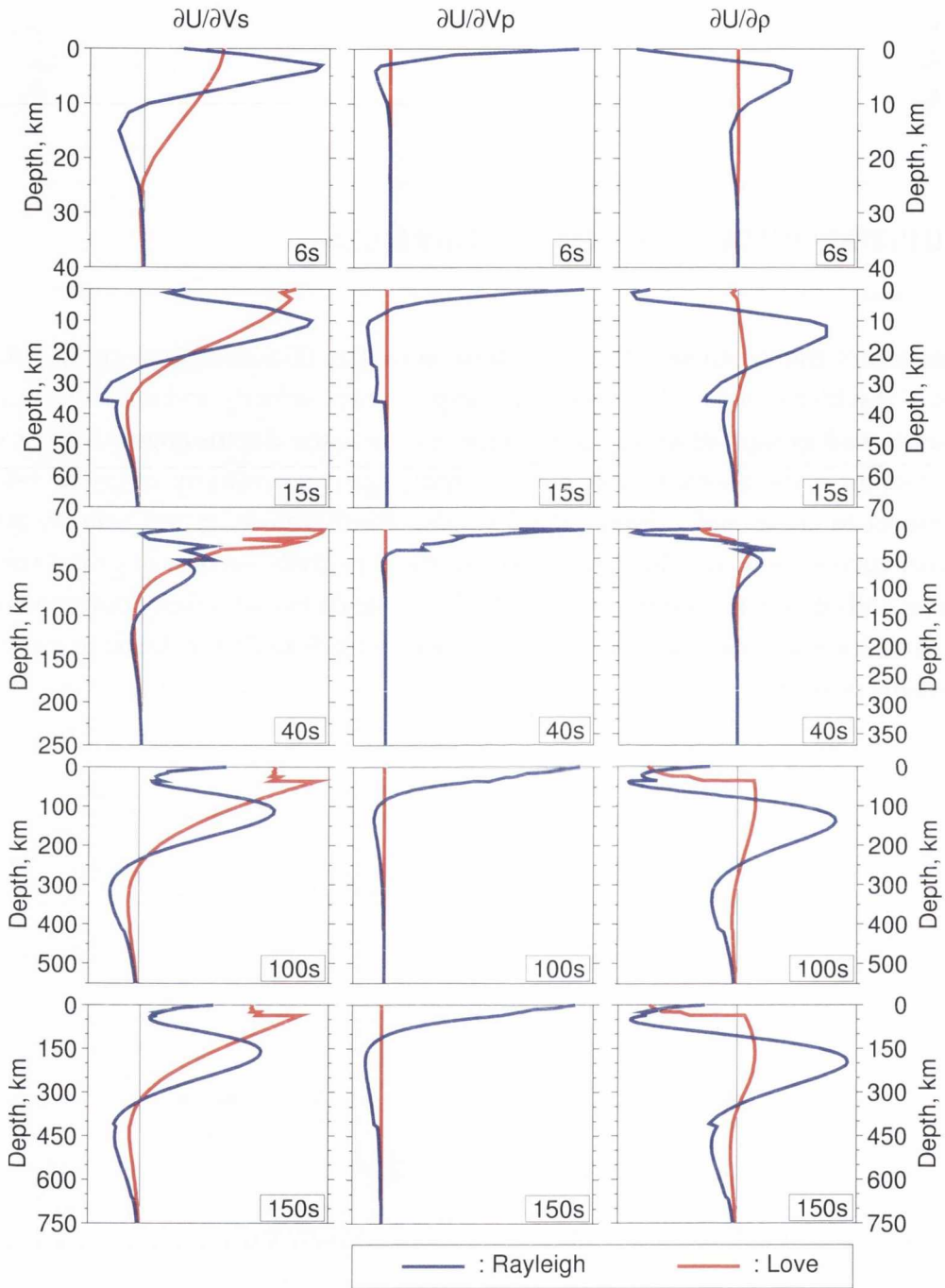


Figure C.1.: Depth sensitivity of surface waves (at 6, 15, 40, 100 and 150 s). The sensitivity curves are the Fréchet derivatives of the group velocities of the fundamental-mode Rayleigh and Love waves with respect to S-wave velocities ( $V_s$ ), P-wave velocities ( $V_p$ ) and density ( $\rho$ ). The derivatives were computed for a continental, 1-D Earth model with a 37 km thick crust, at 4 different periods. Each graph is scaled independently.



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## RESOLUTION AND LIMITS OF THE MOHO-DEPTH RESOLUTION

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### D.1 MOHO DEPTH RESOLUTION WITH NOISE IN THE DATA

To reduce the level of noise in the Rayleigh wave dispersion curves and better resolve the depth of the Moho, we smoothed the phase velocities using an under-damped gradient search. Results using real data show that smoothing helps to reduce the amplitude of the misfit values but may also be able to better constrain the depth of the Moho ([Figure 5.15](#), [Figure 5.16](#), [Figure 5.17](#) and [Figure 5.18](#)). However smoothing the data may introduce biases. We therefore estimated the noise in one of the sub-regions and applied it to a synthetic dispersion curve to confirm that using smoothed data does not alter the resolution of the depth of the Moho. We may in this section only discuss results using a correct starting model. The effects of using a non-correct starting model will be discussed in [Section D.2](#).

We first estimated the "high-frequency" noise as the residual between the measured dispersion curve and the smooth version of this curve. We secondly estimated the "complete estimated" noise as the residual between the measured dispersion curve and the dispersion curve resulting from an inversion for S-wave velocity that best fits the data. This "complete estimated" noise does account the "high-frequency" noise and also noise at broader periods and would represent the "full" noise present in the data. Finally, we approximated the "complete estimated" noise by a "ramp" function of two different maximum amplitudes.

Results in [Figure D.1](#) show that when inverting noisy data for the depth of the Moho, the positive and negative alternation patterns of phase-velocity perturbations are distorted and the amplitude of the misfit values are increased ([Figure 5.19](#)) (confirming the results using real data).

"High-frequency" noise does not seem to interfere with a good resolution of the depth of the Moho. Indeed, with only "high-frequency" noise, the correct depth of the Moho is well recovered. The "complete estimated" noise, however, seems to have a strong effect on

the resolvability of the depth of the Moho as it is estimated 10 km deeper than the correct depth. It is indeed the broader period noise in the "complete estimated" noise that alter the resolution in the manner that results using "ramp noise" (that approximates the broader period noise of the "complete estimated" noise with a simple "ramp") miss-estimated the depth of the Moho by few kilometers.

#### D.2 EFFECTS OF USING AN NON-ACCURATE STARTING MODEL

We have shown that using an inaccurate starting model can alter the resolvability of the depth of the Moho. In [Figure 5.19](#) and [Figure D.1](#) we present results of inversion for the depth of the Moho using synthetic data, with or without noise. Similar synthetic tests have been proceed with incorrect starting S velocity models (denoted as "wrong starting model"). Results show that the depth of the Moho can be miss-estimated by up to 5 km too deep. Surprisingly, even when using perfect (non-noisy) data, the depth of the Moho is not resolved perfectly. It is then very important, while investigating the depth of the Moho, to have a good *a priori* knowledge of the crustal and upper-mantle structures.

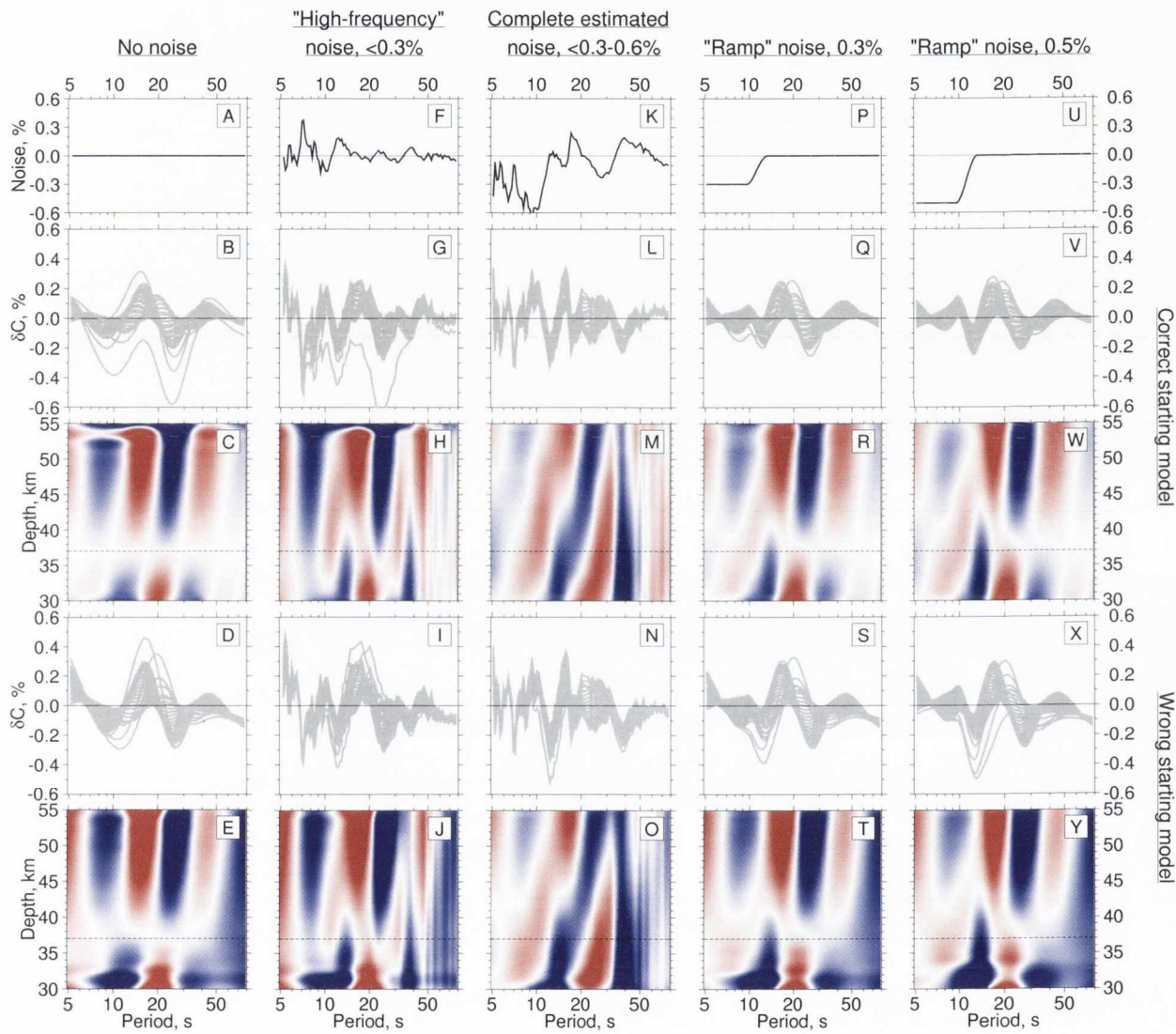


Figure D.1.: Synthetic tests for the resolution of the depth of the Moho with noise in the data and considering whether a correct (B,C,G,H,L,M,Q,R,V,W) or wrong starting model (D,E,I,J,N,O,S,T,X,Y). See text for explanations on how the noise distribution have been estimated.



*La science, mon garçon, est faite d'erreurs,  
mais d'erreurs qu'il est bon de commettre,  
car elles mènent peu à peu à la vérité.*

*[Science, my lad, is made up of mistakes,  
but they are mistakes which it is useful to make,  
because they lead little by little to the truth.]*

— Jules Verne —