

A Teleseismic Receiver Function Study of the Crustal Structure of the British Isles

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by

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Abstract

Teleseismic Receiver Function Study of the Crustal Structure of the British Isles

J. P. Tomlinson

The onshore crustal and upper mantle velocity structure of the British Isles has been investigated by teleseismic receiver function analysis. The results of the study augment the dense offshore and sparse onshore models of the structure beneath the area. In total almost 1500 receiver functions have been analysed, which have been calculated using teleseismic data from 34 broadband and short-period, three-component seismic recording instruments. The crustal structure has primarily been investigated using 1D grid search and forward modelling techniques, returning crustal thicknesses, bulk crustal V_p/V_s ratio and velocity-depth models. Upper mantle structures have been investigated by applying P_s moveout corrections and migration techniques to the observed broadband receiver functions. $H-\kappa$ stacking reveals crustal thicknesses between 25-36 km and V_p/V_s ratios between 1.6-1.9. The crustal thicknesses correlate with the results of previous seismic reflection and refraction profiles to within ± 2 km. The exceptions are the stations close to the lapetus suture where the receiver function crustal thicknesses are up to 5 km less than the seismic refraction Moho. This mismatch has been attributed to the presence of underplated magmatic material at the base of the crust. 1D forward modelling has revealed sub-crustal structures. In northern Scotland these correspond with the Flannan and W-reflectors. The isolated sub-crustal structure at station GIM on the Isle of Man may be related to the closure of the lapetus ocean. P_s conversion from the 410 km and 660 km discontinuities have been identified in the P_s moveout corrected receiver functions. The differential delay time between the phases is close to the global average of 24s, indicating that there is no significant thermal anomaly in the mantle transition zone beneath the British Isles. A discontinuity at ~ 220 km has been identified as the Lehmann discontinuity. A 30 km step in the Lehmann discontinuity close to the lapetus suture may be interpreted as juxtaposition of Laurentian and Avalonian mantle.

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Chapter 1

Introduction

1.1 Project Outline

The geological history of the British Isles is complex. The basement is thought to be composed of a number of unique geological terranes that have been brought together by a series of tectonic events culminating in the Caledonian orogeny (Figure 1.1). There have been many studies aimed at investigating these terranes, in particular the BIRPS seismic reflection profiles provide a great deal of information about the offshore structure (Klemperer and Hobbs, 1991). However, knowledge of the onshore structure is more limited. The LISP and CSSP wide angle seismic refraction profiles provide key deep crustal datasets (Barton, 1992; Al-Kindi, 2002).

This project is concerned with using teleseismic receiver function analysis to augment the onshore dataset of crustal velocity structure. There are 25 permanent short-period and 9 permanent and temporary broadband three-component seismic recording stations throughout the British Isles, the data from which are suitable for teleseismic receiver function analysis. Receiver function analysis is a powerful tool for examining crustal and upper mantle structure. Not only do the models obtained from the stations in this study provide a useful insight to the crustal evolution of the British Isles, but they can also be used to improve the velocity models needed to construct traveltime tables which are used in the monitoring of the Comprehensive Nuclear Test Ban Treaty (CTBT). The project has been funded by the U.K. National Data Centre, the organisation that fulfils the requirements of the U.K. as a member state of the CTBT.

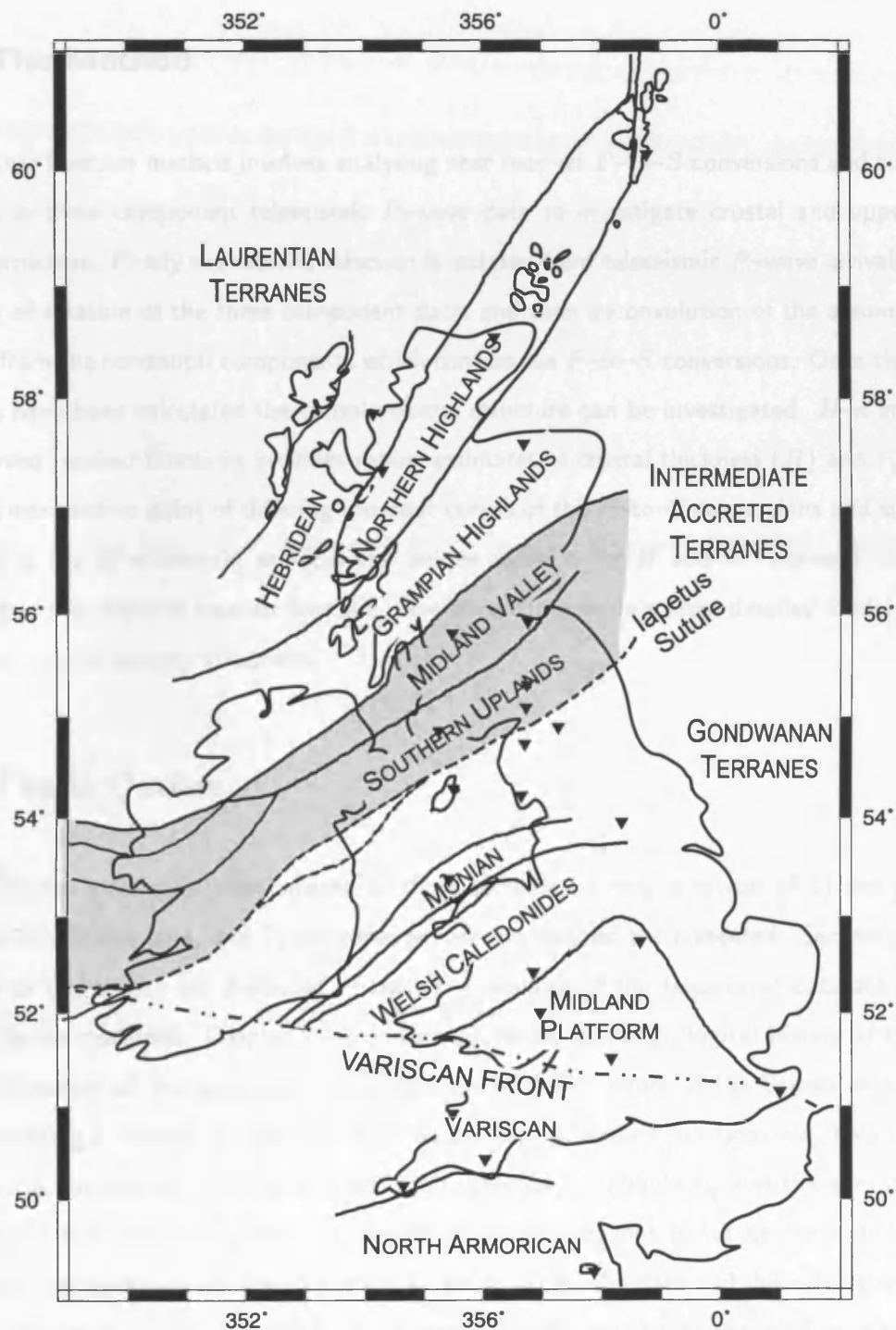


Figure 1.1: A map of the tectonic terranes of the British Isles after Woodcock and Strachan (2000). The seismic recording stations used in this study are marked (Inverted Triangle)

1.2 The Method

The receiver function method involves analysing near receiver P -to- S conversions and subsequent multiples in three component teleseismic P -wave data to investigate crustal and upper mantle velocity structure. Firstly the receiver function is isolated from teleseismic P -wave arrivals through a process of rotation of the three component data, and then deconvolution of the assumed source function from the horizontal components which contain the P -to- S conversions. Once the receiver functions have been calculated the causal velocity structure can be investigated. H - κ stacking of the observed receiver functions provides robust estimates of crustal thickness (H) and V_p/V_s ratio (κ). The intersection point of differing moveout curves of the P -to- S conversions and subsequent multiples in the H - κ domain provides the unique solution for H and κ . Forward and inverse modelling of the observed receiver function waveforms can provide a more detailed model of crustal and upper mantle velocity structures.

1.3 Thesis Outline

This thesis has essentially been written in three sections. Firstly a review of 1) the geological history of the British Isles, and 2) the receiver function method are presented. Secondly the data available to the project are described. Finally the analyses of the teleseismic data are presented and the results discussed. Chapter 2 introduces the review of the geological history of the British Isles. A summary of the geophysical investigations of the structure of the British Isles has been made, providing a number of questions that the analysis of receiver functions may help to answer. In Chapter 3 the receiver function method is fully described, highlighting how the questions raised in Chapter 2 may be investigated. In Chapter 4 the seismic monitoring networks and the data they supply are described along with the tools used to access the data and the catalogue of events used to calculate receiver functions. In Chapters 5-7 the analysis of the receiver functions are presented. In Chapter 5 a grid search modelling method is used to produce robust estimates of crustal thickness and V_p/V_s ratio. In Chapter 6 a more detailed analysis of the crustal and lithospheric mantle receiver function phases is conducted. In Chapter 7 receiver functions derived

from teleseismic events recorded at the broadband instruments are used to investigate the structure of the upper mantle. In Chapter 8 the results of chapters 5-7 are summarised with respect to the questions asked at the end of Chapter 2.

Chapter 2

Crustal Evolution and Physical Properties of the British Isles

2.1 Introduction

The crustal and upper mantle structure of the British Isles developed as the result of several major tectonic events; leaving a complex structure that is primarily dominated by features generated during the Caledonian orogeny that culminated ~ 425 million years ago. Subsequent tectonic events, including compression associated with the Hercynian orogeny, and extension associated with the opening of the North Sea and North Atlantic, have altered but not overprinted the Caledonian trends. Many geological and geophysical studies defining crustal and upper mantle structures have been carried out over the British Isles; of particular relevance to this study are the numerous deep seismic reflection profiles acquired by the British Institutes Reflection Profiling Syndicate (BIRPS) that investigated the offshore structure (Klemperer and Hobbs, 1991). This chapter briefly details the tectonic evolution of the British Isles, and describes studies of the physical properties of the crust and lithospheric mantle providing results that may be used as *a priori* information when modelling receiver function data. The results from previous geophysical studies are presented by geographical region detailing the scientific rationale behind the experiments, the results and findings, and discussion concerning the origin of the structures discovered.

2.2 Tectonic Evolution

The tectonic evolution of the British Isles prior to the Caledonian orogeny is well documented (e.g. Woodcock and Strachan, 2000). The pre-Caledonide British Isles began with Scotland together with Northern England, and Southern England together with Wales situated on different continents, separated by the Iapetus Ocean (Figure 2.1). The northern part was located at the southern margin of the ancient continent of Laurentia, with the southern part located on the northern margin of Gondwana. The two halves of the British Isles were brought together during the Caledonian orogeny, as the micro-continent of Avalonia detached from Gondwana closing the Iapetus Ocean. The Caledonian orogeny was a three way collision between Laurentia, Avalonia, and Baltica. The boundary between Laurentia and Avalonia, the Iapetus Suture, crosses northern England and Ireland (Figure 2.2). The final amalgamation of Laurentia and Avalonia was preceded by the Grampian orogeny on the southern margin of Laurentia. This deformation was the result of a volcanic arc collision with Laurentia. The Tornquist sea was closed as Baltica docked with Laurentia and Avalonia; evidence of this boundary (the Tornquist Suture) has been found beneath the North Sea on deep seismic reflection and refraction profiles (e.g. Abramovitz et al., 1999). As Iapetus closed, the Rheic Ocean was formed between Avalonia and Gondwana. The Caledonian Orogeny was followed by the closing of the Rheic Ocean and the subsequent Variscan Orogeny, as firstly Armorica (another micro-continent rifted from the margin of Gondwana) and then Gondwana collided with Laurussia to form the super-continent of Pangaea.

The structures that resulted from the Caledonian orogeny form the framework of the British Isles. The tectonic blocks that were brought together have been classified relative to their origin and formation. These terranes are attributed as having either; 1) Laurentian, 2) Gondwana or 3) accretionary origins (Bluck et al., 1992) (Figure 2.2a). The terrane classifications have been based upon the geological and geophysical observation of the structures, and their boundaries are normally identified by prominent fault zones. The amalgamation of the terranes was a complex multi-phase event of orthogonal and strike-slip closure that incorporated substantial amounts of accreted material between the continental structures. The terranes have been described by Bluck et al. (1992);

1. **Hebridean Terrane:** Undeformed by the Caledonian orogeny, the Hebridean terrane is part

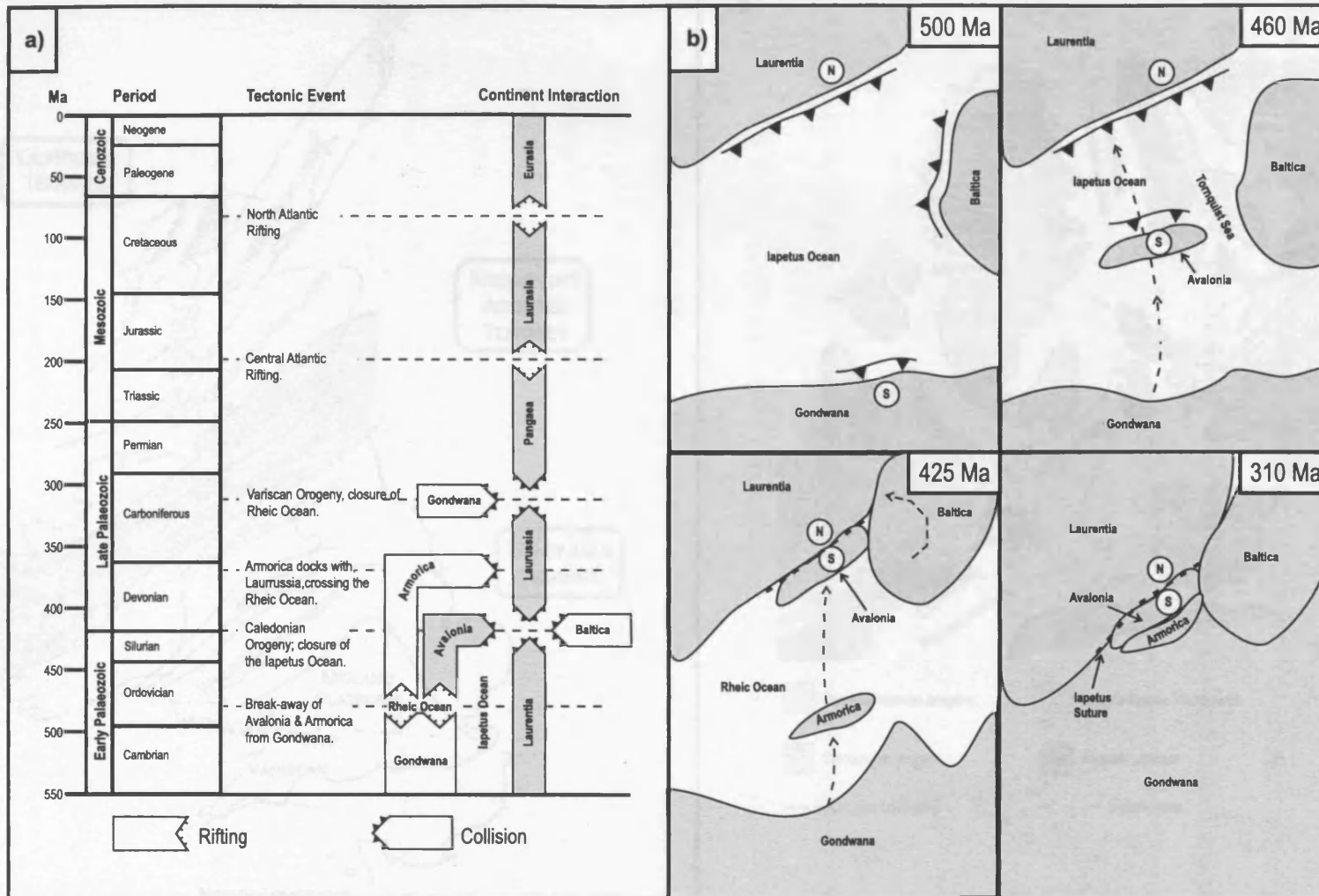


Figure 2.1: a) Schematic diagram of the evolution of the British Isles, showing the major tectonic events and the ancient continents that were involved. b) A series of cartoon palaeogeography maps (after Woodcock and Strachan (2000)) showing the closure of the Iapetus and Rheic Oceans during the Caledonian and Variscan orogenies respectively. The northern (N) and southern (S) sections of the British Isles are labelled.

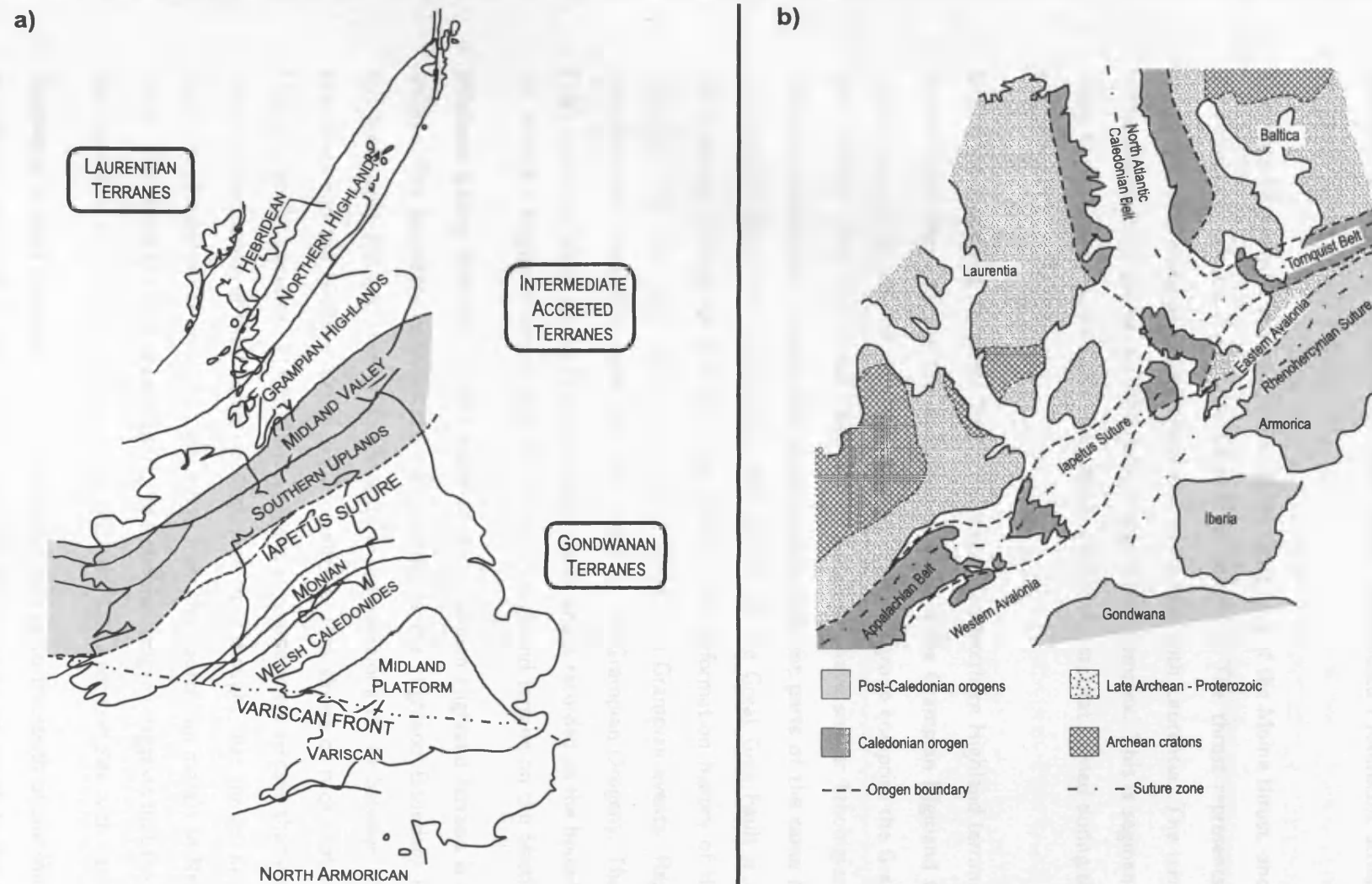


Figure 2.2: a) A tectonic terrane map of the British Isles after Woodcock and Strachan (2000) showing terrane boundaries, identifying those sourced from Laurentia and Avalonia, and those accreted during the Iapetus closure. b) A palaeogeographic map of the British Isles and surrounding areas (after Woodcock and Strachan (2000)) before the opening of the Atlantic Ocean.

of the original Laurentian craton composed of Lewisian gneissose basement. The terrane is bounded to the north by the Outer Isles Fault, a reactivated Proterozoic structure, and to the south by the Moine Thrust.

2. **Northern Highland Terrane:** This is the hanging wall of the Moine thrust, and over-thrust the Hebridean terrane during the Caledonian orogeny. The thrust represents 100 km of orthogonal shortening as Avalonia and Baltica docked with Laurentia. The terrane consists of the Moine super-group, a Proterozoic metamorphic complex. This is segmented into two major thrust nappes by the Sgurr Beag-Naver ductile thrust, that formed during the Grampian orogeny.
3. **Grampian Highland Terrane:** To the southwest, the Northern Highland terrane is bounded by the Great Glen Fault, to the southwest of which is the Grampian Highland terrane. The Dalradian super-group and the deeper Central Highland group comprise the Grampian Highland terrane. The Moine and Central Highland rocks have similar lithologies but have a different metamorphic history, and may represent different parts of the same metamorphic complex that have been juxtaposed. The motion on the Great Glen Fault is suggested to be strike-slip (Woodcock and Strachan, 2000). The deformation history of the terrane is complex, with two major nappes that formed during the Grampian events. Regional metamorphism and substantial uplift was also caused by the Grampian Orogeny. The Great Glen Fault extends offshore as the Walls Boundary Fault, and is recorded as the boundary between the Northern Highland terrane and the Grampian Highland terrane on the Shetland Isles.
4. **Midland Valley Terrane:** To the south of the Grampian Highland terrane is the Midland Valley. The boundary between the two terranes is the Highland Boundary Fault, which has both strike-slip and reverse components of motion along it. In between the Grampian Highlands and the Midland Valley are a number of thin slivers of rock that make up the Highland border complex. The Midland Valley is a magmatic arc terrane that was located on the southern margin of the Laurentian continent. It is thought that the arc developed on a fragment of continental crust that was rifted from the Laurentian margin in Neoproterozoic times. The model that has developed for the Grampian orogeny suggests that the deformation was caused by the oblique collision of the Midland Valley arc complex with Laurentia.
5. **Southern Upland Terrane:** The Southern Uplands lie to the south of the Midland Valley, bounded to the north by the Southern Uplands Fault and to the south by the Solway Line

(the Iapetus Suture). The Southern Uplands have been interpreted as both an accretionary basin, and a thrust-stacked sedimentary basin. It is now thought that the Midland Valley and Southern Uplands were separated by a fore-arc basin in front of the Midland Valley arc complex. The Southern Upland accretionary prism was subsequently thrust over these basinal deposits in the Caledonian deformation (Woodcock and Strachan, 2000).

6. **Iapetus Suture:** The inferred boundary between the Laurentian and Gondwanan terranes, has been associated with a number of north dipping reflectors on deep seismic reflection profiles (Klemperer and Hobbs, 1991). The suture also represents striking lithostratigraphical and faunal contrasts. To the south of the suture the Gondwanan terranes are less well defined than the Laurentian counterparts, due mainly to the lack of surface exposure of these rocks.
7. **Lakesman Terrane:** The Lakesman terrane, directly to the south of the Iapetus Suture, consists of a calc-alkaline volcanic and clastic sequence which formed in the marginal basins of Avalonia. These basins were deepened as Laurentia began to over-ride Avalonia, but were eventually compressed and uplifted as the deformation intensified. The sequence extends westward into Ireland, and is also exposed on the Isle of Man.
8. **Monian Terrane:** Within the Avalonian rocks of England and Wales there is the suspect Monian terrane, which is separated from the Avalonian rocks by the Menai Strait Fault system. The Monian Precambrian basement is unconformably overlain by Palaeozoic rocks. The basement contains high grade gneisses, calc-alkaline granite plutons and a belt of blue-schist facies metamorphic rocks which are indicative of high-pressure / low-temperature metamorphism associated with subduction. The terrane is part of the accretionary prism of an Avalonian subduction zone of unknown polarity.
9. **Avalon Terrane:** This terrane is an amalgamation of the structures in Southern Britain, including the Welsh Basin, the Midland Platform and the concealed Caledonides of Eastern England. The Welsh basin is a >10 km thick sequence of marine clastic sediments that have been deposited in three unconformable super-groups. The basin is underlain by late Precambrian Avalonian basement. The Welsh basin is bounded to the north by the Menai Strait Fault and the Monian terrane. To the south and east it is bounded by the Welsh Border Fault system that leads to the Midland Platform. The relatively undeformed crust of the Midland Platform is comprised of late Precambrian volcanic, volcanoclastic and sedimentary sequences punctuated by plutonic rocks. These are overlain by incomplete Gondwanan shelf

sequences analogous to the those in the Welsh Basin. The Caledonian basement rocks of eastern England are concealed by a blanket of later sedimentary rocks, but are thought to be a southward continuation of the Lakesman terrane.

10. **Variscides:** To the south of the limit of the Variscan deformation, the Variscan Front, is a deformed extension of the Avalon terrane composed of Precambrian metamorphic and volcanic rocks. On the Lizard peninsular in the very southwest of England, the ophiolite complex of the Lizard terrane has been overthrust onto the Avalon terrain during the closure of the Rheic Ocean. Northeastern France and the Channel Islands are located on the North Armorican terrane, part of the Armorican micro-continent that collided with Laurussia during the closure of the Rheic Ocean.

Since the Caledonian Orogeny the British Isles have been characterised by the erosion of uplifted blocks and the deposition of the resultant sediments into basins. This pattern has been driven by the reorganisation of the tectonic blocks due to compressional and extensional events. The tectonic events combined with sea level changes produced a wide range of sedimentary depositional environments. The Caledonian basement has been covered with a blanket of these late Palaeozoic, Mesozoic and Cenozoic sediments. This cover is particularly continuous in the south and east of England.

There have been several phases of extension that have acted in the British Isles since the end of the orogenic episodes described above, and they have not only provided sediment sinks but also substantially thinned the crust. The Jurassic rifting of the Southern Atlantic which initiated North Sea graben formation, and Cretaceous rifting in the North Atlantic and North Sea are the primary phases of the extension. The formation of Iceland on the Atlantic mid-ocean ridge has been associated with uplift and denudation in the British Isles. It has been suggested that uplift has been caused by low density magma sourced directly from the upwelling Iceland Plume underplating parts of the crust of the British Isles (Jones et al., 2002).

2.3 Geophysical Studies

2.3.1 Overview

The complex nature of the structure of the British Isles has been investigated by a large number of geophysical studies. The BIRPS syndicate have carried out many of these, using deep seismic reflection, and wide-angle methods along a number of profiles in the continental shelf seas surrounding the British Isles (Klemperer and Hobbs, 1991). Wide-angle seismic studies providing both crustal and upper mantle velocities have also been carried out on-shore, notably the Lithospheric Seismic Profile in Britain (LISPB) (Bamford et al., 1978), and the Caledonian Suture Seismic Experiment (CSSP) (Bott et al., 1985). The locations of these profiles are shown in Figure 2.3. The seismic velocities and crustal structures recorded by these investigations act as an important constraint to the present receiver function study. The results from these studies are presented as a compilation of bulk crustal physical properties, primarily from the wide-angle seismic reflection/refraction studies. The results of the geophysical investigations are subsequently discussed in relation to the geological structures they investigate. This discussion is divided into four geographical regions concentrating on four individual geological domains. These sections are; The Northwest Highlands (using stations located on Laurentian terranes), The Iapetus Suture (Accreted terranes), Central England and Wales (Avalonian terranes), and Southwest England (to the south of the Variscan front). This group of four regions is maintained throughout the thesis, particularly in Chapter 6, which deals with the modelling of receiver function data from individual stations distributed throughout the British Isles.

2.3.2 Crustal Properties

Deep seismic reflection profiling has been the most commonly used method of investigating crustal and upper mantle structure around the British Isles. The method in particular highlights near horizontal seismic velocity contrasts. The BIRPS deep seismic reflection studies have provided a significant insight into the crustal and upper mantle structure offshore the British Isles as a result of collecting a large number of such profiles. The volume of data collected has allowed the *typical BIRP* structure to be summarised (Table 2.1), which describes a two layer crust above the mantle lithosphere. The classification of this structure is based upon the reflection patterns observed within the BIRPS data. The upper crust, beneath sedimentary basins, is characteristically unreflective and

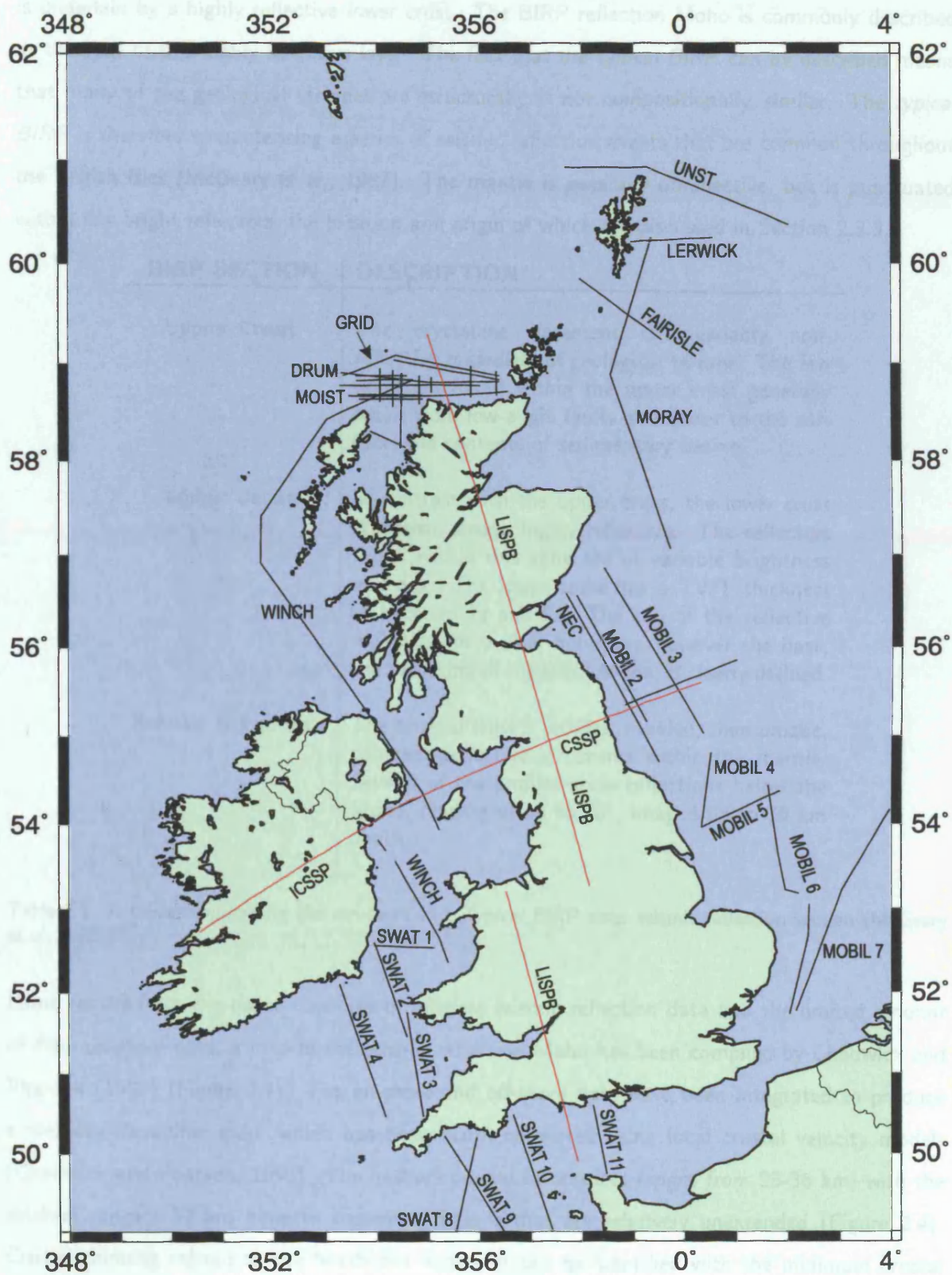


Figure 2.3: A map of the deep seismic reflection and refraction profiles around the British Isles .

is underlain by a highly reflective lower crust. The BIRP reflection Moho is commonly described as the base of this highly reflective layer. The fact that the *typical BIRP* can be described means that many of the geological terranes are structurally, if not compositionally, similar. The *typical BIRP* is therefore characterising a series of seismic reflection events that are common throughout the British Isles (McGeary et al., 1987). The mantle is generally unreflective, but is punctuated with a few bright reflectors, the location and origin of which are discussed in Section 2.3.3.

BIRP SECTION	DESCRIPTION
Upper Crust	The crystalline basement is invariably non-reflective regardless of geological terrane. The few reflectors found within the upper crust generally result from low-angle faults and closer to the surface, the contents of sedimentary basins.
Lower Crust	In contrast with the upper crust, the lower crust is almost always highly reflective. The reflectors found within this zone are of variable brightness and dip. The lower crust has a TWT thickness of between 1s and 5s. The top of the reflective region is not a clear boundary, however the base, the definition of the BIRP Moho, is clearly defined.
Mantle Structure	The original BIRPS profiles revealed, then unique, coherent reflective structures within the mantle. Several of the profiles show reflections below the Moho, dipping at up to 30°, imaged up to 70 km depth.

Table 2.1: A table summarising the structure of a *Typical BIRP* deep seismic reflection section (McGeary et al., 1987)

Using results from the dense coverage of offshore seismic reflection data and the limited amount of deep on-shore data, a map of the seismic reflection Moho has been compiled by Chadwick and Pharaoh (1998) (Figure 2.4). The on-shore and off-shore data have been integrated to produce a two-way-traveltime map, which has been depth converted using local crustal velocity models (Chadwick and Pharaoh, 1998). The onshore crustal thicknesses ranges from 25-36 km, with the thickest range >32 km beneath basement massifs that are relatively unextended (Figure 2.4). Crustal thinning related to the North Sea extension can be identified with the minimum crustal thickness reaching 22 km. Both the northwest of Scotland and the southwest of England show significant thinning of the crust, with thicknesses <28 km. Central and southeast England and



Figure 2.4: A drawing of the seismic reflection Moho map produced by Chadwick and Pharaoh (1998). FR & DR = Intersection respectively of the Flannan Reflector and the Dowsing Reflector with the Moho. Contours show the depth to the Moho in km.

parts of Wales have thicker crust with values rising to ~ 34 km. A thickened root is also present offshore northeast England, trending northeast-southwest, close to the surface exposure of the Iapetus Suture, although this does not appear to extend onshore. Anomalously thick crust is also seen beneath the Midland Valley of Scotland, reaching a maximum of 36 km, the thickest crust in the British Isles. It must be noted that the onshore thicknesses are constrained by very few high quality picks of the Moho and are poorly constrained away from these points. It is likely that the crustal thickness of the British Isles after the Caledonian orogeny reached between 50-60 km, and the process that caused the thinning to the present values has significantly modified the entire crustal column (McGeary et al., 1987). In general, regions that have similar crustal history to the British Isles extended continental crust, are of the order of 30 km thick (Christensen and Mooney, 1995).

Crustal seismic velocity structure is primarily constrained by deep seismic refraction profiles. This velocity information is particularly important as a constraint to the present study because the receiver function method is insensitive to absolute velocity, rather being sensitive to relative delay time and acoustic impedance contrast (Ammon et al., 1990). The LISPb and CSSP profiles provide the main velocity constraint within the British Isles. The LISPb profile transects all of the major tectonic structures providing both P-wave and S-wave velocities for the crust and upper mantle. The LISPb data was initially interpreted by Bamford et al. (1976, 1978) who analysed the crustal P-waves. Assumpção and Bamford (1978) worked on the crustal S-wave data from LISPb, and subsequently Barton (1992) re-interpreted the P-wave data (Figure 2.5). The CSSP profile was shot along strike of the Iapetus suture zone, initially analysed by Bott et al. (1985) and subsequently by Al-Kindi (2002) (Figure 2.6b), and is continued westward into Ireland with the ICSSP (Jacob et al., 1985). Further offshore control on crustal velocity is provided by the JUNE92 survey along the GRID17 BIRPS profile off the north coast of Scotland (Jones et al., 1996; Price and Morgan, 2000; Morgan et al., 2000) (Figure 2.6a).

The resultant velocity models from the wide-angle seismic refraction investigations typically reveal a multi-layer crust (Figure 2.5). The LISPb experiment was shot in two sections; the northern section extending from offshore northwest Scotland south to northern England, and the southern section extending from the north Wales coast to the English Channel (Figure 2.3). The original interpretation of the northern LISPb profile (Bamford et al., 1978) demonstrated a three layer

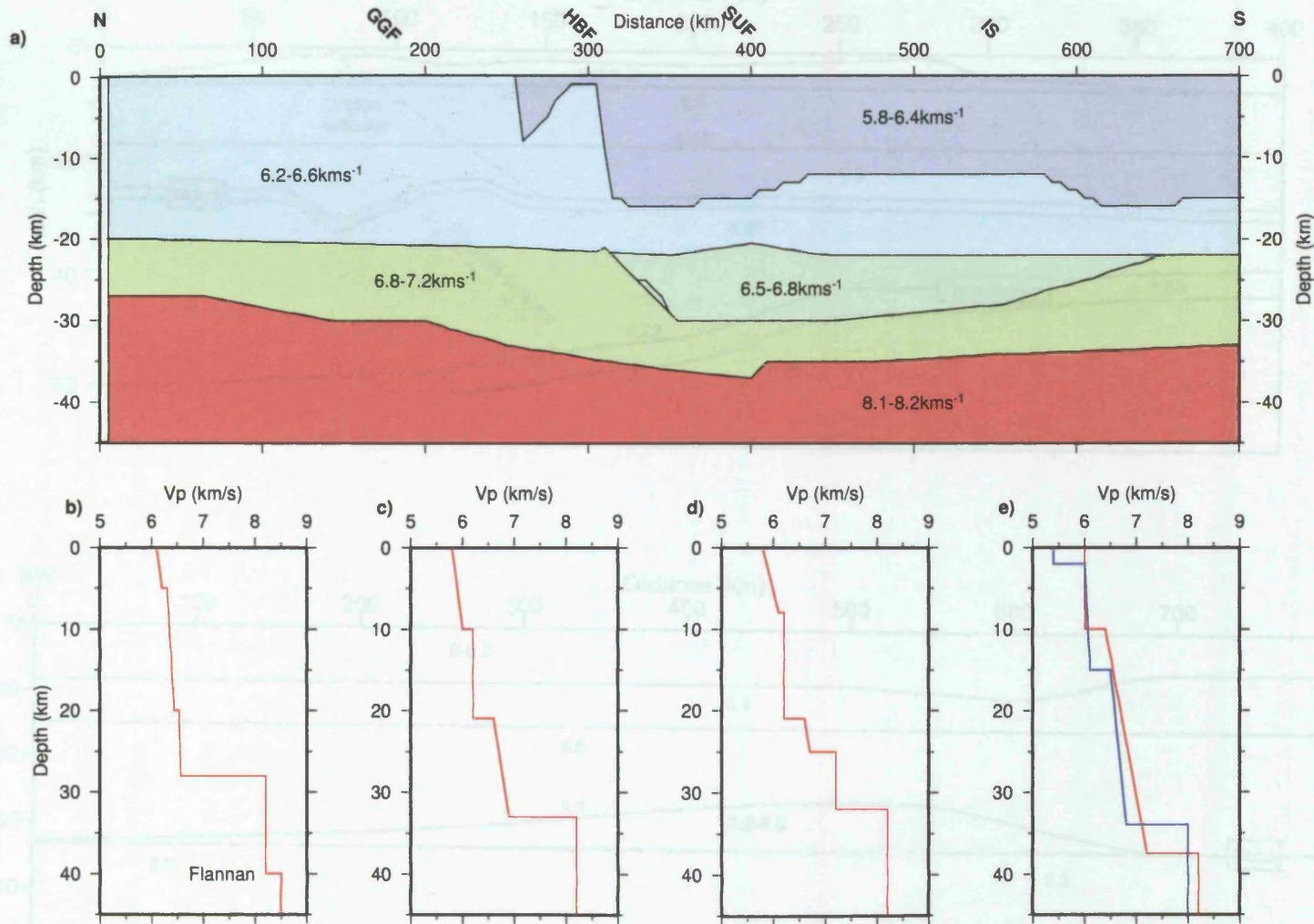


Figure 2.5: Velocity models from seismic refraction studies of the British Isles. a) LISPb northern profile after Barton (1992). b) A velocity-depth model extracted from the eastern section of the model from the JUNE92 experiment along the GRID17 profile (Morgan et al., 2000). c) & d), Models from the LISPb and CSSP profiles respectively. The LISPb model is taken from close to the lapetus Suture zone (Barton, 1992). The CSSP model has a high velocity layer at the base of the crust, which is interpreted as magmatic underplating (Al-Kindy 2002). e) A model from the southern LISPb DELTA profile in Wales, (Bamford et al., 1976).

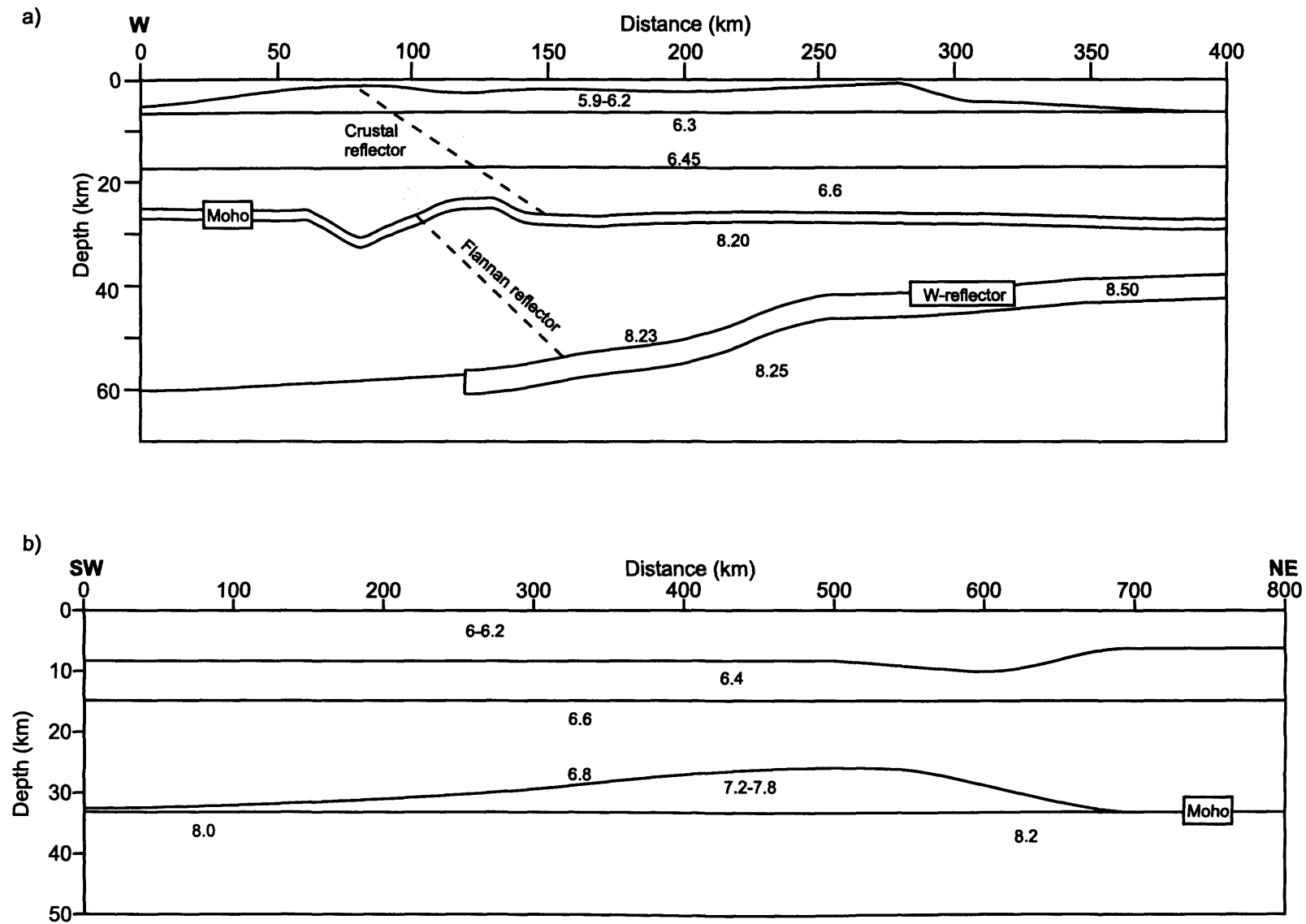


Figure 2.6: Velocity models from the JUNE92 and CSSP seismic refraction profiles. a) The JUNE92 profile model of Morgan et al. (2000), b) the CSSP velocity model of Al-Kindi (2002). The velocities are marked on the sections in km s^{-1} .

crust of variable thickness. This model has been classified into five crustal and mantle lithologies, based upon seismic velocity and geological location (Table 2.2). The upper crust has been divided into two distinct sections, classified by their P -wave and S -wave velocities, to the north and south of the Highland Boundary Fault. To the north is the Caledonian Metamorphic belt, with unmetamorphosed basement to the south. These represent the Laurentian and accreted terranes that were juxtaposed during the Grampian and Caledonian orogenies. The metamorphosed basement has a higher V_p at 6.1-6.2 km s⁻¹ in comparison with 5.8-6.0 km s⁻¹ for the unaltered basement to the south. The mid-crust is thought to represent the pre-Caledonide basement. To the north of the Southern Uplands Fault the mid-crust has $V_p > 6.4$ km s⁻¹ and is probably granulite facies Lewisian basement, while to the south V_p is typically < 6.3 km s⁻¹. The lower-crust ($V_p > 6.9$ km s⁻¹) and mantle (V_p 8.0-8.2 km s⁻¹) are reasonably consistent over the section, with the mantle showing little anisotropy (Bamford et al., 1979). The final LISPB model contains uncertainty in the velocity structure especially where considerable lateral variations are present. The results from the southern LISPB profile (Delta) were only published as a preliminary model (Bamford et al., 1976). The unpublished model of Bamford and Nunn is presented by Edwards and Blundell (1984). This reveals a thicker crust of ~34-35 km. The crust appears to be two layered with the maximum velocity of the lowermost crust reaching only 6.9 km s⁻¹ (Figure 2.5e). The upper layer of the model is ~10 km thick and is suggested to represent the Lower Palaeozoic sediments that were deposited in the Welsh Basin during the closure of Iapetus.

Barton's (1992) remodelling study finds essentially the same lateral discontinuities in the northern profile as Bamford et al.'s (1976) study (Table 2.3). Major geological structures are identified as the boundaries between distinct types of crust. Finer details are resolved in the crustal model, clarifying a number of the uncertain structures in the original model (Figure 2.5a). The crust at the northern end of the profile, 0-280 km, is generally thin (26-30 km), thickening toward the south. A change in the velocity gradient is noted at ~180 km apparently coinciding with the Great Glen Fault, the boundary between the Northern Highland and Grampian Highland terranes. At 280 km an upper crustal low-velocity-zone associated with the Tay Nappe is modelled, which is terminated at 300 km by the Highland Boundary Fault and the start of the Midland Valley. As with Bamford et al.'s (1976) model, the Midland Valley has a significantly different structure to the rest of the profile, with a maximum crustal thickness of 36 km. At the northern boundary there is a change in upper crustal structure, with a high V_p gradient, increasing from 5.8-6.5

Crustal Layer	(V_p (km s ⁻¹))	(Depth (km))	Description
Superficial Layer			Upper Paleozoic and more recent deposits.
Upper Crust	5.8-6.0	0-18	Lower Paleozoic unmetamorphosed upper crust, south of the Highland Boundary Fault.
	6.1-6.2	0-20	Caledonian metamorphic belt north of the Midland Valley.
Mid Crust	>6.4	8-20	Granulite facies Lewisian basement found north of the Southern Uplands Fault.
	<6.3	15-20	Pre-Caledonian basement, underlying Lower Paleozoic upper crust, bounded to the north by the Southern Uplands Fault.
Lower Crust	7.0	20-30	High velocity lower crust, extending over the entire profile.
Mantle	8-8.2	26-36	Mantle. No velocity anisotropy appears to be present, and lateral variations in velocity are insignificant

Table 2.2: A table summarising the crustal layers identified by Bamford et al. (1978)

Unit	Distance	Description
Highlands	0-280 km	The area to the north of the Highland Boundary Fault, which is between 26-30 km thick. North of the Great Glen Fault there is a low velocity gradient in the upper crust from 6.2-6.4 km s ⁻¹ in the upper 20 km, with a much steeper gradient in the lower crust, with V_p reaching 7.3 km s ⁻¹ above the Moho. To the south the velocity gradient is higher in the upper crust and lower in the lower crust, with a similar maximum velocity being found.
Tay Nappe	280-300 km	A low velocity zone in the upper 2-4 km.
Midland Valley	300-400 km	A much different structure to the rest of the profile. There is a high velocity gradient from 5.8-6.5 km s ⁻¹ in the upper 14 km. The mid-crust comprises two layers 5-10 km thick, with velocities of 6.6 km s ⁻¹ and 6.8 km s ⁻¹ respectively. The lower crust is a sliver of high velocity material (V_p 7.3 km s ⁻¹) above the Moho at 36 km. The Poisson's ratio of the mid-crust is anomalously low at 0.22 (Assumpção and Bamford, 1978).
Southern Uplands	400-500 km	To the south of the Southern Uplands Fault, coinciding with the truncation of the Midland valley structures, the Moho is offset by a 2 km decrease in crustal thickness and a decrease in maximum velocity of the lower crust to 6.9 km s ⁻¹ . The upper crustal structures are unaltered, although they show a slight decrease in velocity and velocity gradient.
Iapetus Suture	500-700 km	The south of the profile is least well sampled by the data. There are no significant changes in the model across the Iapetus Suture Zone.

Table 2.3: A table summarising the distinct crustal structures identified by Barton (1992)

in the first 14 km. This is consistent with the terrane interpretation, the Midland Valley being an allochthonous volcanic arc that was accreted on the margin of Laurentia. At the Southern Uplands Fault, the southern boundary of the Midland Valley (~ 400 km), there is a change in lower crustal structure. To the south of this boundary there is a step decrease in the crustal thickness from 36 to 34 km, which also corresponds with a decrease in the P-wave velocity of the lowermost crust from 7.3 to 6.9 km s $^{-1}$. These variations identified in Barton's 1992 model also correspond to the more general upper and lower crustal classifications assigned by Bamford et al. (1978). The southern end of the profile (500-700 km) is the least well sampled, with only two shots reversing the ray coverage over this section. The layered crustal structure is continuous from the Southern Uplands Fault, and there is no recognisable change in seismic velocity structure over the Iapetus Suture Zone. The model includes structure within the mantle at the northern end of the profile with a change in V_p from 8.2 to 8.55 km s $^{-1}$ at a depth of ~ 57 km.

The study of shear-waves provides the opportunity to investigate the compositional makeup of the crust. Variation in Poisson's ratio can be related to variations in the mineralogical content of the crustal column. In general, Poisson's ratio has been linked to the silica content of the rock; between silica contents of 55-70% there is a linear increase in Poisson's ratio with increase in silica, but beyond 70% there is much wider distribution of measurement (Christensen, 1995). The shear-wave velocity is also sensitive to the temperature of the rock, and thus may provide an indication of the thermal state of the crust. Shear-wave velocity information has been presented for the LISPB and JUNE92 experiments by Assumpção and Bamford (1978) and Price and Morgan (2000) respectively. The results are published as Poisson's ratio values rather than directly as V_s values. The northern end of the LISPB profile and the centre of the JUNE92 profile intersect. The crustal values for Poisson's ratio are consistent with one another at 0.246-0.248 and 0.244 respectively. Further south, the LISPB profile reveals a mid-crust with a low Poisson's ratio of 0.224. Other studies have investigated the crustal shear-wave structure. Ward et al. (1992) show that the lower crust beneath Weardale in northern England has a V_p/V_s ratio of 1.84 ($\sigma=0.29$). It is suggested that this anomalously high Poisson's ratio is caused by fluids in the lower crust. Meredith and Pearce (1991) have investigated the dispersion of Rayleigh waves over the U.K. broadband network via the analysis of phase velocity curves. Poisson's ratio for the lower crust of central Britain was found to lie between 0.25-0.27.

Seismic velocity anisotropy on a crustal scale can cause considerable alteration to the form of

observed receiver functions (Levin and Park, 1997). Anisotropy of the seismic velocity of rocks can be caused by a number of plausible geological conditions. Anisotropy may be caused by the layering of different minerals or lithologies, but could also be caused by preferred fabric orientation of minerals that have been aligned by some form of deformation (Levin and Park, 1998). There have been several studies that have concentrated upon the crustal anisotropy of the British Isles. Jones et al. (1996) showed that the crust sampled by the JUNE92 survey includes bulk anisotropy of up to 7%. They noted that Moho depths for deep wide-angle seismic refraction and reflection studies do not correlate when the refraction data is converted into near-offset travel-times, with the reflection Moho being identified at later times than the refraction Moho. There are several possible causes for this mis-match, but Jones et al. (1996) conclude that the 7% anisotropy is the most plausible. With regard to more general studies of the anisotropic behaviour of the crust and upper mantle, Ando et al. (1987) investigated the data from the UKNET broadband seismic network, and concluded that there was no significant anisotropy beneath the British Isles. These findings are consistent with those of Bamford et al. (1979). The findings of Ando et al. (1987) and Bamford et al. (1979), and Jones et al. (1996) do not necessarily contradict each other. Jones et al. (1996) compare methods that are sensitive to horizontal and vertical seismic velocity, and find anisotropy that is associated with horizontal layering which typically has a vertical axis of symmetry. Ando et al. (1987) and Bamford et al. (1979) investigated teleseismic events in order to evaluate azimuthal variation in seismic velocity, and the results of these studies are not sensitive to anisotropy with a vertical axis of symmetry.

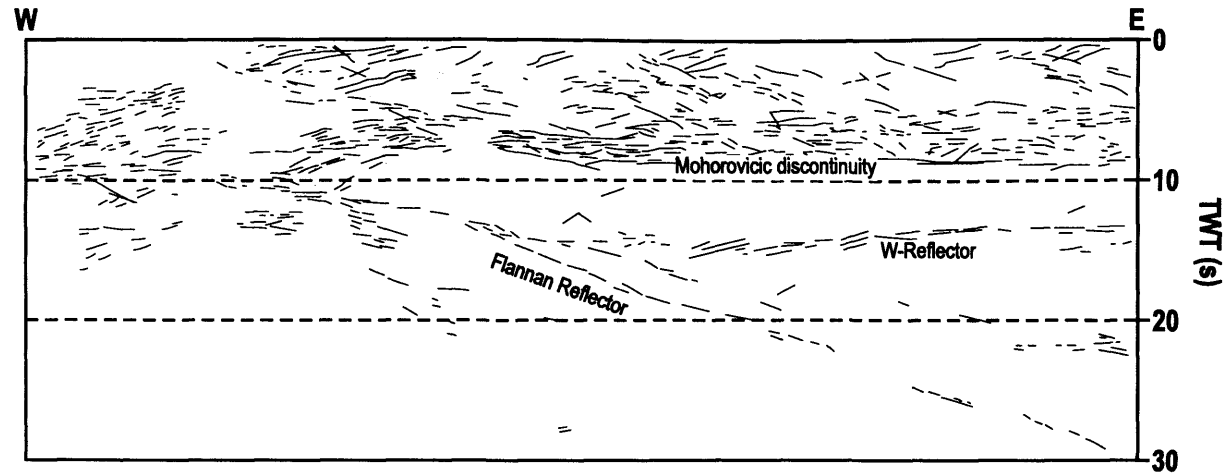
2.3.3 Northwest Highlands

The seas to the north and west of Scotland have been densely covered by BIRPS profiles. The primary surveys include the Moine and Outer Isles Seismic Traverse (MOIST) (Smythe et al., 1982), Deep Reflections from the Upper Mantle (DRUM) (McGeary and Warner, 1985) and SHETland (SHET) (McGeary, 1989) surveys. MOIST was the first of the BIRPS experiments, and was conducted with the aim of imaging the continuation of the Moine Thrust and Outer Isles Fault, and the structures of the Hebridean and Northern Highland terranes. However, the profile also discovered the Flannan reflector, the first bright lithospheric mantle reflector to be imaged by the reflection technique (Klemperer and Hobbs, 1991). The following DRUM and GRID experiments were designed to better map the mantle reflectors found on MOIST. The GRID

survey as the name suggests was formed of a grid of 2D profiles that mapped the continuity of the Flannan reflector northwest of Scotland. The DRUM profile has the longest record length of all the profiles, recording to 30s, and shows the easterly dipping Flannan reflector penetrating from the Moho at ~ 30 km to at least 80 km depth (Figure 2.7a). The flat lying W-reflector is identified at ~ 50 km (McGeary and Warner, 1985). These sub-Moho structures have subsequently been the object of significant further study. The mantle reflectors are particularly high amplitude, with reflection coefficients of 0.08-0.14 estimated by comparing their amplitudes to those of the sea bottom reflection (Snyder and Flack, 1990). There has been much discussion on the origin of the mantle reflectors, with several proposed hypotheses including thrust faulting, extensional shear zones, a relict Moho, relict subduction zones and the presence of fluids (Price and Morgan, 2000). Subsequently a wide-angle profile was shot coincident with the GRID17 profile (Jones et al., 1996). The resultant model shows the mantle with a velocity of 8.20 km s^{-1} above the W-reflector, beneath which the velocity increases to 8.50 km s^{-1} (Figure 2.6a). The velocity contrast between $8.2\text{-}8.5 \text{ km s}^{-1}$ is not enough to produce the observed reflection coefficient of 0.08-0.14. Price and Morgan (2000) analysed the reflection amplitudes and concluded that there must be a low-velocity-zone above the mantle reflector in order to match the amplitude of the reflection. They suggest that this velocity structure could be caused by metasomatised normal mantle above a subducted slab composed of mafic eclogite.

The SHET profiles (UNST, LERWICK and FAIRISLE) cross the Walls Boundary Fault, an extension of the Great Glen Fault, and show a step in the Moho where they cross the fault, with the crust thickening by 2-3 km to the east. The structure was probably formed during the Caledonian orogeny as the Northern Highland and Grampian Highland terranes slipped past one another. The preservation of this topography suggests that the area was not significantly modified during the extension associated with the North Sea rifting and the opening of the North Atlantic Ocean (McGeary, 1989). Further surveys close to the SHET experiment have also revealed upper mantle reflectors similar to those found by MOIST, GRID and DRUM. The NDSP85-8 line, the eastern end of which approaches the Walls Boundary Fault, shows an east dipping upper mantle reflector at between 30-50 km depth, which is an extension of the Flannan reflector found on the MOIST profile (McBride et al., 1995).

a)



b)

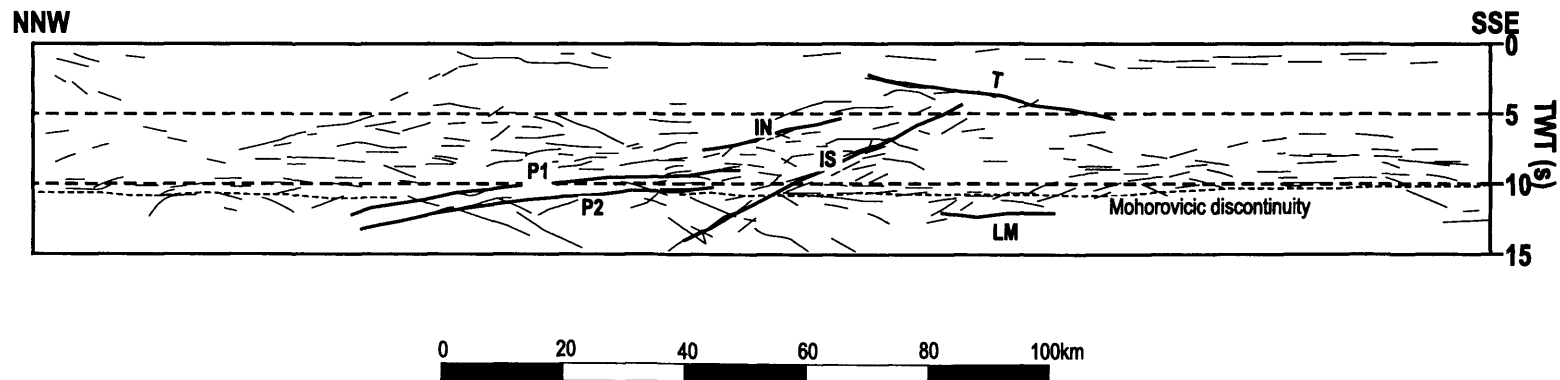


Figure 2.7: a) A line drawing of the DRUM deep seismic reflection profile. This shows the seismic reflection Moho at the base of a highly reflective lower crust. The Flannan Reflector is imaged as an easterly dipping structure, intersecting the W-Reflector at 50 km depth. b) A line drawing of the NEC profile again showing the seismic reflection Moho at the base of a reflective lower crust. The interpretation from Freeman et al. (1988) shows: IN = Structure that maps to the surface at the Iapetus Suture marking a change in reflectivity, IS = Lower boundary of the Iapetus Suture zone complex, T = A reflection possibly from a fault, P1 & P2 = Reflections from within and at the base of the lowermost crust respectively, LM = Layered mantle.

2.3.4 Iapetus Suture

The structures associated with the closure of the Iapetus Ocean have been investigated by BIRPS profiles that run along both the east and west coast of the British Isles (Figure 2.3). On the western side of the country is the Western Isles North Channel profile (WINCH) (Brewer et al., 1983; Hall et al., 1984). There have been several experiments along the eastern margins of the British Isles and in the North Sea including the NorthEast Coast (NEC) (Freeman et al., 1988), and the Measurements Over Basin to Image the Lithosphere (MOBIL) profiles (Klemperer and Hobbs, 1991). These profiles have shown a group of north dipping crustal reflectors which strike parallel to the dominant Caledonian trend and have been identified as the subsurface expression of the Iapetus suture (Freeman et al., 1988; Brewer et al., 1983). An example of a typical Caledonian BIRP, in this case from NEC is given in Figure 2.7b. Again the Moho can be identified as the sharp base of lower crustal reflectivity. On the eastern profiles the crust has been classified into zones of differing reflectivity that represent the different geological terranes (Klemperer and Hobbs, 1991). These zones can be summarised as a) The Midland valley, b) The Iapetus subduction zone complex, c) The Lake district and d) The Midland Platform. There are two north dipping reflectors within the Iapetus subduction zone complex, IS and IN (Figure 2.7b), which have both been identified as the Iapetus Suture by Freeman et al. (1988) and Soper et al. (1992) respectively. The latter argue that, when extrapolated to the surface, IN is almost co-located with the Iapetus Suture as inferred from geological mapping. The crustal thickness on the north-south transects varies between 30-35 km, thickening around the root of the Iapetus subduction complex structure (Figure 2.4). The LISPB profile crosses the Iapetus Suture but reveals no noticeable change in seismic velocity structure (Barton, 1992) (Figure 2.5a). However, the CSSP profile reveals a well shaped high velocity layer at the base of the crust which is attributed to the presence of magmatic underplating (Al-Kindi, 2002) (Figure 2.6b).

2.3.5 Central England & Wales

The deep structure of central and southern England has been the least studied of all of the four regions covered in this chapter. The crustal thickness in England is constrained by short onshore seismic reflection profiles, with some inferences resulting from offshore data (Chadwick and Pharaoh, 1998). The crust in this region is thicker than that beneath the majority of the

British Isles, being between 33-35 km. The seismic Moho map of Chadwick and Pharaoh (1998) shows the extent of the thicker region that is inferred to be the Midland Platform. This is regarded as the relatively unaltered section of the Eastern Avalonia micro-continent (Chadwick and Pharaoh, 1998). The basement structure close to the CWF station used in the present study has been investigated by Whitcombe and Maguire (1980), revealing the high velocity basement ($V_p \approx 6.4 \text{ km s}^{-1}$) is overlain by 2 km of Precambrian sediments. A further P -wave residual study of the area shows the basement velocity is reduced in places by granitic/dioritic intrusions (Maguire et al., 1985). The onshore deep crustal structure of Wales is constrained by the unpublished LISPB model of Bamford and Nunn (Edwards and Blundell, 1984) along with the earlier preliminary model of Bamford et al. (1976). The crust in this area is again thicker than that of the average crust, at $\sim 33 \text{ km}$.

The basement in southern and eastern England is generally concealed by variable thicknesses of Paleozoic, Mesozoic and Cenozoic strata. This sedimentary cover can impair the interpretation of receiver function data because the high acoustic impedance contrast between the sedimentary rocks and the basement rocks generates high amplitude P -to- S wave conversions and multiples that mask the later crustal conversions. There have been several studies of the cover thickness and basement structure of southern England. Busby and Smith (2001) have studied the pre-Variscan basement structure by analysing potential field data. The depth to top Variscan basement has been compiled in England and Wales by Whittaker (1985) and shows sediment thicknesses of greater than 2 km in places.

2.3.6 Southwest England

The crustal structure of the southwest of England and the Celtic basins of Wales has been mapped by the SouthWest Approaches Traverse (SWAT), (Figure 2.3). In these areas the Caledonian structure has been overprinted by deformation associated with the Variscan Orogeny (Figure 2.2). The crust has been thinned in places to $< 28 \text{ km}$ by later extension. The southwest of England is not affected by post Variscan sedimentation. Offshore the SWAT 2 & 3 profiles running N-S from North Wales to the Cornubian Plateau through the Celtic Sea, show an average unextended crust of 30-32 km. The North and South Celtic Sea basins are thinned to a minimum of 25-27

km. Within SWAT profiles 2-3-4, two detachments can be identified extending from these basins to the top of the lower crust at 20 km. This surface has been identified as the Variscan front by a number of authors (e.g. Klemperer and Hobbs, 1991) and may have been partly reactivated during the Permian-Cretaceous extension, forming the Celtic basins. To the southwest of England the SWAT 7-8-9 profiles cross the internal part of the Variscan Orogeny, the northern limit of which is delineated by the thrusting associated with the Lizard ophiolite complex. This thrust can be seen dipping southward beneath the Plymouth basin on the BIRPS profiles at angles of up to 50°.

2.4 Summary

The tectonic evolution of the British Isles has created complex structures within a crust that contains the remnants of at least two continental collisions, leaving juxtaposed crustal segments of contrasting composition and origin. This crust has subsequently been extended during the rifting of the North Sea and North Atlantic Ocean. Many research workers have investigated these crustal structures, collecting a dense set of deep seismic reflection and refraction data profiles. Although these datasets provide a dense coverage of the offshore crust of the British Isles, many questions about its tectonic evolution still remain. The present teleseismic receiver function study was instigated to resolve some of these questions.

1. The onshore crustal thickness is not particularly well constrained, with very few high quality measurements of the Moho. There is a wide distribution of the three-component seismic stations over the British Isles, covering all of the inferred geological terranes. The analysis of receiver functions from these stations using the $H-\kappa$ stacking technique provides robust estimates of the crustal thickness (Chapter 5). This allows the onshore crustal thickness variation to be compared to the terrane distribution, providing particularly important information in the poorly constrained south of England. Furthermore, $H-\kappa$ stacking also produces estimates of V_p/V_s ratio, which is controlled by the mineralogical composition and thermal state of the crust.
2. The most notable anomaly between the deep seismic reflection and refraction methods is across the Iapetus Suture zone. The seismic refraction results do not find any significant variations in seismic velocity over the suture zone, whereas the seismic reflection results provide strong dipping reflectors identifying the suture. The densest distribution of permanent

seismic stations spans the Iapetus Suture zone, and the detailed modelling of the phases of receiver functions from these stations provides an alternative perspective on these structures (Chapter 6). The receiver function method is capable of resolving structures in the upper mantle, allowing the sub-crustal structure of the suture zone to be examined.

3. The onshore continuity of the sub-crustal Flannan and W-reflector structure in northern Britain is uncertain. The LISPb profile has modelled a mantle discontinuity at a similar depth to the W-reflector, and they have been tentatively linked. The depth penetration of the receiver function method allows the presence of any sub-crustal discontinuities to be investigated beneath available stations (Chapter 6). The amount of high quality velocity control in the area constrains the receiver function modelling, enabling investigation of the depth and velocity contrast of the discontinuities through the modelling of the delay times and amplitudes of recorded seismic phases.
4. The upper mantle structure of the British Isles is relatively poorly defined. The receiver function method allows mantle structures such as the 410 km and 660 km discontinuities to be investigated, the results from which may be linked to the current thermal state of the upper mantle. The study of these deeper structures by receiver function analysis is of particular interest in the British Isles. The long and complex geological history has led to the juxtaposition of distinct geological terranes, and analysis of receiver function enables the relationship between the major crustal structures and any lithospheric domains to be investigated.

Chapter 3

Receiver Function Analysis

3.1 Overview

The analysis of teleseismic receiver functions provides a powerful tool when investigating the acoustic impedance contrast structure of the crust and upper mantle (Langston, 1979; Sheehan et al., 1995; Dueker and Sheehan, 1997; Yuan et al., 1997; Gossler et al., 1999). The receiver function method uses teleseismic waves that originate from epicentral distances of between 20-100° to analyse structure beneath the station at which the earthquake is recorded. At such distances, the source P -wave energy has dispersed entirely from any source generated S -waves. However, within the P -wave group recorded at a 3-component station, S -wave energy can be detected on the horizontal component instruments. This shear-wave energy results from P -to- SV conversions as the source P -wave energy passes through seismic discontinuities beneath the station. The receiver function method highlights the P -to- S conversions and subsequent multiples observed within 3-component data by generating a source-equalised time-independent function that represents the interaction of a pulse-like plane-wave with the acoustic impedance structure beneath the station (Figure 3.1). Synthetic receiver functions can be generated to estimate the acoustic impedance model that caused the observed seismograms. This chapter introduces the theory of receiver function analysis, and the methods that are used in modelling and interpreting the data. Much of the data suitable for receiver function analysis within the British Isles have been recorded on short-period instruments, and the implications of using these data are discussed.

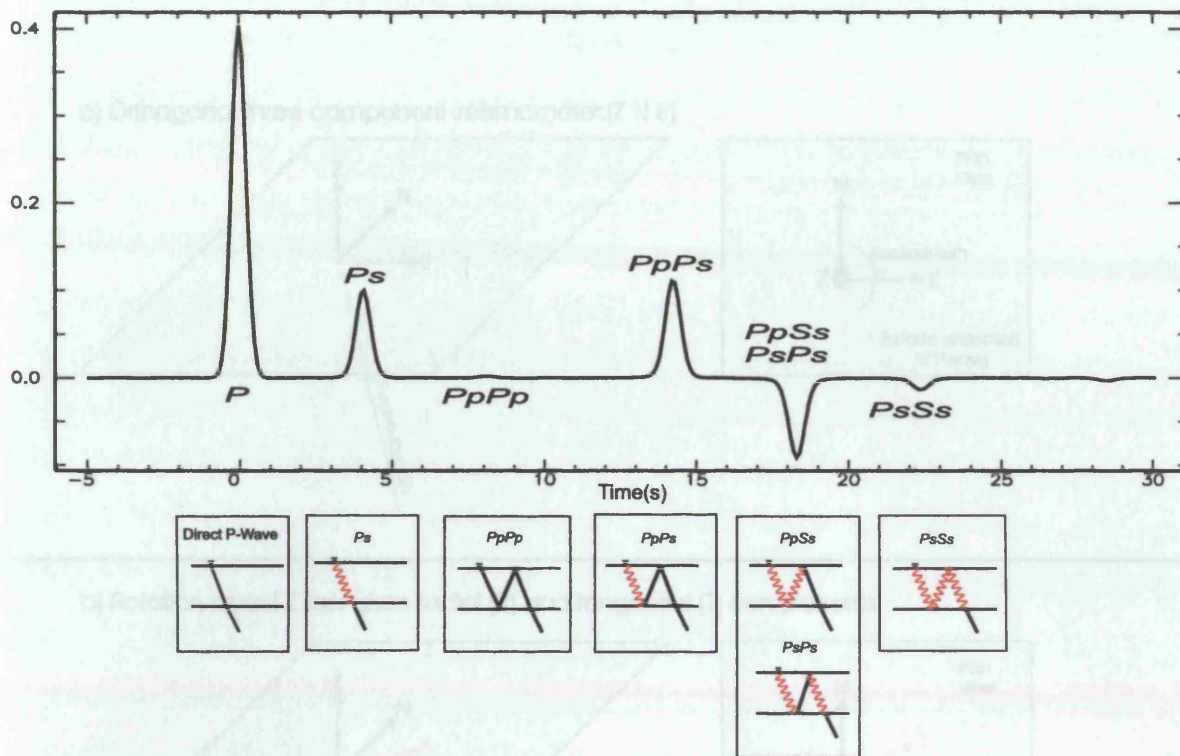


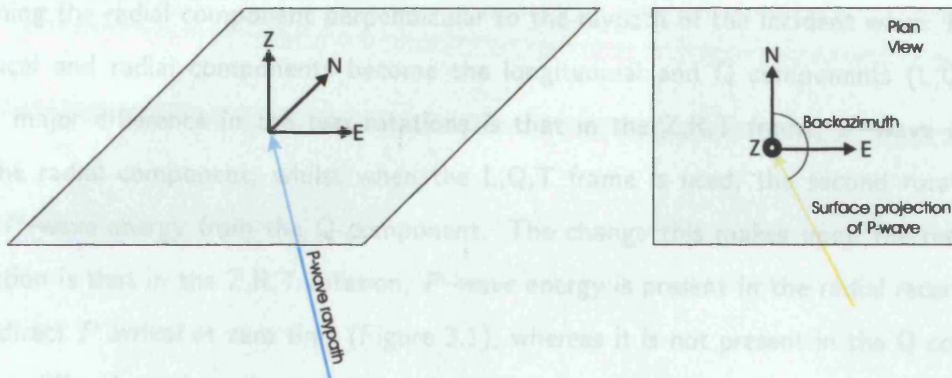
Figure 3.1: A synthetic receiver function, with the converted and multiple phases labelled.

3.2 Receiver Function Calculation

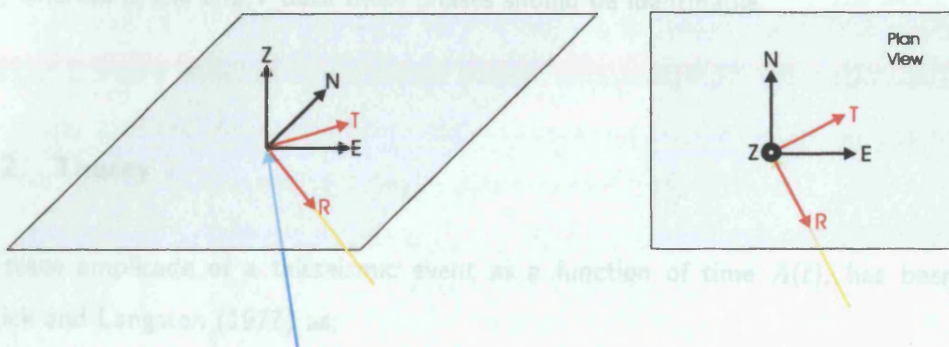
3.2.1 Rotation

Before the seismograms can be source equalised, they must first be rotated to isolate the P -to- SV conversions that the receiver function method utilises. Three-component seismic recording stations are conventionally set up to record the three orthogonal orientations of Vertical, North-South and East-West (Z,N,E) (Figure 3.2). These data must be rotated to separate the SV and SH converted phases on the horizontal N and E components. If a 1D earth model approximation holds, then only P -to- SV conversions should be generated. There are two rotation methods commonly used for isolating the P -to- SV converted energy during receiver function analysis. Firstly, the horizontal components (N,E) can simply be rotated about the vertical axis to give the radial and tangential components respectively (R,T). In this case the radial component is aligned parallel to the backazimuth (the azimuth from the receiver station to the event epicentre) and contains the SV converted energy. The tangential component will hold any SH converted energy if there is any degree of 2 or 3 dimensionality or anisotropy

a) Orthogonal three component seismometer.(Z N E)



b) Rotation about Z axis gives radial (R) and tangential (T) components.



c) A second rotation about the T axis gives the Longitudinal and Q components.

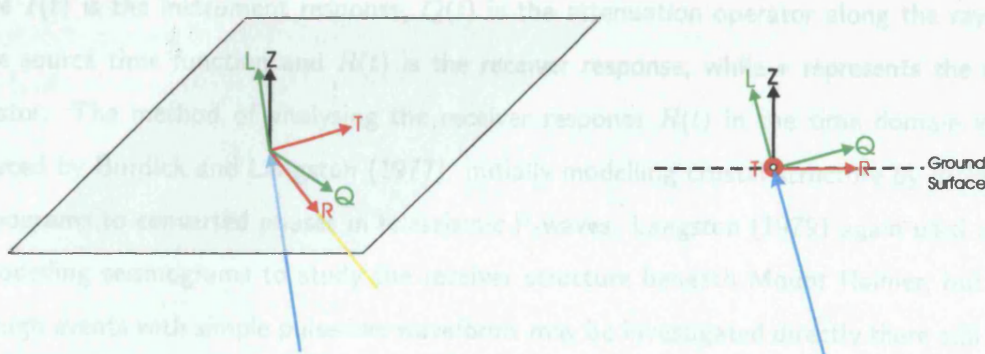


Figure 3.2: Schematic diagram of the three component seismogram rotations used to isolate P -to- SV conversions in receiver function analysis. a) A conventional setup of a three component seismometer, b) the rotation about the Z axis to radial (R) and tangential (T) components and c) a second rotation about the tangential axis can be used to produce the L and Q components.

within the receiver structure. The second scheme again rotates the horizontal components to radial and tangential components, but also includes a further rotation about the tangential axis aligning the radial component perpendicular to the raypath of the incident wave. In this case the vertical and radial components become the longitudinal and Q components (L,Q) respectively. The major difference in the two rotations is that in the Z,R,T frame, *P*-wave energy remains in the radial component, whilst when the L,Q,T frame is used, the second rotation eliminates the *P*-wave energy from the Q component. The change this makes upon the resultant receiver function is that in the Z,R,T rotation, *P*-wave energy is present in the radial receiver function as the direct *P* arrival at zero time (Figure 3.1), whereas it is not present in the Q component. The major difference this makes to the analysis of the receiver function is that *P*-to-*S* conversions from near surface interfaces could be masked within the direct *P* arrival on the Z,R,T rotated data, whereas in the L,Q,T data these phases should be identifiable.

3.2.2 Theory

The trace amplitude of a teleseismic event as a function of time $A(t)$, has been described by Burdick and Langston (1977) as;

$$A(t) = I(t) * Q(t) * S(t) * R(t) \quad (3.1)$$

where $I(t)$ is the instrument response, $Q(t)$ is the attenuation operator along the ray-path, $S(t)$ is the source time function and $R(t)$ is the receiver response, while $*$ represents the convolution operator. The method of analysing the receiver response $R(t)$ in the time domain was first introduced by Burdick and Langston (1977), initially modelling crustal structure by fitting synthetic seismograms to converted phases in teleseismic *P*-waves. Langston (1979) again used this method of modelling seismograms to study the receiver structure beneath Mount Rainier, but notes that although events with simple pulse-like waveforms may be investigated directly there still remain differences in the source time functions $S(t)$, which can cause significant differences in the synthetic seismogram calculation. This observation lead to the conclusion that equalising the seismograms to compensate for the difference in source time functions would be useful. Langston (1979) approached the source equalisation by simplifying Equation 3.1 and defining the vertical, radial and

tangential ground displacements, $D_Z(t)$, $D_R(t)$ and $D_T(t)$ as;

$$D_Z(t) = I(t) * S(t) * E_Z(t) \quad (3.2)$$

$$D_R(t) = I(t) * S(t) * E_R(t) \quad (3.3)$$

$$D_T(t) = I(t) * S(t) * E_T(t) \quad (3.4)$$

where $S(t)$ is the effective source time function, $I(t)$ is the instrument response and $E_Z(t)$, $E_R(t)$ and $E_T(t)$ are the vertical, radial and tangential impulse responses of the station respectively. These are of the form;

$$E(t) = \sum_{i=1}^n \alpha_i \delta(t - \tau_i) + \beta_i H[\delta(t - \tau_i)] \quad (3.5)$$

where α_i and β_i are functions of the reflection-transmission coefficients, $\delta(t)$ is the Dirac delta pulse, τ_i is the travel time of the i^{th} ray and $H[]$ is the Hilbert transform operator. To allow $E_V(t)$, $E_R(t)$ and $E_T(t)$ to be calculated it must be assumed that for near-vertical P-waves, the vertical component of ground motion includes little of any conversion or reverberation from within the crust. From this assumption it can be approximated that;

$$I(t) * S(t) \simeq D_Z(t) \quad (3.6)$$

To calculate $E_R(t)$ and $E_T(t)$, which represent the impulse responses in the radial and tangential directions, the source time function $S(t)$ and the instrument response $I(t)$ must be deconvolved from $D_R(t)$ and $D_T(t)$ respectively. The approximation that $I(t) * S(t) \simeq D_V(t)$ makes the deconvolution possible. The operation is simple in the frequency domain:

$$E_R(\omega) = \frac{D_R(\omega)}{I(\omega)S(\omega)} \simeq \frac{D_R(\omega)}{D_Z(\omega)} \quad (3.7)$$

$$E_T(\omega) = \frac{D_T(\omega)}{I(\omega)S(\omega)} \simeq \frac{D_T(\omega)}{D_Z(\omega)} \quad (3.8)$$

However, this deconvolution method is unstable when $D_Z(\omega) \rightarrow 0$. To control this instability the deconvolution "water level" method is used, filling any troughs in $D_Z(\omega)$ to a specified amplitude. A Gaussian filter is applied to limit the frequency bandwidth of the output receiver function. The modified equation for the receiver function is given by Langston (1979) as;

$$E'_R(\omega) = \frac{D_R(\omega)D_Z^*(\omega)}{\Phi_{ss}(\omega)} \cdot G(\omega) \quad (3.9)$$

where

$$\Phi_{ss}(\omega) = \max \{D_Z(\omega)D_Z^*(\omega), c \max [D_Z(\omega)D_Z^*(\omega)]\} \quad (3.10)$$

and

$$G(\omega) = e^{-\frac{\omega^2}{4a^2}} \quad (3.11)$$

$E'_R(\omega)$ is the deconvolved radial earth response (the receiver function) and $D_Z^*(\omega)$ is the complex conjugate of $D_Z(\omega)$. $\Phi_{ss}(\omega)$ is the autocorrelation of $D_Z(\omega)$, with the spectral troughs filled in up to the fraction c of the maximum amplitude of $D_Z(\omega)D_Z^*(\omega)$, the water level. This function takes the greater of the two values $D_Z(\omega)D_Z^*(\omega)$ and $c \max [D_Z(\omega)D_Z^*(\omega)]$, where the latter is the maximum value of the function $D_Z(\omega)D_Z^*(\omega)$ over all values of ω , scaled by some fraction c . $G(\omega)$ is a Gaussian function, where a is the parameter that controls the frequency bandwidth of the filter. The format of Equation 3.10 is more easily interpreted when re-written in the form

$$\phi_{ss}(\omega) = \max\{D_Z(\omega)D_Z^*(\omega), M\} \quad (3.12)$$

where

$$M = c.\max_{(all\omega)}\{D_Z(\omega)D_Z^*(\omega)\} \quad (3.13)$$

The water level method of deconvolution is similar to adding white noise in the deconvolution of reflection seismic data. The deconvolution is also possible in the time domain. This is achieved by applying the spiking deconvolution filter of D_V to the horizontal component seismograms. Both methods of deconvolution are commonly used in receiver function studies (Ramesh et al., 2002; Gossler et al., 1999; Zhu and Kanamori, 2000; Cassidy, 1992, e.g.). Neither method seems to offer any particular advantage when applied to receiver function analysis.

In early receiver function studies the amplitudes of the receiver functions were normalised by scaling the direct P -wave arrival on the radial receiver function to unit amplitude to eliminate the effect of peak amplitude variation. It is however important to maintain the true amplitude of the receiver function phases. The ratio of the amplitude of the direct P arrival on the radial and vertical components r_0/z_0 is sensitive to the near surface velocity (Ammon, 1991).

$$\frac{r_0}{z_0} = \frac{2\rho\eta\beta}{\beta^2 - 2\rho^2} \quad (3.14)$$

where ρ is the horizontal slowness of the plane wave, η_β is the vertical shear-wave slowness and β is the near surface shear-wave velocity. This reveals that the amplitude of r_0 is reduced when the near surface shear-wave velocity is reduced. Later phases, such as the Moho Ps conversion, are not sensitive to the near surface structure, and therefore in normalised receiver functions, are amplified. It is necessary to maintain the true amplitude of these phases to enable unambiguous determination of the ΔV_s across deeper boundaries (Cassidy, 1992). To enable the receiver functions from separate events to be directly comparable whilst also preserving the true amplitude of the receiver functions, the amplitude of the deconvolution of the vertical component from itself must produce a time series with an amplitude of one. This is applied to the horizontal components by dividing the radial and tangential receiver functions by the maximum amplitude of the deconvolved vertical component.

A receiver function from a 1D single layer crustal model is simple, with clearly identifiable phases (Figure 3.1). If the crustal acoustic impedance contrast model is more complicated, then this complication is reflected in the converted and multiple phases of the receiver functions. Using the crustal models from the British Isles (Figure 2.5) the resultant synthetic receiver functions are obviously more complicated than that resulting from the one layer case (Figure 3.3).

3.2.3 Short-period Data

The majority of the data suitable for receiver function analysis from the British Isles are recorded on short-period seismometers. The limited bandwidth of the data causes the results of the deconvolutions to contain negative lobes ringing the positive phases, and vice-versa. The result of using short-period synthetic seismograms to calculate receiver functions can be seen in Figure 3.4. It can immediately be seen that the data can be severely misinterpreted if the forward model does not account for the bandwidth of the data. When short-period receiver function data are analysed using inversion methods, the artificial negative lobes generated in the deconvolution are modelled as crustal low-velocity-zones. Because the inversion schemes do not account for the limited bandwidth, the negative lobes are likely to be modelled as primary conversions from within the crust.

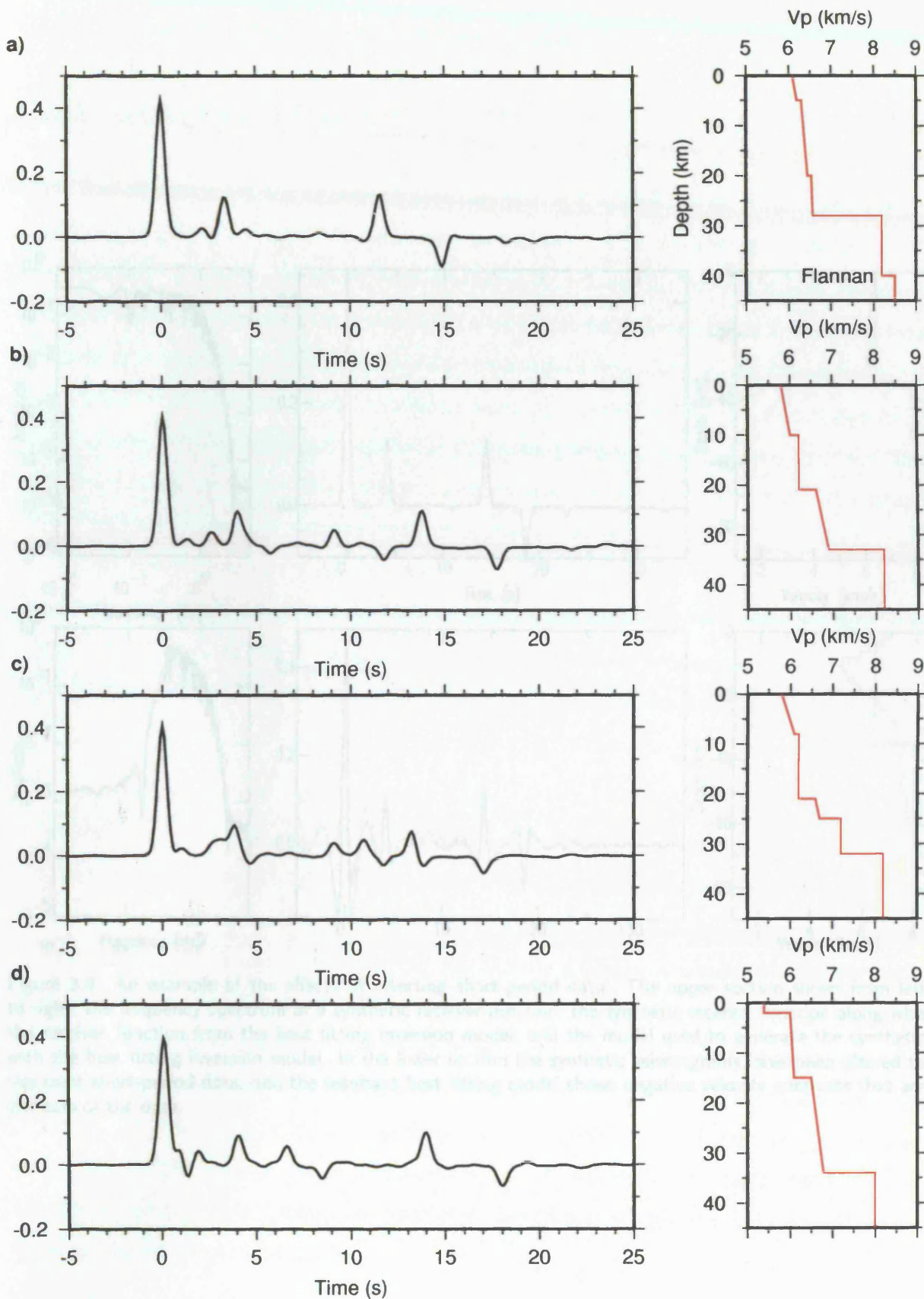


Figure 3.3: Synthetic receiver functions and their respective models from the British Isles; a) a model from the JUNE92 profile sampling the Flannan reflection (Morgan et al., 2000; Price and Morgan, 2000), b) a model from the section of the LISP profile spanning the lapetus Suture, c) a model from the CSSP profile perpendicular to LISP, d) a model from the LISP Delta profile (Edwards and Blundell, 1984).

3.3 Modelling Receiver Function Data

3.3.1 Objectives and Achievability

The time domain modelling of receiver structure was first carried out by Burdick and Langston (1977), who simply modelled the recorded seismograms. The methods of modelling receiver

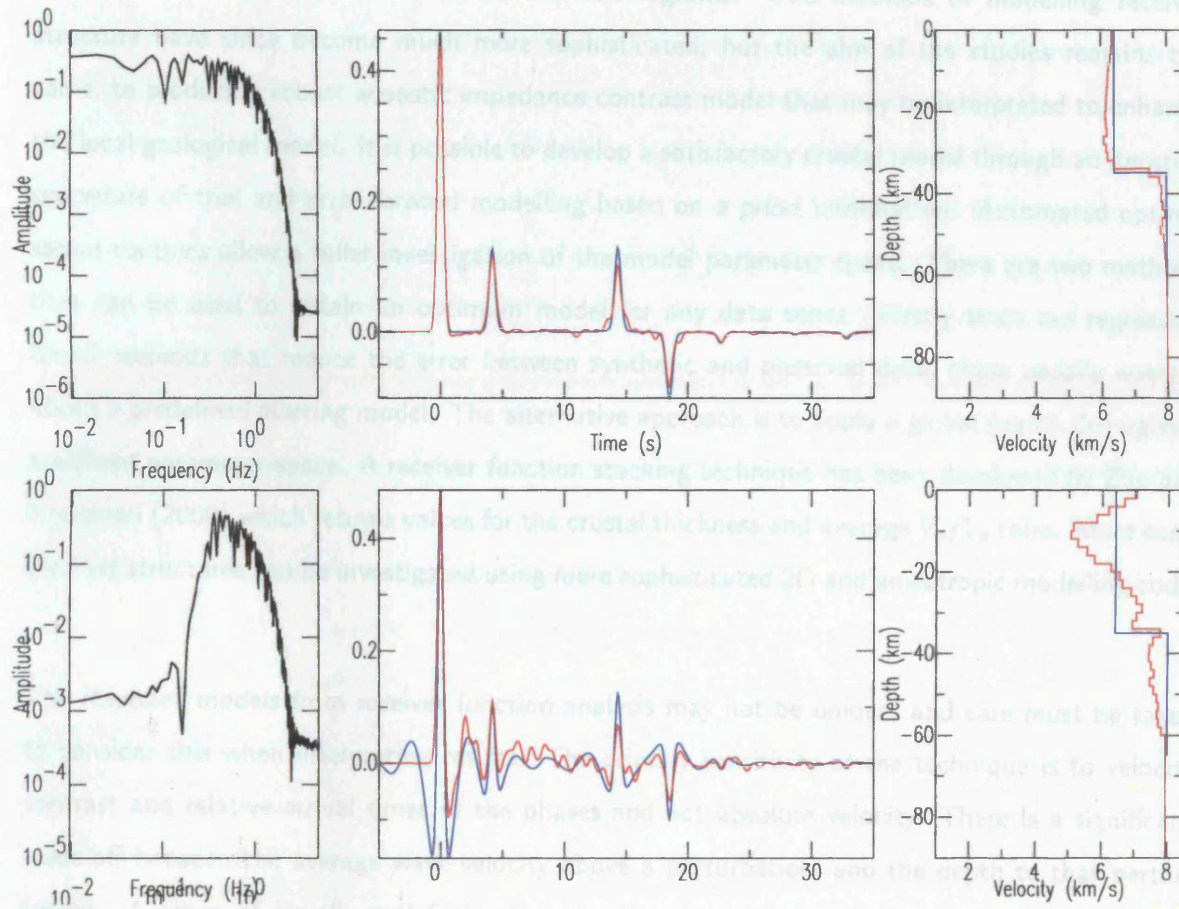


Figure 3.4: An example of the effects of inverting short-period data. The upper section shows from left to right; the frequency spectrum of a synthetic receiver function; the synthetic receiver function along with the receiver function from the best fitting inversion model; and the model used to generate the synthetic with the best fitting inversion model. In the lower section the synthetic seismograms have been filtered to represent short-period data, and the resultant best fitting model shows negative velocity contrasts that are artifacts of the data.

3.3 Modelling Receiver Function Data

3.3.1 Objectives and Achievability

The time domain modelling of receiver structure was first carried out by Burdick and Langston (1977), who simply modelled the recorded seismograms. The methods of modelling receiver structure have since become much more sophisticated, but the aim of the studies remains the same; to produce a robust acoustic impedance contrast model that may be interpreted to enhance the local geological model. It is possible to develop a satisfactory crustal model through an iterative procedure of trial and error forward modelling based on *a priori* information. Automated optimisation routines allow a fuller investigation of the model parameter space. There are two methods that can be used to obtain an optimum model for any data series. Firstly there are regression based methods that reduce the error between synthetic and observed data; these usually operate about a predefined starting model. The alternative approach is to apply a global search throughout a defined parameter-space. A receiver function stacking technique has been developed by Zhu and Kanamori (2000) which returns values for the crustal thickness and average V_p/V_s ratio. More complicated structures can be investigated using more sophisticated 2D and anisotropic modelling code.

The resultant models from receiver function analysis may not be unique, and care must be taken to consider this when interpreting results. The primary sensitivity of the technique is to velocity contrast and relative arrival times of the phases and not absolute velocity. There is a significant trade-off between the average wave-velocity above a perturbation, and the depth to that perturbation. A range of equally well fitting but significantly different velocity depth models can be very similar in the $(T_P - T_{P_s}) - \Delta V_p$ domain (Ammon et al., 1990). The sensitivity to traveltimes and high wave-number variation in velocity parallels the sensitivity of reflection seismology. The absolute velocity resolution is limited in both techniques by the range of horizontal slowness used to sample the medium; a single receiver function only samples one slowness. In regions where refraction velocities are available, *a priori* constraint on the velocity at any depth within the crust or upper mantle may greatly reduce the range of models capable of explaining the observed receiver functions (Ammon et al., 1990).

3.3.2 1D Forward Modelling

The forward modelling of receiver function data allows *a priori* models to be tested against the observed seismograms. In the case of the British Isles there have been numerous previous experiments providing starting models to be input into the receiver function modelling studies. In this study the 1D forward modelling routines developed by Ammon (1991) are used. The synthetic seismograms are generated using the reflection matrix method of Kennett (1983), producing the seismic response of a cylindrically symmetric medium. The method allows the response of the model to be truncated after a specified number of conversions, but does generate all of the arrivals that are excited by the incident plane-wave. The synthetic seismograms may be filtered to a limited bandwidth before the receiver function deconvolution, allowing the British Isles short-period data to be investigated directly. Importantly, using forward modelling allows *a priori* velocity information to be fixed in the model.

3.3.3 1D Linearised Inversion

The method of linearised time domain receiver function inversion is described by Ammon et al. (1990), in which they pose the mathematics of the forward problem, as described by Langston (1979). The forward problem describes the relationship between the various model parameters and the resultant data; in this case the crustal velocity structure and the receiver function respectively. The forward problem is described by Ammon et al. (1990) as

$$d_j = F_j[\mathbf{m}] \quad j = 1, 2, 3 \dots N \quad (3.15)$$

Where d_j represent the N data points of the receiver function and F_j is the function which operates upon the model parameter vector $[\mathbf{m}]$ to produce the synthetic waveform. This is the equation that is used in the forward modelling procedure, where the values within the parameter vector are changed manually. The aim of linearised inversion is to optimise the values within the parameter matrix to generate synthetic data that matches the observed data as closely as possible. Linear inverse problems may be solved by applying a simple regression. In the case of receiver function analysis, equation 3.15 is non-linear and cannot be solved by a process of direct linearisation and regression. This problem is overcome by starting the inversion scheme with a first guess model ($\mathbf{m}^0$), about

which the receiver function ($F_j(\mathbf{m})$) can be approximated as linear for small perturbations (δm).

$$F_j(\mathbf{m}) = F_j(m_1^0 + \delta m_1, m_2^0 + \delta m_2, m_3^0 + \delta m_3 \cdots m_p^0 + \delta m_p) \quad (3.16)$$

From this approximation it is then possible to expand the expression using Taylor's series about $\mathbf{m}^0$

$$F_j[\mathbf{m}] = F_j(\mathbf{m}^0) + \frac{\partial F_j}{\partial m_1} \delta m_1 + \frac{\partial F_j}{\partial m_2} \delta m_2 + \frac{\partial F_j}{\partial m_3} \delta m_3 \cdots + \frac{\partial F_j}{\partial m_p} \delta m_p + \text{higher order terms} \quad (3.17)$$

Simplifying gives

$$F_j[\mathbf{m}] = F_j[\mathbf{m}^0] + (\mathbf{D}, \delta \mathbf{m})_j + O\|(\delta \mathbf{m})\| \quad (3.18)$$

$(\mathbf{D}, \delta \mathbf{m})$ is the product of $\mathbf{D}$, the matrix of partial derivatives from equation 3.17, and the model correction vector $\delta \mathbf{m}$ (Ammon et al., 1990). Assuming that the series converges it is then possible to discard the higher order terms $O\|(\delta \mathbf{m})\|$, completing the linearisation of the problem

$$(\mathbf{D}, \delta \mathbf{m})_j \approx F_j[\mathbf{m}] - F_j[\mathbf{m}^0] \quad (3.19)$$

Substituting from Equation 3.15 simplifies slightly

$$(\mathbf{D}, \delta \mathbf{m})_j \approx d_j - F_j[\mathbf{m}^0] \quad (3.20)$$

This shows that the product of the partial derivative matrix and the model correction vector $(\mathbf{D}, \delta \mathbf{m})$ is simply the residual between $d_j[\mathbf{m}]$ and $d_j[\mathbf{m}^0]$, the observed and calculated waveforms respectively. Once linearised, it is possible to solve for the model parameters using conventional regression techniques.

Applying the linearised receiver function inversion directly often results in velocity models that show a significant amount of variation. Although it is by no means impossible for the Earth's velocity to have large amplitude variations, it is considered likely that these results are caused by under-damping of the inversion (Ammon et al., 1990). To reduce this rapid fluctuation in the velocity model with depth, a smoothness constraint is applied to the inversion. This is achieved by minimising the second derivative of the roughness of the resultant model $m(z)$, along with the

misfit between the original and synthetic data.

$$Roughness = \int (d^2 m / dz^2) dz \quad (3.21)$$

When solving equation 3.20, the solution is the model correction vector. It does not provide the absolute values of the model parameters. To apply a constraint on the smoothness of the model within the inversion scheme it is necessary to solve for the model vector. To achieve this, the linearised problem 3.20 is modified by adding the product of $(\mathbf{D} \cdot \mathbf{m}^0)_j$ to both sides of the equation, giving

$$(\mathbf{D}, \delta \mathbf{m})_j + (\mathbf{D} \cdot \mathbf{m}^0)_j \approx d_j - F_j[\mathbf{m}^0] + (\mathbf{D} \cdot \mathbf{m}^0)_j \quad (3.22)$$

This simplifies to

$$(\mathbf{D} \cdot \mathbf{m})_j \approx d_j - F_j[\mathbf{m}^0] + (\mathbf{D} \cdot \mathbf{m}^0)_j \quad (3.23)$$

This generates a linearised problem that can be solved for the model parameters. To combine the optimisation of the receiver function, and the minimisation of the resultant model roughness, the inverse problem must be altered to include the roughness constraint.

$$\begin{bmatrix} \mathbf{D} \\ \sigma \Delta \end{bmatrix} = \begin{bmatrix} r \\ 0 \end{bmatrix} + \begin{bmatrix} \mathbf{D} \cdot \mathbf{m}^0 \\ 0 \end{bmatrix} \quad (3.24)$$

Where r is the residual vector given by $d_j - F_j[\mathbf{m}^0]$, and the matrix Δ calculates the second derivative of the model file (Ammon et al., 1990). σ is the parameter that weights the bias of the inversion between minimising the misfit and the roughness of the model.

3.3.4 Global Inversion

The use of linearised inversion techniques is dependant upon the initial model $\mathbf{m}^0$ being close enough to the real model for the approximation of linearity to apply. However if there is little *a priori* information the starting model may not be close to the true solution, in which case the linearised inversion method may optimise toward a local minimum. Global inversion methods avoid this phenomenon by searching the entire parameter space. There have been a number of global inversion methods applied to the receiver function problem. The approach of Shibutani et al.

(1996) uses a semi-intelligent genetic algorithm to search the parameter space in a computationally efficient way. These full waveform global inversion techniques have not been used in this study. The $H-\kappa$ stacking method described in Section 3.3.7 has been used to provide a simple grid-search investigation of the global parameter space in the crustal thickness - V_p/V_s ratio domain.

3.3.5 1D Modelling Code

The forward modelling and inversion codes used in this study have different methods of approaching the problem of modelling receiver function data.

1. **Forward Modelling** Developed by Ammon (1991), this code is based upon Kennett's (1983) reflection matrix synthetic seismogram code. The models can be as simple or as complicated as the user desires. The variables that can be changed are; layer thickness, V_p , V_p/V_s and density.
2. **Linearised Inversion** Again developed by Ammon (1991), this regression based inversion uses the forward modelling code to generate the synthetic receiver functions. The optimisation of the code is limited to varying the V_p starting model. As the layer thickness is fixed, it is common practise to use a model with many thin layers so that the depth of velocity discontinuities can vary. The code does not allow for the use of synthetic seismograms with limited bandwidth in the forward model.

As discussed in Section 3.3.1, receiver function data is sensitive to velocity contrast and the relative arrival time of the phases, and therefore there is no totally unique solution when modelling the data. The way that the model parameter space is defined affects the number of possible solutions that are investigated, and with this in mind, it is important to constrain the model with *a priori* information wherever possible. The forward modelling code allows full control over all the model parameters. The linearised inversion does not allow any control on the parameter space. When using a starting model with many thin layers it is possible to produce a well fitting receiver function with many different velocity-depth models. The inversion routine does not allow the bandwidth of the seismograms to be limited in the forward model. As discussed in Section 3.2.3 the inversion routines can produce phantom structures that are artifacts of the short-period data, so care must be taken when analysing the resultant velocity models.

3.3.6 Real Earth: Anisotropic and Multidimensional Modelling

In the previous sections the modelling of 1D crustal structures has been discussed. However, geological structures are naturally complex, and rarely conform to a 1D approximation. The teleseismic receiver function method uses the radial component of the deconvolved seismograms to analyse the 1D acoustic impedance structure beneath a station. When using the 1D approximation, it is assumed that all primary conversions are P -to- SV , and as such are recorded on the radial component seismogram. P -to- SH conversions can be produced by plane P -wave interaction with 2 or 3D structures, but may also be produced by seismic velocity anisotropy within horizontal layers. The tangential components of receiver functions contain any SH conversions that may have occurred. The amplitude of any P -to- SH conversions created by either of these mechanisms varies with backazimuth. This variation is also reflected in the amplitude of the phase within the radial component. The cyclicity of the amplitude variation can characterise the type of structure that has caused the SH conversion (Figure 3.5a). In the case of a 2D planar dipping layer, the amplitude of the P -to- SH conversion for that layer cycles once between positive and negative polarity through 360° (Cassidy, 1992). Receiver function modelling studies have used a hexagonally symmetric model to calculate the synthetic seismograms (Levin and Park, 1998), which describes the angular variation of seismic velocity (V_p and V_s) away from an axis of symmetry $\hat{\mathbf{w}}$:

$$\rho V_p^2 = A + B \cos 2\theta + C \cos 4\theta \quad (3.25)$$

$$\rho V_s^2 = D + E \cos 2\theta \quad (3.26)$$

Where θ is the angle away from $\hat{\mathbf{w}}$. The tilt and orientation of $\hat{\mathbf{w}}$ may also be defined. A, B, C, D and E are elastic constants, $A = \lambda + 2\mu$ and $D = \mu$, where λ and μ are the Lamé constants. In an isotropic media $B = C = E = 0$. If B and $E > 0$ then $\hat{\mathbf{w}}$ defines the fast axis for propagation, conversely if B and $E < 0$ then $\hat{\mathbf{w}}$ is the slow axis. The elastic parameter $C \ll 0$ and has been set to $C = 0$ by Levin and Park (1998). If $\hat{\mathbf{w}}$ is vertical, which may be the case in horizontal strata, then there will be no P -to- SH conversion. When a horizontal axis of symmetry is used in this hexagonal model the variation in the P -to- SH phase cycles twice between positive and negative values through 360° (Figure 3.5b).

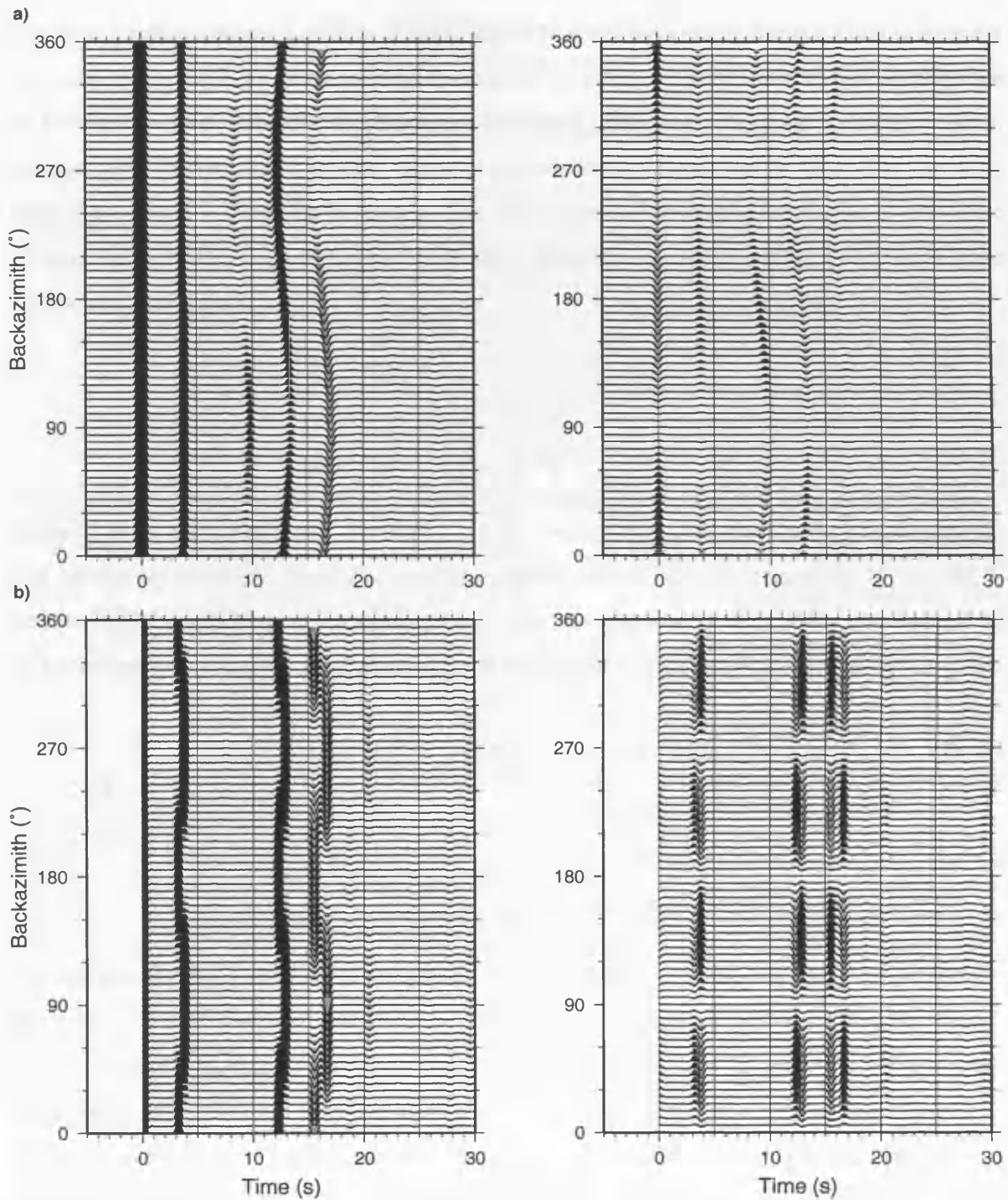


Figure 3.5: Synthetic radial (left) and tangential (right) receiver functions plotted relative to backazimuth for a) a 1 layer crust (30 km) with a Moho dip of 10°E, and b) a 1 layer crust with 7% seismic velocity anisotropy, with a fast velocity axis parallel to a horizontal axis of anisotropy striking N-S.

3.3.7 H- κ Stacking

The H - κ stacking method is applied to large receiver function datasets to produce a robust estimate of crustal thickness (H) and average V_p/V_s ratio (κ) for any three-component station. A precursor to the method was developed by Zandt and Ammon (1995) where they estimated the crustal thickness and Poisson's ratio of many receiver functions at several locations by measuring the delay time of the Moho P_s and $PpPs$ phases. The arrival times of these phases relative to the direct P -wave can provide a unique solution for H and κ given the average crustal V_p . The travel times for P_s and $PpPs$ are;

$$T_{P_s} = \frac{H}{V_p} \left(\sqrt{\frac{V_p^2}{V_s^2} - \rho^2 V_p^2} - \sqrt{1 - \rho^2 V_p^2} \right) \quad (3.27)$$

$$T_{PpPs} = \frac{H}{V_p} \left(\sqrt{\frac{V_p^2}{V_s^2} - \rho^2 V_p^2} + \sqrt{1 - \rho^2 V_p^2} \right) \quad (3.28)$$

where H is the crustal thickness, V_p and V_s are the average crustal P -wave and S -wave velocities and ρ is the ray parameter. Both T_{P_s} and T_{PpPs} are dependent upon the crustal V_p , V_p/V_s and H . Equations 3.27 and 3.28 can be rearranged to provide two solutions for H , and V_s can subsequently be calculated by using an average crustal V_p and setting the two values of H equal to one another.

$$H = \frac{T_{P_s}}{\sqrt{\frac{1}{V_s^2} - \rho^2} - \sqrt{\frac{1}{V_p^2} - \rho^2}} = \frac{T_{PpPs}}{\sqrt{\frac{1}{V_s^2} - \rho^2} + \sqrt{\frac{1}{V_p^2} - \rho^2}} \quad (3.29)$$

$$V_s = \frac{1}{\left[\frac{\left(\frac{1}{V_p^2} - \rho^2 \right) (T_{P_s} + T_{PpPs})^2}{(T_{P_s} - T_{PpPs})^2} \right] + \rho^2} \quad (3.30)$$

The method of picking the time of the P_s and $PpPs$ phases, and calculating multiple solutions for any station adopted by Zandt and Ammon (1995) is time consuming. Zhu and Kanamori (2000) have developed a method to produce a single solution for the crustal H and κ from the receiver functions for any station. They calculate the arrival times for the P_s , $PpPs$ and P_sPs (and $PpSs$) phases for a grid of H - κ values for each receiver function of ray parameter ρ and average crustal P -wave velocity V_p . The amplitudes of each receiver function at each of these traveltimes, are summed and added to the stack $S(H, \kappa)$.

$$S(H, \kappa) = \sum_{j=1-n} w_1 r_j(t_1) + w_2 r_j(t_2) - w_3 r_j(t_3) \quad (3.31)$$

where r_j is a radial receiver function, t_1 , t_2 and t_3 are the predicted P_s , $PpPs$ and P_sPs times, and w_1 , w_2 and w_3 are weighting functions of the stack. Due to the differing moveout curves of P_s , $PpPs$ and P_sPs in the H - κ domain the point at which the phases intersect gives the unique crustal H and κ solution. Figure 3.6 shows the intersection of the P_s and $PpPs$ phases. This point is expressed in the stack as the maximum amplitude.

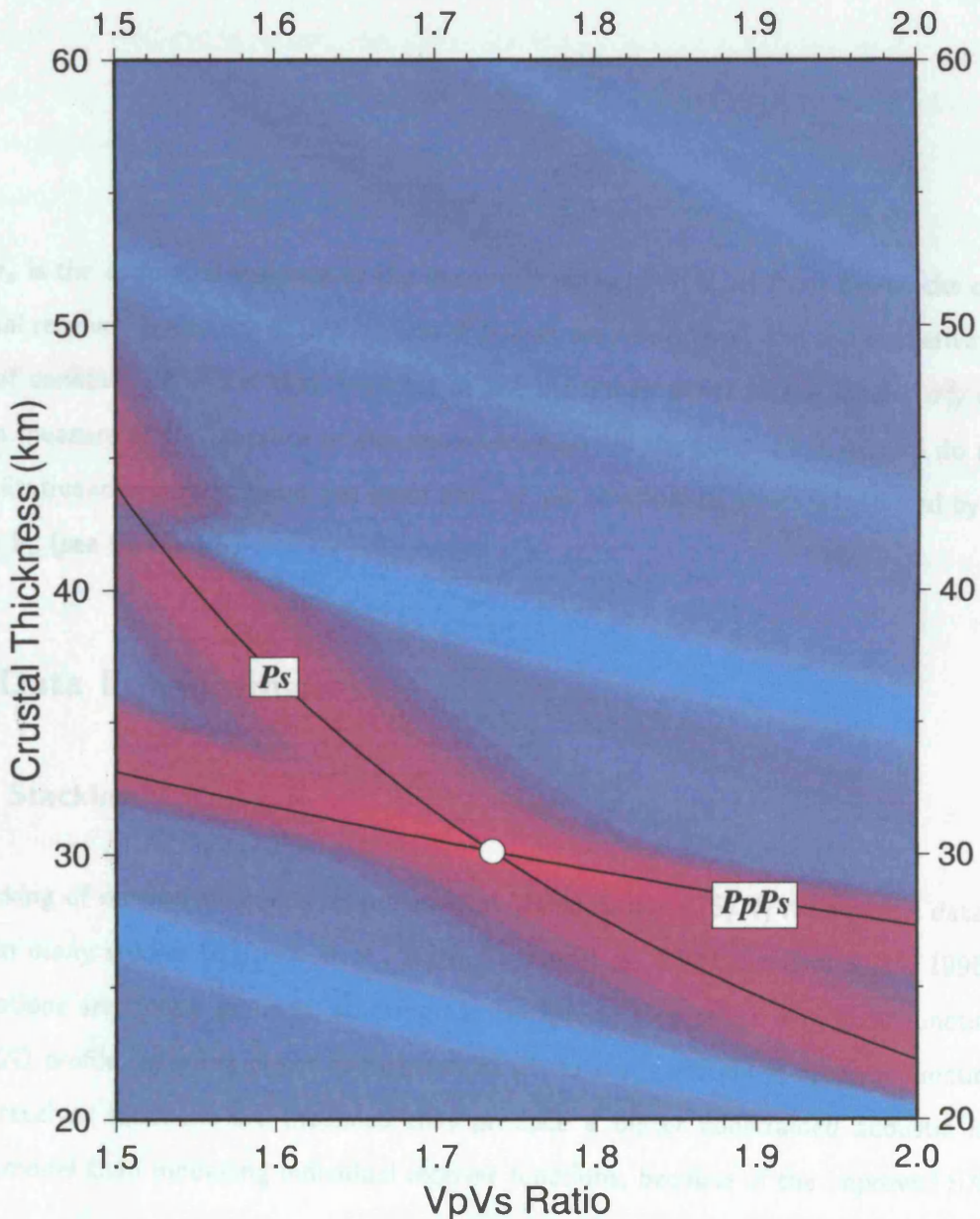


Figure 3.6: An example stack $S(H, \kappa)$ for a synthetic receiver function from a 1 layer crustal model of $H = 30$ km, $V_p/V_s = 1.73$. The moveout of the P_s and $PpPs$ phases are highlighted, and the point of intersection gives the H - κ solution for the synthetic seismograms.

The H - κ stacking method has been used successfully on many data from around the globe, for

example Cheverot and van der Hilst (2000) and Ramesh et al. (2002). The primary advantages of this method are that; 1) the stacking is automated, rapidly producing results, negating the need to identify and pick the times of the Ps , $PpPs$ and $PsPs$ phases, and 2) the uncertainties of the maximum point of the stack can be calculated. Zhu and Kanamori (2000) state that the Taylor expansion of $S(H, \kappa)$ at the maximum point of the stack, omitting the higher order terms, gives the variance of the stack σ in both H and κ .

$$\sigma_H^2 = 2\sigma_s / \frac{\partial^2 S}{\partial H^2} \quad (3.32)$$

$$\sigma_\kappa^2 = 2\sigma_s / \frac{\partial^2 S}{\partial \kappa^2} \quad (3.33)$$

where σ_s is the estimated variance of the maximum value of $S(H, \kappa)$ from the stacks of all of the individual receiver functions. $\partial^2 S / \partial H^2$ and $\partial^2 S / \partial \kappa^2$ are respectively the second derivatives of the planes of constant H and κ that intersect at the maximum point of the stack. σ_H^2 and σ_κ^2 are simply a measure of the variance of the maximum point of the stack $S(H, \kappa)$ and do not include any qualitative information about the input data or any estimate of errors introduced by the choice of input V_p (see Chapter 5).

3.4 Data Enhancement

3.4.1 Stacking

The stacking of receiver functions to enhance the signal-to-noise (S/N) ratio of the data has been applied in many studies (Ramesh et al., 2002; Gossler et al., 1999; Sandvol et al., 1998). Where many stations are closely grouped, stacking can be applied to produce a receiver function section along a 2D profile, allowing direct comparison of the lateral variation in receiver function. When stacked receiver functions are modelled they produce a better constrained acoustic impedance contrast model than modelling individual receiver functions, because of the improved S/N ratio.

When stacking receiver functions from a single station, the parameters that must be considered are; 1) the local variation in geological structure, and 2) the ray parameter of the events to be summed. The receiver function data can be classified by the backazimuth from which the seismic waves

approach the recording station, and therefore by the geological structure that they might sample. By summing data from bins with a wide azimuthal aperture, variations in the data are likely to be smoothed. Conversely if narrow bins are used then less smoothing will occur, but the resultant data may have a poorer S/N ratio. Variations of the seismograms with respect to ray parameter (ρ) not only includes any geological variation that the data might sample, but also variations in the timing of the phases (Equations 3.27 and 3.28) and the amplitude of the phases. Figure 3.7a shows the variation in receiver function with ρ for a simple 1D crustal velocity-depth model, revealing a small variation in the timing of the P_s and $PpPs$ phases at crustal depths. There is a more notable variation in the amplitude of the phases, with the closer events having greater amplitudes due to the larger inclination incident angle of the P -wave group at each velocity discontinuity.

In order to position receiver function data correctly when producing a 2D profile, the point at which each receiver function raypath pierces the Moho is calculated given the backazimuth and epicentral distance. If the target structure is not the Moho then the calculations are altered accordingly. Once every pierce point has been calculated the receiver functions are then projected perpendicularly from the crustal pierce points on to a 2D profile. At this point the receiver functions may be binned, and then stacked to produce a 2D profile, with regularly spaced seismograms if required.

3.4.2 Moveout Corrections

When stacking receiver function data the ray parameter of the seismogram must be considered because of the variation that this can cause to the timing (Equations 3.27 and 3.28) and amplitude of the phases. The P_s phase arrives earlier in time with increased ρ , whereas the $PpPs$ phase is delayed with increasing ρ (Figure 3.7). It is possible to correct the receiver function data for the variation in the timing of the phases due to changes in ray parameter. Correcting the receiver function data for the moveout in P_s phases caused by the change in ray parameter is similar to the NMO correction in reflection seismology. Applying these corrections produces stronger P_s phases in the stacked receiver functions, whilst reducing the amplitude of the $PpPs$ phase in the stack. The amplitude of the $PpPs$ phase is reduced because the moveout curve with respect to ρ is opposite to that of the P_s phase. Therefore when the P_s moveout correction is applied, the time

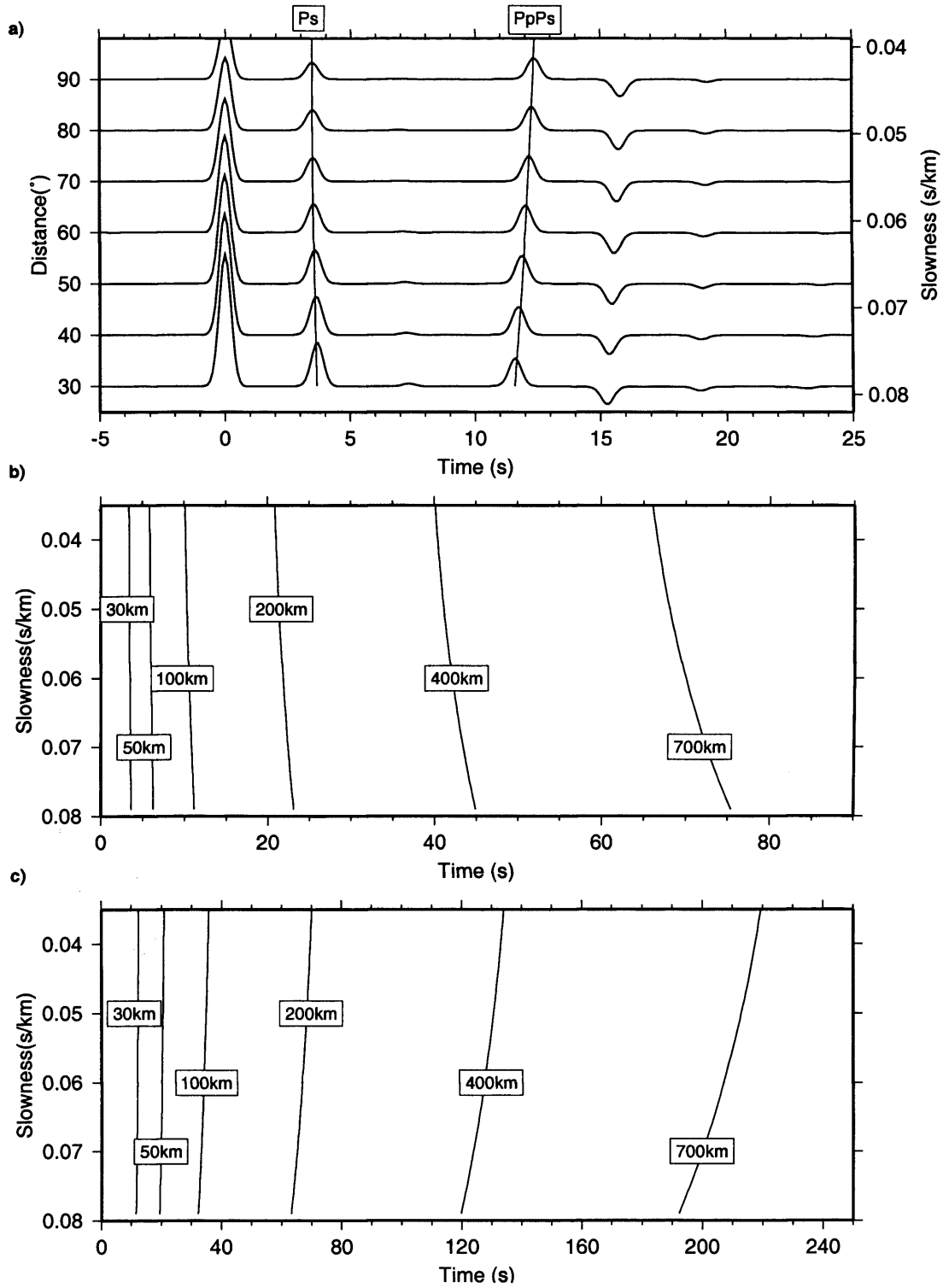


Figure 3.7: a) Synthetic receiver functions from a 1 layer crust of 30 km thickness, $V_p = 6.5$ and $V_p/V_s = 1.73$. The P_s and $PpPs$ phase moveout curves for the model are overlaid. b) & c) The P_s and $PpPs$ moveout curves respectively for a range of conversion depths from the IASPEI91 velocity model.

variation with respect to ρ of the $PpPs$ phase is increased. This method of data enhancement is of particular importance when structures within the upper mantle are being investigated. At the depth of mantle discontinuities, e.g. 410 km or 660 km, the moveout of Ps phases resulting from events at epicentral distances of between 30° and 100° is considerable (Figure 3.7 b&c), and therefore becomes more important in studies that target these deeper structures. Many receiver function studies of the upper mantle have used the moveout correction to improve the quality of their data (e.g. Dueker and Sheehan, 1997; Gossler et al., 1999).

3.4.3 Migration

Producing a 2D profile by calculating the pierce points of the receiver function data through a particular interface is a very simple form of migration, repositioning the data relative to the source of that particular phase. The moveout correction is also a form of depth migration. Once the times have been corrected for moveout from a given velocity model, the depth of the phases can be directly interpreted from the arrival time. A more sophisticated migration technique has been employed to image crustal and upper mantle structures (e.g. Gossler et al., 1999; Yuan et al., 1997; Knapmayer and Harjes, 2000). In this method three-component data is used to calculate the backazimuth and incident angle of the Ps phase for each receiver function. The data is then projected to the true 3D position along this raypath, for a given velocity model. In order to produce a 2D section the data are projected on to a linear profile, and then binned and stacked on a distance-depth grid. Where stations are very closely spaced, Kirchhoff migration techniques have been successfully applied.

3.5 Summary

The analysis of teleseismic receiver functions can reveal much about the seismological properties of the crust and upper mantle. The method investigates the locally generated P -to- S conversions and subsequent multiples following the P -wave group. This provides the opportunity to study the acoustic impedance structure, and in particular the shear-wave velocity structure beneath the recording station. In this study a suite of tools are used to investigate the receiver functions from a set of seismic stations in the British Isles;

1. **1D Forward Modelling:** The code allows the interactive iterative testing of models against

receiver function data from single stations. The models can be constrained by *a priori* velocity-depth structure information.

2. **$H-\kappa$ Stacking** A simple method of analysing all of the data for one seismic station. The stack returns robust values for the crustal thickness and average V_p/V_s ratio, but must be constrained by using an *a priori* average crustal V_p to calculate the stack.
3. **1D Inversion** Based on forward modelling code, the inversion scheme uses an automated routine to optimise the synthetic data. This gives a non-subjective search of the parameter space, returning the model that produces seismograms that best fit the observed data.
4. **Real Earth Modelling** The 2D and anisotropic forward modelling code allow more complex models to be tested against the receiver function data. The effects of bulk scale crustal anisotropy can be tested on the results that the $H-\kappa$ stacking and 1D modelling produce.
5. **Imaging** The data enhancement techniques can produce images of crustal and upper mantle structures. Data from a number of stations can be directly compared by generating 2D profiles of equally spaced receiver functions. Structures within the upper mantle can be investigated by applying moveout corrections and migration to the data.

Applying these techniques to the data from the British Isles will help investigate the questions raised in Chapter 2. The $H-\kappa$ stacking results will help constrain the onshore crustal thickness, whilst providing information on the variation of the average crustal V_p/V_s ratio. Using the forward modelling and inversion routines will provide more detailed crustal velocity models, which can be constrained by *a priori* information from previous deep seismic reflection and seismic refraction studies. The imaging techniques can provide 2D cross-sections, especially where the stations are closely spaced. Analysis of broadband data allows upper mantle structures to be investigated and compared with the structural variations from within the crust.

Chapter 4

British Isles Data

4.1 Introduction

Receiver function analysis requires three component teleseismic data. There are 34 seismic recording stations within the British Isles from which the data are suitable for teleseismic receiver function analysis. This chapter details those stations and the data that are available from them; covering the event catalogue, data retrieval from archive, data format conversion, data preparation and receiver function calculation. The catalogue of events that produce high quality receiver functions suitable for analysis is compared with the original catalogue of events. The event parameters which control the suitability of seismic data for use in the present receiver function study are discussed.

4.2 Data Sources

The seismic recording stations used in the present receiver function study span the British Isles, recording data that sample all of the major tectonic terranes (Figure 4.1). The majority of data have been recorded on short-period instruments, with only a sparse coverage of broadband instruments. The main source of data is from the British Geological Survey's (BGS) seismic monitoring network. In addition to these stations there are a small number of instruments maintained by; The Atomic Weapons Establishment (AWE), The Incorporated Research Institutes for Seismology (IRIS), and also a temporary array named SPICED run by the University of Leeds.

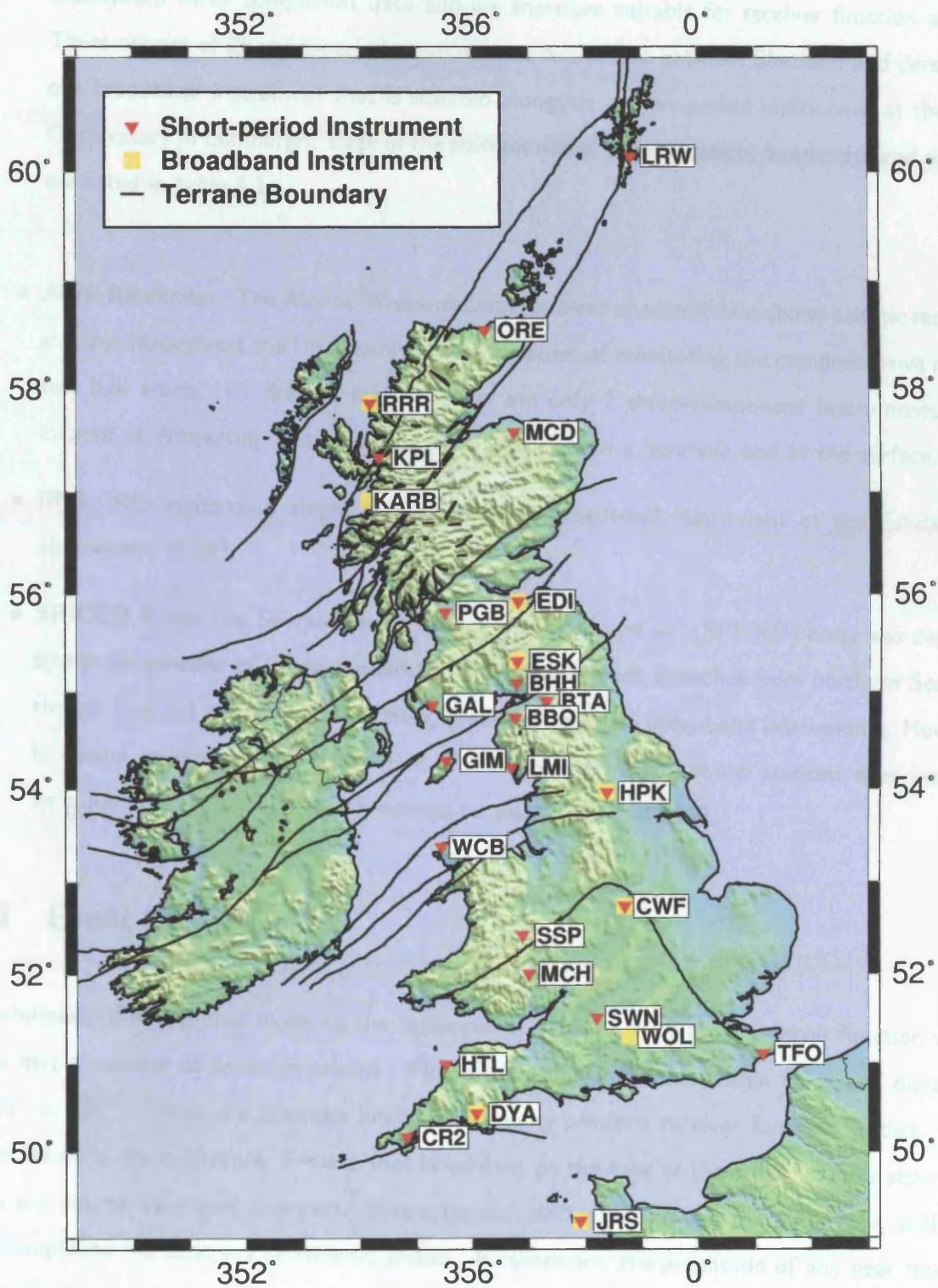


Figure 4.1: A map of all the seismic monitoring stations used in the receiver function project.

- **BGS:** Of the 141 instrument BGS seismological monitoring network there are 26 instruments that record three component data and are therefore suitable for receiver function analysis. These consist of 25 short-period seismometers distributed between Shetland and Jersey, and one broadband instrument that is situated alongside a short-period instrument at the Royal Observatory in Edinburgh. Each of the stations name, code, location, bandwidth and operator are listed in table 4.1.
- **AWE Blacknest:** The Atomic Weapons Establishment operate 9 broadband seismic recording stations throughout the UK, mainly for the purposes of monitoring the comprehensive nuclear test ban treaty. Of these 9 stations there are only 2 three-component instruments, both located at Wolverton in southern England which are in a borehole and at the surface.
- **IRIS:** IRIS maintain a single three-component broadband instrument at the Eskdalemuir observatory (ESK).
- **SPICED Array** The Seismic Profile of the Inner CorE and D'' (SPICED) array was deployed by the Universities of Leeds and Bristol. The profile which stretches from northern Scotland through England to France utilises BGS, Blacknest and IRIS broadband instruments. However, to ensure an even spatial distribution of stations 6 extra broadband stations were installed for a period of approximately 18 months between 1998 and 2000.

4.3 Event Selection

The teleseismic events that make up the catalogue of data for the present receiver function study have met a number of selection criteria. Firstly the events must fall within epicentral distances of 25° to 100° . These are common limits set in many previous receiver function studies. The lower bound is set so that the P -wave that is incident on the base of the crust has fully separated from any source generated S -waves. Events beyond 100° pass through the outer core, and this can complicate the observed teleseismic phases. Furthermore, the amplitude of any near receiver P -to- S conversions from events at distances $>100^\circ$ are low amplitude, because the incidence angle of the P -wave is approaching vertical.

Name	Network	Code	Latitude	Longitude	Bandwidth
Lerwick	BGS	LRW	60.13600	-1.17790	SP
Reay	BGS	ORE	58.54800	-3.76220	SP
Rubha Reidh	BGS	RRR	57.85770	-5.80670	SP
Coleburn Distilary	BGS	MCD	57.58270	-3.25410	SP
Plockton	BGS	KPL	57.33910	-5.65270	SP
Edinburgh	BGS	EDI	55.92330	-3.18610	SP BB
Glenifferbraes	BGS	PGB	55.81000	-4.47800	SP
Eskdalemuir	BGS	ESK	55.31670	-3.20500	SP
Talkin	BGS	BTA	54.90570	-2.68410	SP
Howatts Hill	BGS	BHH	55.09280	-3.21870	SP
Galloway	BGS	GAL	54.86640	-4.71140	SP
Bothel	BGS	BBO	54.7367	-3.2464	SP
Isle of Man	BGS	GIM	54.29230	-4.46700	SP
Millom	BGS	LMI	54.22060	-3.30700	SP
Haverah Park	BGS	HPK	53.95540	-1.62400	SP
Church Bay	BGS	WCB	53.37820	-4.54650	SP
Charnwood Forest	BGS	CWF	52.73820	-1.30710	SP
Michael Church	BGS	MCH	51.99770	-2.99830	SP
Stoney Pound	BGS	SSP	52.41770	-3.11190	SP
Swindon	BGS	SWN	51.51300	-1.80050	SP
Folkstone	BGS	TFO	51.11360	1.14060	SP
Hartland	BGS	HTL	50.99440	-4.48500	SP
Yardsworthy	BGS	DYA	50.43520	-3.93090	SP
Rosemanowes2	BGS	CR2	50.16690	-5.16870	SP
Maison St Louis	BGS	JRS	49.19240	-2.09170	SP
Eskdalemuir	IRIS	ESK	55.31670	-3.20500	BB
Wolverton	Blacknest	WOL	51.31	-1.22	BB
Rubha Reidh	SPICED	RRRB	57.85770	-5.80670	BB
Arisaig	SPICED	KARB	56.9188	-5.8290	BB
Haverah Park	SPICED	HPKB	53.95540	-1.62400	BB
Charnwood Forest	SPICED	CWFB	52.73820	-1.30710	BB
Yardsworthy	SPICED	DYAB	50.43520	-3.93090	BB
St. Aubins	SPICED	JSAB	49.1879	-2.1709	BB

Table 4.1: Station name and code information for the main seismic recording stations used in the receiver function project. Latitude and longitude are in degrees, SP = Short Period, BB = Broadband.

The minimum magnitude of events included in the catalogue was M_b 6.0. The minimum magnitude must be applied to maintain a reasonable signal-to-noise ratio for the incident P -wave and subsequent conversions and multiples. The list of events suitable for the catalogue has been generated using the USGS National Earthquake Information Centre (NEIC) search facility (<http://neic.usgs.gov/neis/epic/epic.html>). This search allows events from limited magnitude and date ranges from a circular area of specified radius and centre point to be included in the catalogue. The event list produced using this facility has been used to extract the teleseismic data required for receiver function analysis from the digital data archives at BGS, AWE and IRIS (see Section 4.4).

A list of approximately 800 events of $M_b \geq 6.0$ from between 1990 and 2001 has been generated. Figure 4.2 shows the global distribution of the events in the list. It can be seen from this that the majority of the suitable events originate around the Pacific rim. More proximal events are also located through Asia, and along the Atlantic spreading ridge. When the list is transformed into epicentral distance and backazimuth it can be seen that many of the events fall at distances between 70° and 100° (Figures 4.3 & 4.4). There is a notable lack of events from southerly backazimuths (Figure 4.5).

4.4 Data Retrieval

To compile a catalogue of events from the various data archives several different retrieval methods have been used. For the BGS and AWE data the e-mail based Automatic Data Request Manager (Auto DRM) must be used to access their digital event based catalogues. The IRIS data have been accessed by the Windows Extracted from Event Data (WEED) software. The data from the SPICED array have been accessed using custom software at the University of Leeds.

The data from the majority of stations have been acquired using the Auto DRM system at the BGS. With this system an individual e-mail request must be submitted for the time window of each required event. These messages must conform to the following syntax:

BEGIN GSE

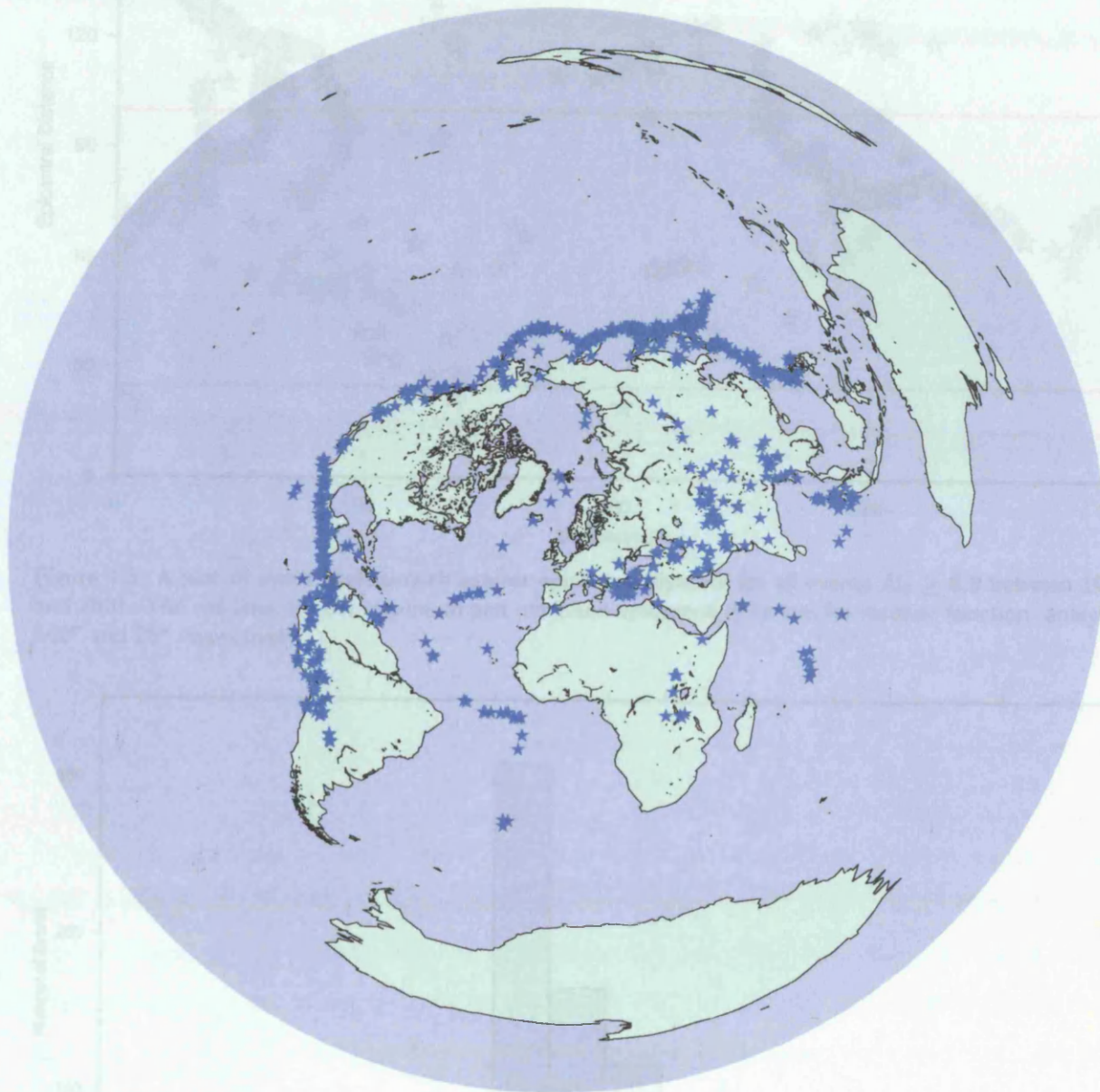


Figure 4.2: A map of events of $M_b \geq 6.0$ at epicentral distances less than 100° from 1990 to 2001. The map is plotted using an azimuthal equidistant projection centred on the British Isles.

Figure 4.3: A histogram of the range of epicentral distances for a set of events of $M_b \geq 6.0$ at epicentral distances between 10° and 100° from 1990 to 2001.

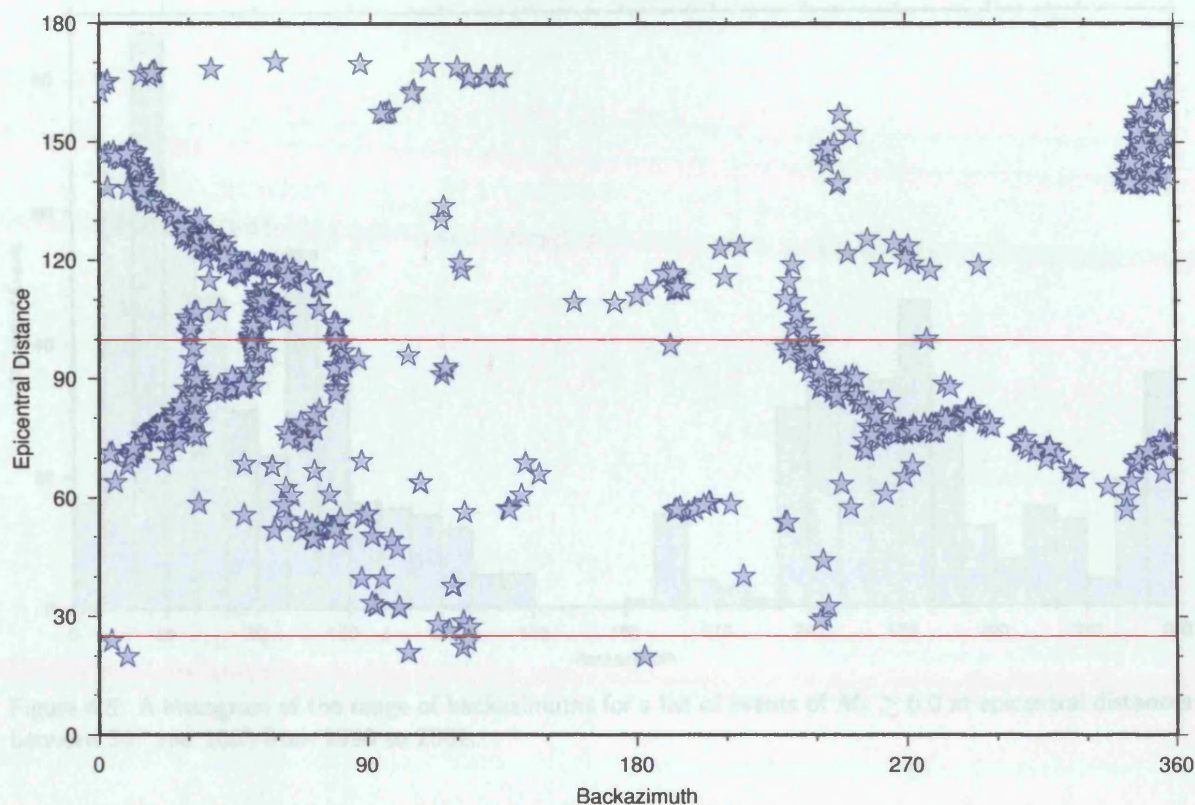


Figure 4.3: A plot of event backazimuth against epicentral distance for all events $M_b \geq 6.0$ between 1990 and 2001. The red lines are the maximum and minimum epicentral distances for receiver function analysis, 100° and 25° respectively.

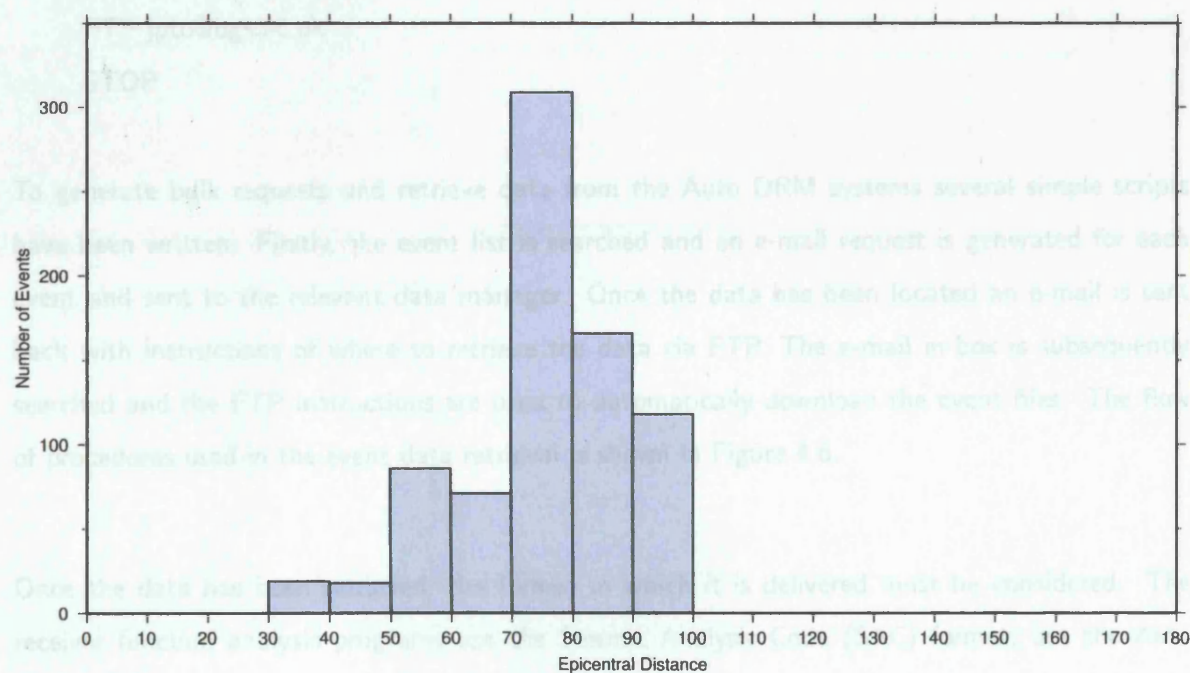


Figure 4.4: A histogram of the range of epicentral distances for a list of events of $M_b \geq 6.0$ at epicentral distances between 30° and 100° from 1990 to 2001.

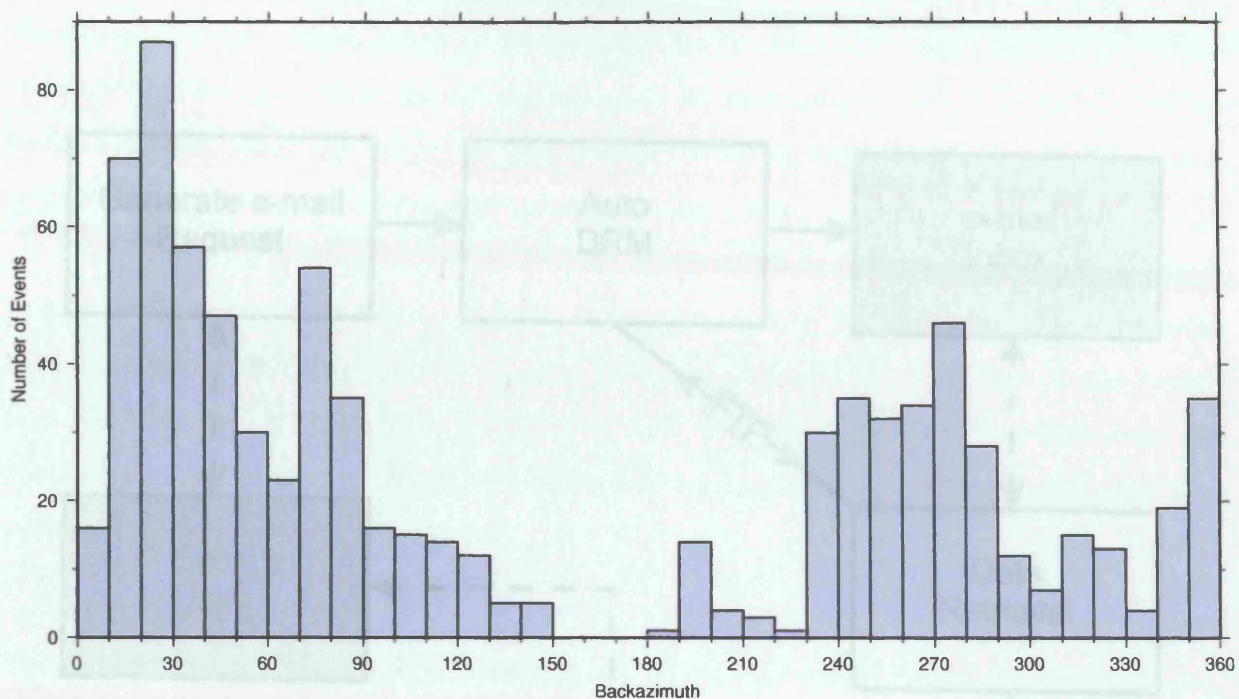


Figure 4.5: A histogram of the range of backazimuths for a list of events of $M_b \geq 6.0$ at epicentral distances between 30° and 100° from 1990 to 2001.

```

TIME START_DATE START_TIME END_DATE END_TIME
STA_LIST EDI, CWF ...
CHAN_LIST SH* or BH*
FTP jpto@bgs.ac.uk
STOP

```

To generate bulk requests and retrieve data from the Auto DRM systems several simple scripts have been written. Firstly, the event list is searched and an e-mail request is generated for each event and sent to the relevant data manager. Once the data has been located an e-mail is sent back with instructions of where to retrieve the data via FTP. The e-mail in-box is subsequently searched and the FTP instructions are used to automatically download the event files. The flow of procedures used in the event data retrieval is shown in Figure 4.6.

Once the data has been retrieved, the format in which it is delivered must be considered. The receiver function analysis programs use the Seismic Analysis Code (SAC) format, but the Auto DRM delivers the data in GSE2.0 format, which is the format of the International Seismic Monitoring Stations (ISMS). The GSE files contain all of the data from all of the stations for

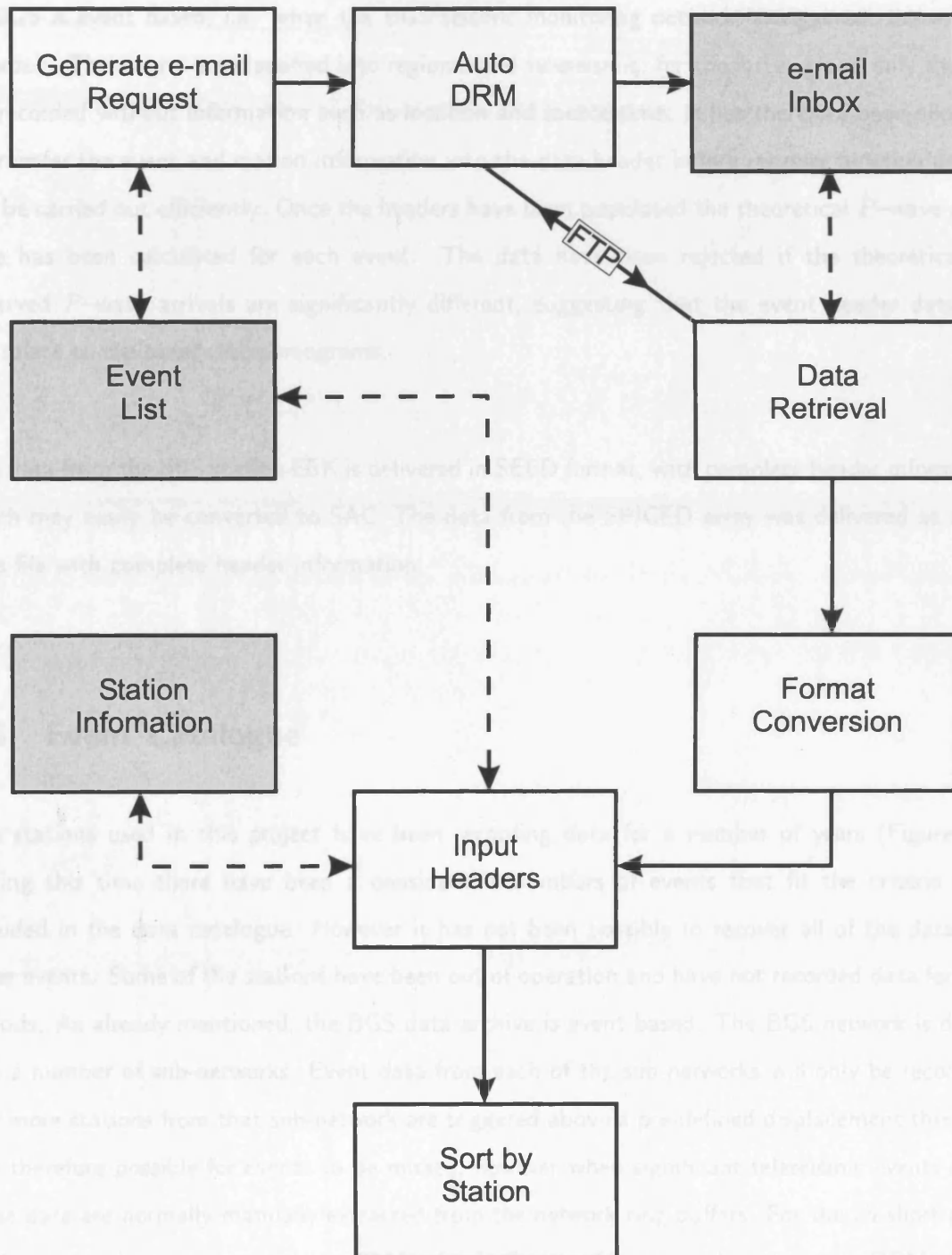


Figure 4.6: A flow chart of the stages involved in the data acquisition from the BGS and Blacknest data archives.

the requested event. When converted into SAC format the data have been stored in a directory structure that contains all of the event information for one station. The data that is archived by BGS is event based, i.e. when the BGS seismic monitoring network is triggered, the event is recorded. The events are classified into regional and teleseismic; for the latter group only the data are recorded without information such as location and source time. It has therefore been necessary to transfer the event and station information into the data header before receiver function analysis can be carried out efficiently. Once the headers have been populated the theoretical P -wave arrival time has been calculated for each event. The data have been rejected if the theoretical and observed P -wave arrivals are significantly different, suggesting that the event header data does not relate to the observed seismograms.

The data from the IRIS station ESK is delivered in SEED format, with complete header information, which may easily be converted to SAC. The data from the SPICED array was delivered as a SAC data file with complete header information.

4.5 Event Catalogue

The stations used in this project have been recording data for a number of years (Figure 4.7). During this time there have been a considerable numbers of events that fit the criteria to be included in the data catalogue. However it has not been possible to recover all of the data from these events. Some of the stations have been out of operation and have not recorded data for some periods. As already mentioned, the BGS data archive is event based. The BGS network is divided into a number of sub-networks. Event data from each of the sub-networks will only be recorded if 2 or more stations from that sub-network are triggered above a pre-defined displacement threshold. It is therefore possible for events to be missed, however when significant teleseismic events occur, those data are normally manually extracted from the network ring buffers. For the 25 short-period BGS stations there were a total of ~ 7500 teleseismic records recovered via the auto DRM system. During the test of the Auto DRM system, it was found that little data was recovered from before 1990. This was the reason that the event list was truncated at 1990. The exception to this was the data from ESK, where some data were available from before 1990. In this case an expanded

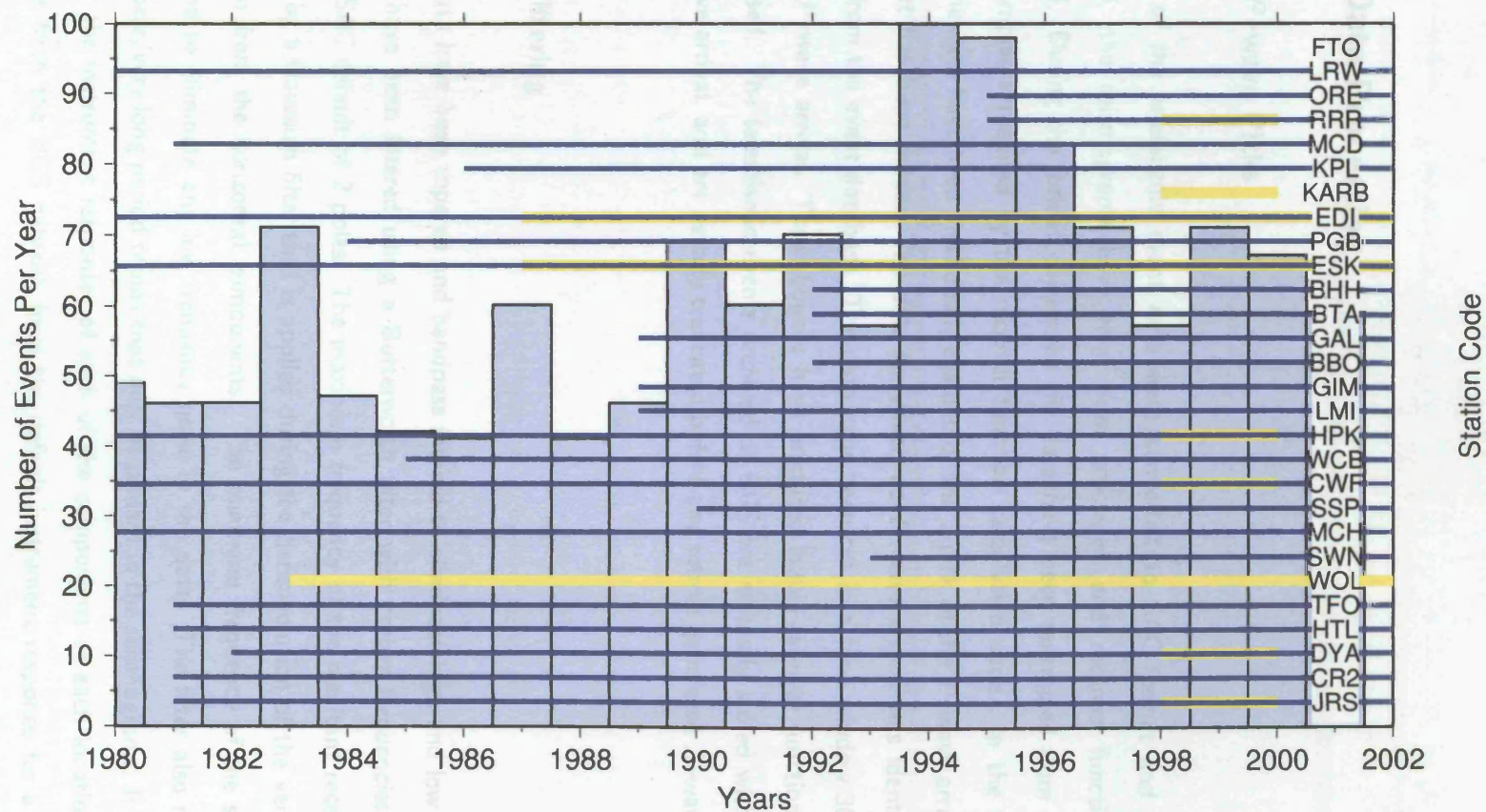


Figure 4.7: A graph to show the duration for which all instruments used in the present receiver function study have been operational (Blue = short-period, Yellow = broadband). The plot is overlaid on a histogram of the number of events of $M_b \geq 6.0$ per year.

event list was used to request data.

4.6 Data Processing

4.6.1 *P*-wave Picks

Once all of the teleseismic event data were converted to SAC format and the headers fully populated, the seismograms have then been processed and receiver functions subsequently calculated. During the format conversion the data have been resampled from 100 samples/s to the 10 samples/s required by the receiver function calculation code. In the first stage of the analysis the data have been individually evaluated, and a pick of the *P*-wave arrival in the vertical component has been made. The data for which no *P*-wave arrival was identifiable have been removed from the event database. The data have been cut in a time window 30s before and 90s after the *P*-wave arrival. These bounds have produced stable receiver functions from all of the stations used. The teleseismic events archived at BGS are normally stored with only 10s before the *P*-wave arrival, and are usually truncated before any source generated *S*-waves arrive.

4.6.2 Filtering

The cut data have been tapered and bandpass filtered to eliminate high and low frequency noise. The data have been filtered using a Butterworth filter with corner frequencies at 0.1s and 4s using the SAC default of 2 poles. The maximum frequency of the resultant receiver functions is controlled by a Gaussian filter that is applied during the deconvolution of the vertical component seismogram from the horizontal components. The minimum frequency of the seismograms has been limited to eliminate any low frequency noise in the data. The filter also removes any DC offset or linear/very long period trends that may be present in the seismograms. It is also necessary to consider the instrument responses of each of the components at each location. However the instruments from the BGS network have the default instrument response for a Wilmore MK-II seismometer, and therefore removing these instrument responses will produce no change in the relative amplitude of the orthogonal component seismograms. Removing the instrument response

from the short-period seismograms will minimise the negative lobes caused by limited bandwidth deconvolution that ring the major phases of the receiver functions. In studies aimed at imaging mantle structures with receiver function analysis, the instrument response of short-period stations have commonly been removed to enhance the long-periods signals needed to image such deep structures (e.g. Yuan et al., 1997). The short-period BGS data have been recorded using 16-bit digitisers for which the dynamic range is limited, and when the instrument response has been removed the longer-period signal is normally very noisy. Furthermore, due to the short window of data before the P -wave arrival, the end effects associated with deconvolution performed when removing the instrument response, degrade the teleseismic signal. Therefore the instrument response has not been removed from the short-period BGS seismograms.

4.6.3 Receiver Function Calculation

The calculation of the receiver functions is simple, and is performed using the `pwaveqn` code of Ammon et al. (1990). A typical teleseismic event used to calculate receiver functions recorded on a short-period seismometer is shown in Figure 4.8a. The resultant receiver functions from `pwaveqn` are controlled by 3 parameters; 1) c , the deconvolution water level parameter, 2) a , the Gaussian filter parameter, and 3) the length of data window. The frequency domain deconvolution procedure is fully discussed in Section 3.2.2. The effect that variation in deconvolution parameters have on the resultant receiver functions from KPL has been demonstrated in Figures 4.8b and 4.9. To calculate the receiver functions presented in these figures a has been varied between 2 and 3, which approximately represent low-pass filters with maximum frequencies of 1Hz and 1.5Hz respectively. c has been varied between 0.1 and 0.0001. The data have been cut to three time windows, Cut 1 contains 90s of data after the P -wave arrival, Cut 2 contains 60s and Cut 3 contains 30s. The data have all been cut 30s in front of the P -wave arrival but many of the short-period seismograms were archived with a shorter time window before the P -wave arrival. In Figure 4.9 all of the receiver functions from KPL are stacked regardless of backazimuth or epicentral distance. If the underlying geology departs from the 1D case then stacking all of the receiver functions from one station will result in interference between the receiver function phases. For the purposes of demonstrating the variation in stacked receiver function with deconvolution parameter it is not necessary to account for this problem. From Figures 4.8b and 4.9 it is possible to note that;

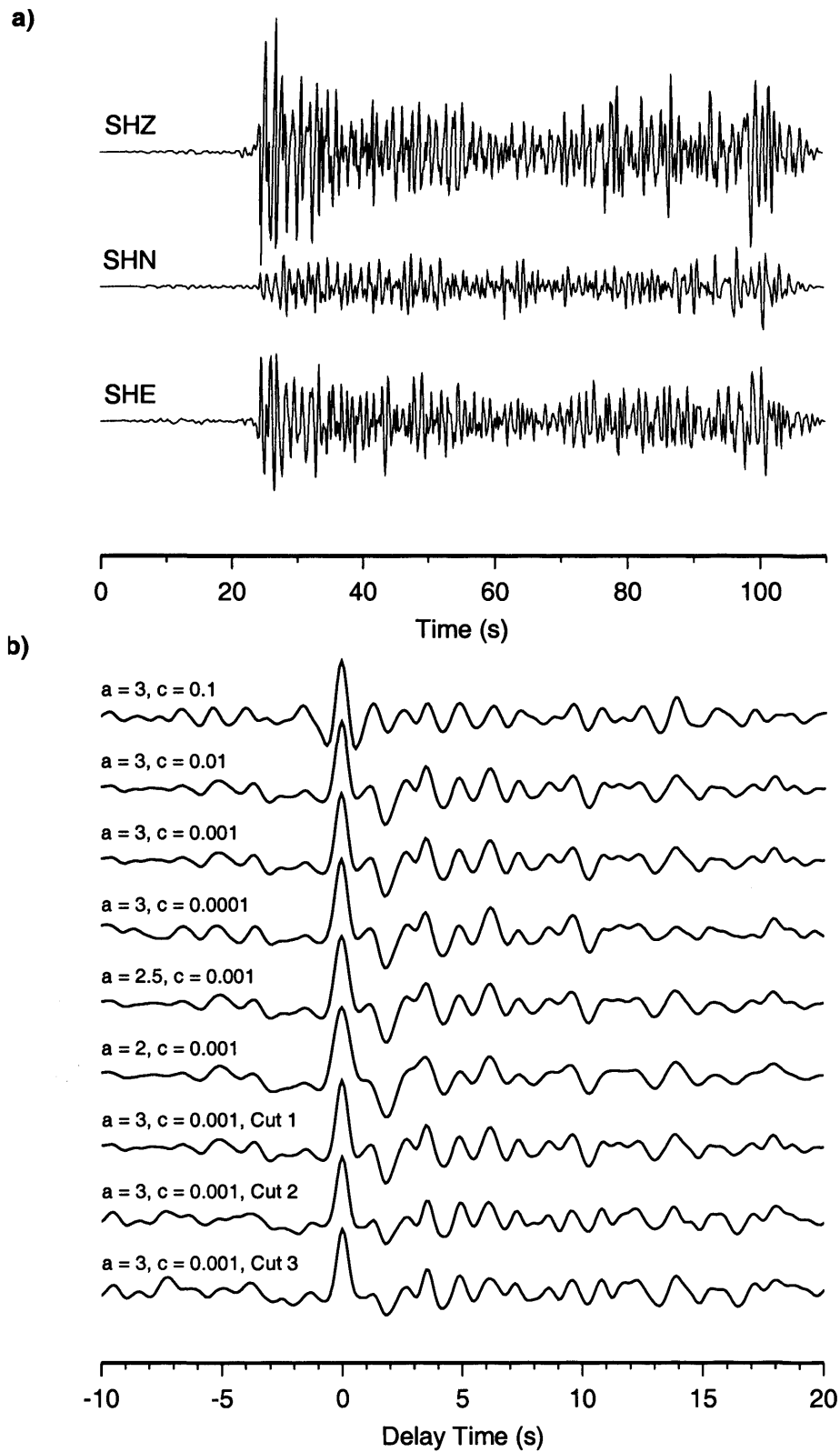


Figure 4.8: a) The starting three component short-period seismograms used to calculate receiver functions in b), the magnitude 7.0 event occurred on in the Hindu Kush region of Afghanistan on 9th August 1993. b) A series of radial receiver functions from the event in a) recorded at station KPL calculated with varying a , c and cut window length. Cut 1 contains 90s of data after the P -wave arrival, Cut 2 contains 60s and Cut 3 contains 30s.

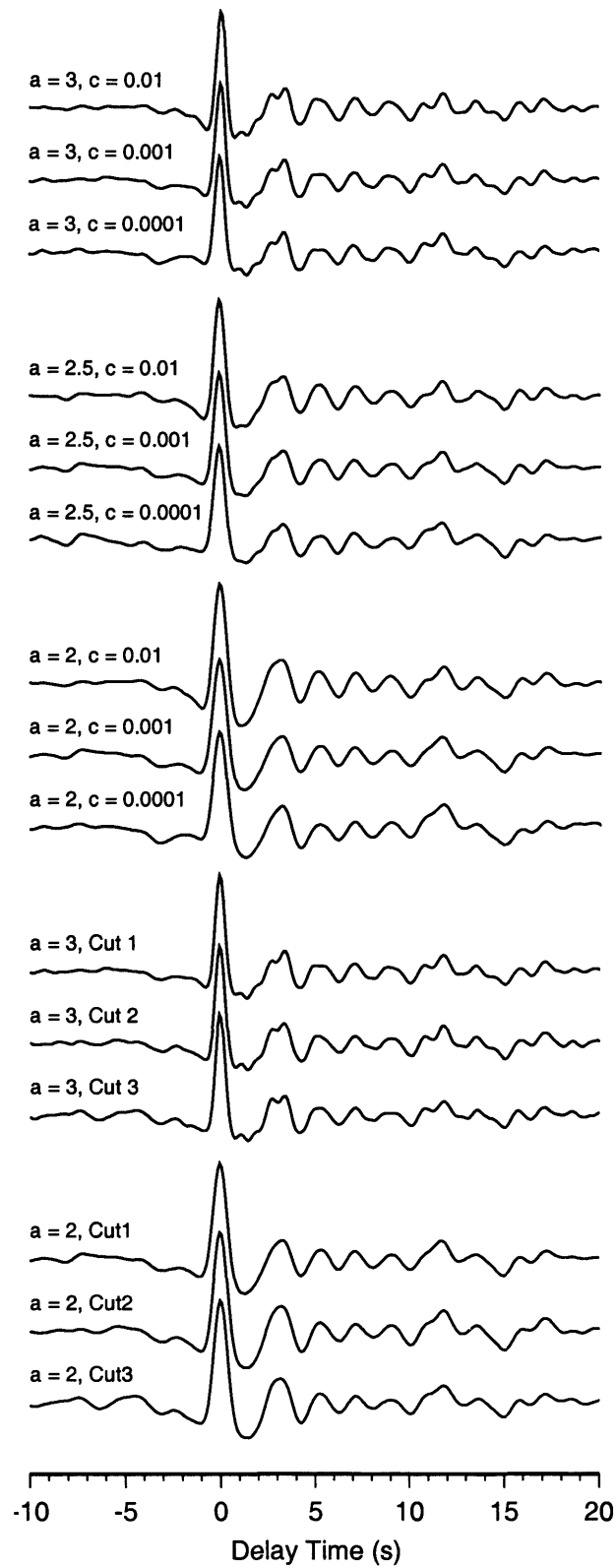


Figure 4.9: A series of stacked receiver functions from station KPL. The stacks include all events recorded at KPL and are calculated with varying a , c and cut window length. Cut 1 contains 90s of data after the P -wave arrival, Cut 2 contains 60s and Cut 3 contains 30s.

1. The maximum frequency of the resultant receiver functions is reduced by reducing a , the Gaussian filter parameter. It is clear that as the frequency is reduced the Moho P_s and $PpPs$ phases become better resolved. However there is a trade-off between the maximum frequency of the resultant receiver functions and the resolution to which the data may be interpreted. As the maximum frequency of the Gaussian filter is lowered there is a reduction in the detail of the receiver function phases.
2. Increasing c above 0.001 increases the noise before the direct P -wave arrival in the receiver functions. Furthermore, the negative lobes that ring the direct P -wave arrival are increased with the increase in this water level parameter.
3. Decreasing the time window of data used to calculate the receiver functions increases the noise level before the direct P -wave arrival. This increase in noise is also particularly noticeable in the reduced resolution of the Moho $PpPs$ at ~ 11.5 s.

4.7 Receiver Function Catalogue

The receiver functions investigated in Chapters 5 and 6 have been calculated using a Gaussian filter parameter of $a=3$ and a water level parameter of $c=0.001$, giving the resultant receiver functions a maximum frequency of approximately 1.5Hz. The teleseismic event data have been cut to 30s before and 90s after the P -wave arrival, where the data was available. These values have been shown to produce stable receiver functions. The higher Gaussian filter parameter of $a = 3$ has been used to allow the resultant receiver functions to contain more higher frequency information, and to therefore improve the resolution of the subsequent modelling studies in Chapters 5 and 6. The raw radial receiver functions calculated for all of the 34 instruments from the British Isles are shown in Appendix A (Figures A.1-A.66). From the ~ 7500 teleseismic event records recovered from the various archives, a total of ~ 1500 good receiver functions have been analysed. Initially the raw seismograms were visually checked and P -wave arrivals were picked. Many of the seismograms were rejected at this stage for a number of reasons including faulty instrumentation and high levels of noise. The observed P -wave arrival times of the data were checked with the corresponding theoretical values. Finally the receiver function data shown in Appendix A were visually checked and the poor quality receiver functions removed from the receiver function catalogue.

Chapter 5

H– κ Stacking Analysis

5.1 Introduction

The H – κ stacking method has been applied to three-component data from 34 seismic stations operating in the British Isles. These analyses have resulted in estimates of the crustal thickness (H) and V_p/V_s ratio (κ) beneath each of the stations. An *a priori* average crustal velocity (V_p) has been used to calculate the stacks. The sensitivity of the H – κ results to stacking V_p is discussed and the rationale behind the chosen input value described. Of the 34 seismometers used, 12 are at sites which have co-located broadband and short period instruments, giving a total of 28 unique measurements of the crustal thickness and average crustal V_p/V_s . These data sample all of the inferred geological terranes of the British Isles (Bluck et al., 1992). The results show considerable variation in both H and κ , ranging between 25–38 km and 1.6–1.85 respectively. The results are described and discussed with respect to the geographical regions outlined in Chapter 2.

5.2 Method

5.2.1 H – κ Stacking

The H – κ stacking technique has been fully described in Section 3.3.7. Briefly, the method exploits the differing move-out curves of the Ps , $PpPs$ and $PpSs$ phases of a receiver function in the H –

κ domain to provide a unique solution for crustal thickness and V_p/V_s ratio. The method applies a very simple grid search forward modelling routine. For a range of one layer crustal velocity models with variable H and κ , the arrival times of Ps , $PpPs$ and $PpSs$ are calculated. The amplitudes of the receiver functions at the theoretical arrival times of the three phases are added to the stack $S(H, \kappa)$;

$$S(H, \kappa) = \sum_{j=1-n} w_1 r_j(t_1) + w_2 r_j(t_2) - w_3 r_j(t_3) \quad (5.1)$$

where r_j is a radial receiver function, t_1 , t_2 and t_3 are the predicted Ps , $PpPs$ and $PpSs$ times, and w_1 , w_2 and w_3 are weighting functions of the stack. The standard deviation of the maximum value of the H - κ stack can be estimated;

$$\sigma_H^2 = 2\sigma_s / \frac{\partial^2 S}{\partial H^2} \quad (5.2)$$

$$\sigma_\kappa^2 = 2\sigma_s / \frac{\partial^2 S}{\partial \kappa^2} \quad (5.3)$$

where σ_s is the estimated variance of the value of $S(H_{max}, \kappa_{max})$ (the point giving the solution for H and κ) from the H - κ stacks of all of the individual receiver functions. $\partial^2 S / \partial H^2$ and $\partial^2 S / \partial \kappa^2$ are the second derivatives of $S(H, \kappa)$ along the planes of constant H and κ that intersect at $S(H_{max}, \kappa_{max})$.

5.2.2 Sensitivity to Stacking Velocity

The primary sensitivity of the receiver function method is to velocity contrast and relative arrival times of the converted and multiple phases, and not to the absolute velocity (Ammon et al., 1990). The V_p/V_s ratio derived from H - κ stacking is relatively insensitive to change in the input V_p . However, there is a considerable velocity-depth trade-off in the measurement. This trade-off is show in Figure 5.1. In this example three synthetic receiver functions of variable κ have been analysed by H - κ stacking using increasing input V_p . The true solution for the synthetic receiver functions is $H = 30$ km and $V_p = 6.3$ km s⁻¹ with the three values of κ 1.65, 1.73 and 1.84 respectively. This diagram also demonstrates the relatively low sensitivity of the stack to variation of V_p in the κ domain. The results show that for these particular earth models (of a similar thickness to the crust of the British Isles) there can be an approximately 1 km variation in H for a

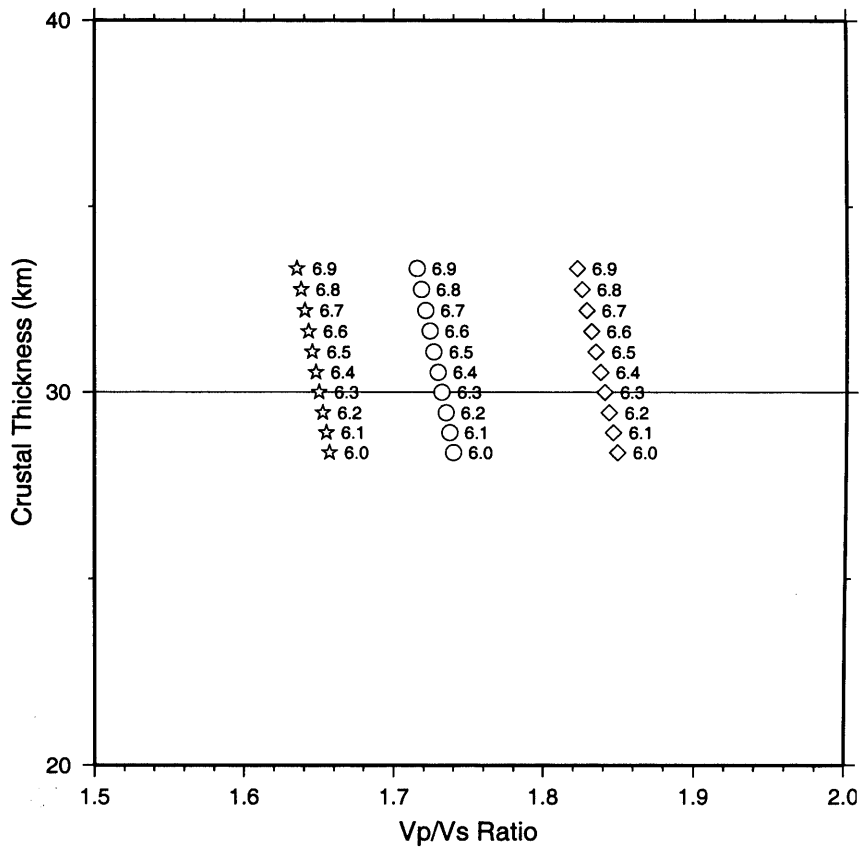


Figure 5.1: The maximum points of a range of H - κ stacks using different input V_p for three synthetic receiver functions derived from a crust with $H = 30$ km, $V_p = 6.3$ km s⁻¹ and κ of 1.65(star), 1.73(circle) and 1.84(diamond). The maximum points are labelled with the V_p used to calculate the stack. This example shows the dependence of the H - κ stacking method upon the input V_p used.

0.2 km s⁻¹ change in V_p . For a thicker crust the change in depth with velocity is greater, because the calculated depth is dependent upon the travel times of the Ps , $PpPs$ and $PpSs$ phases through the crust. For example for a 60 km thick crust a 0.1 km s⁻¹ change in V_p results in a 1 km change in H . It is therefore important where possible, to constrain the input stacking V_p with *a priori* velocity information.

5.3 Data & Results

5.3.1 Data Analysis

H - κ stacking analysis has been applied to all of the 34 broadband and short period three-component instruments deployed throughout the British Isles. A total of 1493 receiver functions have been stacked for the 34 instruments. The majority of these data have been generated using a water level parameter (c) of 0.001 and a Gaussian filter (a) of 3. In some cases the Gaussian filter parameter has been lowered to $a=2$ to stabilise the maximum point of the stack. Using the lower frequency low-pass filter gives resultant receiver functions with a better resolved $PpPs$ phase (see Section 4.6.3), leading to a more stable solution for the H - κ stack. The trade-off in using the lower frequency data is that the standard deviation of the maximum point of the H - κ stack is increased. The stacking Ps , $PpPs$ and $PpSs$ times have been calculated for a range of models which increment every 0.1 km in the H domain and every 0.005 in the κ domain. The receiver functions used in the H - κ stack have been re-sampled from the original sample rate of 0.1s to a sample interval of 0.01s. This was done to stabilise the standard deviation calculations by minimising any sawtooth/step effects in the data. However, because the original raw seismograms were recorded at 0.01s and were subsequently reformatted to 0.1s for the receiver function calculation, this re-sampling to 0.01s does mean that some information from the original seismograms has been lost. The amplitudes of the H - κ stacks are normalised relative to the amplitude at their maximum value, which defines the measured H and κ , and the normalised stack values above 0.6 are plotted as shaded contours (Figures 5.2-5.7).

The results of the H - κ analyses have been compiled in Table 5.1. This table shows the results of two separate analyses. In the first a constant average crustal P -wave velocity of 6.3 km s⁻¹ has

been used, whereas in the second the P -wave velocities used have been derived from the results of crustal seismic refraction profiles. These two distinct analyses have been performed to study the sensitivity of the H - κ stacks from the British Isles to the input V_p . The crustal velocity is primarily constrained by the LISPB (Bamford et al., 1978; Barton, 1992), CSSP (Bott et al., 1985) and JUNE92 (Jones et al., 1996; Morgan et al., 2000) deep seismic refraction profiles. The structures and range of velocities found in these studies are discussed in Section 2.3.2. Analysis 1 gives a non-subjective investigation of the receiver function data by using a constant V_p of 6.3 km s^{-1} as the input H - κ stacking P -wave velocity. This represents a mean average velocity of the crust of the British Isles calculated from the primary seismic refraction investigations. This figure is slightly higher than that of the average crustal velocity for extended continental crust of 6.2 km s^{-1} reported by Christensen and Mooney (1995). Their model does however include a layer of recent lower velocity sedimentary rocks that do not exist beneath many of the stations investigated in this study. In Analysis 2, when a station samples the crust close to a seismic refraction profile, the input P -wave velocity is simply the mean crustal velocity at that point along the profile. However, many of the seismic recording stations do not lie directly upon any seismic refraction profile; in this case the V_p has either been left at 6.3 km s^{-1} , or if this is not an appropriate value then a regional mean velocity has been used. In most areas the seismic velocity remains close to $6.3 \pm 0.1 \text{ km s}^{-1}$, with the exception of the Midland Valley where the average increases to $>6.5 \text{ km s}^{-1}$. This is caused by a high velocity gradient in the upper crust, where the P -wave velocity reaches a value of 6.5 km s^{-1} at about 15 km depth (Barton, 1992).

The resultant stacks derived from Analysis 1 are shown in Figures 5.2-5.7. The receiver functions used to calculate the stack are presented in Appendix A, Figures A.1 to A.66. Firstly the data are presented as the raw radial receiver functions. The data for each station have also been summed in 20° backazimuth bin windows and are presented together with the a plot of each H - κ stack. The theoretical Ps and $PpPs$ phase times for the maximum point of the H - κ stack are labelled on both the stacked radial and tangential receiver functions, and the move-out curves of these phases are overlain on the H - κ plots. For clarification, the stacked radial receiver functions presented in Appendix A will always be referred to as stacked receiver functions. The results of the receiver function H - κ analysis will be referred to as H - κ stacks.

The standard deviations of the solutions derived from the H - κ stacks are shown in Table 5.1.

These values are estimated during the stacking procedure and only reflect the measured standard deviation of the maximum point of the stack, and do not include any estimate of other errors that may occur; for example the use of the inappropriate stacking V_p . Furthermore, this does not include any qualitative information about the input receiver functions, for example as to whether the phases that produce the maximum stacking point are in fact the Moho Ps , $PpPs$ and $PpSs$ phases. The errors introduced by the use of an incorrect stacking velocity have been investigated by performing Analyses 1 and 2 using both constant and variable V_p on the data from each station. In order to provide a qualitative description of the results each $H-\kappa$ stack has been categorised as either;

1. A high quality stack where both the Ps and $PpPs$ phases can be identified in the stacked receiver functions.
2. An intermediate quality stack where Ps is still strong, but $PpPs$ is less well defined, and results in a stringing effect in the stack meaning the maximum point of the stack, must be inferred rather than being clearly identifiable.
3. A poor quality stack with multiple highs that provide no stable or reasonable solution in the $H-\kappa$ domain.

As an example the $H-\kappa$ stack from BBO (Figure 5.4) has been classified as high quality (1); the Ps and $PpPs$ phases from the stack can be clearly identified in the stacked receiver functions (Figure A.31). The $H-\kappa$ stack from CWF (Figure 5.5) has been classified as an intermediate quality (2) stack. In the stacked receiver functions (Figure A.43) the Ps phase is clearly identifiable but any $PpPs$ phases are of much lower amplitude, bringing ambiguity to location of the maximum point of the $H-\kappa$ stack. The stack for SWN (Figure 5.6) has been classified as quality 3. The stacked receiver functions are dominated by many high amplitude phases that cannot be clearly identified (Figure A.51). This classification scheme is similar to that used by Cheverot and van der Hilst (2000) from which they provide an estimate of the quality of the $H-\kappa$ stacks from their study.

	n	Analysis 1						Analysis 2							Q
		V_p	H	σ_H	κ	σ_κ	σ	V_p	$Ref.$	H	σ_H	κ	σ_κ	σ	
LRW	52	6.30	32.8	1.88	1.66	0.10	0.215	6.30	x	33.0	1.97	1.65	0.10	0.210	2
ORE	41	6.30	26.1	1.20	1.74	0.07	0.253	6.37	ab	26.4	1.28	1.74	0.06	0.253	1
RRR	48	6.30	24.3	1.08	1.75	0.07	0.258	6.37	b	24.6	1.17	1.75	0.07	0.258	1
RRRB	28	6.30	24.2	0.94	1.76	0.07	0.264	6.37	b	24.6	1.04	1.76	0.07	0.262	1
MCD	59	6.30	31.2	1.45	1.76	0.07	0.262	6.24	b	30.7	1.46	1.76	0.06	0.264	1
KARB	11	6.30	28.3	0.84	1.83	0.06	0.289	6.37	b	28.7	0.92	1.82	0.06	0.285	2
KPL	67	6.30	28.0	0.96	1.75	0.06	0.256	6.37	b	28.3	0.98	1.75	0.06	0.256	1
EDI	73	6.30	32.6	1.39	1.71	0.07	0.238	6.54	b	34.0	1.58	1.70	0.07	0.235	1
EDIB	27	6.30	34.3	1.84	1.68	0.09	0.226	6.54	b	36.3	2.00	1.67	0.08	0.218	2
PGB	23	6.30	30.3	1.86	1.77	0.11	0.266	6.54	b	26.2	1.92	1.88	0.13	0.303	2
ESK	39	6.30	29.5	0.80	1.81	0.05	0.282	6.35	b	30.2	0.92	1.79	0.05	0.275	1
ESKB	134	6.30	30.8	1.53	1.77	0.09	0.268	6.35	b	31.0	1.63	1.77	0.08	0.268	1
BTB	53	6.30	29.3	1.16	1.75	0.07	0.256	6.25	b	29.1	1.23	1.75	0.07	0.256	1
BHH	26	6.30	27.3	1.33	1.81	0.09	0.280	6.25	b	27.0	1.35	1.81	0.08	0.282	1
GAL	48	6.30	31.6	2.71	1.60	0.12	0.179	6.25	b	29.0	2.82	1.65	0.15	0.210	2
BBO	85	6.30	28.9	0.80	1.72	0.05	0.242	6.25	b	28.6	0.84	1.72	0.05	0.245	1
GIM	49	6.30	29.8	1.23	1.69	0.06	0.228	6.25	b	29.6	1.15	1.68	0.05	0.226	1
LMI	60	6.30	27.9	1.20	1.75	0.07	0.260	6.25	b	27.6	1.23	1.75	0.07	0.260	1
HPK	35	6.30	32.0	1.09	1.78	0.06	0.271	6.25	b	31.8	1.17	1.78	0.06	0.271	1
HPKB	20	6.30	31.7	0.91	1.78	0.05	0.271	6.25	b	31.4	0.94	1.78	0.05	0.271	1
WCB	37	6.30	27.6	1.40	1.85	0.09	0.295	6.36	c	27.7	1.45	1.85	0.08	0.295	2
CWF	66	6.30	35.8	1.71	1.71	0.07	0.240	6.36	c	36.2	1.83	1.71	0.07	0.240	2
CWFB	25	6.30	36.1	1.66	1.71	0.07	0.240	6.36	c	36.5	1.74	1.71	0.07	0.240	2
MCH	42	6.30	35.3	1.99	1.77	0.08	0.268	6.36	c	35.9	2.00	1.76	0.07	0.264	2
SSP	60	6.30	36.0	1.67	1.75	0.09	0.256	6.36	c	36.4	1.80	1.75	0.08	0.256	2
SWN	31	6.30	48.8	1.48	1.72	0.05	0.245	6.24	c	48.3	1.43	1.72	0.05	0.245	3
WOB	53	6.30	38.9	1.99	1.79	0.07	0.273	6.24	c	39.0	2.03	1.78	0.06	0.269	2
TFO	18	6.30	34.1	1.84	1.61	0.08	0.186	6.30	x	34.1	1.96	1.61	0.08	0.186	3
HTL	42	6.30	31.5	1.37	1.66	0.08	0.212	6.30	x	31.3	1.47	1.67	0.09	0.218	2
CR2	55	6.30	28.1	0.73	1.67	0.05	0.218	6.30	x	28.1	0.75	1.67	0.04	0.218	1
DYAB	8	6.30	28.4	0.76	1.79	0.05	0.275	6.30	x	28.5	0.85	1.79	0.05	0.273	1
DYA	29	6.30	28.4	0.81	1.80	0.06	0.277	6.30	x	28.4	0.85	1.79	0.06	0.275	1
JRS	34	6.30	32.3	1.60	1.74	0.07	0.251	6.30	x	32.3	1.60	1.74	0.07	0.251	2
JSAB	15	6.30	31.3	1.04	1.76	0.05	0.264	6.30	x	31.3	1.09	1.76	0.05	0.264	1

Table 5.1: A table of the H - κ stacking results for all 34 instruments from the British Isles, each stack includes n receiver functions. Listed for both analyses are V_p (km s⁻¹), the crustal thickness H (km), V_p/V_s ratio κ , and their respective standard deviations defined by Zhu and Kanamori (2000) σ_H and σ_κ . Poisson's ratio σ is also presented. Analysis 1 uses a constant V_p whereas Analysis 2 uses V_p derived from crustal seismic refraction studies. The source publication of the V_p used in Analysis 2 is listed; a =Jones et al. (2002), b =Barton (1992) and c =Edwards and Blundell (1984). Finally Q describes the quality of the resultant stack.

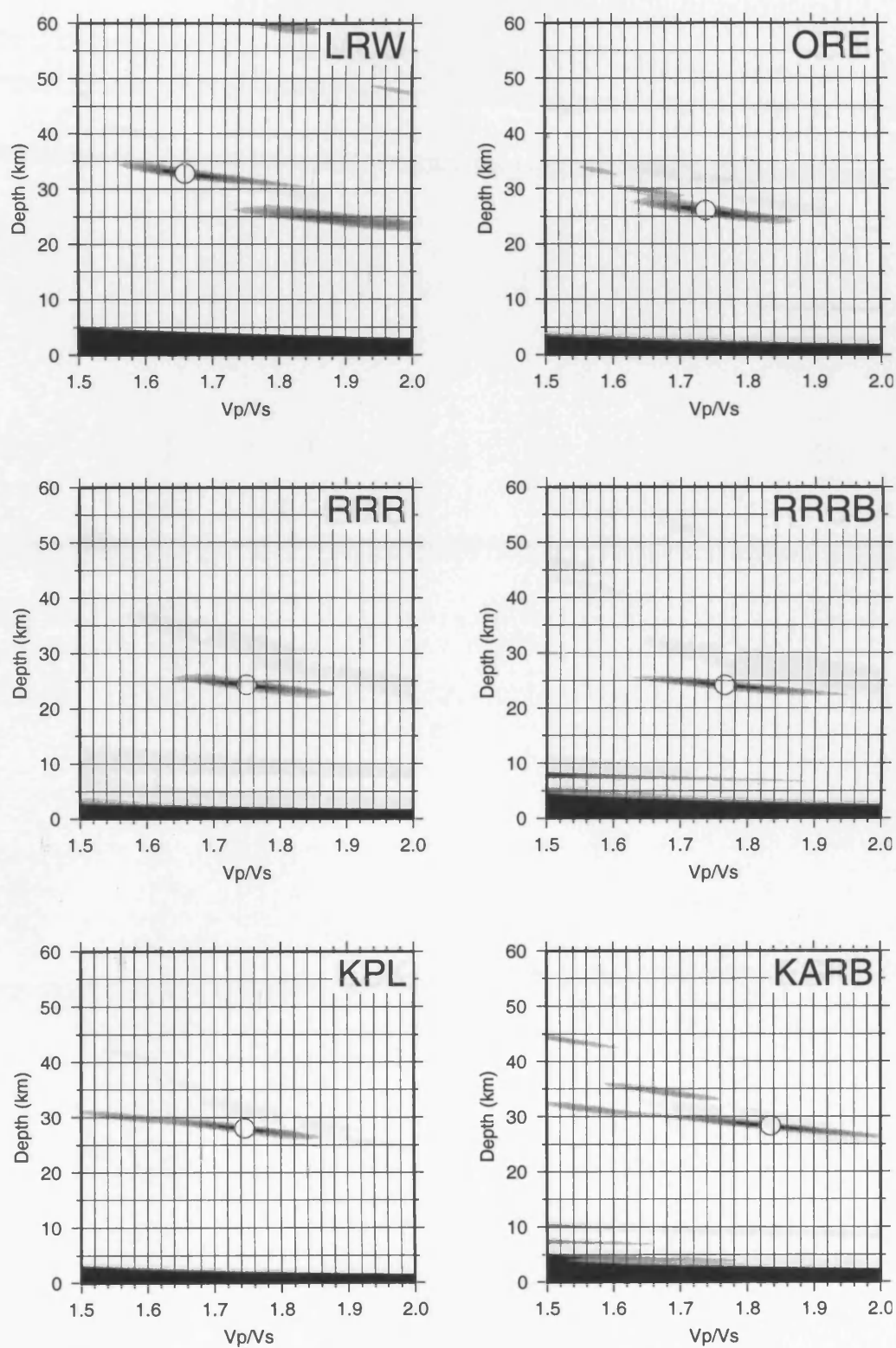


Figure 5.2: $H-\kappa$ stacks for stations LRW, ORE, RRR, RRRB, KARB and KPL.

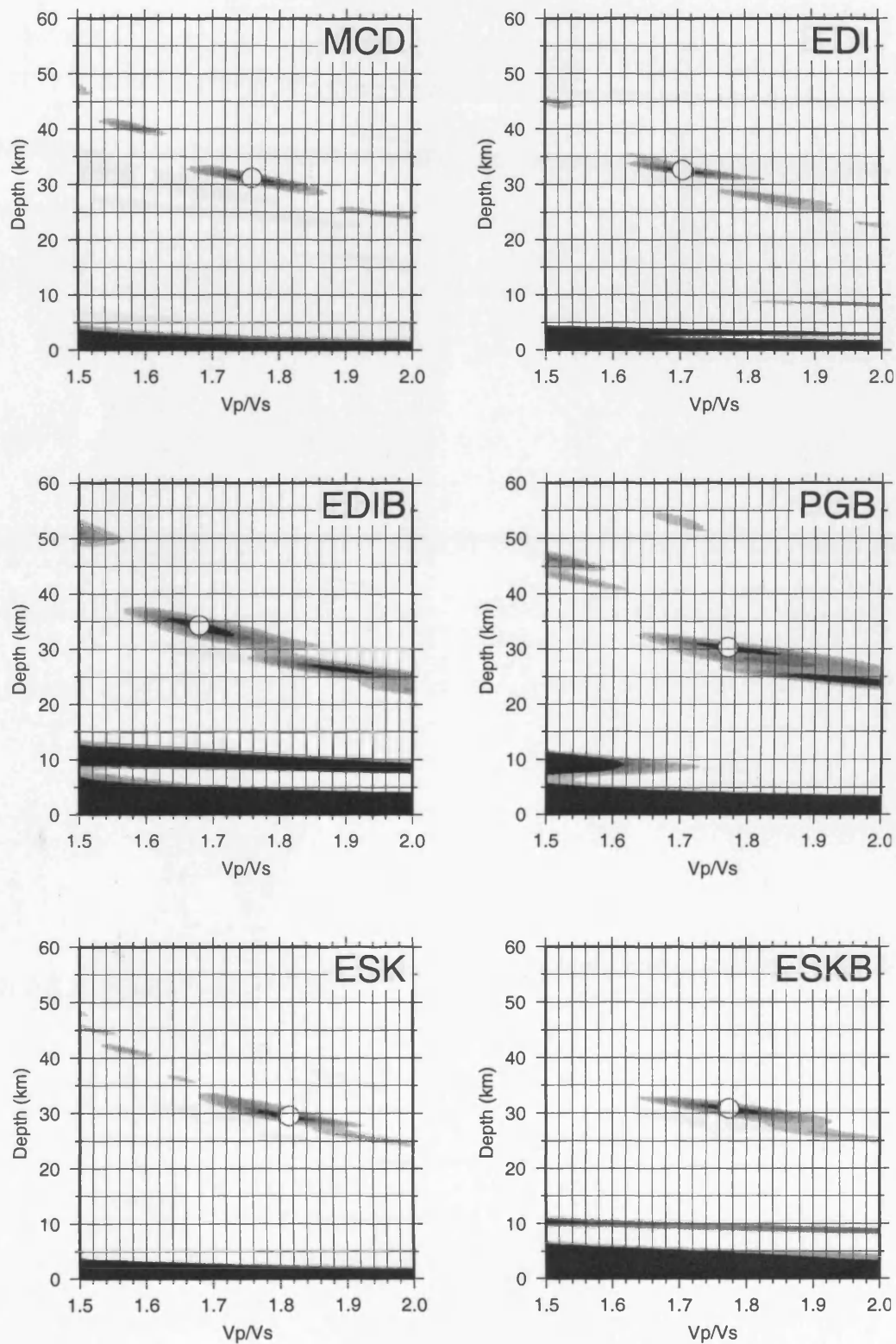


Figure 5.3: $H-\kappa$ stacks for stations MCD, EDI, EDIB, PGB, ESK and ESKB.

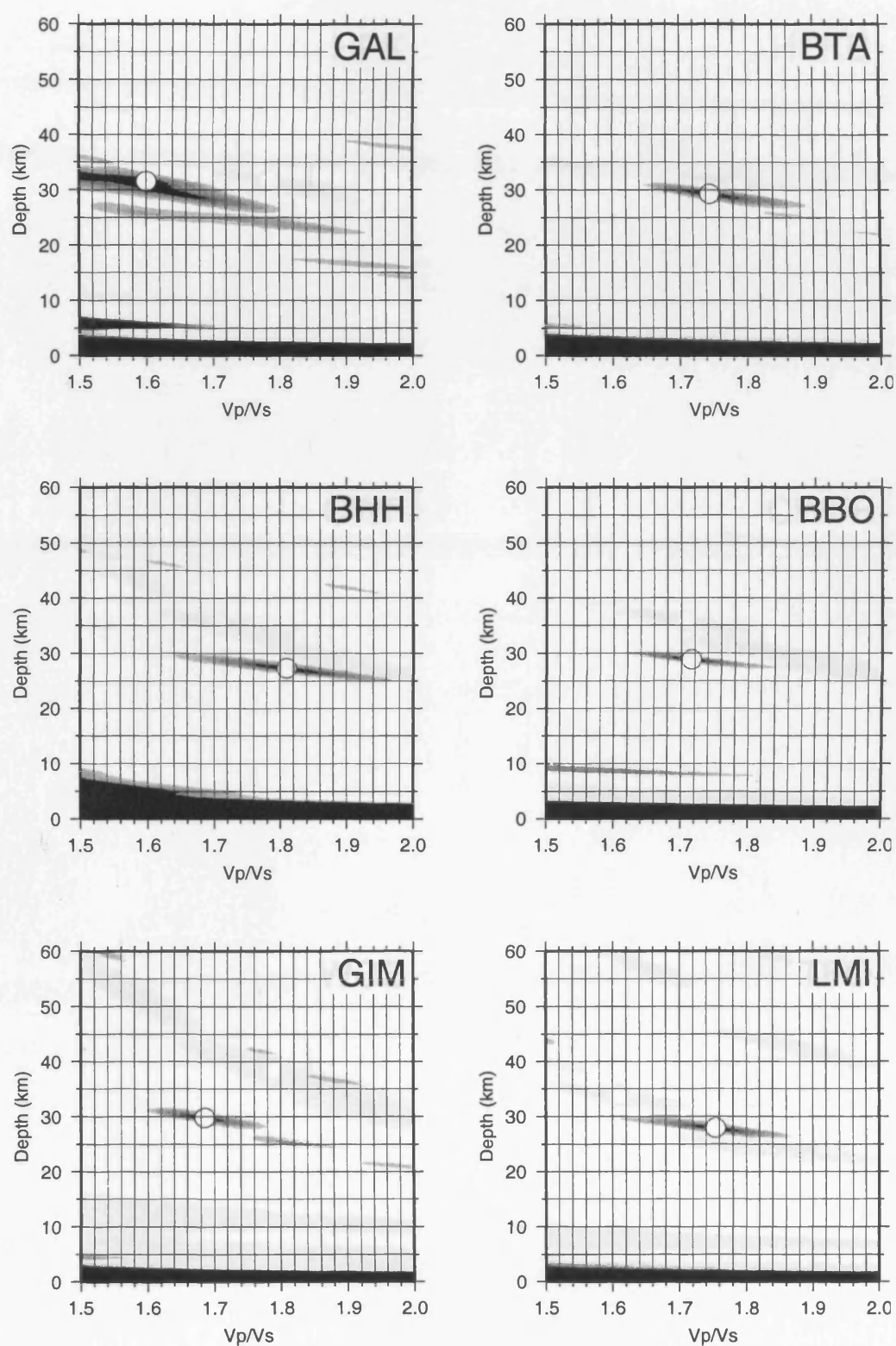


Figure 5.4: $H-\kappa$ stacks for stations GAL, BTA, BHH, BBO, GIM and LMI.

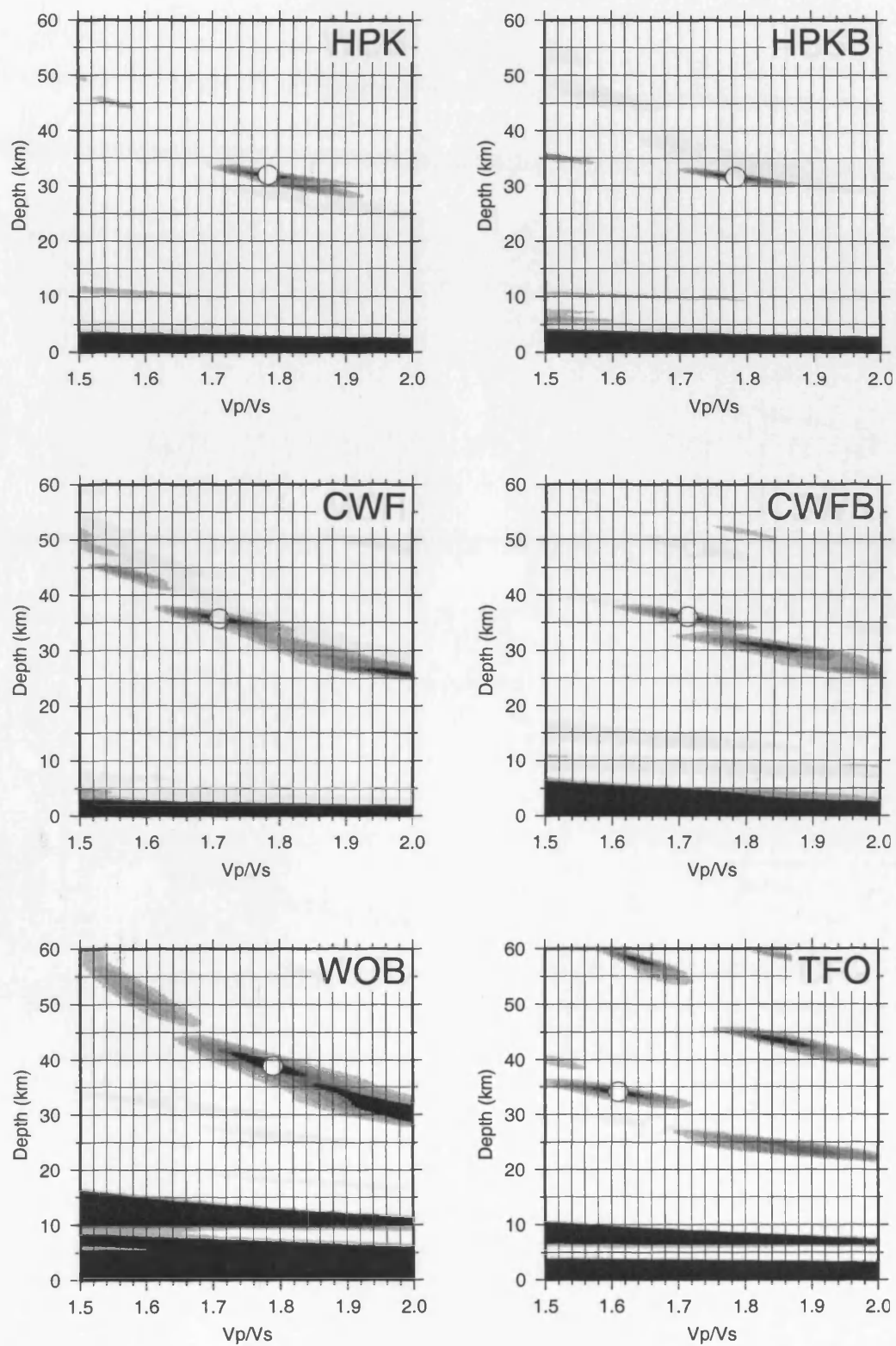


Figure 5.5: H - κ stacks for stations HPK, HPKB, CWF, CWFB, WOB and TFO.

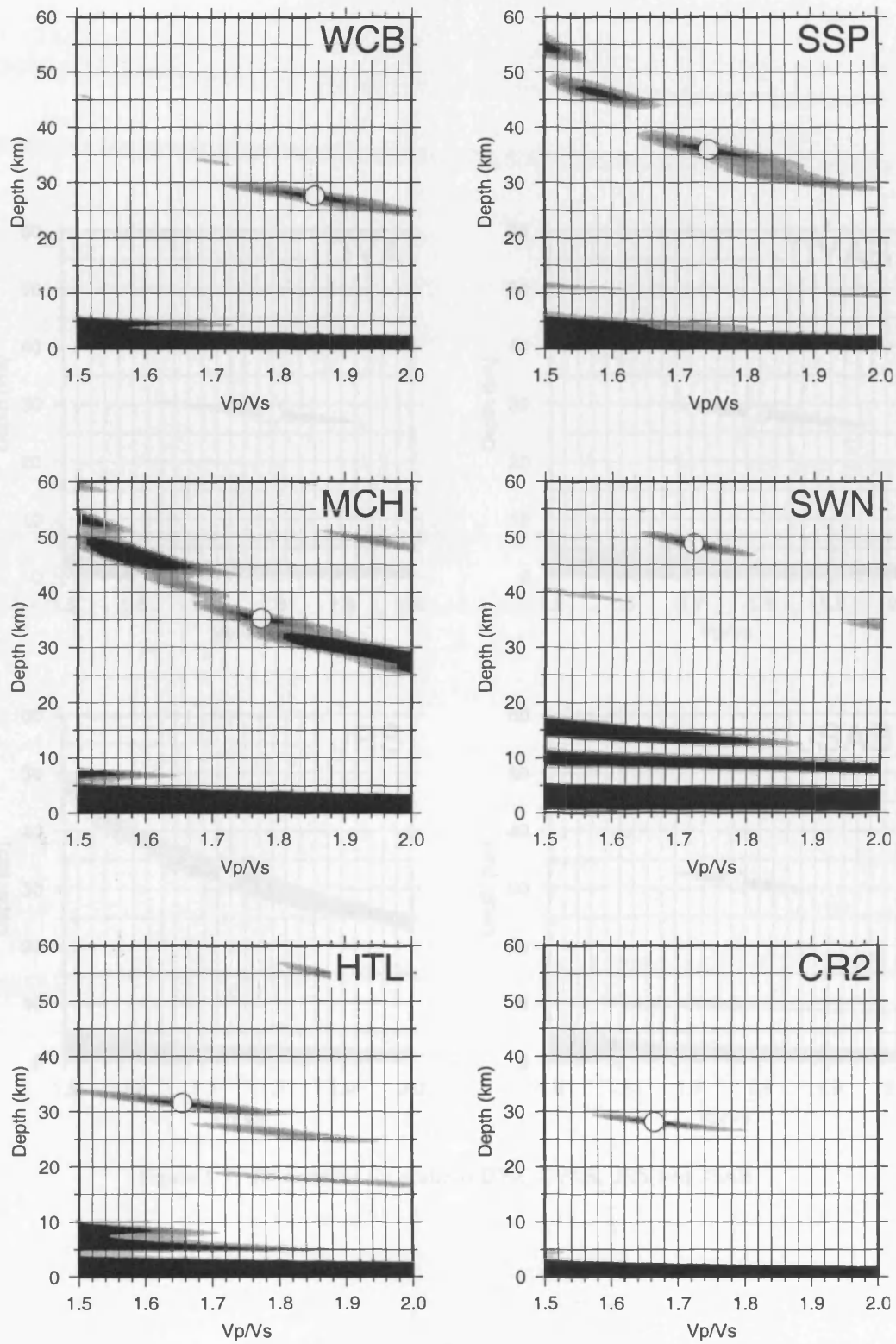


Figure 5.6: $H-\kappa$ stacks for stations WCB, SSP, MCH, SWN, HTL and CR2.

5.3.2 Results

Northwest Highlands

The Northwest Highlands region has $H-\kappa$ stacks from the stations LNW, GPF, MCD, FRR, RRRB, KPL and KARU (Figures 5.2 & 5.3). The stations sample Laurentian crust from the Hebridean, Florida

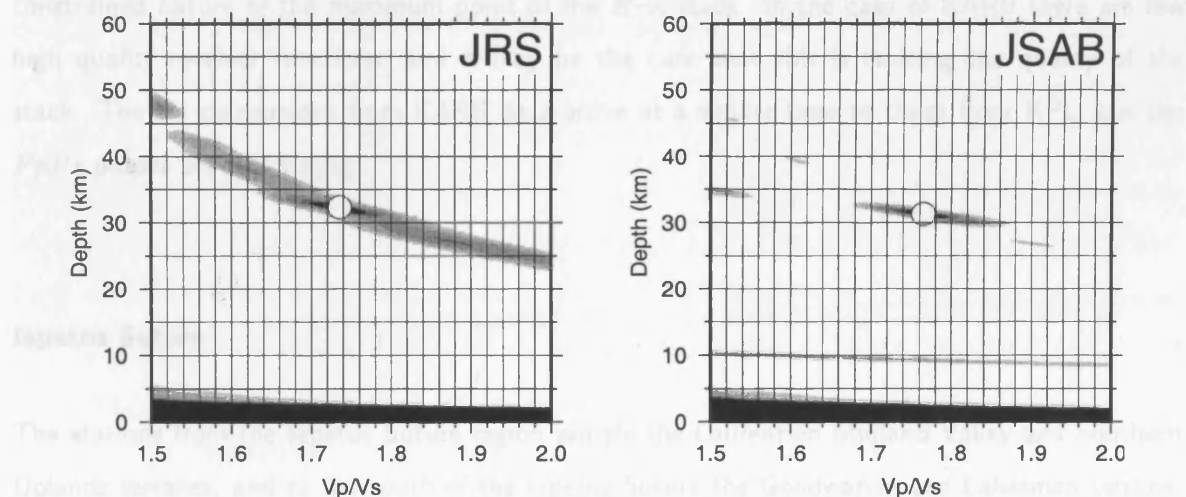
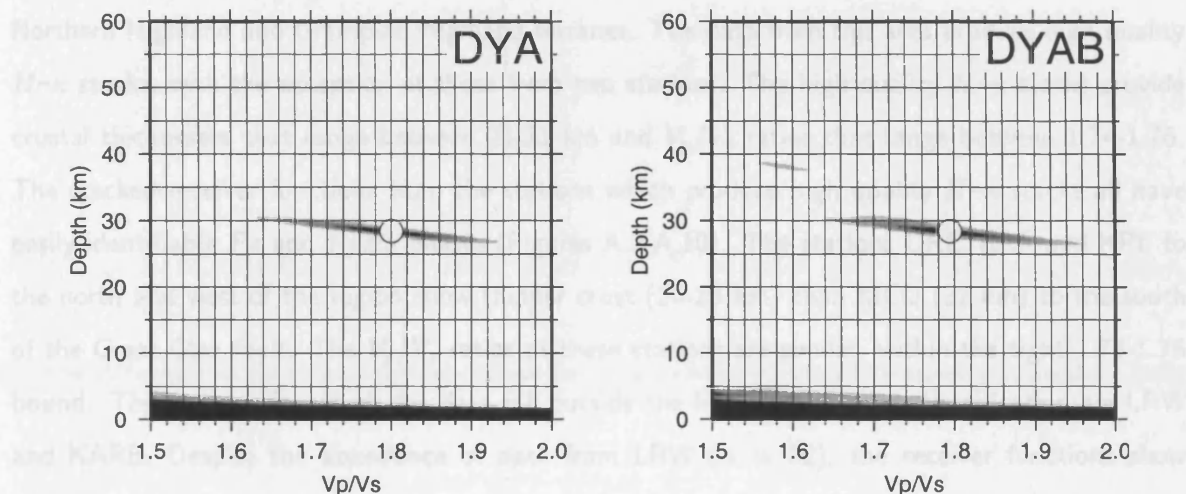


Figure 5.7: $H-\kappa$ stacks for stations DYA, DYAB, JRS and JSAB.

5.3.2 Results

Northwest Highlands

The Northwest Highlands region has $H-\kappa$ stacks from the stations LRW, ORE, MCD, RRR, RRRB, KPL and KARB (Figures 5.2 & 5.3). The stations sample Laurentian crust from the Hebridean, Northern Highland and Grampian Highland terranes. The data from this area produce high quality $H-\kappa$ stacks, with the exception of those from two stations. The high quality $H-\kappa$ stacks provide crustal thicknesses that range between 24-31 km and V_p/V_s ratios that range between 1.74-1.76. The stacked receiver functions from the stations which produce high quality $H-\kappa$ stacks all have easily identifiable P_s and $PpPs$ phases (Figures A.2-A.10). The stations ORE, RRR and KPL to the north and west of the region show thinner crust (24-28 km) than MCD (31 km) to the south of the Great Glen fault. The V_p/V_s ratios at these stations are similar, within the tight 1.74-1.76 bound. The stations for which the data fall outside the high quality stack classification are LRW and KARB. Despite the abundance of data from LRW ($n = 52$), the receiver functions show very low amplitude P_s and $PpPs$ conversions (Figure A.2) and this is reflected in the poorly constrained nature of the maximum point of the $H-\kappa$ stack. In the case of KARB there are few high quality receiver functions, and it may be the case that this is limiting the quality of the stack. The P_s conversions from KARB data arrive at a similar time to those from KPL, but the $PpPs$ phases are less strong.

Iapetus Suture

The stations from the Iapetus Suture region sample the Laurentian Midland Valley and Southern Uplands terranes, and to the south of the Iapetus Suture the Gondwanan and Lakesman terrane. The stations include EDI, EDIB, PGB, ESK, ESKB, GAL, BHH, BBO, BTA, GIM, LMI, HPK and HPKB, the results from which are presented in Figures 5.3-5.5. The crustal thicknesses range between 27-34 km and the V_p/V_s ratios between 1.60-1.81. Again the majority of the $H-\kappa$ stacks are of high quality, with the exception of EDIB, PGB and GAL. The high quality stack from BHH was produced using receiver function data that was calculated using the lower frequency Gaussian filter ($a = 2$) to stabilise the stack that was initially calculated, (Figure A.27). The two stations from the Midland Valley, EDI and PGB produce differing crustal thicknesses at 30 km and 34 km

respectively. The stacked receiver functions for both stations have reasonably clear P_s phases but at PGB the $PpPs$ phase is less well defined, introducing ambiguity to the $H-\kappa$ stack (Figures A.15 & A.19). The stations directly spanning the Iapetus Suture are of high quality, and in particular ESK and BBO show very clear P_s and $PpPs$ phases (Figures A.21 & A.31). GAL has a clear P_s phase but again suffers from a poorly resolved $PpPs$ phase that results in smeared maximum point of the stack (Figure A.25).

Central England & Wales

The stations from central England and Wales (WCB, CWF, CWFB, MCH, SSP, SWN, WOB and TFO) sample the crust of the Midland Platform and Welsh basin, part of the Avalonian terrane. Station WCB lies on the suspect Monian terrane, on the north Wales island of Anglesey. None of the $H-\kappa$ stacks from the stations within this region fall into the high quality category (Table 5.1, Figures 5.5-5.6). Of the intermediate quality results, the crustal thicknesses are the largest measured within the British Isles, ranging between 35-39 km, with the V_p/V_s ratio ranging between 1.71-1.79. The one exception to this range of values is WCB which shows a much thinner 27 km thick crust and a high V_p/V_s ratio of 1.85. For the intermediate quality data the P_s phase is generally well defined, the $PpPs$ phases are, however, unclear. In the case of CWF there is a high amplitude, well-defined P_s phase, and a very low amplitude $PpPs$ phase (Figure A.43). Two of the stations, SWN and TFO, are classified as poor quality (3) stacks. The stack for SWN reveals an anomalously thick crust of 48 km, with a V_p/V_s ratio of 1.72. The stacked receiver functions from both stations contain many high amplitude phases. It is possible to identify consistent phases at arrival times that correlate with Moho P_s phases, but there are too many high amplitude reverberations to identify which, if any, is the Moho $PpPs$ phase that is required to constrain the $H-\kappa$ stack (Figures A.51 & A.55). The crustal thicknesses for these stations have been directly calculated from T_{P_s} assuming a chosen V_p/V_s ratio (Equation 3.29). Using a V_p of 6.3 km s^{-1} and a V_p/V_s ratio of 1.73, this method renders thicknesses of 48 km and 29 km for SWN and TFO respectively. There is a large uncertainty with these estimates due to the ambiguity in the receiver function phase identification. It is therefore reasonable to exclude these measurements from any further interpretation or discussion.

Southwest England

The few stations from the southwest of England (HTL, CR2, DYA, DYAB, JRS and JSAB) are located on Avalonian crust that has been deformed by the Variscan orogeny. In the case of the stations on the Channel Islands, JRS and JSAB, the underlying crustal geology is from the Armorican terrane. These stations return high quality $H-\kappa$ stacks with the exception of HTL and JRS (Table 5.1, Figures 5.6-5.7) . The crustal thicknesses measured by the stations located on the mainland range between 28-31 km, and the V_p/V_s ratios between 1.66-1.80. The $H-\kappa$ stacks for the two stations on the Channel Islands return similar $H-\kappa$ values of ~ 32 km and ~ 1.75 respectively.

5.4 Discussion

5.4.1 $H-\kappa$ Stacking Method

The analysis of receiver function data using $H-\kappa$ stacking allows non-subjective investigation of crustal structure. The aim of the method is to produce robust estimates of crustal thickness and V_p/V_s ratio. There are however a number of difficulties in the analysis of receiver function data using the $H-\kappa$ stacking technique. 1) There is a considerable velocity-depth trade-off in the results. 2) Although the standard deviations of $S(H_{max}, \kappa_{max})$ may be estimated, it is not possible to quantify the other errors in the procedure. In the present study constraint on the velocity-depth trade-off is offered by the onshore deep seismic refraction profiles. The estimates of the standard deviation of H are between 1 and 2 km. The range of observed crustal thicknesses is between 25 and 35 km; this variation is significantly larger than the standard deviations. The range of observed V_p/V_s ratios is between 1.60 and 1.85, but many of the measurements are in the middle of this range. The estimates of standard deviation of κ are between 0.05 and 0.1. This means that the observed variations in κ are not as significant as the variations in H .

5.4.2 Observed $H-\kappa$ Stack Quality

In the majority of receiver functions from the stations from within the British Isles the Moho P_s phase can be clearly identified. The positions of the maxima of the $H-\kappa$ stacks are dependent

upon how well resolved the $PpPs$ phases are. A qualitative estimate of the resolution of the receiver function phases and therefore stack quality has been used. From this analysis only two stations have failed to be classified as either high or intermediate quality and 19 of the 34 stations have been classified as high quality. Applying a lower pass Gaussian filter when calculating the receiver functions has helped to stabilise those $H-\kappa$ stacks where the $PpPs$ phase is poorly resolved. The $PpPs$ phase is better resolved when more of the higher frequency signal is minimised, but using this data increases the estimates of the standard deviation of $S(H_{max}, \kappa_{max})$. The qualitative assessments reveal that the $H-\kappa$ stacks for the data from the stations in the north of the British Isles are generally of better quality than those in the south. In particular the $PpPs$ phase for the stations in central England and Wales are poorly resolved.

5.4.3 Non-Geological Variation

Before any geological interpretation can result from the $H-\kappa$ stacking analyses, an evaluation of any possible non-geological sources of stack variation must be made. Variation in the $H-\kappa$ stack results that do not reflect the true velocity structure of the crust may result from;

1. The phases that contribute to form the maximum point of the $H-\kappa$ stack not being the Moho Ps , $PpPs$ and $PpSs$ phases.
2. The Moho Ps , $PpPs$ and $PpSs$ phases not arriving at the expected times derived from a 1D crustal velocity model.

In most studies the Moho Ps and $PpPs$ are usually the strongest phases of the receiver function following the direct P arrival (Zhu and Kanamori, 2000). In case 1, the most likely cause of anomalous phases in the observed receiver functions is the presence of high amplitude reverberations from a strong velocity contrast close to the surface which mask the Moho phases. However other anomalous high amplitude phases may occur. If the Moho phases cannot be resolved from reverberations or noise within the observed receiver functions, the results of the $H-\kappa$ stacks of these data will not reflect the true crustal thickness and V_p/V_s ratio beneath the station. The presence of unidentified high amplitude phases in the receiver functions has been noted in the qualitative analysis of the $H-\kappa$ stacks.

In case 2, the relative timing of the P_s , $PpPs$ and $PpSs$ phases may be altered if the geological structure departs from the 1D model. In most $H-\kappa$ stacking studies authors follow Zhu and Kanamori (2000), suggesting that by stacking receiver functions from a range of different distances and directions, the effects of lateral structural variation are suppressed and an average crustal model may be obtained. To test this argument a series of $H-\kappa$ stacks have been performed for synthetic receiver functions entering a simple 3-D model from different backazimuths. The synthetic receiver functions have been calculated using a three-dimensional raytracing program, used by Cassidy (1992). The model used to produce the synthetic receiver functions analysed in the $H-\kappa$ stacks consists of a single layer crust ($H=30$ km, $\kappa=1.73$ and $V_p=6.3$ km s⁻¹) with a Moho dipping 10°E. The results show that the $H-\kappa$ stacks of the individual 3-D receiver functions from different backazimuths do not oscillate about the 1D solution as might be expected (Figure 5.8a). The maximum up-dip receiver functions (BAZ 270°) produce a lower H , and higher κ than the reference 1D stack. Interestingly the down dip receiver functions produce only a slightly greater H than the reference model, whilst κ still remains slightly higher than the true 1D solution. It is therefore clear that if receiver functions from an even distribution of backazimuths were stacked, the correct 1D solution would not be obtained through the averaging process suggested by Zhu and Kanamori (2000).

These findings may be explained by the results of Cassidy's (1992) study, in which the difference in lateral sampling of the P_s phase and subsequent reverberations on horizontal and dipping interfaces is examined (Figure 5.8b&c). This investigation finds that dipping structure significantly alters the raypath geometry of receiver function phases. The alteration to the raypath geometry and relative timing of the receiver function phases results in the misinterpretation of crustal structure by the $H-\kappa$ stacking technique. The up-dip receiver functions sample a much greater volume of crust, whereas the down dip events actually sample a smaller volume of crust than those from the 1-D model. The synthetic receiver functions were calculated with an epicentral distance of 67°. The distortion of the stack would be increased with decreasing epicentral distance, and vice-versa. Further 2D or 3D structure on the Moho allows for many more possible variations in the resultant receiver functions and $H-\kappa$ stacks. The up-dip migration of the maximum point of the $H-\kappa$ stack is difficult to eliminate. A review of the results of any previous seismic experiments in the study area may allow comment on the likelihood of dip effects occurring. However, geological structures are rarely simple and some alteration of the $H-\kappa$ stack results due to deviation from

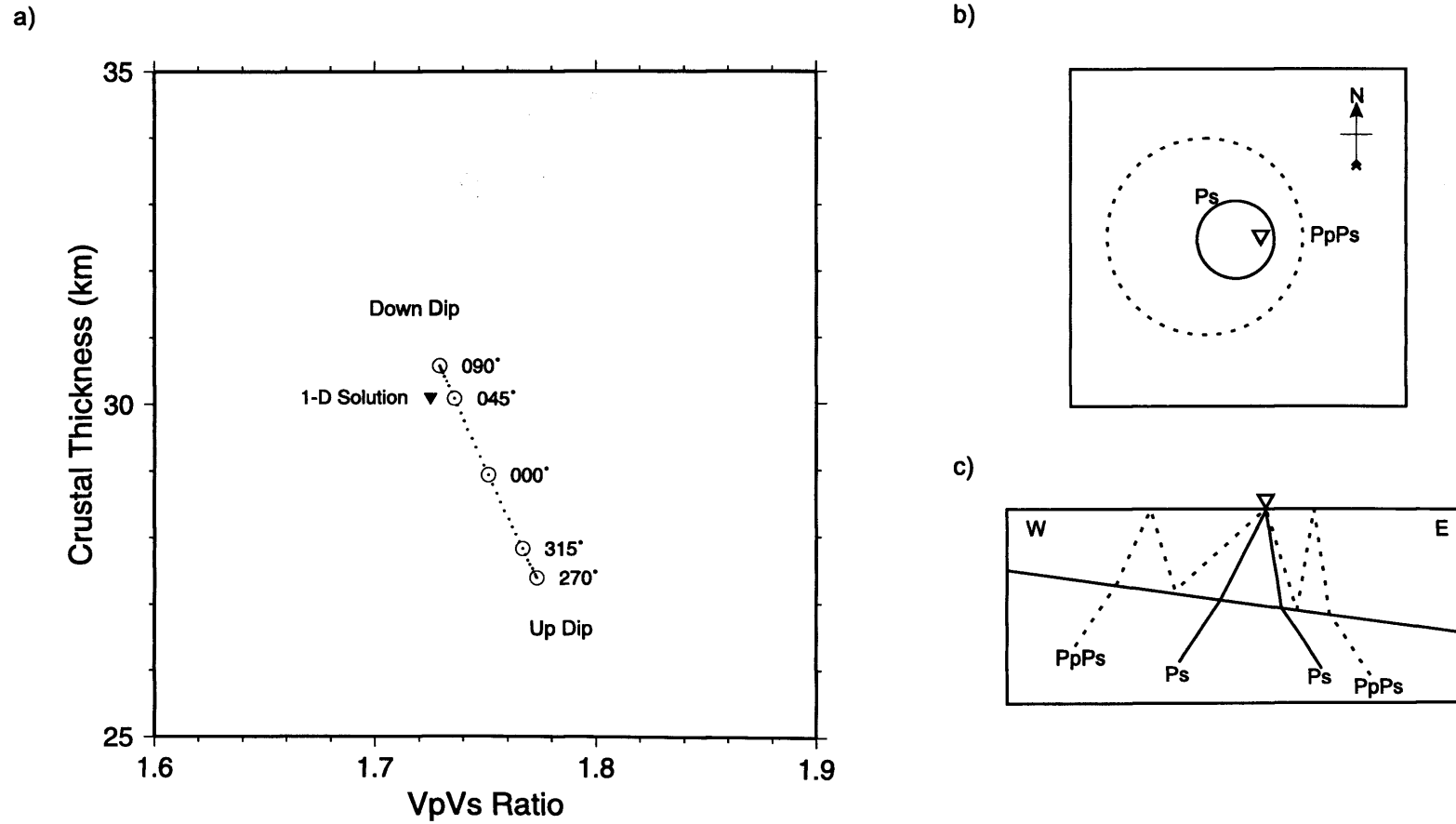


Figure 5.8: a) The maximum points of $H-\kappa$ stacks for the synthetic receiver functions from a dipping Moho model. The stacks use data from variable backazimuths (labelled next to the maximum point). The 3D model consists of a single layer crust of $H=30$ km, $\kappa=1.73$, $V_p=6.3$ km s $^{-1}$ and a dipping Moho of 10° striking 000. The dots are 5° increment models in between the labelled results. The inverted triangle is the $H-\kappa$ stacking result from a 1D synthetic receiver function using identical parameters. b) A sketch of the plan view of the lateral extent of the crustal raypaths of the P_s and $PpPs$ phases over a dipping layer model (Cassidy, 1992). c) A sketch of the cross sectional view of the lateral extent of the P_s and $PpPs$ phases through the dipping layer model.

1D structures must be expected.

The crustal thickness within the British Isles varies between 25 and 35 km, but these changes occur over large distances (Chadwick and Pharaoh, 1998). It seems unlikely that there are Moho dips greater than 5° beneath any of the stations from which data are analysed in this study. Below this threshold the variation in the $H-\kappa$ stack results caused by misinterpretation of receiver functions passing through dipping structure is <1 km for crust with a similar velocity structure to the British Isles. This difference is of a similar magnitude to the measured standard deviations of the $H-\kappa$ stacks. A more likely scenario that may cause raypath deviation in the British Isles is that significant dipping structures are present in the crust or upper mantle. This type of crustal structure has been imaged in numerous studies on the BIRPS profiles (e.g. Hall et al., 1984; Freeman et al., 1988; Klemperer and Hobbs, 1991). An intra-crustal interface with around 30° dip only produces small changes in the $H-\kappa$ stack result away from the 1D approximation. This is because the change in seismic velocity across such crustal interfaces is small (normally $<0.5 \text{ km s}^{-1}$), and the raypath geometry of the receiver functions is therefore only perturbed slightly by the dipping structure. Significant dipping structures in the upper mantle have been found to the north and west of Scotland (e.g. McBride et al., 1995; Price and Morgan, 2000; Morgan et al., 2000). The observed velocity increase over these boundaries is $\sim 0.3 \text{ km s}^{-1}$, but in places a significant low-velocity-zone has been modelled above the reflecting interface (Price and Morgan, 2000). Applying the $H-\kappa$ stacking technique to receiver function data that has been generated using a model similar to Price and Morgan's (2000) produces only a small spread (≤ 1 km) of measured crustal thicknesses relative to backazimuth. Although the low-velocity-zone has a considerable dip and velocity change across it, the zone is relatively thin (≤ 5 km). The alteration of the raypath geometry in this mantle structure does not significantly alter the estimate of the crustal thickness and V_p/V_s ratio obtained through $H-\kappa$ stacking.

5.4.4 Geological Implications

Overview

The results of the receiver function H - κ stacking investigation reveal significant variations in the measurements of crustal thickness and bulk V_p/V_s ratio. The magnitude of these variations are outwith those that might be expected by mis-interpretation due to erroneous input velocity or 3D effects. The variations in H and κ are related to the complex crustal history and evolution of the British Isles. Since the Caledonian and subsequent Variscan orogenies, the crust of the British Isles has been substantially thinned, through a process of isostatic uplift and erosion interspersed by periods of basin forming extension. The derived crustal thicknesses provide insight into the variation in thinning that has occurred, and how this relates to the terrane structure of the British Isles. The V_p/V_s ratios obtained by H - κ stacking are sensitive to several variables; 1) bulk crustal composition, 2) crustal temperature and 3) crustal fluid content. The variation of Poisson's ratio due to changes in pressure and temperature is small, and therefore laboratory based measurements of Poisson's ratio are directly comparable over a wide range of crustal depths (Christensen, 1995). The V_p/V_s ratios again provide the opportunity to study how the variation in crustal properties is related to the terrane structure of the British Isles.

H - κ Maps

A series of maps of the results derived from the H - κ stacks are presented (Figures 5.9a,b & 5.10a,b). The crustal thickness maps show considerable variation over the British Isles, which are consistent with the seismic reflection crustal thickness variations presented by Chadwick and Pharaoh (1998). The variations recorded by this receiver function study and Chadwick and Pharaoh (1998) do to some extent correlate with the underlying terrane geology. In general the thickest areas of the British Isles are those of Central England and Wales, the Midland Platform and Welsh Basin. Regions of northwest Scotland and southwest England show crust thinned by extension related to the Atlantic and North Sea opening. The variation in V_p/V_s ratio over the British Isles follow a less consistent pattern than the crustal thickness values.

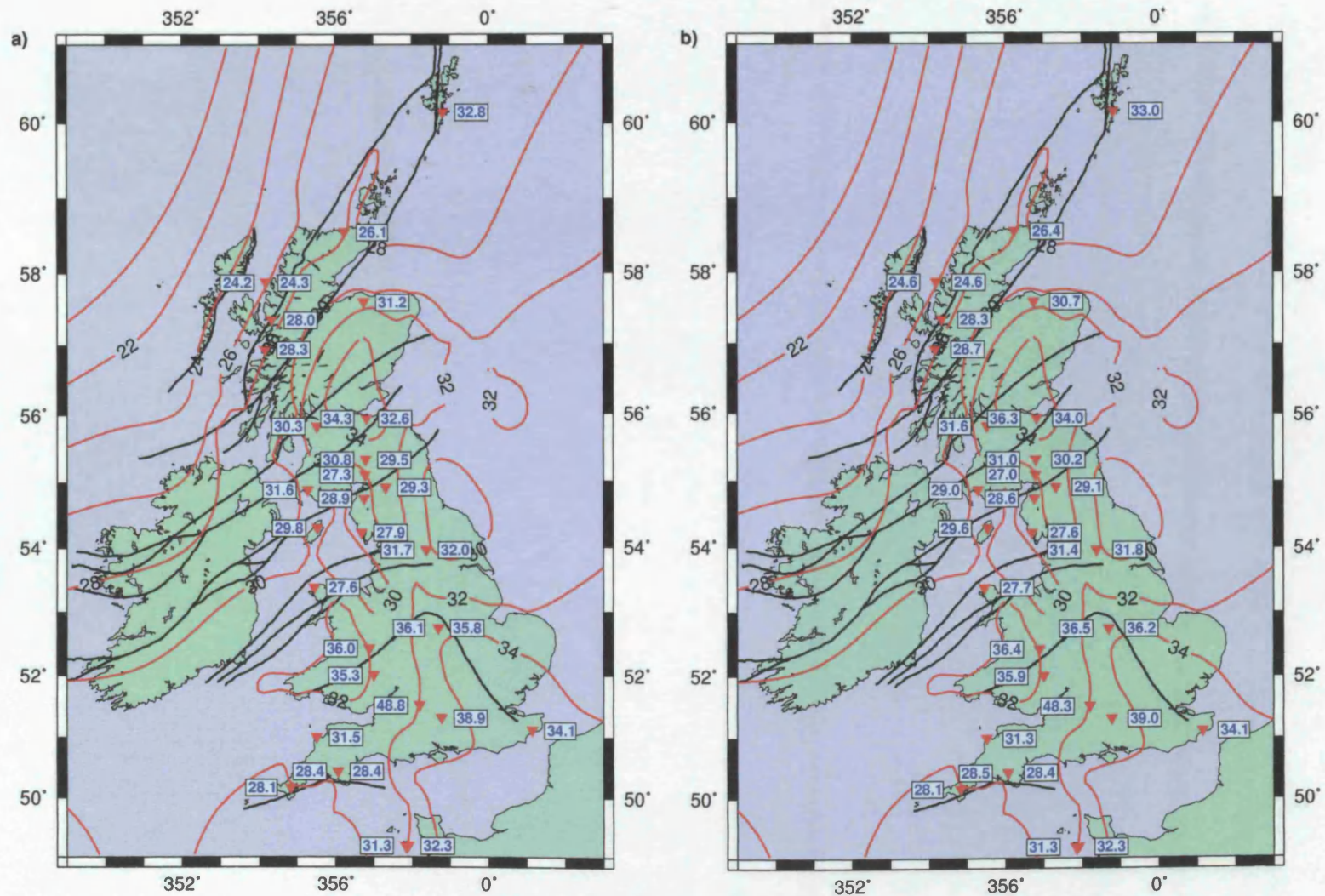


Figure 5.9: a) A map of crustal thicknesses output from Analysis 1, using a constant V_p of 6.3 km s^{-1} . The crustal thicknesses reported by Chadwick and Pharaoh (1998) are contoured and plotted beneath the receiver function values. The black lines represent the terrane boundaries presented in Figure 1.1. b) A map of crustal thicknesses from Analysis 2, using input V_p defined by seismic refraction studies (Table 5.1).

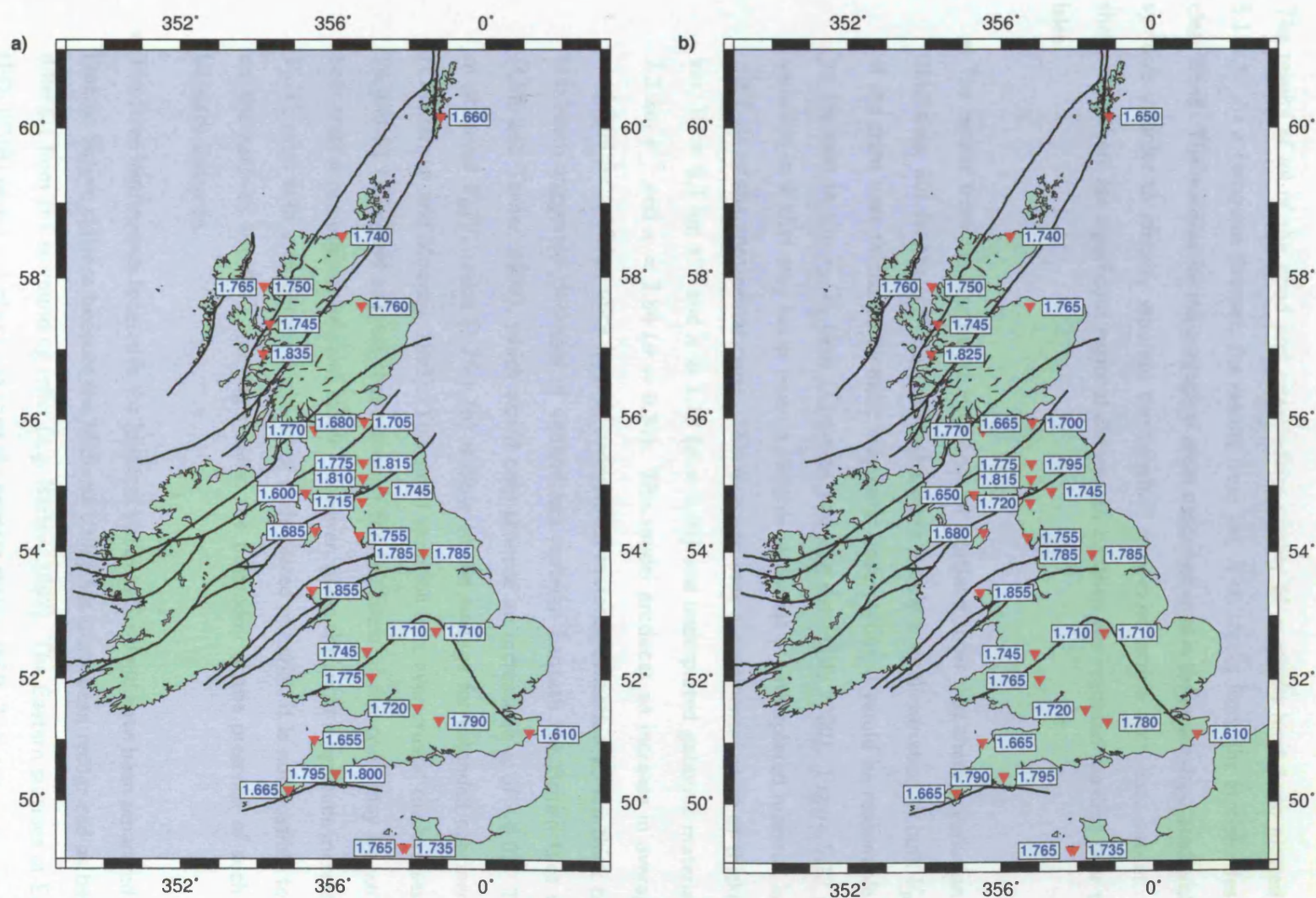


Figure 5.10: a) A map of average crustal V_p/V_s ratios output from Analysis 1, using a constant V_p of 6.3 km s^{-1} . The black lines represent the terrane boundaries presented in Figure 1.1. b) A map of average crustal V_p/V_s ratios output from Analysis 2, using input V_p defined by seismic refraction studies (Table 5.1)

***H*- κ Correlation**

The results of all of the good and intermediate quality *H*- κ stacks have been plotted in Figure 5.11a,b. As a complete dataset, the results from the *H*- κ stacks from the British Isles reveal no clear trend. The maxima for the geological areas described in this thesis are decorated with different symbols in order to identify separate trends which may exist within each data subset. This plot shows that there are significant regional differences between the recorded maxima over the British Isles.

- The results from the Northwest Highlands of Scotland show that there is variation in crustal thickness, but for the high quality results there is only a small variation in bulk V_p/V_s ratio. If the crust were thickened by basic magmatic underplating it would be reasonable to expect an increase in the V_p/V_s ratio (Cheverot and van der Hilst, 2000). Figure 5.12 shows the variation in κ that may occur when a variable thickness of underplated material is added to the base of the continental crust. This example uses a continental crust of thickness = 25 km, $V_p = 6.3 \text{ km s}^{-1}$ and $\kappa = 1.73$ ($\sigma = 0.25$) and underplated gabbroic material of $V_p = 7.3 \text{ km s}^{-1}$ and $\kappa = 1.84$ ($\sigma = 0.30$). This model produces an increase in average crustal κ of 0.04 for a 10 km thick layer of underplated material beneath a 25 km thick crust. The maximum suggested thickness of underplated material beneath the British Isles is ~ 4 km (Clift and Turner, 1998), which would only produce an increase in κ of ~ 0.02 . The range of observed V_p/V_s ratios (1.74-1.76) is close to the average for extended continental crust (Christensen and Mooney, 1995). This small variation in κ over crustal thicknesses between 24 and 31 km would suggest that although the thickness of the crust may be varying, the bulk crustal composition remains stable. However, given the small magnitude increase in bulk V_p/V_s ratio with a moderate thickness of underplated material, it is not possible to exclude, on the basis of the *H*- κ stacking results, the possibility of the presence of such material beneath this area.
- The three instruments located in the Midland Valley of Scotland have been separated from the Iapetus Suture stations because the Midland Valley has often been recognised as being quite different from the surrounding crust (e.g. Barton, 1992). The Eastern stations of Edinburgh (EDI, EDIB) reveal a thicker crust than the western station PGB. This may indicate that there is significant lateral change in the structure of the Midland Valley. However the quality of

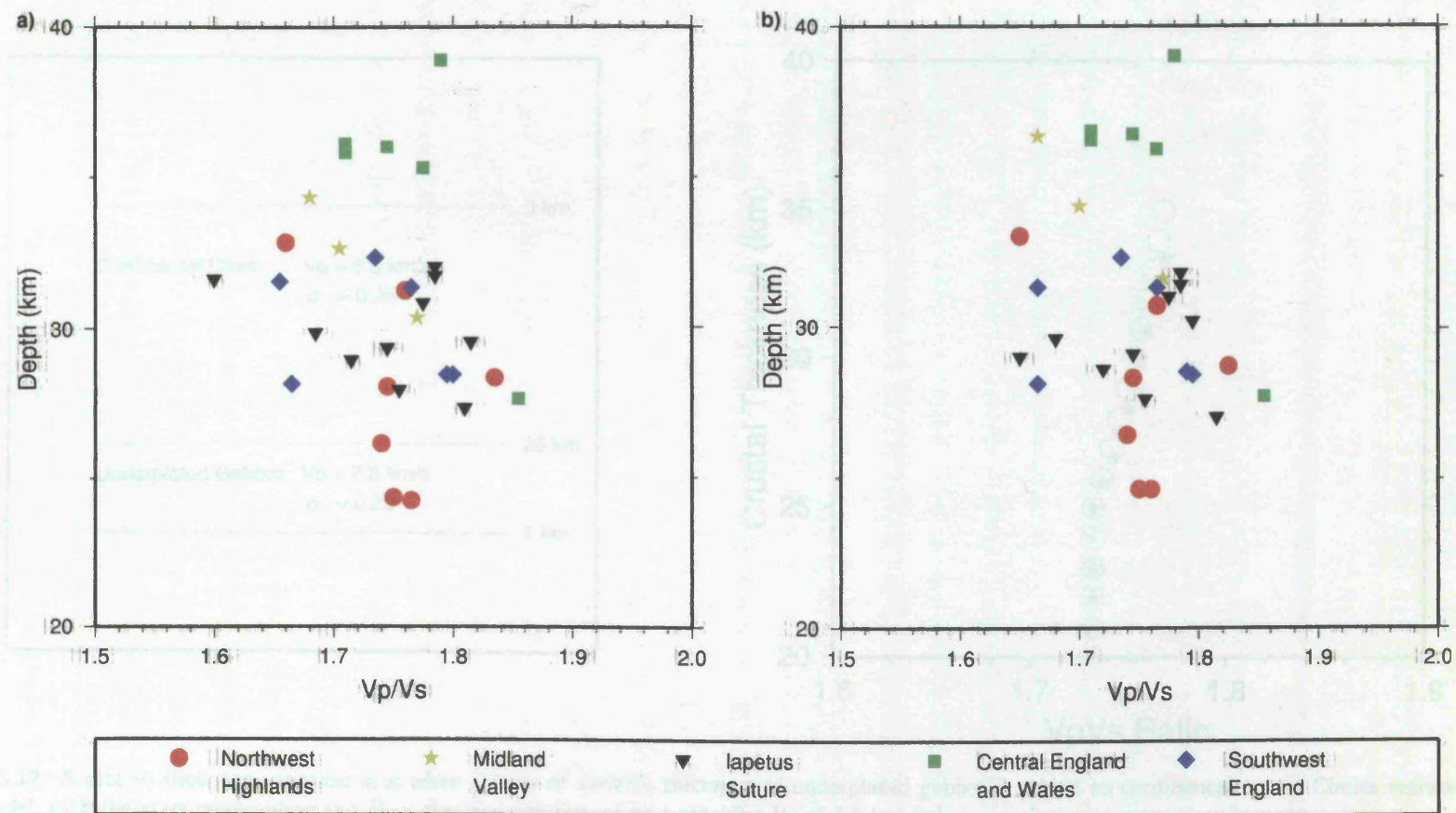


Figure 5.11: a) A plot of all of the $H-\kappa$ stacking values from the receiver function study calculated using a V_p of 6.3 km s^{-1} b) A plot of all of the $H-\kappa$ stacking values from the receiver function study calculated using the V_p values listed in Table 5.1

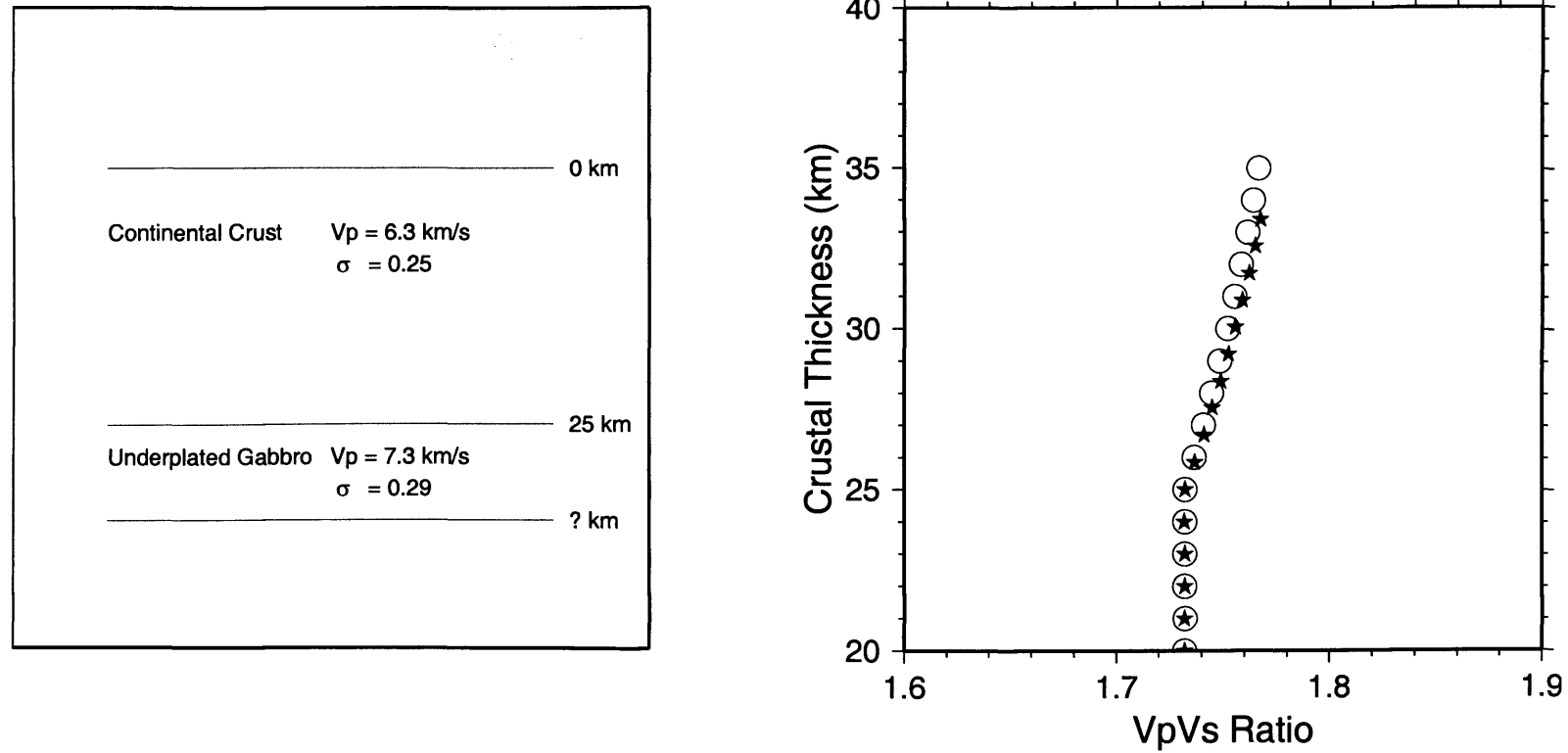


Figure 5.12: A plot to show the variation in κ when a layer of variable thickness of underplated gabbro is added to continental crust. Circles represent the true model, with the stars representing the $H-\kappa$ stacking solution using a stacking V_p of 6.3 km s^{-1} .

the data from PGB are relatively poor and therefore the possibility of errors being introduced into the $H-\kappa$ analysis must be taken into account. The crustal thickness data from the WINCH profile do not reveal substantially thickened crust beneath the westerly extension of the Midland Valley (Hall et al., 1984), and it is therefore possible that the thicker crust is limited in extent to the eastern regions.

- The results from the lapetus Suture area reveal a crust that has only a small variation in crustal thickness, between 28-31 km. The crustal thicknesses do not correlate with the values ~ 33 km observed by the LISPB and CSSP profiles (Barton, 1992; Al-Kindi, 2002). There is more variation in the V_p/V_s ratio (1.60-1.81). This variation does not seem to be systematic in either geographical distribution or in relation to the measured crustal thicknesses. Because the variation in V_p/V_s ratio does not correlate with any geographical pattern it seems unlikely that its cause is related to lateral variation in the mineralogical composition of the crust. It is also possible that the variations in V_p/V_s ratio could be caused by 3-D structure within the crust. This cause of the variation is plausible as the lapetus Suture area contains bright dipping reflectors in the BIRPS seismic sections of the area. However it has already been noted that such structures do not cause significant alteration of the $H-\kappa$ stacks of the resultant receiver functions (see Section 5.4.3). The low amplitude of the velocity contrast over the dipping structures means that the raypath geometry of the teleseismic phases and the relative arrival time of the P_s , $PpPs$ and $PpSs$ phases are only slightly modified.
- Central England and Wales clearly have thicker crust than the majority of the British Isles. There are however variations within the measured V_p/V_s ratio. The variation in κ may relate to variation in the bulk crustal composition, but again there seems to be no systematic geographical variation to these changes. The $H-\kappa$ stacks of the receiver functions from this area are not high quality due to the poor resolution of the $PpPs$ phases. The variation in the measured V_p/V_s ratio may therefore be caused by the uncertain nature of the receiver function $PpPs$ phases. The southernmost station from the lapetus Suture group (HPK) shows crust that is thicker (~ 32 km) than the majority of the observed H values (28-31 km) for the area. This may represent a thickening of the crust toward the Midland Platform. The most notable anomaly in the $H-\kappa$ results from this area is from WCB, where H is ~ 28 km and $\kappa \sim 1.85$. WCB is located close to the East Irish Sea and the crustal thickness is close to the values recorded at other stations bounding the Sea, such as GIM and LMI. The V_p/V_s ratio is however unusually high for this study. High values of κ are characteristic

of suites of basic igneous rocks such as oceanic crust. The Monian terrane is part of an Iapetus subduction accretionary prism (Bluck et al., 1992), and the high V_p/V_s ratio may be explained if the prism is underlain by significant amounts of oceanic crust. A value of $\kappa = 1.85$ would be difficult to explain with a model of continental crust underplated with basic igneous material. Figure 5.12 shows that the addition of 10 km of underplated material to the base of the continental crust will only increase the bulk crustal V_p/V_s ratio by ~ 0.04 .

- The results from the stations from southwest England show that the crust is becoming progressively thinner toward the Atlantic ocean. There are only a small number of stations in the area and there is no clear pattern in the distribution of the stack results in the H - κ domain.

Receiver Function vs Controlled Source Seismic Crustal Thicknesses

The maps of crustal thickness plotted above the crustal thickness contours of Chadwick and Pharaoh (1998) reveal that both the receiver function and seismic reflection data highlight similar variations over the British Isles (Figures 5.9a&b). To further examine the similarity between the two measurements of crustal thickness, the receiver function thicknesses have been cross plotted against seismic reflection and refraction thicknesses (Figure 5.13a,b). The two figures show the receiver function results from Analysis 1 & 2 respectively. In Analysis 2, where separate input V_p 's for each station have been used, the majority of stations fall within ± 2 km of a direct correlation between the receiver function and seismic crustal thicknesses. The use of individual stacking velocities for each station reduces the number of stations which fall outside the ± 2 km limits relative to the constant V_p analysis. The average RMS misfit between the receiver function and seismic thicknesses has been reduced from 1.75 km in Analysis 1 to 1.65 km in Analysis 2.

There are several errors that have been introduced into the dataset of crustal thicknesses that have been correlated. Firstly there are the errors associated with the H - κ stacking technique which include the standard deviation of the maximum point of the stack, and the trade-off between stacking V_p and depth. The standard deviations on the H domain are in the order of ± 1 -2 km. A change in stacking V_p of 0.2 km s^{-1} can result in a change in depth of ~ 1 km (when the crust is ~ 30 km thick). Secondly there are the errors associated with the seismic reflection and refraction techniques which may amount to up to ± 1 km (Barton, 1992). Finally, the extrapolation of the

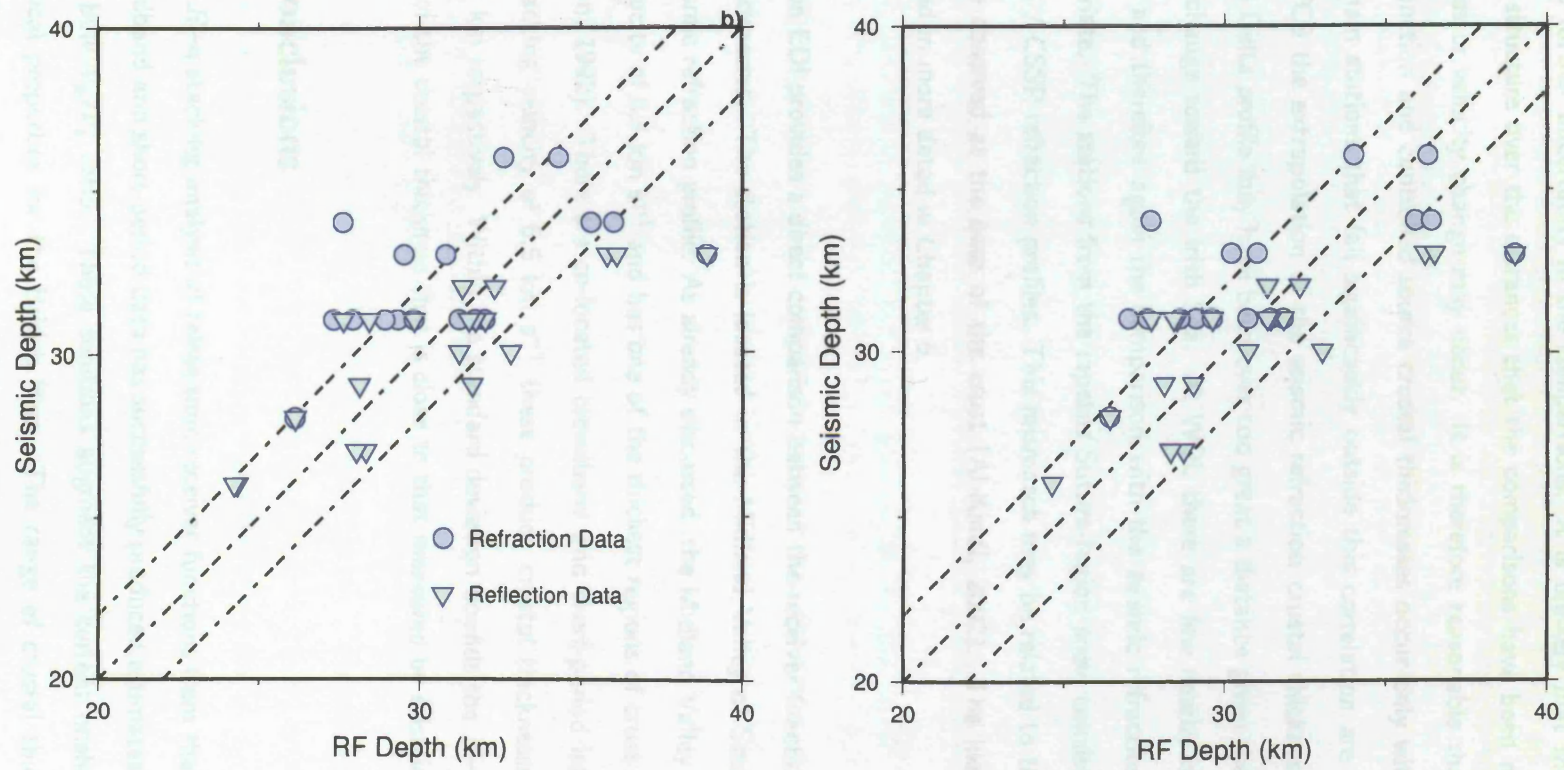


Figure 5.13: a) A cross plot of crustal thicknesses from the receiver function study against crustal thicknesses from seismic refraction and reflection methods. The stacks were calculated using a V_p of 6.3 km s^{-1} b) A cross plot of crustal thicknesses from the receiver function study against crustal thicknesses from seismic refraction and reflection methods. The stacks were calculated using the V_p values listed in Table 5.1. The dashed lines show a 1:1 correlation, and a variation of ± 2 km.

controlled source seismic crustal thicknesses to the location of the receiver function stations has introduced further uncertainty in the comparisons. It is unlikely that there is significant change in crustal structure over the distances that the comparisons have been made, but small changes in thickness or velocity change may occur. It is therefore reasonable that the correlation of the receiver function and controlled source crustal thicknesses occur only within ± 2 km. In the plot the two main stations that fall significantly outside this correlation are WCB and WOL. In the case of WCB the extrapolation of the seismic refraction crustal thickness from the north end of the LISPB Delta profile may have been over too great a distance given the proximity of the crustal thickness change toward the Irish Sea. At WOL there are few nearby measurements for crustal thickness, and therefore again the comparison with the seismic refraction measurement may not be appropriate. The stations from the Iapetus Suture region show consistently thinner crust than the LISPB / CSSP refraction profiles. This mismatch may be related to the presence of magmatic underplate observed at the base of the crust (Al-Kindi, 2002). The likelihood of this has been investigated in more detail in Chapter 6.

The Station EDI provides a direct comparison between the receiver function and seismic refraction crustal thicknesses. The station is located in the Midland Valley of Scotland, directly upon the LISPB seismic refraction profile. As already discussed, the Midland Valley has an elevated average crustal velocity of 6.5 km s^{-1} and has one of the thickest regions of crust in the British Isles at 36 km (Barton, 1992). There are co-located broadband and short-period instruments at EDI. With a $H-\kappa$ stacking velocity of 6.5 km s^{-1} these produce crustal thicknesses of 36.3 ± 2.0 km and 34.0 ± 1.58 km respectively. Within the standard deviation bounds the $H-\kappa$ stacks have produced estimates of the crustal thickness that is close to that measured by seismic refraction.

5.5 Conclusions

- The $H-\kappa$ stacking analysis of teleseismic receiver functions from the British Isles using both broadband and short-period data has successfully produced estimates of the crustal thickness and bulk V_p/V_s ratio. These solutions augment the current onshore database of crustal physical properties for the British Isles. The range of crustal thicknesses (25-36 km) is significant compared with the estimates of the associated standard deviations (1-2 km). However, the range of V_p/V_s ratios (1.60-1.85) is less significant compared to the associated

standard deviations (0.05-0.1).

- The quality of the $H-\kappa$ stacking results is dependent upon how well resolved the Moho P_s phase and subsequent reverberations are. It is not possible to fully quantify the variability of the $H-\kappa$ stacks caused by poor resolution of the receiver function phases. Comparing the resultant $H-\kappa$ stacks and input receiver functions from all of the stations has allowed a qualitative description of the data to be made.
- The results of $H-\kappa$ stacking can be affected by using receiver functions that sample 3-D structures. For example given an even azimuthal distribution of raypaths passing through a moderately dipping (10°) Moho the crustal thickness is underestimated, whilst the V_p/V_s ratio is overestimated. The affect of more complicated 3D structure on the results of $H-\kappa$ stacking has not been tested.
- Synthetic tests show that the effect of simple 2D or 3D structure on the $H-\kappa$ stacks does not cause significant problems in the interpretation of the receiver function data from the British Isles. Any structure on the Moho is likely to be of low amplitude and high wavenumber, and therefore low dip angle. This results in only a small perturbation of the raypath geometry of the teleseismic events, and therefore the dip effect is not significant. Dipping intra-crustal and upper mantle structures, as have been imaged by the BIRPS seismic profiles, do not cause major alteration to $H-\kappa$ estimates. This is due to the low amplitude of the velocity contrast over the reflecting boundary which again does not cause significant alteration of the raypath geometry.
- The $H-\kappa$ stacking crustal thickness results are, within moderate error bounds, consistent with models from both previous seismic reflection and refraction experiments, with the exception of the stations from the Iapetus Suture region. This leads to the conclusion that the contrasting seismic methods are resolving the same velocity discontinuity. The crustal thicknesses obtained from the $H-\kappa$ stacking method are sensitive to the input V_p . Two separate $H-\kappa$ analyses have been performed using an average crustal V_p for all stations and locally derived V_p . The second analysis using individual V_p for each station slightly improved the correlation between the receiver function crustal thicknesses and those found by controlled source seismic methods.
- The crustal thickness estimates from both teleseismic receiver functions and controlled source

seismic experiments correlate to some extent with the underlying geological terrane structure. Central England and Wales are the thickest regions of the British Isles, and the east of the Midland Valley also has anomalously thick crust. Elsewhere the crust is ~ 30 km thick, thinning to ~ 25 km in the Northwest Highlands and ~ 28 km in Southwest England.

- The estimates of V_p/V_s ratio are not as consistent over the British Isles as the crustal thickness estimates. In the Northwest Highlands where there are good quality receiver functions the V_p/V_s ratio values show that the composition of the crust is stable despite varying crustal thicknesses. However in other areas where κ is less stable there is no direct correlation between crustal thickness and V_p/V_s ratio.

Chapter 6

Modelling

6.1 Introduction

This chapter covers the detailed crustal modelling of both the short-period and broadband data from the British Isles, providing a more detailed investigation of the receiver function data than $H-\kappa$ stacking. The chapter is presented as sections covering geographical areas from which the receiver function data samples distinct geological structures. As in the previous chapters in this thesis the geographical-geological areas are; The Northwest Highlands, The Iapetus Suture Zone, Central England and Wales, and Southwest England. The data from the stations within these regions provide onshore constraint of the structures imaged by the extensive BIRPS profiles. Firstly, the data from the co-located broadband and short-period instruments at Eskdalemuir (ESKB & ESK) have been analysed. This demonstrates that the modelled crustal structure can be affected by the bandwidth of the recorded teleseismic events. Where short-period data is analysed, the approach taken is to produce the best fitting 1D forward model using the relevant *a priori* information to constrain the models. In this case the forward model for the synthetic seismograms uses limited bandwidth to reproduce the negative lobes that are generated when performing the receiver function deconvolution with short-period data.

6.2 Modelling Example

6.2.1 Eskdalemuir ESK/ESKB

The co-located short-period and broadband instruments at Eskdalemuir provided an opportunity to investigate the effect of using short-period rather than broadband data in receiver function analysis. Eskdalemuir is located in the Southern Uplands of Scotland, with the nearby LISP profile providing constraint on the crustal velocity structure. The raw receiver function data for ESK and ESKB (Figure 4.1) are shown in Figures A.20 & A.22. Both the radial and tangential receiver functions have been stacked in 20° backazimuth bins, and the bin window has been rotated through 360° at 5° intervals (Figures A.21 & A.23). The binning has been performed to enhance the receiver function phases by minimising noise in the signal through destructive interference when data are stacked. As can be seen in the raw receiver function plots the majority of these bins contain several receiver functions. Applying this technique, the receiver functions may be smoothed over a maximum of $\pm 10^\circ$ from their true backazimuth position. Azimuthal variation in receiver functions may be caused by dipping structure, seismic velocity anisotropy in 1D media, or more complicated 3D structure. The $H-\kappa$ stack of the receiver function data for each instrument is also presented, and the times of the Moho P_s and $PpPs$ phases calculated from the stack are indicated on the plots of both the radial and tangential receiver function data. The stacked radial and tangential data plots for other stations in Appendix A are produced using the same binning parameters.

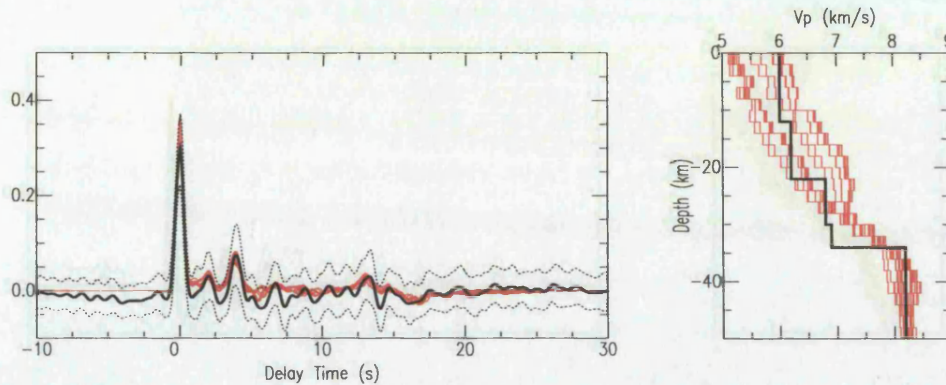
On both the ESK and ESKB receiver functions the Moho P_s phase arrives at a consistent delay time of ~ 4 s for all backazimuths. The Moho $PpPs$ phase is weaker, but is still consistently identifiable on both the broadband and short-period data at between 12 and 13s. In the tangential components, there are no consistent phases seen through all backazimuths in the short-period data. In the broadband tangential data there is a phase at between 0 and 1s that changes polarity, appearing positive between 330° - 120° and negative between 180° - 310° .

The receiver function data from ESKB and ESK have been analysed using the linear inversion scheme of Ammon et al. (1990). The models presented in receiver function analysis are normally described by the S -wave velocity. In this study the models are described by their P -wave velocity to allow direct comparison with the models from the crustal seismic refraction studies from

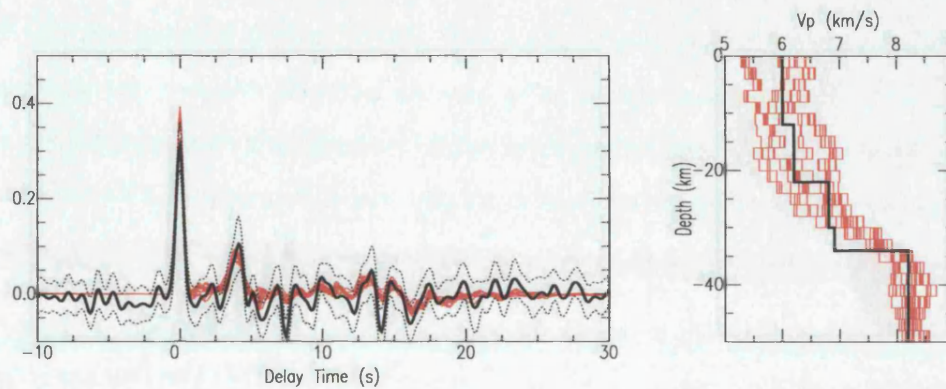
around the British Isles. The 1D forward modelling and inversion code of Ammon et al. (1990) use a default Poisson's ratio of 0.25 ($V_p/V_s = 1.73$), and therefore the S -wave model is simply linearly related to the P -wave velocity model. A stack of the radial receiver functions from all backazimuths have been inverted for ESKB and ESK to analyse the average crustal structure beneath Eskdalemuir (Figures 6.1a & 6.2a). Stacking the receiver functions from a wide range of backazimuths may result in degradation of the receiver function phases if there is any degree of 2/3D structure beneath the recording station. To examine the detail of azimuthal variation in the radial receiver functions, further inversions have been performed on stacks of data in the backazimuthal ranges 0° - 30° , 90° - 120° and 270° - 300° for both the broadband (Figures 6.1b-d) and short-period data (Figures 6.2b-d). These stacks have been limited to contain events from epicentral distances of 60 - 100° . The amplitude and timing of the receiver function phases are dependent upon the ray parameter of the source equalised events (Figure 3.7); limiting the size of the epicentral distance bin minimises the variation in receiver function due to changes in ray parameter. The standard deviation of the stacked receiver functions has been calculated from the variance of the raw receiver functions included in the stacks. 20 separate inversions have been performed on each stacked receiver function using a range of starting models that have been calculated by random perturbation of a discrete 1D velocity model, extracted from the velocity model of Barton (1992). The perturbations have been made to examine the dependence of the best fitting inversion model on the starting model. The models have been parameterised as a number of thin layers. The inversion code optimises the velocity within a fixed layer structure. Therefore, parameterising the starting model with many thin layers allows as unconstrained an inversion as possible. Each of these 20 inversions has been run for 5 iterations, using a smoothing parameter of $\sigma=0.2$. This parameter controls the tradeoff between producing a good fitting model, with the roughness of the resultant model (see Section 3.3.3, Equation 3.24). The results of the inversion investigations (e.g. Figure 6.1) show the best fitting (lowest R.M.S error) model for each of the separate 20 inversions. Where there is a wide range of models, there are a number of local minima to which the inversion routine is optimising. When the models do not vary greatly the inversion has optimised to a global minimum.

As discussed in section 3.2.3, analysing receiver functions calculated from short-period teleseismic data without considering the bandwidth of the seismograms, can result in misinterpretation of the crustal and upper mantle structure. Comparing the results of the inversions of the data recorded

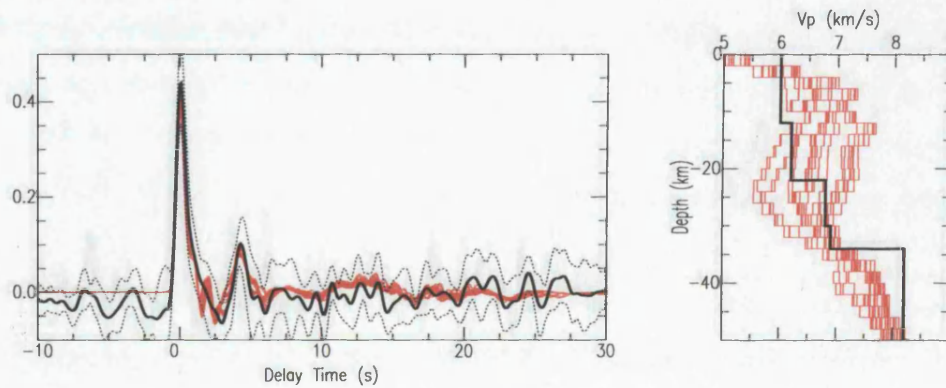
a) ALL



b) 000-030



c) 090-120



d) 270-300

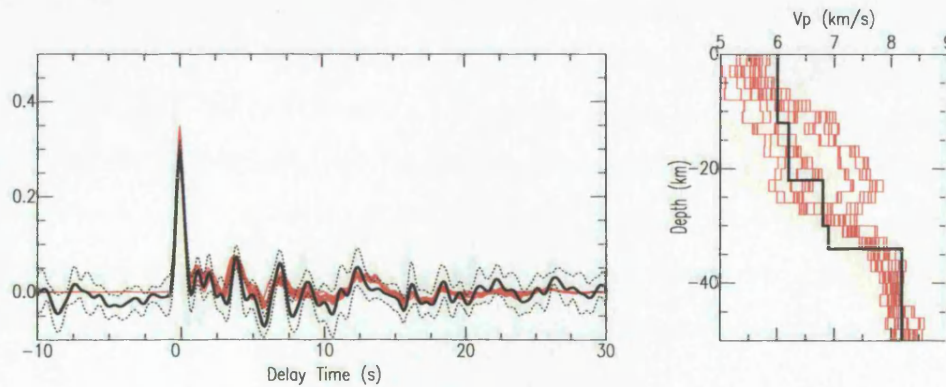
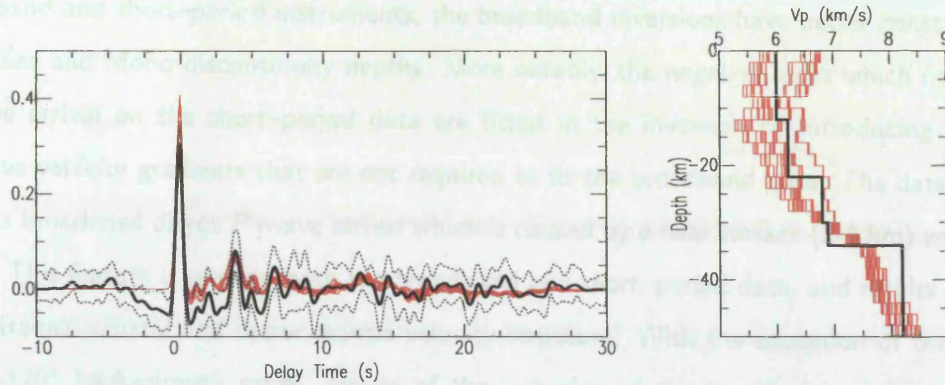
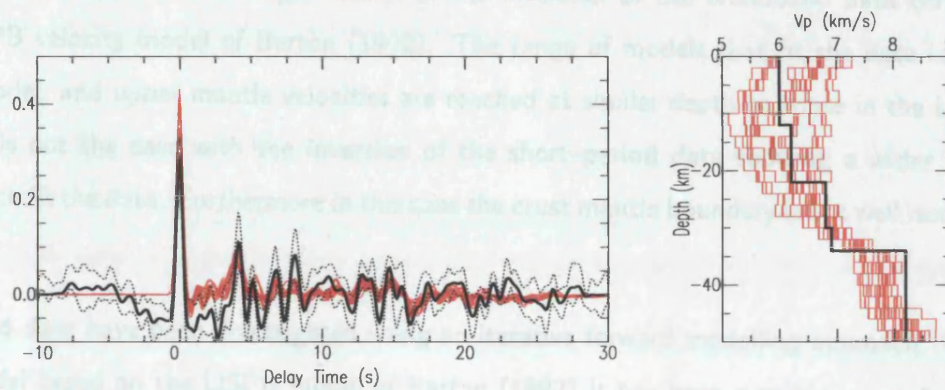


Figure 6.1: Results of inversion modelling of receiver function data from the Eskdalemuir broadband instrument (ESKB). Observed stacked receiver functions (black, $\pm 1\sigma$ dotted) and synthetic receiver functions (red). The starting model (black) and inverted models (red) plotted with each set of receiver functions; the starting model is derived from the LISPB profile (Barton, 1992). Inversion of a) receiver functions from all backazimuths, b) receiver functions from 0-030°, c) receiver functions from 090-120° and d) receiver functions from 270-300°.

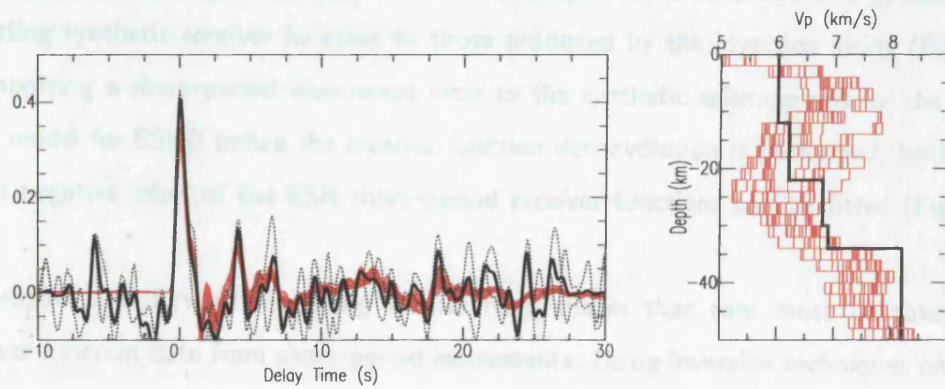
a) ALL



b) 000-030



c) 090-120



d) 270-300

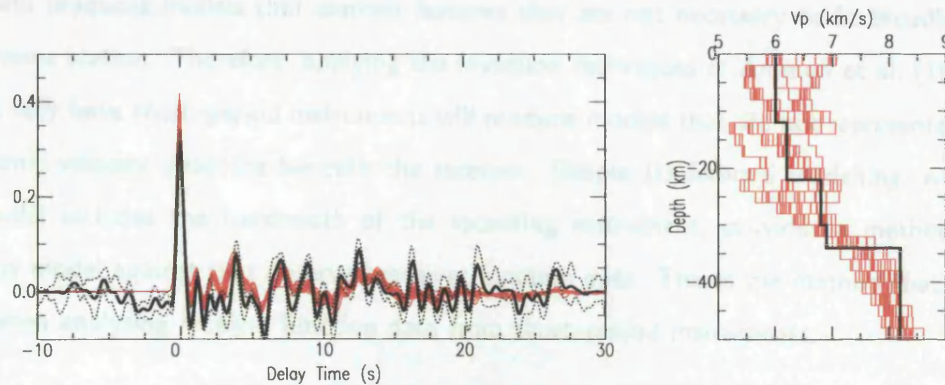


Figure 6.2: Results of inversion modelling of receiver function data from the Eskdalemuir short-period instrument (ESK). Observed stacked receiver functions (black, $\pm 1\sigma$ dotted) and synthetic receiver functions (red). The starting model (black) and inverted models (red) plotted with each set of receiver functions; the starting model is derived from the LISPB profile (Barton, 1992). Inversion of a) receiver functions from all backazimuths, b) receiver functions from 0-030°, c) receiver functions from 090-120° and d) receiver functions from 270-300°.

on the broadband and short-period instruments, the broadband inversions have better constrained mantle velocities and Moho discontinuity depths. More notably, the negative lobes which ring the direct P -wave arrival on the short-period data are fitted in the inversion by introducing upper crustal negative velocity gradients that are not required to fit the broadband data. The data from 90° - 120° has a broadened direct P -wave arrival which is caused by a near surface (2-4 km) velocity discontinuity. This feature is seen on both the broadband and short-period data, and results in the less well constrained crustal and upper mantle velocity structure. With the exception of the data from the 90° - 120° backazimuth range, results of the inversion of the broadband data correlate with the LISP model of Barton (1992). The range of models that fit the data include the LISP model, and upper mantle velocities are reached at similar depth to those in the LISP model. This is not the case with the inversion of the short-period data showing a wider range of models which fit the data. Furthermore in this case the crust mantle boundary is not well resolved.

The broadband data have been investigated using an iterative forward modelling approach. Using a starting model based on the LISP model of Barton (1992) it has been possible to produce an equally well fitting synthetic receiver function to those produced by the inversion study (Figures 6.3a&b). By applying a short-period instrument filter to the synthetic seismograms for the best fitting forward model for ESK before the receiver function deconvolution is performed, both the phases and the negative lobes of the ESK short-period receiver functions can be fitted (Figures 6.3c&d).

The 1D inversion and forward modelling studies have shown that care must be taken in analysing receiver function data from short-period instruments. Using inversion techniques on the short-period data produces models that contain features that are not necessary to fit broadband data from the same station. Therefore, applying the inversion techniques of Ammon et al. (1990) at stations that only have short-period instruments will produce models that are not representative of the true seismic velocity structure beneath the receiver. Simple 1D forward modelling, where the forward model includes the bandwidth of the recording instrument, provides a method of testing a velocity model against that observed receiver function data. This is the method that has been adopted when analysing receiver function data from short-period instruments.

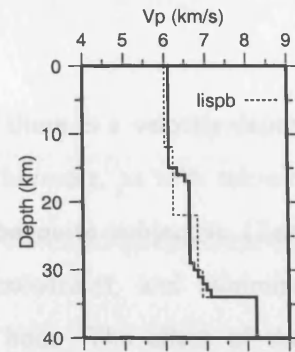
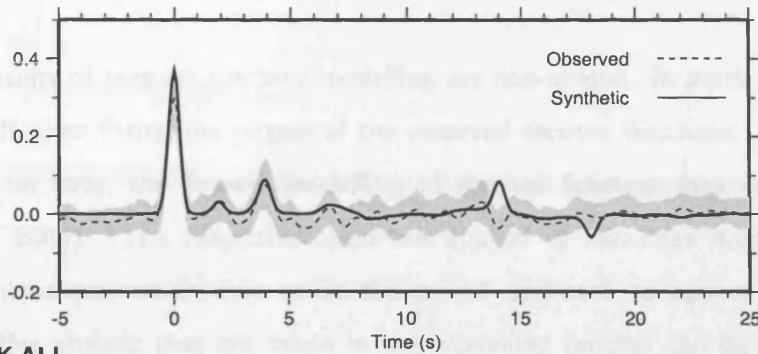
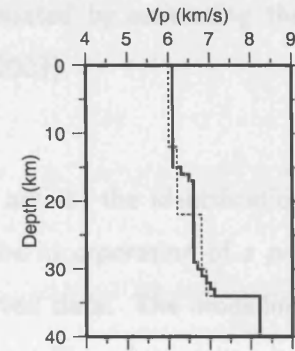
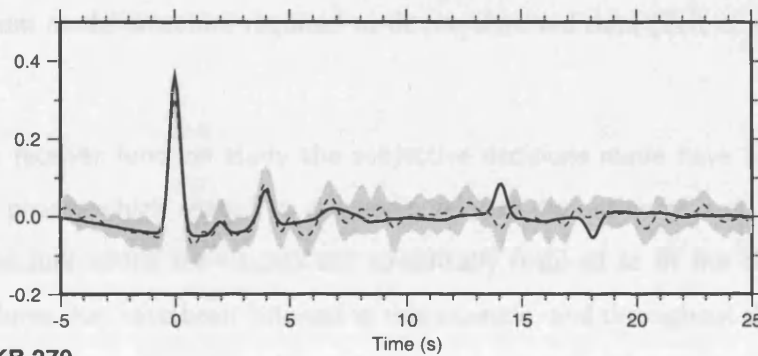
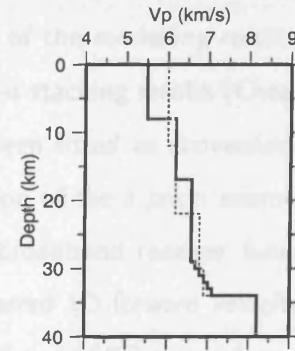
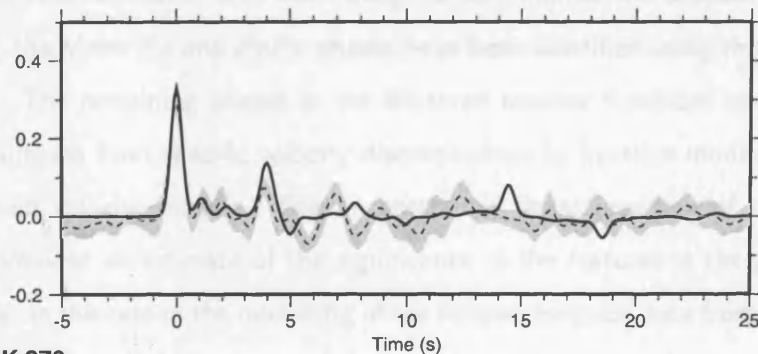
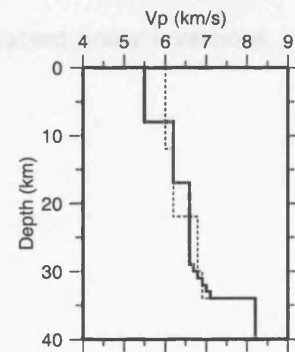
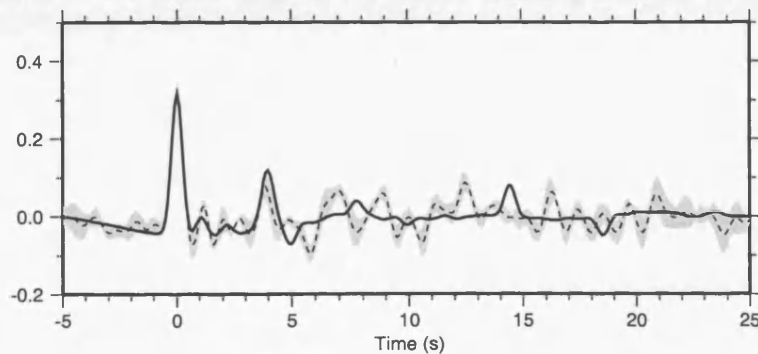
a) ESKB ALL**b) ESK ALL****c) ESKB 270****d) ESK 270**

Figure 6.3: Results of 1D forward receiver function data from the ESKB and ESK. Observed stacked receiver functions (dotted, $\pm 1\sigma$ grey shaded) and synthetic receiver functions (black). The LISPb starting model (dotted) as used in the inversion modelling and the preferred 1D model (black) are plotted with each set of receiver functions. Modelling a) ESKB receiver functions from all backazimuths, b) ESK receiver functions from all backazimuths, c) ESKB receiver functions from 270-300° and d) ESK receiver functions from 270-300°.

6.2.2 Modelling Strategy

The results of receiver function modelling are non-unique. In particular there is a velocity-depth tradeoff when fitting the phases of the observed receiver functions. Furthermore, as with seismic refraction data, the forward modelling of receiver function data can be quite subjective (Zelt et al., 2003). The subjective steps are applied to maximise model constraint, and minimise non-uniqueness, which may cause the overall approach to appear ad hoc. The effect of the subjective choices that are made in the modelling process can be evaluated by estimating the minimum model structure required to fit the observed data (Zelt et al., 2003).

In this receiver function study the subjective decisions made have been are; 1) the identification of the phases which related to a specific velocity discontinuity, and 2) the incorporation of a *a priori* structure within the models not specifically required to fit the observed data. The modelling procedures that have been followed in this example, and throughout the modelling of the remaining receiver function data, have been designed to minimise the subjectivity of the modelling results. Firstly, the Moho P_s and $PpPs$ phases have been identified using the $H-\kappa$ stacking results (Chapter 5). The remaining phases in the observed receiver functions have been fitted as conversions and multiples from specific velocity discontinuities by iterative modification of the *a priori* seismic refraction velocity models. The unconstrained linear inversion of the broadband receiver functions provides an estimate of the significance of the features in the preferred 1D forward velocity models. In the case of the modelling of the receiver function data from ESK and ESKB the preferred 1D models lie within the range of the models resulting from the unconstrained linear inversions.

6.3 Northwest Highlands

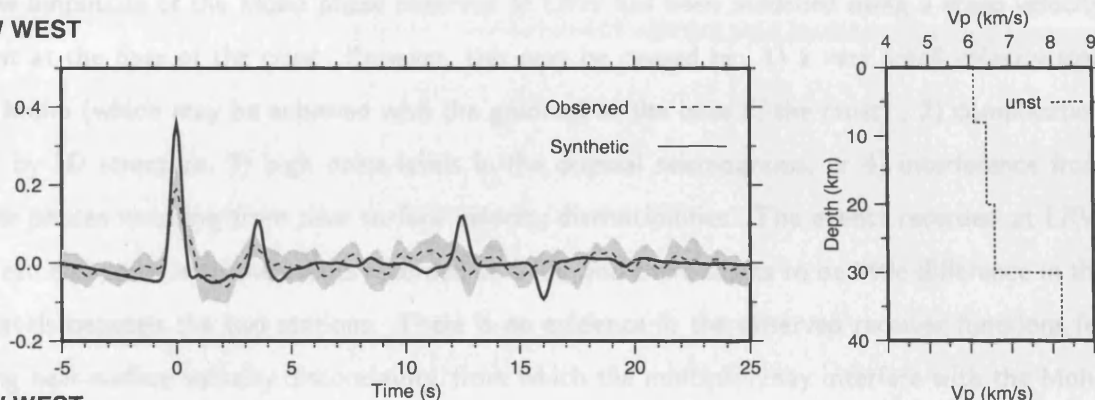
6.3.1 Introduction

The lithospheric structure of the Northwest Highlands has been the subject of numerous geophysical investigations, primarily from the BIRPS experiments (e.g. Smythe et al., 1982; Brewer et al., 1983; Klemperer and Hobbs, 1991; McGeary, 1989; McBride et al., 1995; Price and Morgan, 2000; Morgan et al., 2000). These profiles include DRUM, WINCH, GRID, SHET and MOIST. The receiver function data cover 6 stations, LRW, ORE, MCD, RRR, KPL and KARB (Figure 4.1). These stations are located primarily along the coast of northwestern Scotland, and provide the opportunity to constrain the onshore variation of the seismic models that have resulted from the BIRPS experiments. In particular the BIRPS profiles reveal upper mantle structures, which are found from the base of the crust to around 80 km depth and for which there have been various geological interpretations (Section 2.3.3). The onshore continuity of these structures is unconstrained, although the LISP profile has identified upper mantle reflectors in northern Scotland at similar depths to those identified by BIRPS (Bamford et al., 1978; Barton, 1992).

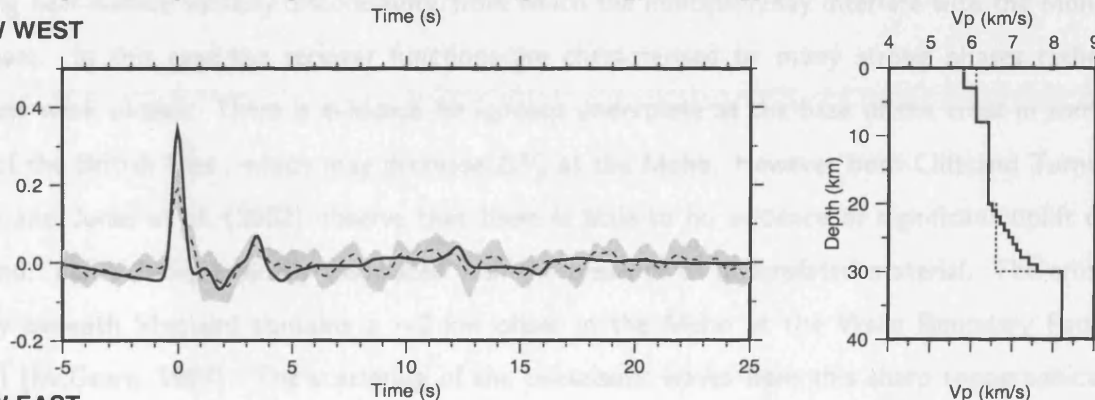
6.3.2 Lerwick (LRW)

LRW is located at the south of Shetland (Figure 4.1), and is situated on Devonian Old Red sandstone, overlying Dalradian schist. The stacked radial and tangential receiver functions contain only a few weak phases (Figure A.2). These data are unlike the receiver functions associated with typical crustal velocity structures from around the British Isles (Figure 3.3). The tangential component receiver functions show no evidence for any significant azimuthal variation. The data from 210° to 360° have a slightly higher amplitude phase at a time corresponding to the Moho P_s phase calculated from the $H-\kappa$ stacking results. A forward model based on the seismic reflection structure from McGeary (1989) and P -wave velocities from Morgan et al. (2000) show that the weak phase at ~ 3.5 s corresponds to the time of the Moho P_s phase (Figure 6.4a). The amplitude of the observed phase is however much lower than that calculated from the *a priori* velocity model. The amplitude of the P_s phase has been fitted by reducing ΔV_p at the Moho by introducing a velocity gradient into the lower crust (Figure 6.4b). The lower amplitude of the observed Moho P_s phase in the data from 0° to 150° can also only be matched within

a) LRW WEST



b) LRW WEST



c) LRW EAST

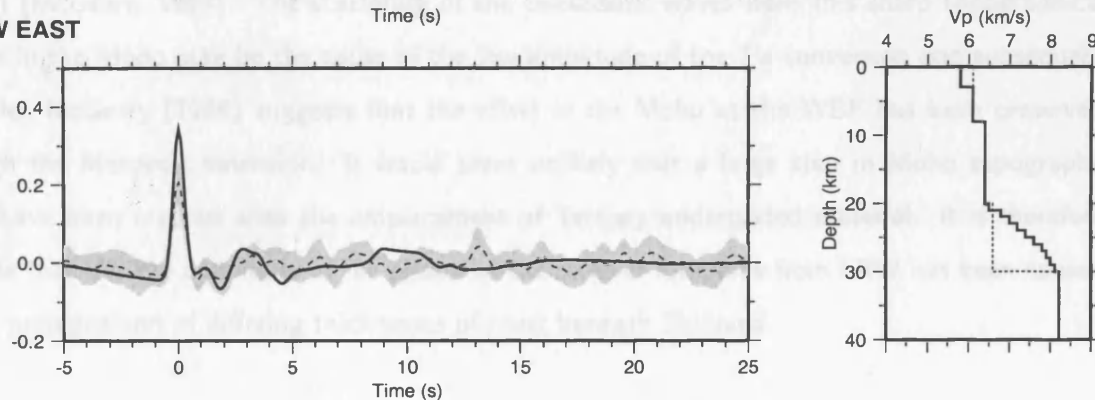


Figure 6.4: Results of 1D forward modelling of receiver function data from LRW for; a) the westerly receiver functions compared with the crustal model from the UNST profile (McGeary, 1989), b) the westerly data modelled with a velocity gradient at the base of the crust, c) the easterly data modelled with a velocity gradient at the base of the crust. The data are presented in the same way as Figure 6.3.

the *a priori* structures by introducing a velocity gradient at the Moho, removing the step in the velocity depth function (Figure 6.4c).

The low amplitude of the Moho phase observed at LRW has been modelled using a steep velocity gradient at the base of the crust. However, this may be caused by: 1) a very small velocity step at the Moho (which may be achieved with the gradient at the base of the crust), 2) complication caused by 3D structure, 3) high noise levels in the original seismograms, or 4) interference from multiple phases resulting from near surface velocity discontinuities. The events recorded at LRW have been compared with the events recorded at ORE, and there seems to be little difference in the noise levels between the two stations. There is no evidence in the observed receiver functions for a strong near-surface velocity discontinuity, from which the multiples may interfere with the Moho P_s phase. In this case the receiver functions are characterised by many strong phases rather than few weak phases. There is evidence for igneous underplate at the base of the crust in some areas of the British Isles, which may decrease ΔV_p at the Moho. However both Clift and Turner (1998) and Jones et al. (2002) observe that there is little to no evidence of significant uplift of Shetland, 'uplift' being typically associated with the presence of underplated material. The crust directly beneath Shetland contains a ~ 2 km offset in the Moho at the Walls Boundary Fault (WBF) (McGeary, 1989). The scattering of the teleseismic waves from this sharp topographical feature in the Moho may be the cause of the low amplitude of the P_s conversion and subsequent multiple. McGeary (1989) suggests that the offset in the Moho at the WBF has been preserved through the Mesozoic extension. It would seem unlikely that a large step in Moho topography could have been created after the emplacement of Tertiary underplated material. It is therefore possible that the low amplitude of the phases of the receiver functions from LRW has been caused by the juxtaposition of differing thicknesses of crust beneath Shetland.

6.3.3 Reay (ORE)

The station ORE lies on the north coast of Scotland (Figure 4.1) and is particularly close to the eastern end of the MOIST, DRUM and GRID BIRPS seismic surveys. The station is located directly upon the crystalline basement of the Moine sequence, close to the margin of the Devonian sediments of the Shetland-Orkney Platform. The results of the $H-\kappa$ modelling of the receiver

function data for ORE reveal a crustal thickness of 26.7 km and a V_p/V_s of 1.73, using an average crustal V_p of 6.3 km s^{-1} (Figure A.4). The Moho Ps conversion corresponding with the $H-\kappa$ stacking model can clearly be seen at 3s across all backazimuths in the radial receiver function section (Figure A.4). The most notable feature of the data after the Moho Ps and associated $PpPs$ phases, is a non-continuous phase that can be seen at 5s. This phase is limited in azimuth distribution; it is clearly visible between $0-90^\circ$ and is less clear between $240-360^\circ$.

The seismic velocity and amplitude modelling of Morgan et al. (2000) and Price and Morgan (2000) provides a detailed velocity model to test against the receiver function data from ORE. Their crustal velocity model for the area is consistent with that from the LISPb experiment (Barton, 1992) and $H-\kappa$ stacking, giving a crustal thickness of ~ 27 km. The LISPb model does include a significant velocity gradient at the base of the crust that is not present in the model for the JUNE92 profile. The synthetic receiver functions generated from an initial velocity depth model based upon the Morgan et al. (2000) P -wave velocity model confirm the phases at ~ 3 s and ~ 11 s are in fact the Moho Ps and $PpPs$ phases (Figure 6.5a). The amplitude of the Moho phase requires a velocity discontinuity of $6.6-8.2 \text{ km s}^{-1}$ ($\Delta V_p = 1.6 \text{ km s}^{-1}$). This corresponds with the velocities in the JUNE92 model, and is inconsistent with the velocity gradient at the base of the crust in Barton's LISPb model. The velocity model from Morgan et al. (2000) shows the P -wave velocity beneath the sub-crustal W-reflector is $8.5 \pm 0.05 \text{ km s}^{-1}$, increasing from the normal mantle velocity of 8.2 km s^{-1} . When the W-reflector is included in the velocity model for the synthetic receiver functions it is clear that the amplitude of the Ps conversion caused by the velocity discontinuity is not large enough to fit the 5s phase (Figure 6.5a). The receiver function data from $180-300^\circ$ backazimuth show no clear phase at 5s. However, they do contain a phase in the first 2s following the direct P arrival. This is caused by a near surface velocity discontinuity (Figure 6.5b). The occurrence of the phase seems unusual because the station is located directly upon the Moinian basement, and such near surface velocity discontinuities are usually associated with the presence of a near surface sedimentary layer. The multiple phases from the near surface discontinuity do not arrive at a time corresponding to the post Moho Ps phase at 5s.

The phase that is present at ~ 5 s has two possible sources; 1) it is a primary Ps conversion that originates from within the upper mantle, or, 2) it is a multiple from a discontinuity from within the crust. Simply calculating the depth of the arrival from its traveltime using an average velocity

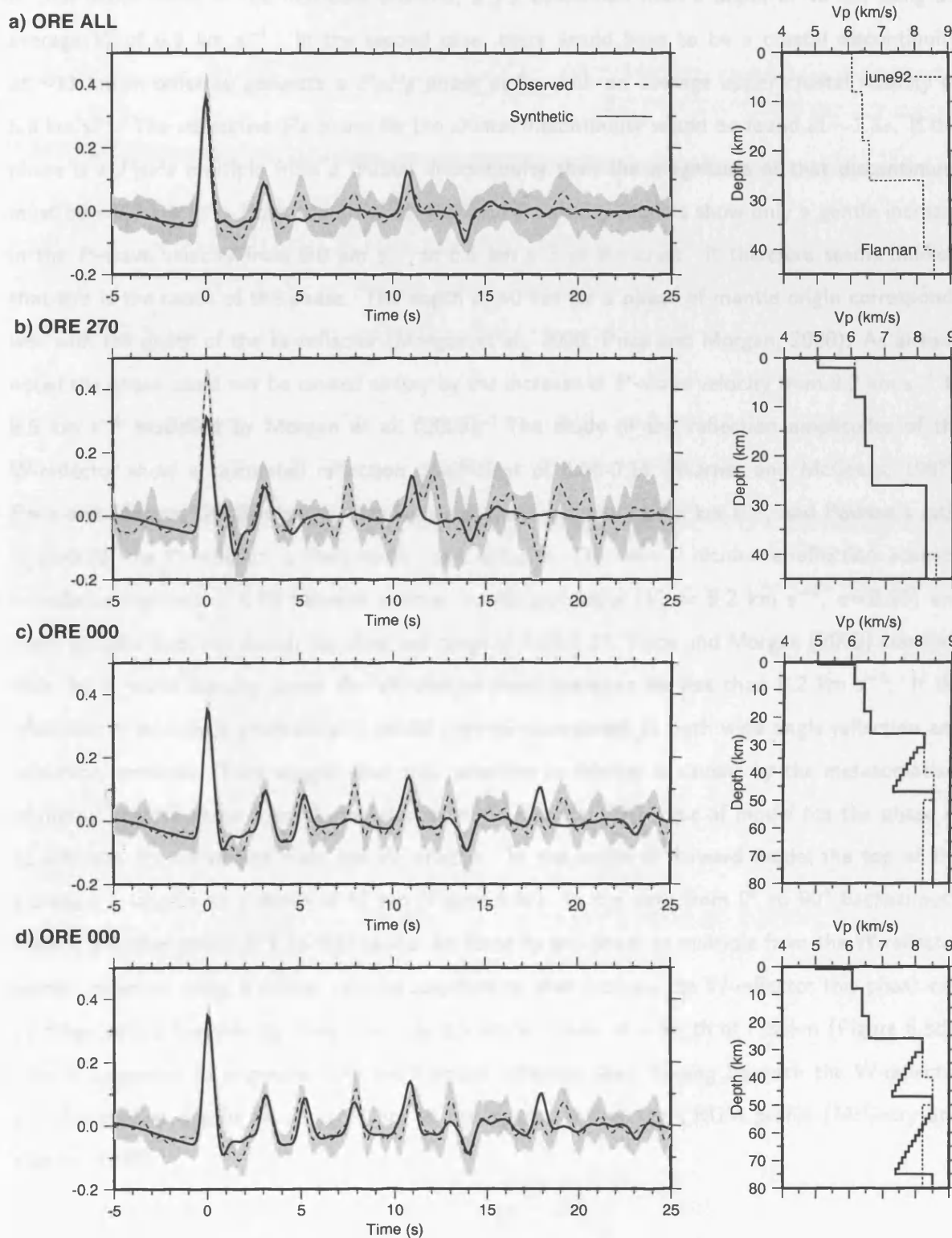


Figure 6.5: The 1D forward modelling results from ORE for: a) a stack of a the receiver function data. b) the data from 270°-300°. c) the data from 0°-30°. d) the data from 0°-30° with a model with two mantle discontinuities. The data are presented in the same way as Figure 6.3.

to that depth yields in the first case scenario, a Ps conversion from a depth of 40 km using an average V_p of 6.9 km s^{-1} . In the second case, there would have to be a crustal discontinuity at $\sim 11 \text{ km}$ in order to generate a $PpPs$ phase at 5s, with an average upper crustal velocity of 5.8 km s^{-1} . The respective Ps phase for the crustal discontinuity would be found at $\sim 1.5 \text{ s}$. If the phase is a $PpPs$ multiple from a crustal discontinuity then the magnitude of that discontinuity must be $\sim 0.8 \text{ km s}^{-1}$. Both the JUNE92 and LISPB velocity models show only a gentle increase in the P -wave velocity from 6.0 km s^{-1} to 6.6 km s^{-1} in the crust. It therefore seems unlikely that this is the cause of the phase. The depth of 40 km for a phase of mantle origin corresponds well with the depth of the W-reflector (Morgan et al., 2000; Price and Morgan, 2000). As already noted the phase could not be caused simply by the increase in P -wave velocity from 8.2 km s^{-1} to 8.5 km s^{-1} modelled by Morgan et al. (2000). The study of the reflection amplitudes of the W-reflector show a calculated reflection co-efficient of 0.08-0.14 (Warner and McGeary, 1987). Price and Morgan (2000) observe that with a P -wave velocity of 8.5 km s^{-1} , and Poisson's ratio of $\sigma=0.29$, the W-reflector is likely to be mafic eclogite. The normal incidence reflection acoustic impedance contrast of 0.06 between normal mantle peridotite ($V_p = 8.2 \text{ km s}^{-1}$, $\sigma=0.25$) and mafic eclogite does not match the observed range of 0.08-0.14. Price and Morgan (2000) conclude that the P -wave velocity above the W-reflector must therefore be less than 8.2 km s^{-1} . If the reduction in velocity is gradational it would then be transparent to both wide angle reflection and refraction methods. They suggest that this reduction in velocity is caused by the metasomatism of normal mantle above a subducted slab of mafic eclogite. This type of model fits the phase at 5s with the Ps conversion from the W-reflector. In the preferred forward model the top of the 8.5 km s^{-1} layer is at a depth of 47 km (Figure 6.5c). In the data from 0° to 90° backazimuth there is a further phase at 7.5s that cannot be fitted by any phase or multiple from the W-reflector model. However using a similar velocity structure to that used on the W-reflector this phase can be fitted with a low-velocity zone above an 8.5 km s^{-1} layer at a depth of $\sim 80 \text{ km}$ (Figure 6.5d). This is suggested to originate from the Flannan reflector, seen dipping beneath the W-reflector and detected at depths of up to 80 km at the eastern end of the DRUM profile (McGeary and Warner, 1985).

6.3.4 Rubha Reidh (RRR-RRRB)

Rubha Reidh is sited upon Torridonian sandstone, overlying Lewisian gneiss on the northwest coast of Scotland (Figure 4.1). The stacked radial and tangential receiver function data for both the broadband and short-period instruments show clear Moho P_s and P_pP_s phases (Figures A.6 & A.8). The $H-\kappa$ stacks of the data reveal a crust of ~ 24 km using a stacking velocity of 6.3 km s^{-1} . The Moho P_s and P_pP_s phases corresponding to this crustal model arrive at ~ 3 s and ~ 10 s respectively. The most notable phase after the Moho P_s and subsequent multiples, arrives at ~ 4.5 s and can be most clearly identified in the radial receiver functions from between 210° and 300° backazimuth on both the broadband and short-period instruments. There must be some departure from the 1D case as this phase is clearly identified in the tangential receiver functions. The amplitude of the Moho P_s phase also varies, showing the lowest amplitude between 0° and 90° backazimuth. Stacks of the radial receiver functions from RRRB have been inverted for the backazimuth ranges 0° - 360° , 0° - 30° and 270° - 300° (Figures 6.6a, b & c respectively). The inversions have been performed using a starting model extracted from the LISPB profile (Barton, 1992). The inversion produces a crust consistent with the input velocity model, showing a steep velocity gradient at the base of the crust. The phase at ~ 4.5 s on the 270° stack is fitted with a complication of the velocity model beneath the Moho.

Stacks of the broadband and short-period data have been investigated using 1D forward modelling (Figures 6.7 & 6.8). The stacks of all the radial receiver function data (0° - 360°) (Figures 6.7a & 6.8a) and the northerly receiver function data (0° - 30°) (Figures 6.7b & 6.8b) require a high velocity gradient at the base of the crust. This has been introduced to reduce ΔV_p at the Moho, and therefore the amplitude of the Moho P_s phase. The reduction in ΔV_p could be produced by a lower velocity gradient through a greater thickness of crust. The reduction in ΔV_p could also be produced by a series of larger steps in the velocity-depth function, but there is no evidence to support this in the observed receiver functions. The velocity gradient at the base of the crust is consistent with the range of models produced by the inversion of the broadband data. In the stacks of the westerly receiver function data the amplitude of the Moho P_s phase is greater. A simple two layer model was sufficient to model the Moho P_s phase. The ~ 4.5 s phase has been modelled using a sub-crustal low-velocity-zone (Figures 6.7c,d & 6.8c,d). The distribution of the sub-crustal Flannan reflector has been mapped by a number of the BIRPS profiles along the

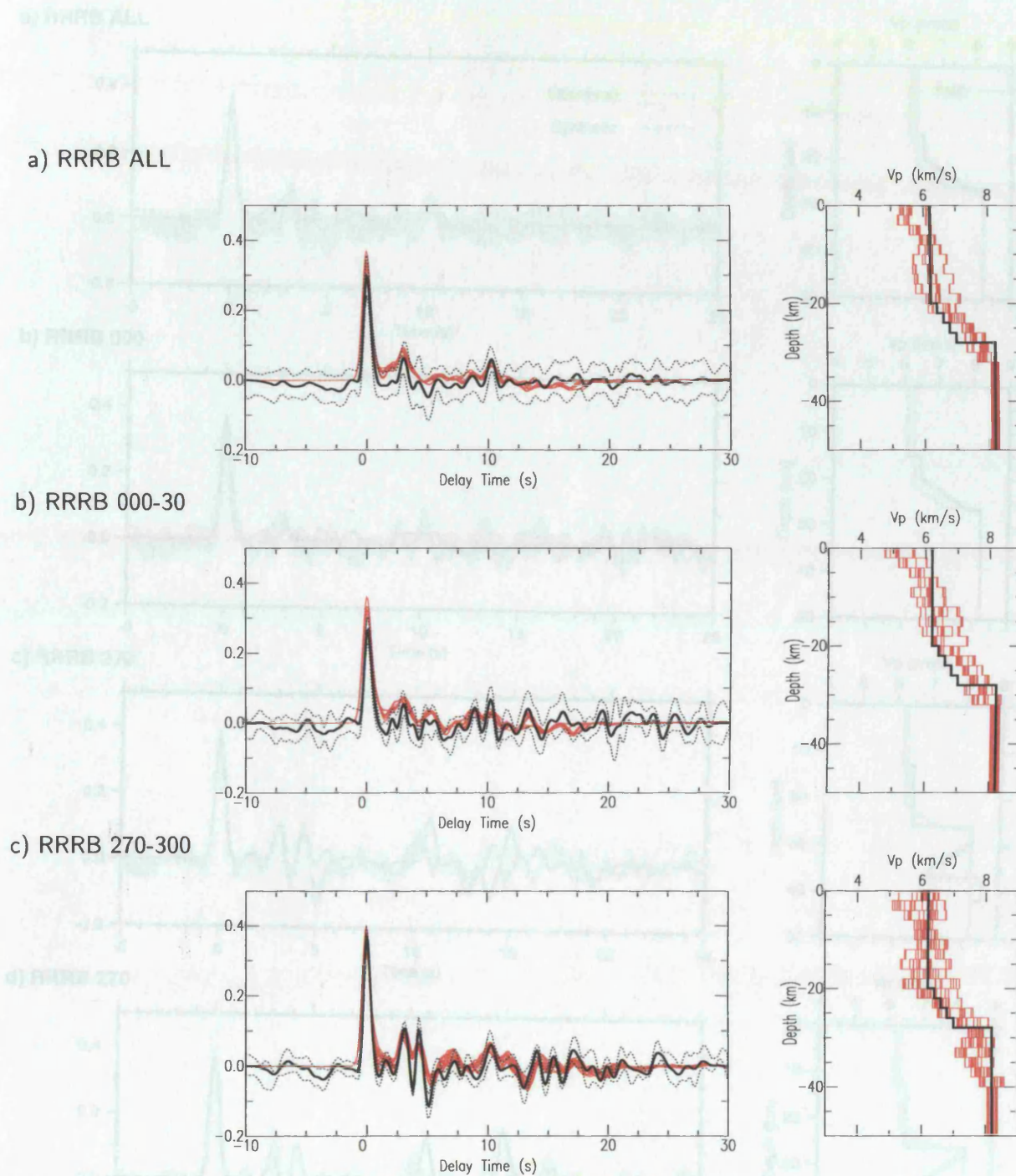


Figure 6.6: Results of 1D inversion of receiver function data from RRRB between; a) 0°-360°, b) 0°-30° and 270°-300°. The data are presented in the same way as Figure 6.1.

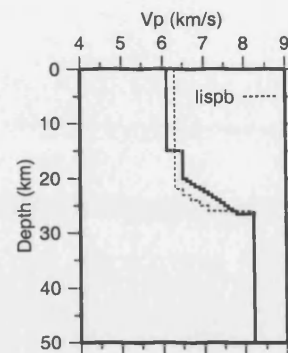
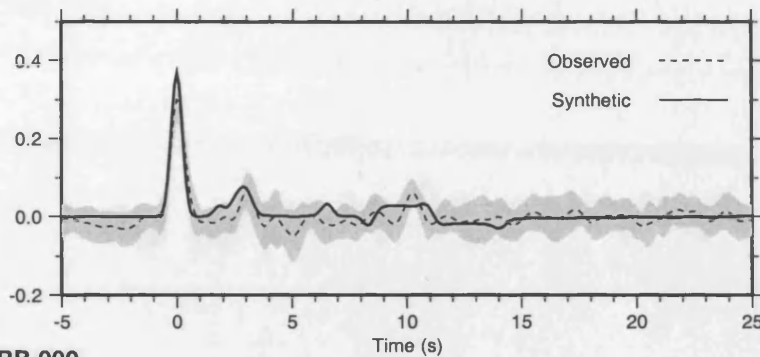
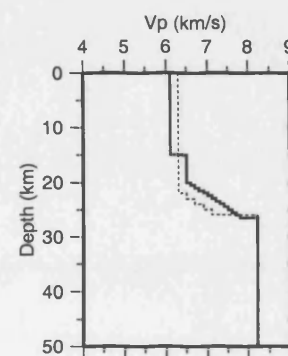
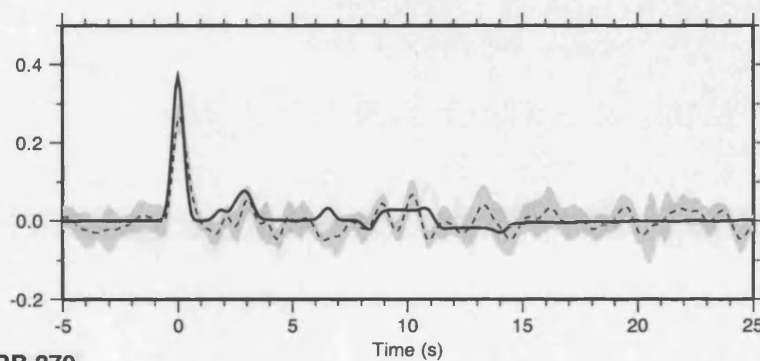
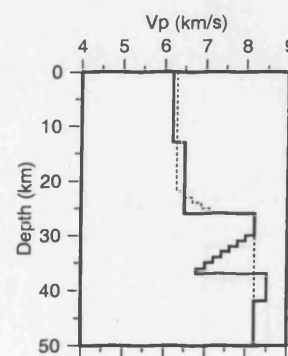
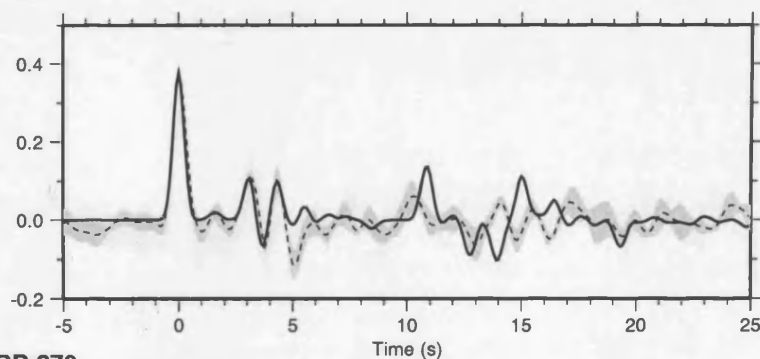
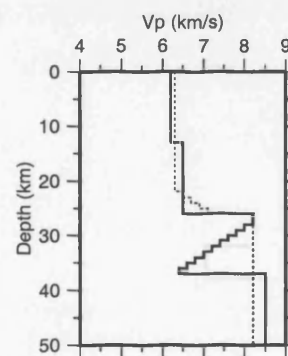
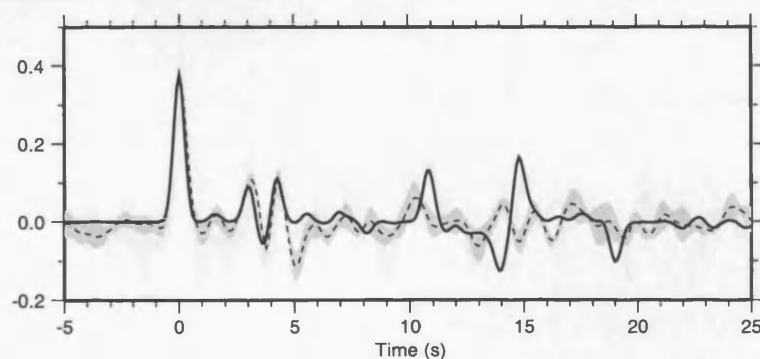
a) RRRB ALL**b) RRRB 000****c) RRRB 270****d) RRRB 270**

Figure 6.7: Results of 1D forward modelling of receiver function data from RRRB for; a) the stack of all the receiver function data, b) the data from 0° - 30° , c) the data from 270° - 300° and d) the data from 270° - 300° . The data are presented in the same way as Figure 6.3.

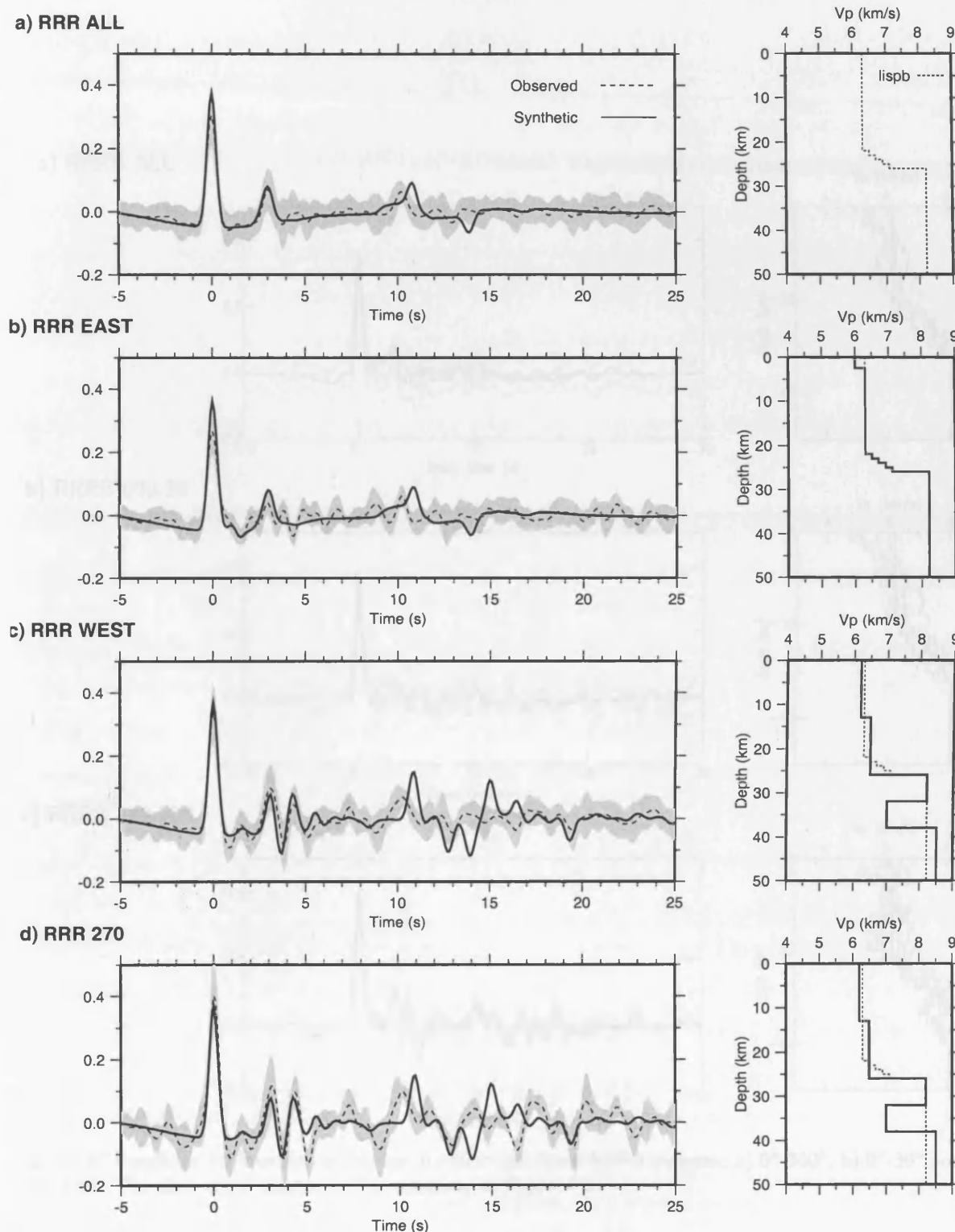
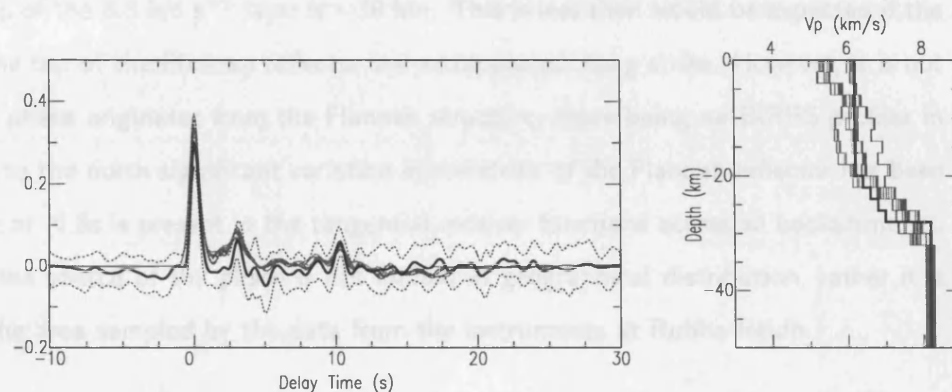


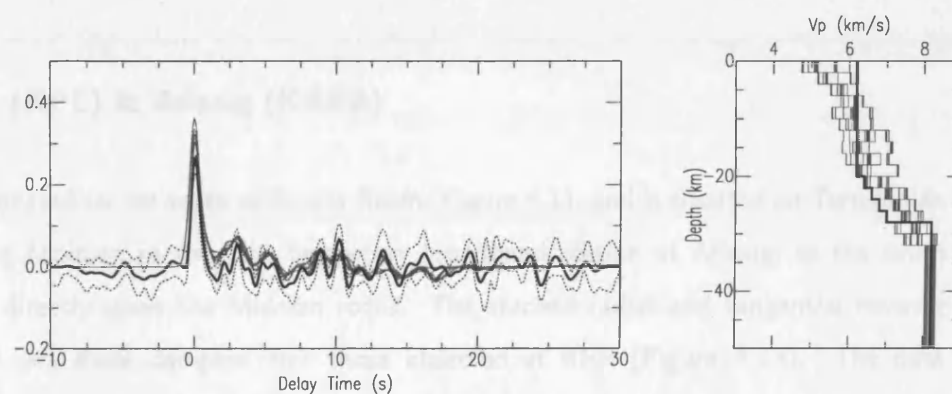
Figure 6.8: Results of 1D forward modelling of receiver function data from RRR for; a) the stack of all the receiver function data, b) the westerly receiver function data, c) the easterly receiver function data and d) the data from 270°-300°. The data are presented in the same way as Figure 6.3.

Northwest coast of Scotland. A depth migrated display of this data reveals that the Flannan reflector beneath the Moray Firth is located at depths between 30 and 42 km, dipping to the west (Foster and Park, 1999; McBride et al., 1999). The low velocity zone is the velocity anomaly for the 270°-300° receiver function is consistent with the observations of Pratt and Morgan (2000).

a) RRRB ALL



b) RRRB 000-30



c) RRRB 270-300

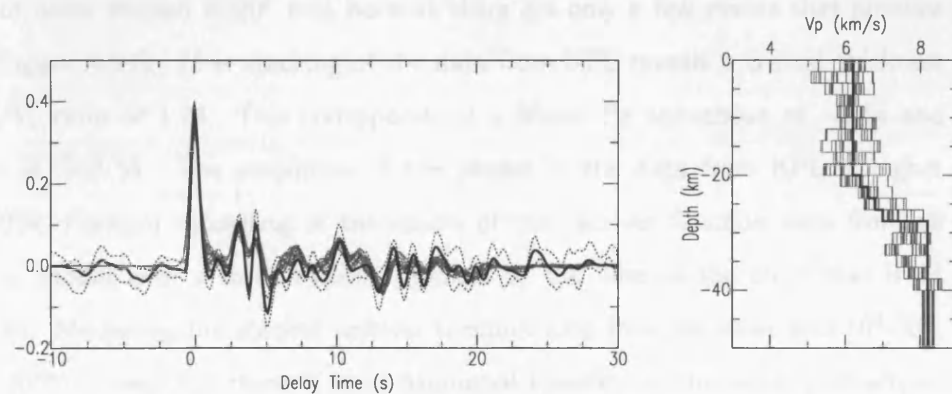


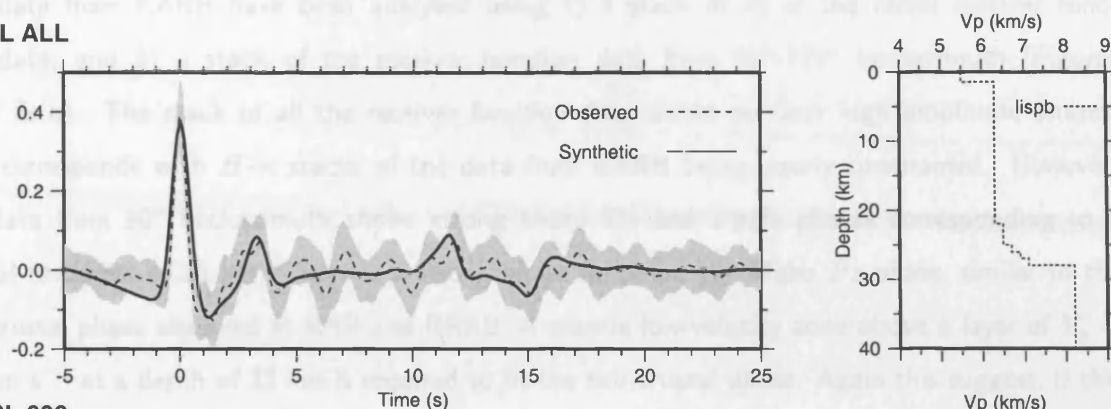
Figure 6.6: Results of 1D inversion of receiver function data from RRRB between; a) 0°-360°, b) 0°-30° and 270°-300°. The data are presented in the same way as Figure 6.1.

northwest coast of Scotland. A depth migrated compilation of this data reveals that the Flannan reflector beneath the Minch basin is found at depths between 30 and 42 km, dipping to the east (Snyder and Flack, 1990; McBride et al., 1995). The low velocity zone in the velocity models for the 270°-300° receiver functions is consistent with the observations of Price and Morgan (2000). The depth to the top of the 8.5 km s⁻¹ layer is ~38 km. This is less than would be expected if the depth contours to the top of the Flannan reflector are extrapolated along strike. However, it is not impossible that this phase originates from the Flannan structure, there being no BIRPS profiles in this region. Further to the north significant variation in the strike of the Flannan reflector has been mapped. The phase at 4.5s is present in the tangential receiver functions across all backazimuths. This suggests that the source of the phase is not limited in geographical distribution, rather it is found under all of the area sampled by the data from the instruments at Rubha Reidh.

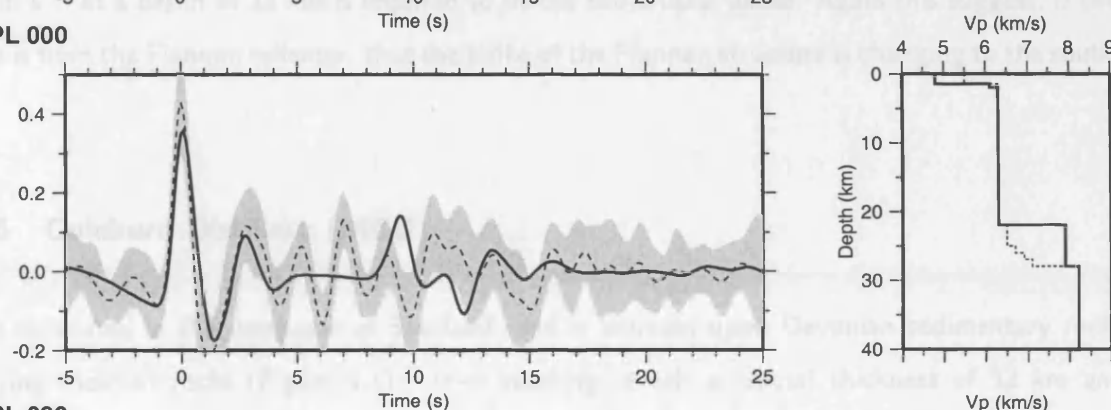
6.3.5 Plockton (KPL) & Arisaig (KARB)

Plockton (KPL) is located to the south of Rubha Reidh (Figure 4.1), and is situated on Torridonian Sandstone, overlying Moinian rocks. The temporary broadband station at Arisaig, to the south of KPL, is located directly upon the Moinian rocks. The stacked radial and tangential receiver functions from KPL are more complex than those observed at RRR (Figure A.13). The data from KARB have not been stacked in 20° bins because there are only a few events that produce receiver functions (Figure A.11). $H-\kappa$ stacking of the data from KPL reveals a crustal thickness of ~28 km and V_p/V_s ratio of 1.74. This corresponds to a Moho P_s conversion at ~3.5s and a $PpPs$ conversion at ~11.5s. The amplitude of the phases in the data from KPL is higher than observed at RRR. Forward modelling of the stacks of the receiver function data from all backazimuths gives a model with a steep velocity gradient at the base of the crust that is 28 km thick (Figure 6.9). Modelling the stacked receiver function data from narrower bins (0°-30°, 90°-120° and 270°-300°) shows that there is some azimuthal variation in the velocity structure. The data from 90° shows a strong Moho P_s conversion, whereas the Moho P_s conversion from 0° is much more complicated. The data from 270° requires a similar model to the stack of all of the data from KPL. As for the data from RRR the modelling suggests no departure from the 1D model.

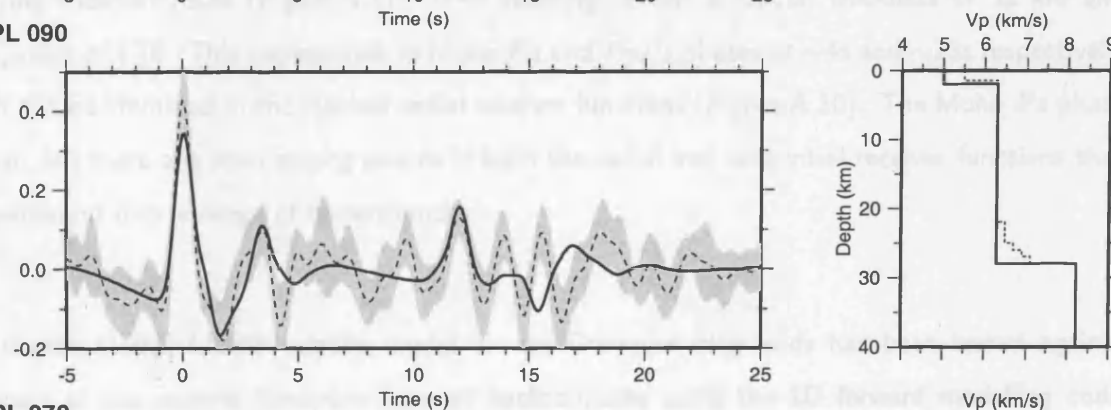
a) KPL ALL



b) KPL 000



c) KPL 090



d) KPL 270

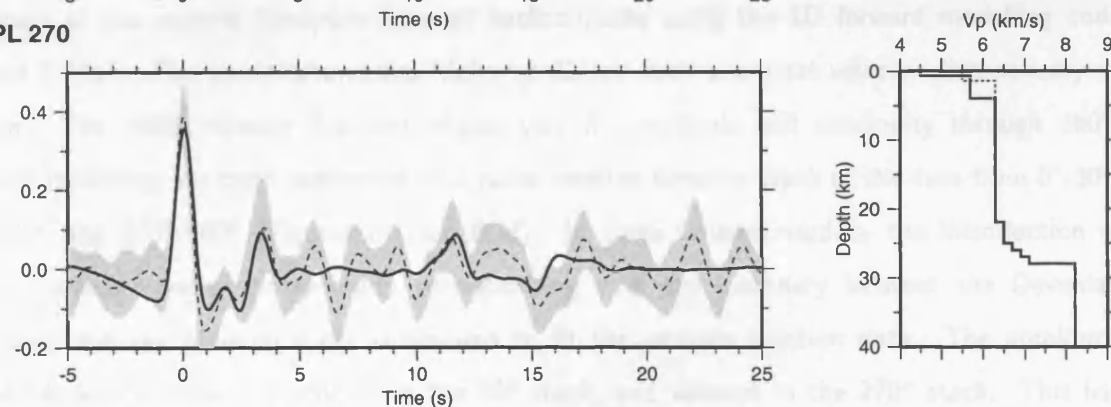


Figure 6.9: Results of 1D forward modelling of receiver function data from KPL for; a) a stack of all the receiver functions data, b) the data from 0° - 30° , c) the data from 90° - 120° and d) the data from 270° - 300° . The data are presented in the same way as Figure 6.3.

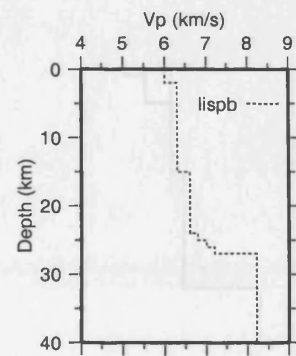
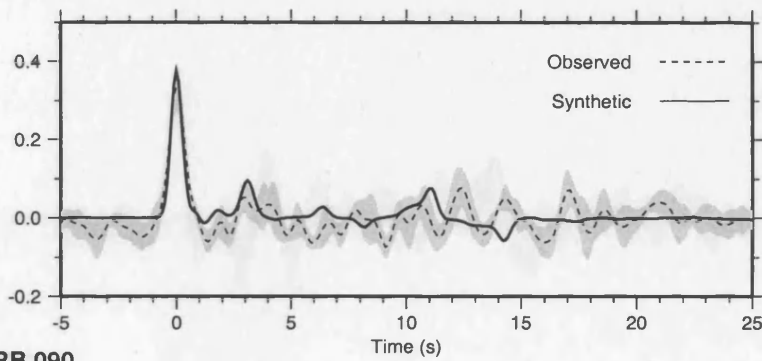
The data from KARB have been analysed using 1) a stack of all of the radial receiver function data, and 2) a stack of the receiver function data from 90°-120° backazimuth (Figures 6.10a & b). The stack of all the receiver function data shows no clear high amplitude phases. This corresponds with $H-\kappa$ stacks of the data from KARB being poorly constrained. However, the data from 90° backazimuth shows strong Moho P_s and PpP_s phases corresponding to a crustal thickness of 25 km. There is a strong phase following the Moho P_s phase, similar to the sub-crustal phase observed at RRR and RRRB. A mantle low-velocity zone above a layer of $V_p = 8.5 \text{ km s}^{-1}$ at a depth of 33 km is required to fit the sub-crustal phase. Again this suggest, if this phase is from the Flannan reflector, that the strike of the Flannan structure is changing to the south.

6.3.6 Coleburn Distillery (MCD)

MCD is located in the northeast of Scotland, and is situated upon Devonian sedimentary rocks overlying Moinian rocks (Figure 4.1). $H-\kappa$ stacking reveals a crustal thickness of 32 km and V_p/V_s ratio of 1.76. This corresponds to Moho P_s and PpP_s phases at $\sim 4\text{s}$ and $\sim 13\text{s}$ respectively, which can be identified in the stacked radial receiver functions (Figure A.10). The Moho P_s phase is clear, but there are other strong phases in both the radial and tangential receiver functions that are consistent over a range of backazimuths.

The Barton (1992) LISPB velocity model for the Grampian Highlands has been tested against the stack of the receiver functions from all backazimuths using the 1D forward modelling code (Figure 6.11a). The model shows the Moho at 32 km with a crustal velocity discontinuity at 18 km. The radial receiver function phases vary in amplitude and continuity through 360°. Forward modelling has been performed on a radial receiver function stack of the data from 0°-30°, 90°-120° and 270°-300° (Figures 6.11b,c & d). In these forward models, the introduction of a near surface velocity discontinuity corresponding with the boundary between the Devonian sediments and the Moinian rocks is required to fit the receiver function data. The amplitude of the Moho P_s phase is strongest in the 90° stack, and weakest in the 270° stack. This has been modelled by variations in ΔV_p at the crust mantle boundary, but it is more likely that the variation is caused by departure from the 1D case, for example dipping structure or seismic velocity anisotropy. The data from 90° contains a strong negative-positive phase after the

a) KARB ALL



b) KARB 090

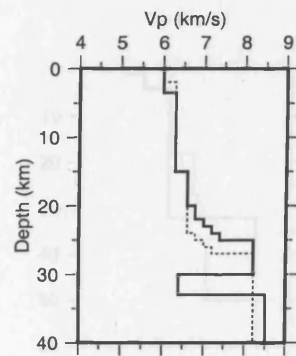
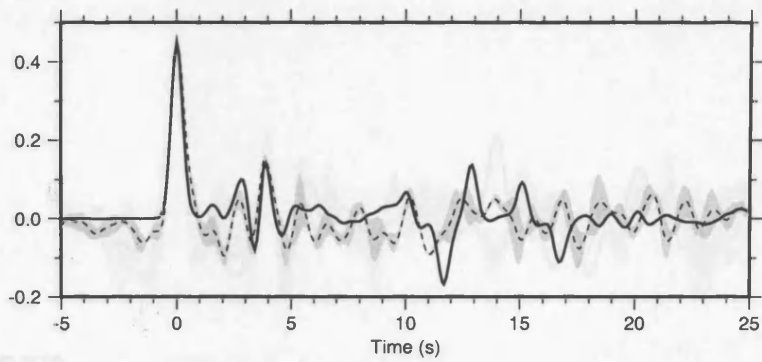


Figure 6.10: Results of 1D forward modelling of receiver function data from KARB for; a) a stack of all of the receiver function data, and b) a stack of the data from 90° - 120° . The data are presented in the same way as Figure 6.3.

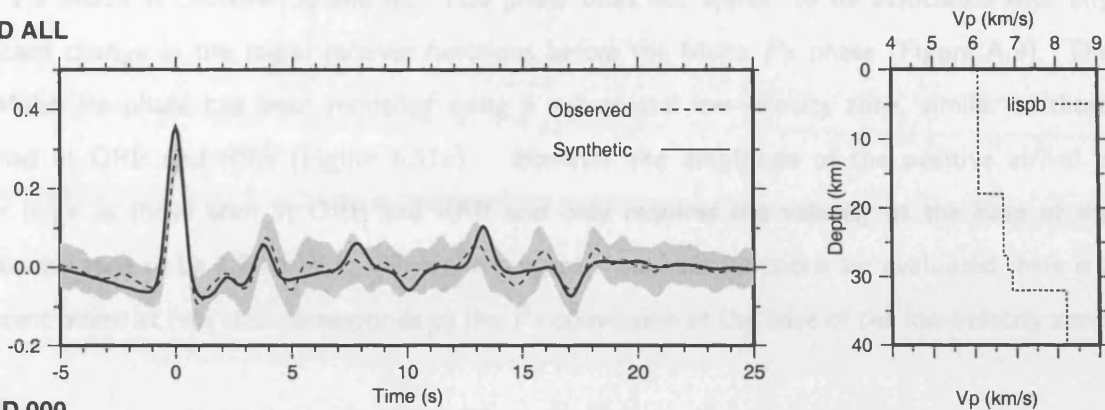
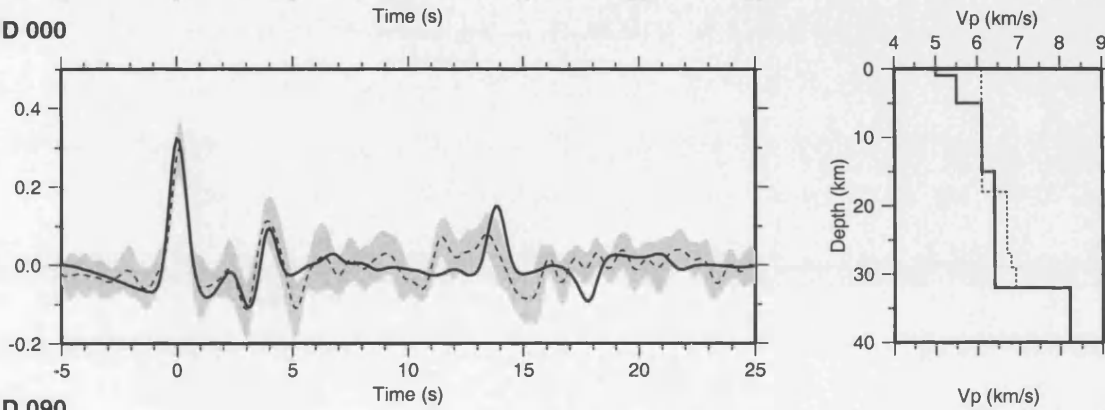
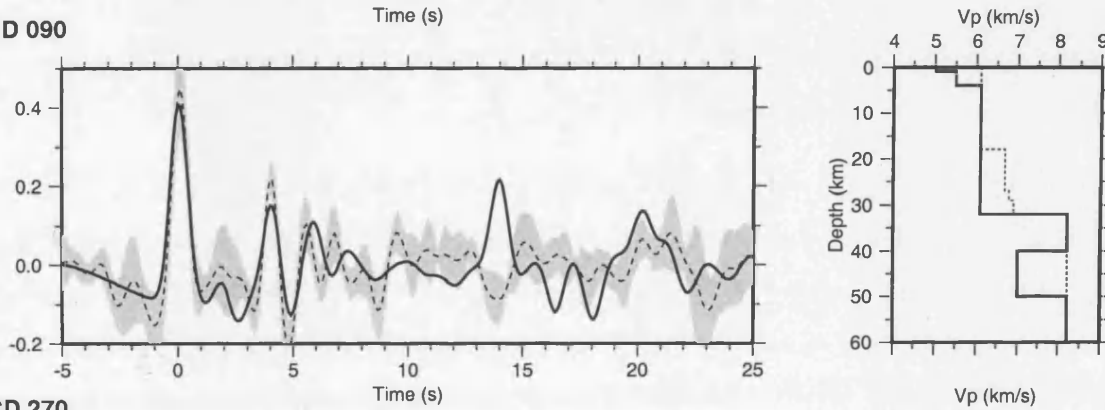
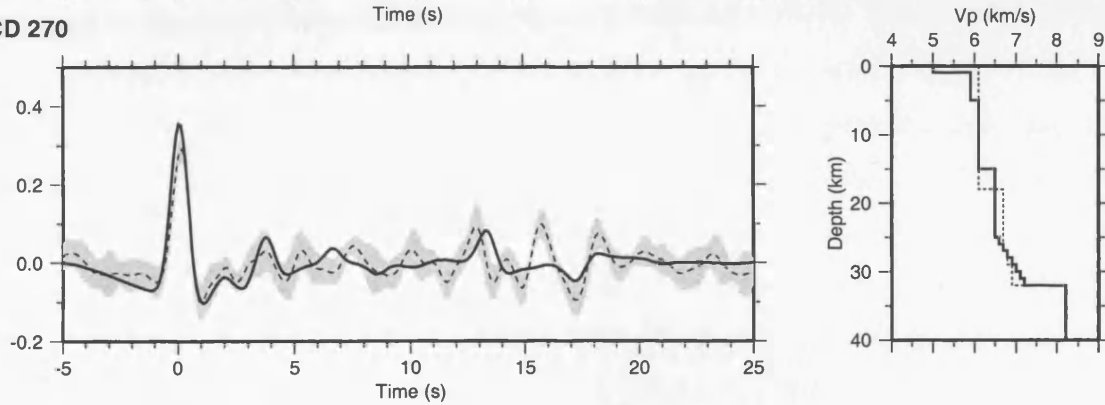
a) MCD ALL**b) MCD 000****c) MCD 090****d) MCD 270**

Figure 6.11: Results of 1D forward modelling of receiver function data from MCD for; a) a stack of all the receiver functions data, b) the data from 0° - 30° , c) the data from 90° - 120° and d) the data from 270° - 300° . The data are presented in the same way as Figure 6.3.

Moho P_s phase at between 5s and 6s. This phase does not appear to be associated with any significant change in the radial receiver functions before the Moho P_s phase (Figure A.9). The post-Moho P_s phase has been modelled using a sub-crustal low-velocity zone, similar to those modelled at ORE and RRR (Figure 6.11c) . However the amplitude of the positive arrival is not as large as those seen at ORE and RRR and only requires the velocity at the base of the low-velocity zone to be 8.2 km s^{-1} . When the tangential receiver functions are evaluated there is a consistent phase at $\sim 6\text{s}$ that corresponds to the P_s conversion at the base of the low-velocity zone.

6.4 Iapetus Suture Zone

6.4.1 Introduction

The Iapetus Suture area has been investigated by numerous onshore and offshore deep seismic reflection and refraction profiles. The LISPB and CSSP wide angle deep seismic refraction profiles provide *a priori* velocity constraint for the crust and upper mantle of the area. The BIRPS offshore WINCH, MOBIL and NEC profiles map the dipping intra-crustal reflectors that are inferred to mark the Iapetus Suture, but there is little constraint on the location of this feature onshore (see Chapter 2). This area has dense station spacing with PGB, EDI, ESK, BHH, BTA, BBO, LMI, HPK, GAL, WCB and GIM in close proximity (Figure 6.12). Not only have the receiver function data from these stations been analysed using 1D forward modelling, but their close spacing has allowed the data to be projected onto a profile along the line of LISPB, enabling the lateral variation of the receiver functions to be examined.

6.4.2 Receiver Function Profile

The receiver functions from nine stations in the Iapetus Suture region (PGB, EDI, ESK, BHH, BBO, BTA, GIM, LMI, HPK) have been projected onto a 2D profile. To produce the section, firstly the point at which each receiver function raypath pierces the Moho has been calculated. The projection along the P_s raypath has been made using the backazimuth and epicentral distance of the event and the IASPE91 velocity model (Kennett and Engdahl, 1991). This is a simple form of migration, repositioning the receiver function event to the point at which the Moho P_s phase enters the crust. A profile has been constructed by projecting each receiver function from its crustal pierce point perpendicularly onto the LISPB profile. The line of the LISPB profile was originally chosen because it cut the Caledonian northeast-southwest trend almost perpendicular to strike (Bamford et al., 1976). By projecting the receiver functions perpendicularly onto the line of the LISPB profile the data have simply been projected along strike. The final section was produced by stacking the receiver functions into 20 km wide bins at 5 km increments along the profile. This has resulted in the smoothed regularly spaced profile presented in Figure 6.13a. A synthetic receiver function profile has been generated along the LISPB profile by using Barton's 1992 P -wave velocity model (Figure 2.5a) digitised into discrete 1D velocity models at 5 km

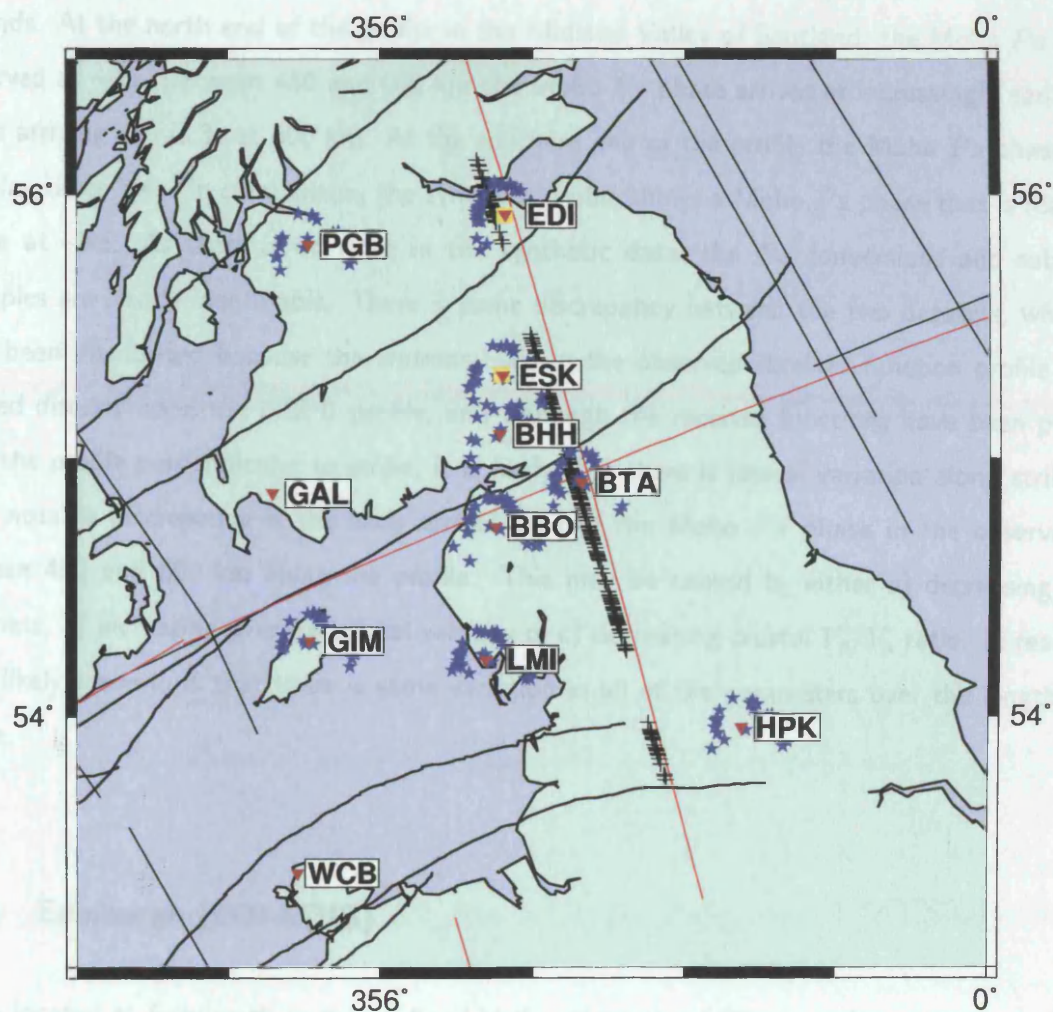


Figure 6.12: A map of the stations in the Iapetus Suture Zone region. The Moho pierce points for the receiver functions used to produce the receiver functions section are marked (stars) along with the projected positions of the receiver functions along the profile (crosses).

intervals (Figure 6.13b). As with the linear inversions and previous 1D forward modelling the default V_p/V_s ratio of 1.73 ($\sigma=0.25$) has been used. The synthetic receiver functions were produced using identical parameters to those used to calculate the observed receiver functions.

The observed receiver function section shows a Moho P_s phase at delay times of between 3 and 4 seconds. At the north end of the profile in the Midland Valley of Scotland, the Moho P_s phase is observed at ~ 4 s. Between 450 and 600 km the Moho P_s phase arrives at increasingly earlier delay times arriving at ~ 3.3 s at 600 km. At the southern end of the profile the Moho P_s phase arrives at delay times > 4 s. In comparison, the synthetic profile shows a Moho P_s phase that is reasonably stable at ~ 4 s. As there is no noise in the synthetic data, the P_s conversions and subsequent multiples are clearly identifiable. There is some discrepancy between the two datasets, which may have been introduced because the stations used in the observed receiver function profile are not located directly upon the LISPB profile, and although the receiver functions have been projected onto the profile perpendicular to strike, it is likely that there is lateral variation along strike. The most notable discrepancy is the early arrival time of the Moho P_s phase in the observed data between 450 and 600 km along the profile. This may be caused by either a) decreasing crustal thickness, b) increasing average crustal velocity or c) decreasing crustal V_p/V_s ratio. In reality, the most likely scenario is that there is some variation in all of the parameters over the length of the profile.

6.4.3 Edinburgh (EDI-EDIB)

EDI is located at Edinburgh in the Midland Valley of Scotland (Figure 6.12), and has co-located broadband and short-period instruments (EDIB & EDI). The stacked receiver functions for these stations are shown in Figures A.15 & A.17. There is an azimuthal variation in the receiver function data from EDI and EDIB. On the westerly data (BAZ 210° - 360°) there is a strong positive arrival at ~ 2 s that is preceded by a strong negative arrival on both the broadband and short-period receiver functions. In the easterly receiver functions the strong phase at ~ 2 s is not seen, but there is some broadening of the direct P -wave arrival, suggesting the presence of a P -to- S conversion close to the receiver. The Moho P_s phase is consistent with that identified by H - κ stacking, and arrives at ~ 4 s for events from all backazimuths.

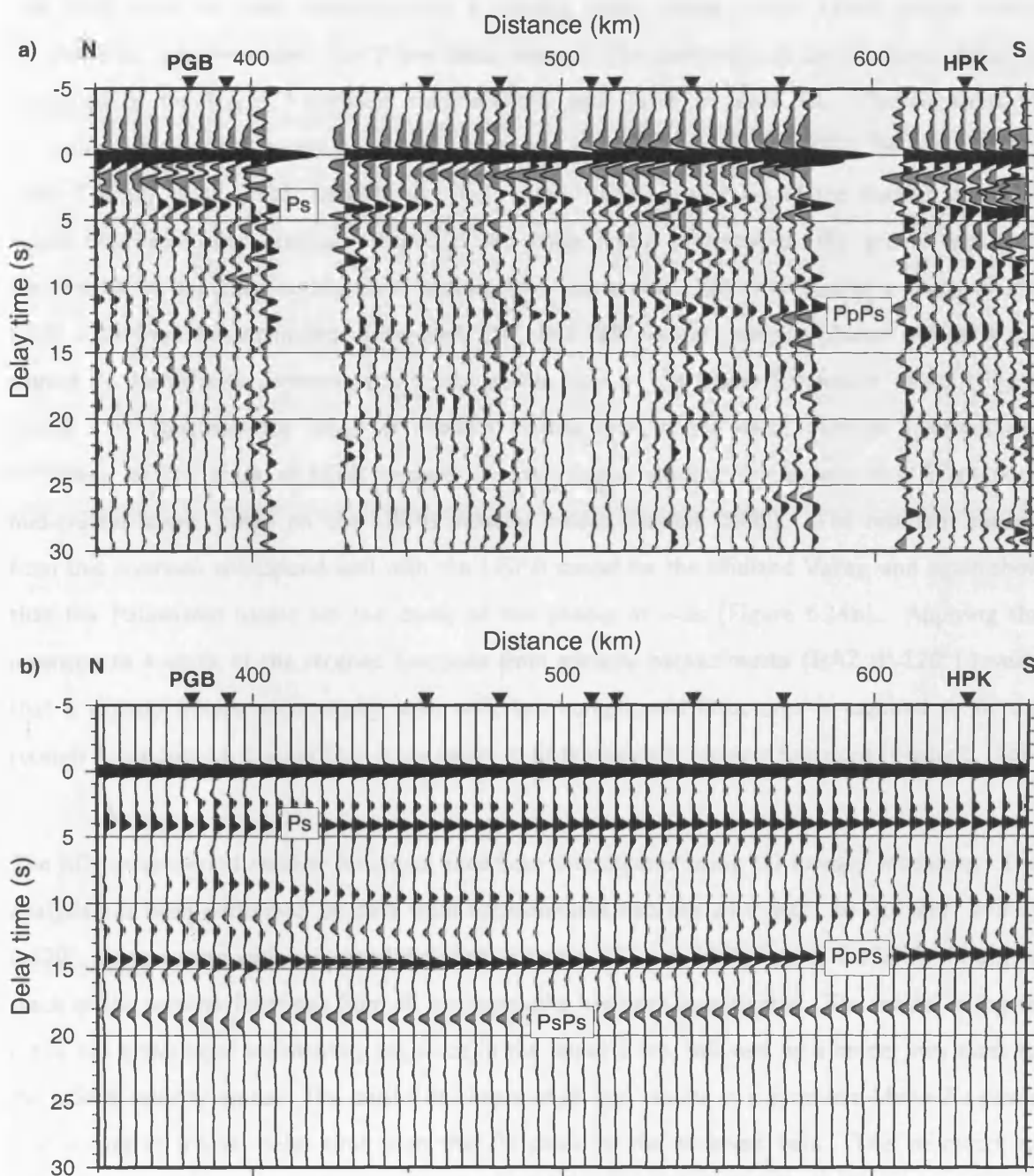
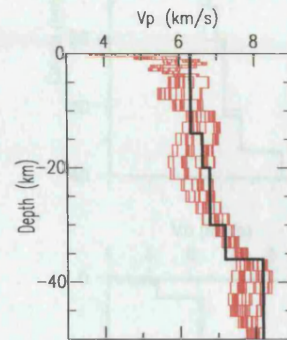
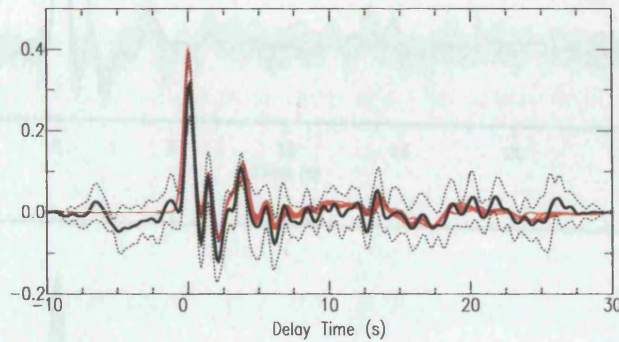


Figure 6.13: a) The data from PGB EDI ESK BTA BBO BHH GIM LMI and HPK presented as a receiver function section. Data have been projected onto the LISPb profile, and stacked in 20 km wide bins every 5 km along the profile. b) A synthetic receiver function section along the LISPb profile. The synthetic seismograms have been calculated using the velocity model of Barton (1992). The distance along the profile increases south from the start of the LISPb profile.

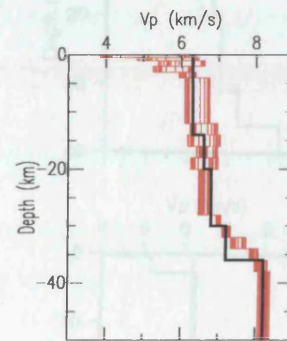
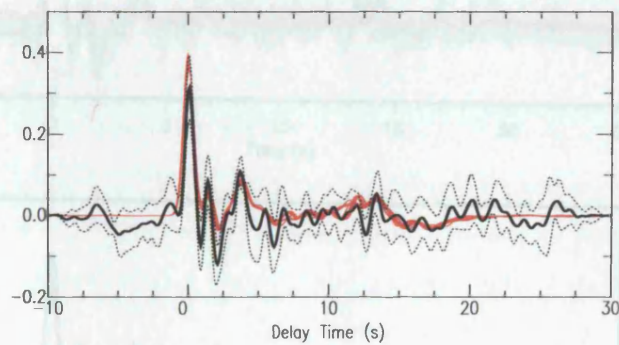
The EDIB data has been inverted using a starting model based on the LISPB profile model for the area, parameterised into 2 km thick layers. The basement of the Midland Valley is concealed by the Upper Palaeozoic Carboniferous and Devonian basin fill. The thickness of the sediments in these layers is not greater than 4 km (Bamford et al., 1977; Davidson et al., 1984; Conway et al., 1987; Dentith and Hall, 1989). The upper 4 km of the starting inversion model has been parameterised using 0.25 km thick layers to allow for the greater variation that might be expected within these sedimentary sequences. The inversion of a stack of the EDIB data from backazimuths of between 210° and 360° reveal that the phases before 2s are caused by the velocity contrasts within and at the base of the Upper Palaeozoic basins (Figure 6.14a). To constrain the range of possible models that fit the data, another inversion was performed on the stack of EDIB westerly events using a starting model with two 8 km thick mid-crustal layers, based on the LISPB velocity model (Barton, 1992). The resultant models from this inversion correspond well with the LISPB model for the Midland Valley, and again show that the Palaeozoic basins are the cause of the phases at ~2s (Figure 6.14b). Applying the inversion to a stack of the receiver functions from easterly backazimuths (BAZ 0°-120°) reveals that a slightly thicker sedimentary layer with less complicated structures is required to fit the receiver functions; i.e. thicker than is necessary to fit the westerly receiver functions (Figure 6.14c).

The EDI short-period receiver functions have been investigated using 1D forward modelling. The analysis has been performed on data from backazimuths between a) 0-360°, b) 210-360° and c) 0-120°, using events with epicentral distances between 60° and 100° (Figures 6.15a-d). Firstly, a stack of the receiver functions from all backazimuths has been investigated. The model in Figure 6.15a has a two layer sedimentary sequence in the upper 3 km, followed by a model very close to the LISPB velocity model. The crustal thickness of 36 km, results in a synthetic Moho P_s phase that arrives at a later delay time than the P_s phase in the observed data. This mismatch in the delay time is increased when the observed and synthetic Moho P_pP_s phases are compared. Assumpção and Bamford (1978) report a reduced Poisson's ratio for the crust of the Midland Valley of Scotland in comparison with the rest of the LISPB profile. Using a σ of 0.22 rather than the modelling code default of $\sigma = 0.25$ produces a synthetic receiver function in which the delay time of the Moho P_s phase matches that of the observed data whilst maintaining the crustal thickness of 36 km from the LISPB profile defined by Bamford et al. (1978) and Barton

a) EDI WEST



b) EDI WEST



c) EDI EAST

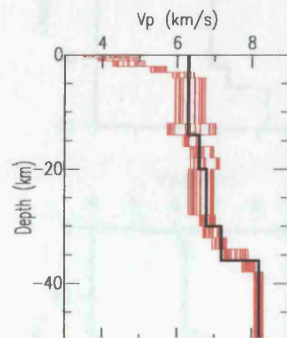
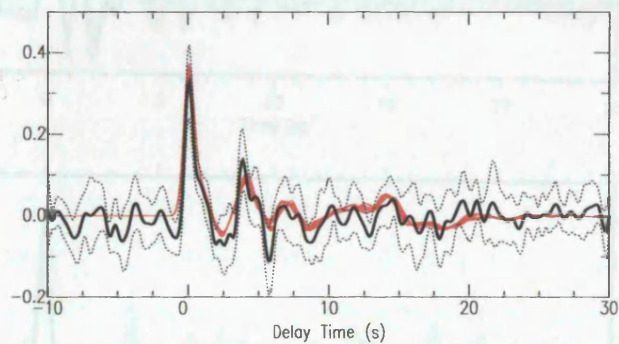
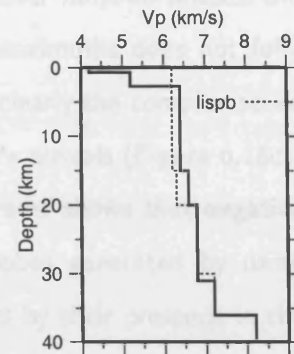
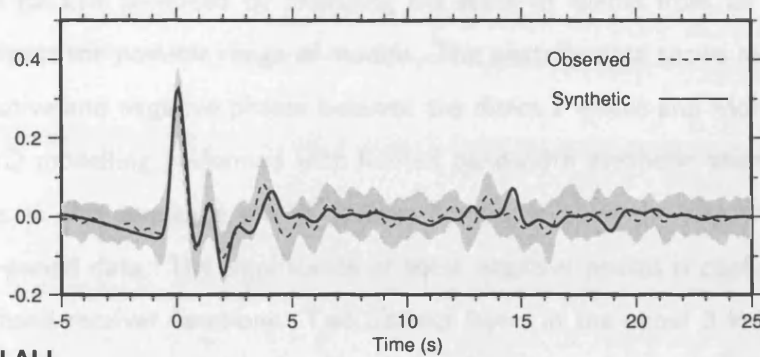
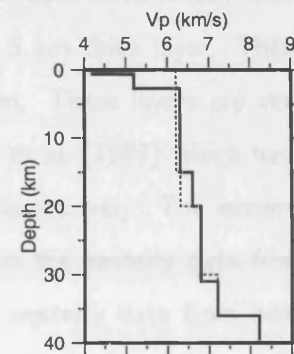
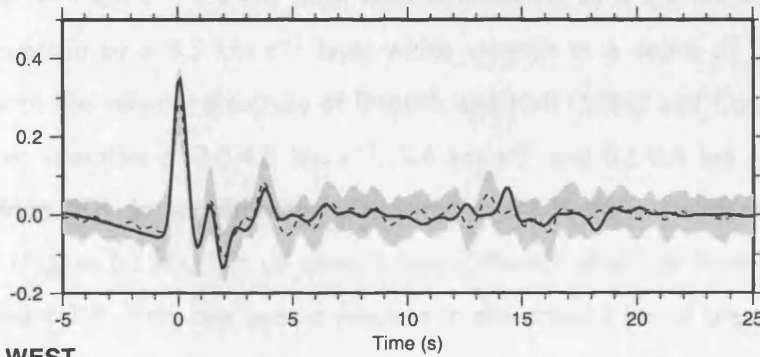


Figure 6.14: Results of 1D inversion of receiver function data from EDIB for; a) the westerly receiver function data, b) the westerly receiver function data (with a limited parameter starting model) and c) the easterly receiver function data. The data are presented in the same way as Figure 6.1.

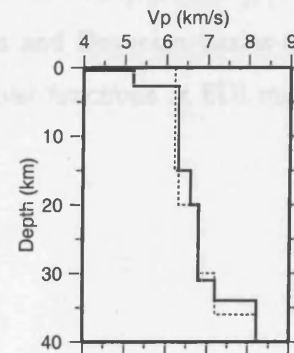
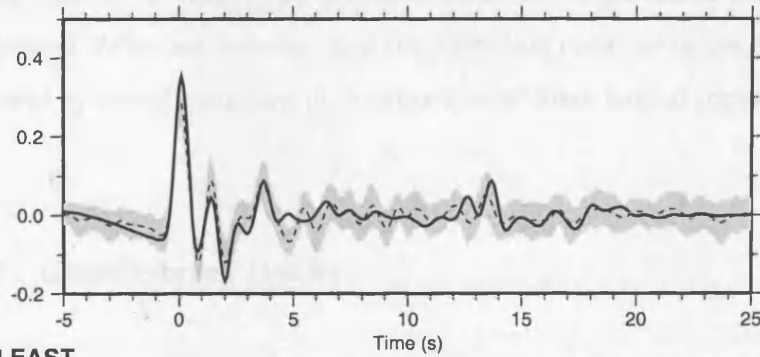
a) EDI ALL



b) EDI ALL



c) EDI WEST



d) EDI EAST

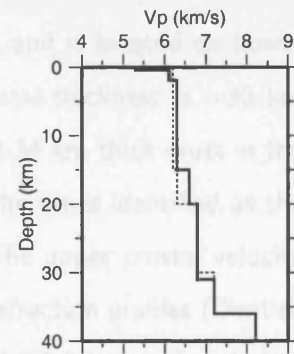
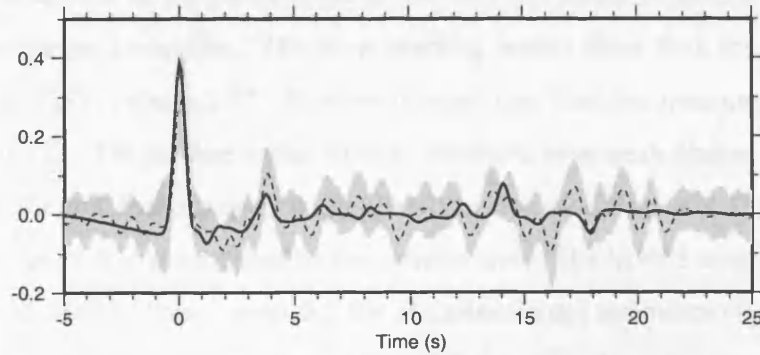


Figure 6.15: Results of 1D forward modelling of receiver function data from EDI for; a) a stack of all of the receiver functions, b) a stack of all of the receiver functions, c) the westerly receiver function data and d) the easterly receiver function data. The data are presented in the same way as Figure 6.3.

(1992) (Figure 6.15b). Due to the azimuthal variation in the crustal receiver function phases, the model for EDI produced by analysing the stack of events from all backazimuths does not fully investigate the possible range of models. The westerly data shows most clearly the complex series of positive and negative phases between the direct P -wave and Moho P_s arrivals (Figure 6.15c). The 1D modelling performed with limited bandwidth synthetic seismograms shows that negative phases in this sequence are of greater amplitude than the negative lobes generated by using short-period data. The significance of these negative phases is confirmed by their presence in the broadband receiver functions. Two distinct layers in the upper 3 km are required to match these phases. A 4 km s^{-1} 0.5 km thick layer is underlain by a 5.2 km s^{-1} 2.5 km thick layer. These are underlain by a 6.3 km s^{-1} layer which extends to a depth of 15 km. These layers are very similar to the velocity structure of Dentith and Hall (1989) and Conway et al. (1987) which have P -wave velocities of $3.0\text{--}4.5 \text{ km s}^{-1}$, 5.4 km s^{-1} and $6.1\text{--}6.4 \text{ km s}^{-1}$ respectively. The easterly data from EDI do not show the broadened direct P -wave arrival seen in the easterly data from EDIB (Figure 6.15d), but do show a very different structure from the westerly data from both EDI and EDIB. Only one layer is required in the upper 3 km of the model to fit the phases of the stacked receiver function. The geological structure of the Carboniferous and Devonian basins of the Midland Valley are complex, and the azimuthal variation in the receiver functions at EDI may be caused by lateral variations in the structure of these basinal rocks.

6.4.4 Glenifferbraes (PGB)

PGB is located at the western end of the Midland Valley (Figure 6.12), and is located on Lower Carboniferous Limestone. The $H\text{--}\kappa$ stacking results show that the crustal thickness is $\sim 30 \text{ km}$ and the V_p/V_s ratio is 1.77. This is a thinner crust than the measured 32–34 km thick crust in the east at EDI. The stacked radial receiver functions have weak phases at the times identified as the Moho P_s and P_pP_s phases in the $H\text{--}\kappa$ stacking plot (Figure A.19). The upper crustal velocity model for PGB is constrained by the western end of the MAVIS seismic refraction profiles (Dentith and Hall, 1989). These reveal 2–3 km of Carboniferous sediments ($V_p = 4.0\text{--}5.0 \text{ km s}^{-1}$) underlain by $\sim 1 \text{ km}$ of Devonian sediments ($V_p = 5.5 \text{ km s}^{-1}$). The velocity structure in the MAVIS model does not fit the observed receiver functions, but when the near surface velocity is increased to $\sim 5.4 \text{ km s}^{-1}$ the near surface phases can be fitted (Figure 6.16). PGB does not lie directly upon

the MAVIS profile, which samples the Upper Carboniferous Westphalian coal measures and delta sequences, and it would be expected that the Lower Carboniferous limestone on which PGB lies is of higher velocity than the Westphalian units.

The LISPB velocity model suggests that the crust of the Midland Valley is ~ 36 km at its thickest point. This model does not fit the observed receiver functions from PGB (Figure 6.16a). The Moho P_s phase of the synthetic receiver function is later in time and greater in amplitude than in the observed receiver function. The amplitude of the Moho P_s phase at PGB is lower than that at EDI. There appears to be a double phase between 3.5-4.5s. The timing and amplitude of this double phase may be fitted by increasing the velocity of the high velocity lower crustal layer identified from the LISPB model for the Midland Valley (Barton, 1992) to ~ 7.6 km s $^{-1}$ (Figures 6.16b-d). This velocity is similar to that of the high velocity lower crustal layer seen by Al-Kindi (2002) beneath the Irish Sea. The top of this layer is at 28 km, with the seismic velocity increasing to 8.2 km s $^{-1}$ at 36 km (Figure 6.16b). In the stacked receiver functions there are no strong phases that could be identified as the Moho $PpPs$ phase. The increase in V_p at the base of the crust has generated two discontinuities with a moderate ΔV_p rather than a single discontinuity with a large ΔV_p . This may be one of the reasons that the Moho $PpPs$ phase is not resolvable in the observed receiver function data.

6.4.5 Eskdalemuir (ESK-ESKB) & Galloway (GAL)

Galloway and Eskdalemuir are located on the Silurian rocks of the Southern Uplands. The data from ESK and ESKB (Figure 6.12) have been analysed in Section 6.2. The data from Eskdalemuir show consistent Moho P_s and $PpPs$ phases, and the $H-\kappa$ stacking reveals a crustal thickness of 30-31 km. The 1D modelling of the data from Eskdalemuir indicate a crustal thickness of ~ 33 km, using the velocities from the LISPB profile model (Barton, 1992).

Although the instrument at Galloway (GAL) (Figure 6.12) samples the Southern Uplands terrane, and is located on similar Silurian rocks, the stacked receiver functions are somewhat different to those observed at Eskdalemuir. The strongest phase following the direct P -wave arrivals is observed at between 2.5-3s (Figure A.25). There is also a phase that can be identified on some,

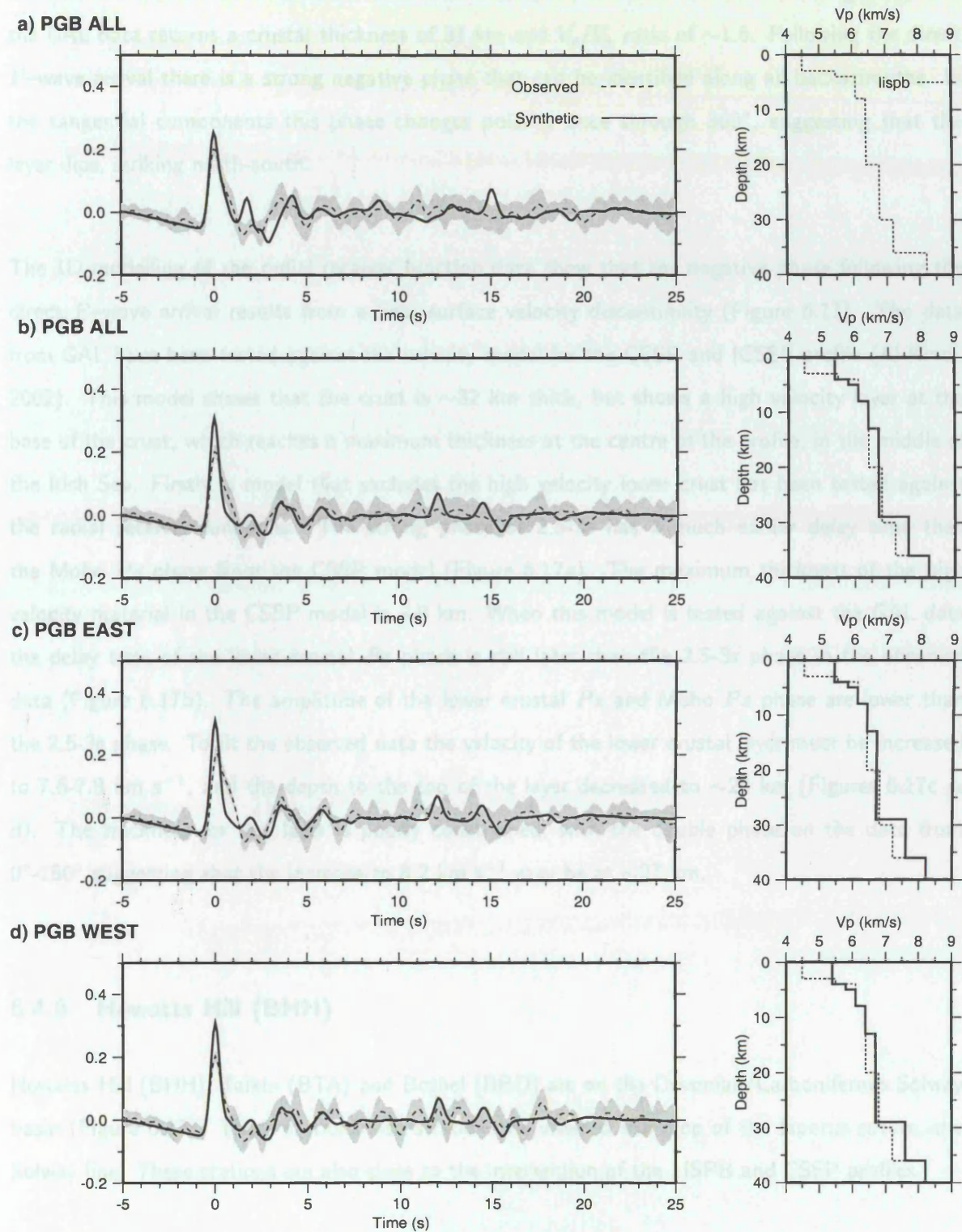


Figure 6.16: Results of 1D forward modelling of receiver function data from PGB for; a) & b) a stack of all of the data compared in a) against the LISPb model, and in b) with the preferred model to fit the observed phases, c) the easterly data and d) the westerly data. The data are presented in the same way as Figure 6.3.

but not all backazimuths in the stacked receiver functions, at ~ 12 s. The $H-\kappa$ stacking analysis of the GAL data returns a crustal thickness of 31 km and V_p/V_s ratio of ~ 1.6 . Following the direct P -wave arrival there is a strong negative phase that can be identified along all backazimuths. In the tangential components this phase changes polarity once through 360° , suggesting that the layer dips, striking north-south.

The 1D modelling of the radial receiver function data show that the negative phase following the direct P -wave arrival results from a near surface velocity discontinuity (Figure 6.17). The data from GAL have been tested against the velocity model for the CSSP and ICSSP profile (Al-Kindi, 2002). This model shows that the crust is ~ 32 km thick, but shows a high velocity layer at the base of the crust, which reaches a maximum thickness at the centre of the profile, in the middle of the Irish Sea. Firstly, a model that excludes the high velocity lower crust has been tested against the radial receiver functions. The strong phase at 2.5-3s has a much earlier delay time than the Moho P_s phase from the CSSP model (Figure 6.17a). The maximum thickness of the high velocity material in the CSSP model is ~ 8 km. When this model is tested against the GAL data the delay time of the lower crustal P_s phase is still later than the 2.5-3s phase in the observed data (Figure 6.17b). The amplitude of the lower crustal P_s and Moho P_s phase are lower than the 2.5-3s phase. To fit the observed data the velocity of the lower crustal layer must be increased to $7.6-7.8 \text{ km s}^{-1}$, and the depth to the top of the layer decreased to ~ 20 km (Figures 6.17c & d). The thickness for this layer is poorly constrained, with the double phase on the data from $0^\circ-180^\circ$ suggesting that the increase to 8.2 km s^{-1} may be at ~ 27 km.

6.4.6 Howatts Hill (BHH)

Howatts Hill (BHH), Talkin (BTA) and Bothel (BBO) are on the Devonian/Carboniferous Solway basin (Figure 6.12). These stations also straddle the inferred sub-crop of the Iapetus suture, the Solway line. These stations are also close to the intersection of the LISPb and CSSP profiles.

BHH is located in the centre of the Solway basin, and is situated on Devonian Old Red Sandstone. The stacked receiver functions show phases at ~ 3.5 s and ~ 11 s that correspond to the Moho P_s and $PpPs$ phases derived from the $H-\kappa$ stacking solution of a crustal thickness of 27 km and

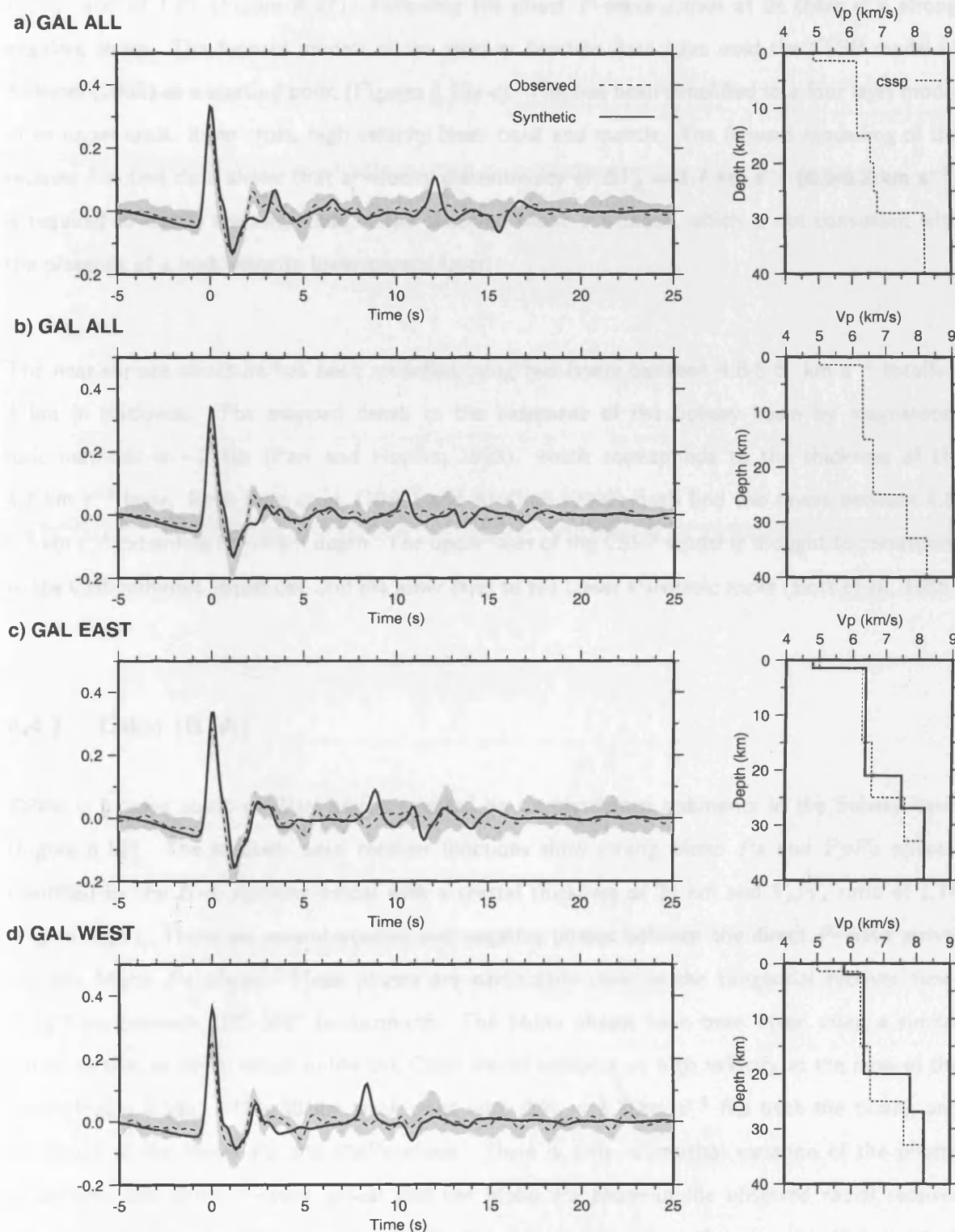


Figure 6.17: Results of 1D forward modelling of receiver function data from GAL for; a) a stack of all of the receiver function data compared with the CSSP model, b) a stack of all of the receiver function data compared with the preferred model, c) the easterly data and d) the westerly data. The data are presented in the same way as Figure 6.3.

V_p/V_s ratio of 1.81 (Figure A.27). Following the direct P -wave arrival at 0s there is a strong negative phase. The forward models of the receiver function data have used the CSSP model of Al-Kindi (2002) as a starting point (Figures 6.18a-c). This has been simplified to a four layer model of an upper crust, lower crust, high velocity lower crust and mantle. The forward modelling of the receiver function data shows that a velocity discontinuity of $\Delta V_p = 1.7 \text{ km s}^{-1}$ ($6.5\text{-}8.2 \text{ km s}^{-1}$) is required to match the amplitude of the observed Moho P_s phase, which is not consistent with the presence of a high velocity lower crustal layer.

The near surface structure has been modelled using two layers between $4.8\text{-}5.8 \text{ km s}^{-1}$ totalling 4 km in thickness. The mapped depth to the basement of the Solway basin by magnetotelluric methods is $\sim 2 \text{ km}$ (Parr and Hutton, 1993), which corresponds to the thickness of the 4.7 km s^{-1} layer. Both Bott et al. (1985) and Al-Kindi (2000) both find two layers between $4.5\text{-}5.7 \text{ km s}^{-1}$ extending to $\sim 4 \text{ km}$ depth. The upper layer of the CSSP model is thought to correspond to the Carboniferous sequences, and the lower layer to the Lower Paleozoic rocks (Bott et al., 1985).

6.4.7 Talkin (BTA)

Talkin is located south of BHH and is situated on Carboniferous sediments in the Solway basin (Figure 6.12). The stacked radial receiver functions show strong Moho P_s and PpP_s phases, identified by the $H\text{-}\kappa$ stacking model with a crustal thickness of 29 km and V_p/V_s ratio of 1.74 (Figure A.29). There are several positive and negative phases between the direct P -wave arrival and the Moho P_s phase. These phases are particularly clear in the tangential receiver functions from between $210^\circ\text{-}360^\circ$ backazimuth. The Moho phases have been fitted using a similar model to that at BHH, which unlike the CSSP model contains no high velocity at the base of the crust (Figure 6.19a). This 30 km thick crust with $\Delta V_p = 1.7 \text{ km s}^{-1}$ fits both the timing and amplitude of the Moho P_s and PpP_s phase. There is some azimuthal variation of the phases in between the direct P -wave arrival and the Moho P_s phase in the observed radial receiver functions. The two models required to fit these phases show that there is azimuthal variation in the near surface structure, but the crustal velocity model required to fit the data are the same (Figures 6.19b & c).

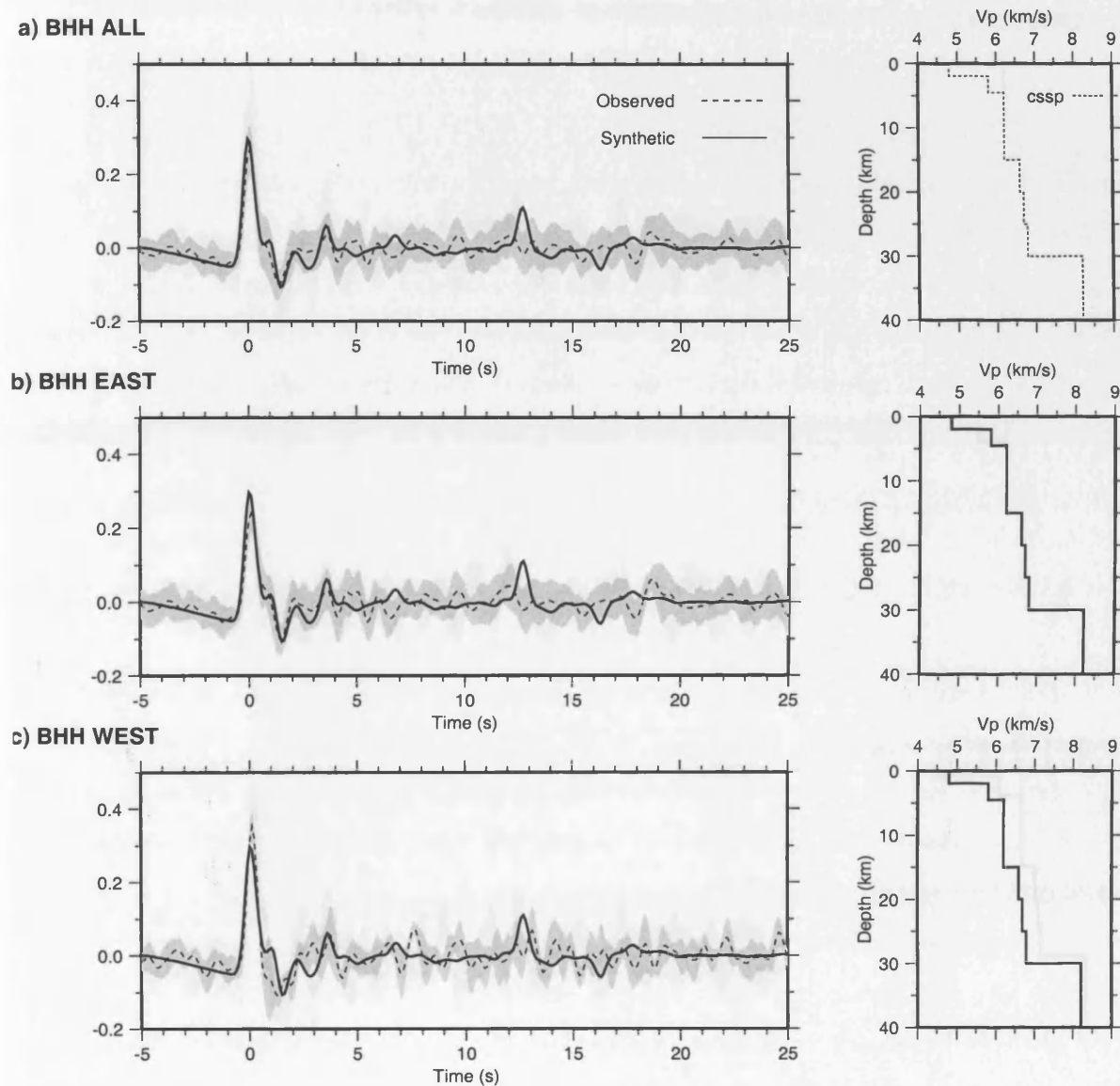


Figure 6.18: Results of 1D forward modelling of receiver function data from BHH for; a) a stack of all of the receiver function data, b) the easterly data and c) the westerly data. The data are presented in the same way as Figure 6.3.

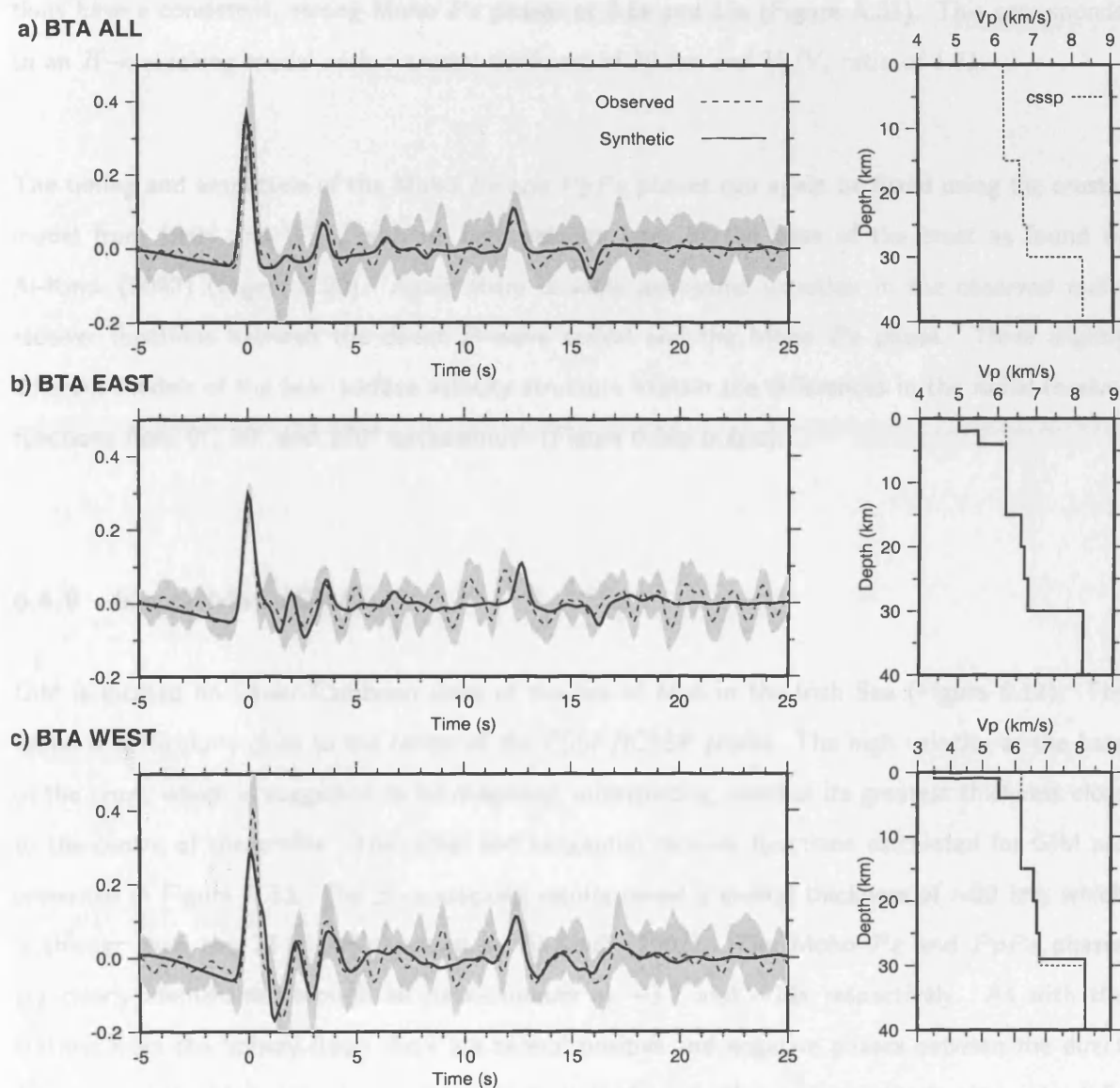


Figure 6.19: Results of 1D forward modelling of receiver function data from BTA for; a) a stack of all of the receiver function data, b) the easterly data and c) the westerly data. The data are presented in the same way as Figure 6.3.

6.4.8 Bothel (BBO)

BBO is located on Carboniferous limestone at the edge of the Solway basin, on the margin of Silurian/Ordovician sedimentary volcanic sequences (Figure 6.12). The stacked receiver functions have a consistent, strong Moho P_s phases at 3.5s and 12s (Figure A.31). This corresponds to an $H-\kappa$ stacking model with a crustal thickness of 29 km and V_p/V_s ratio of 1.71.

The timing and amplitude of the Moho P_s and $PpPs$ phases can again be fitted using the crustal model from BHH and BTA, with no high velocity layer at the base of the crust as found by Al-Kindi (2002) (Figure 6.20). Again there is some azimuthal variation in the observed radial receiver functions between the direct P -wave arrival and the Moho P_s phase. Three slightly different models of the near surface velocity structure explain the differences in the radial receiver functions from 0° , 90° and 270° backazimuth (Figure 6.20a,b & c).

6.4.9 Isle of Man (GIM)

GIM is located on Lower Cambrian units of the Isle of Man in the Irish Sea (Figure 6.12). The island is particularly close to the centre of the CSSP/ICSSP profile. The high velocity at the base of the crust, which is suggested to be magmatic underplating, reaches its greatest thickness close to the centre of the profile. The radial and tangential receiver functions calculated for GIM are presented in Figure A.33. The $H-\kappa$ stacking results reveal a crustal thickness of ~ 30 km, which is thinner than the 32-33 km reported by Al-Kindi (2002). The Moho P_s and $PpPs$ phases are clearly identifiable through all backazimuths at ~ 3.2 and ~ 12 s respectively. As with the stations from the Solway Basin there are several positive and negative phases between the direct P -wave and the Moho P_s phases. However there is also a further high amplitude phase at ~ 5.5 s between 0° and 90° backazimuth (Figure 6.21a,b & c).

The radial receiver functions from GIM have been modelled using the CSSP model of Al-Kindi (2002) as the *a priori* velocity information. However, as with the stations from the Solway Basin, the amplitude of the Moho P_s phase cannot be modelled with a high velocity layer at the base of the crust (Figure 6.21a). The strong negative phase following the direct P -wave arrival has been fitted with a thin surface layer of 4.7 km s^{-1} . The secondary high amplitude phase on the

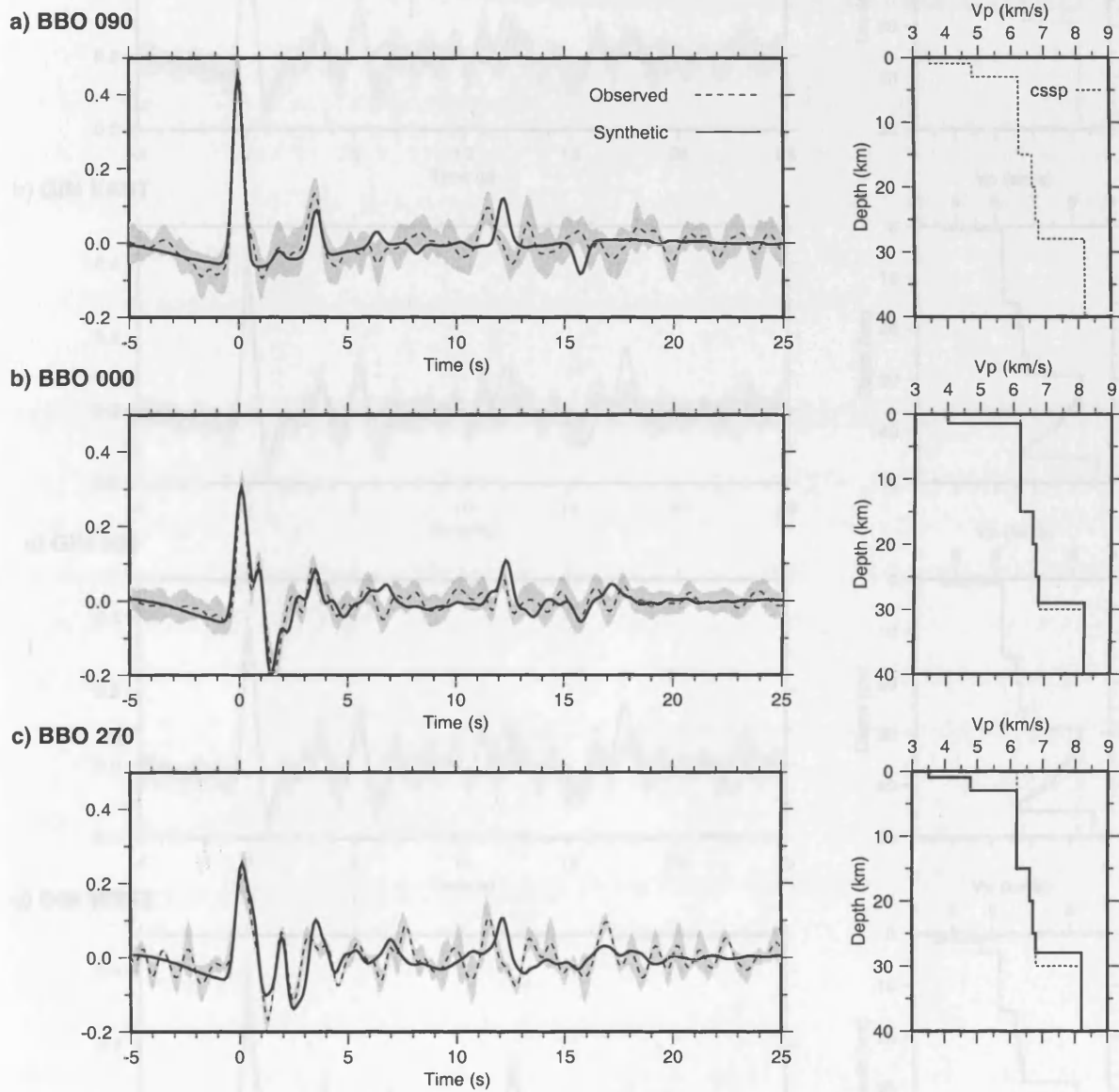
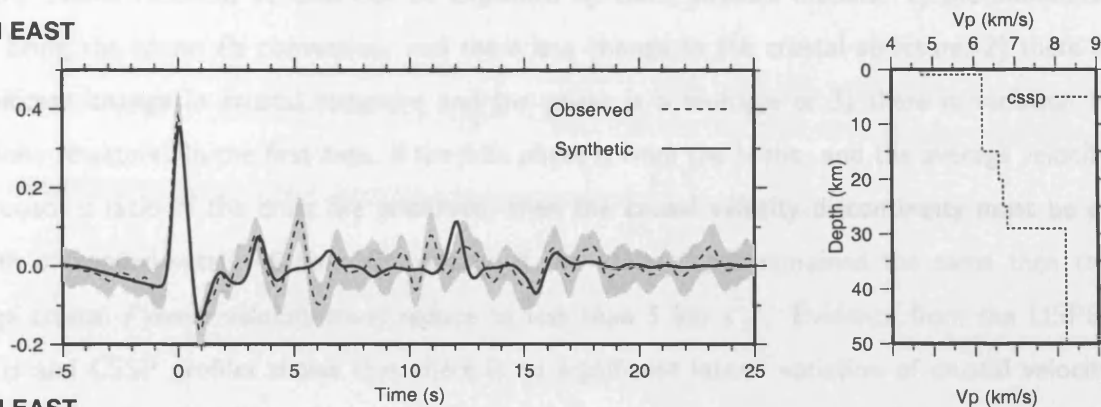
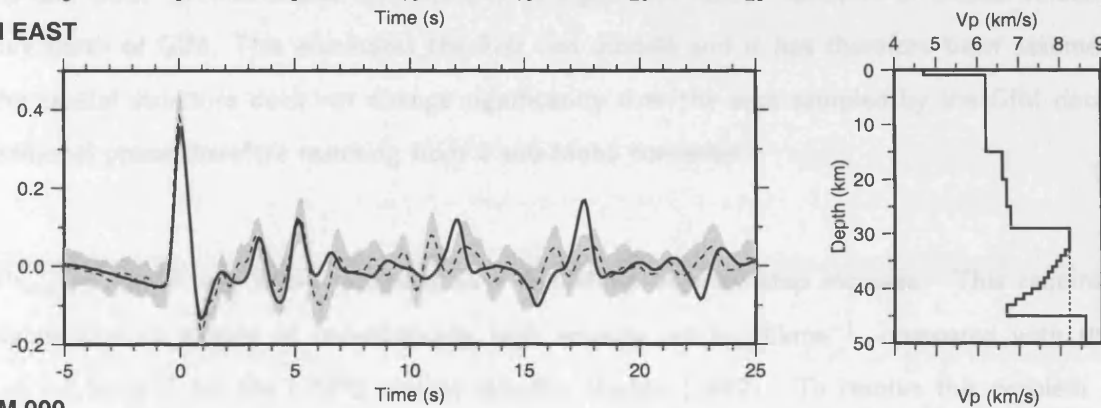


Figure 6.20: Results of 1D forward modelling of receiver function data from BBO for; a) data from 90°-120°, b) 0°-30° and c) 270°-300°. The data are presented in the same way as Figure 6.3.

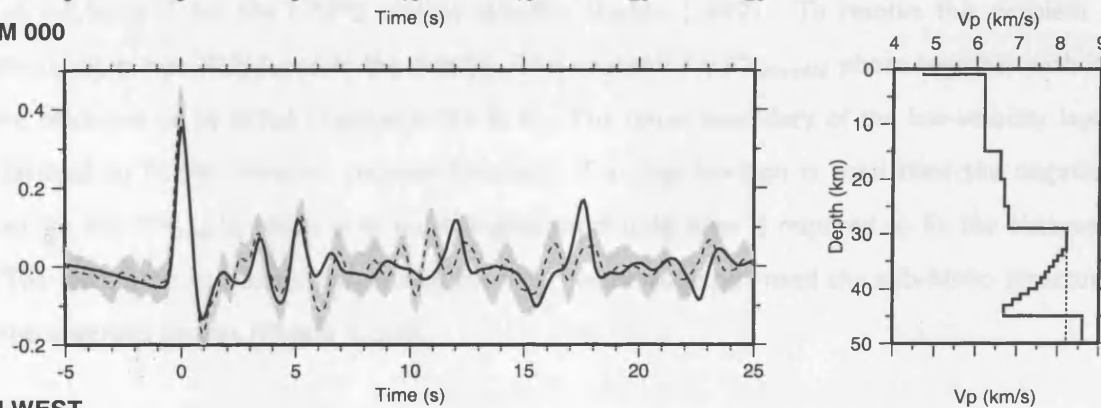
a) GIM EAST



b) GIM EAST



c) GIM 000



d) GIM WEST

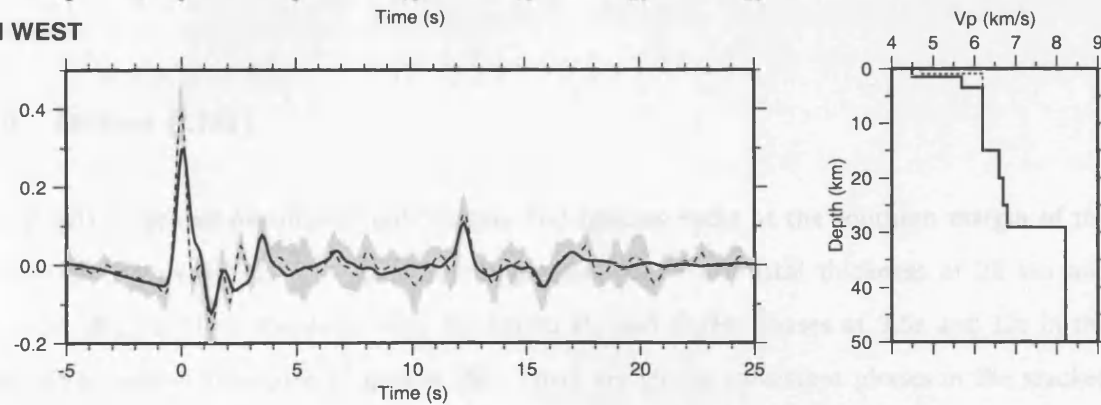


Figure 6.21: Results of 1D forward modelling of receiver function data from the GIM for; a & b) the easterly data, c) the data from 0° - 30° and d) the westerly data. The data are presented in the same way as Figure 6.3.

northerly events recorded at GIM can be explained by three possible models: 1) the additional phase being the Moho P_s conversion, and there is a change in the crustal structure, 2) there is a significant change in crustal structure and the phase is a multiple or 3) there is variation in sub-Moho structure. In the first case, if the 5.5s phase is from the Moho, and the average velocity and Poisson's ratio of the crust are preserved, then the causal velocity discontinuity must be at a depth of approximately 45 km. Conversely, if the Moho depth remained the same then the average crustal P -wave velocity must reduce to less than 5 km s^{-1} . Evidence from the LISPB, WINCH and CSSP profiles shows that there is no significant lateral variation of crustal velocity structure north of GIM. This eliminates the first two models and it has therefore been assumed that the crustal structure does not change significantly over the area sampled by the GIM data, the additional phase therefore resulting from a sub-Moho converter.

The $P_{s_{\text{mantle}}}$ phase was firstly modelled as a sub-Moho velocity step increase. This required the introduction of a layer of unrealistically high velocity, up to 10 km s^{-1} , compared with the value of $\sim 8.5 \text{ km s}^{-1}$ for the LISPB mantle reflector Barton (1992). To resolve this problem a low-velocity layer was introduced in the mantle. This enabled the $P_{s_{\text{mantle}}}$ phase together with its negative precursor to be fitted (Figures 6.21a & b). The upper boundary of the low-velocity layer is gradational to fit the observed receiver function. If a step function is used then the negative precursor to the $P_{s_{\text{mantle}}}$ phase is of much higher amplitude than is required to fit the observed data. The data from backazimuths outside the $0\text{-}90^\circ$ window do not need the sub-Moho structure to fit the observed phases (Figure 6.21d).

6.4.10 Millom (LMI)

Millom (LMI) is located on Silurian sedimentary and igneous rocks at the southern margin of the Lake District (Figure 6.12). The $H\text{-}\kappa$ stacking results show a crustal thickness of 28 km and V_p/V_s ratio of 1.75. This correlates with the Moho P_s and PpP_s phases at 3.5s and 12s in the stacked radial receiver functions (Figure A.35). There are strong consistent phases in the stacked tangential receiver function between the direct P -wave and Moho P_s phase.

As with the stations from the Solway Basin and the Isle of Man, the amplitude of the P_s phase

can be fitted without the high velocity layer at the base of the crust found on the CSSP profile (Figure 6.22). Although there is azimuthal variation in the observed radial and tangential receiver functions between the direct P -wave and Moho P_s phases, the model for the deep crustal structure remains the same (Figures 6.22b & c).

6.4.11 Haverah Park (HPK-HPKB)

Haverah Park is located on the Westphalian rocks of the Pennine Basin. The observed receiver functions are different from those seen in the Southern Uplands, and Lake District area. In the receiver functions section along the LISP profile, the Moho P_s at HPK arrives at a larger delay time than at the stations directly to the north (Figure 6.13a). The $H-\kappa$ stacking values for both the broadband and short-period instruments provide a crustal thickness of 32 km and V_p/V_s ratio of 1.78, which correlates with Moho P_s and PpP_s phases at ~ 4.2 s and ~ 13.5 s in the observed radial receiver functions (Figures A.37 & A.39). In both the radial and tangential receiver functions there are significant phases in between the direct P -wave arrival and the Moho P_s phase.

The stacks of all of the broadband data from HPKB have been inverted using a starting model based on the velocity model from southern end of the LISP profile (Barton, 1992). Two inversions have been performed; firstly using a model parameterised with numerous thin layers to give a minimum constraint result (Figure 6.23a), and secondly a model with a small number of thicker layers based on the velocity structure observed beneath the LISP profile (Figure 6.23b). Both of the inversions show an increase to mantle velocities at between 30-35 km, along with relatively low velocities in the near surface. Where the model has been constrained to three crustal layers, the near surface structure has been parameterised as a number of thin layers, and the resulting structure of the near surface in the inversion is poorly constrained.

The 1D modelling of the short-period data from HPK shows that using the LISP velocity model, the Moho P_s phase can be fitted with a crustal thickness of ~ 34 km (Figures 6.24a-c). The first positive and negative phases between the direct P -wave arrival and the Moho P_s phase have been modelled with a 2 km thick surface layer of $V_p = 4.2-5.0$ km s $^{-1}$. The phase preceding the Moho P_s phase has been fitted as a mid-crustal discontinuity at ~ 22 km, which is consistent with

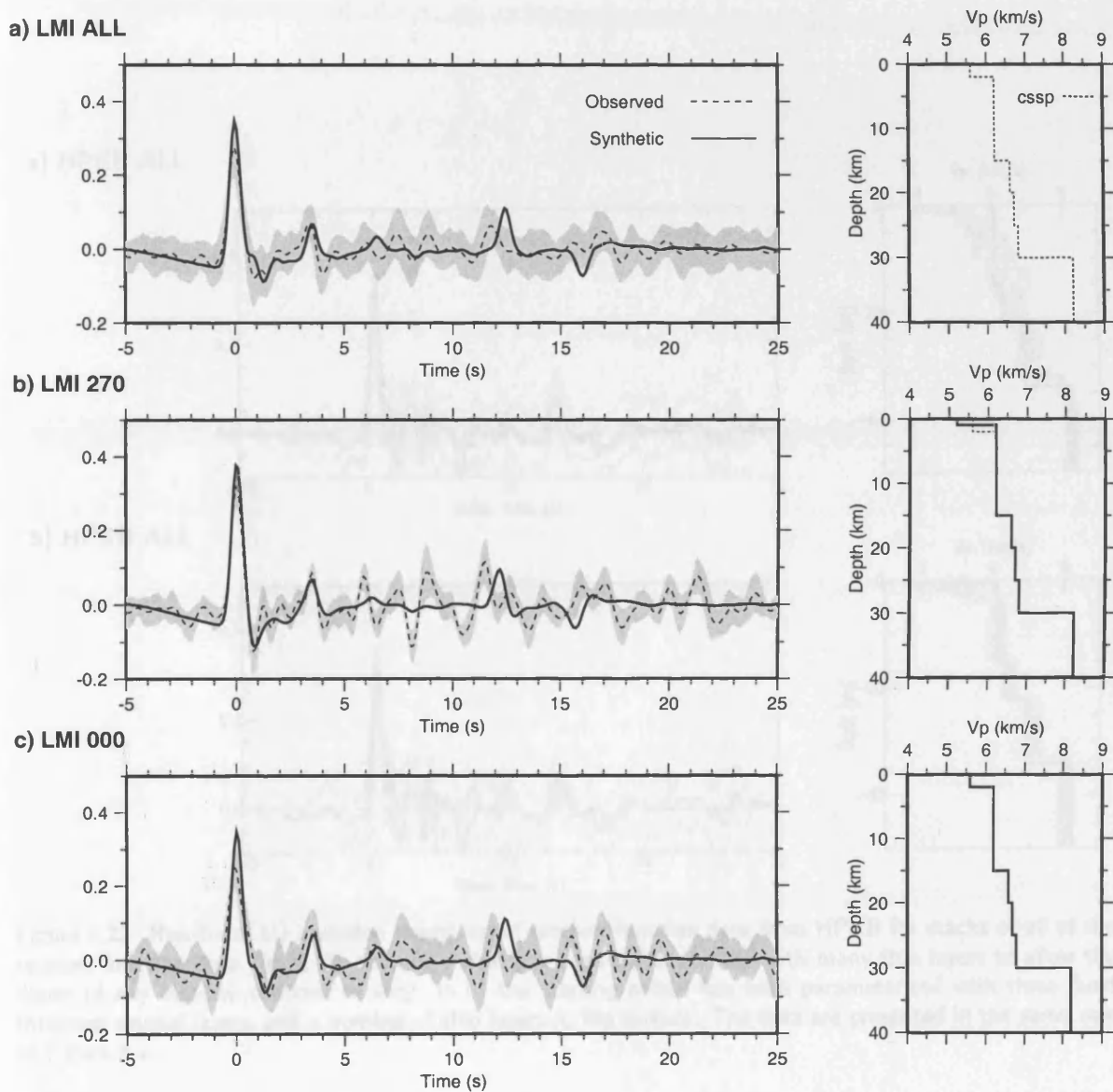
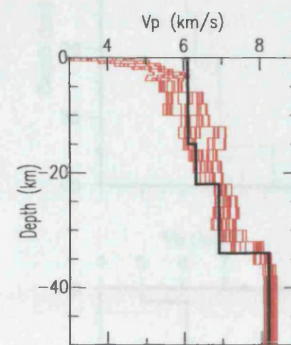
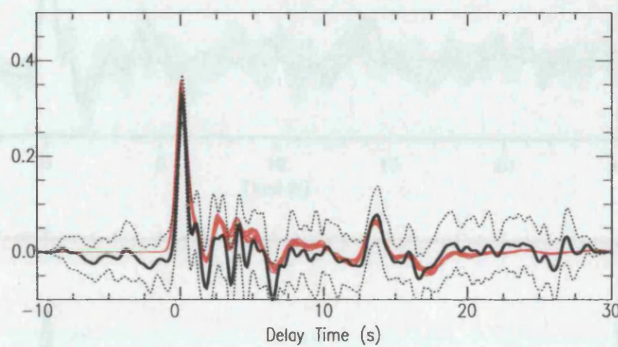


Figure 6.22: Results of 1D forward modelling of receiver function data from the LMI for; a) a stack of all of the receiver function data, b) the data from 270°-300° and c) the data from 0°-30°. The data are presented in the same way as Figure 6.3.

a) HPKB ALL



b) HPKB ALL

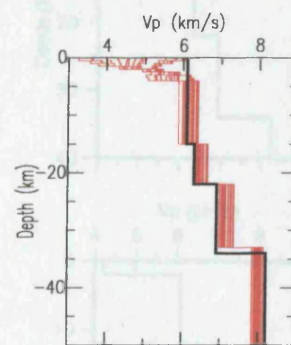
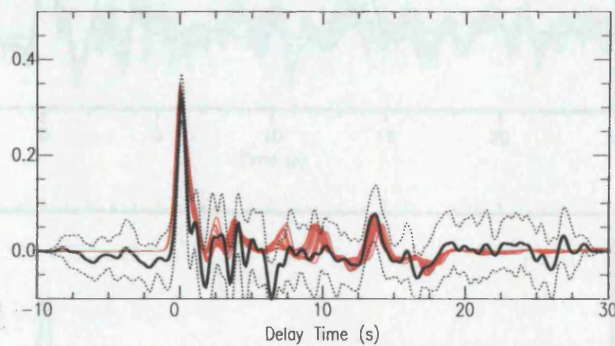


Figure 6.23: Results of 1D inversion modelling of receiver function data from HPKB for stacks of all of the receiver function data. In a) the starting model has been parameterised with many thin layers to allow the depth of any velocity contrast to vary. In b) the starting model has been parameterised with three fixed thickness crustal layers, and a number of thin layers at the surface. The data are presented in the same way as Figure 6.1.

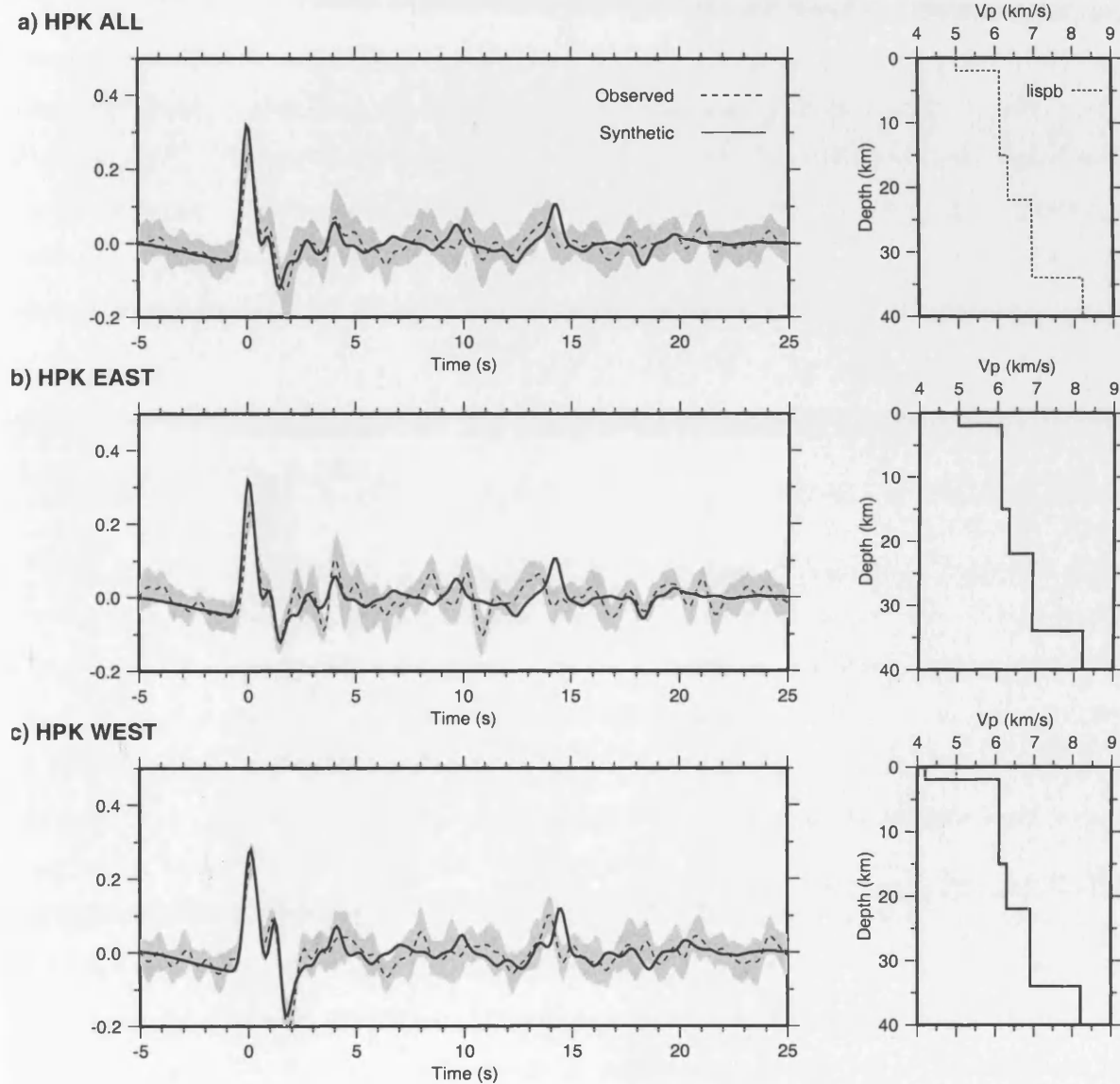


Figure 6.24: Results of 1D forward modelling of receiver function data from the HPK for; a) a stack of all of the receiver function data, b) the easterly data and c) the westerly data. The data are presented in the same way as Figure 6.3.

the 6.8 km s^{-1} layer at the southern end of the LISPb profile. However the $PpPs$ phase from this layer does not fit the later phases in the observed receiver functions. Other stations that are located on Carboniferous basins, for example EDI, PGB and BTA show phases between the direct P -wave and Moho $PpPs$ phase that can be attributed to a more complex near surface structure.

6.5 Central England & Wales

6.5.1 Introduction

The crustal structure of Central England and Wales is poorly constrained in comparison with the margins of the British Isles. The Moho has been imaged by a number of short normal incidence seismic reflection profiles, and these reveal a crust of between 33-35 km thick (Chadwick and Pharaoh, 1998). The deep seismic velocity structure of this region is constrained by the unpublished model presented in Edwards and Blundell (1984). Parts of the area are masked by a blanket of sedimentary rocks providing difficult conditions for the modelling of receiver function data. The stations used in this area are WCB, CWF, SSP, MCH, SWN, WOL and TFO (Figure 4.1).

6.5.2 Church Bay (WCB)

WCB is located on the Precambrian rocks of the Monian terrane in Anglesey. The stacked radial receiver functions from WCB show a strong Moho P -to- S conversion at ~ 4 s (Figure A.41). Between 11-14s there are two phases that could be identified as the Moho $PpPs$ phase. The $H-\kappa$ stacks of the WCB data produce a crustal thickness of 28 km and V_p/V_s ratio of 1.85. In Chapter 5 it has been noted that these values, in particular the V_p/V_s ratio are un-typical of the area. The 1D forward modelling has been based on the LISP Delta profile velocity model described in Edwards and Blundell (1984) (Figure 6.25). At the northern end of LISP Delta, the model has a crustal thickness of 34 km. The crust is divided into two layers, an upper crust of 6.1 km s^{-1} to a depth of ~ 15 km, and a lower crust of 6.4-6.6 km s^{-1} . Using this two layer crustal velocity structure, the Moho Ps phase at ~ 4 s can be fitted with a 32 km thick crust (Figures 6.25a,b). In the stack of the radial receiver function data from all backazimuths, a 5.4 km s^{-1} , 2 km thick layer above the 6.1 km s^{-1} upper crust is required to fit the negative phase between 1-2s. The boundary between the upper and lower crust corresponds with the LISP Delta model at 15 km. This has been constrained by fitting the intra-crustal $PpPs$ phase at ~ 6.5 s. The amplitude of the modelled Moho $PpPs$ phase is much greater than any in the observed receiver function data. There is clearly some mismatch between the crustal thicknesses calculated from forward modelling and $H-\kappa$ stacking. The Moho $PpPs$ phase in the WCB data is weak,

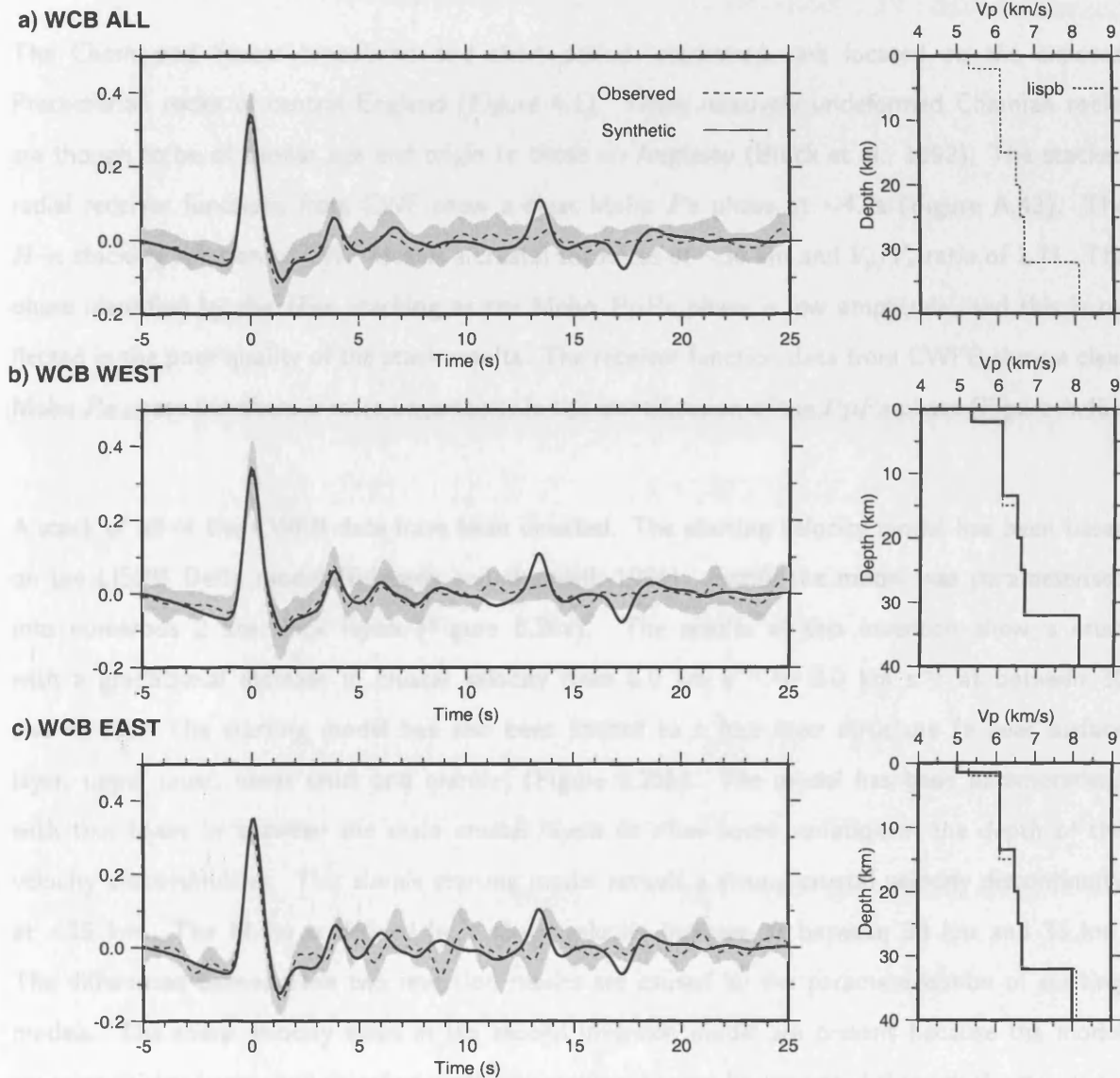


Figure 6.25: Results of 1D forward modelling of receiver function data from WCB for; a) a stack of all of the receiver function data, b) the westerly data and c) the easterly data. The data are presented in the same way as Figure 6.3.

and the maximum point of the $H-\kappa$ stack may have been distorted by other more prominent phases.

6.5.3 Charnwood Forest (CWF-CWFB)

The Charnwood Forest broadband and short-period instruments are located on the exposed Precambrian rocks of central England (Figure 4.1). These relatively undeformed Charnian rocks are thought to be of similar age and origin to those on Anglesey (Bluck et al., 1992). The stacked radial receiver functions from CWF show a clear Moho P_s phase at ~ 4.2 s (Figure A.43). The $H-\kappa$ stacking solution of CWF reveals a crustal thickness of ~ 36 km and V_p/V_s ratio of 1.71. The phase identified by the $H-\kappa$ stacking as the Moho $PpPs$ phase is low amplitude, and this is reflected in the poor quality of the stack results. The receiver function data from CWFB show a clear Moho P_s phase but there is more uncertainty in the identification of the $PpPs$ phase (Figure A.45).

A stack of all of the CWFB data have been inverted. The starting velocity model has been based on the LISPB Delta model (Edwards and Blundell, 1984). Firstly the model was parameterised into numerous 2 km thick layers (Figure 6.26a). The results of this inversion show a crust with a gradational increase in crustal velocity from 6.0 km s^{-1} to 8.0 km s^{-1} at between 30 and 40 km. The starting model has also been limited to a four layer structure (a near surface layer, upper crust, lower crust and mantle) (Figure 6.26b). The model has been parameterised with thin layers in between the main crustal layers to allow some variation in the depth of the velocity discontinuities. This simple starting model reveals a strong crustal velocity discontinuity at ~ 15 km. The Moho is defined by a sharp velocity increase at between 33 km and 35 km. The differences between the two inversion results are caused by the parameterisation of starting models. The sharp velocity steps in the second inversion model are present because the model contains thicker layers, and therefore a velocity gradient cannot be generated through the structure.

The limited parameter inversion model was based upon the LISPB Delta velocity model. This model has been tested against the broadband and short-period receiver function data by 1D forward modelling (Figure 6.27). This reveals a 35 km thick crust, similar to the value obtained by $H-\kappa$ stacking. In both the broadband and short-period data the addition of a ~ 2 km thick surface layer ($V_p = 5.5 \text{ km s}^{-1}$) is required to fit the observed receiver functions. The upper

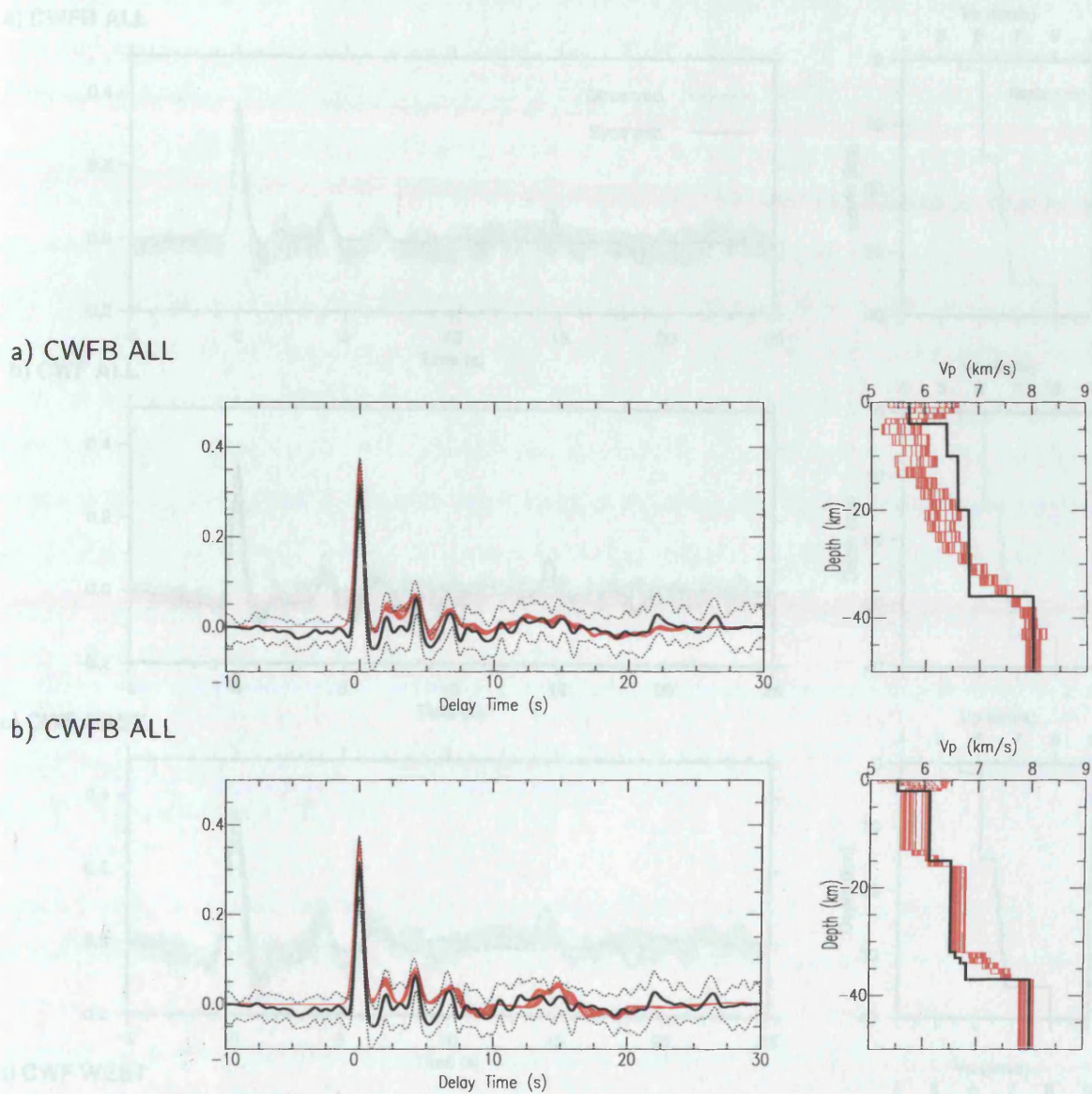


Figure 6.26: Results of 1D inversion modelling of receiver function data from CWFB for stacks of all of the receiver function data. a) Shows an inversion using a starting model with many 2 km thick layers. b) Shows an inversion performed using a starting model with fewer thick layers based on the LISPB model. The data are presented in the same way as Figure 6.1.

Figure 6.27: Results of 1D forward modelling of receiver function data from LWF & CWFB for: a) a stack of all the CWFB receiver functions, b) a stack of all the LWF receiver functions, c) a stack of the CWFB receiver functions, and d) a stack of the LWF receiver functions. The data are presented in the same way as Figure 6.1.

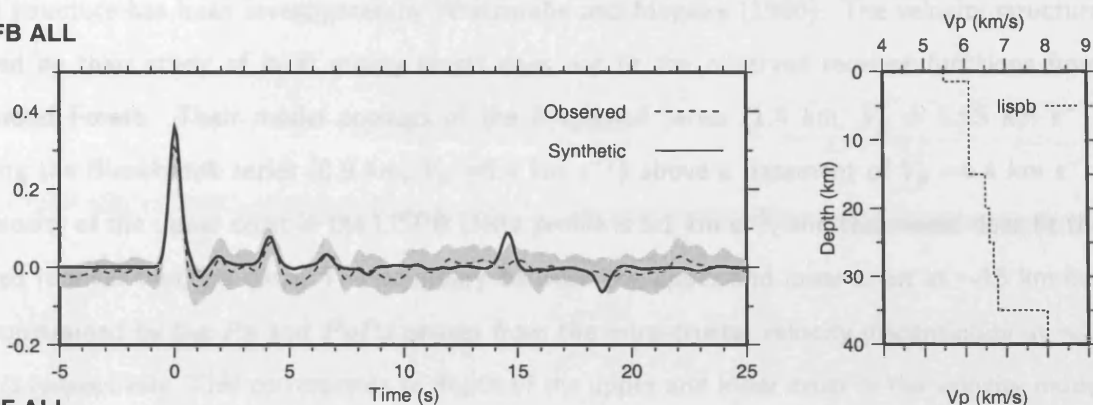
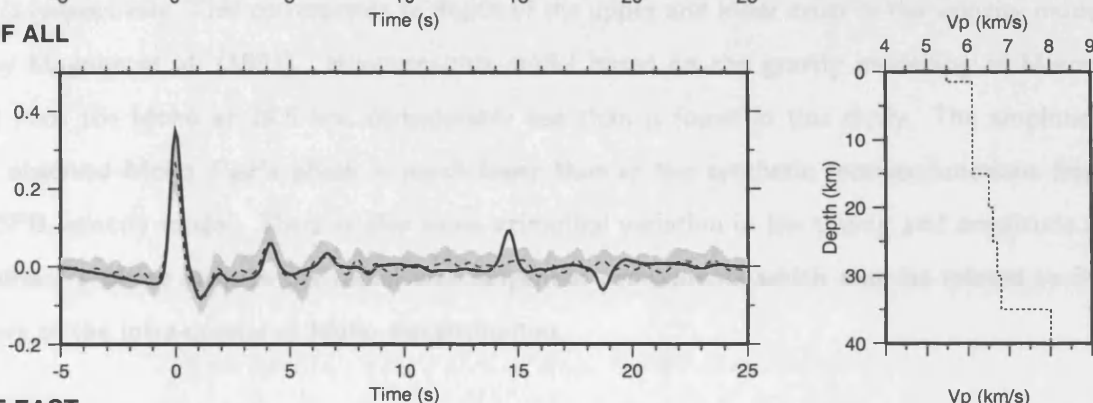
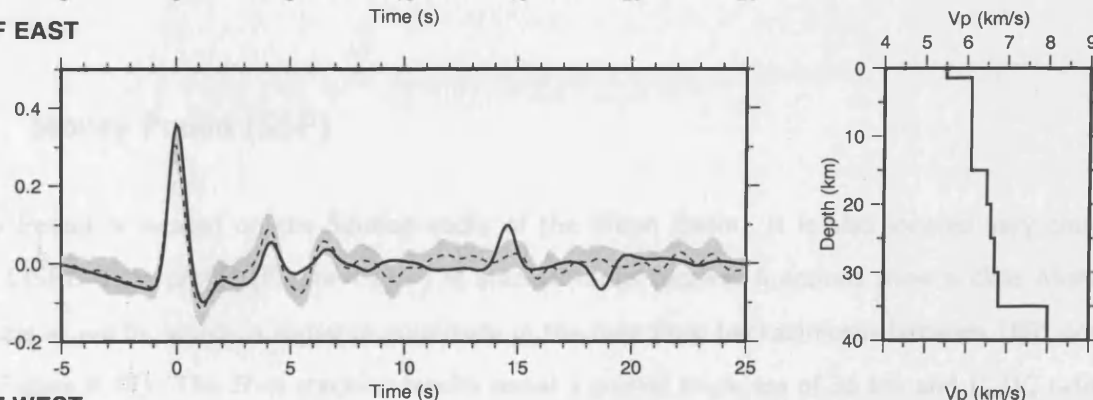
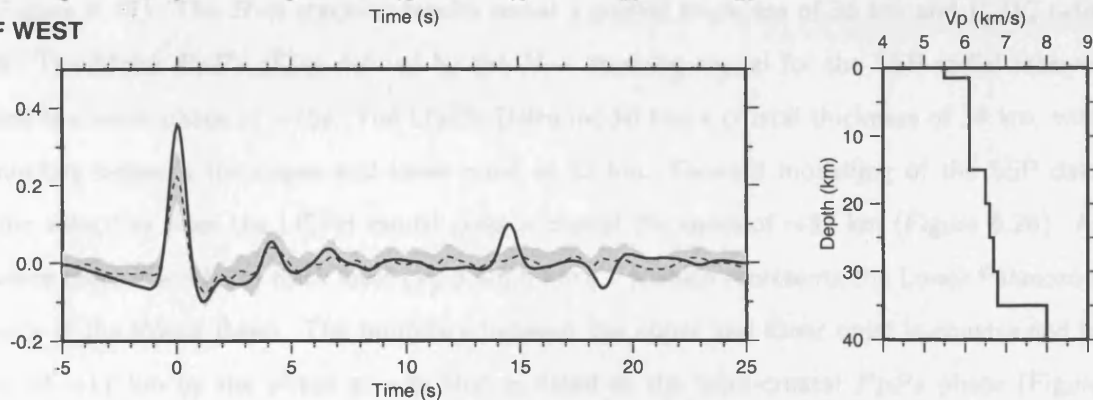
a) CWFB ALL**b) CWF ALL****c) CWF EAST****d) CWF WEST**

Figure 6.27: Results of 1D forward modelling of receiver function data from CWF & CWFB for; a) a stack of all the CWFB receiver functions, b) a stack of all the CWF receiver functions, c) a stack of the CWF easterly receiver functions and d) a stack of the CWF westerly receiver functions. The data are presented in the same way as Figure 6.3.

crustal structure has been investigated by Whitcombe and Maguire (1980). The velocity structure obtained by their study of local quarry blasts does not fit the observed receiver functions from Charnwood Forest. Their model consists of the Maplewell series (1.4 km, $V_p = 5.65 \text{ km s}^{-1}$) overlying the Blackbrook series (0.9 km, $V_p = 5.4 \text{ km s}^{-1}$) above a basement of $V_p = 6.4 \text{ km s}^{-1}$. The velocity of the upper crust in the LISP Delta profile is 6.1 km s^{-1} , and this model does fit the observed receiver function data. The boundary between the upper and lower crust at $\sim 15 \text{ km}$ has been constrained by the P_s and $PpPs$ phases from the intra-crustal velocity discontinuity at $\sim 2\text{s}$ and $\sim 7\text{s}$ respectively. This corresponds to depth of the upper and lower crust in the velocity model used by Maguire et al. (1981). However, this model based on the gravity modelling of Maroof (1973) finds the Moho at 29.5 km, considerably less than is found in this study. The amplitude of the observed Moho $PpPs$ phase is much lower than in the synthetic receiver functions from the LISP velocity model. There is also some azimuthal variation in the timing and amplitude of the Moho P_s phase in both the radial and tangential components which may be related to 3D structure of the intra-crustal or Moho discontinuities.

6.5.4 Stoney Pound (SSP)

Stoney Pound is located on the Silurian rocks of the Welsh Basin. It is also located very close to the LISP Delta profile (Figure 4.1). The stacked radial receiver functions show a clear Moho P_s phase at $\sim 4.5\text{s}$, which is higher in amplitude in the data from backazimuths between 180° and 360° (Figure A.47). The $H-\kappa$ stacking results reveal a crustal thickness of 36 km and V_p/V_s ratio of 1.75. The Moho $PpPs$ phase defined by the $H-\kappa$ stacking model for the SSP radial receiver functions is a weak phase at $\sim 15\text{s}$. The LISP Delta model has a crustal thickness of 34 km, with the boundary between the upper and lower crust at 15 km. Forward modelling of the SSP data using the velocities from the LISP model gives a crustal thickness of $\sim 37 \text{ km}$ (Figure 6.28). At the surface there is a $\sim 5 \text{ km}$ thick layer ($V_p 5.4\text{--}5.6 \text{ km s}^{-1}$) which represents the Lower Palaeozoic sediments of the Welsh Basin. The boundary between the upper and lower crust is constrained to a depth of $\sim 17 \text{ km}$ by the phase at $\sim 8\text{s}$ that is fitted as the intra-crustal $PpPs$ phase (Figure 6.28b). The amplitude of the modelled Moho $PpPs$ phase is much larger than any phases between 12-17s in the observed receiver functions.

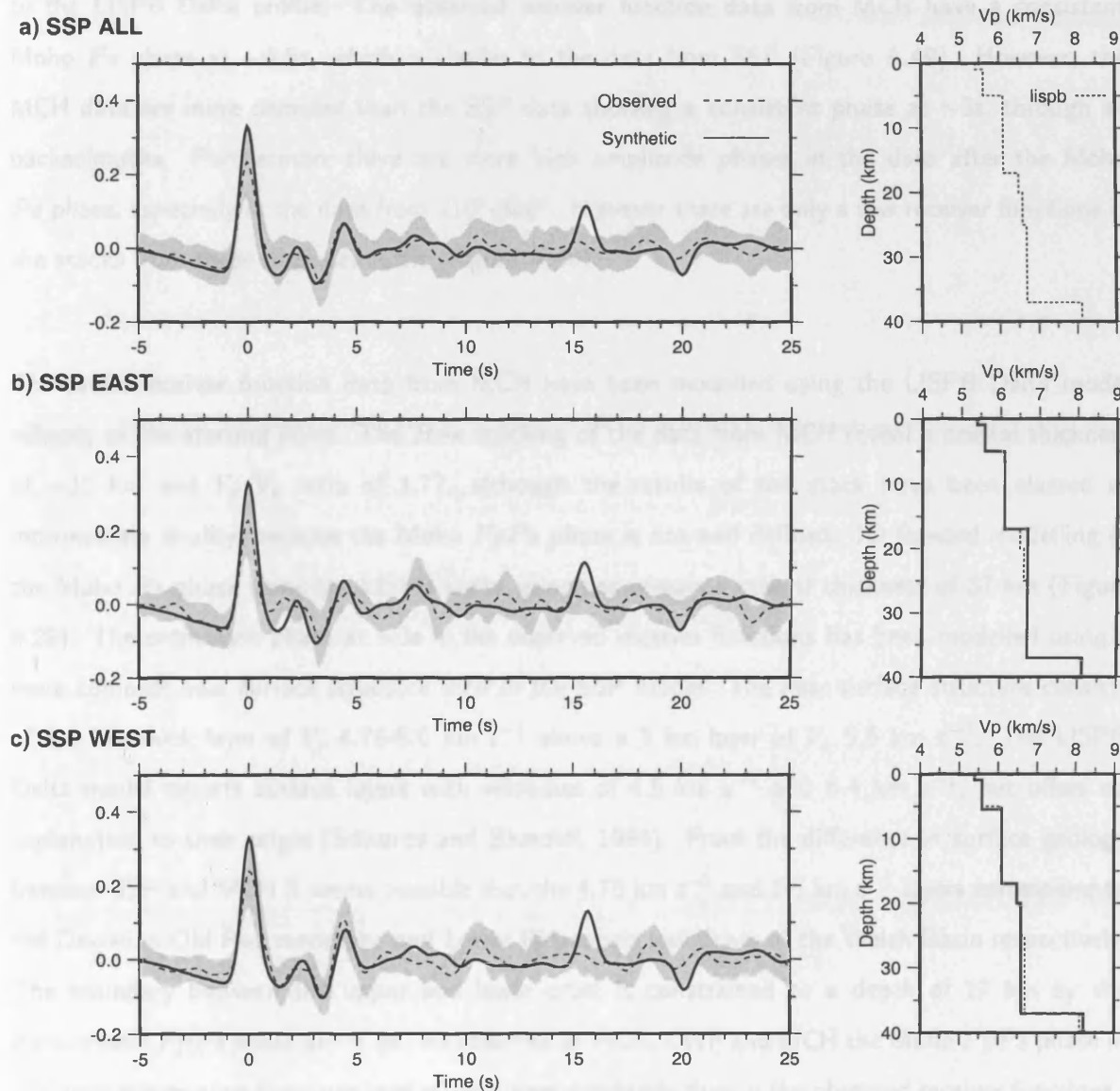


Figure 6.28: Results of 1D forward modelling of receiver function data from SSP for; a) a stack of all the receiver function data, b) the easterly data and c) the westerly data. The data are presented in the same way as Figure 6.3.

6.5.5 Michael Church (MCH)

Michael Church is located on Devonian Old Red sandstone of the Anglo-Welsh Basin (Figure 4.1), that overlies the Silurian strata mapped at Stoney Pound. The station is again located very close to the LISPB Delta profile. The observed receiver function data from MCH have a consistent Moho P_s phase at ~ 4.5 s, which is similar to the data from SSP (Figure A.49). However, the MCH data are more complex than the SSP data showing a consistent phase at ~ 3 s, through all backazimuths. Furthermore there are more high amplitude phases in the data after the Moho P_s phase, especially in the data from 210° - 300° . However there are only a few receiver functions in the stacks from these back azimuths (Figure A.48).

The radial receiver function data from MCH have been modelled using the LISPB Delta model velocity as the starting point. The $H-\kappa$ stacking of the data from MCH reveal a crustal thickness of ~ 35 km and V_p/V_s ratio of 1.77, although the results of the stack have been classed as intermediate quality, because the Moho $PpPs$ phase is not well defined. 1D forward modelling of the Moho P_s phase using the LISPB Delta velocities reveals a crustal thickness of 37 km (Figure 6.29). The prominent phase at ~ 3 s in the observed receiver functions has been modelled using a more complex near surface structure than in the SSP model. The near surface structure consists of a 3 km thick layer of V_p 4.75 - 5.0 km s^{-1} above a 3 km layer of V_p 5.5 km s^{-1} . The LISPB Delta model reports surface layers with velocities of 4.5 km s^{-1} and 5.4 km s^{-1} , but offers no explanation to their origin (Edwards and Blundell, 1984). From the difference in surface geology between SSP and MCH it seems possible that the 4.75 km s^{-1} and 5.5 km s^{-1} layers correspond to the Devonian Old Red sandstone and Lower Palaeozoic sediments of the Welsh Basin respectively. The boundary between the upper and lower crust is constrained to a depth of 17 km by the intra-crustal $PpPs$ phase at ~ 7.5 s. As observed at WCB, CWF and MCH the Moho $PpPs$ phase in the synthetic receiver functions is of much higher amplitude than in the observed receiver functions.

6.5.6 Swindon (SWN), Wolverton (WOL WOB) & Folkestone (TFO)

The stations at Swindon, Wolverton and Folkestone are located on post-Variscan Mesozoic sedimentary sequences (Figure 4.1). The large velocity contrast between the young sediments

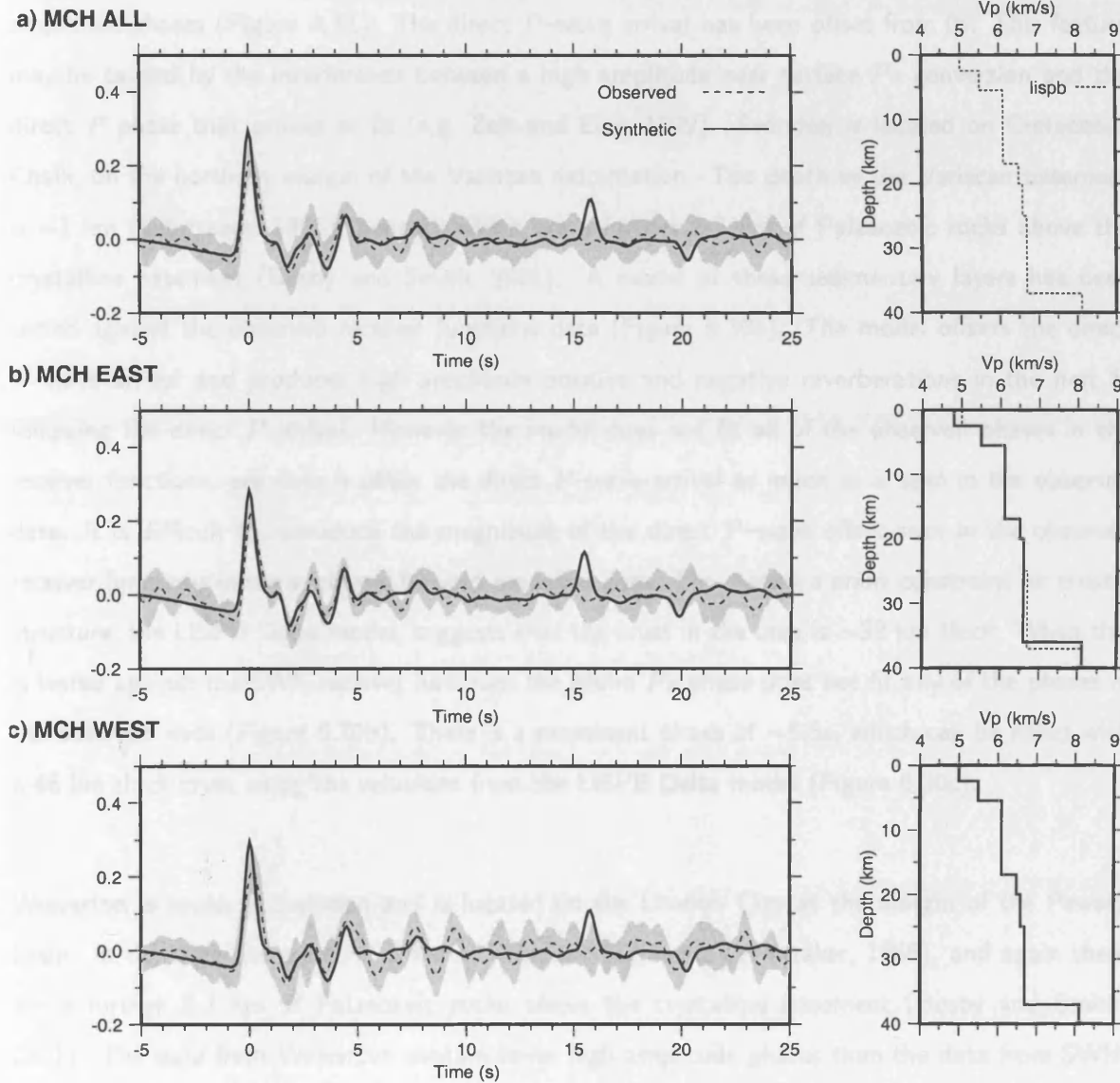


Figure 6.29: Results of 1D forward modelling of receiver function data from MCH for; a) a stack of all the receiver function data, b) the easterly data and c) the westerly data. The data are presented in the same way as Figure 6.3.

and the Variscan basement causes high amplitude multiples that mask the phases from deeper structures, making the data from these stations difficult to interpret.

The stacked radial and tangential receiver functions from SWN are characterised by numerous high amplitude phases (Figure A.51). The direct P -wave arrival has been offset from 0s. This feature may be caused by the interference between a high amplitude near surface Ps conversion and the direct P phase that arrives at 0s (e.g. Zelt and Ellis, 1999). Swindon is located on Cretaceous Chalk, on the northern margin of the Variscan deformation. The depth to the Variscan basement is ~ 1 km (Whittaker, 1985), beneath which are a further 2-3 km of Palaeozoic rocks above the crystalline basement (Busby and Smith, 2001). A model of these sedimentary layers has been tested against the observed receiver functions data (Figure 6.30a). The model offsets the direct P -wave arrival and produces high amplitude positive and negative reverberations in the first 5s following the direct P arrival. However the model does not fit all of the observed phases in the receiver functions, nor does it offset the direct P -wave arrival as much as is seen in the observed data. It is difficult to reproduce the magnitude of the direct P -wave offset seen in the observed receiver functions in the synthetic forward modelled data. The closest *a priori* constraint on crustal structure, the LISPB Delta model, suggests that the crust in the area is ~ 33 km thick. When this is tested against the SWN receiver functions the Moho Ps phase does not fit any of the phases in the observed data (Figure 6.30b). There is a prominent phase at ~ 5.5 s, which can be fitted with a 48 km thick crust using the velocities from the LISPB Delta model (Figure 6.30c).

Wolverton is south of Swindon and is located on the London Clay at the margin of the Pewsey basin. In this area there are ~ 1.5 km of Mesozoic sediments (Whittaker, 1985), and again there are a further 2-3 km of Palaeozoic rocks above the crystalline basement (Busby and Smith, 2001). The data from Wolverton contain fewer high amplitude phases than the data from SWN, but the direct P arrival again remains offset from 0s resulting from interference with a high amplitude phase from a near surface velocity discontinuity (Figure A.53). The H - κ stacking of the data from Wolverton produces a crustal thickness of 39 km and V_p/V_s ratio of 1.79. This identifies the Moho Ps phase at ~ 5 s, with the $PpPs$ phase at ~ 16 s. In the radial receiver functions there is a higher amplitude phase at ~ 13 s, but this produces an unrealistic crustal model which is ~ 32 km thick with a V_p/V_s ratio of ~ 1.95 . 1D forward modelling has been used to fit the offset direct P -wave. As for the model for SWN, this produces a series

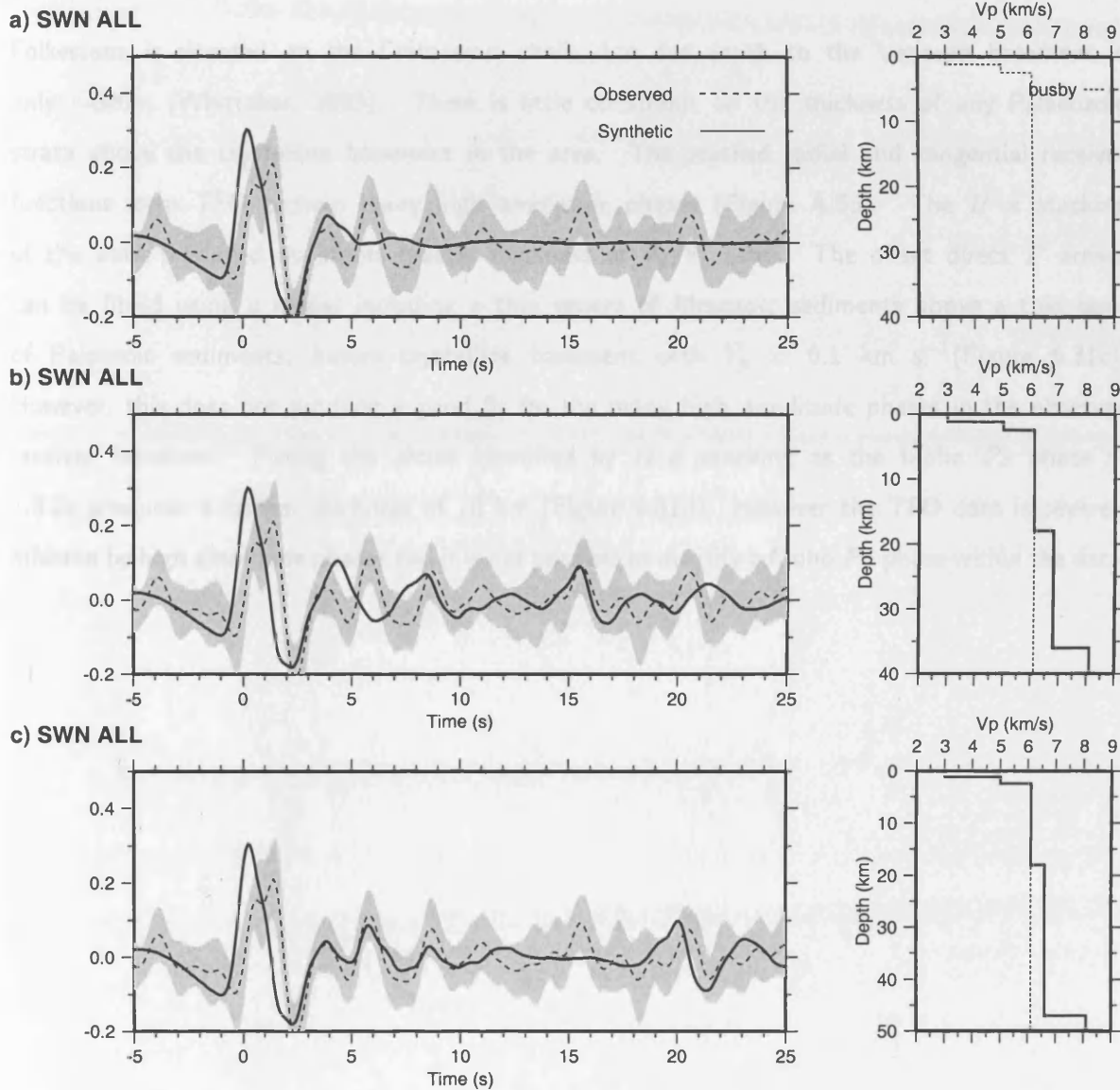
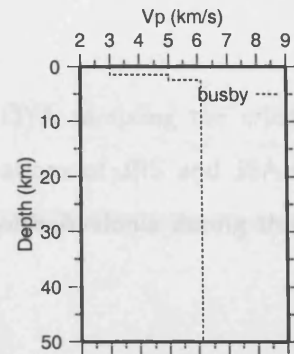
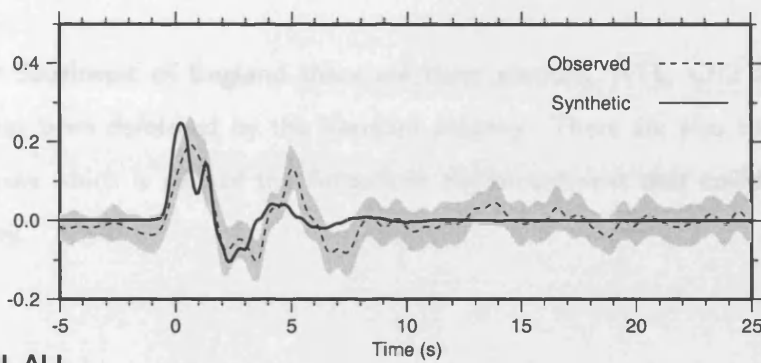


Figure 6.30: Results of 1D forward modelling of receiver function data from SWN for a) - c) of stacks of all of the receiver function data. a) shows the phases resulting from a near surface velocity discontinuity. b) shows the synthetic data from the LISP Delta model. c) shows the model that fits the ~6s phase as the Moho. The data are presented in the same way as Figure 6.3.

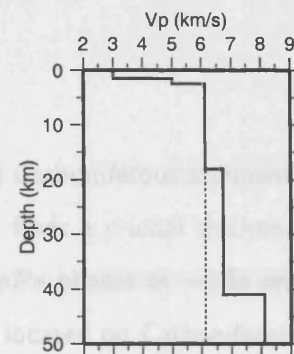
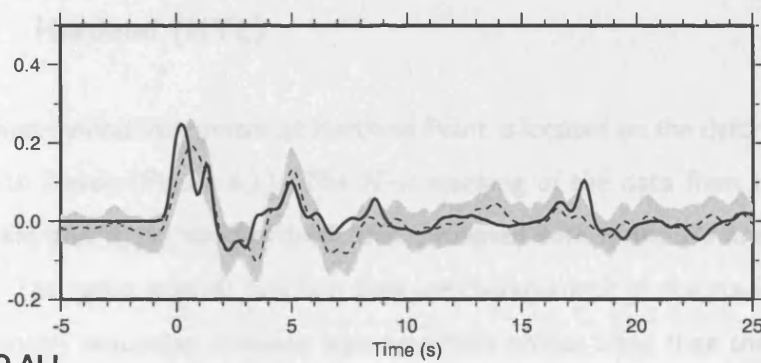
of negative and positive reverberations up to and beyond 5s (Figure 6.31a). The synthetic receiver function from this model does not fit the phase at ~ 5 s. Using the velocities from the LISPB Delta model, this phase can again be fitted using a crustal thickness of 40 km (Figure 6.31b).

Folkestone is situated on the Cretaceous chalk, but the depth to the Variscan basement is only ~ 600 m (Whittaker, 1985). There is little constraint on the thickness of any Palaeozoic strata above the crystalline basement in the area. The stacked radial and tangential receiver functions from TFO contain many high amplitude phases (Figure A.55). The $H-\kappa$ stacking of the data produced no stable crustal thickness or V_p/V_s ratio. The offset direct P arrival can be fitted using a model including a thin veneer of Mesozoic sediments above a thin layer of Paleozoic sediments, before crystalline basement with $V_p = 6.1 \text{ km s}^{-1}$ (Figure 6.31c). However, this does not produce a good fit for the many high amplitude phases in the observed receiver functions. Fitting the phase identified by $H-\kappa$ stacking as the Moho P_s phase at ~ 3.2 s produces a crustal thickness of 28 km (Figure 6.31d). However the TFO data is severely affected by high amplitude phases and it is not possible to identify a Moho P_s phase within the data.

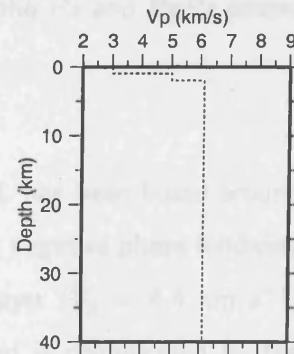
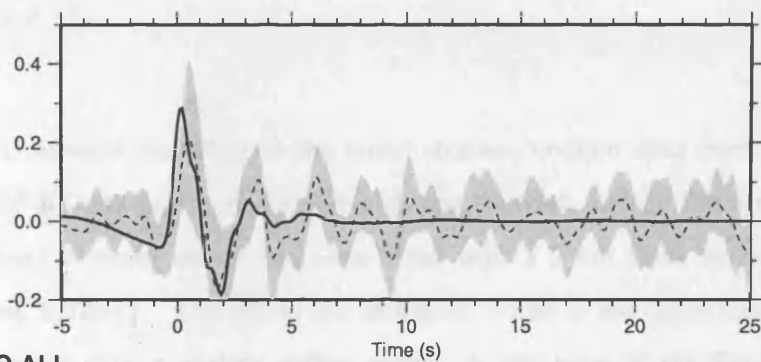
a) WOL ALL



b) WOL ALL



c) TFO ALL



d) TFO ALL

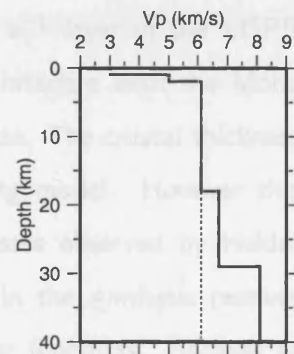
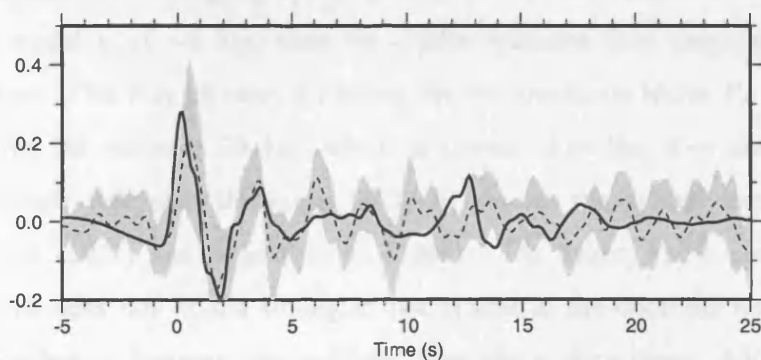


Figure 6.31: Results of 1D forward modelling of receiver function data from WOL & TFO for; a) & b) stacks of all of the WOL receiver functions, and c) & d) stacks of all of the TFO receiver functions. In a) and c) the offset direct *P*-wave arrival is fitted. In b) and d) the phases at 5s and 3.5s are fitted as the Moho *P*_s conversion respectively. The data are presented in the same way as Figure 6.3.

6.6 Southwest England

In the Southwest of England there are three stations, HTL, CR2 and DYA sampling the crust that has been deformed by the Variscan orogeny. There are also the stations of JRS and JSAB on Jersey which is part of the Armorican micro-continent that collided with Avalonia during this orogeny.

6.6.1 Hartland (HTL)

The short-period instrument at Hartland Point is located on the deformed Carboniferous sediments of North Devon (Figure 4.1). The $H-\kappa$ stacking of the data from HTL finds a crustal thickness of 31 km and V_p/V_s ratio of 1.65; this correlated with Moho P_s and $PpPs$ phases at ~ 3.5 s and ~ 13 s. The radial receiver function data are characteristic of the stations located on Carboniferous sedimentary sequences, showing high amplitude phases other than the Moho P_s and $PpPs$ phases (Figure A.57).

The 1D forward modelling of the radial receiver function data from HTL has been based around the LISPB Delta profile model (Edwards and Blundell, 1984). The strong negative phase following the direct P -wave arrival has been fitted with a 2 km thick surface layer ($V_p = 4.4 \text{ km s}^{-1}$) (Figures 6.32a-c). The Moho P_s phase at ~ 3.5 s is low amplitude and is complicated by the interference with a slightly earlier phase. If the base of the 5.4 km s^{-1} layer in the LISPB Delta model is at ~ 5 km, then the $PpPs$ multiples from that layer interfere with the Moho P_s phase. This may be what is causing the low amplitude Moho P_s phase. The crustal thickness from the 1D model is 28 km, which is thinner than the $H-\kappa$ stacking model. However this corresponds well with the range of 27-30 km for the crustal thicknesses observed by Holder and Bott (1971) and Brooks et al. (1984). The Moho $PpPs$ phase in the synthetic receiver functions does not fit the timing of the phases in the observed receiver functions. Because of the interference between the multiples from the surface layers, ΔV_p at the Moho is difficult to constrain using the amplitude of the P_s phase. The boundary between the upper and lower crust in the models is taken from the LISPB Delta model. It is not possible to constrain the depth of this boundary directly from a P_s conversion, but the $PpPs$ phase from this intra-crustal

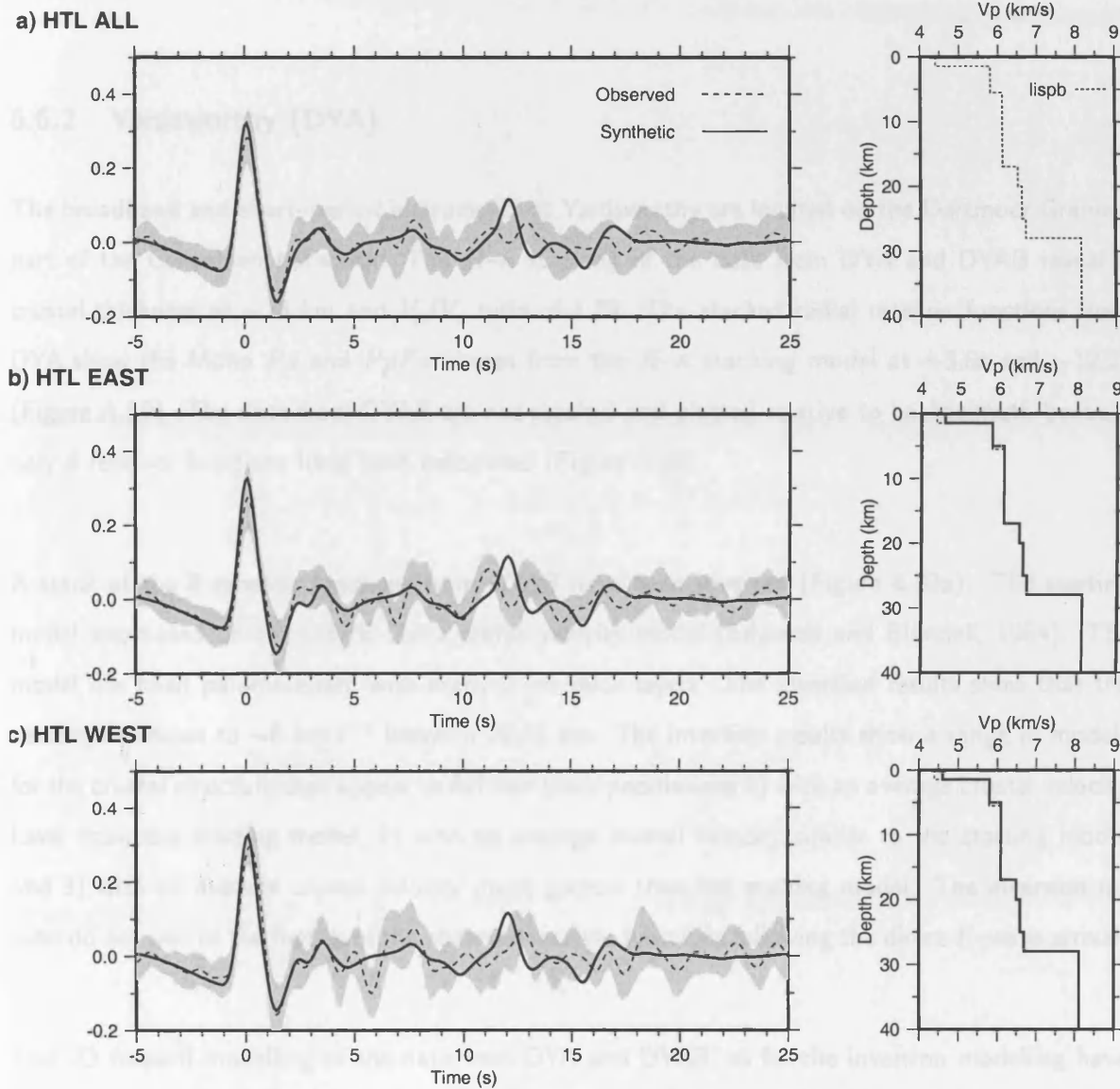


Figure 6.32: Results of 1D forward modelling of receiver function data from HTL for; a) a stack of all the receiver function data, b) the easterly data and c) the westerly data. The data are presented in the same way as Figure 6.3.

discontinuity fits the timing of the phase at ~ 7.5 s in the observed receiver functions. This results in a discontinuity at ~ 17 km. This interface is slightly deeper than the base of the granite found by Holder and Bott (1971), but corresponds with the intra-crustal interface from Brooks et al. (1984) at 10-15 km, which they interpreted as a major thrust relating to the Variscan orogeny.

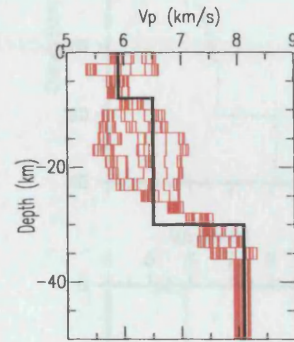
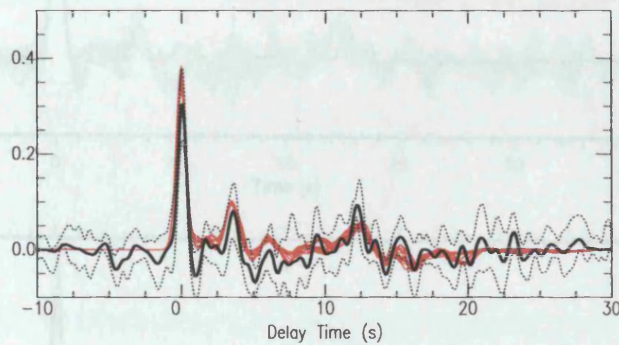
6.6.2 Yardsworth (DYA)

The broadband and short-period instruments at Yardsworth are located on the Dartmoor Granite, part of the Cornubian batholith. The $H-\kappa$ stacking of the data from DYA and DYAB reveal a crustal thickness of ~ 28 km and V_p/V_s ratio of 1.79. The stacked radial receiver functions from DYA show the Moho P_s and $PpPs$ phases from the $H-\kappa$ stacking model at ~ 3.8 s and ~ 12.2 s (Figure A.59). The data from DYAB are not stacked and plotted relative to backazimuth because only 8 receiver functions have been calculated (Figure A.60).

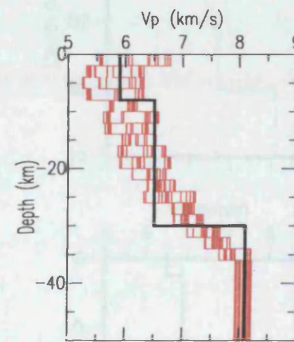
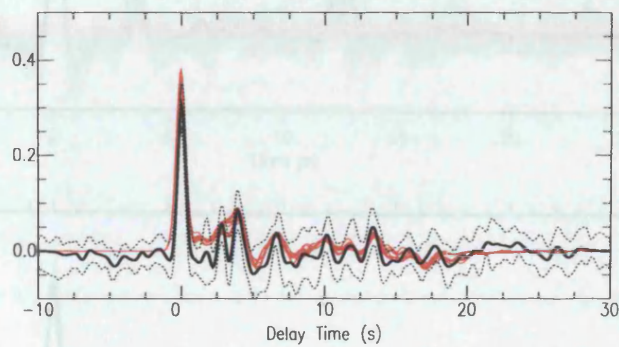
A stack of the 8 receiver functions from DYAB have been inverted (Figure 6.33a). The starting model was based on the LISPB Delta profile velocity model (Edwards and Blundell, 1984). The model has been parameterised with many 2 km thick layers. The inversion results show that the velocity increases to ~ 8 km s $^{-1}$ between 25-35 km. The inversion results show a range of models for the crustal structure that appear to fall into three populations 1) with an average crustal velocity lower than the starting model, 2) with an average crustal velocity similar to the starting model and 3) with an average crustal velocity much greater than the starting model. The inversion results do not well fit the first 2s of the observed receiver functions following the direct P -wave arrival.

The 1D forward modelling of the data from DYA and DYAB, as for the inversion modelling have been based around the LISPB Delta profile velocity model. The velocity of the upper crustal layer has been constrained by the model from Holder and Bott (1971). The modelling of the stack of all of the receiver functions from DYAB shows a crustal thickness of 30 km and an intra-crustal velocity discontinuity at 14 km (Figure 6.34a). The intra-crustal discontinuity corresponds well with the depth of the R2 reflector of Brooks et al. (1984) which is associated with Variscan orogeny, the crustal thicknesses also fit well with the range 27-30 km from Brooks et al. (1984). The same model is used to fit the Moho P_s phase for the stack of all data from DYA (Figure

a) DYAB ALL



b) JSAB ALL



c) JSAB ALL

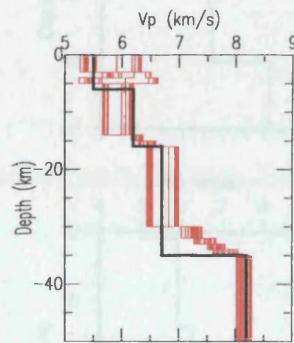
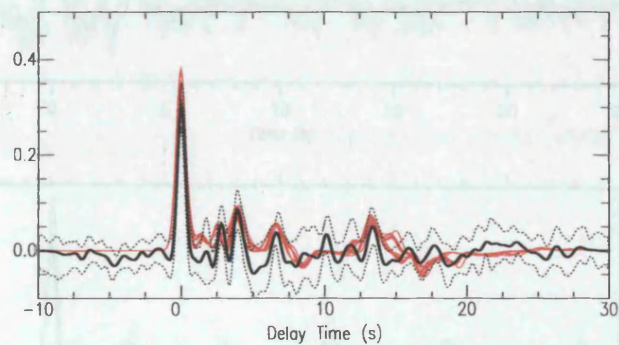


Figure 6.33: Results of 1D inverse modelling of receiver function data from DYA & JSAB for a) a stack of all of the DYA receiver functions, and b) & c) stacks of all of the JSAB receiver functions. In b) the starting model uses many thin layers, whereas in c) the starting model has been parameterised with fewer layers of fixed thickness. The data are presented in the same way as Figure 6.1.

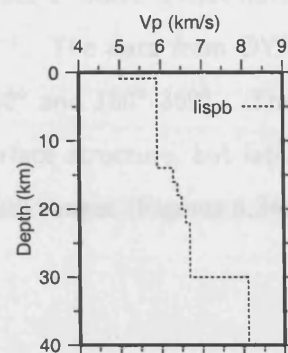
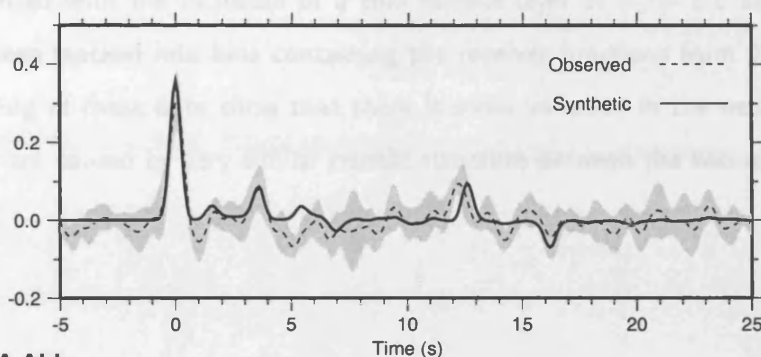
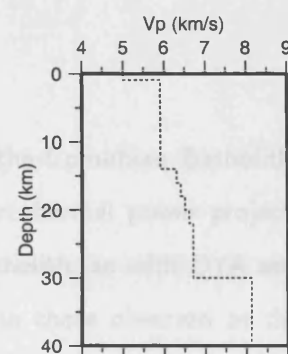
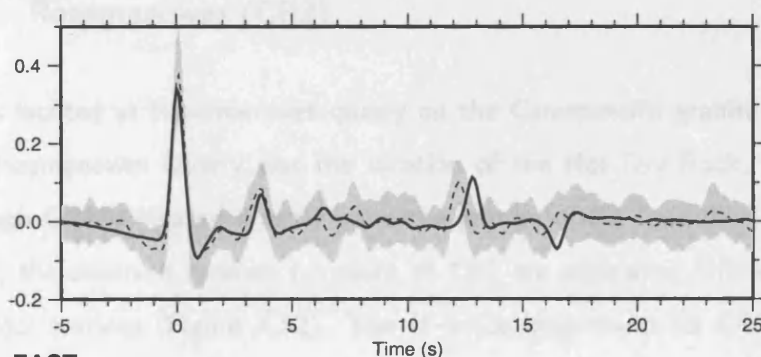
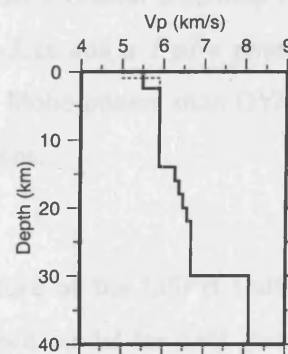
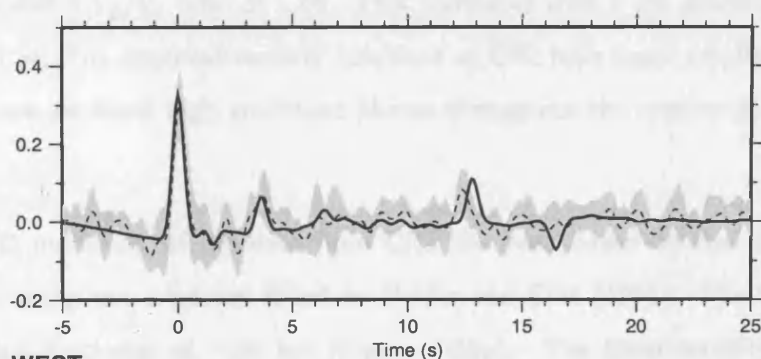
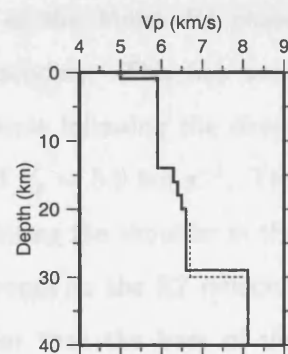
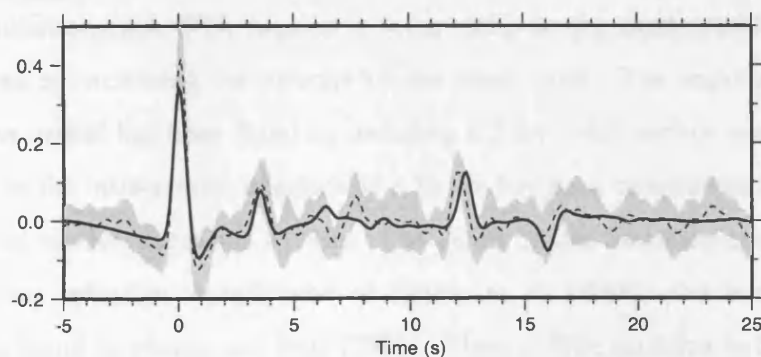
a) DYAB ALL**b) DYA ALL****c) DYA EAST****d) DYA WEST**

Figure 6.34: Results of 1D forward modelling of receiver function data from DYA for; a) & b) stacks of all of the receiver functions, c) the easterly data and d) the westerly data. The data are presented in the same way as Figure 6.3.

6.34b). The first 2s of the observed receiver functions following the direct P -wave arrival have been fitted with the inclusion of a thin surface layer of $V_p = 5.0 \text{ km s}^{-1}$. The data from DYA have been stacked into bins containing the receiver functions from 0° - 180° and 180° - 360° . The modelling of these data show that there is some variation in the near surface structure, but later phases are caused by very similar crustal structure between the two azimuth ranges (Figures 6.34c & d).

6.6.3 Rosemanowes (CR2)

CR2 is located at Rosemanowes quarry on the Carnmenellis granite of the Cornubian Batholith. The Rosemanowes Quarry was the location of the Hot Dry Rock, hydrothermal power project. Although CR2 is located directly upon the granite of the Cornubian batholith, as with DYA and DYAB, the observed receiver functions at CR2 are somewhat different to those observed at the Dartmoor stations (Figure A.62). The $H-\kappa$ stacking results for CR2 show a crustal thickness of 28 km and a V_p/V_s ratio of 1.65. This correlates with a Ps phase at $\sim 3.1\text{s}$ and a $PpPs$ phase at $\sim 11.5\text{s}$. The observed receiver functions at CR2 have lower amplitude Moho phases than DYA, and there are fewer high amplitude phases throughout the receiver functions.

The 1D modelling of the data from CR2 has been based on the structure of the LISP Delta profile, using the velocities found by Holder and Bott (1971). The forward model for CR2 gives a crustal thickness of $\sim 25 \text{ km}$ (Figure 6.35a). The lower amplitude of the Moho Ps phase in comparison with DYA requires a lower ΔV_p at the crust mantle boundary. This has been achieved by increasing the velocity of the lower crust. The negative phase following the direct P -wave arrival has been fitted by including a 2 km thick surface layer of $V_p = 5.0 \text{ km s}^{-1}$. The depth to the intra-crustal interface at $\sim 15 \text{ km}$ has been constrained by fitting the shoulder in the observed receiver functions at $\sim 2\text{s}$. The depth of this boundary corresponds to the R2 reflector in seismic reflection investigation of Brooks et al. (1984), and is deeper than the base of the granite found by Holder and Bott (1971). There is little variation in the models required to fit the observed data from a range of backazimuths (Figures 6.35b & c).

6.6.4. Mahon St. Louis (JHS) & St. Andrew (JSA2)

The short-period broadband at Mahon St. Louis and broadband instrument at St. Andrew are located on Jersey. A deep seismic reflector profile along the east of the BOST 3D reflection profile

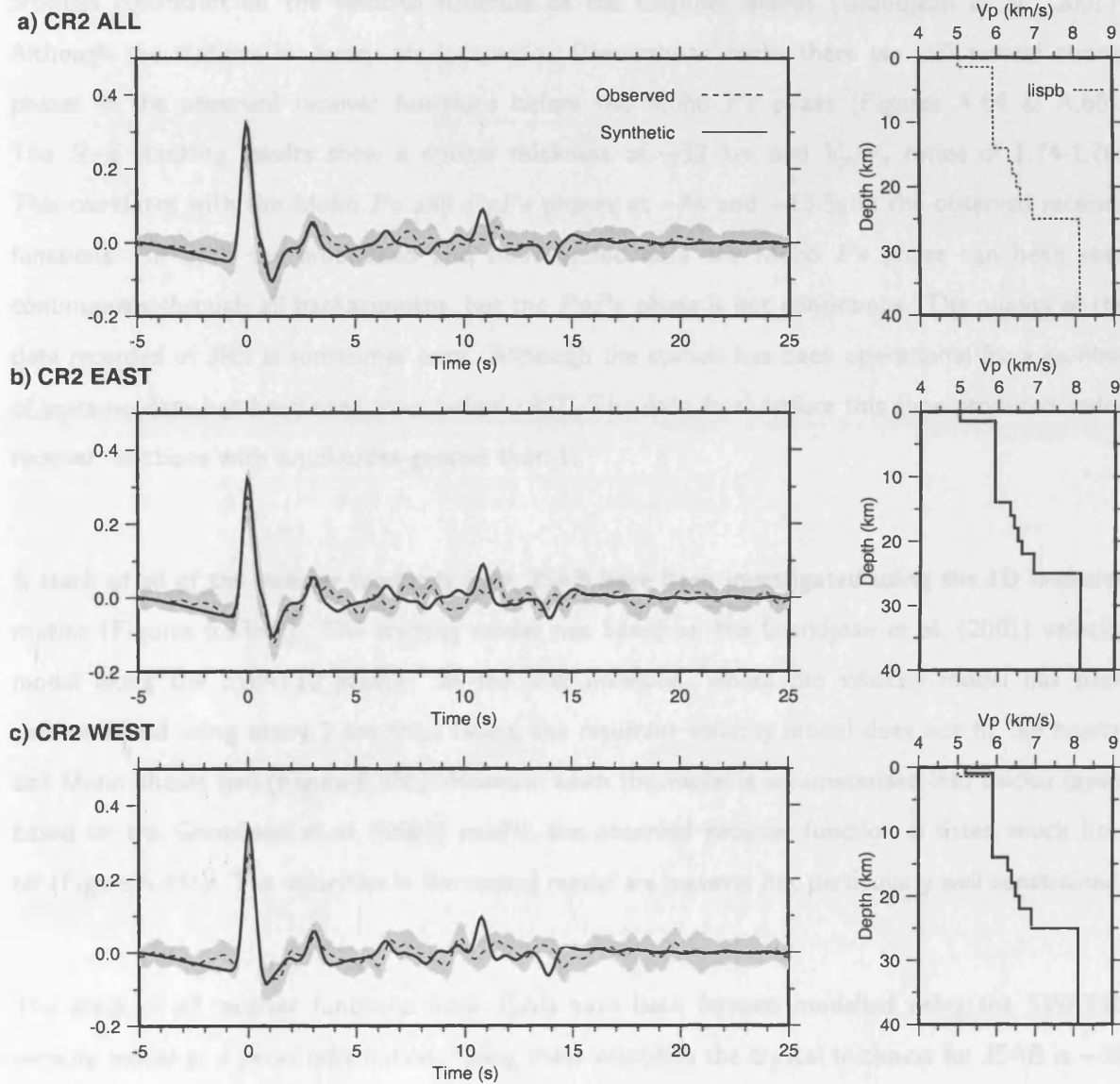


Figure 6.35: Results of 1D forward modelling of receiver function data from CR2 for; a) a stack of all of the receiver functions, b) the easterly data and c) the westerly data. The data are presented in the same way as Figure 6.3.

6.6.4 Maison St. Louis (JRS) & St. Aubins (JSAB)

The short-period instrument at Maison St. Louis and broadband instrument at St. Aubins are located on Jersey. A deep seismic refraction profile along the line of the SWAT10 reflection profile provides constraint on the velocity structure of the Channel Islands (Grandjean et al., 2001). Although the stations in Jersey are located on Precambrian rocks there are still several strong phases in the observed receiver functions before the Moho P_s phase (Figures A.64 & A.66). The $H-\kappa$ stacking results show a crustal thickness of ~ 32 km and V_p/V_s ratios of 1.74-1.76. This correlates with the Moho P_s and $PpPs$ phases at ~ 4 s and ~ 13.5 s in the observed receiver functions. In both the broadband and short-period data the Moho P_s phase can be seen continuously through all backazimuths, but the $PpPs$ phase is not continuous. The quality of the data recorded at JRS is sometimes poor. Although the station has been operational for a number of years no data has been used from before 1997. The data from before this time produced radial receiver functions with amplitudes greater than 1.

A stack of all of the receiver functions from JSAB have been investigated using the 1D inversion routine (Figures 6.33b,c). The starting model was based on the Grandjean et al. (2001) velocity model along the SWAT10 profile. In the first inversion, where the velocity model has been parameterised using many 2 km thick layers, the resultant velocity model does not fit the crustal and Moho phases well (Figure 6.33b). However when the model is parameterised into thicker layers based on the Grandjean et al. (2001) model, the observed receiver function is fitted much better (Figure 6.33c). The velocities in the second model are however not particularly well constrained.

The stack of all receiver functions from JSAB have been forward modelled using the SWAT10 velocity model as *a priori* information. Using these velocities the crustal thickness for JSAB is ~ 33 km (Figure 6.36a). The depth to the top of the lower crust has been constrained at ~ 15 km by the $PpPs$ phase at ~ 7 s. The stacks of the JRS data from 270° - 300° , 0° - 30° and 90° - 120° show that there is some azimuthal variation in the observed receiver functions (Figures 6.36a-c). There is some variation in the phases between the direct P -wave arrival and the Moho P_s phase that can be explained by changes in the near surface structure. However, there are large changes in the amplitude of the Moho P_s phase, which when modelled by 1D structures require large changes in ΔV_p at the base of the crust. It is therefore likely that these changes in amplitude are related to

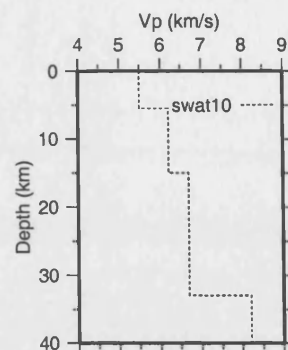
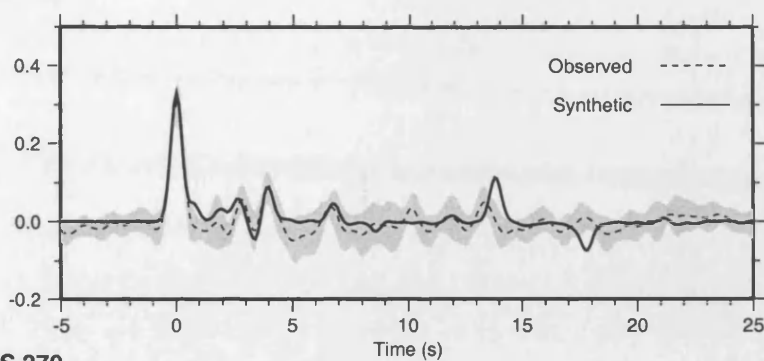
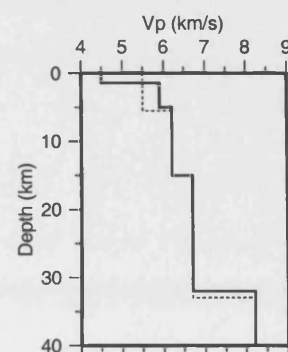
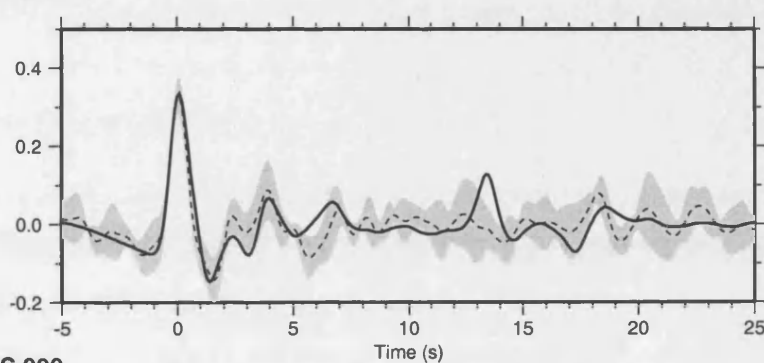
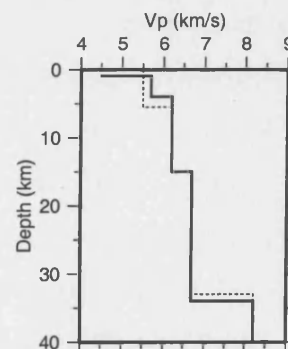
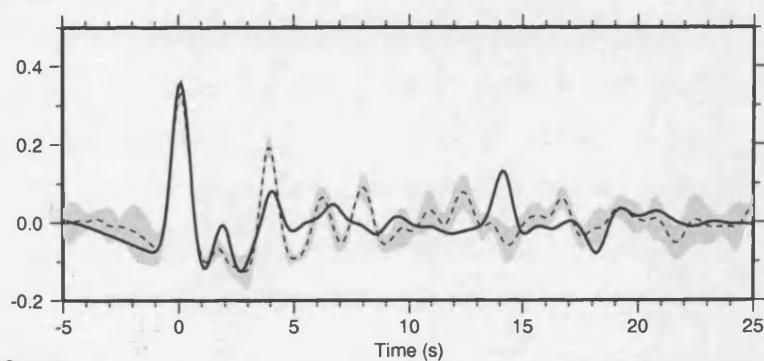
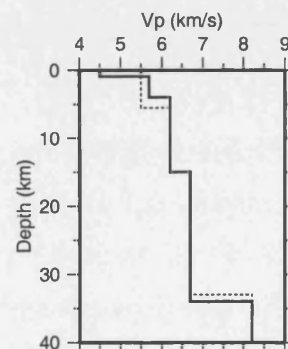
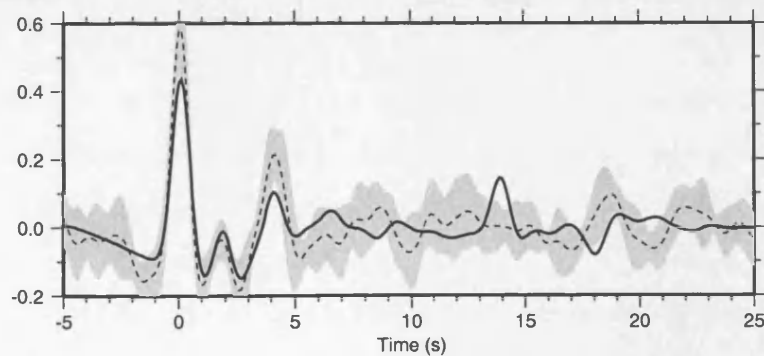
a) JSAB ALL**b) JRS 270****c) JRS 000****d) JRS 090**

Figure 6.36: Results of 1D forward modelling of receiver function data from JSAB & JRS for; a) a stack of all of the receiver functions from JSAB, b) the receiver functions between 270°-300° at JRS, c) the receiver functions between 0°-30° at JRS and d) the receiver functions between 90°-120° at JRS. The data are presented in the same way as Figure 6.3.

non-1D structure.

6.7 Discussion

6.7.1 Receiver Function Modelling

The 1D modelling that has been performed in this study has used velocity models based upon those derived from the LISPB, CSSP, JUNE92 and SWAT10 wide-angle seismic refraction profiles (Barton, 1992; Al-Kindi, 2002; Price and Morgan, 2000; Morgan et al., 2000; Grandjean et al., 2001). There are three crustal model variables which control the observed and synthetic receiver functions: 1) the depth to seismic velocity interfaces, 2) the P -wave velocity 3) the S -wave velocity (or V_p/V_s ratio). A change in any of these variables will result in a change in the synthetic receiver function (Figures 6.37a-c).

Figure 6.37d shows the non-uniqueness in the modelling of receiver functions due to the velocity-depth tradeoff, displaying that it is possible to fit a Moho Ps phase at 3.5s with a range of models with differing crustal thicknesses and V_p/V_s ratios. However, the use of *a priori* velocity information from seismic refraction experiments in this modelling study helps to constrain the range of possible models fitting the observed receiver function data. The strategy of incorporating this *a priori* information into the forward models is subjective, and may result in models that contain structures which are not specifically required to fit the observed data. Performing unconstrained linear inversions on the broadband receiver function data has provided an estimate of the significance of the *a priori* structures which have been incorporated into the 1D forward models. The variability of the stacked receiver functions that have been modelled in this study has been estimated by calculating the standard deviation of each point in the receiver function stack.

At some of the stations used in this study there are co-located broadband and short-period instruments. It has been shown that by modelling short-period data as broadband data it is possible to develop false velocity structures. To avoid this happening when modelling short-period receiver functions, the synthetic seismograms have been filtered to reproduce the frequency spectrum of the short-period data. Using this method it has been shown that for the data from the stations where short-period and broadband instruments are co-located, it is possible to fit the receiver functions from both instruments with the same model. This shows that the models derived from receiver functions from stations with only short-period instruments can be representative of crustal velocity structure.

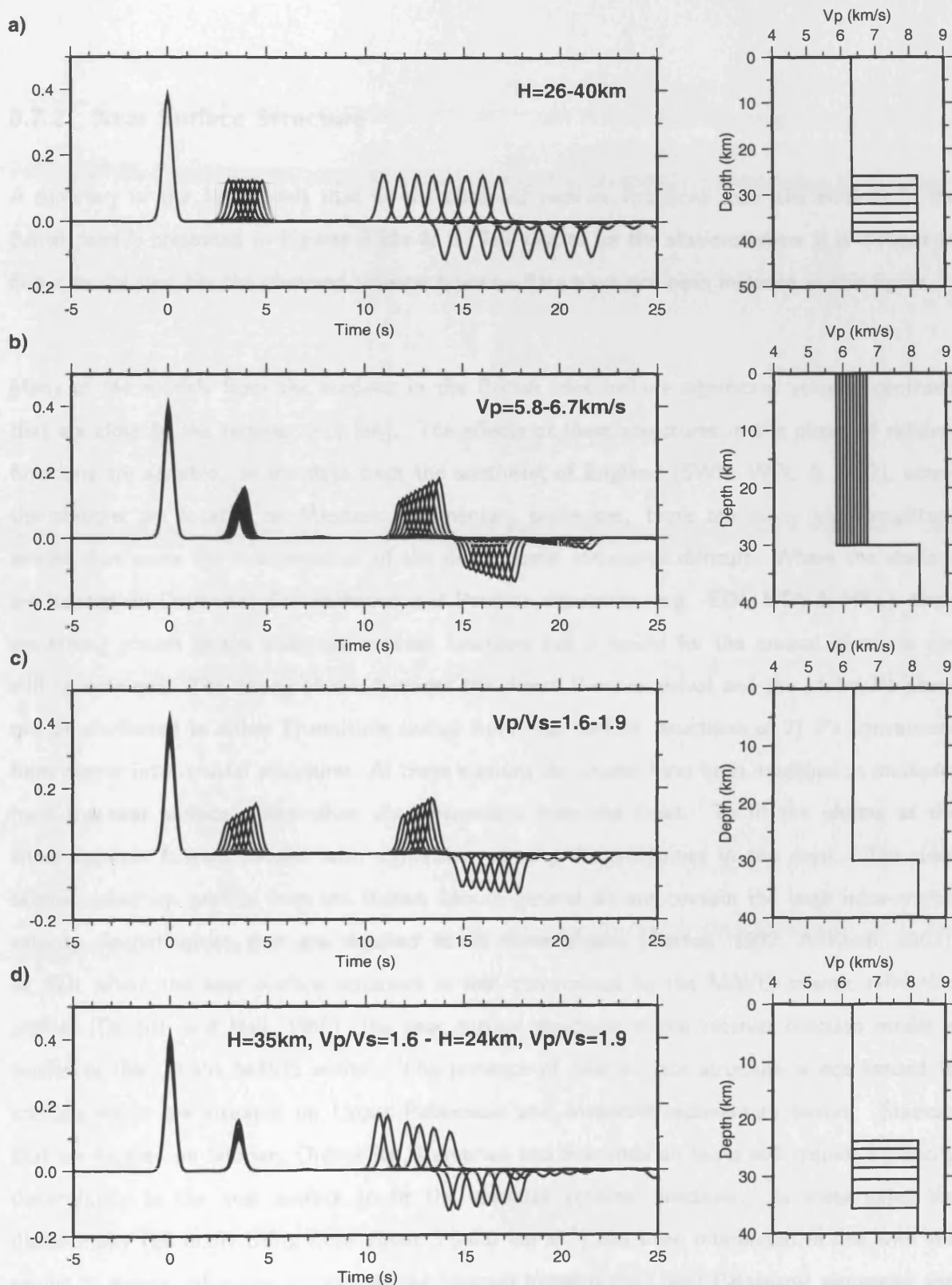


Figure 6.37: a) The variation of synthetic receiver function with crustal thickness. b) The variation of synthetic receiver function with crustal velocity. c) The variation of synthetic receiver function with V_p/V_s ratio. d) The variation of synthetic receiver function with crustal thickness and V_p/V_s ratio. This shows that a Moho P_s phase at 3.5s can be fitted using a range of models from $H=35\text{km}$ and $V_p/V_s=1.6$ through to $H=24\text{km}$ and $V_p/V_s=1.9$.

6.7.2 Near Surface Structure

A summary of the 1D models that fit the observed receiver functions from the stations in the British Isles is presented in Figures 6.38a & b. The results for the stations where it is difficult to find a model that fits the observed receiver function data have not been included in this figure.

Many of the models from the stations in the British Isles include significant velocity contrasts that are close to the receiver (<5 km). The effects of these structures in the observed receiver functions are variable. In the data from the southeast of England (SWN, WOL & TFO), where the stations are located on Mesozoic sedimentary sequences, there are many high amplitude phases that make the interpretation of the deep crustal structures difficult. Where the stations are located on Devonian, Carboniferous and Permian sequences (e.g. EDI, BTA & HTL), there are strong phases in the observed receiver functions but a model for the crustal structure can still be obtained. The strong phases between the direct P -wave arrival and the Moho P_s phase can be attributed to either 1) multiple energy from near surface structures or 2) P_s conversions from deeper intra-crustal structures. At these stations the phases have been modelled as multiples from the near surface, rather than P_s conversions from the crust. To fit the phases as the latter requires forward models with significant velocity discontinuities in the crust. The deep seismic refraction profiles from the British Isles in general do not contain the large intra-crustal velocity discontinuities that are required to fit these phases (Barton, 1992; Al-Kindi, 2002). At EDI where the near surface structure is well constrained by the MAVIS seismic refraction profiles (Dentith and Hall, 1989), the near surface structure in the receiver function model is similar to that in the MAVIS model. The presence of near surface structure is not limited to stations which are situated on Upper Palaeozoic and Mesozoic sedimentary basins. Stations that are located on Silurian, Ordovician, Cambrian and Precambrian rocks still require a velocity discontinuity in the near surface to fit the observed receiver functions. In these cases the discontinuity (generally being from about $5.5\text{--}6.0$ km s^{-1}) has been interpreted in line with the results of seismic refraction surveys, as the contrast between the Lower Palaeozoic sequences and the crystalline basement rocks (e.g. Bott et al., 1985; Bamford et al., 1978; Grandjean et al., 2001).

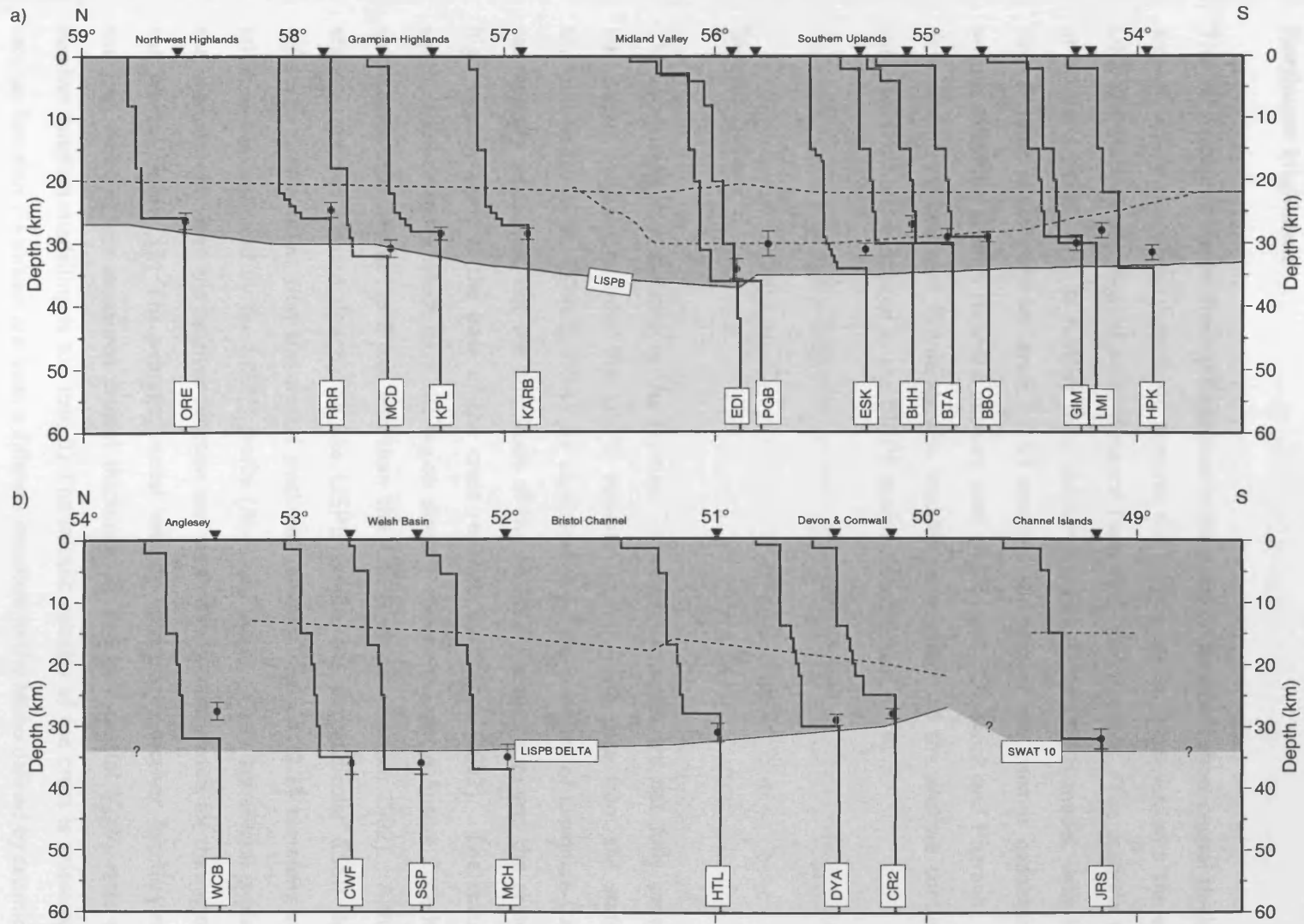


Figure 6.38: The results of the 1D receiver function modelling plotted as two north-south sections. The models have been plotted with the mantle layer located directly beneath the station. The LISP velocity model (a) (Barton, 1992) and the LISP Delta profile (b) (Edwards and Blundell, 1984) and SWAT10 velocity model (Grandjean et al., 2001) have been plotted beneath the receiver function 1-D models. The black dots show the results of the $H-\kappa$ stacking for each station $\pm 1\sigma$.

6.7.3 Crustal Structure

Northwest Highlands

The 1D modelling results from the stations in the north of Scotland show crustal thicknesses that are not totally consistent with those beneath the LISPb profile. However, with the exception of ORE the stations are located some distance from the LISPb profile. The stations to the west of Scotland (RRR, KPL & KARB) show thinner crust than the LISPb model, with MCD in the east showing slightly thicker crust. This westerly thinning of the crust is consistent with the seismic reflection profiles recorded offshore west of Scotland (Chadwick and Pharaoh, 1998). The data from the Northwest Highlands also enables investigation of the onshore continuity of the sub-crustal structure imaged in the BIRPS profiles (see Section 6.7.4)

Iapetus Suture

The models for the stations in the Iapetus Suture region again are not fully consistent with the crustal thicknesses from the LISPb velocity model. The data from the stations in the Midland Valley (EDI, EDIB & PGB) are complicated by the presence of Devonian-Carboniferous sedimentary sequences, but the amplitude of the Moho P_s phases supports the presence of the high velocity layer at the base of the crust reported by Barton (1992). The stations to the south of Eskdalemuir which span the region directly above the Iapetus Suture Zone have crustal thicknesses that are up to 5 km less than the LISPb model (Barton, 1992). Although these stations are not located directly on the LISPb profile, the perpendicular CSSP deep seismic refraction profile shows that the crustal thickness remains between 32-34 km along strike of the structures investigated by the LISPb profile (Al-Kindi, 2002). There are several explanations for the mismatch between the receiver function and seismic refraction models for the region including the Iapetus Suture: 1) The average crustal velocity used in the receiver function modelling is too low, reducing the measured crustal thickness; 2) The bulk crustal V_p/V_s ratio used in the receiver function modelling is too low; 3) The seismic velocity of the crust is anisotropic; 4) The receiver function P_s phases are from a different interface to the Moho defined by seismic refraction.

By subtracting Equation 3.28 from Equation 3.27 and rearranging it is possible to define the average crustal V_p with respect to the arrival time of the Moho P_s (T_{Ps}) and $PpPs$ (T_{PpPs}) phases, the crustal thickness (H) and the ray parameter of the teleseismic event (ρ);

$$V_p = \frac{1}{\frac{(T_{PpPs} - T_{Ps})^2}{4H^2} + \rho^2} \quad (6.1)$$

The V_p/V_s ratio can be calculated using Equation 3.27 once V_p has been calculated. By fixing the value of H and using the times of the Moho P_s and $PpPs$ phase from the results of $H-\kappa$ stacking it is possible to test the P -wave velocity that is required to make the receiver function data from the Iapetus Suture area fit the crustal thickness data from the LISPb and CSSP seismic refraction experiments. To fit a crustal thickness of 33 km with the data from BTA the average crustal P -wave velocity must be $\sim 7.1 \text{ km s}^{-1}$, with a V_p/V_s ratio of 1.74. To fit the same crustal thickness model with the data from BHH, V_p must increase to $\sim 7.5 \text{ km s}^{-1}$, with a V_p/V_s ratio of 1.79. These values for V_p represent a significant increase in average crustal P -wave velocity. In the results of the deep seismic refraction studies of the British Isles the average crustal P -wave velocity varies between $6.2\text{--}6.6 \text{ km s}^{-1}$, and it therefore seems very unlikely that an average velocity of between $7\text{--}7.5 \text{ km s}^{-1}$ is reached beneath the Iapetus Suture region. Fixing the crustal thickness, H , allows a unique solution for both V_p and V_p/V_s to be calculated. These calculations show that the V_p/V_s ratios for BTA ($H-\kappa$ stacking $V_p/V_s = 1.75$) and BHH ($H-\kappa$ stacking $V_p/V_s = 1.81$) are only altered slightly relative to the $H-\kappa$ stacking results when H is fixed. This again highlights the fact that although there is a velocity-depth trade off in the modelling of receiver function data, the calculated V_p/V_s ratio is relatively insensitive to changes in H and V_p .

The possibility of seismic velocity anisotropy causing the mismatch between the seismic refraction and receiver function models has been tested using anisotropic synthetic receiver function code (Levin and Park, 1997, 1998). The raypaths of the phases in receiver function studies are close to vertical, whereas the raypaths of phases constraining the velocities in seismic refraction studies are predominantly horizontal. If the seismic velocity of the rocks is anisotropic then it is likely a mismatch between the receiver function and seismic refraction results may occur if the data are interpreted without considering this anisotropy. As shown by the fixed H calculation, the observed receiver functions require a V_p of $7\text{--}7.5 \text{ km s}^{-1}$ to fit the crustal thickness of 33 km. This is much greater than the velocities observed in the LISPb and CSSP profiles. If the rocks in the Iapetus Su-

ture area are anisotropic this suggests that the fast axis of the of the anisotropy ellipsoid is vertical, therefore decreasing the traveltimes of the near vertical receiver function phases through the crust. The receiver function data from BTA, BBO and BHH have been modelled using the anisotropic code (Figure 6.39a-c). The aim of the modelling was to produce an estimate of bulk crustal velocity anisotropy by fitting the delay times of the Moho phases. These models show that timing of the Moho phases can be fitted using a crustal thickness of 33 km and horizontal P -wave velocity of 6.4 km s^{-1} . To fit the receiver functions using this model there must be a bulk crustal P -wave and S -wave anisotropy of 20-25%, with the fast axis of the anisotropy ellipsoid being vertical.

It is common for rocks to show up to 10% seismic velocity anisotropy, and in some lithologies this can exceed 15% (Levin and Park, 1998). The bulk crustal velocity anisotropy of 20-25% found in the receiver function models is somewhat greater than these likely maximum anisotropy values. Furthermore, if the level of anisotropy varies within the crust then some of the crust must exceed 20-25% anisotropy. Seismic velocity anisotropy can be caused by several mechanisms. In the upper crust the main cause of anisotropy is thought to be the presence of aligned cracks and pore space. In the lower crust it is assumed that the cracks are closed and lattice preferred orientation of mineral crystals is the main cause of anisotropy (Levin and Park, 1998). Jones et al. (1996) study the mismatch between seismic reflection and seismic refraction results offshore the north of Scotland. They find that the seismic reflection Moho is deeper than the seismic refraction Moho. A bulk crustal seismic velocity anisotropy of $\sim 7\%$ is required to eliminate the mismatch between the two datasets. The fast axis of anisotropy ellipsoid is horizontal. If the axis of symmetry of the ellipsoid is vertical then the anisotropy would be characteristic of that caused by horizontal layering (Levin and Park, 1998). This is in accordance with the Typical BIRP which shows a lower crust that contains many sub-horizontal layers (McGeary et al., 1987). The anisotropy required to explain the miss-match between the crustal thicknesses derived from the receiver function and wide angle seismic refraction data would have to be caused by a different mechanism to the anisotropy observed beneath the North of Scotland.

It seems unlikely, given the change in bulk crustal V_p and the magnitude of the seismic velocity anisotropy required to fit the observed receiver functions with a crustal thickness of 33 km, that receiver function phases are caused by the seismic refraction Moho defined by Barton (1992) and Al-Kindi (2002). The area around the Irish Sea, close to the Iapetus Suture Zone, has been

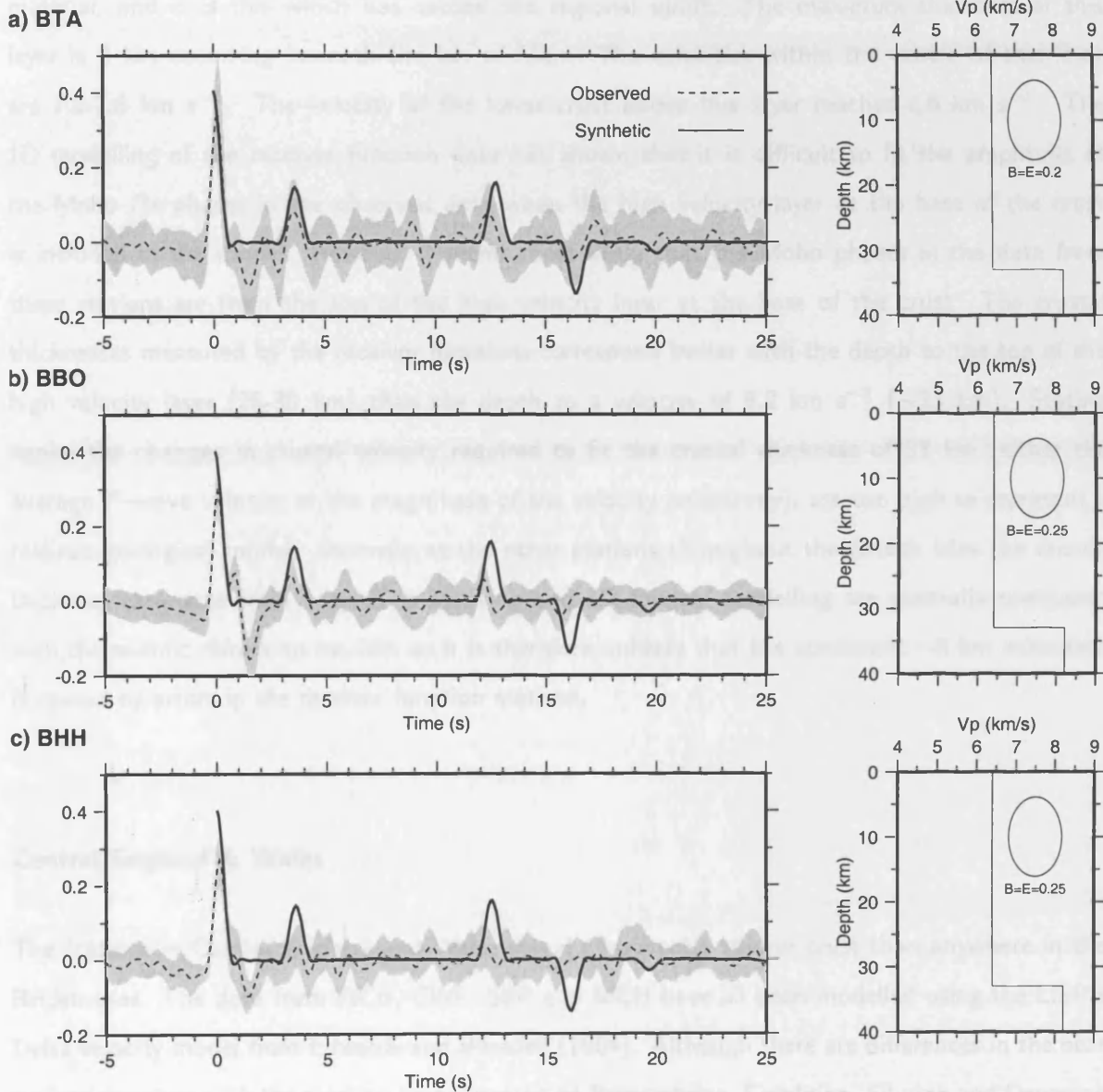


Figure 6.39: The results of the modelling of receiver function data from the Iapetus Suture zone using the seismic velocity anisotropy code of Levin and Park (1998). a) A model with anisotropy ellipse with a vertical axis of symmetry and $B=E=0.2$ is required to fit the timing of the phases in the observed receiver functions at BTA if the crustal thickness = 33 km and $V_p = 6.4 \text{ km s}^{-1}$. b) The data from BHH requires an anisotropy model with $B=E=0.25$. c) The data from BBO requires an anisotropy model with $B=E=0.25$.

significantly uplifted during the Cenozoic (Jones et al., 2002). Al-Kindi (2002) models a well shaped high velocity layer in the CSSP/ICCSP data which is centred upon the uplift maximum in the middle of the Irish Sea. He concludes that this high velocity body is underplated magmatic material, and it is this which has caused the regional uplift. The maximum thickness of this layer is 8 km occurring beneath the Isle of Man. The velocities within the centre of this layer are $7.2\text{--}7.8 \text{ km s}^{-1}$. The velocity of the lower crust above this layer reaches 6.6 km s^{-1} . The 1D modelling of the receiver function data has shown that it is difficult to fit the amplitude of the Moho P_s phases in the observed data when the high velocity layer at the base of the crust is included in the model. However it seems most likely that the Moho phases in the data from these stations are from the top of the high velocity layer at the base of the crust. The crustal thicknesses measured by the receiver functions correspond better with the depth to the top of the high velocity layer (25–30 km) than the depth to a velocity of 8.2 km s^{-1} ($\sim 33 \text{ km}$). Stating again, the changes in crustal velocity required to fit the crustal thickness of 33 km (either the average P -wave velocity or the magnitude of the velocity anisotropy), are too high to represent a realistic geological model. Secondly at the other stations throughout the British Isles the crustal thickness estimates from both $H\text{--}\kappa$ stacking and 1D forward modelling are generally consistent with the seismic refraction models, so it is therefore unlikely that the consistent $\sim 5 \text{ km}$ mismatch is caused by errors in the receiver function method.

Central England & Wales

The stations in Central England and Wales reveal consistently thicker crust than anywhere in the British Isles. The data from WCB, CWF, SSP and MCH have all been modelled using the LISP Delta velocity model from Edwards and Blundell (1984). Although there are differences in the near surface structure with the stations being located on Precambrian, Cambrian, Silurian and Devonian rocks, the crustal structure remains similar. The crust from the LISP Delta model is divided into two layers, an upper and lower crust. This structure is maintained in the receiver function models (Figure 6.38b). The boundary between the upper and lower crust has been constrained by the intra-crustal $PpPs$ phases at between 7–8s. A characteristic of the data from these stations is that the Moho $PpPs$ phase is low amplitude and difficult to identify in the observed receiver functions. In some cases this has made the results of the $H\text{--}\kappa$ stacking unstable. In particular it was noted

that the $H-\kappa$ stacking of the data from WCB produces an unusually thin, high V_p/V_s ratio crustal model. The 1D modelling of the data from WCB using the default V_p/V_s ratio of 1.73 reveals a crustal thickness of ~ 32 km, which is much closer to the observed values at the northern end of the LISPB Delta profile (34 km) and seismic reflection data reported by Chadwick and Pharaoh (1998) (32 km). If the crust at WCB is similar to that observed at the other stations in Wales then it would be expected, as observed, that the Moho $PpPs$ phase is weak. This may be the cause of the unusual $H-\kappa$ stacking result from WCB. As already noted in Section 6.7.2 the interpretation of the data from the southeast of England has been made difficult owing to the presence of near surface sedimentary layers. It is not possible to comment on the continuity of the thickened crust into these areas using the receiver function data.

Southwest England

The stations to the south of the Variscan front show a crust that is 4-8 km thinner than the 36 km reached in central England and Wales. This thinner crust is consistent with the observations along the LISPB Delta profile (Edwards and Blundell, 1984) and the seismic reflection Moho map (Chadwick and Pharaoh, 1998). At HTL the near surface Carboniferous sedimentary sequences complicate the observed receiver functions. At DYA and CR2 where the instruments are located on the Cornubian granite batholith, there is still some near surface structure. Holder and Bott (1971) find that the granite extends to between 10-12 km depth, so these structures must occur within the granite. In the models for HTL, DYA and CR2 there is an intra-crustal structure between 14-17 km depth. This range is somewhat deeper than the granite observed by Holder and Bott (1971), and may correlate with the R2 reflector of Brooks et al. (1984), which was suggested to represent a major thrust of Variscan age.

The stations located on Jersey are the only ones from which the data sample the crust of the Armorican micro-continent. These stations show a crust that is thicker than the stations in the southwest of England. The two layer crust is similar to that observed at CWF, SSP and MCH located on the thick Avalonian crust. The Armorican crust, like the Avalonian crust of central England and Wales is relatively undeformed and may represent the relict structure from a Gondwanan craton.

6.7.4 Upper Mantle Structure

The 1D modelling of the receiver function phases in this study have allowed the velocity structure of the lithospheric mantle to be investigated. The BIRPS profiles reveal several structures in the lithospheric mantle beneath the margins of the British Isles (e.g. Hall et al., 1984; McGeary and Warner, 1985). The data from ORE located on the north coast of Scotland has provided a correlation between the receiver function data in this study and the known structures imaged by the BIRPS profiles offshore northern and western Scotland.

ORE lies close to the eastern end of the JUNE92 profile which constrains the velocity structure of the Flannan and W-reflectors imaged by the DRUM and GRID profiles (Morgan et al., 2000; Price and Morgan, 2000). Following the Moho P_s phase the receiver function data from ORE contains phases that when modelled using the *a priori* velocity information are found to originate from discontinuities at 47 km and 75 km depth (Figure 6.40a). The upper discontinuity is correlated with the W-reflector from the JUNE92 profile. The easterly dipping Flannan reflector is seen at a depth of up to 80 km on the DRUM profile (McGeary et al., 1987), and may correlate with the discontinuity at 75 km in the receiver function model. The amplitude of the receiver function phases has been fitted with a low velocity zone above the W and Flannan reflectors. This is consistent with the P -wave reflection amplitude modelling of Price and Morgan (2000) which suggest that there must be a transparent gradational low-velocity zone above the W-reflector. They suggest that the W-reflector is a subducted slab of mafic eclogite, and the low-velocity zone has been caused by metasomatism of the mantle material above the slab.

The data from RRR and KARB require upper mantle structures to fit the observed receiver functions. At these stations the depths to the top of the sub-crustal structures are 37 km and 33 km respectively. This is much shallower than the W-reflector seen at ORE. Offshore northwest Scotland the dipping Flannan reflector is found at between 26-40 km depth. However, the dip of the Flannan reflector is quite steep, and extrapolation of the reflector depth contours would suggest that the sub-crustal structure is significantly deeper than the structures in the receiver function models. If the structures seen at RRR and KARB are the Flannan reflector then the strike or dip of the reflector must change southward to fit the observations from the receiver function modelling.

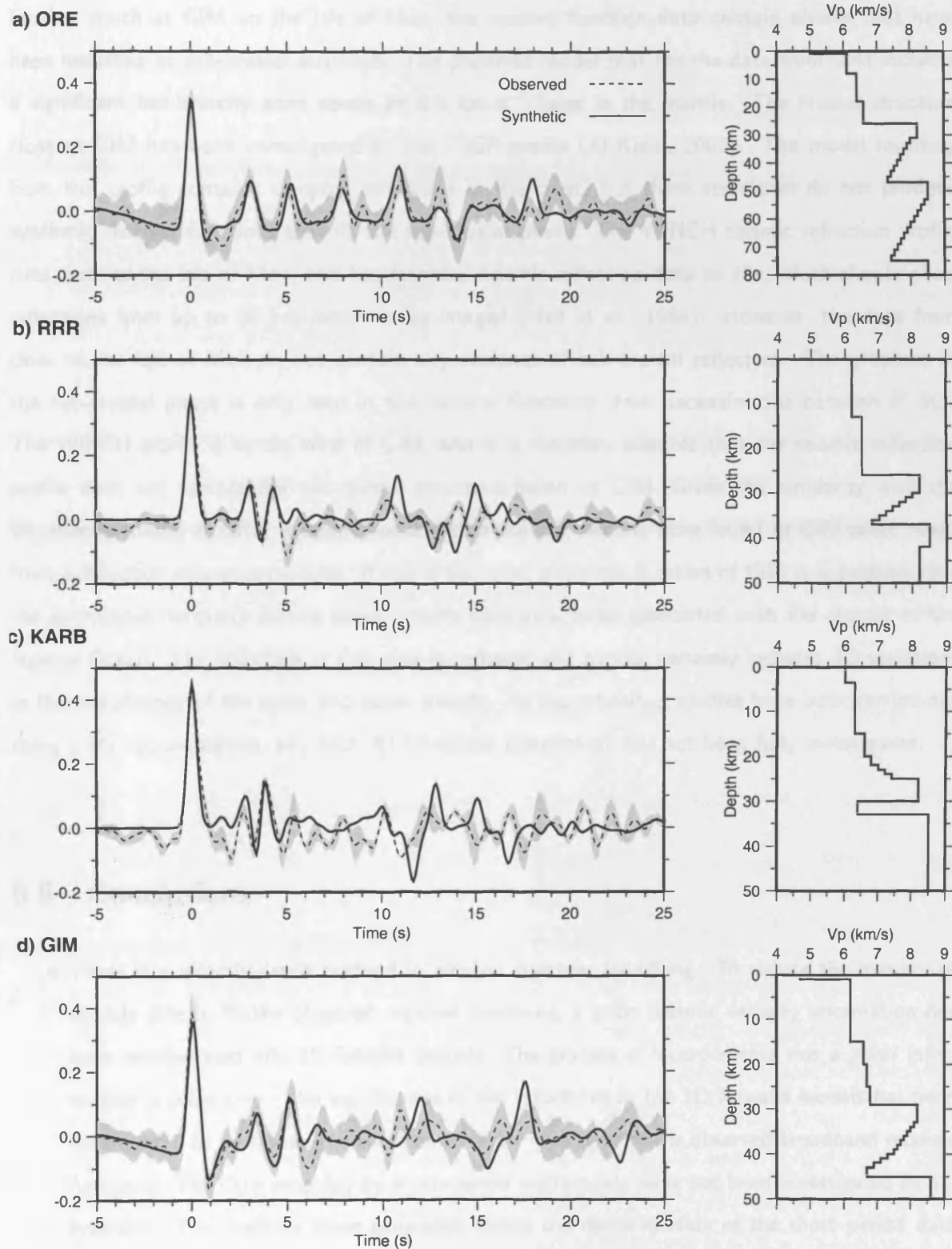


Figure 6.40: The 1D receiver function models from the British Isles including upper mantle structure a) The receiver function data and model from ORE identifying the W and Flannan reflectors, b) The receiver function data and model from RRR, c) The receiver function data and model from KARB, d) The receiver function data and model from GIM.

Further south at GIM on the Isle of Man, the receiver function data contain phases that have been modelled as sub-crustal structure. The preferred model that fits the data from GIM includes a significant low-velocity zone above an 8.5 km s^{-1} layer in the mantle. The crustal structure close to GIM has been investigated by the CSSP profile (Al-Kindi, 2002). The model resulting from this profile contains complex structures in the crust, but these structures do not produce synthetic receiver functions that fit the sub-crustal phase. The WINCH seismic reflection profile runs close to the Isle of Man, and has recorded seismic reflection data to 15s, which should allow reflections from up to 50 km depth to be imaged (Hall et al., 1984). However, the data from close to the Isle of Man do not contain any evidence of sub-crustal reflectors. The presence of the sub-crustal phase is only seen in the receiver functions from backazimuths between 0° - 90° . The WINCH profile is to the west of GIM, and it is therefore possible that the seismic reflection profile does not sample the sub-crustal structure found at GIM. Given the similarity with the W-reflector model at ORE, it seems possible that the low-velocity zone found at GIM could result from subduction related structures. If this is the case, given the location of GIM it is possible that the anomalous northerly mantle phase results from structures associated with the closure of the Iapetus Ocean. The structure in this area is complex and almost certainly includes 3D variations in the morphology of the crust and upper mantle. As the modelling studies have been carried out using a 1D approximation, any such 3D structural morphology has not been fully investigated.

6.8 Conclusions

- There is a velocity-depth tradeoff in receiver function modelling. To reduce the number of models able to fit the observed receiver functions, *a priori* seismic velocity information has been incorporated into 1D forward models. The process of incorporating this *a priori* information is subjective. The significance of the structures in the 1D forward models has been investigated by performing unconstrained linear inversions of the observed broadband receiver functions. The data recorded by short-period instruments have not been investigated by 1D inversion. The negative lobes generated during the deconvolution of the short-period data result in false structures in the resultant inversions models.
- The 1D forward modelling of the receiver function data from the British Isles has produced similar crustal thickness values to the $H-\kappa$ stacking technique (Chapter 5). The crustal

thicknesses from both of these methods are generally consistent with models from controlled source seismic methods. Exceptions to this include the modelled crustal thicknesses from the Iapetus Suture area.

- Many of the stations in this study have strong velocity discontinuities close to the surface. This is reflected in the quality of the observed receiver functions. High amplitude multiple phases from near surface layers interfere with the primary conversions from deeper structures making the receiver functions difficult to interpret. This is particularly the case with the data from the stations in the southeast of England, where the instruments are located on Mesozoic sedimentary rocks.
- With respect to the mismatch of up to 5 km between the models for seismic refraction and receiver function data in the Iapetus suture region, various explanations have been considered. The changes in average V_p and bulk crustal anisotropy required to fit the seismic refraction crustal thicknesses are large. It has therefore been concluded that the receiver function Moho in this area is in fact the top of the underplated layer found by Al-Kindi (2002), rather than the seismic refraction Moho.
- The stations in the Welsh Basin and on the Midland Micro-craton have the thickest crust in the British Isles at ~ 36 km. The models for SSP, MCH and CWF are all similar. A low amplitude Moho $PpPs$ phase is characteristic of the data from all of these stations.
- Structures in the lithospheric mantle have been identified. At ORE these correlate with the W and Flannan reflectors (47 km and 75 km respectively) from the JUNE92 profile (Price and Morgan, 2000). To fit the amplitudes of the phases it has been necessary to include a gradational low-velocity zone above an 8.5 km s^{-1} layer. This is consistent with the model of Price and Morgan (2000), suggesting the structure could represent a metasomatised layer above a subducted slab. Beneath RRR and KARB on the northwest coast of Scotland sub-crustal low-velocity zones have been found at 37 km and 33 km. These phases are shallower than the Flannan reflector imaged by the BIRPS profiles in this area. At GIM a significant low-velocity zone is required to fit the sub-crustal phase from 0° - 90° backazimuth. Given the similarity to the ORE model it seems possible that the structures at GIM are associated with subduction related to the closure of the Iapetus Ocean. However the features all vary with backazimuth, indicating that 3D structure exists. Therefore, these structures may not have been fully investigated with the 1D modelling study.

Chapter 7

Upper Mantle Structure

7.1 Introduction

The receiver function method not only enables the investigation of the structure of the lithosphere, but also provides information about the velocity structure of the upper mantle. The 410 km and 660 km discontinuities that bound the mantle transition zone (MTZ) have been investigated in numerous studies (e.g. Yuan et al., 1997; Dueker and Sheehan, 1997; Kosarev et al., 1999; Nyblade et al., 2000; Li et al., 2000b; Ramesh et al., 2002). In these studies the data from both broadband and short-period instruments have been used. To resolve discontinuities at the depth of the MTZ, it is necessary to apply a low pass filter to the seismograms. In previous studies the 3db corner frequency of the low pass filter is generally between 10-20s. In the present study the data from the short-period instruments have been recorded using 16bit digitisers, limiting the dynamic range of the seismograms. When such low-pass filters have been applied to the short-period seismograms, the resultant receiver functions are poor. Therefore only the data from the broadband instruments have been analysed with respect to the MTZ. The receiver functions have been calculated using seismograms that have been low-pass filtered below a period of 12s, before any rotation or deconvolution has been performed. The analysis of the upper mantle structure has used the techniques and code of Yuan et al. (1997) in which the receiver functions are calculated in the LQT ray coordinate system, rather than the ZRT component data used in the crustal structure study. In total 238 receiver functions have been calculated that are suitable for analysing the MTZ.

7.2 Moveout Corrected Receiver Functions

The delay time of the P_s conversions from the discontinuities that bound the MTZ can vary by several seconds with changing epicentral distance (Figure 3.7). As discussed in Section 3.4.2 it is possible to apply moveout correction to the receiver function data, which is similar to the NMO correction applied during seismic reflection processing. The receiver functions in this study have been corrected for the moveout of the P_s phase. The data have been corrected to an arbitrary epicentral distance (Δ) of 67° ($\rho=6.4\text{s}/^\circ$). Applying P_s moveout corrections to normalise the receiver functions to a constant Δ has resulted in the enhancement of P_s phases and reduction of the amplitude of the $PpPs$ phases in the stack of all receiver functions from each station.

The P_s phases from the 410 km and 660 km discontinuities can be seen in the stacks of the moveout corrected receiver functions from the broadband stations within the British Isles (Figure 7.1). The 410 km phase can be seen between 42.6-45.6s and the 660 km phase is seen between 65.9-69.8s. The time picks for these phase have been made on the stack of all of the P_s moveout corrected receiver functions from each station (Table 7.1). The sample interval of the receiver functions is 0.1s. The theoretical arrival times of the P_{s410} and P_{s660} through the IASPEI91 velocity model at an epicentral distance of 67° are 44s and 68s respectively.

7.3 Migration

The broadband receiver functions have been depth migrated onto a linear profile (Figure 7.2). The depth migration has been performed by back projecting the receiver function data along the incident P_s phase raypath. The reference velocities that have been used for the migration are taken from the IASPEI91 model. The depth migrated data have then been projected onto a line along the northern LISPb profile. The migrated section has been generated by dividing the linear projection of the depth migrated receiver functions into a grid of 1 km x 1 km distance-depth cubes. The mean amplitude of the receiver functions passing through each 1 km x 1 km cube is presented as the migrated receiver function section.

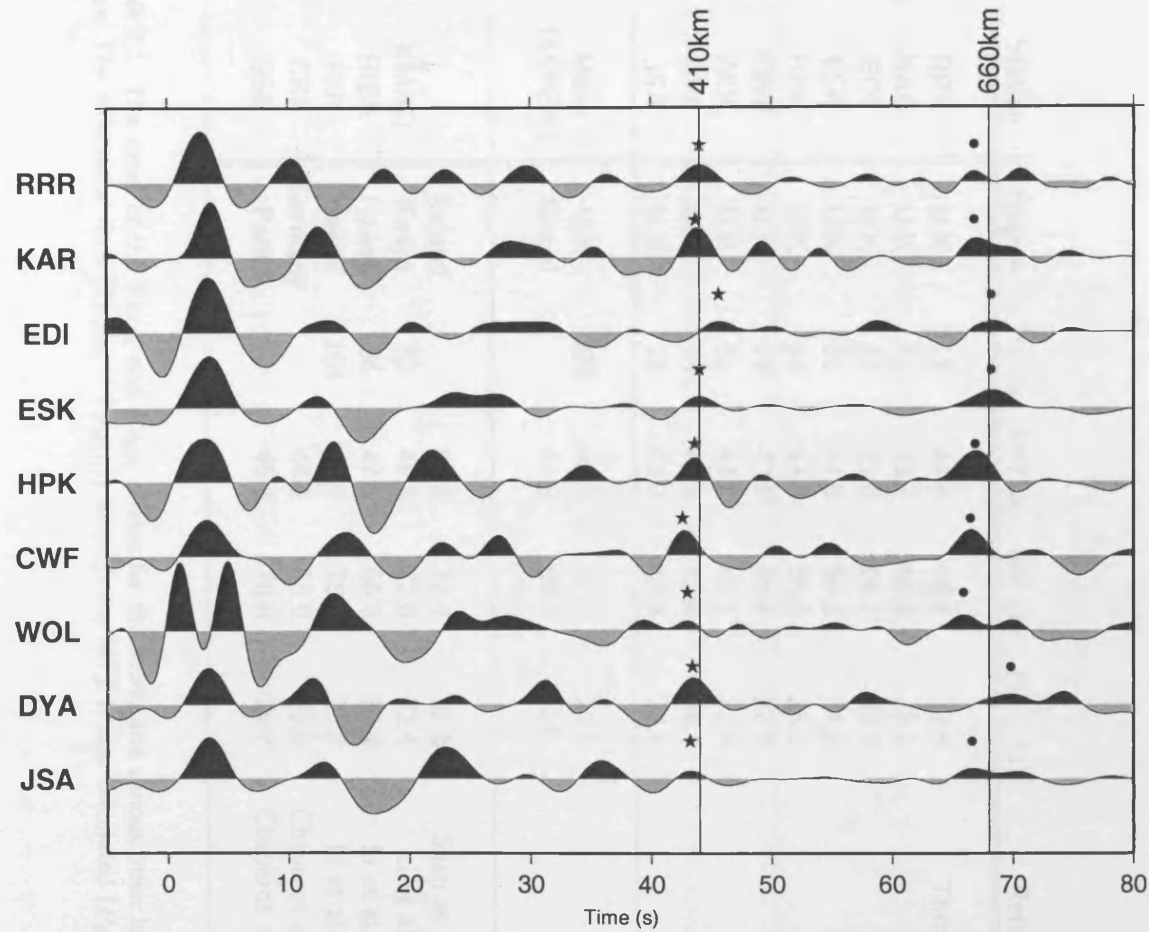


Figure 7.1: Stacked receiver functions for each of the broadband instruments. The receiver functions have been generated using seismograms that have been low-pass filtered below a corner frequency of 12s, and have been corrected for the moveout of the P_s phase to an epicentral distance of 67° . The time of the P_{s410} (stars) and P_{s660} (circles) phases have been marked. The theoretical arrival times of the P_{s410} and P_{s660} through the IASPEI91 at $\Delta=67^\circ$ have been marked at 44s and 68s.

Station	Region	Traces	410 (s)	660 (s)	P_{diff} (s)	Reference
RRR	U.K.	18	44.0	66.8	22.8	This study
KAR	U.K.	11	43.7	66.8	23.1	
EDI	U.K.	12	45.6	68.2	22.6	
ESK	U.K.	85	44.0	66.2	24.2	
HPK	U.K.	16	43.6	66.8	23.3	
CWF	U.K.	14	42.6	66.4	23.8	
WOL	U.K.	55	43.0	65.1	22.9	
DYA	U.K.	13	43.4	69.8	26.4	
JSA	U.K.	14	43.2	66.6	23.4	
Mean IASPEI91	U.K.	225	43.7	67.0	23.3	
	Global		44.0	68.0	24	
KMBO HIBS KIP GRF SSB	Iceland	52	50.8	72.4	21.6	Shen et al. (1998)
	Kenya		49.6	72.0	22.4	Li et al. (2003)
	Hawaii		47.2	66.6	19.4	Li et al. (2000a)
	Hawaii		47.2	70.4	23.2	Li et al. (2000a)
	Germany		44.2	68.0	23.8	Cheverot et al. (1999)
	Paris		45.3	70.0	24.7	Cheverot et al. (1999)

Table 7.1: The times of the P_{s410} and P_{s660} phases for this study and various other locations around the globe. The differential time ($T_{P_{s660}} - T_{P_{s410}}$) through the MTZ is also calculated (P_{diff})

In the migrated section there are several consistent phases in the observed data. The two phases in the top 100 km correspond to the Moho P_s and P_pP_s phases. Following this there are phases at 200-220 km, 380-420 km and 650-680 km. The quality of the migrated section is not particularly good. However, in previous studies migrated sections have been produced using a great deal more receiver functions than the 283 used in this study.

7.4 Discussion

The upper and lower boundaries of the MTZ at the global average depths of 410 km and 660 km are usually attributed to the mineralogical phase transformations of olivine to wadsleyite, and spinel to perovskite and magnesiowüstite (Lebedev et al., 2002). The equilibrium depth of these mineral phases is dependent upon the mantle temperature and pressure. The Clapeyron slope is positive at the 410 km discontinuity and negative at the 660 km discontinuity (Li et al., 2003). Therefore, if a high temperature anomaly exists in the MTZ the depth of the 410 km discontinuity will increase and the depth of the 660 km discontinuity will decrease. This will result in an overall thinning of the MTZ. By analysing the differential between the arrival times of the P_{s410} and P_{s660} phases it is possible to infer the thermal state of the MTZ (Cheverot et al., 1999; Shen et al., 1998; Lebedev et al., 2003; Li et al., 2003, e.g.). Cheverot et al. (1999) have analysed the P_{s410} and P_{s660} phases for numerous stations around the globe. They find that although there is considerable variation in the absolute arrival times of MTZ phases, almost all of the differential times ($P_{diff} = T_{P_{s660}} - T_{P_{s410}}$) are within the range $\pm 1s$ of the global average of 24s. This corresponds to a variation of ± 10 km of the standard MTZ thickness of 250 km.

The results from the British Isles show that the average P_{diff} time is 23.3s (Table 7.1). This is slightly lower than the 24.0s global average. ESK has the largest number of broadband receiver functions, and it would be expected that the results from this station are the most reliable. The P_{diff} time at ESK is 24.2s which is very close to the global average. Beneath Iceland the P_{diff} time is greatly reduced (Table 7.1) and this has been linked to the increase in mantle temperature associated with the plume beneath the Icelandic spreading ridge (Shen et al., 1998). The normal thickness of the MTZ beneath the British Isles shows that there is no significant thermal anomaly

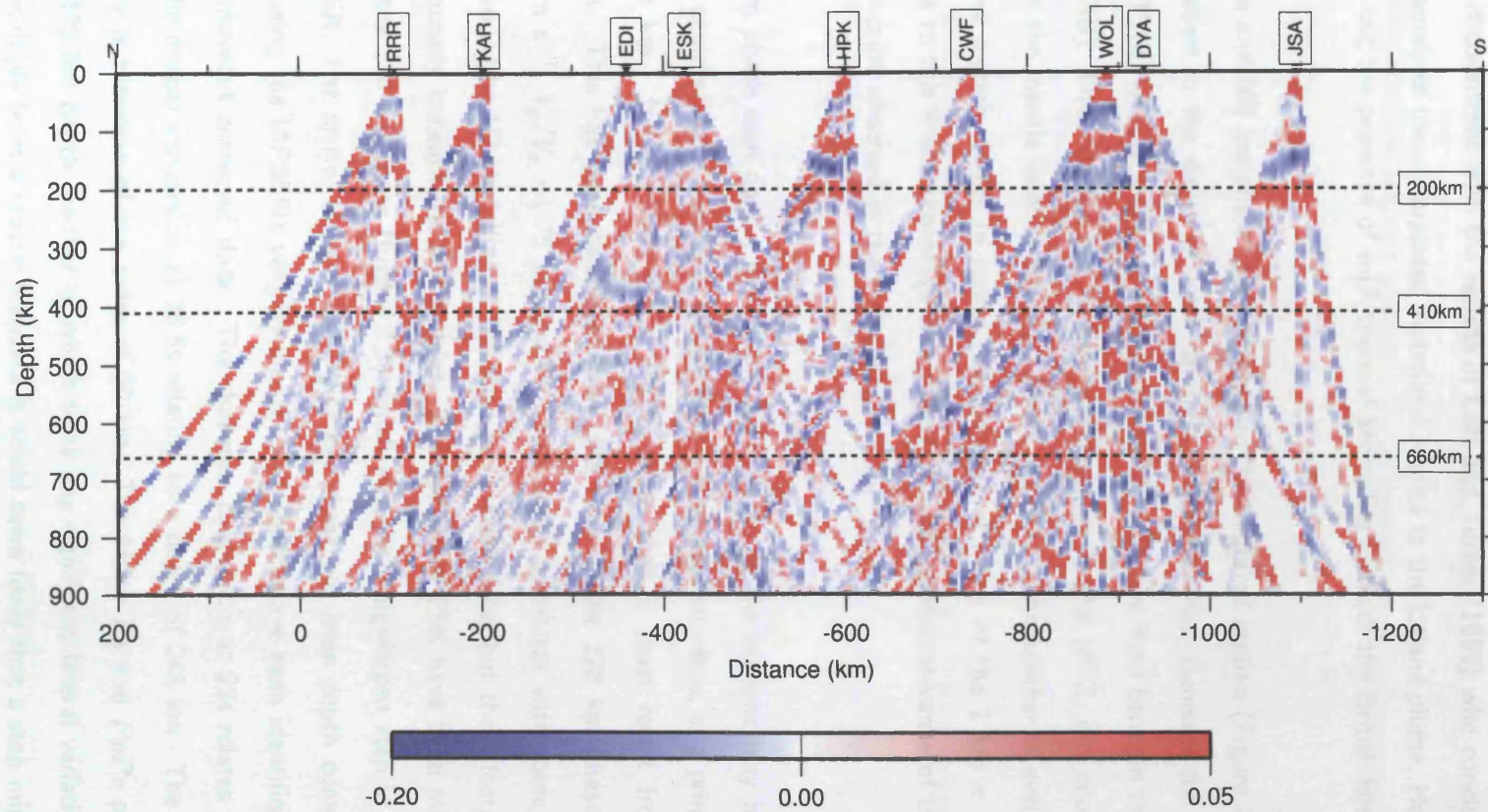


Figure 7.2: A profile of migrated receiver functions for the British Isles. The data have been projected onto the LISP profile. The distance axis is plotted relative to the northern end of the LISP profile (59.29°N , 5.02°E).

in this part of the mantle. It is therefore unlikely that the underplated material at the base of the crust is related to melting caused by a persistent thermal anomaly in the MTZ beneath the British Isles. This is consistent with the results of Clift and Turner (1998) who conclude that the causal thermal anomaly of the underplated material is related to the Iceland plume. However these results do not rule out the presence of such thermal anomalies beneath the British Isles in the past.

The 410 km and 660 km phases are identified in the migrated section (Figure 7.2). There is some lateral variation in the depth of both of these discontinuities. Beneath the stations in central England (HPKB, CWFB, WOL) the P_{s660} is slightly shallower than beneath the northern stations (ESKB, EDIB). This may be due to lateral differences in the MTZ, but could also result from variations in the mantle velocity structure above the MTZ. The number of events used to produce the migrated section results in sparse ray coverage of many of the 1 km x 1 km cubes in the profile. Due to this limited coverage it is difficult to discuss the relevance of the subtle variations in MTZ structure observed in the migrated profile.

The 220 km phase seen on the data from the majority of the stations may be more significant. This phase could either be multiple energy from a shallower interface, or a primary P_s conversion from ~ 220 km. If the phase was a $PpPs$ phase then it must result from a discontinuity at ~ 60 km. This has been calculated from the time of the 220 km phase using an average $V_p = 7.1 \text{ km s}^{-1}$, $V_p/V_s = 1.75$ at $\Delta = 67^\circ$. This depth correlates with some of the sub-crustal features seen in the 1D modelling of the receiver function data, but these features have not been seen continuously beneath the British Isles. The data from ESK have been migrated individually to look at the 220 km phase (Figure 7.3) which includes a significant step in delay times to the north of ESK. The arrival time of the ~ 220 km phase has been depth converted for the data from ESK using the IAPSEI91 velocity model. The phases have been identified at 23s and 26.5s in the P_s moveout corrected data. The shallower conversion at 23s relates to a depth of 213 km, with the deeper conversion at 26.5s relating to a depth of 243 km. The total offset at the discontinuity is therefore of the order of 30 km. The Moho P_s and $PpPs$ phases can be seen between 0-100 km depth, neither of which show any significant lateral variation. If the 220 km phase is a multiple from a crustal interface it would seem likely that a step might also be seen in the migrated crustal phases. Therefore it seems likely that the 220 km phase is a primary conversion.

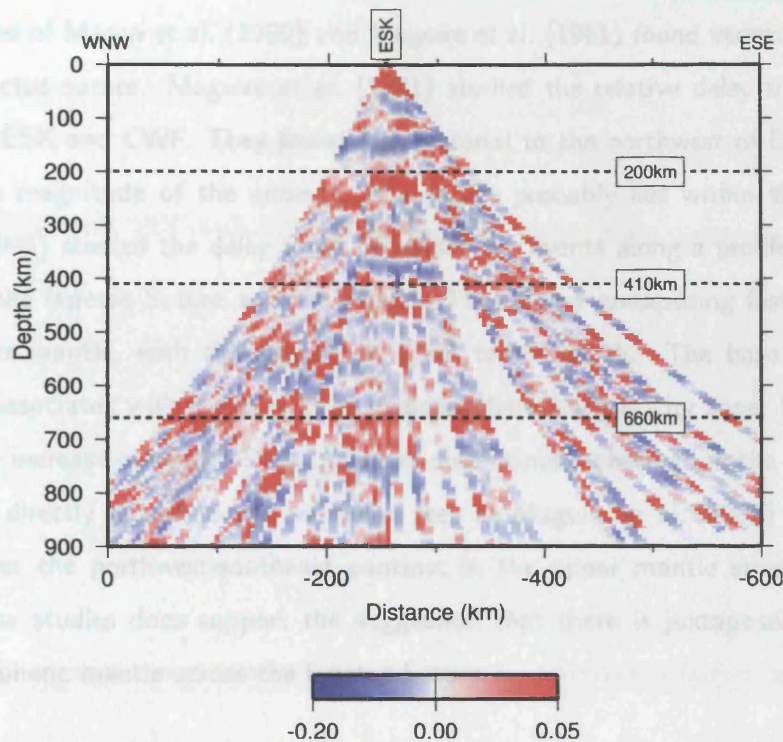


Figure 7.3: Migrated receiver functions from ESK showing a step in the phase at between 200-250 km depth.

The most documented mantle discontinuity at ~ 220 km is the Lehmann discontinuity (Deuss and Woodhouse, 2002). The origin and continuity of the Lehmann discontinuity are open to some discussion. It has been identified in the PREM velocity model, but not the IASPEI91 velocity model. Gu et al. (2001) find the Lehmann discontinuity in only 25% of oceanic regions and 50% of continental regions. They note that the Lehmann discontinuity is intermittent in nature, and has strong depth variation. It has frequently been associated with a rheological boundary separating a rigid continental plate from plastic, convecting mantle below (Gu et al., 2001). Gaherty and Jordan (1995) conclude that the Lehmann discontinuity may be the base of an anisotropic layer. Beneath continents upper mantle anisotropy appears to be inherited from major episodes of orogenic deformation. In the global study of SS-precursors Deuss and Woodhouse (2002) find discontinuities at 220 ± 20 km and 260 ± 10 km beneath the British Isles.

The step in the ESK migrated data is maximised when projected onto a WNW-ESE profile. This profile is almost perpendicular to the strike of the structures created during the Caledonian orogeny. The location and trend of the step suggest that it may be related to the Caledonian orogeny, resulting from the juxtaposition of Laurentian and Avalonian Mantle. The teleseismic

delay-time studies of Mason et al. (1999) and Maguire et al. (1981) found variation in their results close to the lapetus suture. Maguire et al. (1981) studied the relative delay time of teleseismic events between ESK and CWF. They found fast material to the northwest of ESK and conclude that due to the magnitude of the anomaly, the source probably lies within the upper mantle. Mason et al. (1999) studied the delay times of teleseismic events along a profile of seismometers which spanned the lapetus Suture zone in Ireland. They found juxtaposing fast and slow blocks within the upper mantle, with the fast block again to the north. The base of the Lehmann discontinuity is associated with the base of an asthenospheric low-velocity zone. It therefore seems unlikely that the increase in depth of the Lehmann discontinuity observed in the migrated receiver function data is directly related to the anomalies seen by Maguire et al. (1981) and Mason et al. (1999). However the northwest-southeast contrast in the upper mantle structure observed in all three of these studies does support the suggestion that there is juxtaposed Laurentian and Avalonian lithospheric mantle across the lapetus Suture.

7.5 Conclusions

- The broadband teleseismic data from the British Isles have been processed to analyse mantle structure by applying a 12s corner frequency low-pass filter to the seismograms. In the resultant moveout corrected and migrated receiver functions phases have been identified at ~ 220 km, ~ 410 km and ~ 660 km.
- The average differential travel-time P_{diff} between the top and bottom of the MTZ for the British Isles is 23.3s. At ESK where the greatest number of broadband receiver functions have been calculated P_{diff} is 24.2s. These values are close to the global average of 24.0s and suggest that there is no significant thermal anomaly in the MTZ below the British Isles.
- The most notable phase in the migrated receiver functions section is that at ~ 220 km. If this is a primary P_s conversion from the mantle then it correlates well with the reported depths of the Lehmann discontinuity. There is a 30 km step in this phase beneath Eskdalemuir. If the Lehmann discontinuity is the boundary between rigid continental mantle, and the convecting mantle beneath, then the step observed beneath Eskdalemuir may represent the juxtaposition of Laurentian and Avalonian mantle.

Chapter 8

Summary of Findings

8.1 Introduction

This final chapter summarises the results of the receiver function project discussing how the results of Chapters 5-7 help answer the questions raised at the end of Chapter 2. Specifically the aims of the study set out in Chapter 2; by using the available teleseismic receiver function data were to 1) extend the knowledge of the structures beneath the available stations, therefore increasing the constraint of onshore crustal properties, 2) closely examine the structure of the Iapetus suture region, concentrating in particular on the contrast between the Laurentian and Avalonian crusts and the magmatic underplated material found beneath the Irish Sea, 3) constrain the onshore extent of the sub-crustal Flannan and W-reflectors found in northwest Scotland, and 4) investigate the structure of the upper mantle, looking for links between the crustal and upper mantle structures. Finally future work that could follow on from the research in this project is discussed.

8.2 Summary

In total teleseismic data from 34 broadband and short-period instruments have been investigated using receiver function analysis. $H-\kappa$ stacking has been used to produce a series of point values for crustal thickness and average crustal V_p/V_s ratio. 1-D modelling of the receiver function phases has provided more detailed information about the velocity-depth structure beneath each of the

seismic monitoring stations. Where broadband data have been available the structure of the mantle transition zone has been investigated using migration and stacking techniques.

Before the results of the analyses are reviewed, it is important to consider the difficulties found when interpreting teleseismic receiver function data. The most notable problem when investigating the observed receiver function phases using $H-\kappa$ stacking and the 1-D forward and inverse modelling techniques is a velocity-depth tradeoff. This velocity-depth tradeoff equates to approximately a 1 km change in depth with a 0.2 km s^{-1} change in V_p for the crustal thicknesses observed within the British Isles. Using *a priori* velocity information in the analyses greatly reduces the number of models which may fit the observed receiver functions. In the case of the $H-\kappa$ stacking, the dependency on the stacking V_p has been investigated by performing two sets of analyses. In the first case a constant $V_p = 6.3 \text{ km s}^{-1}$ is used for the data from all stations; in the second the V_p for each location is defined from pre-existing seismic refraction velocity models. The 1-D forward modelling of the crustal velocity structure has used the pre-existing seismic refraction models as a starting point. The process of iterative forward modelling using the *a priori* velocity information is subjective. The results of unconstrained 1-D linear inversions performed on the broadband receiver functions have been used to provide an estimate of the significance of the structures within the 1-D forward models.

The results of the modelling from the receiver functions have increased the knowledge of the onshore crustal structure of the British Isles. This has shown that the crustal thicknesses vary between 24-36 km. The features of the crustal morphology that have been seen are; 1) thinning of the crust to ~ 25 km in northwest Scotland, 2) a 36 km thick crust in the Midland Valley of Scotland, 3) a ~ 36 km thick crust through central England and Wales and 4) thinner crust (~ 28 km) in the southwest of England. In general the models from $H-\kappa$ stacking and 1-D forward modelling correlate with the observations from the previous deep seismic reflection and refraction profiles. The exception are the models from the lapetus suture region, where the base of the crust is up to 5 km shallower than the seismic refraction Moho. This mismatch is thought to occur because the strongest Ps phases in the receiver functions have originated from the top of a layer of underplated material at the base of the crust observed by Al-Kindi (2002), rather than from the Moho. The crustal thickness of the Welsh Basin and Midland Micro-craton is significantly thicker than the majority of the British Isles. The deep crustal structure of this area has only been sparsely sampled by previous studies, but the new models concur with the *a priori* information.

The constraint on the structure of southeast England has not been improved because the crustal phases of the observed receiver functions are masked by multiple phases from near surface velocity discontinuities between the Mesozoic sediments and the Variscan basement.

The detailed crustal structure of the Iapetus suture zone was of particular interest because deep seismic reflection profiling has found strongly dipping reflectors in the crust, which are suggested to represent the boundary between the Laurentian and Avalonian crust (e.g. Freeman et al., 1988; Soper et al., 1992). The seismic refraction models in the area do not find significant velocity variations across the Iapetus suture. The receiver function study has provided an opportunity to investigate the velocity structure and in particular the S -wave velocities of the region. The results from H - κ stacking and 1D forward modelling do not show any consistent variations in crustal thickness, intra-crustal structure or V_p/V_s ratio across the suture zone. The study of the intra-crustal velocity structure has been hampered at some stations by the presence of multiple energy from the Carboniferous and Devonian sedimentary basins on which they are situated.

Structures have been modelled in the uppermost mantle (<100 km depth) from the data recorded at a number of stations. At ORE on the northern coast of Scotland, the Flannan and W-reflectors have been identified in the observed receiver functions. This shows that it is possible to identify these known structures within the receiver function data, therefore increasing the plausibility of the other structures modelled within the upper mantle. The model for the W-reflector at ORE supports the seismic refraction model of Price and Morgan (2000), showing a gradational low-velocity zone above an 8.5 km s^{-1} layer. The model is therefore compatible with their theory that the 8.5 km s^{-1} layer represents a subducted slab, and the low-velocity zone above it was caused by metasomatism. The data from the stations RRR, KARB (NW Scotland) and GIM (Isle of Man) have been fitted with similar models to that for ORE, indicating the presence of further sub-crustal low-velocity zones. The depth and distribution of the sub-crustal structures at RRR and KARB link them to the Flannan reflector, seen dipping beneath the northwest of Scotland. The southward continuity of the structures found at ORE, RRR and KARB is unknown. The data from the stations in the Midland Valley contain no sub-crustal structures suggesting that the Flannan and W-reflectors seen in northwest Scotland must be truncated or pinch out toward the south. The distribution of the sub-crustal structure away from GIM is unknown. The models from the other stations in the Iapetus suture zone, as well as the WINCH seismic reflection profile do

not show structures similar to that found at GIM, which suggest that this is an isolated feature. If the sub-crustal low-velocity zone at GIM is linked to subduction related metasomatism, then the structure may represent a relic subduction zone formed during the closure of the Iapetus Ocean.

The final objective of the study was to investigate the upper mantle, with the aim of identifying any significant structures which may or may not correlate with the crustal structure. This has been possible using the limited amount of broadband data. The results of this investigation have identified the 410 km and 660 km discontinuities in P_s moveout corrected receiver functions. The differential arrival times between these two phases is close to the global average of 24s. This indicates that there is no significant thermal anomaly in the mantle transition zone beneath the British Isles. Secondly a phase originating from ~ 220 km has been identified. If this phase is not a multiple from shallower structures then it may represent the Lehmann discontinuity. In the data from Eskdalemuir there is a significant step in the ~ 220 km phase. This step is maximised when the data are projected onto a WNW-ESE profile. This is almost perpendicular to the strike of the Caledonian orogeny. It has been interpreted that this step in the Lehmann discontinuity may highlight the juxtaposition of Laurentian and Avalonian mantle during the Caledonian orogeny.

8.3 Future Work

If this study had been carried out using data from an array of seismic recording stations specifically designed to answer the questions laid out in the objectives, rather than using the existing dataset from the BGS, then some of the methods used may have been different. There are two specific experiments which would help to answer some of the questions raised in this study and other works.

1. **2D Profiles** Recording teleseismic events with a closely spaced profile of broadband instruments allows the lateral continuity of the crustal and upper mantle structures to be investigated. This can be achieved by applying more sophisticated moveout and migration techniques than have been used in the present study. With a closely spaced receiver function dataset it is possible to apply Kirchhoff migration techniques, enabling a detailed investigation of the mantle transition zone. The two most useful profiles which could be recorded in the British Isles would be a north-south profile cutting all of the major

geological structures, and an east-west profile running from Wales to Eastern England. The north-south profile would allow the southward continuity of the Flannan and W-reflectors to be investigated, and, like the LISPB profile, would provide information about the lateral contrasts between the juxtaposed geological terranes. The east-west profile would help constrain the eastward extent of the thicker crust in central England and Wales, and also extend the knowledge of the crust of eastern England where there is currently little information. At present (August 2003) there is a 27 station profile of broadband instruments installed in northern Scotland running approximately along the line of the LISPB profile, which is maintained by the University of Bristol (G Helffrich, pers. comm. 2003). The data from this experiment may help to constrain the southward extent of the Flannan and W-reflectors.

2. **3D Array** In the present study the effects of dipping and anisotropic structures have been tested against the observed receiver functions. However, with the limited distribution of stations available it has not been possible to investigate 3D variability of the crust and upper mantle. In areas where complex 3D structure is known to exist, as for example the Iapetus Suture zone, the installation of a grid of broadband stations rather than a 2D profile would help to provide a better constraint on the 3D variation in structure. If such a dataset were collected then more complex 3D modelling codes would have to be applied to the receiver function method. However, the number of variables used in the 3D modelling of receiver function data may make the procedure prohibitively non-unique. It may therefore be the case that it would only be possible to test specific models against the observed receiver function data, rather than constructing a unique model based around the observations. Recording teleseismic data from a grid of broadband stations over the Iapetus Suture zone would also allow the distribution of the underplated magmatic material found in this area to be investigated more thoroughly. By calculating the extent and volume of the underplated material it would be possible to estimate the subsequent uplift that was caused by its emplacement. It would also be possible to calculate the size of the thermal anomaly required to generate the volume of underplated material, which may then provide evidence as to the origin of the underplated material.

Appendix A

Receiver Functions

A.1 Receiver Function Plots

The raw receiver functions from all of the 34 instruments from within the British Isles are plotted. The receiver functions have been generated using teleseismic events cut 30s before and 90s after the P -wave arrival. The deconvolution has been performed using a Gaussian filter parameter of $a = 3$ and a water level parameter of $a = 0.001$. The receiver function data are plotted relative to the event backazimuth. For many of the stations, plots of both the radial and tangential receiver functions are presented. The receiver functions have been stacked in 20° backazimuth bins, and the bin window has been rotated through 360° at 5° intervals. The binning has been performed to enhance the receiver functions phases by minimising noise in the signal through destructive interference when that data are stacked. As can be seen in the raw receiver functions the majority of these bins contain several receiver functions. Applying this technique the receiver functions may be smoothed over a maximum of $\pm 10^\circ$ from their true backazimuthal position. The $H-\kappa$ stack of the receiver function data for each instrument is also presented, and the times of the Moho P_s and $P_p P_s$ phases calculated from the stack are indicated on the plots of both the radial and tangential receiver function data.

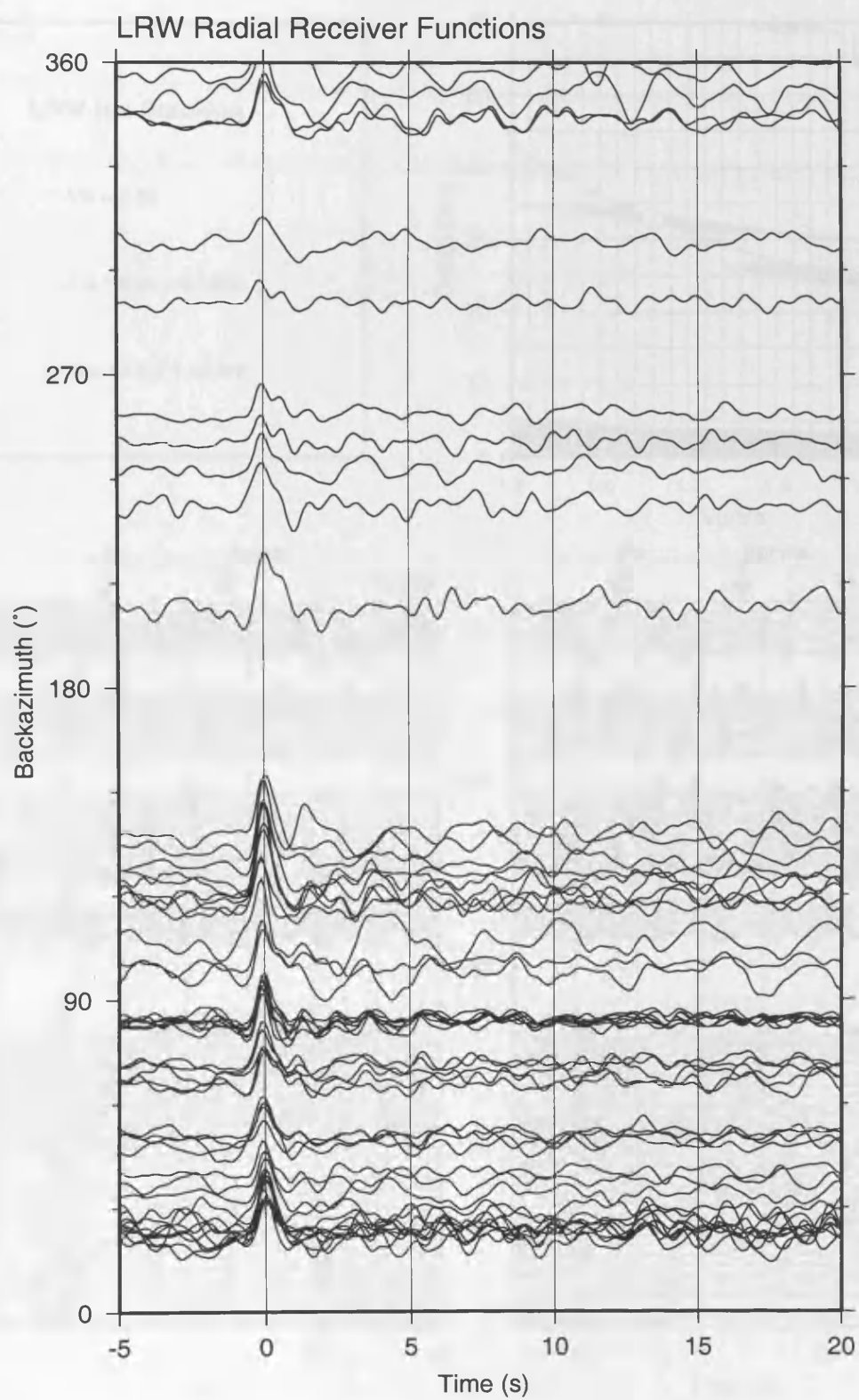


Figure A.1: Raw receiver function data from LRW plotted relative to backazimuth.

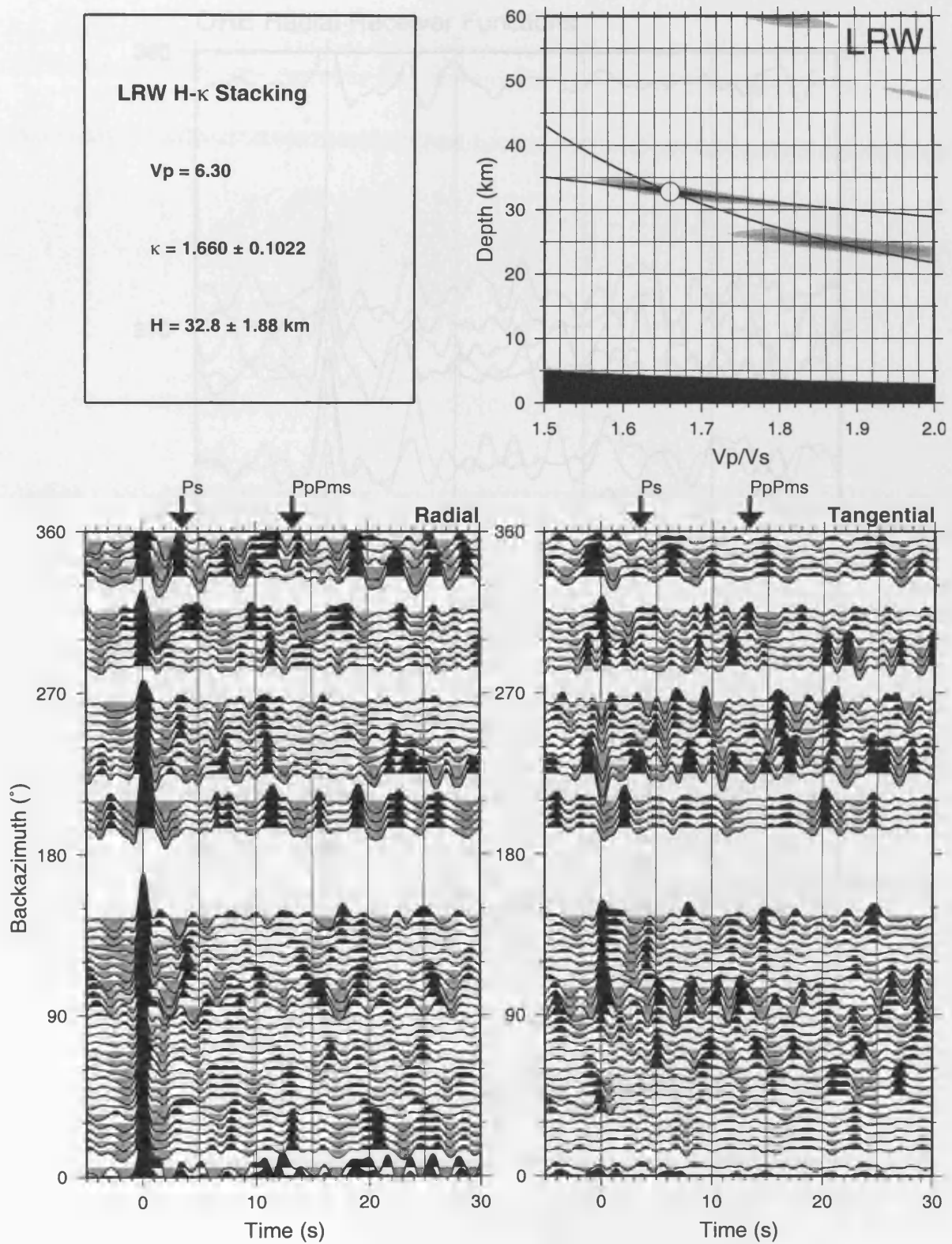


Figure A.2: The radial and tangential receiver functions for LRW, with the results of the H - κ stacking.

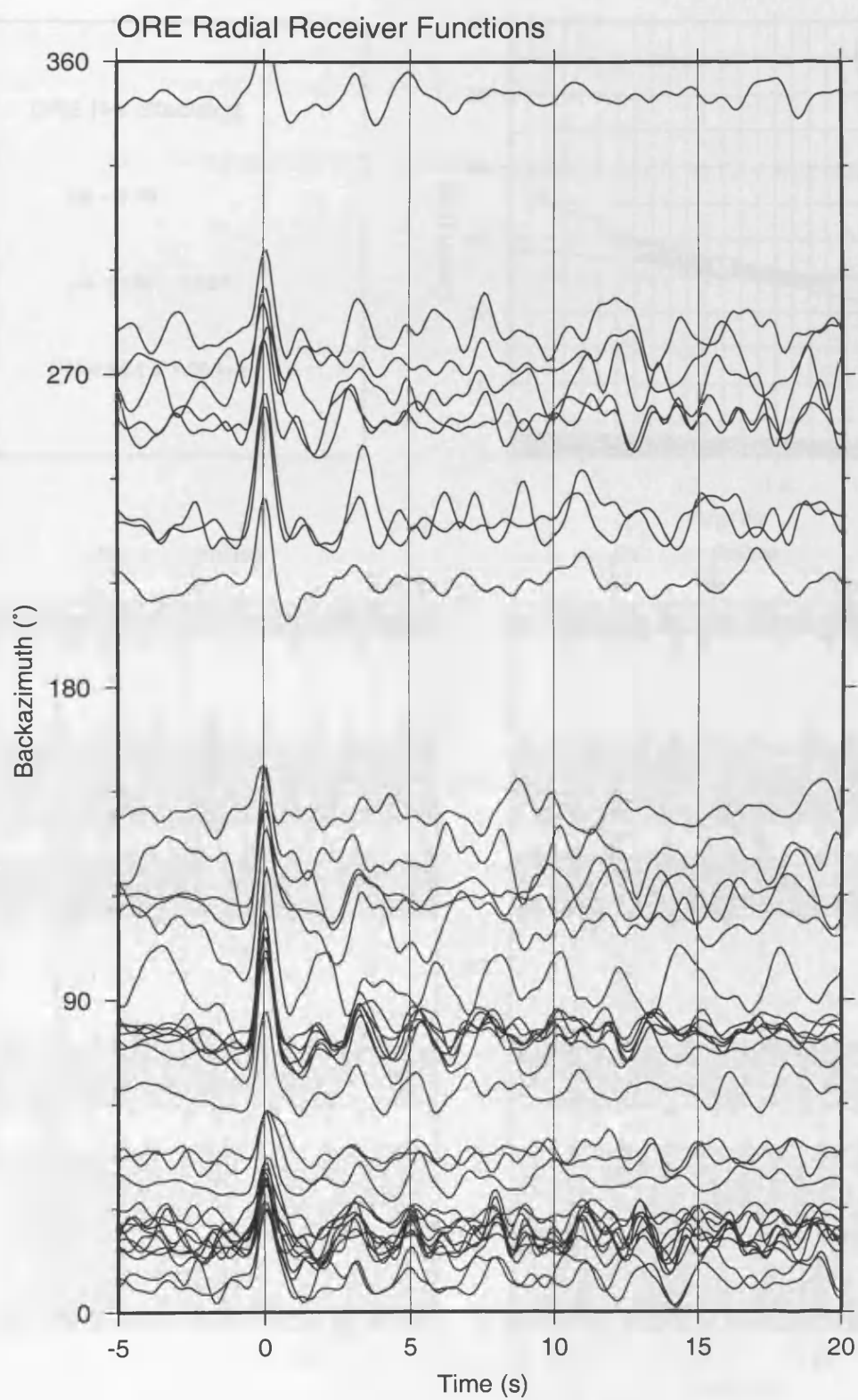


Figure A.3: Raw receiver function data from ORE plotted relative to backazimuth.

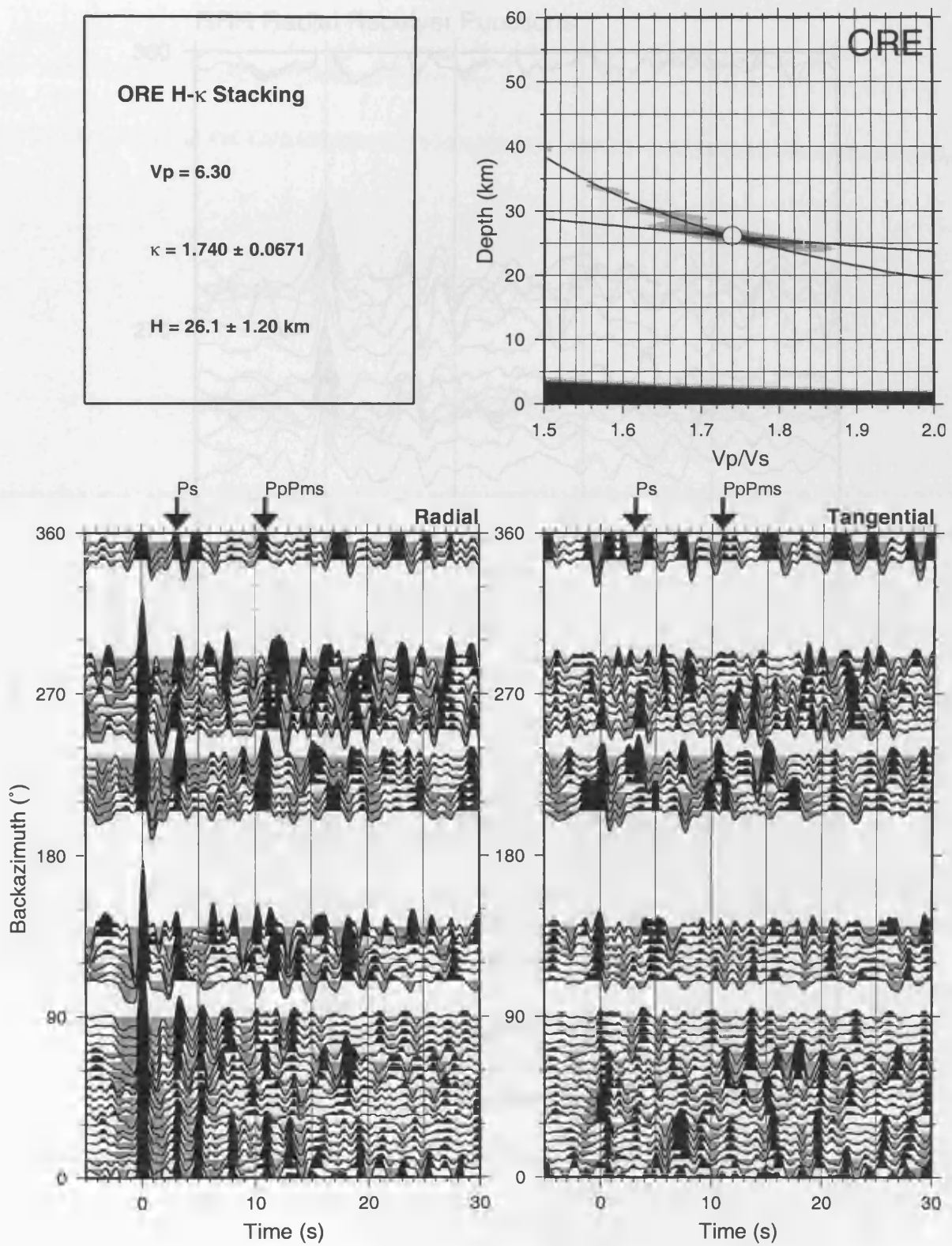


Figure A.4: The radial and tangential receiver functions for ORE, with the results of the H - κ stacking.

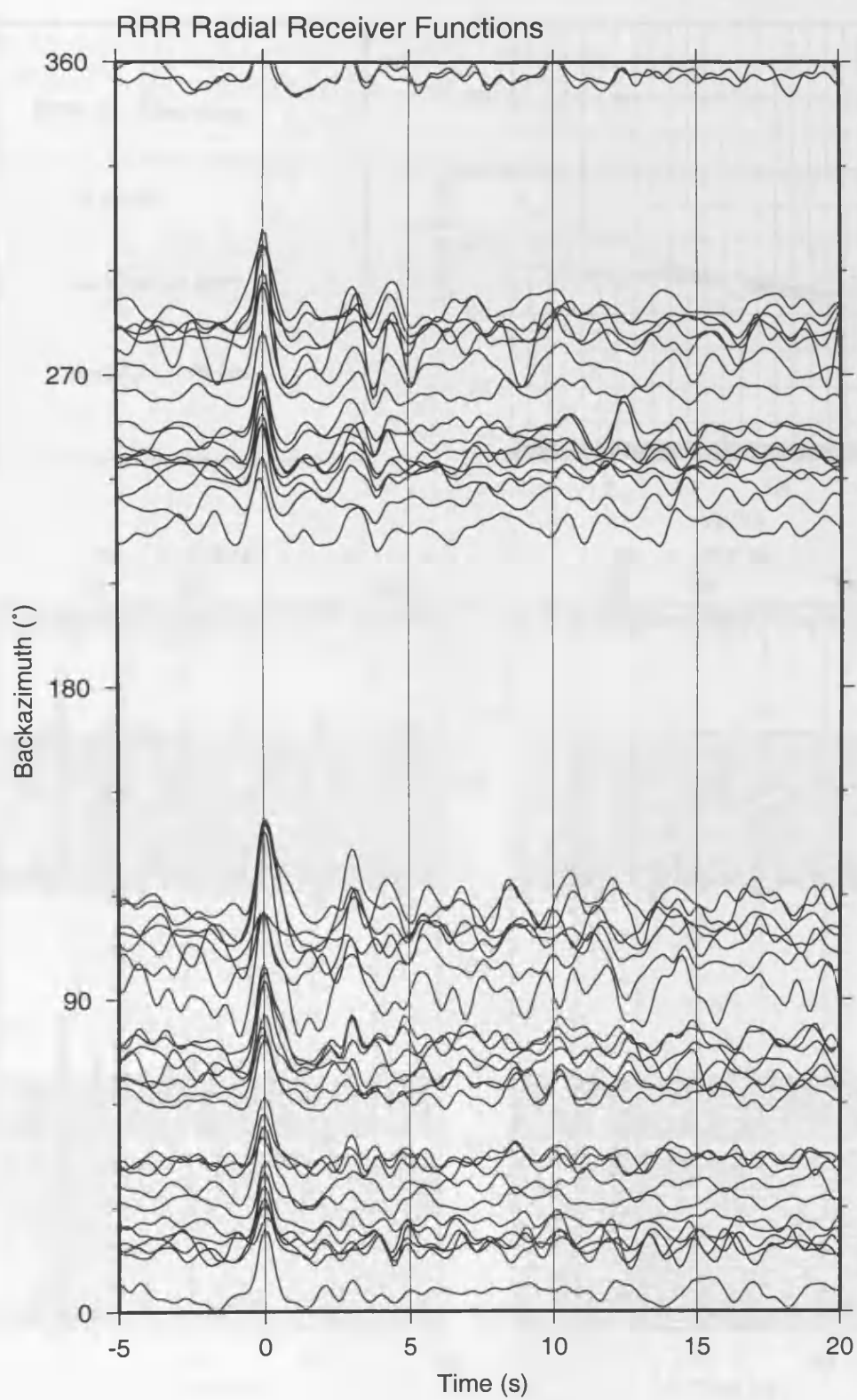


Figure A.5: Raw receiver function data from RRR plotted relative to backazimuth.

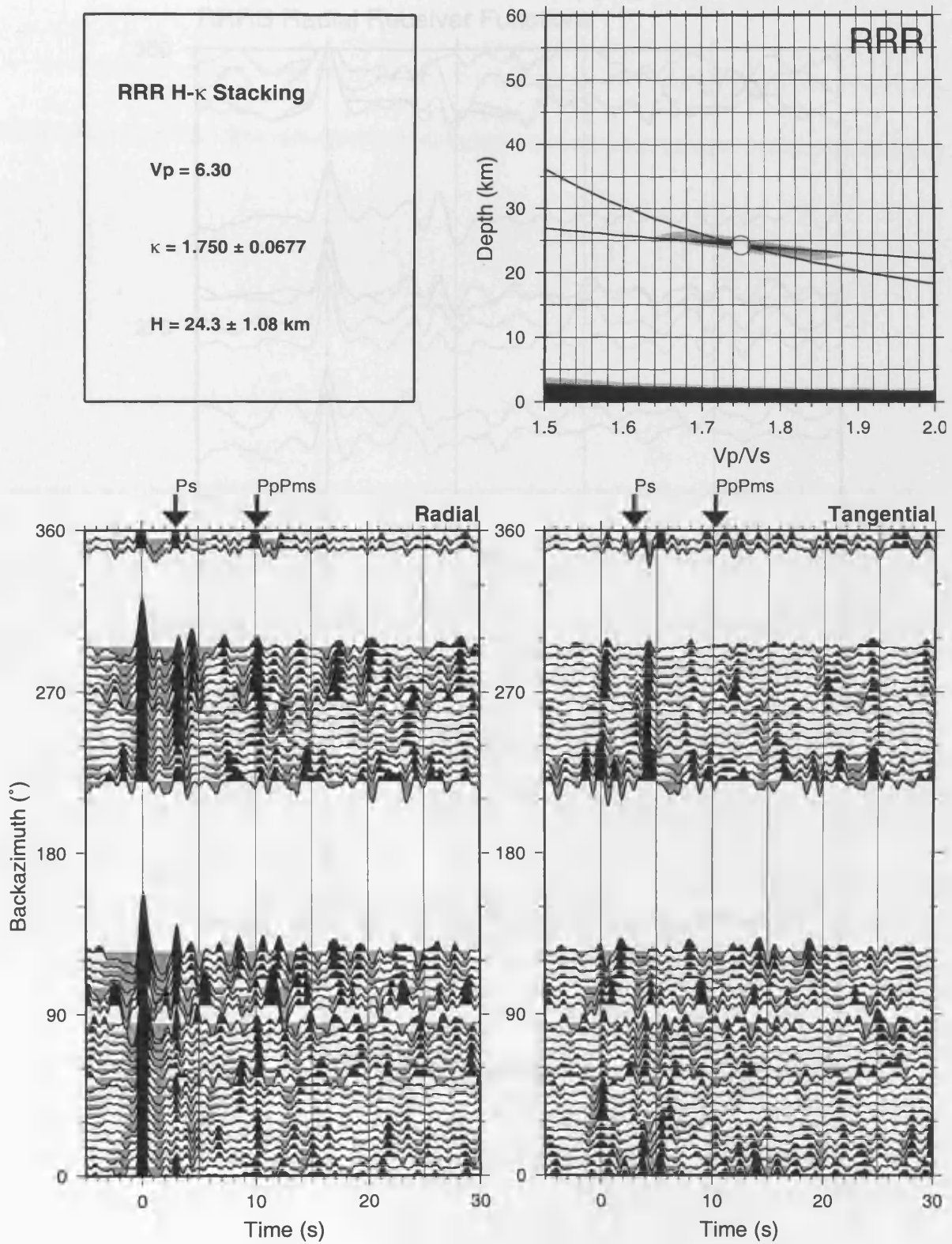


Figure A.6: The radial and tangential receiver functions for RRR, with the results of the H - κ stacking.

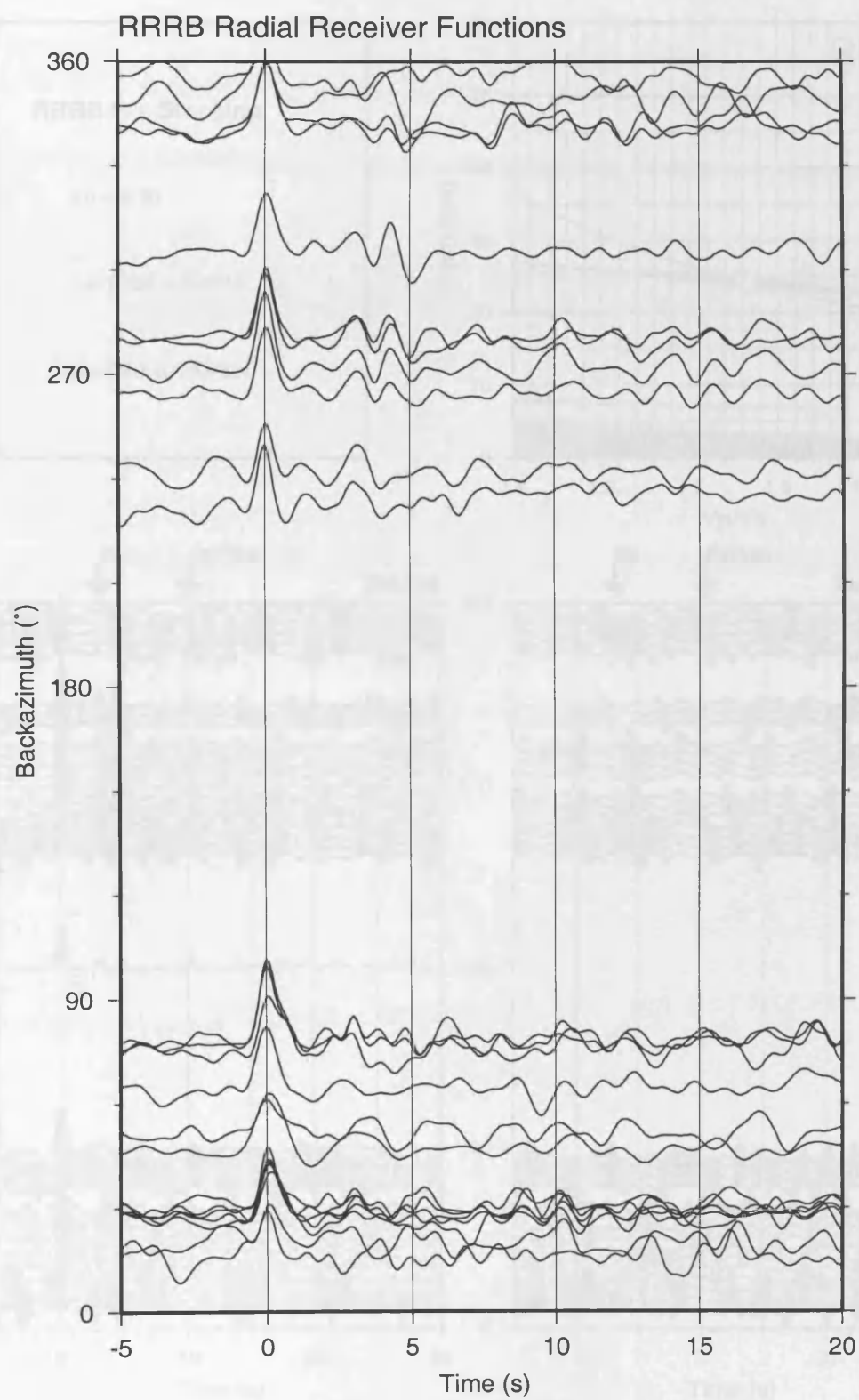


Figure A.7: Raw receiver function data from RRRB plotted relative to backazimuth.

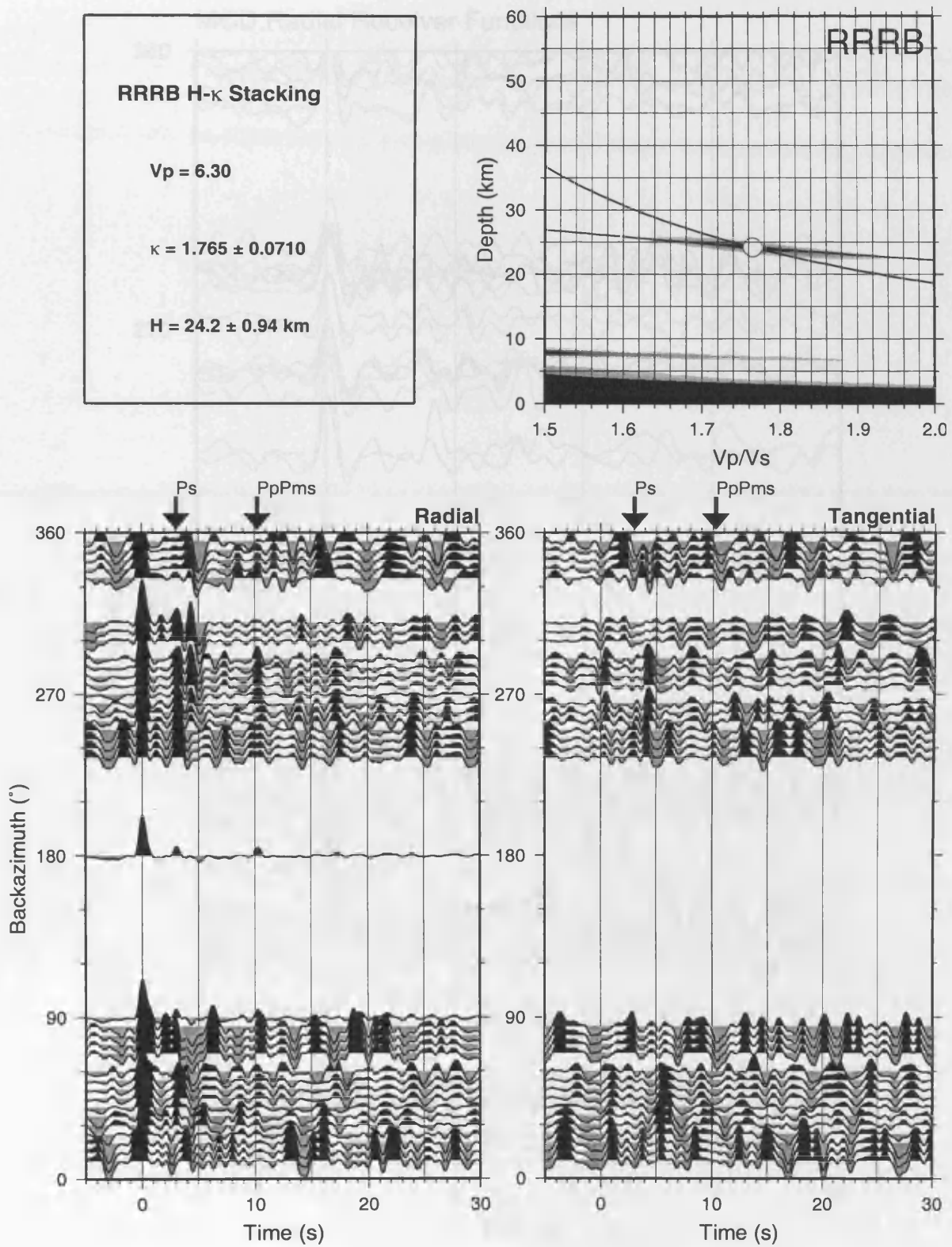


Figure A.8: The radial and tangential receiver functions for RRRB, with the results of the H - κ stacking.

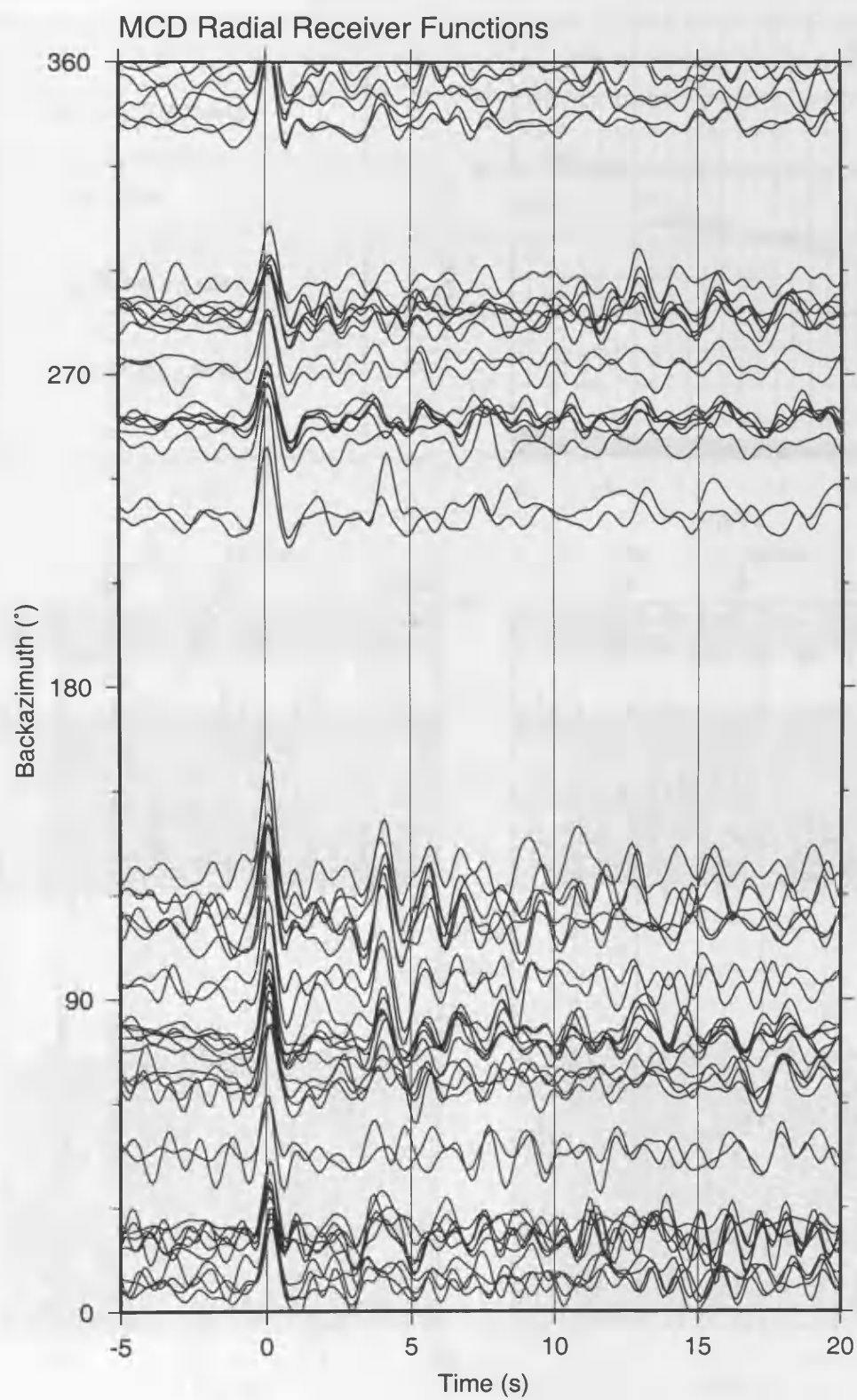


Figure A.9: Raw receiver function data from MCD plotted relative to backazimuth.

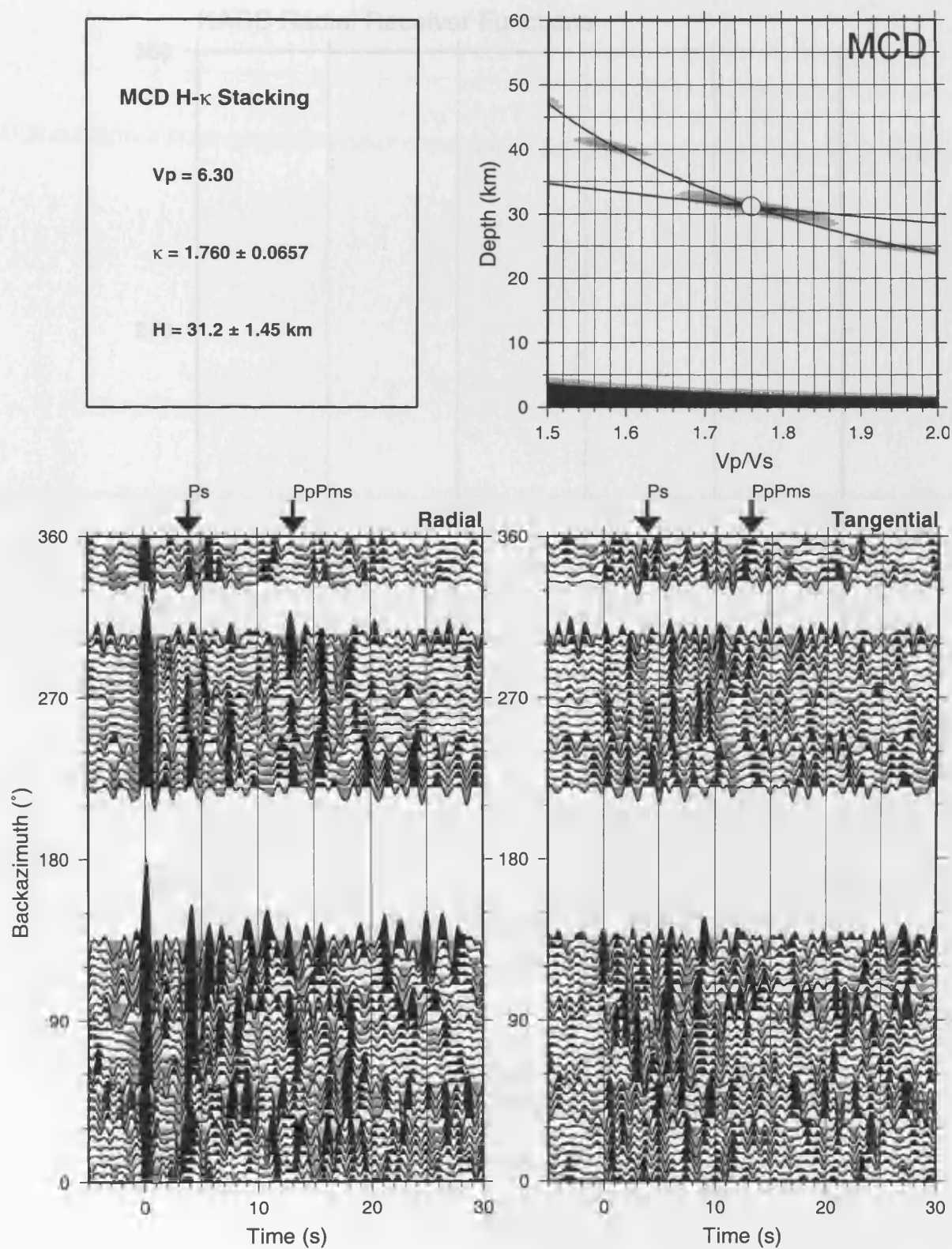


Figure A.10: The radial and tangential receiver functions for MCD, with the results of the H - κ stacking.

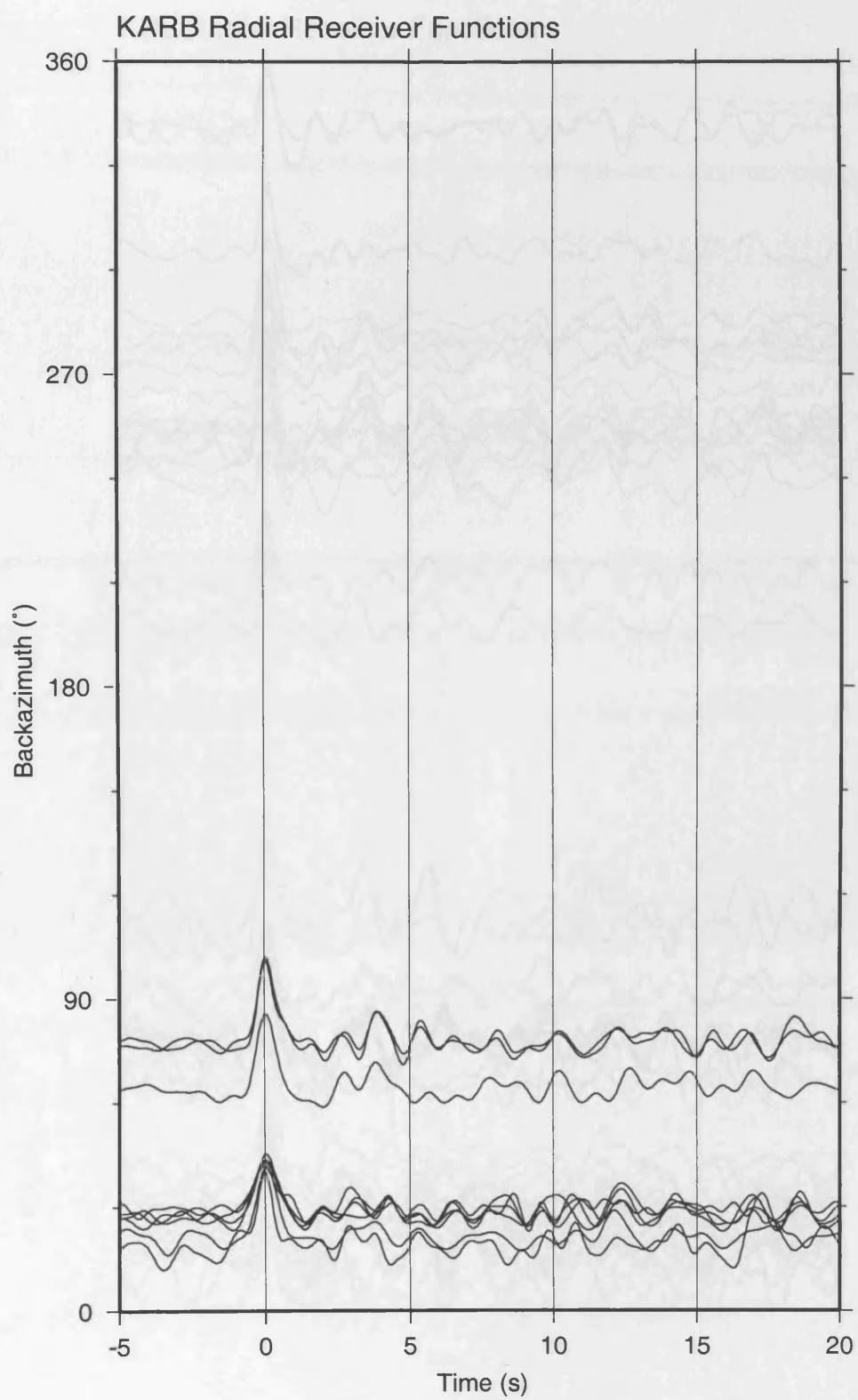


Figure A.11: Raw receiver function data from KARB plotted relative to backazimuth.

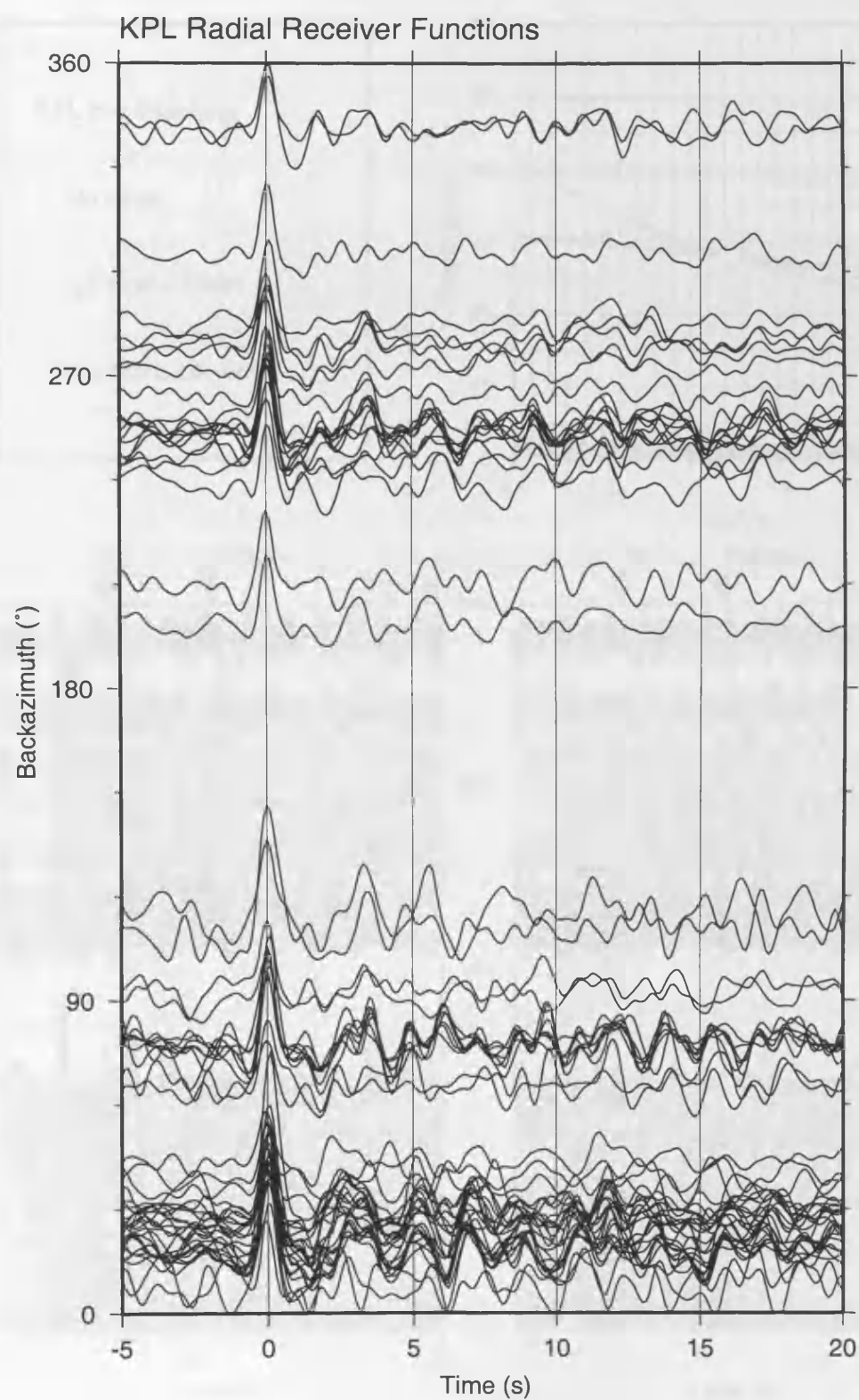


Figure A.12: Raw receiver function data from KPL plotted relative to backazimuth.

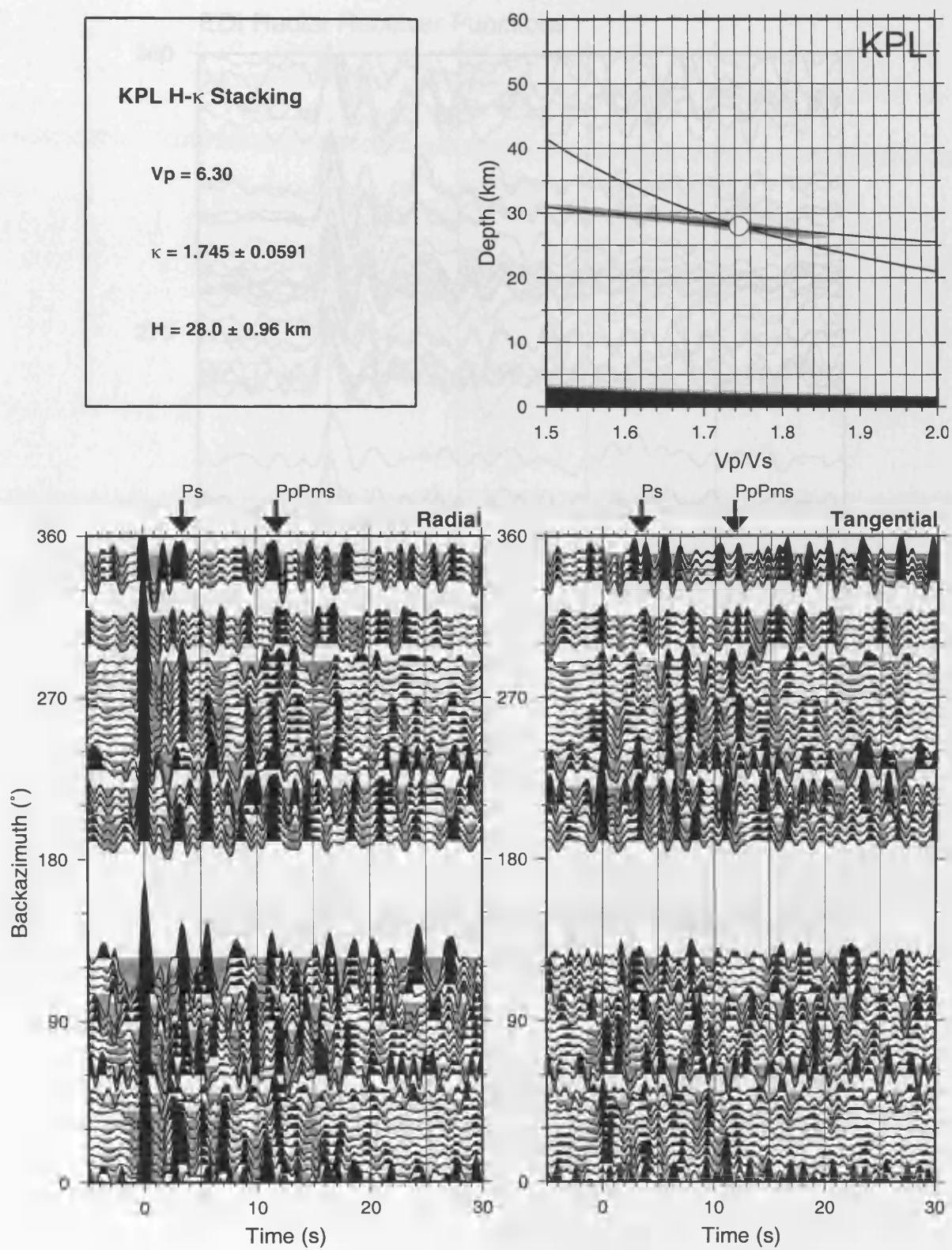


Figure A.13: The radial and tangential receiver functions for KPL, with the results of the H - κ stacking.

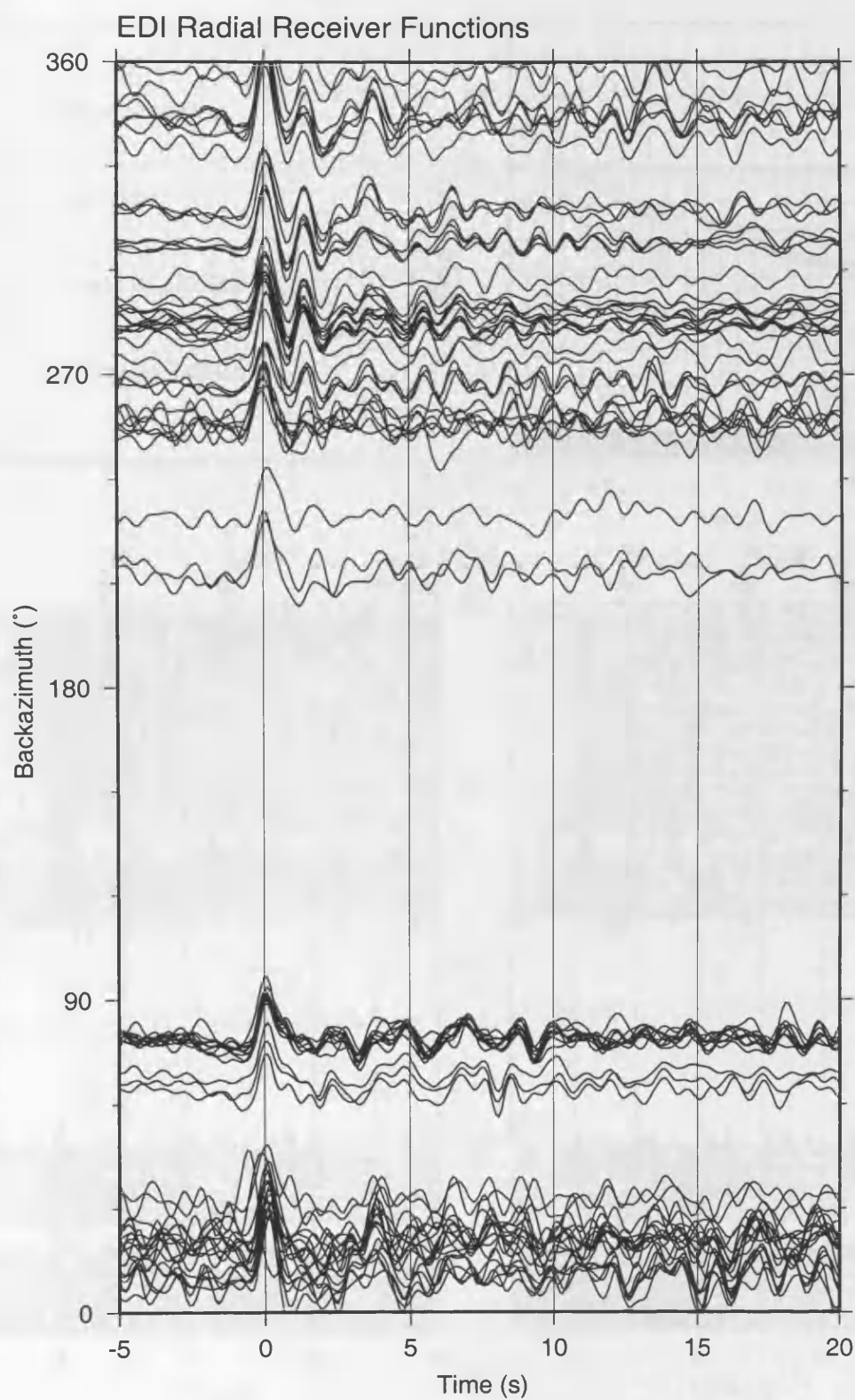


Figure A.14: Raw receiver function data from EDI plotted relative to backazimuth.

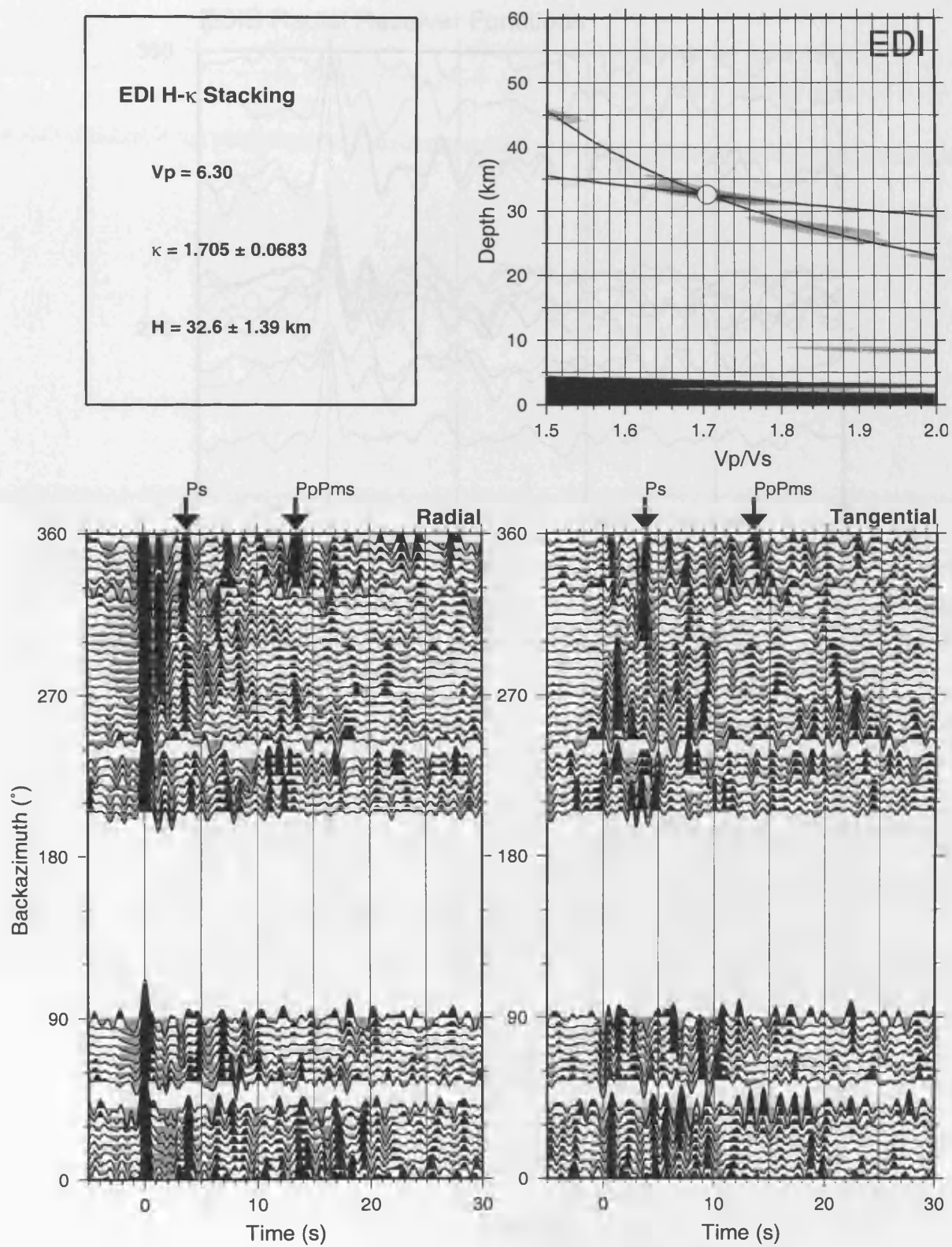


Figure A.15: The radial and tangential receiver functions for EDI, with the results of the H - κ stacking.

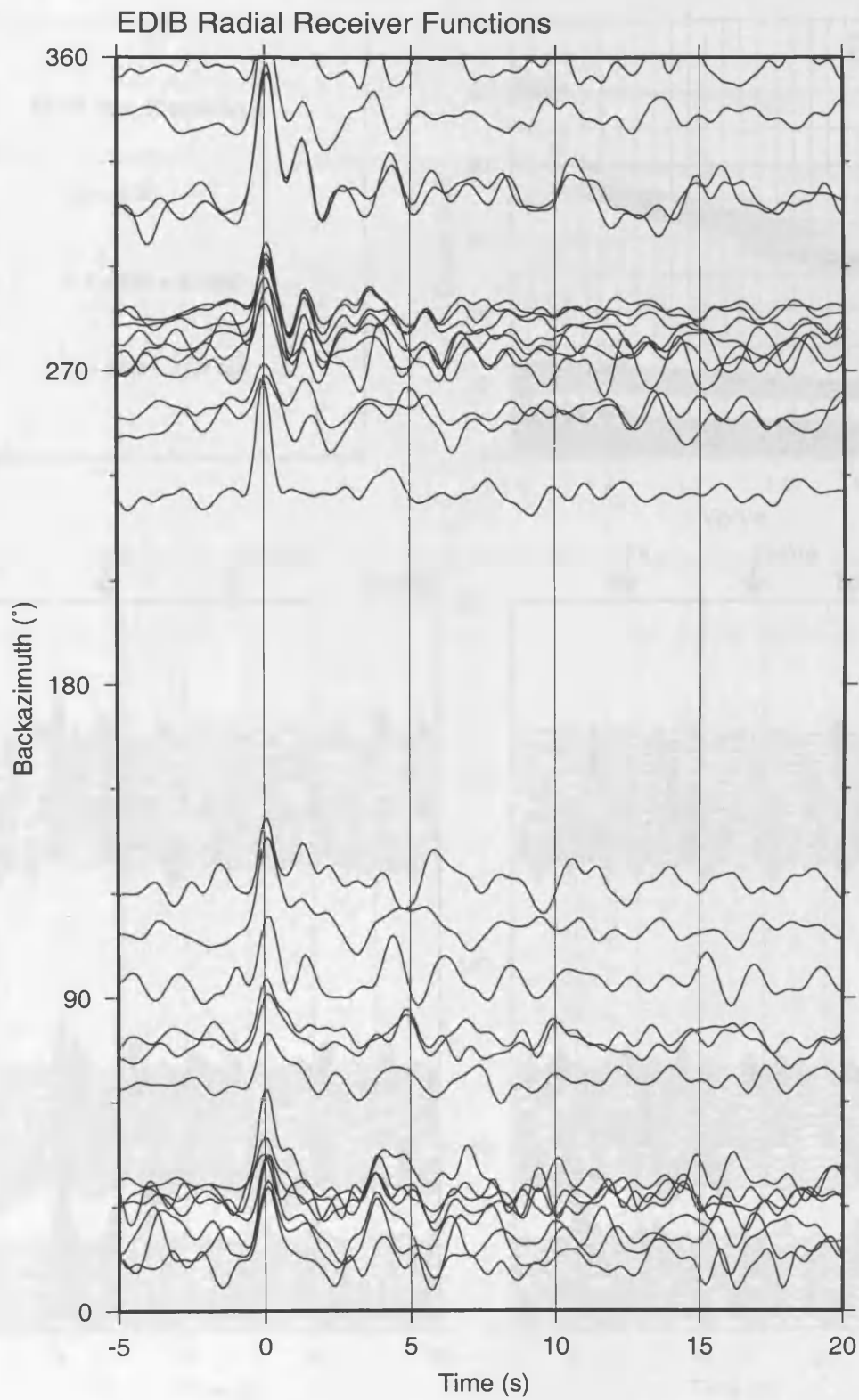
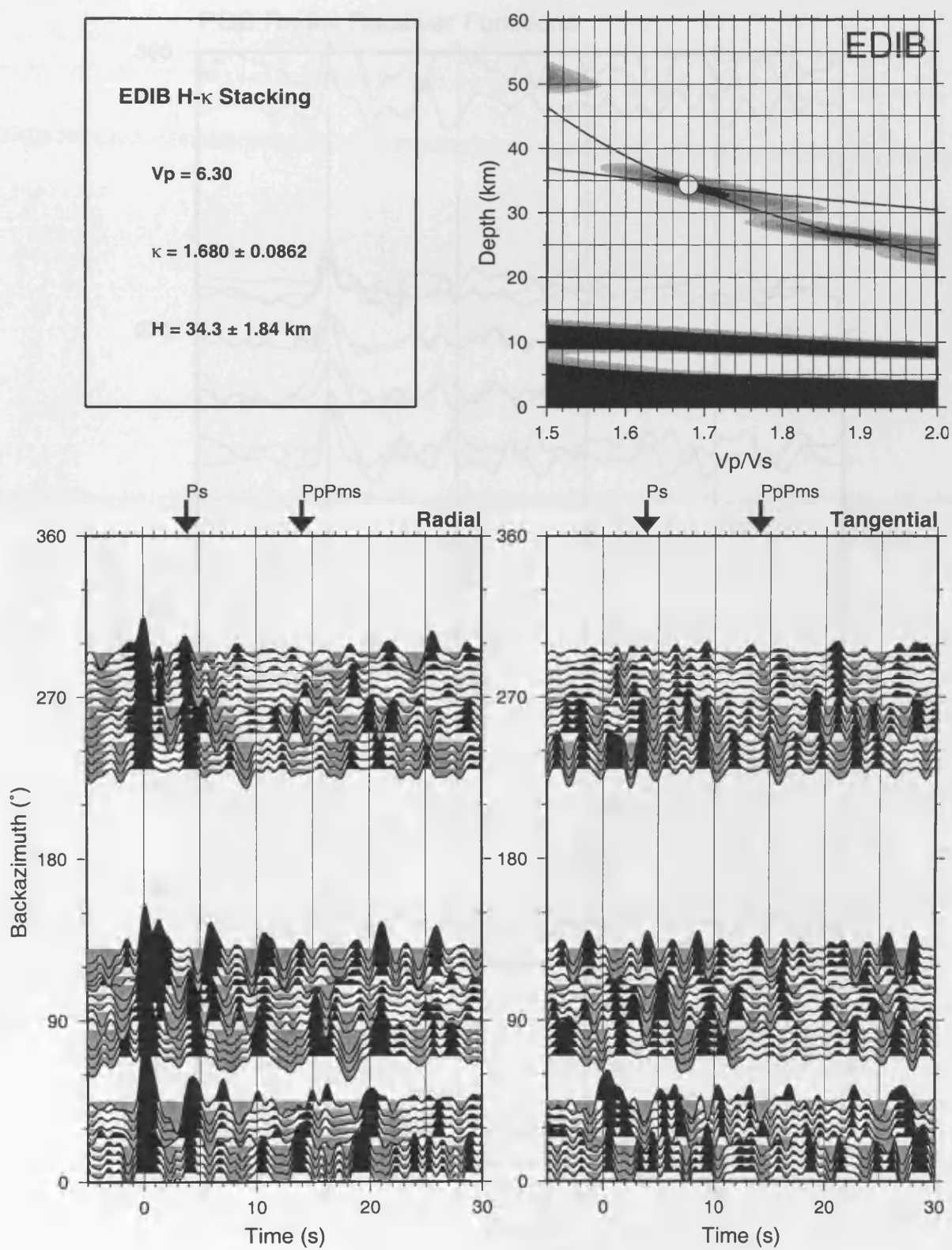


Figure A.16: Raw receiver function data from EDIB plotted relative to backazimuth.

Figure A.17: The radial and tangential receiver functions for EDIB, with the results of the H - κ stacking.

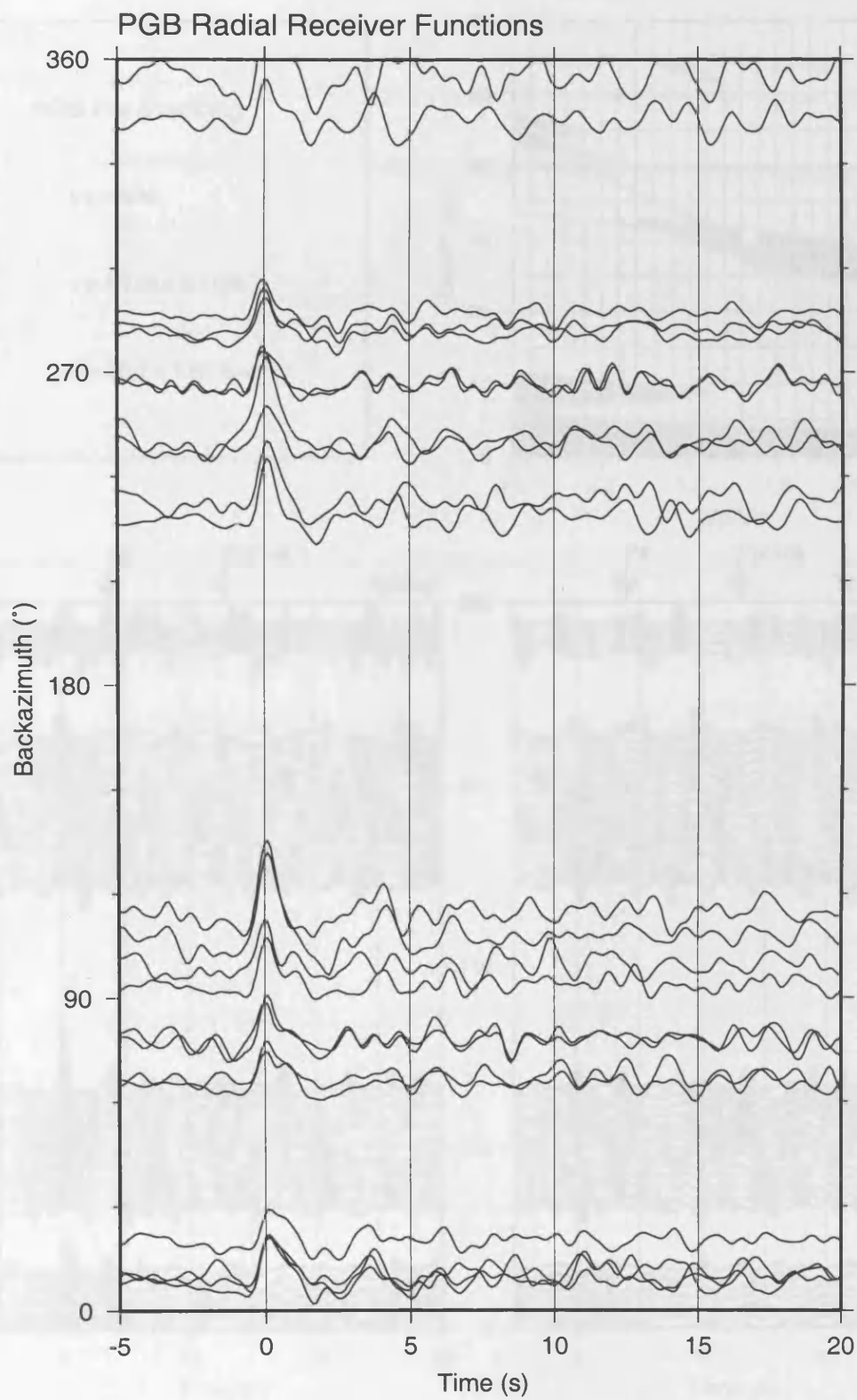


Figure A.18: Raw receiver function data from PGB plotted relative to backazimuth.

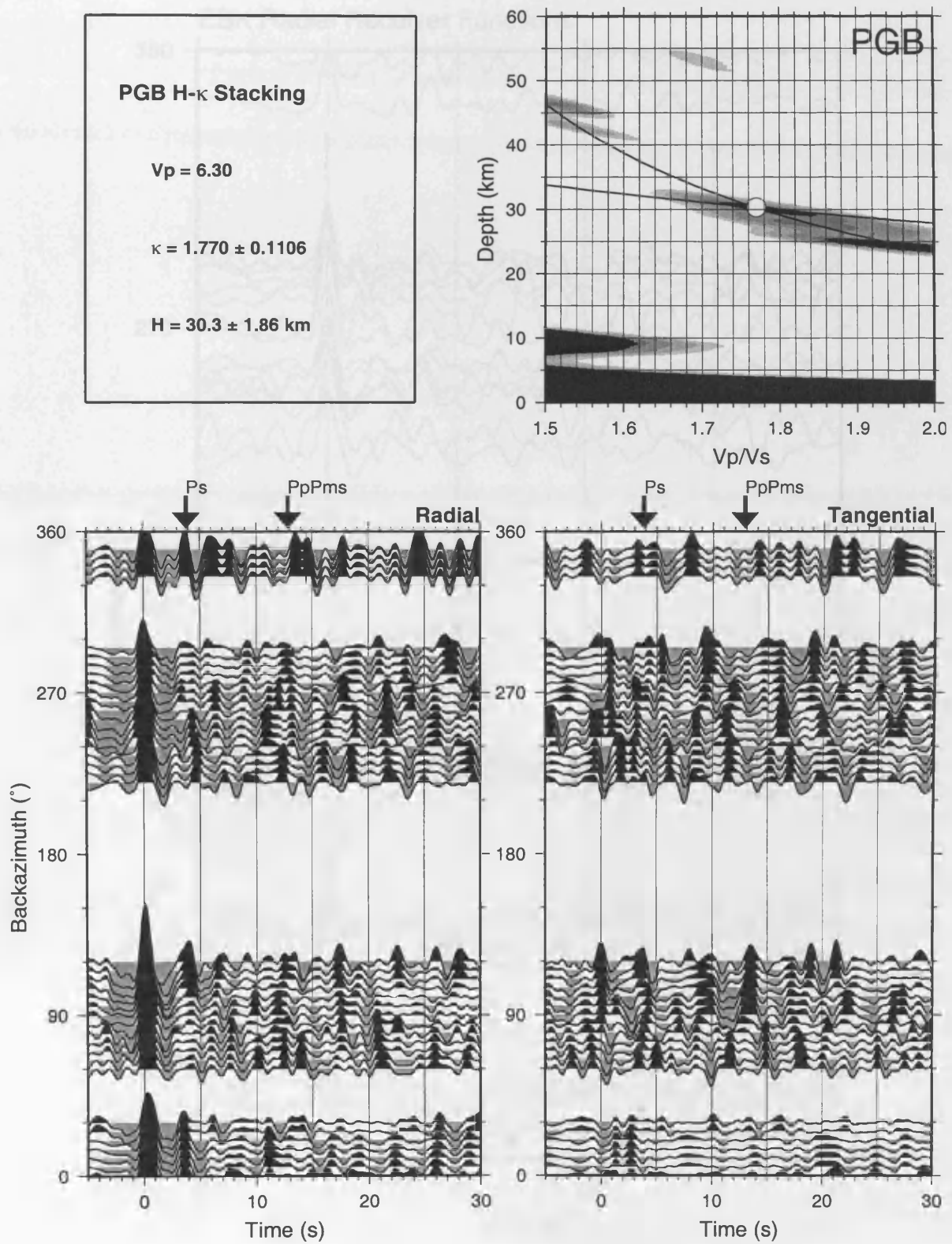


Figure A.19: The radial and tangential receiver functions for PGB, with the results of the $H-\kappa$ stacking.

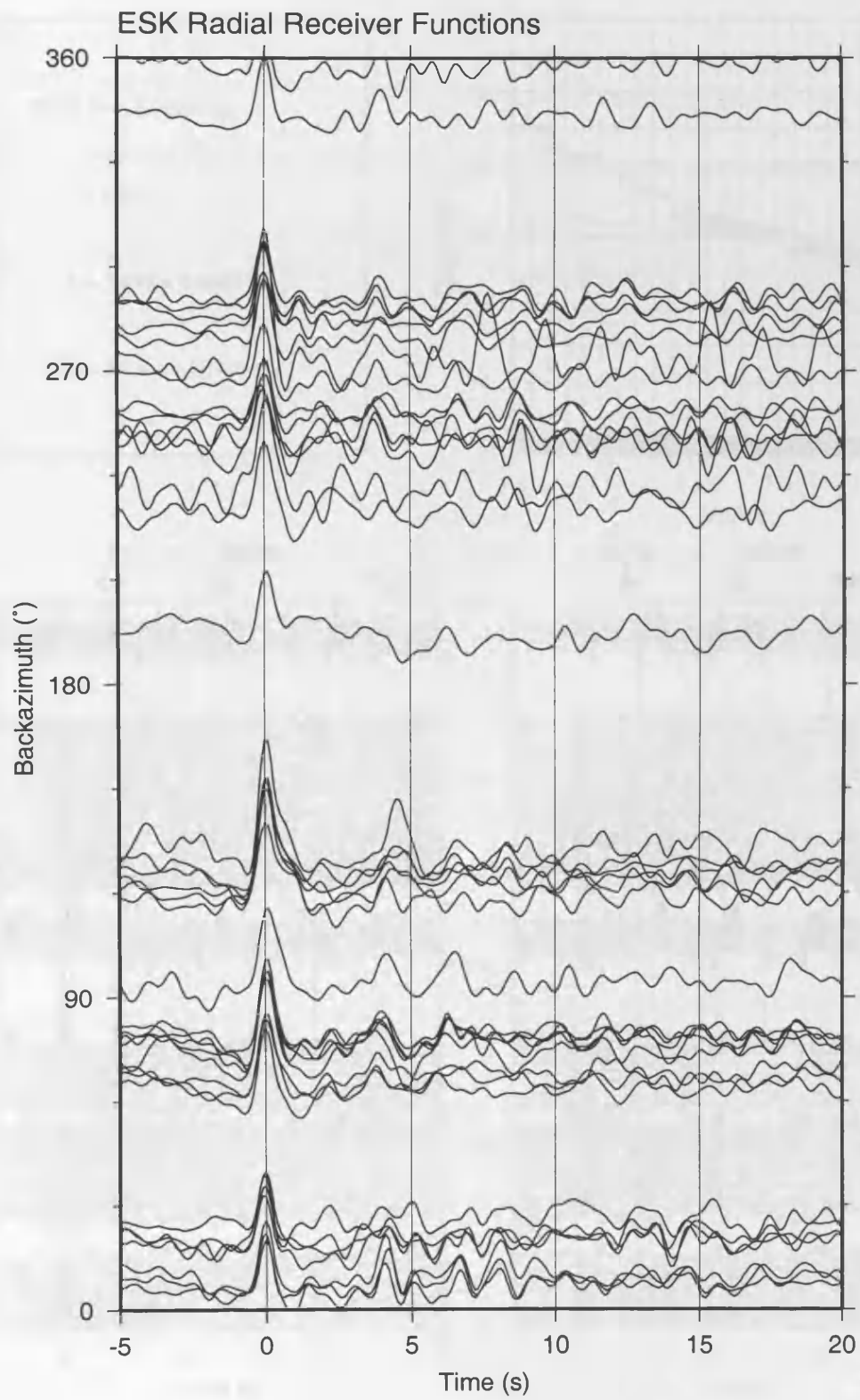


Figure A.20: Raw receiver function data from ESK plotted relative to backazimuth.

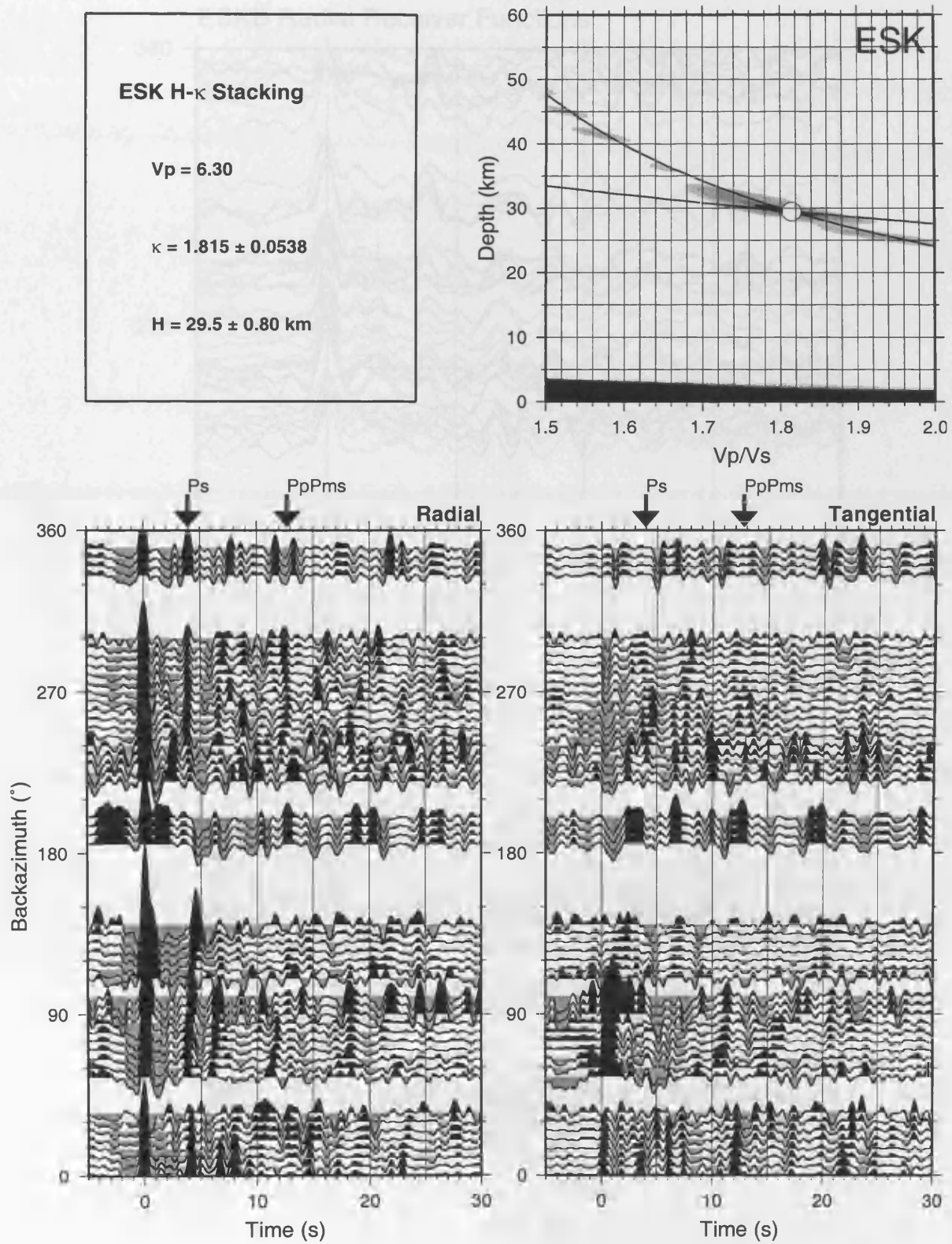


Figure A.21: The radial and tangential receiver functions for ESK, with the results of the H - κ stacking.

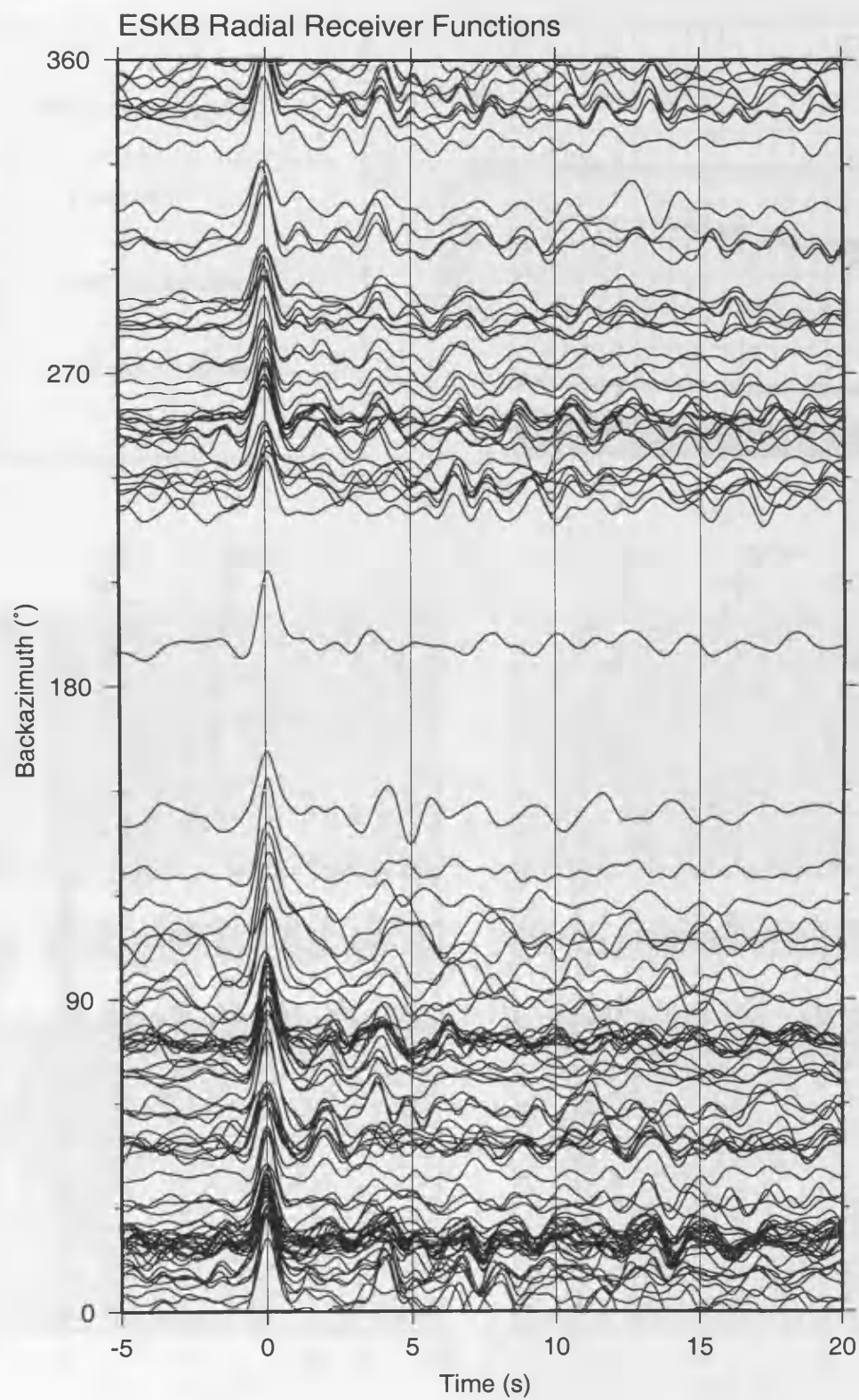
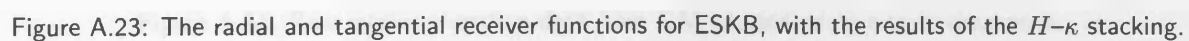


Figure A.22: Raw receiver function data from ESKB plotted relative to backazimuth.



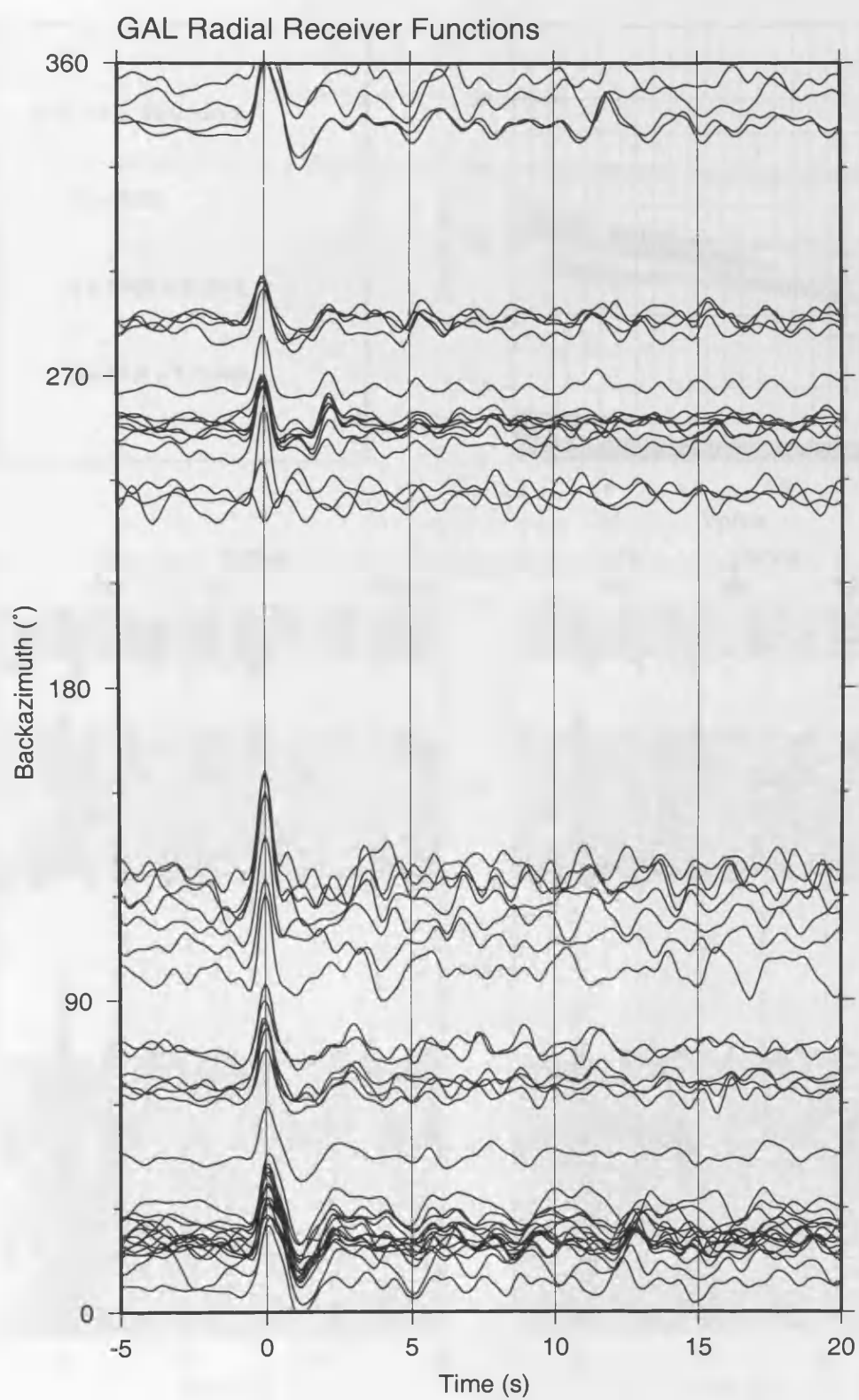


Figure A.24: Raw receiver function data from GAL plotted relative to backazimuth.

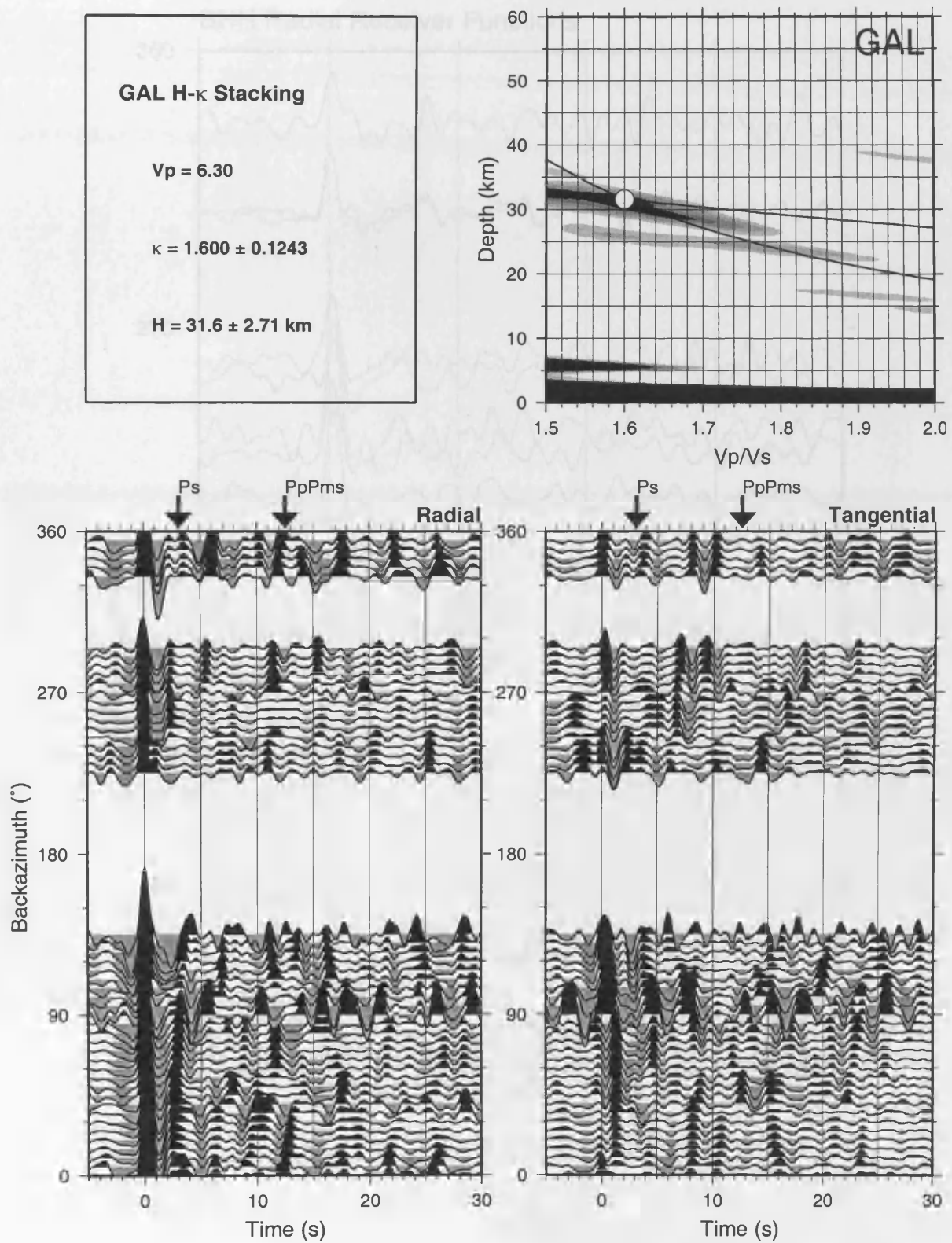


Figure A.25: The radial and tangential receiver functions for GAL, with the results of the $H-\kappa$ stacking.

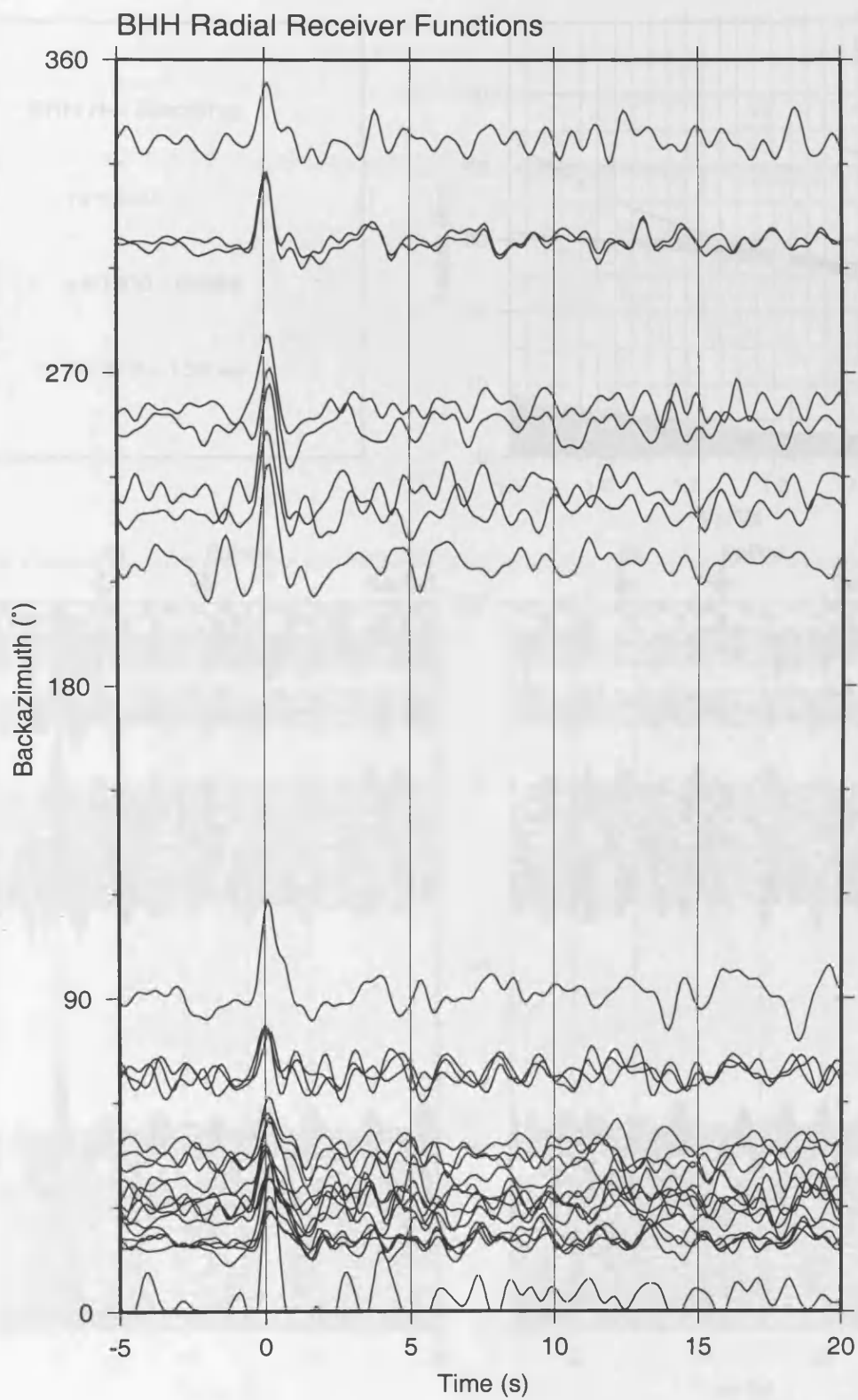


Figure A.26: Raw receiver function data from BHH plotted relative to backazimuth.

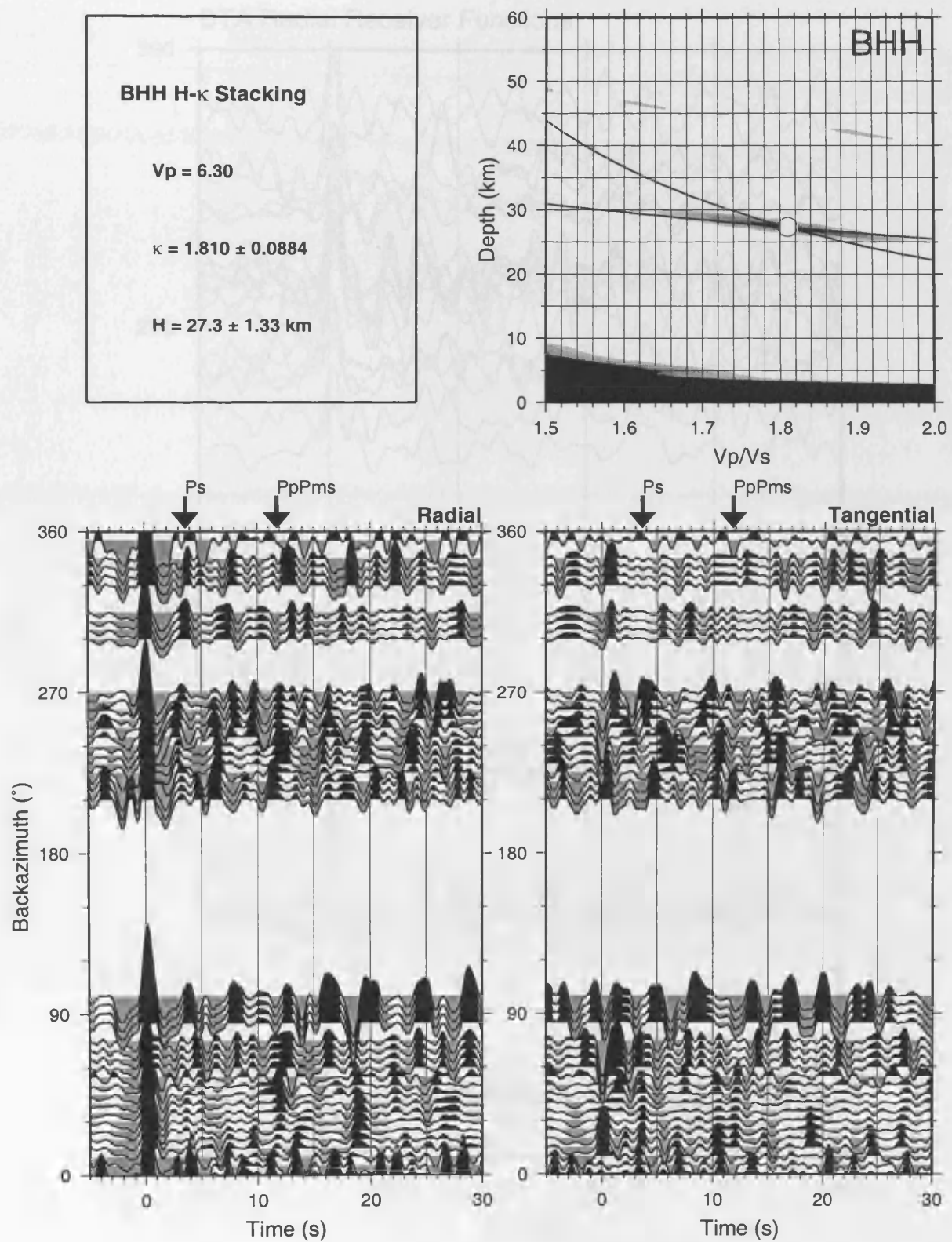


Figure A.27: The radial and tangential receiver functions for BHH, with the results of the H - κ stacking.

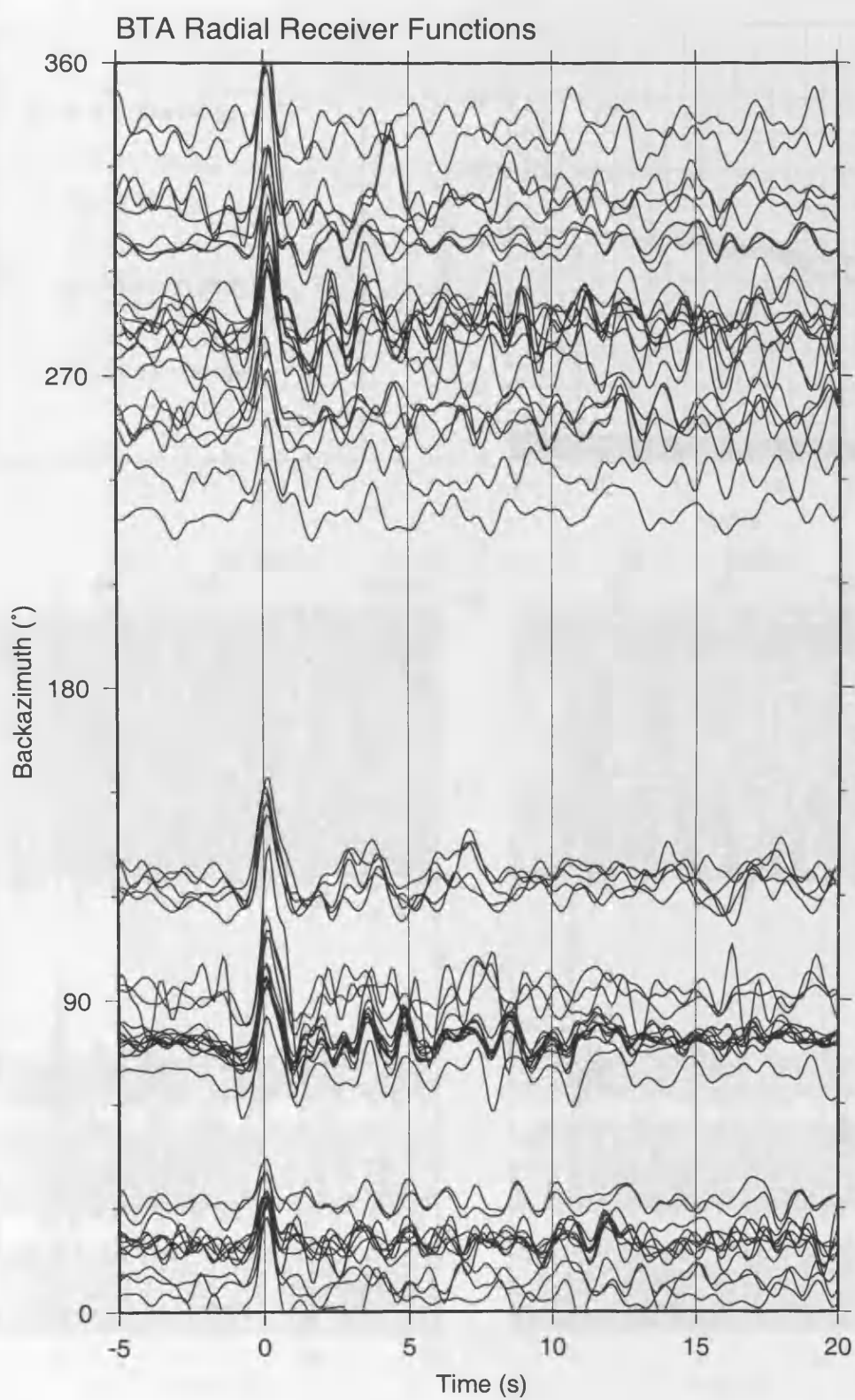


Figure A.28: Raw receiver function data from BTA plotted relative to backazimuth.

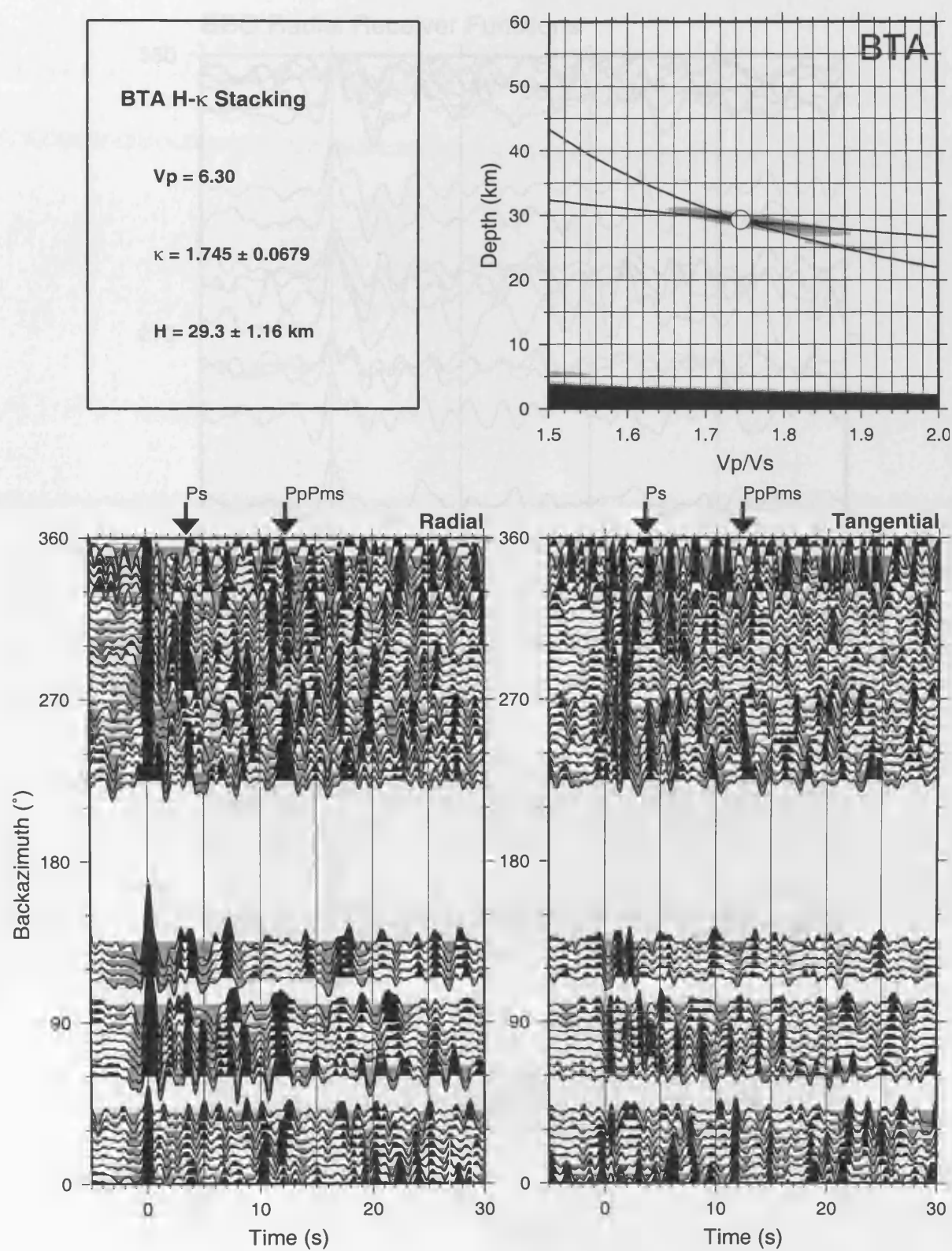


Figure A.29: The radial and tangential receiver functions for BTA, with the results of the H - κ stacking.

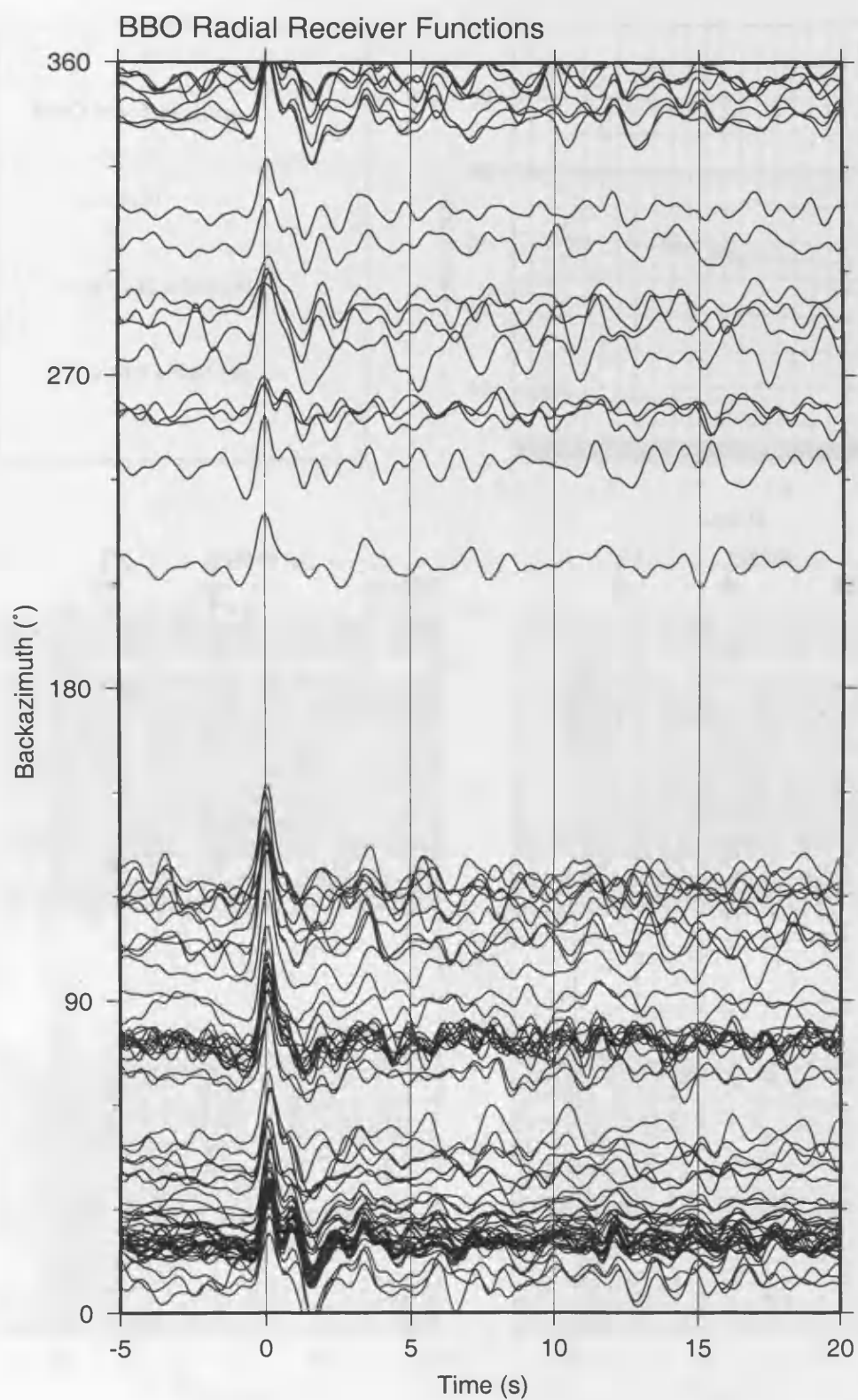


Figure A.30: Raw receiver function data from BBO plotted relative to backazimuth.

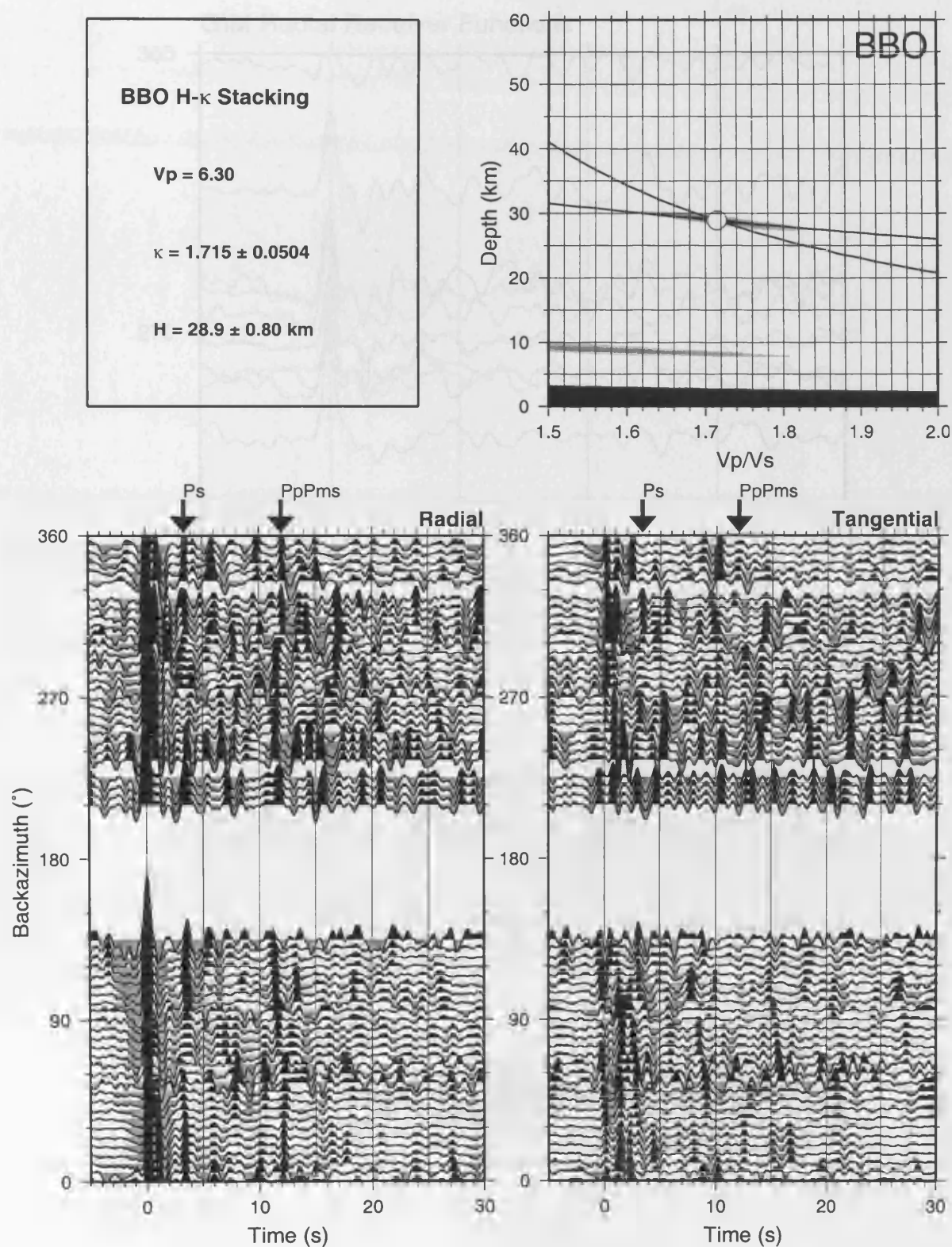


Figure A.31: The radial and tangential receiver functions for BBO, with the results of the H - κ stacking.

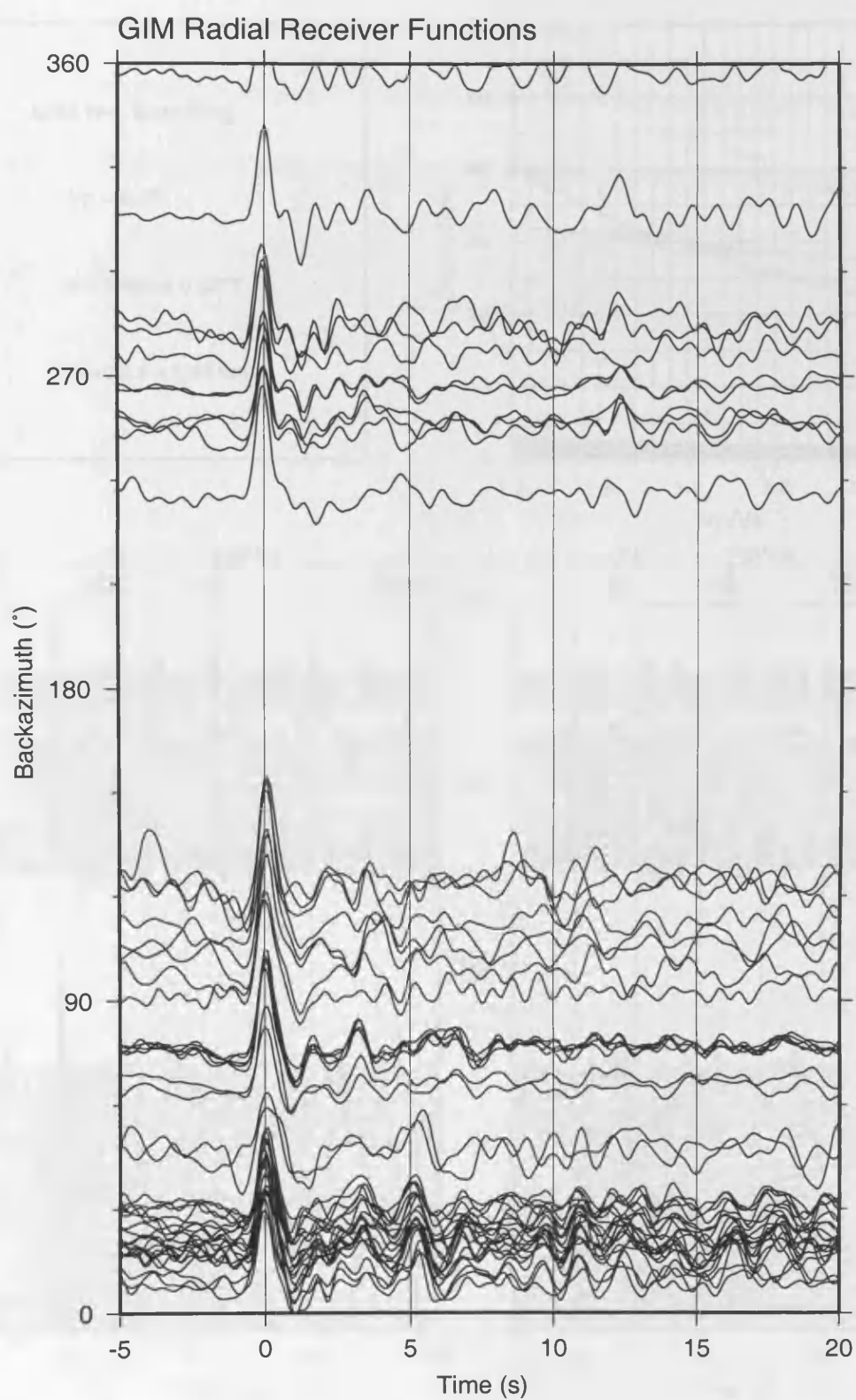


Figure A.32: Raw receiver function data from GIM plotted relative to backazimuth.

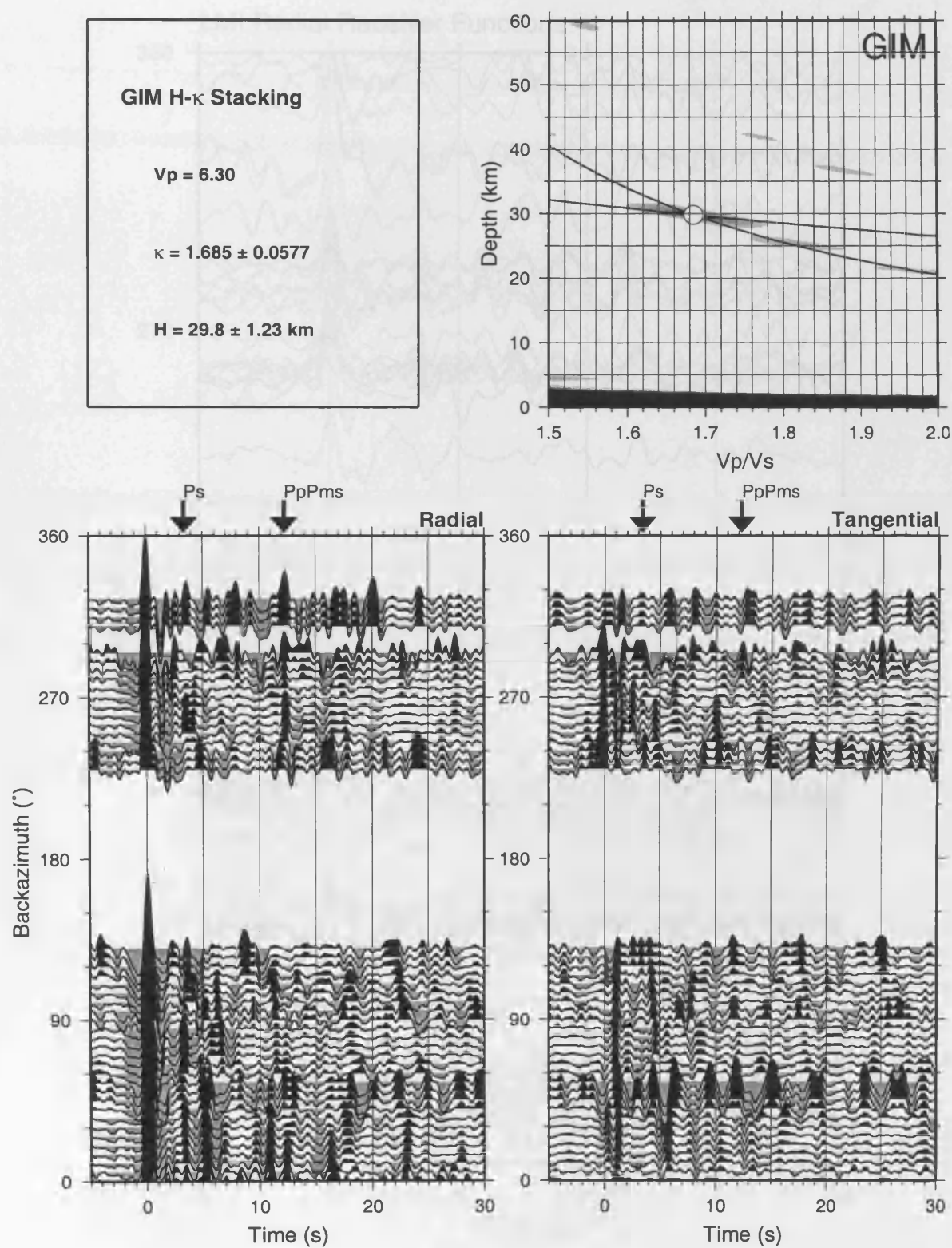


Figure A.33: The radial and tangential receiver functions for GIM, with the results of the H - κ stacking.

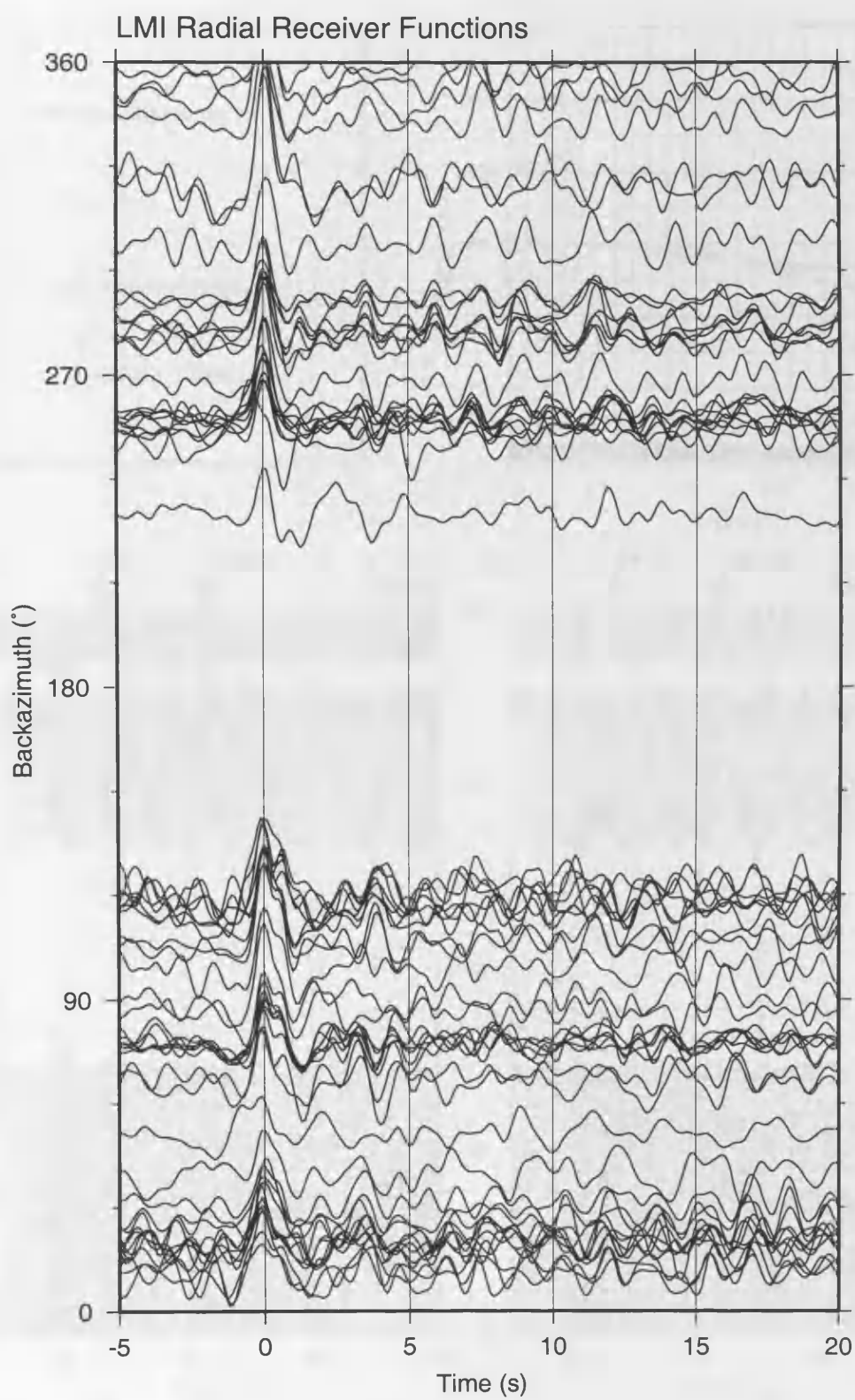


Figure A.34: Raw receiver function data from LMI plotted relative to backazimuth.

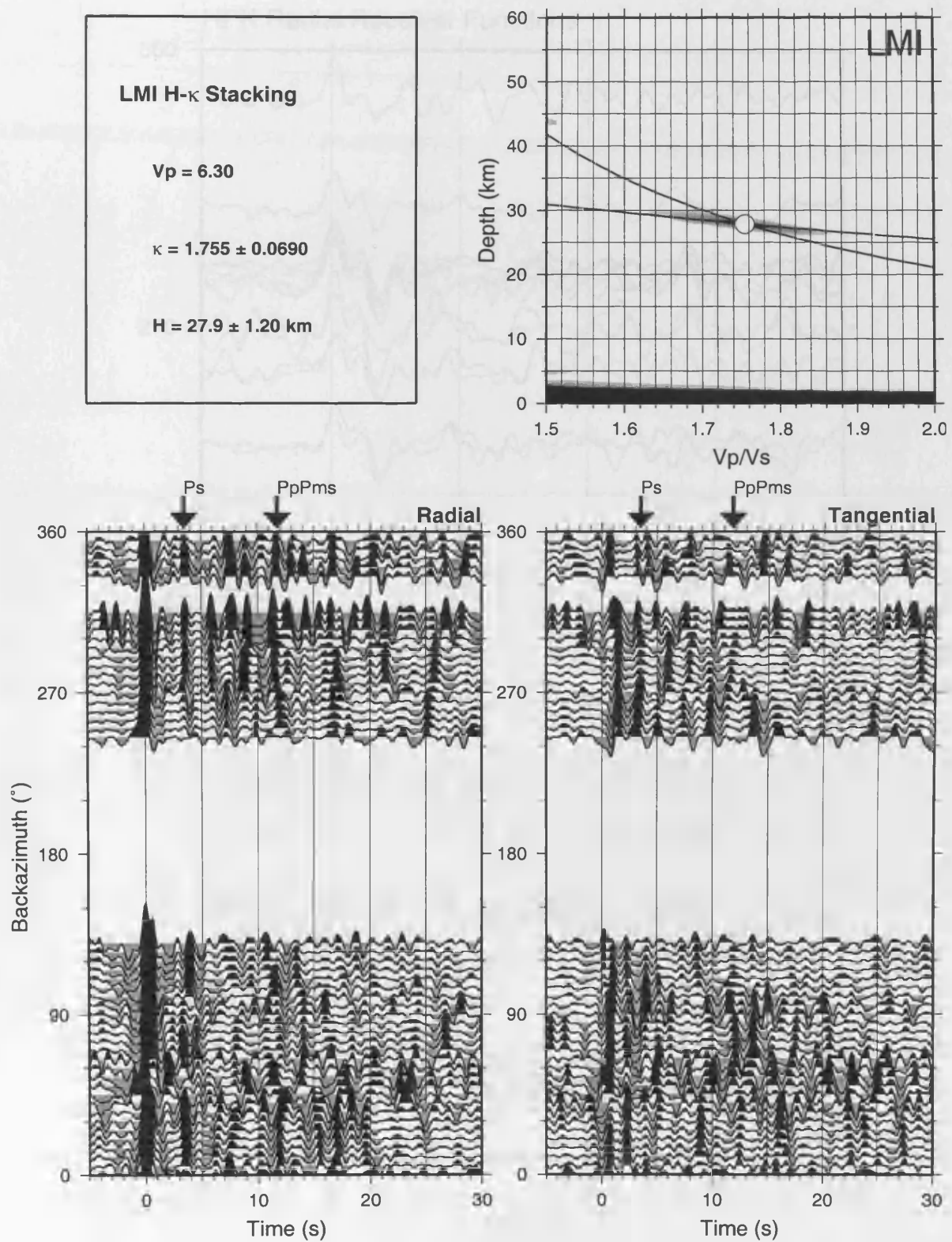


Figure A.35: The radial and tangential receiver functions for LMI, with the results of the H - κ stacking.

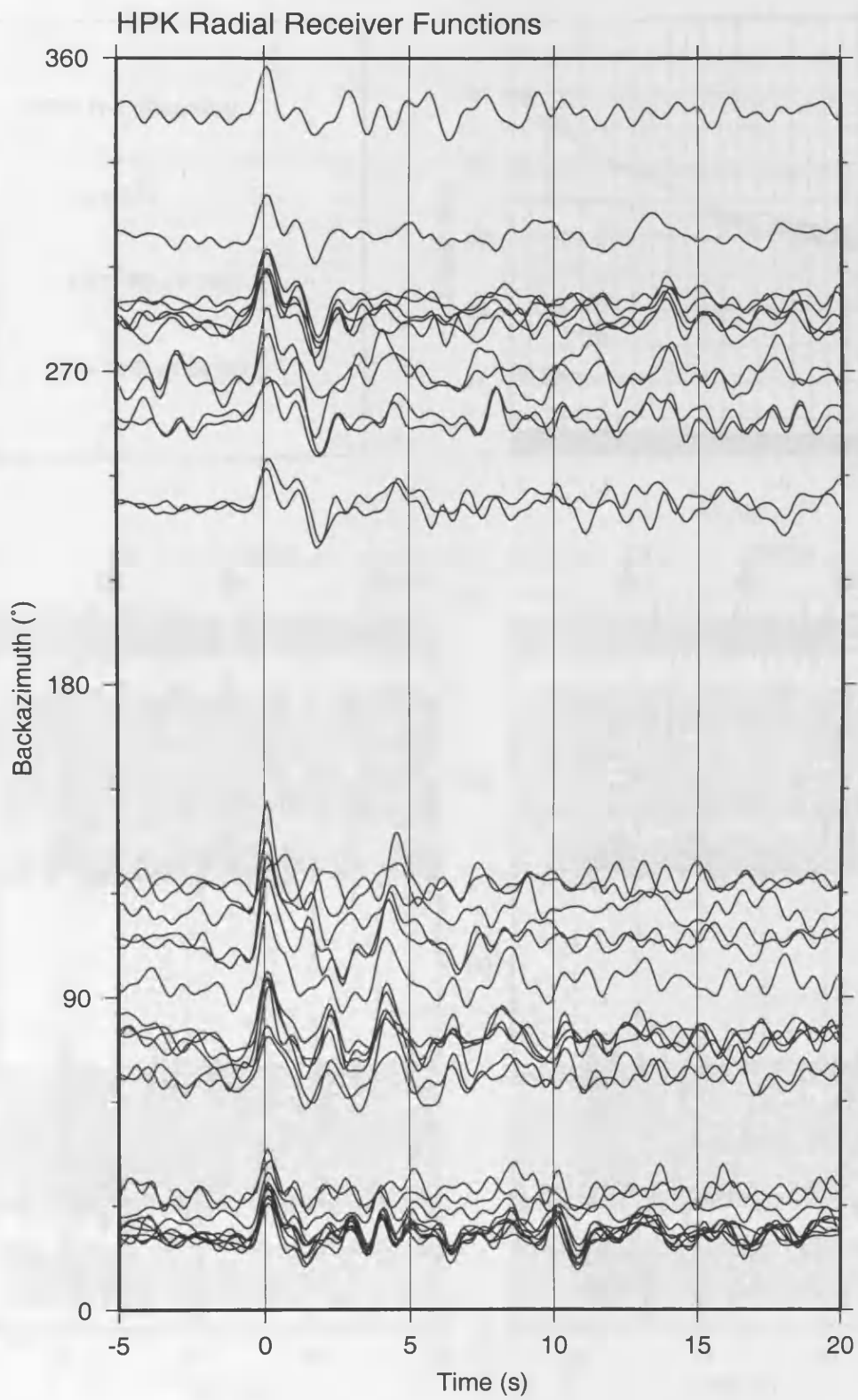


Figure A.36: Raw receiver function data from HPK plotted relative to backazimuth.

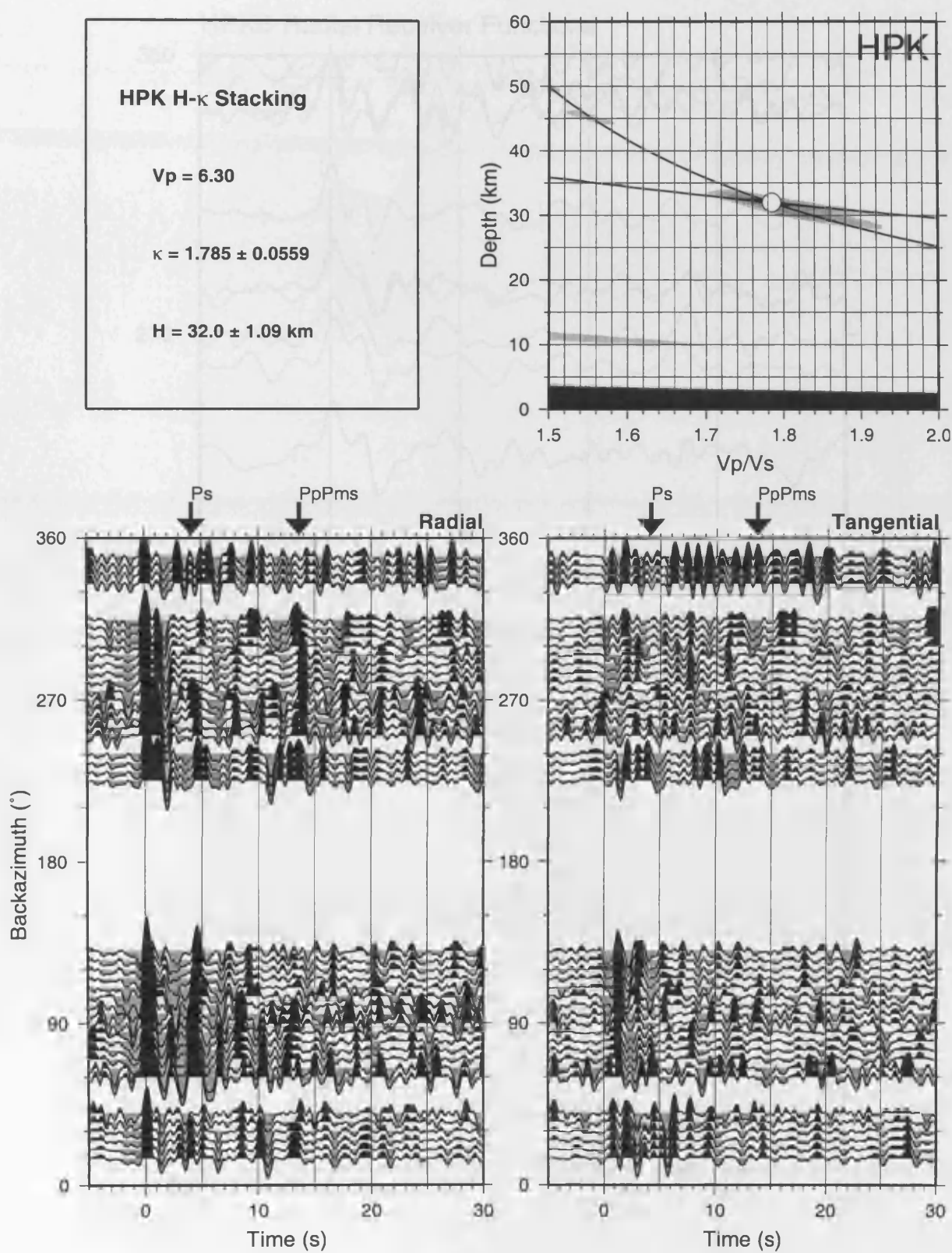


Figure A.37: The radial and tangential receiver functions for HPK, with the results of the H - κ stacking.

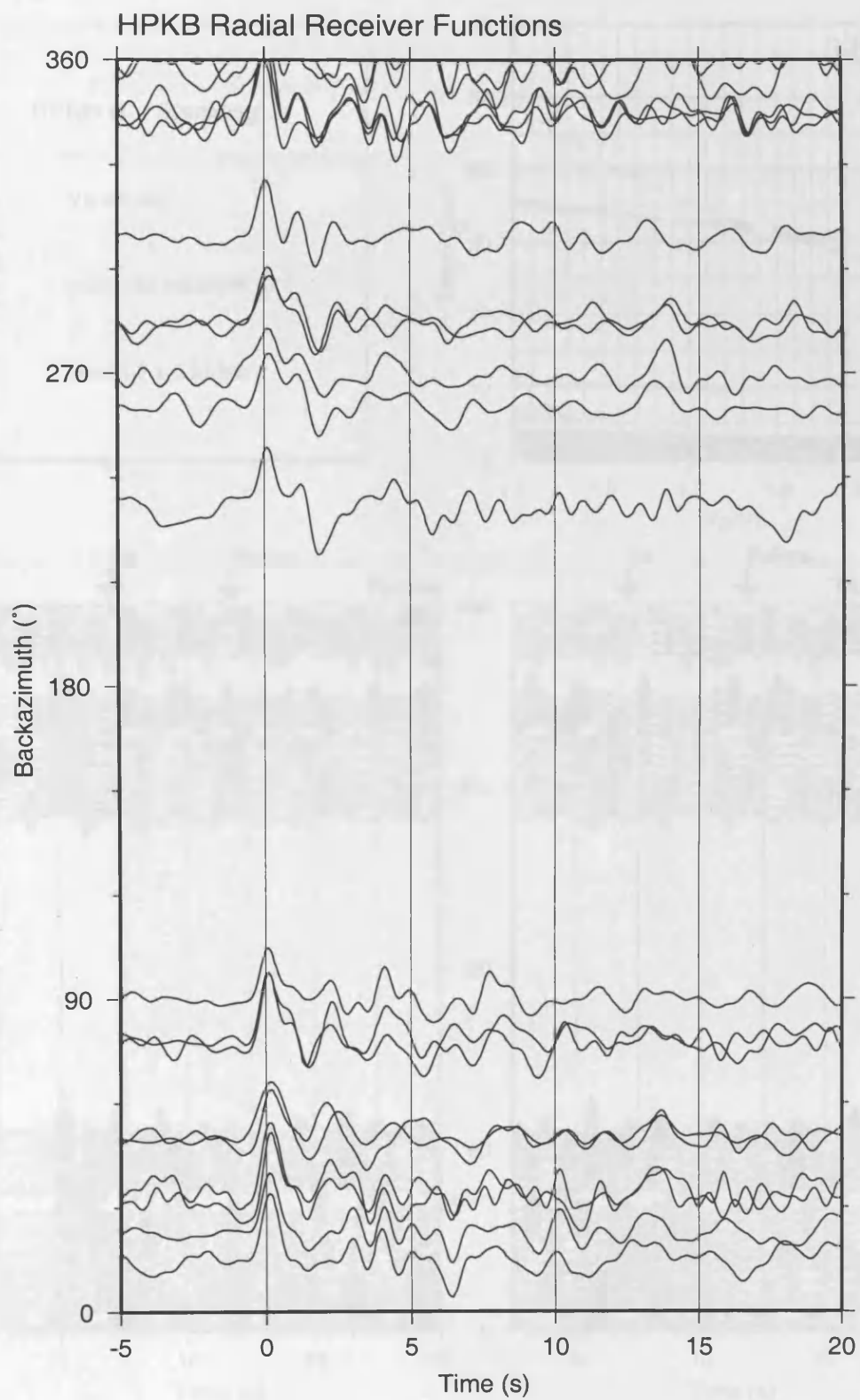


Figure A.38: Raw receiver function data from HPKB plotted relative to backazimuth.

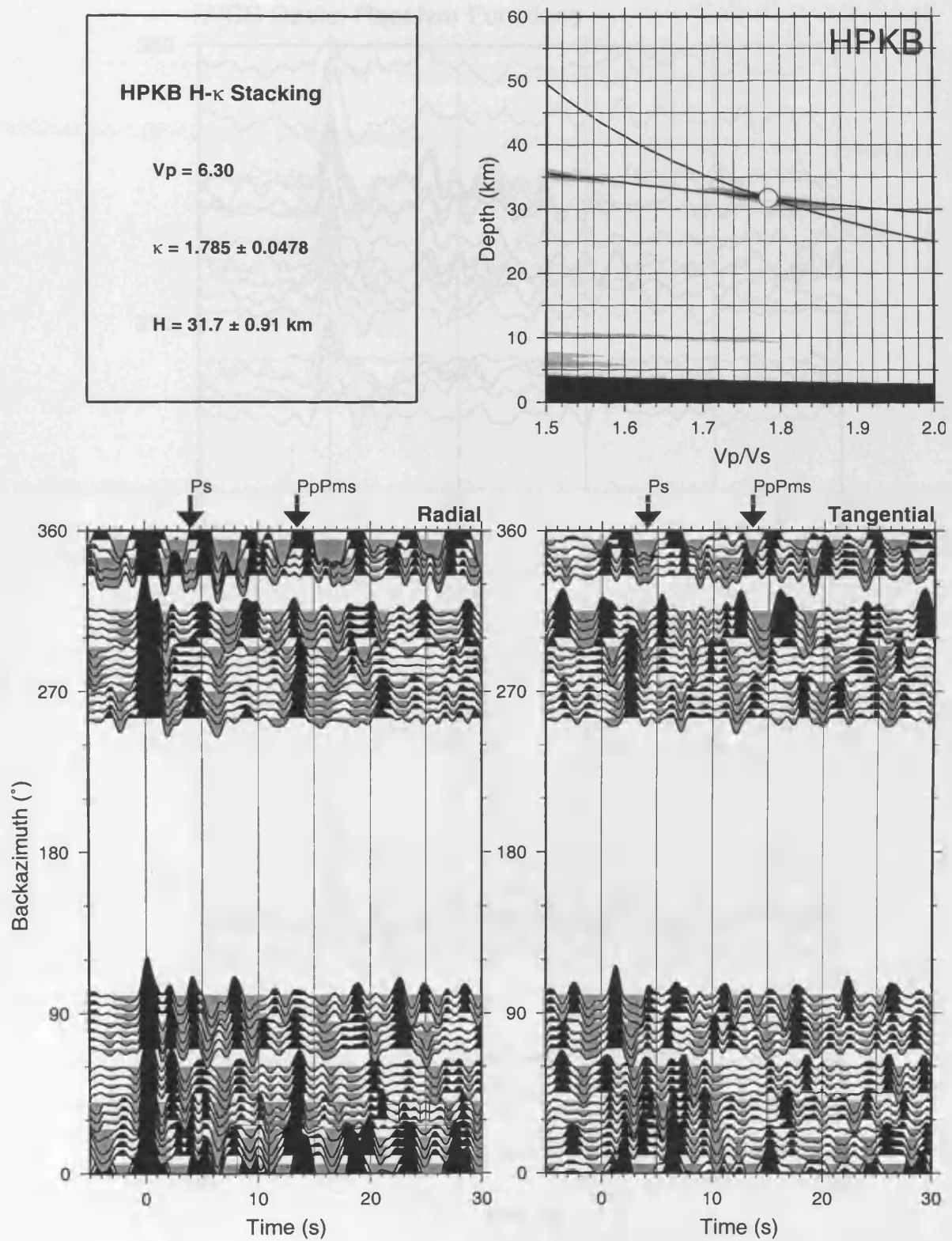


Figure A.39: The radial and tangential receiver functions for HPKB, with the results of the H - κ stacking.

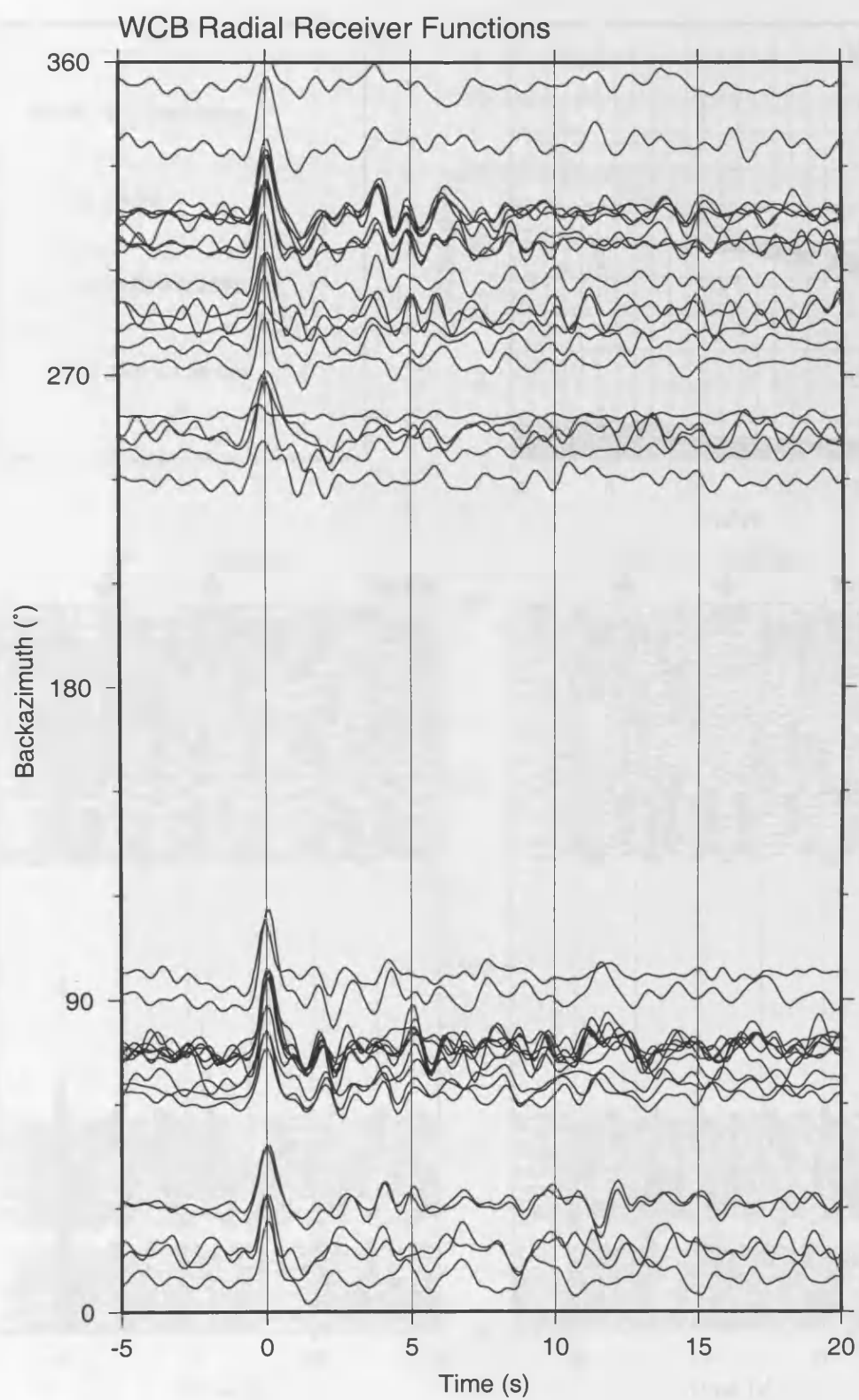
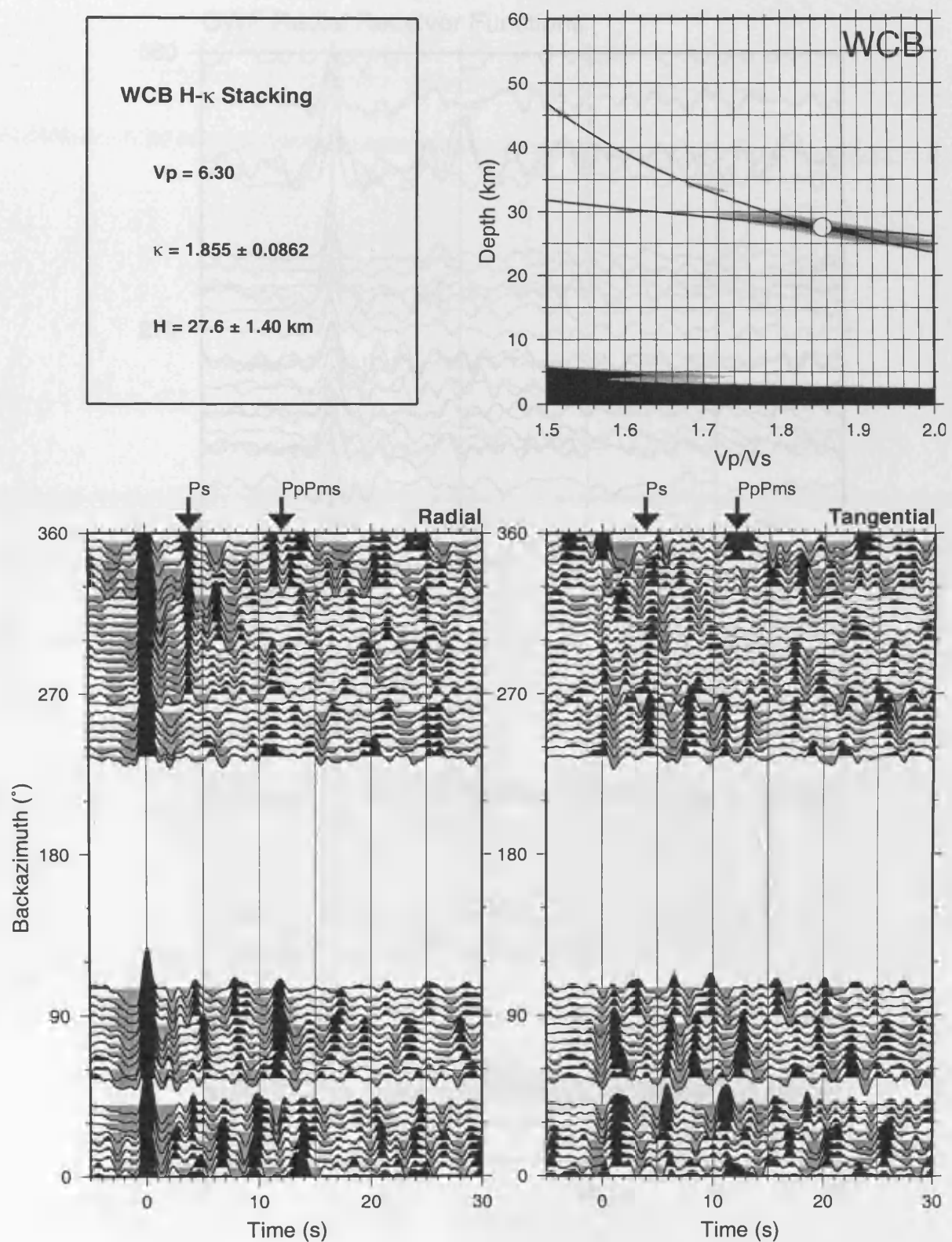


Figure A.40: Raw receiver function data from WCB plotted relative to backazimuth.

Figure A.41: The radial and tangential receiver functions for WCB, with the results of the H - κ stacking.

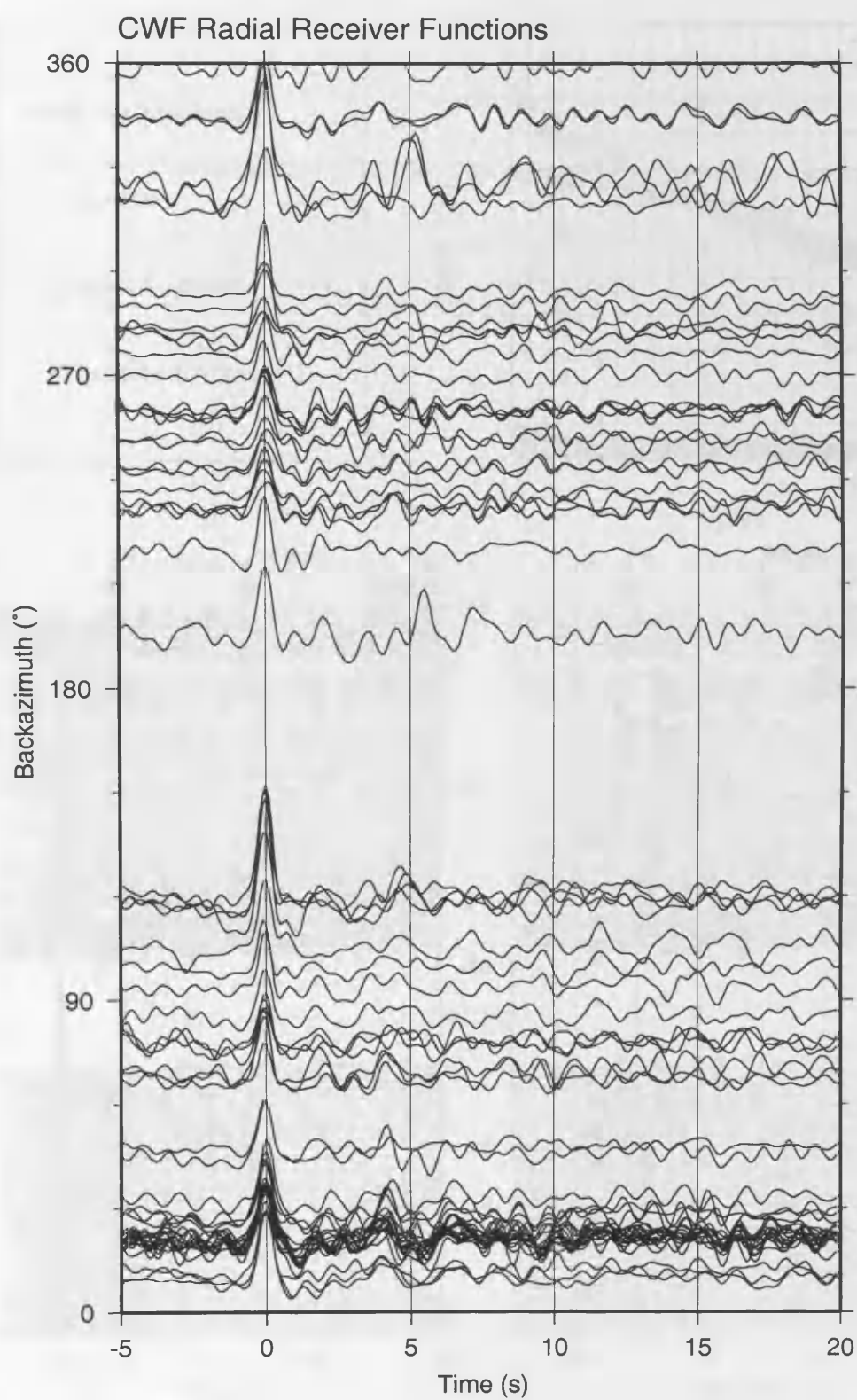


Figure A.42: Raw receiver function data from CWF plotted relative to backazimuth.

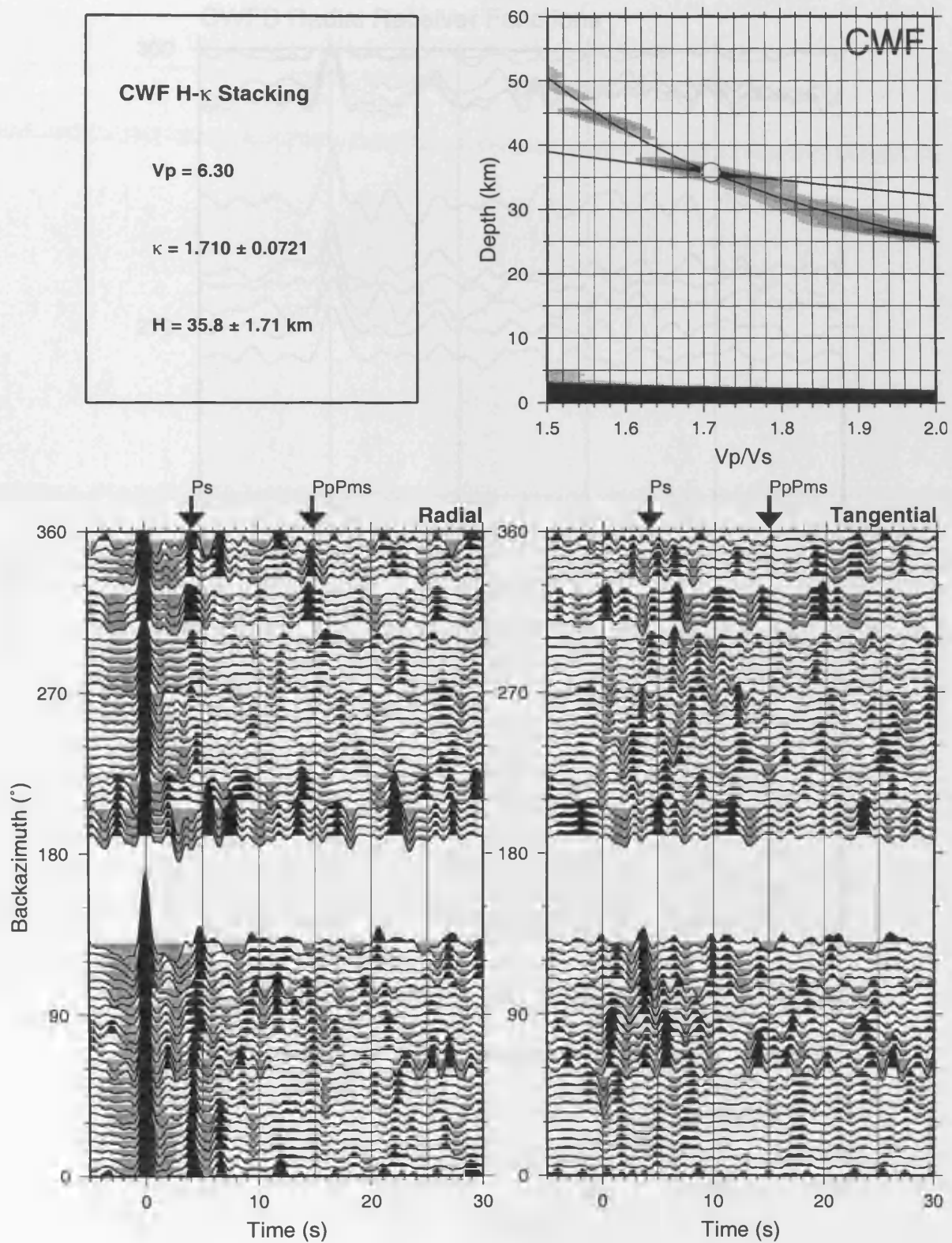


Figure A.43: The radial and tangential receiver functions for CWF, with the results of the H - κ stacking.

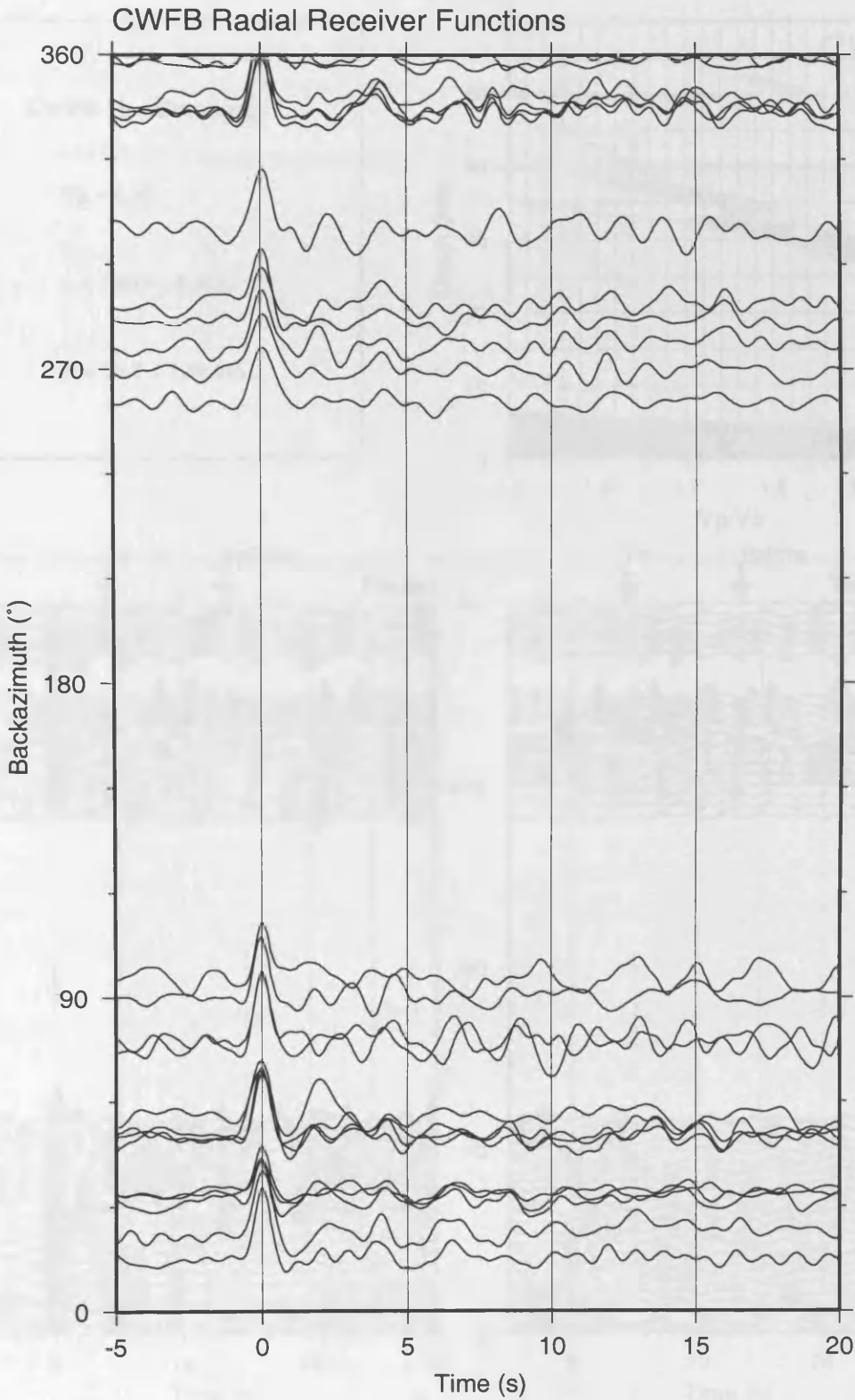


Figure A.44: Raw receiver function data from CWFB plotted relative to backazimuth.

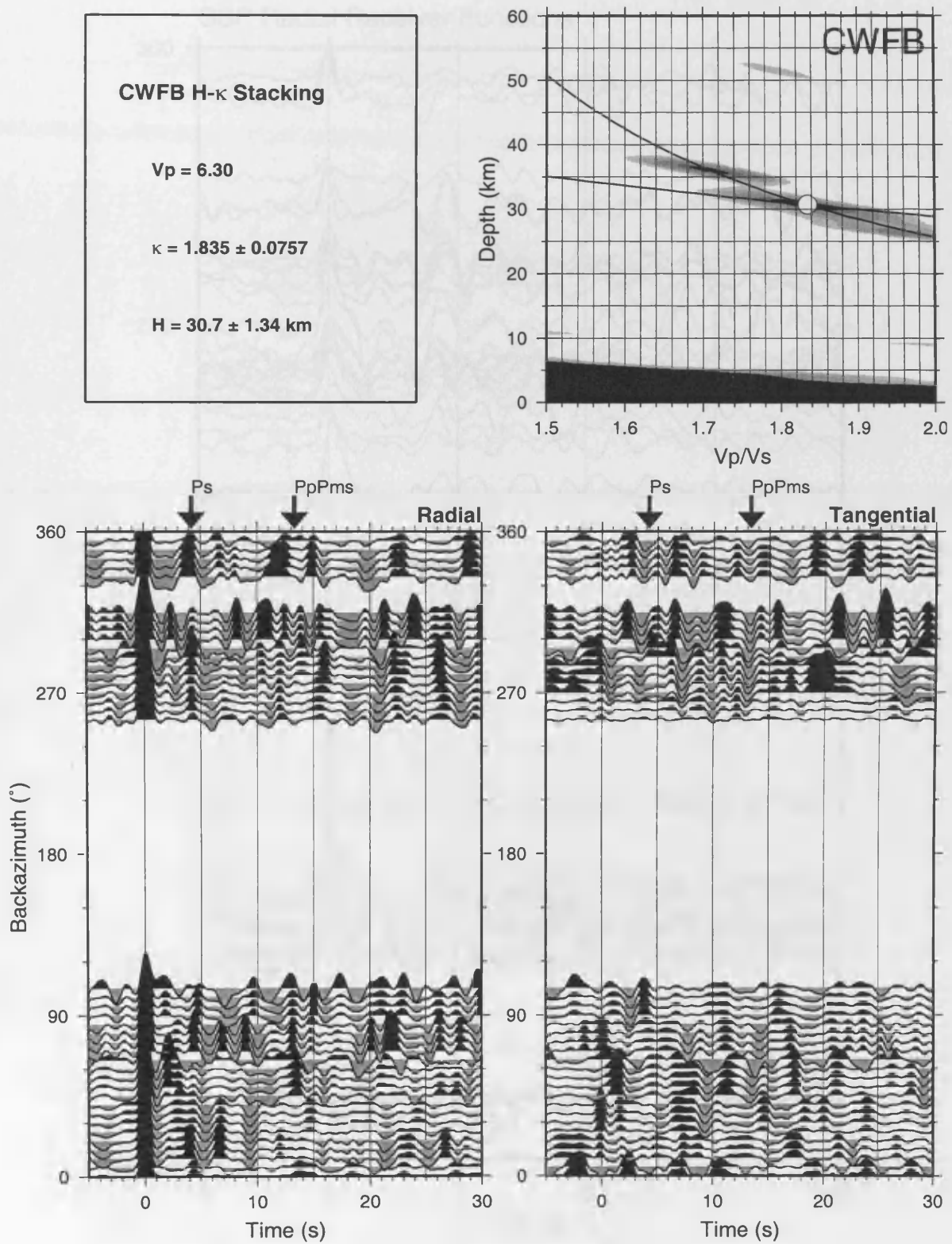


Figure A.45: The radial and tangential receiver functions for CWFB, with the results of the H - κ stacking.

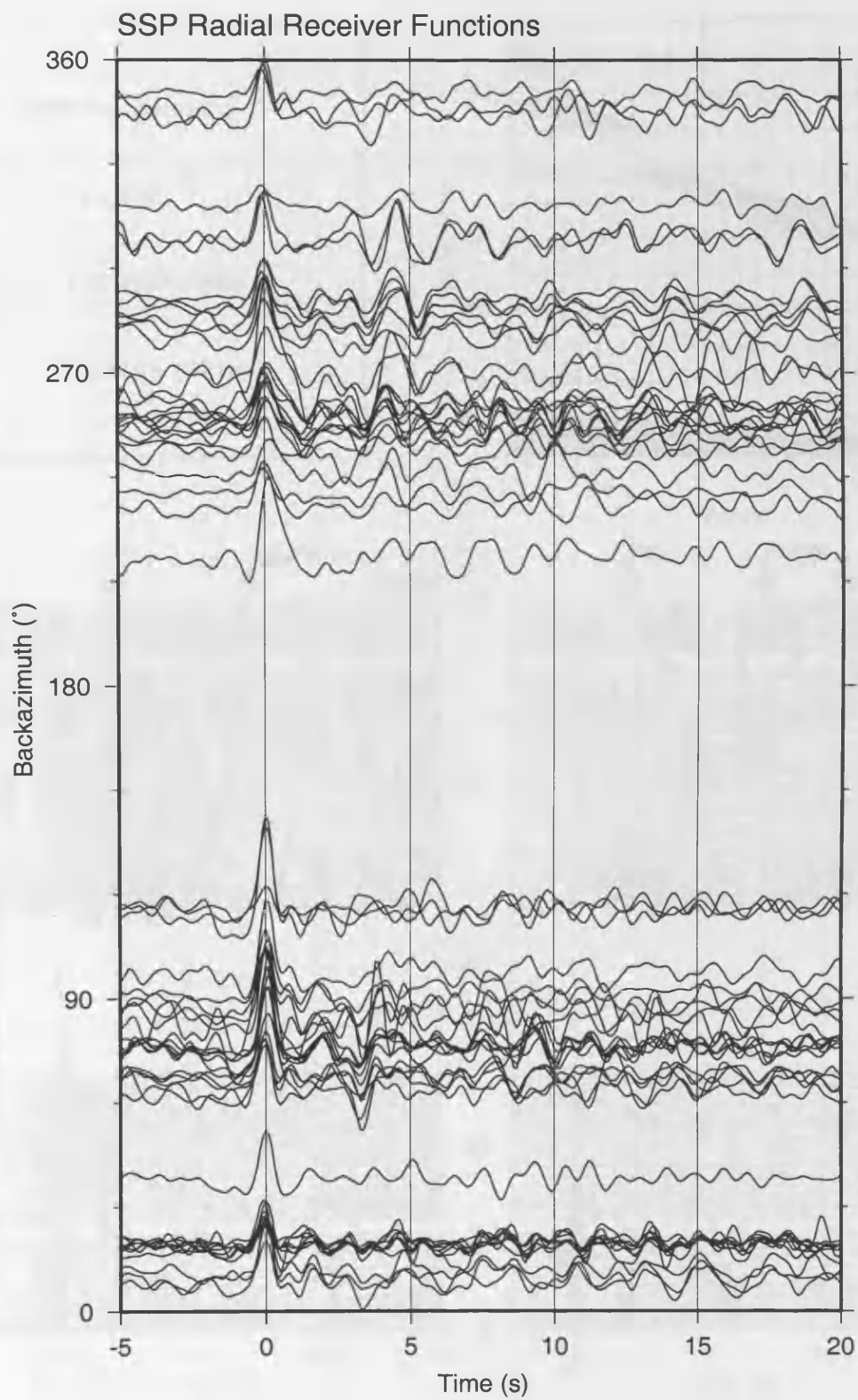


Figure A.46: Raw receiver function data from SSP plotted relative to backazimuth.

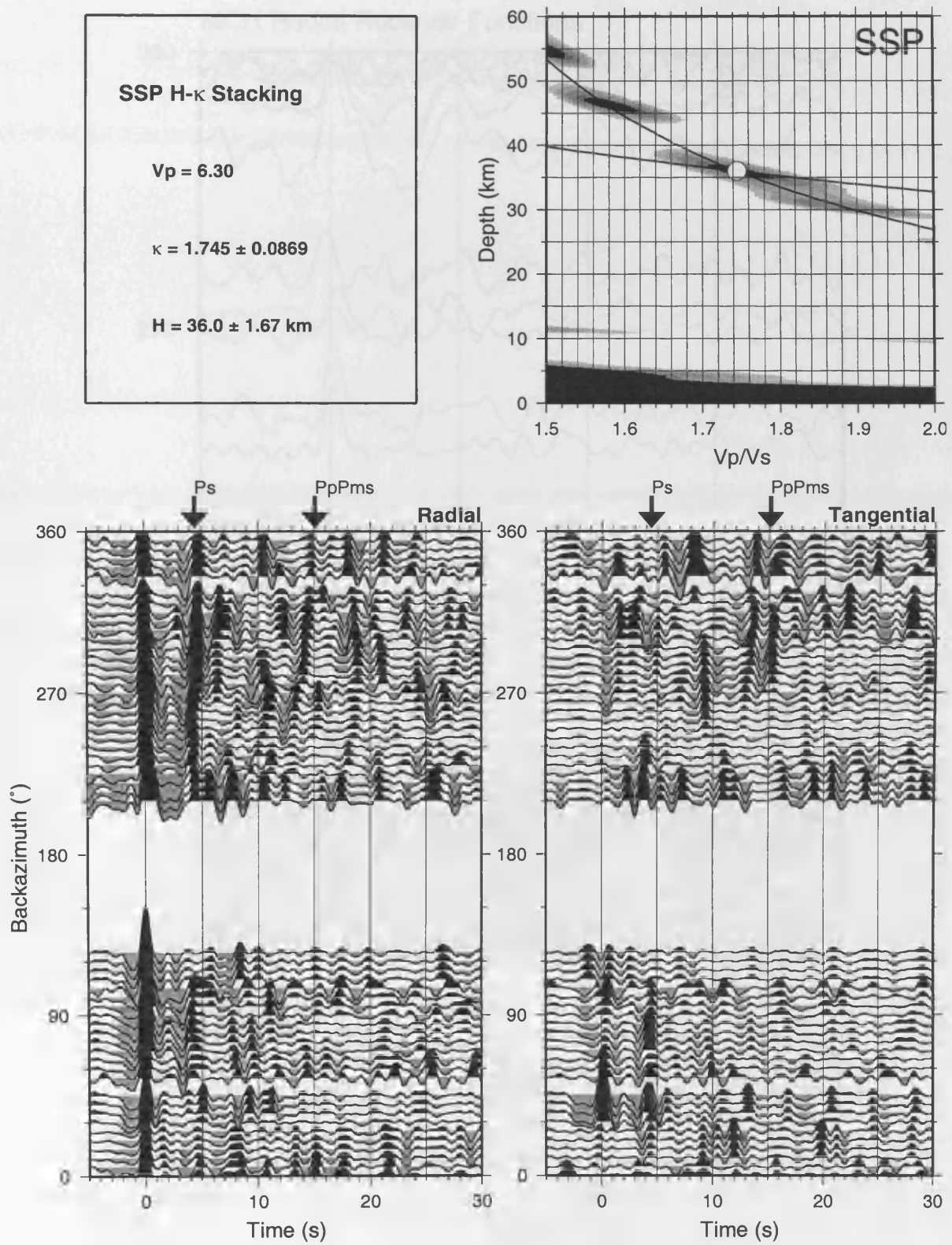


Figure A.47: The radial and tangential receiver functions for SSP, with the results of the H - κ stacking.

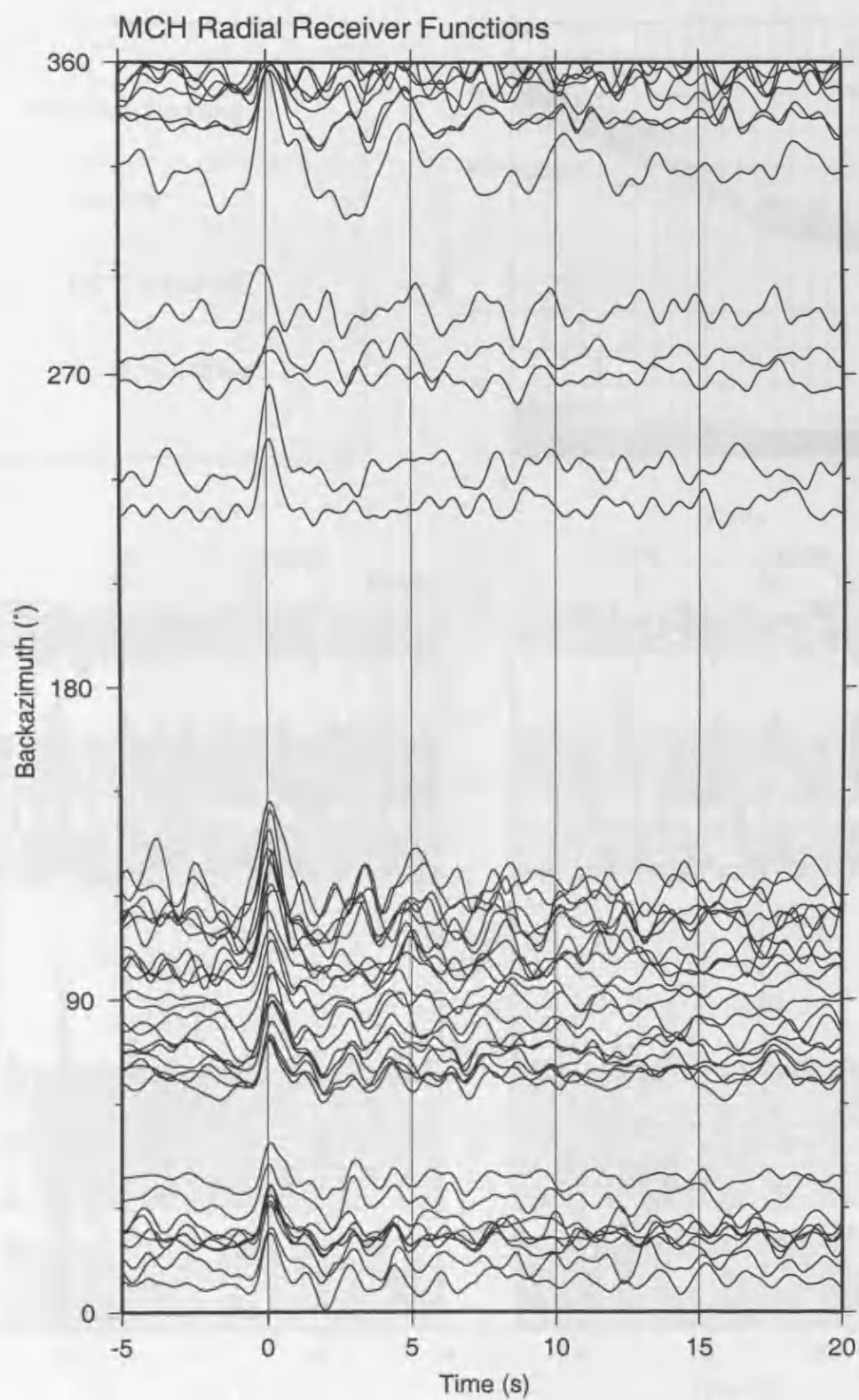


Figure A.48: Raw receiver function data from MCH plotted relative to backazimuth.

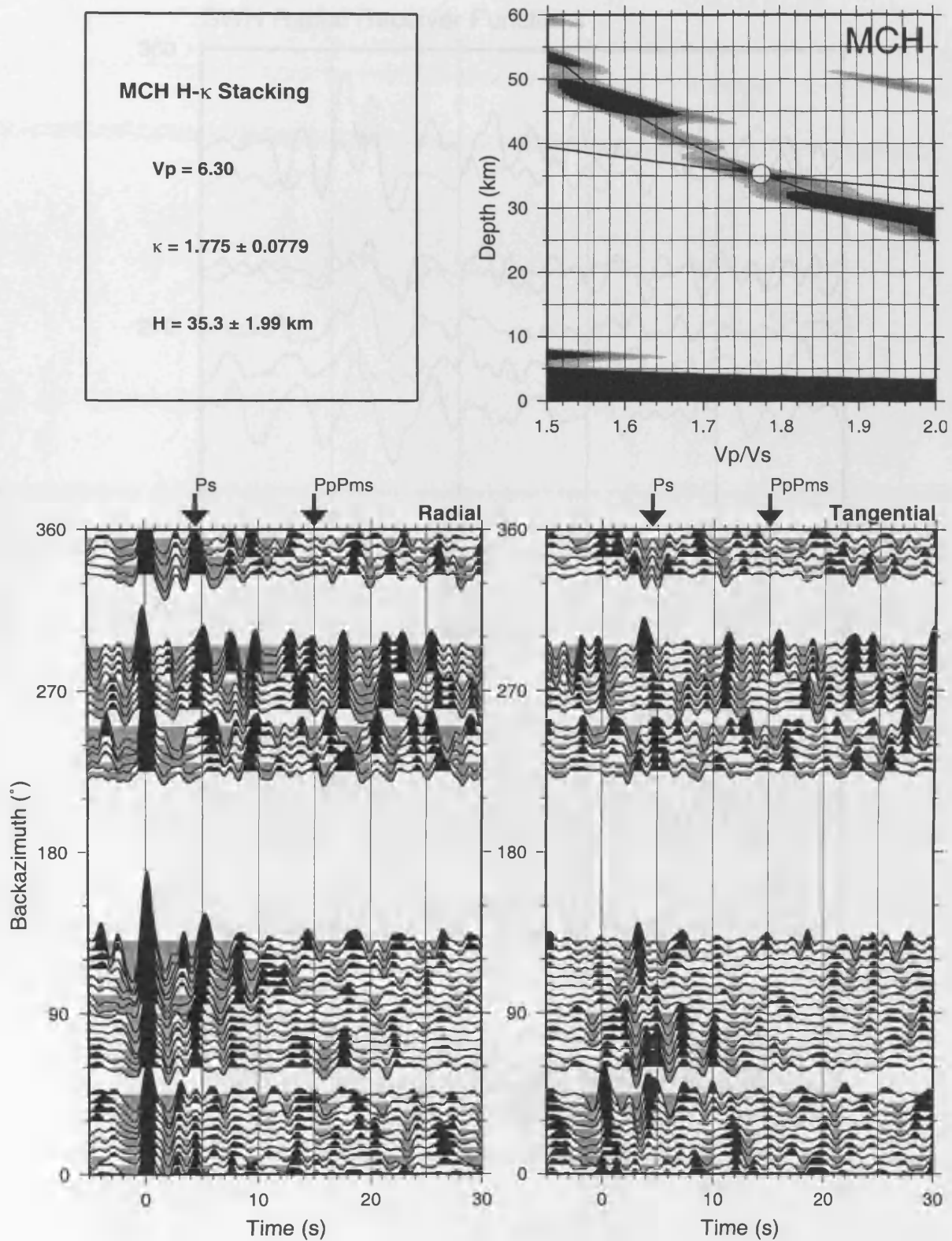


Figure A.49: The radial and tangential receiver functions for MCH, with the results of the H - κ stacking.

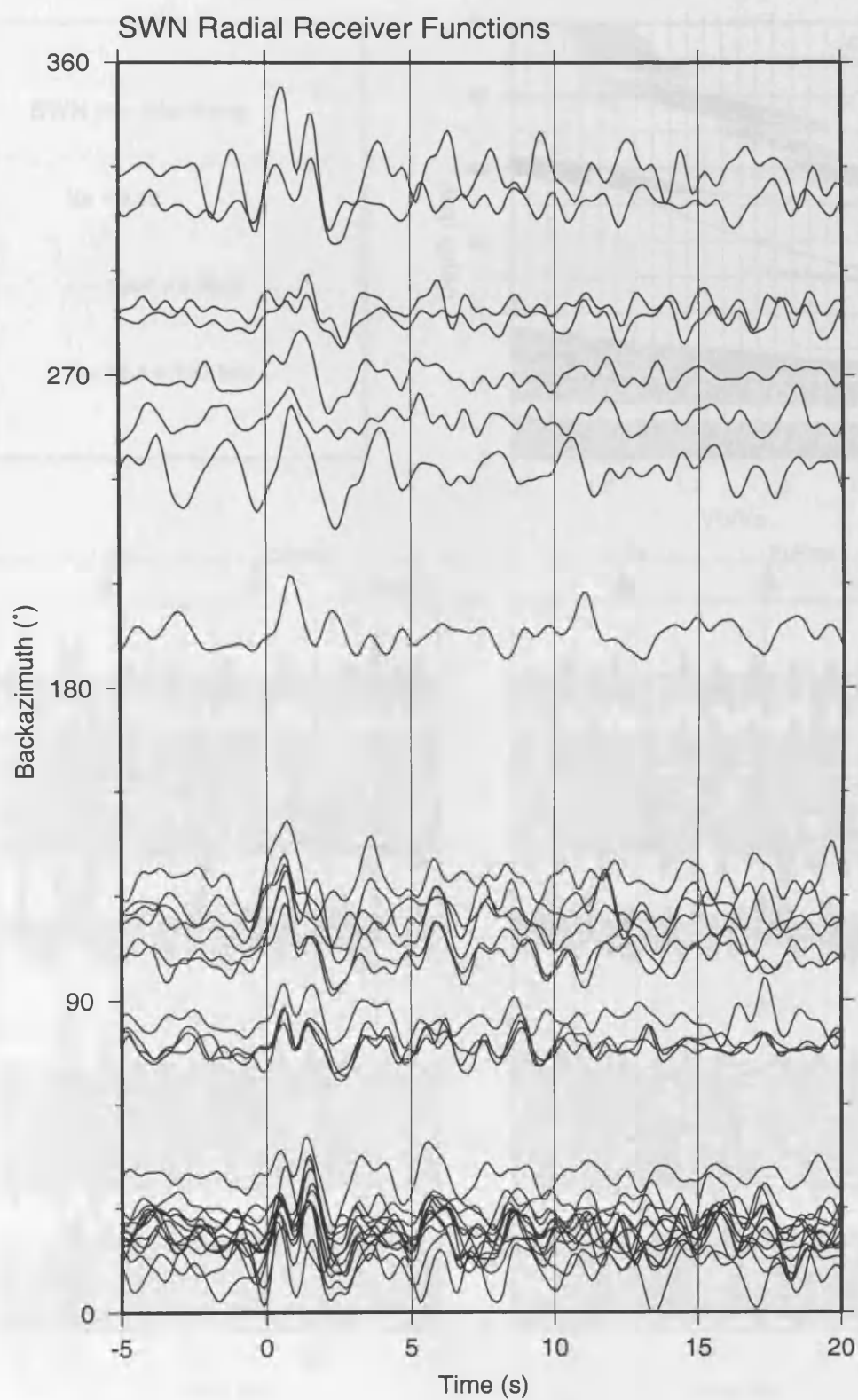


Figure A.50: Raw receiver function data from SWN plotted relative to backazimuth.

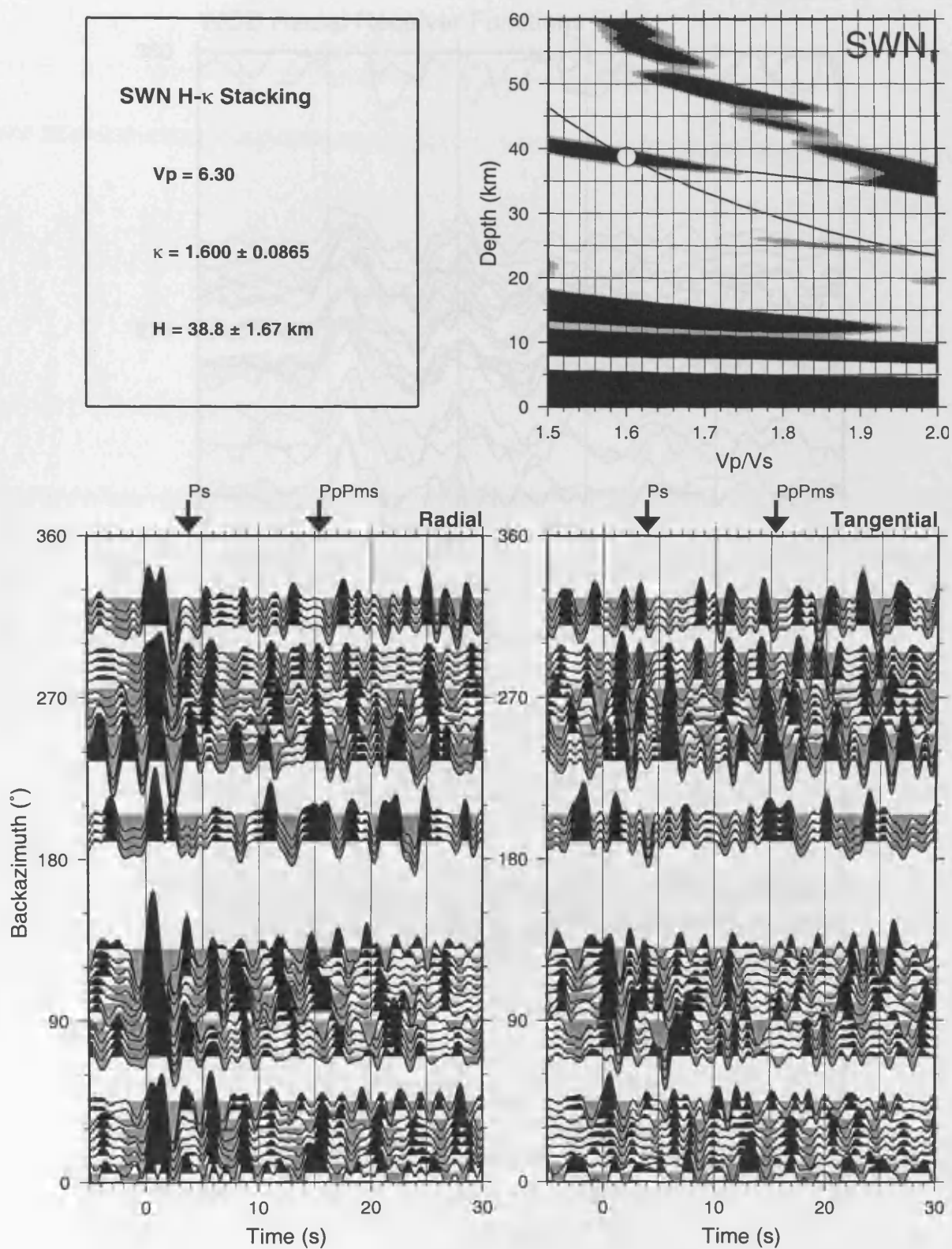


Figure A.51: The radial and tangential receiver functions for SWN, with the results of the H - κ stacking.

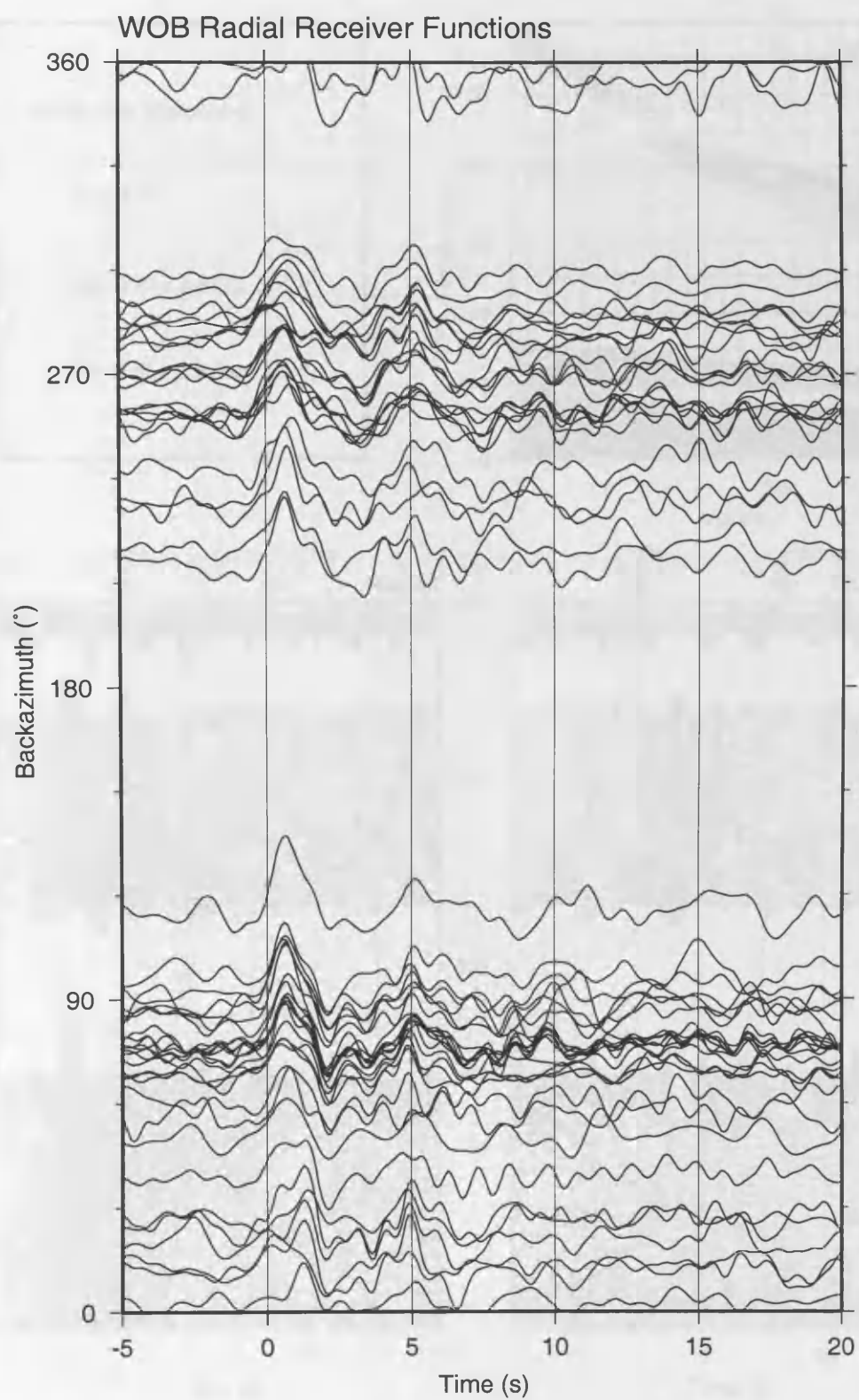


Figure A.52: Raw receiver function data from WOB plotted relative to backazimuth.

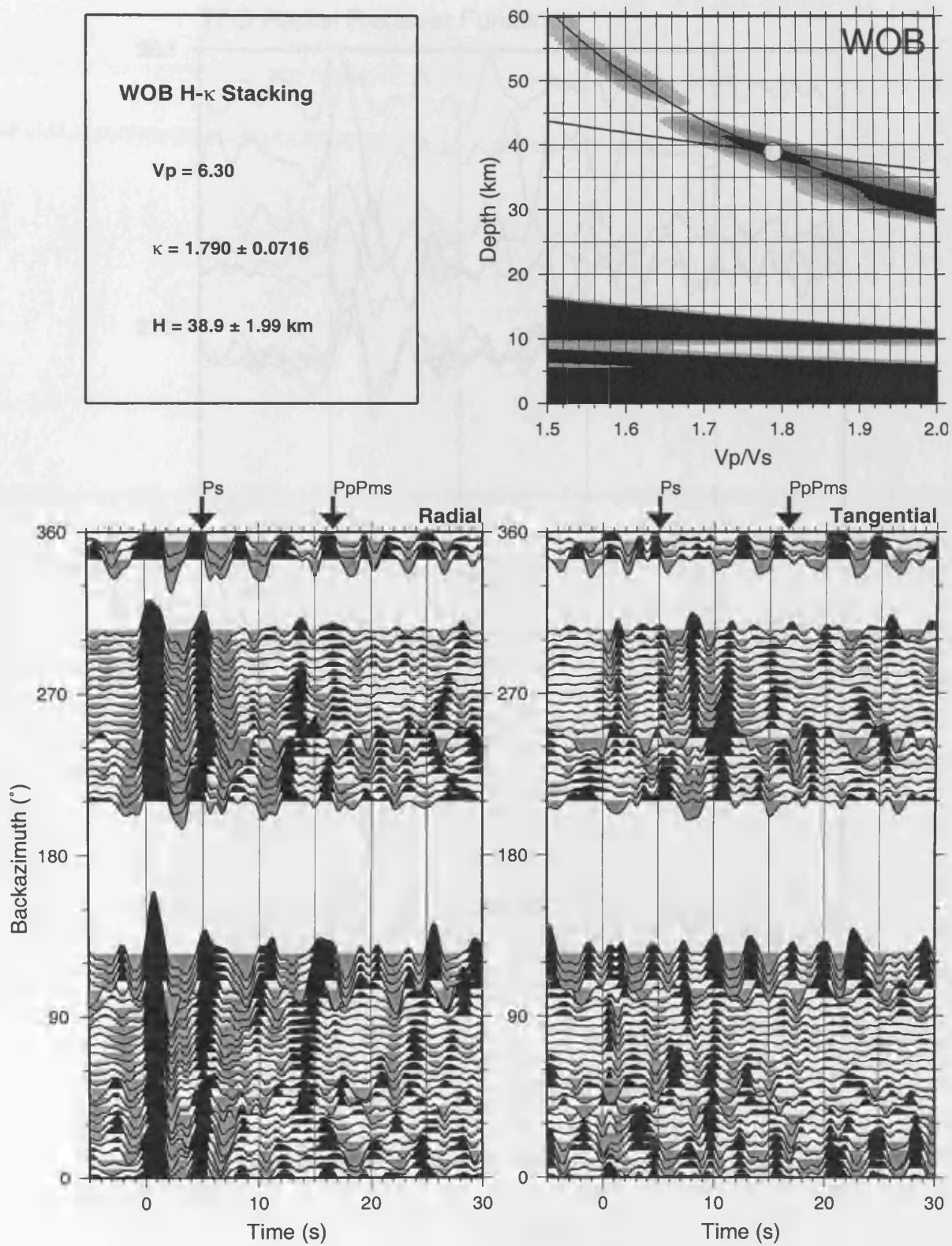


Figure A.53: The radial and tangential receiver functions for WOB, with the results of the H - κ stacking.

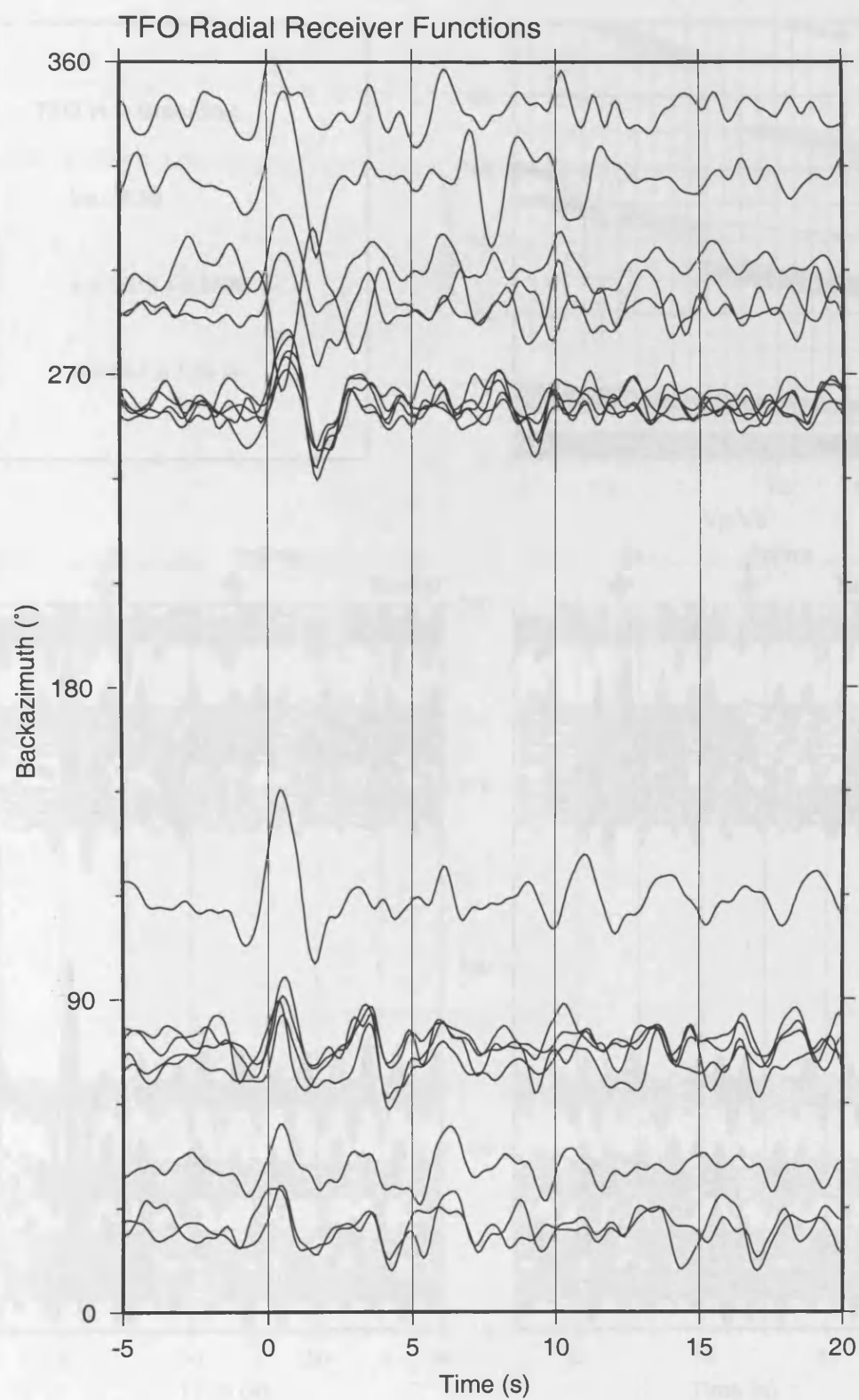


Figure A.54: Raw receiver function data from TFO plotted relative to backazimuth.

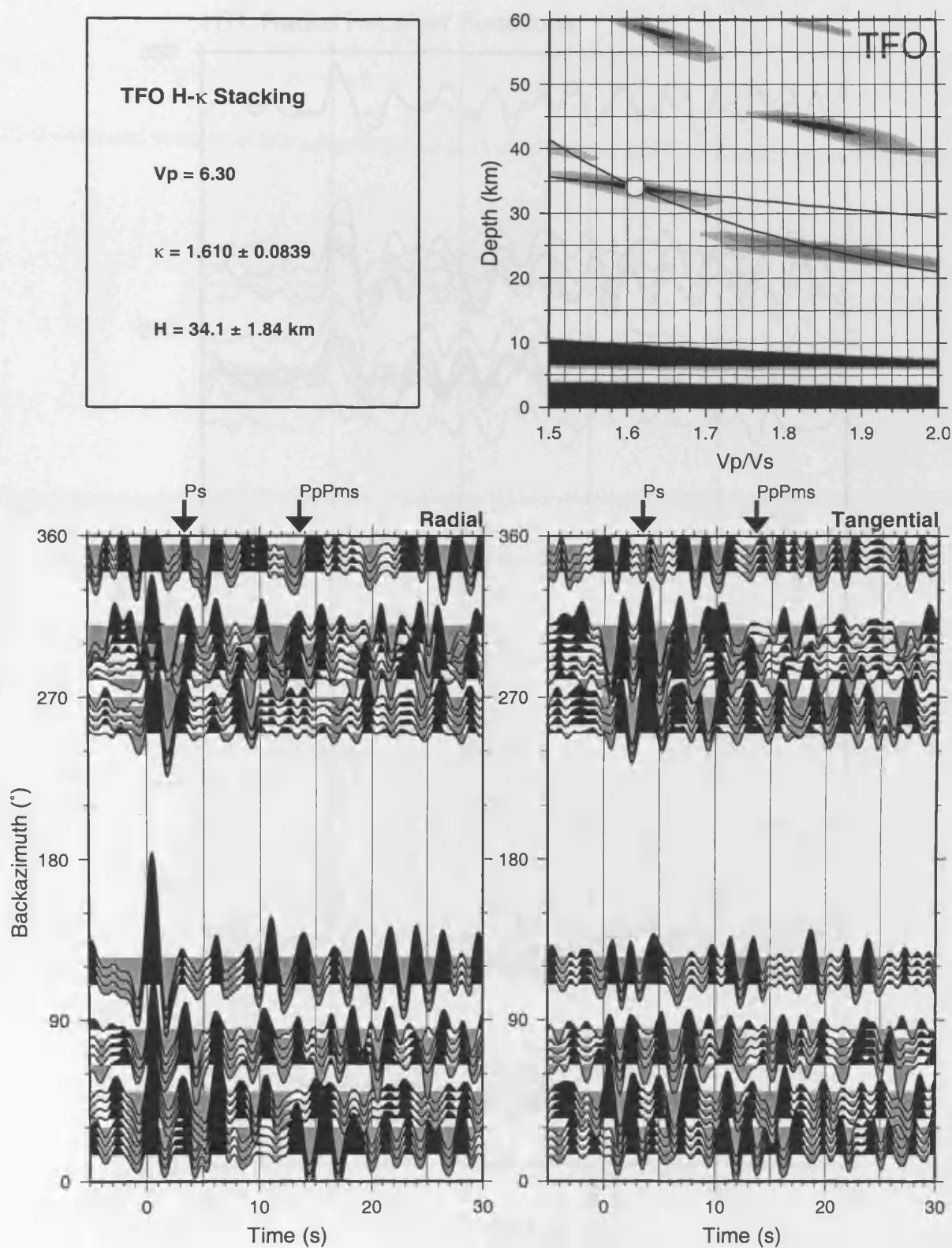


Figure A.55: The radial and tangential receiver functions for TFO, with the results of the H - κ stacking.

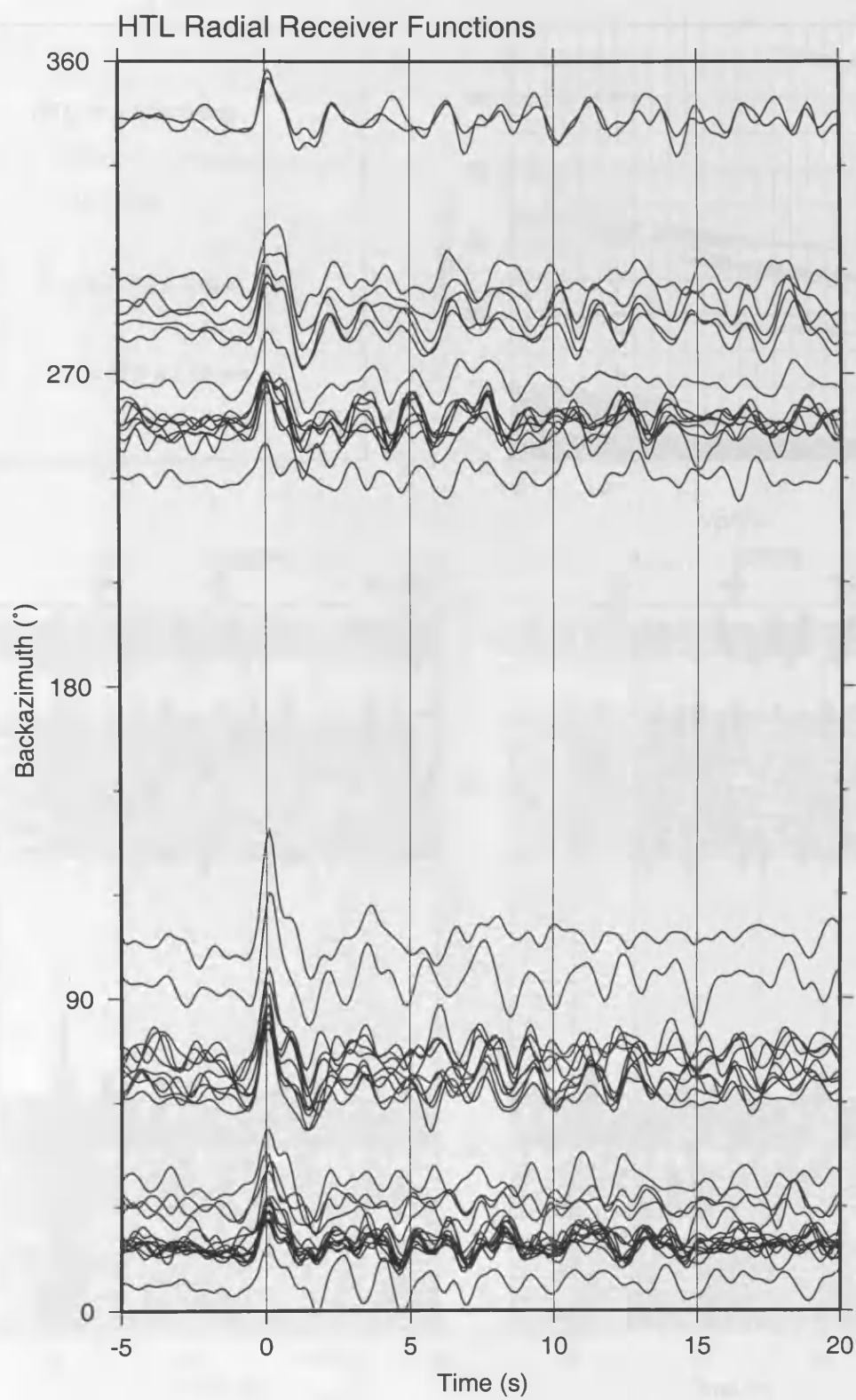


Figure A.56: Raw receiver function data from HTL plotted relative to backazimuth.

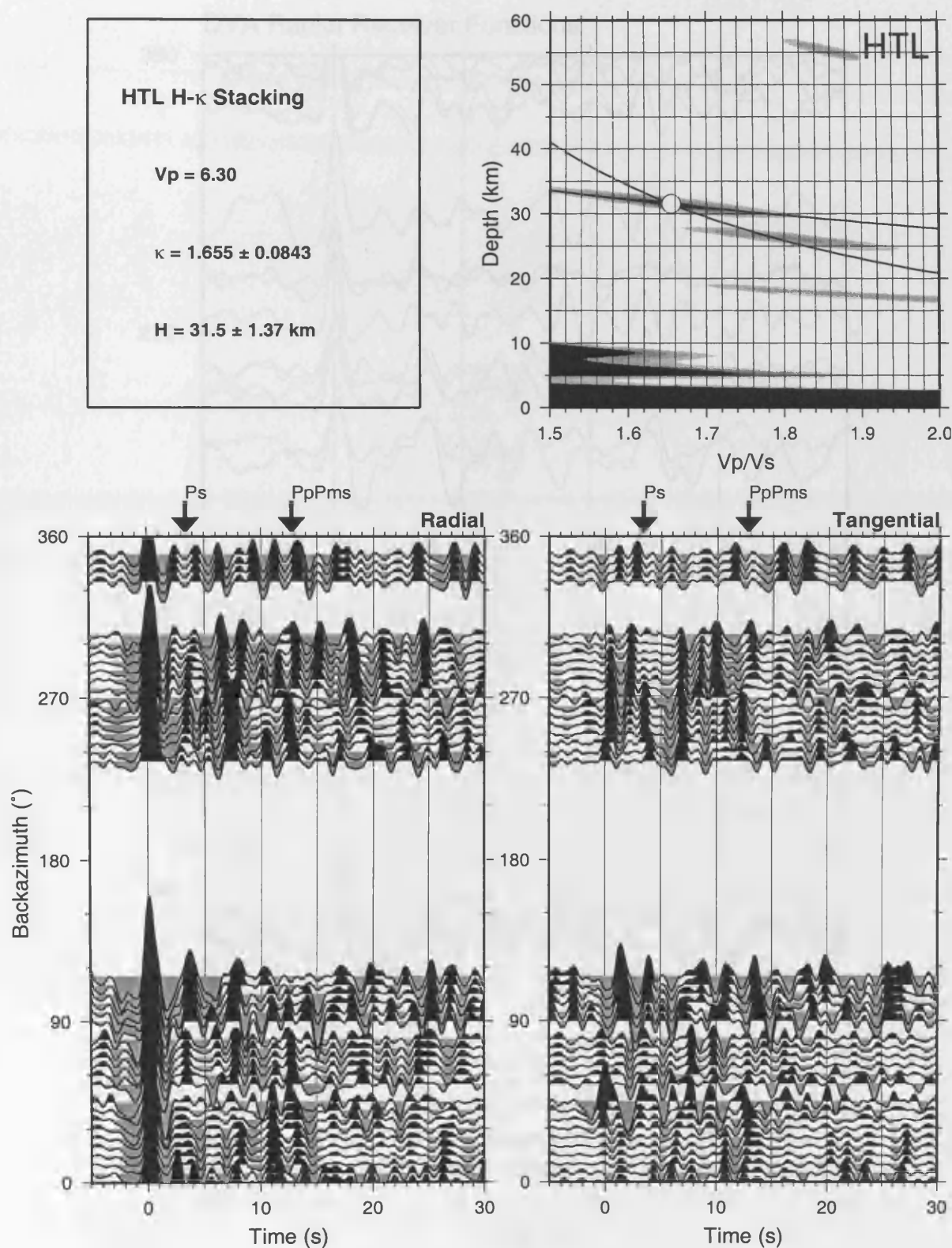


Figure A.57: The radial and tangential receiver functions for HTL, with the results of the H - κ stacking.

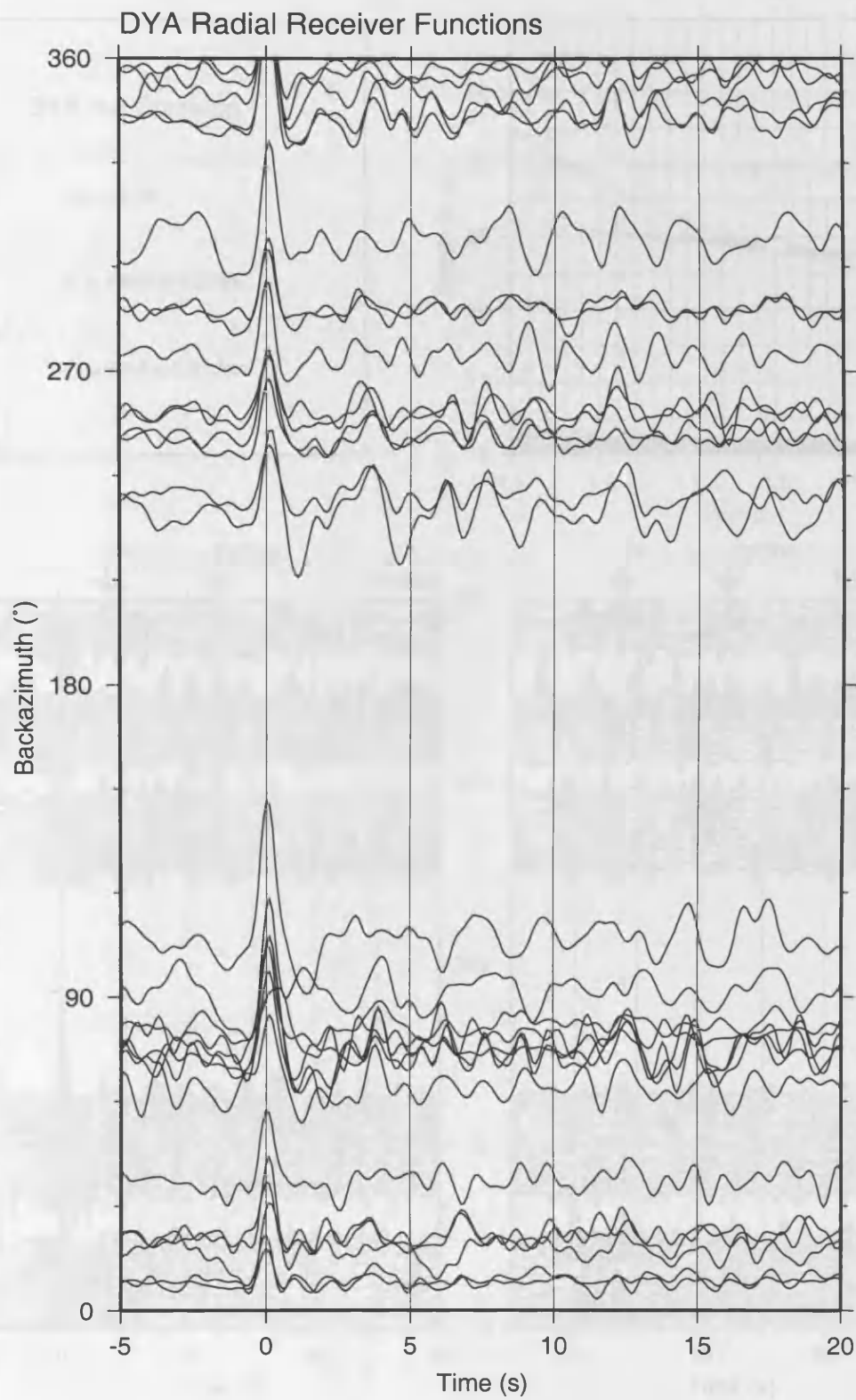


Figure A.58: Raw receiver function data from DYA plotted relative to backazimuth.

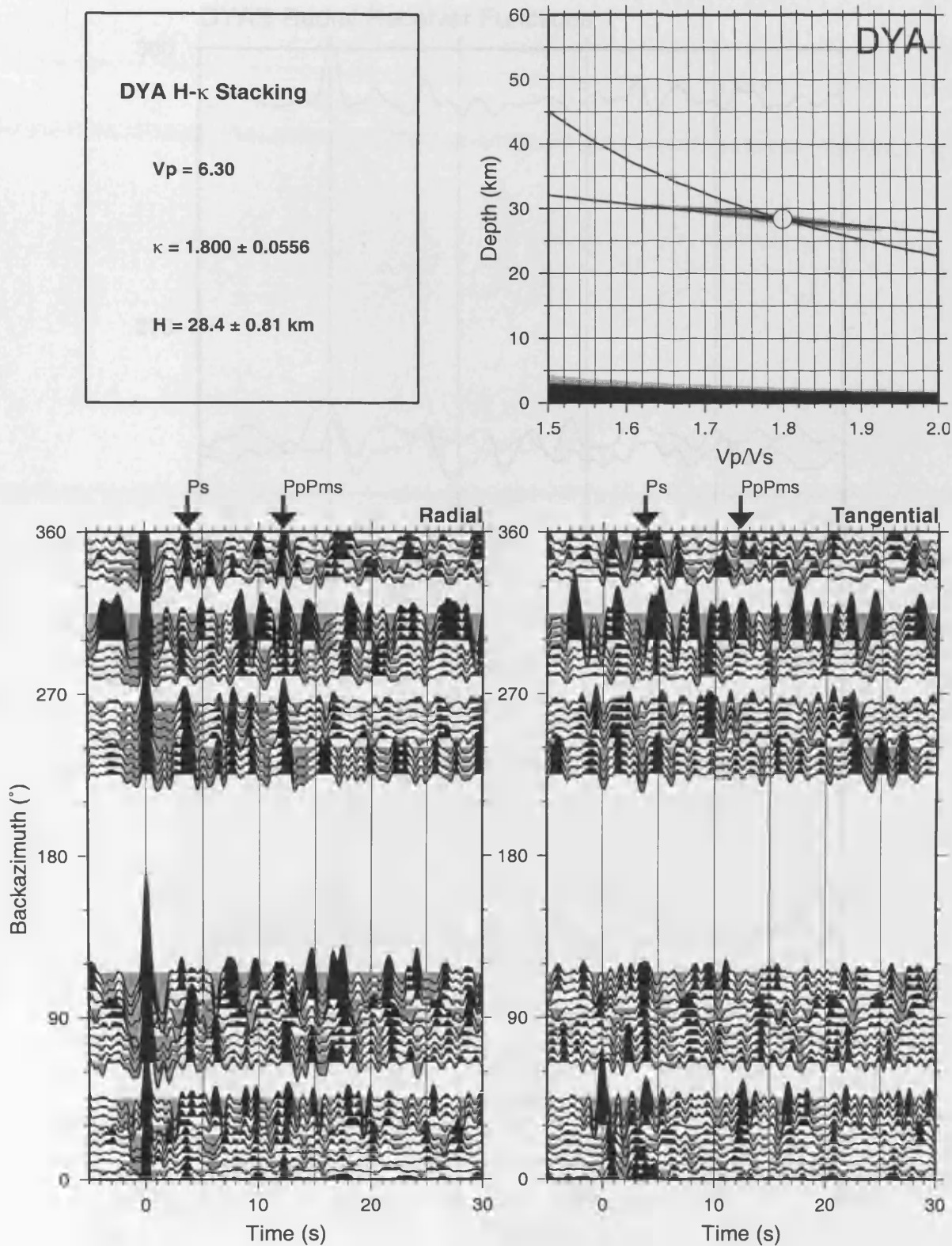


Figure A.59: The radial and tangential receiver functions for DYA, with the results of the H - κ stacking.

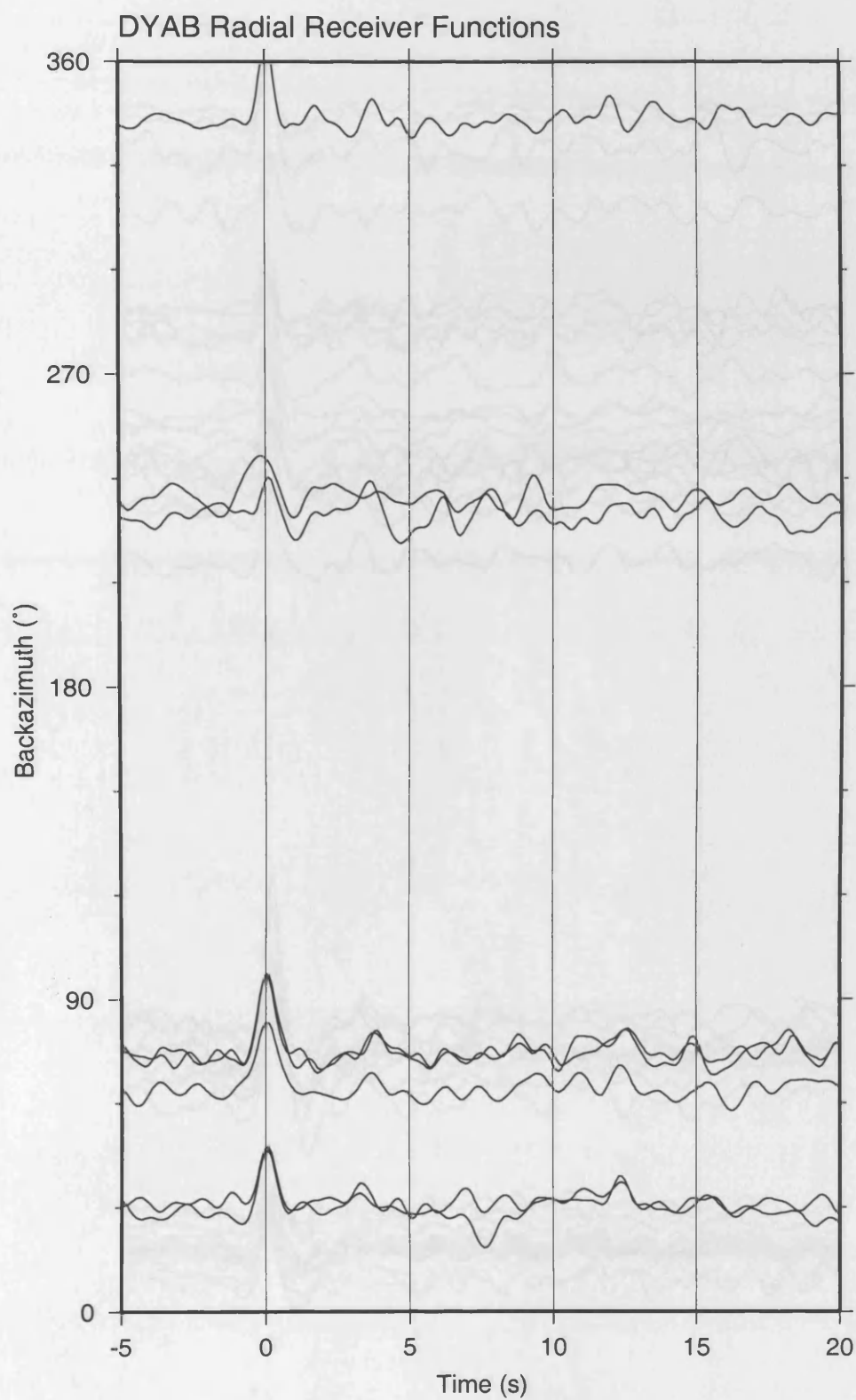


Figure A.60: Raw receiver function data from DYAB plotted relative to backazimuth.

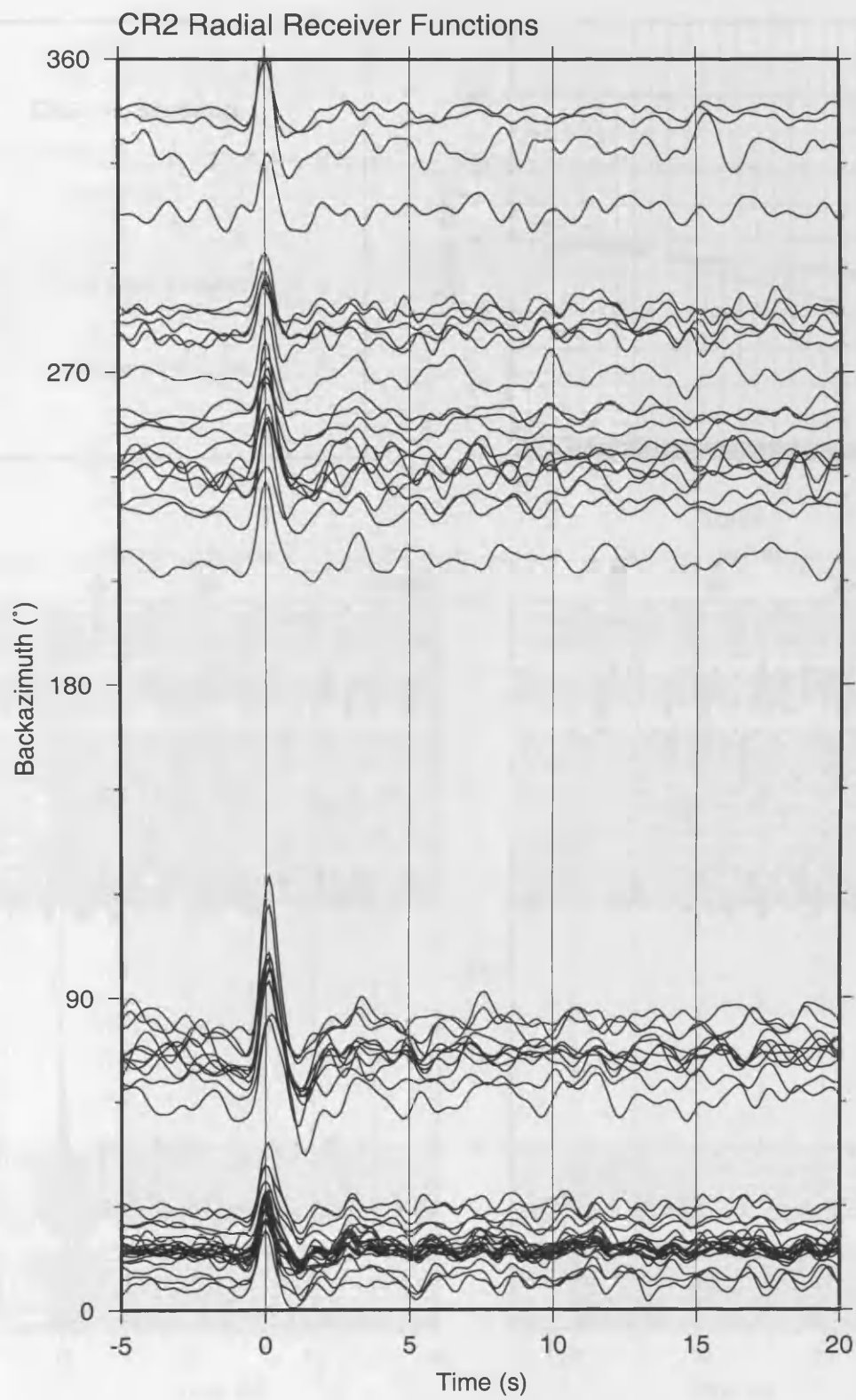


Figure A.61: Raw receiver function data from CR2 plotted relative to backazimuth.

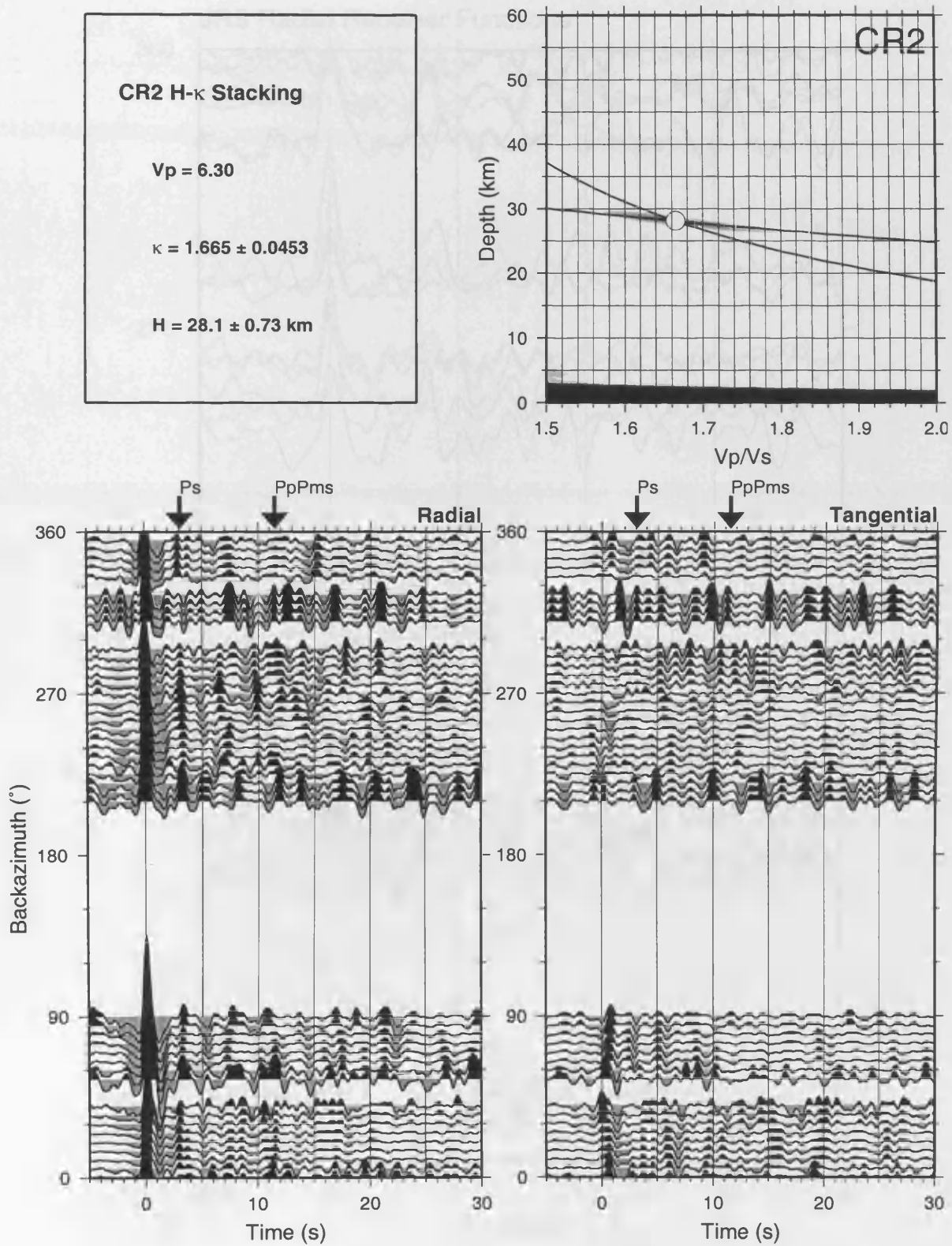


Figure A.62: The radial and tangential receiver functions for CR2, with the results of the H - κ stacking.

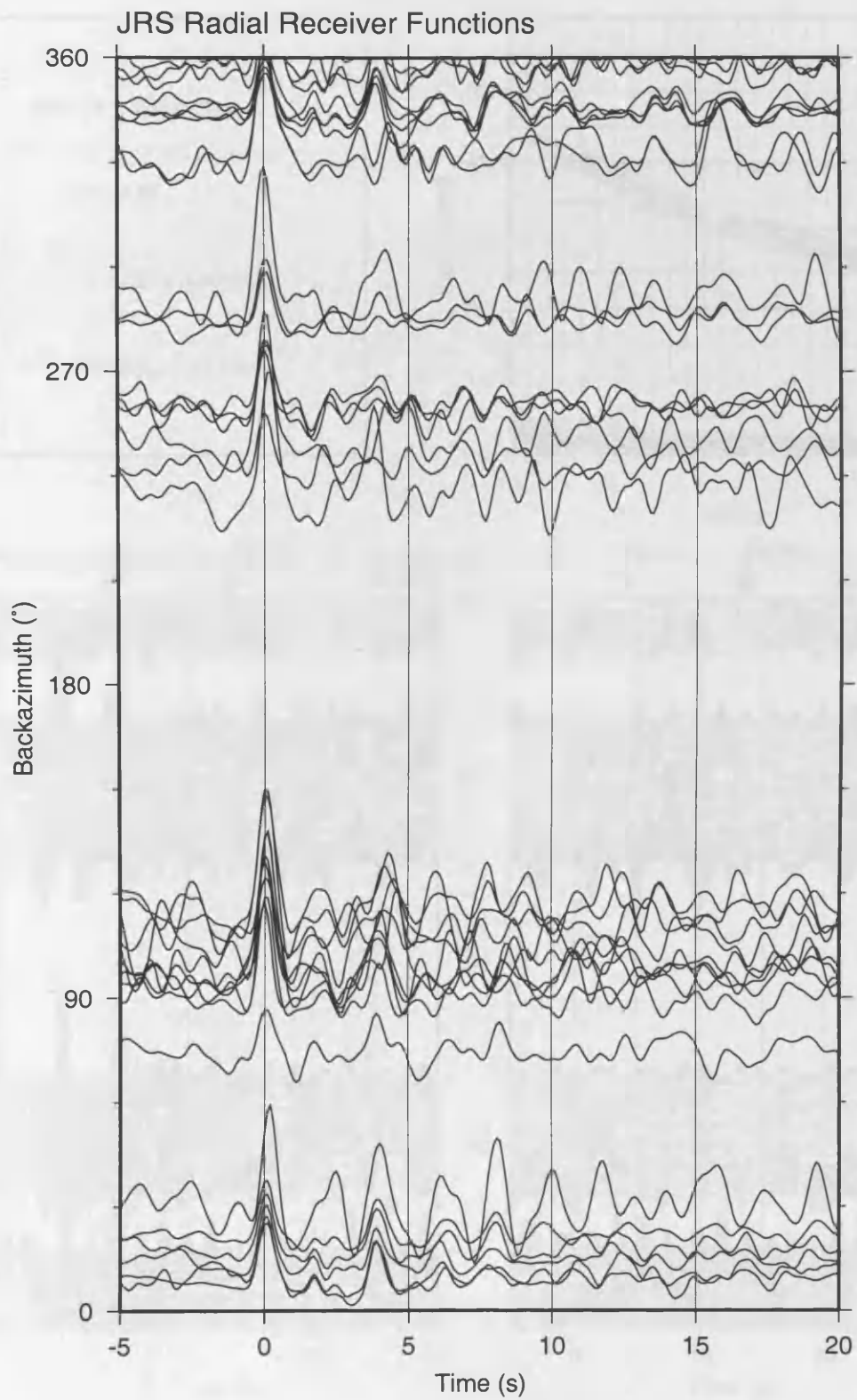


Figure A.63: Raw receiver function data from JRS plotted relative to backazimuth.

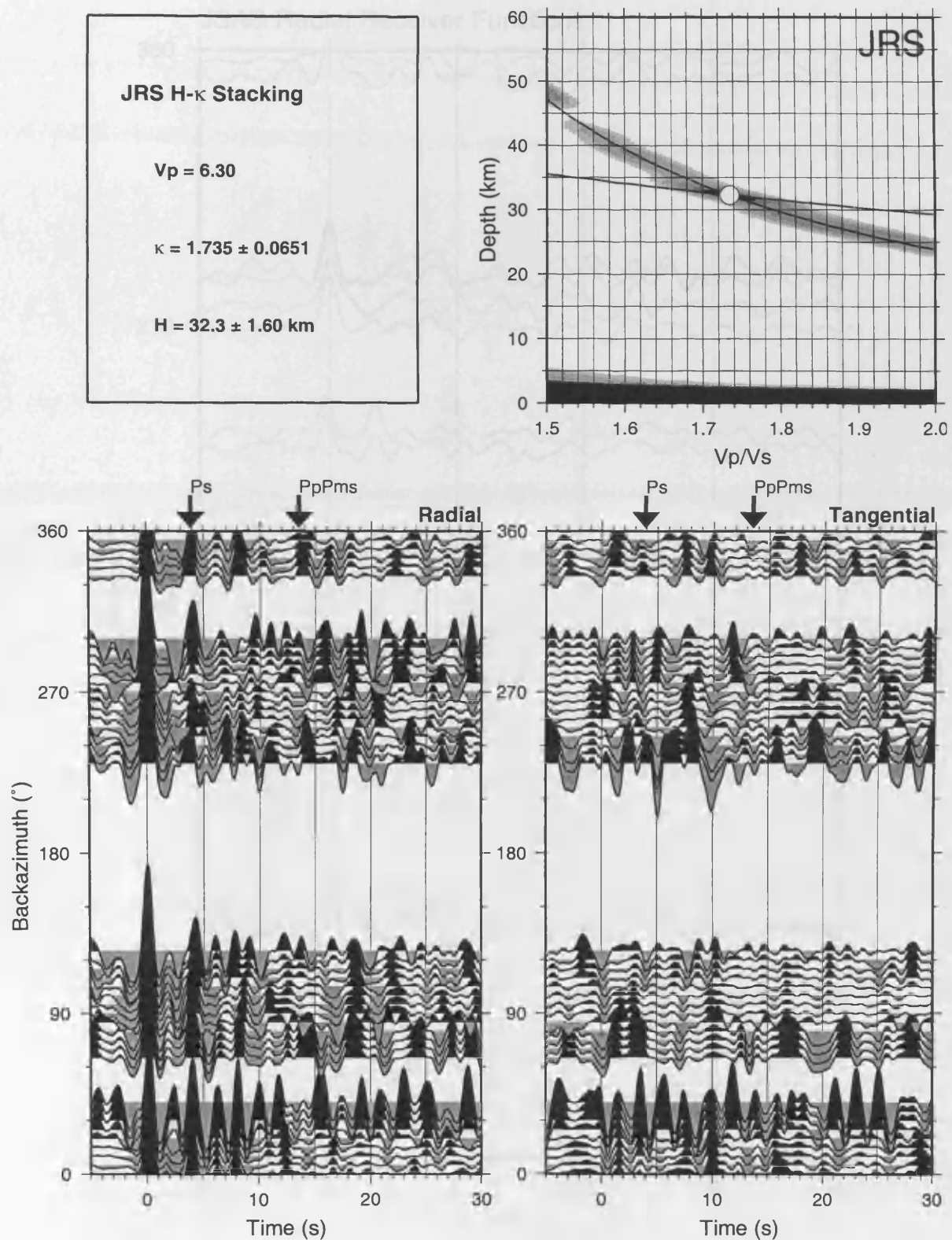


Figure A.64: The radial and tangential receiver functions for JRS, with the results of the H - κ stacking.

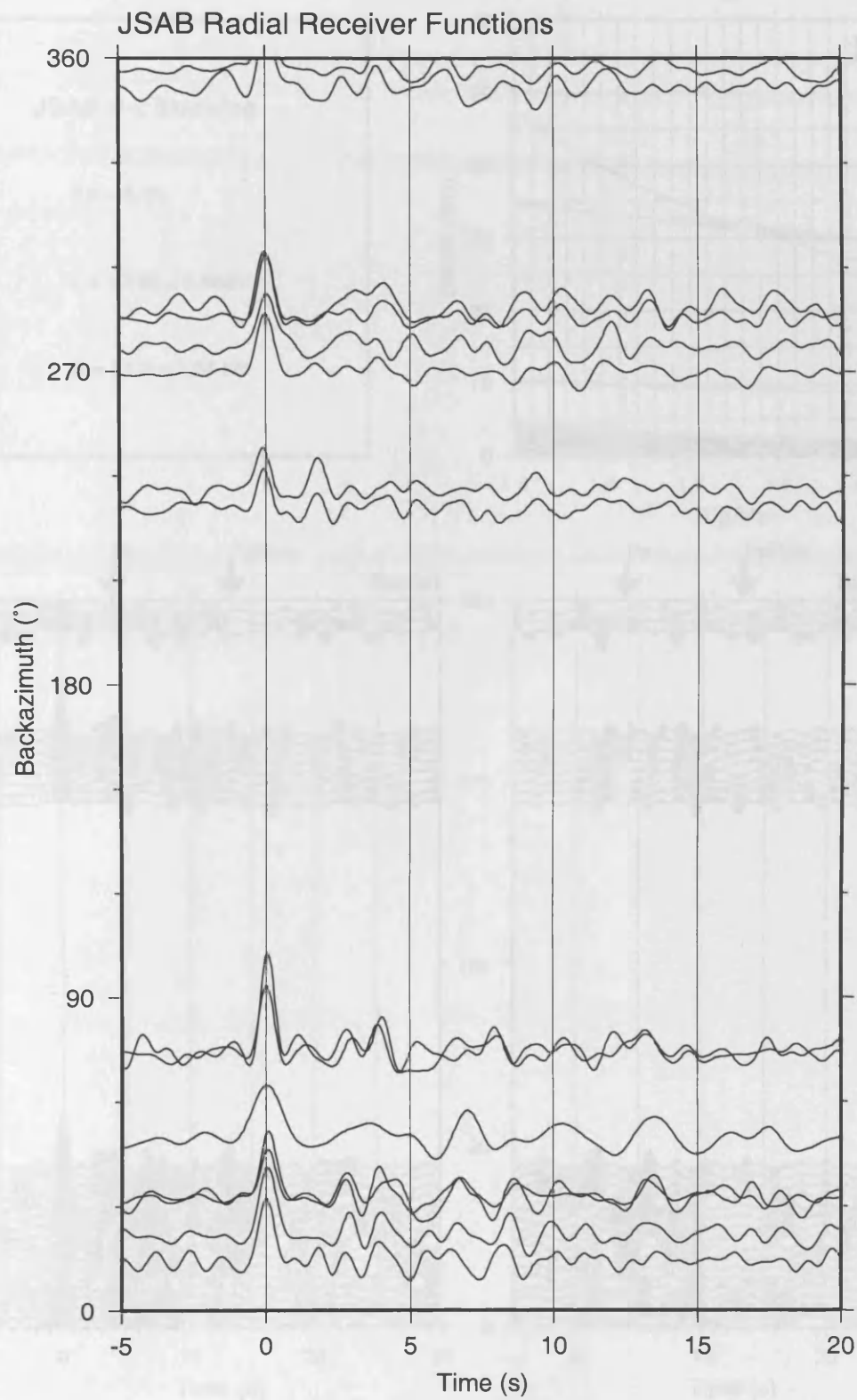


Figure A.65: Raw receiver function data from JSAB plotted relative to backazimuth.

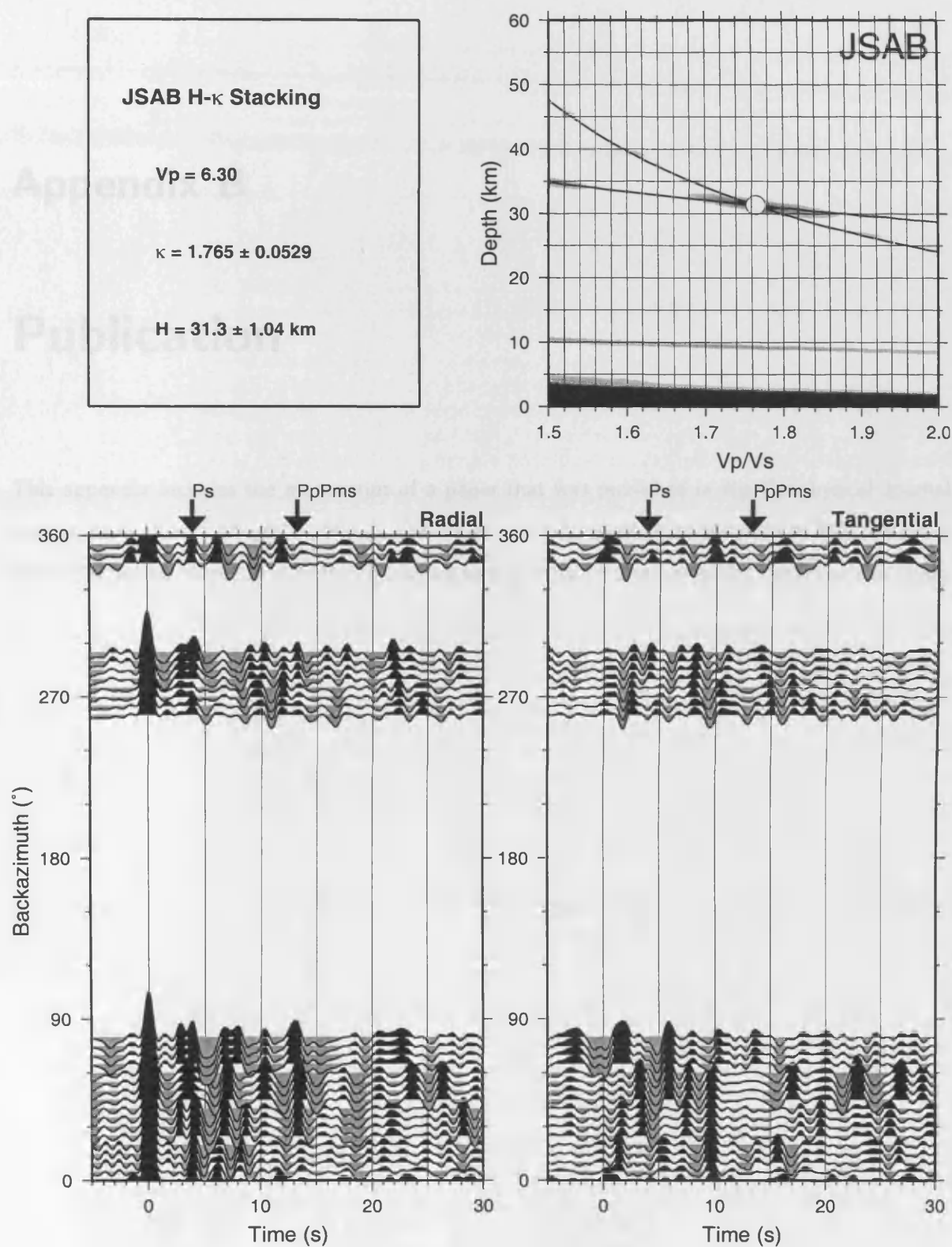


Figure A.66: The radial and tangential receiver functions for JSAB, with the results of the $H-\kappa$ stacking.

Appendix B

Publication

This appendix includes the manuscript of a paper that was published in the *Geophysical Journal International* (Tomlinson et al., 2003). The paper was submitted for publication in June 2002 and was accepted for publication in March 2003 reporting some of the preliminary results of this study.

UK crustal structure close to the Iapetus Suture: a receiver function perspective

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SUMMARY

Teleseismic receiver functions have been calculated for data from 10 short-period three-component seismic recording stations across northern Britain to investigate variations in crustal and upper-mantle velocity structure. The stations straddle the Iapetus Suture zone, the inferred boundary between two of the continents fused together during the Caledonian Orogeny. The receiver function data shows that there is considerable azimuthal variation in both crustal and upper-mantle structure beneath several stations. The data are projected on to a 2-D profile, showing laterally continuous Moho *Ps* conversions at delay times between 3 and 4 s. Synthetic receiver functions, generated using the velocity model from a previous deep seismic reflection/refraction survey show *Ps* and *PpPms* phases comparable to the observed data. 1-D forward modelling of the data gives crustal thicknesses of ~30 km. There is a significant velocity–depth trade-off in the receiver function method, and the crustal thicknesses have been constrained by *a priori* velocity information. Investigation of the data from station GIM close to the Iapetus Suture shows a sub-Moho phase, which is found only on data from northerly backazimuths. Phase modelling is consistent with the presence of a gradational low-velocity zone with a minimum V_p of 6.5 km s⁻¹ at a depth of ~43 km. This feature has similar characteristics to the wide-angle seismic reflection and refraction velocity model of the W-reflector in northern Scotland. The large-scale heterogeneity at GIM is attributed to structures associated with the Iapetus Suture. However, modelling has been performed using 1-D approximations, while the phase signature could result from complex 3-D morphology. We therefore conclude only that the results provide evidence of significant lateral variation in subcrustal structure.

Key words: British Isles, crust, Moho discontinuity, receiver functions.

1 INTRODUCTION

The basement of the British Isles is composed of a number of geological terranes that have been brought together through a series of tectonic events culminating in the Caledonian Orogeny, some 470 Ma (Armstrong & Owen 2001) (Fig. 1). The offshore deep geology of the British Isles has been extensively investigated using deep normal incidence seismic surveys by the British Institutions Reflection Profiling Syndicate (BIRPS) (see, for example, Brewer *et al.* 1983; Hall *et al.* 1984; Freeman *et al.* 1988). However, our knowledge of the deep structure of on-shore Britain is less detailed. The Lithospheric Seismic Profile in Britain (LISPB), a wide angle reflection/refraction profile undertaken in 1974 provides a key deep crustal data set (Bamford *et al.* 1976).

There are 25 permanent three-component short-period seismic recording stations throughout the UK that are suitable for receiver function analysis. These stations form part of the British Geological Survey (BGS) seismological monitoring network. Digital teleseismic event data from this network extend over more than 10 years.

Receiver function analysis is a powerful tool in the examination of crustal and mantle structures (see, for example, Gossler *et al.* 1999; Knapmeyer & Harjes 2000; Zhu & Kanamori 2000). 1-D receiver function analysis of the teleseismic data from these stations are now being used to produce a series of Moho spot-depths over the British Isles, augmenting the Moho depth data sets obtained by conventional reflection and refraction studies. Furthermore, careful analysis of the azimuthal variation in receiver function for each station will allow more information to be gained concerning local variability in crustal structure beneath each individual receiver location.

2 GEOLOGICAL SETTING

The dominant structural trend throughout the northern British Isles is northeast–southwest. This trend was developed during the closure of the Iapetus Ocean and the Caledonian Orogeny, an oblique collision between the two continents bounding the Iapetus Ocean, Laurentia and Eastern Avalonia (a microcontinent on the margin of Baltica). The suture between Laurentia and Eastern Avalonia crosses

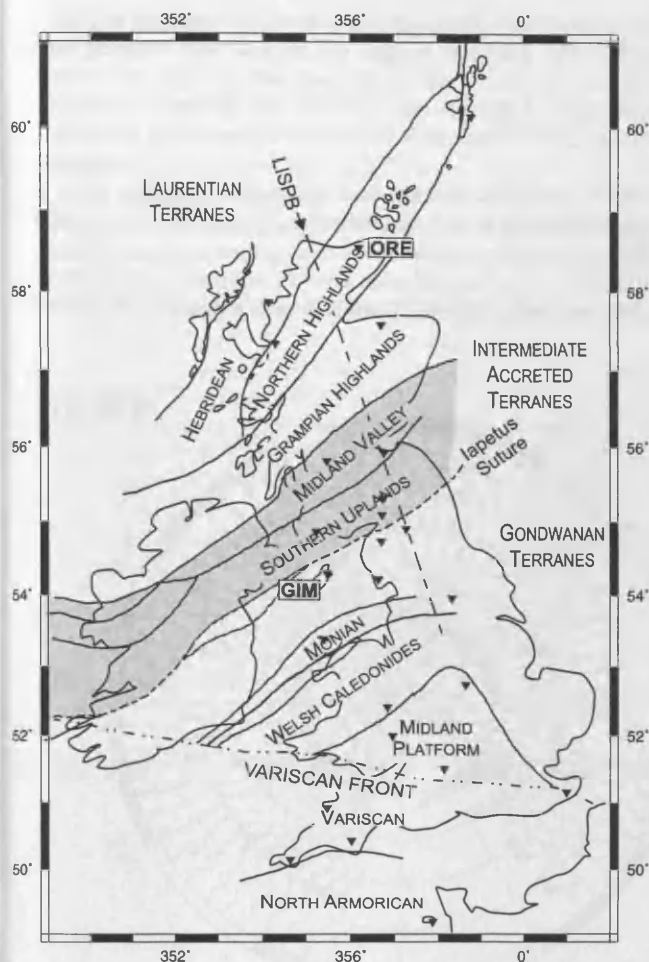


Figure 1. A geological terrane map of the British Isles, after British Geological Survey (1996). The BGS three-component short-period instruments are marked by inverted triangles, with stations GIM and ORE labelled. The northern part of the LISP seismic refraction profile is annotated.

northern England and continues through Ireland (Fig. 1). This has been imaged as a dipping crustal discontinuity both east and west of the British Isles by the deep seismic reflection NEC (Freeman *et al.* 1988) and WINCH (Brewer *et al.* 1983; Hall *et al.* 1984) profiles, respectively. The Iapetus Suture is less well mapped onshore, with the LISP profile showing little change in seismic velocity structure over the inferred boundary (Barton 1992). It is also important to note that the structure mapped by the LISP profile is assumed to be 2-D. The Southern Uplands Seismic Profile (SUSP) (Hall *et al.* 1983), which was orientated normal to the LISP line, shows velocities that are significantly different from those for LISP over the same area, demonstrating marked azimuthal variability in the crustal structure. This difference has been interpreted in terms of broad scale anisotropy (Barton 1992).

The surface geology over the British Isles is highly variable, ranging from Caledonian basement through thick Carboniferous basins to Quaternary drift. The near-surface geology can have a significant effect upon the ease of interpretation of the resultant receiver functions (Zelt & Ellis 1999); for example, the presence of a 2 km thick sedimentary basin beneath a station could result in high-amplitude multiples overprinting the signature of deeper structures. Moho depths derived from seismic reflection profiles have been mapped across the British Isles (Chadwick & Pharaoh 1998); typi-

cal values range between 25–35 km. Sub-Moho discontinuities have been identified beneath northern Scotland on both the LISP and the BIRPS profiles. The BIRPS data identifies the Flannan and the W-reflectors offshore of the northwest Highlands, and the LISP profile shows a reflector at 40 km depth beneath northern Scotland (Barton 1992). The Flannan reflector has been imaged as a discontinuity dipping to the east at 20°–30°, located to the west of the northwest Highlands; the W-reflector is flat lying at 50 km depth and on some profiles intersects the Flannan reflector (McBride *et al.* 1995). One of the aims of the present study is to investigate the continuity and extent of these mantle structures beneath northern Britain.

3 DATA AND METHOD

The BGS has an extensive seismic monitoring network incorporating 25 three-component short-period instruments, of which the most densely covered area bridges the Iapetus Suture Zone in northern England and southern Scotland (Fig. 1). Receiver functions have been calculated for 10 three-component stations across the Iapetus Suture from the BGS network, using over 700 separate events $\geq 6.0 m_b$, occurring in the period 1990–2001. As an initial step the horizontal component seismograms were rotated into the radial and tangential orientations. The receiver functions were then calculated using the source equalization technique of Langston (1979), in which the vertical component $[D_V(\omega)]$ of the seismogram is deconvolved from the radial $[D_R(\omega)]$ and tangential components $[D_T(\omega)]$ in the frequency domain to produce radial (RRF) and tangential (TRF) receiver functions, respectively. The radial receiver function can be written as

$$\text{RRF}(\omega) = \frac{D_R(\omega)D_V^*(\omega)}{\Phi_{ss}(\omega)}G(\omega), \quad (1)$$

where

$$\Phi_{ss}(\omega) = \max\{D_V(\omega)D_V^*(\omega), c \max[D_V(\omega)D_V^*(\omega)]\} \quad (2)$$

and

$$G(\omega) = e^{-\omega^2/4a^2}. \quad (3)$$

$\Phi_{ss}(\omega)$ is defined so as to fill in spectral troughs in the denominator, minimizing problems caused by dividing by small numbers. The water level parameter c defines the level to which the spectral troughs are filled as a percentage of the maximum amplitude of the denominator function $[D_V(\omega)D_V^*(\omega)]$. $G(\omega)$ is a Gaussian filter, limiting the bandwidth of the resulting receiver function, with the parameter a controlling the frequency limits of the Gaussian function. The true amplitudes of the receiver functions are maintained by ensuring that the amplitude of the deconvolution of the vertical component of the seismogram from itself is equal to one; this is achieved by dividing the horizontal receiver functions by the maximum amplitude of the vertical deconvolution (Ammon 1991). If a 1-D approximation holds then the Psv conversions should be contained in the radial receiver functions. The entire data set has been processed using a water level parameter of $c = 0.001$ and a Gaussian limit of $a = 2$. These values give consistent, stable receiver functions for all of the stations in this study.

4 AZIMUTHAL VARIATION OF RECEIVER FUNCTIONS

Given the complex geological history of the British Isles, the 3-D variability of the crust beneath each station must be considered; this can be done by examining the azimuthal variability in radial

receiver function. To do this the data have been stacked in 20° backazimuth bins, and the bin window has been rotated through 360° at 5° intervals. The majority of bins contain several receiver functions. Applying this technique the receiver functions may be smoothed over a maximum of $\pm 10^\circ$ from their true backazimuthal position.

The receiver function data have not been corrected for the variation due to changes in ray parameter. This is standard practice in studies aimed at imaging deep mantle structures such as the 440 and 670 km discontinuities (for example, Kosarev *et al.* 1999; Li *et al.* 2000), but the corrections that would be applied to *Ps* conversions

at depths of 25–35 km would produce negligible improvement in the stacked data quality. Furthermore, the method corrects all the data as though they were *P*-to-*S* conversions through a given velocity model, however, multiple phases have move-out in the opposite sense to the *Ps* phase. Therefore, applying such pre-stack corrections to data from a range of epicentral distances will result in a degradation of the signal from multiple phases.

The resulting stacked receiver functions have been plotted radially relative to backazimuth (baz). The plotting of receiver function data in this manner facilitates the examination of the 3-D nature of the crust. For example, Fig. 2 shows the variation in receiver function

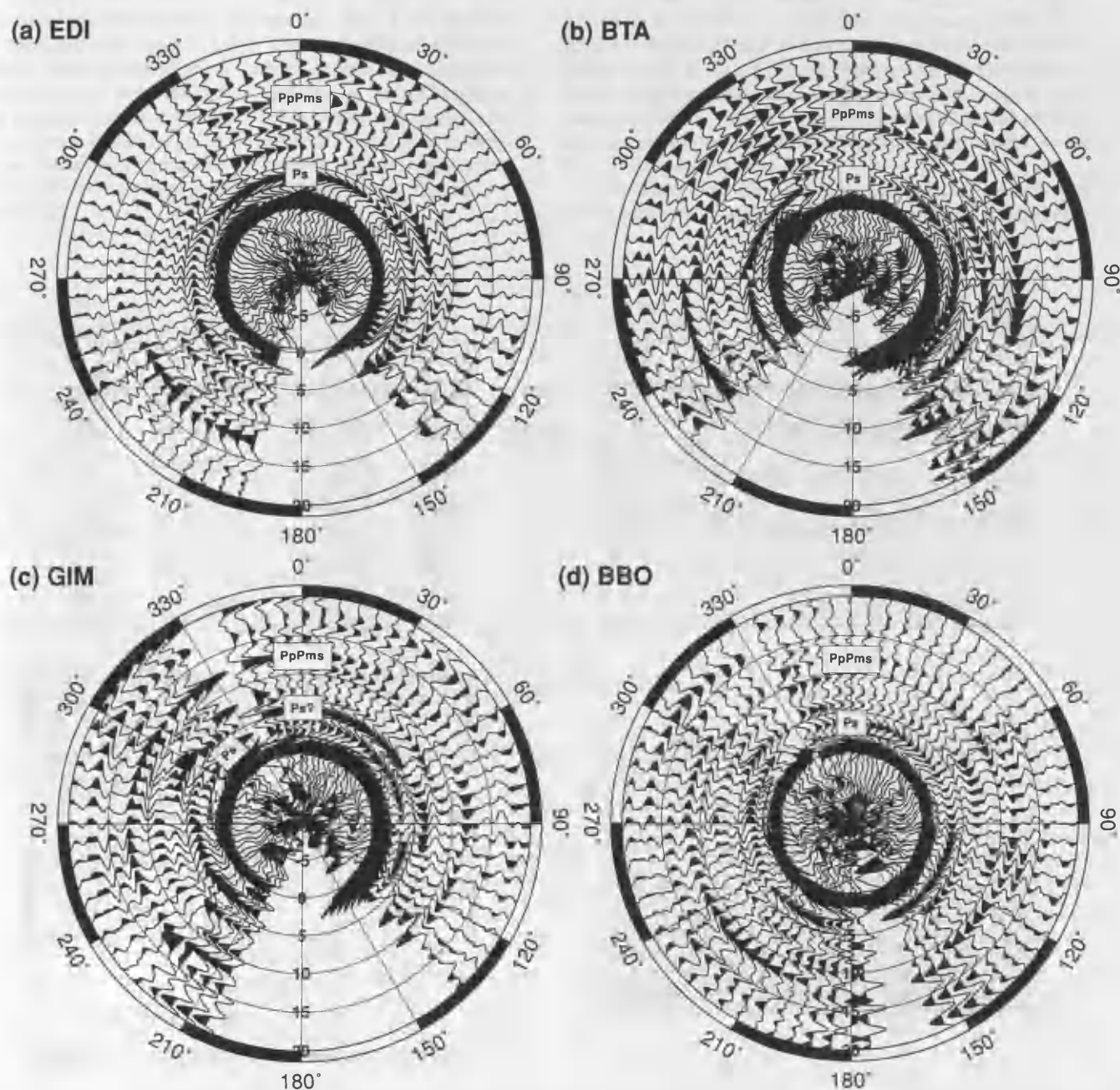


Figure 2. (a) Stacked EDI radial receiver function (RRF) bins plotted relative to backazimuth, the positive loops of the receiver functions are shaded. Delay time is plotted from -10 to 20 s from the centre of the plot outward. (b) RRFs for BTA, showing a consistent Moho *Ps* phase at 4 s. (c) GIM RRFs again showing consistent Moho *Ps* phase along with an additional high-amplitude phase at 5.5 s. (d) BBO RRFs.

for four separate stations. For EDI (Edinburgh) (Fig. 2a), situated upon the Carboniferous sedimentary basin of the Midland Valley of Scotland, the receiver functions show a strong Moho P_s phase at a delay time of around 3.8 s (the time between the direct P arrival and the P_s phase). There is also at least one earlier phase representing a P -to- S conversion from a mid-crustal layer. The arrival time of this phase varies between 1 and 2 s, demonstrating considerable variation in structure with azimuth. Stations BTA, BBO and GIM lie along strike from one another, close to the Iapetus Suture (Fig. 3c). Examination of the westerly arrivals (baz 240° – 300°) at stations BTA and GIM shows strong similarity (Figs 2b and c), with a Moho P_s phase at between 3 and 4 s and a $PpPms$ multiple phase at around 12 s. However, comparison of the northerly data (baz 330° – 030°) shows a marked difference between the two stations. BTA maintains the same phases at 4 and 11 s, if of slightly smaller amplitude. In contrast GIM still shows the 4 s P_s phase, but also includes an additional phase at around 5.5 s. Furthermore, the $PpPms$ phase can be seen to be more complex, with a double arrival at 11 and 13 s between backazimuths of 0° and 030° . The data from BBO shows similar phase timing to that from BTA, with the westerly data showing the most prominent P_s and $PpPms$ phases at 3.5 and 12 s, respectively.

5 LATERAL VARIATION OF RECEIVER FUNCTIONS

The receiver function data from the stations close to the Iapetus Suture have been compared with the LISP profile. Given the event backazimuth and epicentral distance, the point at which each receiver function raypath pierces the Moho was calculated for the IASP91 velocity model (Kennett & Engdahl 1991). A profile was constructed by projecting each receiver function from its crustal pierce point, perpendicularly on to the chosen line of the LISP profile (Fig. 3). The final section was produced by stacking the receiver functions into 20 km wide bins at 5 km intervals along the profile. A synthetic receiver function profile has been generated along the LISP profile, using the velocity model from Barton (1992). This was achieved by digitizing the LISP velocity model into a series of individual 1-D models. Synthetic seismograms were generated using the reflection matrix technique (Kennett 1983), using a V_p/V_s value of 1.73. The receiver functions were produced by applying identical processing steps to these synthetic seismograms, as had been used on the observed data.

Examination of the two profiles shows that there is a consistently strong Moho P_s conversion across northern Britain at delay times

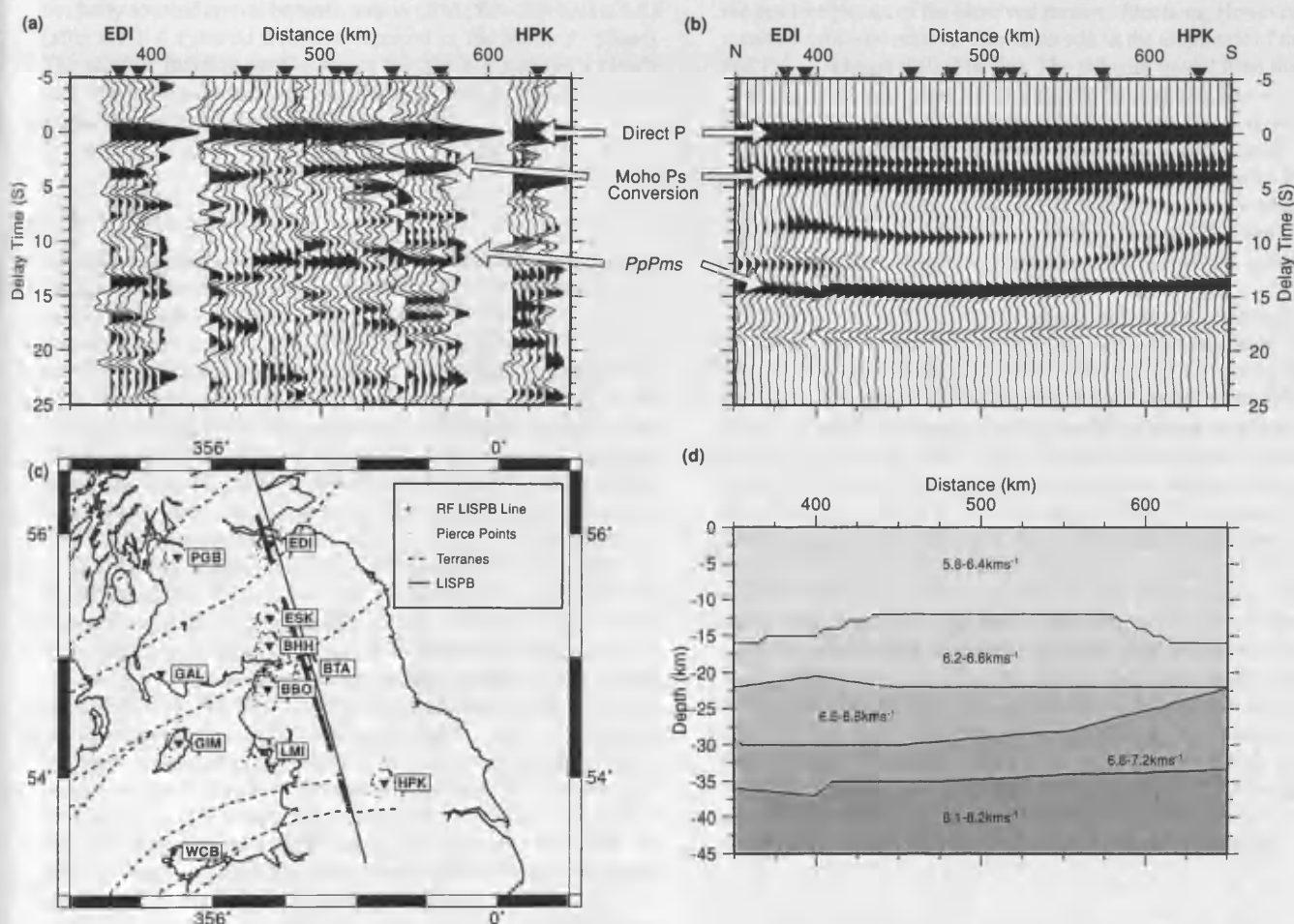


Figure 3. (a) Receiver function data projected on to the LISP profile. The data are binned in 20 km bins and stacked into a 2-D profile. Traces are plotted relative to distance along the LISP profile. (b) The synthetic receiver function profile generated using the LISP velocity model of Barton (1992). (c) Location map of the stations used to generate the receiver function profile, along with the pierce points of all receiver functions. (d) The LISP velocity model (Barton 1992).

of between 2.5 and 4 s. Closer examination reveals that the northern end of the profile is most closely matched by the LISPb synthetics, where both the P_s and $PpPms$ are apparent at 4 and 14 s, respectively. However, the synthetic data has a P_s phase that remains at a constant time across the section, whereas the real data has a P_s phase that arrives earlier towards the south of the section. The northerly observed data are perhaps more consistent with the synthetic data because the stations that have been projected on to the LISPb line are much closer to the profile than those in the south, indicating some change in crustal structure along strike from the LISPb profile. The earlier arrival times of the P_s phase in the south may be caused by either: (1) decreasing crustal thickness; (2) increasing average crustal velocity or (3) decreasing crustal V_p/V_s . In reality, the most likely scenario is that there is some variation in all three parameters over the area covered by the stations. Lateral changes in crustal thickness and average crustal velocity along the 2-D LISPb velocity model are shown in Fig. 3(d) (Barton 1992). The observed receiver function data reveal a change in the mid-crustal structure. A prominent phase is present at ~ 2 s within the northern data (station EDI), the amplitude of this phase gradually reduces, until between 500 and 525 km along the LISPb profile it can barely be resolved. This indicates a change in crustal velocity structure, which could explain the decrease in Moho P_s conversion time southwards. The most noticeable anomalous feature on the section is the high-amplitude northerly sourced arrival beneath station GIM (525–550 km) at 5.5 s (after the 3–4 s arrival that is interpreted as the Moho P_s phase). The receiver function profile shows that the 3–4 s phase is clearly part of a continuous-velocity discontinuity, whereas the 5.5 s phase suggests a localized velocity change.

6 MODELLING

Modelling techniques have been applied to the data to further investigate the unusual structure beneath GIM. The modelling was carried out using simple 1-D forward models to fit the primary converted phases on each of the receiver functions. The observed receiver functions were calculated using short-period seismograms. As a result of the limited bandwidth of the deconvolution of the vertical from the radial and tangential components, negative lobes ringing the primary positive phases have been generated. To reproduce these features in the modelled receiver functions, the synthetic seismograms have been convolved with a short-period instrument response prior to the calculation of the synthetic receiver functions.

The starting models have been constructed using *a priori* information gained from the LISPb models (Barton 1992) and the Caledonian Suture Seismic Profile (CSSP) (Bott *et al.* 1985), which runs parallel to the Iapetus Suture. Data from BBO and the westerly events from GIM have been chosen as representative of the normal crust of the area. The modelling results show the data to be consistent with a Moho depth of ~ 30 km (Figs 4a and b). A velocity gradient has been introduced into the base of the crust to reduce the velocity contrast at the Moho, and therefore the amplitude of the associated P_s conversion. The crustal thickness of 30 km is somewhat thinner than that found in the LISPb model, but correlates well with the depths of between 29 and 31 km as interpreted on normal incidence seismic reflection sections (Chadwick & Pharaoh 1998).

The secondary high-amplitude phase on the northerly events recorded at GIM can be explained by two possible models: (1) the additional phase is, in fact, the Moho P_s conversion, and there is a change in the crustal structure or (2) there is variation in sub-Moho structure. In the first case, if the 5.5 s phase is from the Moho, and

the average velocity and Poisson's ratio of the crust are preserved, then the causal velocity discontinuity must be at a depth of approximately 45 km. Conversely, if the Moho depth remained the same then the average crustal P -wave velocity must reduce to less than 5 km s^{-1} . Evidence from the LISPb, WINCH and CSSP profiles shows that there is no significant lateral variation of crustal velocity structure north of GIM. We have therefore assumed that the crustal structure does not change significantly over the area sampled by the GIM data and concluded that the additional phase represents a sub-Moho converter.

The $P_{s\text{mantle}}$ phase was first modelled as a sub-Moho velocity step increase. This required the introduction of a layer of unrealistically high velocity, up to 10 km s^{-1} , compared with the value of $\sim 8.5 \text{ km s}^{-1}$ for the LISPb mantle reflector (Barton 1992). To resolve this problem a low-velocity layer was introduced in the mantle. This enabled the $P_{s\text{mantle}}$ phase together with its negative precursor to be fitted (Fig. 4c). The upper boundary of the low-velocity layer is gradational to fit the observed receiver function. If a step function is used then the negative precursor to the $P_{s\text{mantle}}$ phase is of much higher amplitude than is required to fit the observed data. To test the robustness of the 1-D modelling a constrained genetic algorithm inversion has been performed (Shibutani *et al.* 1996). The inversion uses a forward model with synthetic receiver functions of unlimited bandwidth, and therefore do not reproduce the negative ringing of the positive phases of the observed receiver functions. However, the resultant synthetic receiver functions still fit the amplitude of the P_s and $P_{s\text{mantle}}$ phases well (Fig. 4c). The velocity model from the genetic algorithm inversion produces a similar magnitude of velocity discontinuity to the 1-D modelling. The receiver function method is sensitive to relative arrival times of the phases and velocity contrast, not to absolute velocity. Therefore, a significant velocity–depth trade-off exists (Ammon *et al.* 1990). The present study relies on the *a priori* velocity information provided by the LISPb and CSSP seismic refraction studies to constrain the depth of the modelled velocity discontinuities.

The receiver functions from station ORE on the north coast of Scotland provide valuable comparison with the data from GIM (Fig. 1). The receiver functions from ORE show a mantle P_s phase similar to that seen at GIM (Fig. 4d). An extension of the DRUM profile, on which the mantle Flannan and W-reflectors have been imaged, passes close to ORE. Seismic reflection waveform modelling of the W-reflector in this area reveals a negative velocity contrast at the reflecting interface (Price & Morgan 2000). The measured reflection coefficient requires that the velocity above the discontinuity is substantially lower than that of the reflector. There is no imaged velocity discontinuity above the W-reflector, and therefore a transparent negative velocity gradient is required to match the observed amplitude data. Using this model, synthetic receiver functions have been generated for ORE (Fig. 4d). The synthetic data match the observed receiver functions well, and the velocity model is similar to that of the northerly data from GIM. Importantly, the velocity contrast over the W-reflector is that which generates the mantle phase of the observed receiver functions. Price & Morgan (2000) suggest that the W-reflector is a subducted slab, and the low-velocity zone above it was caused by metasomatism during the subduction.

7 DISCUSSION AND CONCLUSIONS

The observed data show a broad correlation with the synthetic receiver functions produced using Barton's (1992) LISPb velocity model. This is not unexpected, as the two studies investigated the

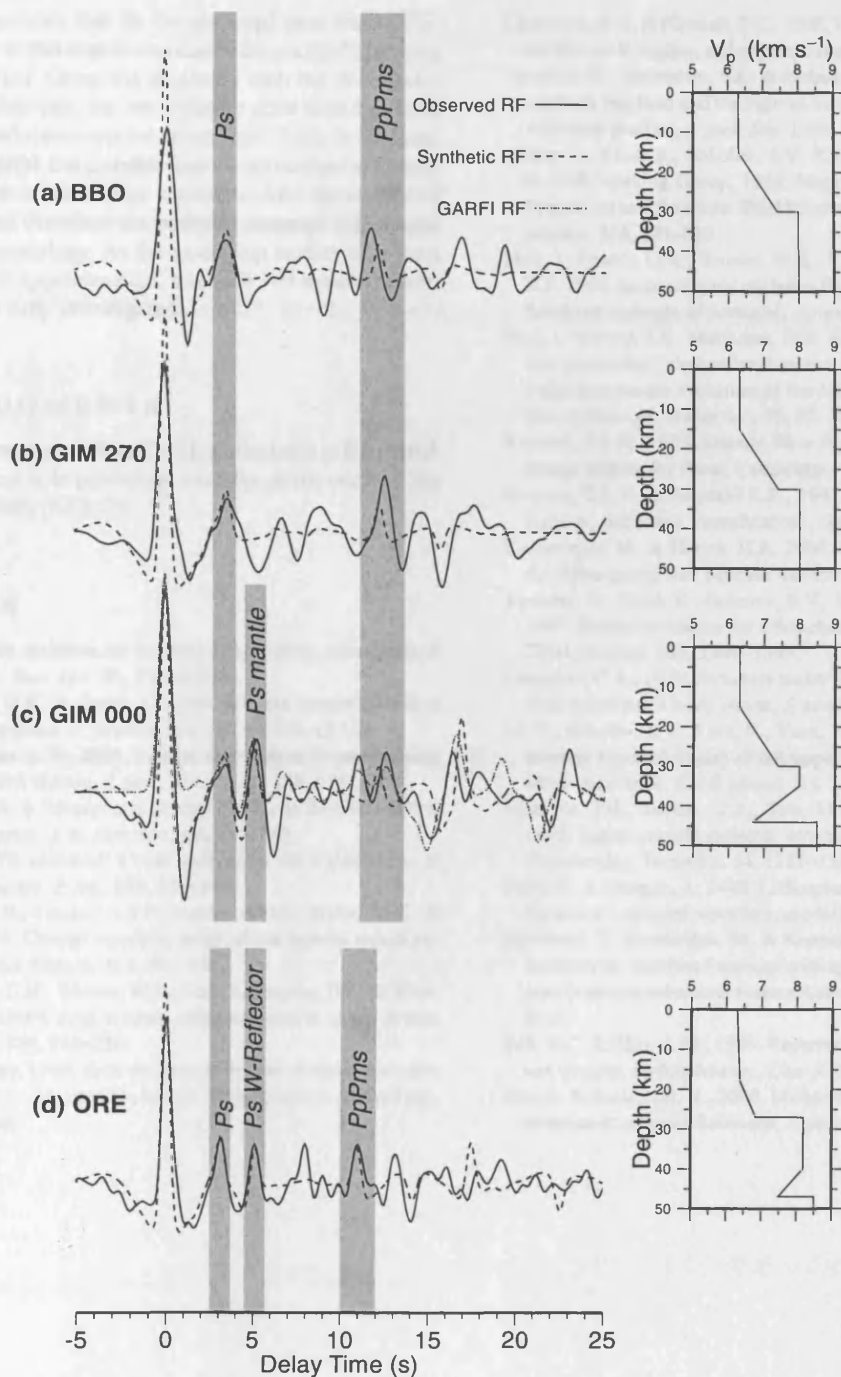


Figure 4. Results of the forward modelling of radial receiver functions. (a) BBO data and the subsequent velocity model. (b) Westerly GIM data, representing the normal velocity structure. (c) Northerly GIM data, along with the optimum velocity model. (d) ORE observed and synthetic receiver functions. The synthetic data have been calculated using a velocity model from Price & Morgan (2000). The sub-Moho structure is the W-reflector. The shaded areas highlight various phases of the receiver functions.

same structures. However, it does confirm that the short-period receiver function data are producing results consistent with the wide-angle seismic refraction models of the area. Closer examination of the data from individual stations shows that there is variation in the receiver functions relative to backazimuth. With reference to the data from EDI this is interpreted as variation in the intracrustal structure. Strong azimuthal variation in receiver function is observed in the data from GIM, with the northerly data indicating a prominent sub-

crustal phase. In contrast, the nearby stations BTA and BBO show none of the variations found at GIM. The preferred model beneath GIM producing a fit of the subcrustal phases includes a significant low-velocity zone in the upper mantle. Receiver function data from ORE that sample the W-reflector (Price & Morgan 2000) have been compared with those from GIM. The velocity model of Price & Morgan (2000) that is suggested to represent a subducted slab of oceanic crust with a low-velocity metasomatized layer above it

produces receiver functions that fit the observed data from ORE. This model is similar to that required to match the northerly receiver function data from GIM. Given the similarity with the W-reflector model it seems possible that the low-velocity zone found at GIM could result from subduction-related structures. If this is the case, given the location of GIM it is possible that the anomalous northerly mantle phase results from structures associated with the closure of the Iapetus Ocean, and therefore has probably sampled 3-D crustal and upper-mantle morphology. As the modelling studies have been carried out using a 1-D approximation, any such 3-D structural morphology has not been fully investigated.

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REFERENCES

- Ammon, C.J., 1991. The isolation of receiver effects from teleseismic *P* waveforms, *Bull. seis. Soc. Am.*, **81**, 2504–2510.
- Ammon, C.J., Randall, G.E. & Zandt, G., 1990. On the nonuniqueness of receiver function inversions, *J. geophys. Res.*, **95**, 15 303–15 318.
- Armstrong, H.A. & Owen, A.W., 2001. Terrane evolution of the paratectonic Caledonides of northern Britain, *J. geol. Soc. Lond.*, **158**, 475–486.
- Bamford, D. *et al.*, 1976. A lithospheric seismic profile in Britain—I. Preliminary results, *Geophys. J. R. astr. Soc.*, **44**, 145–160.
- Barton, P.J., 1992. LISP-B revisited: a new look under the Caledonides of northern Britain, *Geophys. J. Int.*, **110**, 371–391.
- Bott, M.H.P., Long, R.E., Green, A.S.P., Lewis, A.H.J., Sinha, M.C. & Stevenson, D.L., 1985. Crustal structure south of the Iapetus suture beneath northern England, *Nature*, **314**, 724–727.
- Brewer, J.A., Matthews, D.H., Warner, M.R., Hall, J., Smythe, D.K. & Whittington, R.J., 1983. BIRPS deep seismic reflection studies of the British Caledonides, *Nature*, **305**, 206–210.
- British Geological Survey, 1996. *Tectonic map of Britain, Ireland and adjacent areas*, Pharaoh T.C., Morris J.H., Long C.B. & Ryan P.D. (compilers), Keyworth, Nottingham.
- Chadwick, R.A. & Pharaoh, T.C., 1998. The seismic reflection Moho beneath the United Kingdom and adjacent areas, *Tectonophysics*, **299**, 255–279.
- Freeman, B., Klemperer, S.L. & Hobbs, R.W., 1998. The deep structure of northern England and the Iapetus Suture zone from BIRPS deep seismic reflection profiles, *J. geol. Soc. Lond.*, **145**, 727–740.
- Gosler, J., Kind, R., Sobolev, S.V., Kampf, H., Wylegalla, K., Stiller, M. & TOR Working Group, 1999. Major crustal features between the Harz Mountains and the Baltic Shield derived from receiver functions, *Tectonophysics*, **314**, 321–333.
- Hall, J., Powell, D.W., Warner, M.R., El-Isa, Z.H., Adesanya, O. & Bluck, B.J., 1983. Seismological evidence for shallow crystalline basement in the Southern Uplands of Scotland, *Nature*, **305**, 418–420.
- Hall, J., Brewer, J.A., Matthews, D.H. & Warner, M.R., 1984. Crustal structure across the Caledonides from the 'WINCH' seismic reflection profile: influences on the evolution of the Midland Valley of Scotland, *Trans. R. Soc. Edinburgh, Earth Sci.*, **75**, 97–109.
- Kennett, B.L.N., 1983. *Seismic Wave Propagation in Stratified Media*, Cambridge University Press, Cambridge.
- Kennett, B.L.N. & Engdahl, E.R., 1991. Traveltimes for global earthquake location and phase identification., *Geophys. J. Int.*, **105**, 429–465.
- Knapmeyer, M. & Harjes, H.P., 2000. Imaging crustal discontinuities and the down-going slab beneath western Crete, *Geophys. J. Int.*, **143**, 1–21.
- Kosarev, G., Kind, R., Sobolev, S.V., Yuan, X., Hanka, W. & Oreshin, S., 1999. Seismic evidence for a detached Indian lithospheric mantle beneath Tibet, *Science*, **283**, 1306–1309.
- Langston, C.A., 1979. Structure under Mount Rainier, Washington, inferred from teleseismic body waves, *J. geophys. Res.*, **84**, 4749–4762.
- Li, X., Sobolev, S.V., Kind, R., Yuan, X., Estabrook, Ch., 2000. A detailed receiver function image of the upper mantle discontinuities in the Japan subduction zone, *Earth planet. Sci. Lett.*, **183**, 527–541.
- McBride, J.H., Snyder, D.B., Tate, M.P., England, R.W. & Hobbs, R.W., 1995. Upper mantle reflector structure and origin beneath the Scottish Caledonides, *Tectonics*, **14**, 1351–1367.
- Price, C. & Morgan, J., 2000. Lithospheric structure north of Scotland—II. Poisson's ratios and waveform modelling, *Geophys. J. Int.*, **142**, 737–754.
- Shibutani, T., Sambridge, M. & Kennett, B.N.L., 1996. Genetic algorithm inversion for receiver functions with application to the crust and uppermost mantle structure beneath eastern Australia, *Geophys. Res. Lett.*, **23**, 1829–1832.
- Zelt, B.C. & Ellis, R.M., 1999. Receiver–function studies in the Trans Hudson Orogen, Saskatchewan., *Can. J. Earth Sci.*, **36**, 585–603.
- Zhu, L. & Kanamori, H., 2000. Moho variation in southern California from teleseismic receiver functions, *J. geophys. Res.*, **105**, 2969–2980.

Appendix C

Data CD

Inserted into the back page of this thesis is a CD containing the original seismograms and receiver functions used in the receiver function project. In directory `data` there is a gzipped tar file for each station. The directory structure within each tar file is easy to follow. The unfiltered seismograms are stored in the `good` and `trash` directories. In the root directory of the CD `readme.txt` details the contents of the directory structure comprehensively.

Bibliography

- Abramovitz, T., Landes, M., Thybo, H., Jacob, B., Prodehl, C., 1999. Crustal velocity structure across the Tornquist and Iapetus Suture Zones – a comparison based on MONA LISA and VARNET data. *Tectonophysics* 314, 69–82.
- Al-Kindi, S., 2000. A study into underplating as a cause of Tertiary epeirogenic uplift and denudation in the East Irish Sea. *Lithos Science Report* 2, 95–100.
- Al-Kindi, S., 2002. A seismological study of Cenozoic epeirogeny across the British Isles. Ph.D. thesis, University of Cambridge.
- Ammon, C. J., 1991. The isolation of receiver effects from teleseismic P-waveforms. *Bull. Seis. Soc. Am.* 81 (6), 2504–2510.
- Ammon, C. J., Randall, G. E., Zandt, G., 1990. On the nonuniqueness of receiver function inversions. *J. Geophys. Res.* 95 (B10), 15303–15318.
- Ando, M., Crampin, S., Booth, D., Redmayne, D., 1987. A study of anisotropy beneath Great Britain using the polarizations of teleseismic shear-waves recorded by BGS/UKNET. *J. Phys. Earth* 35, 469–485.
- Assumpção, M., Bamford, D., 1978. LISPB - V. Studies of crustal shearwaves. *Geophys. J. R. astr. Soc.* 54, 60–73.
- Bamford, D., Faber, S., Jacob, B., Kaminski, W., Nunn, K., Prodehl, C., Fuchs, K., King, R., Wilmore, P., 1976. A Lithospheric Seismic Profile in Britain - I Preliminary Results. *Geophys. J. R. astr. Soc.* 44, 145–160.
- Bamford, D., Jentsch, M., Prodehl, C., 1979. P_n anisotropy studies in northern Britain and the eastern and western United States. *Geophys. J. R. astr. Soc.* 57, 397–429.

- Bamford, D., Nunn, K., Prodehl, C., Jacob, B., 1977. LISP-B-III. Upper crustal structure of northern Britain. *J. Geol. Soc. Lon.* 133, 481–488.
- Bamford, D., Nunn, K., Prodehl, C., Jacob, B., 1978. LISP-B - IV. Crustal Structure of Northern Britain. *Geophys. J. R. astr. Soc.* 54, 43–60.
- Barton, P. J., 1992. LISP-B revisited: a new look under the Caledonides of northern Britain. *Geophys. J. Int.* 110, 371–391.
- Bluck, B. J., Gibbons, W., Ingham, J. K., 1992. Terranes. In: Cope, J. C., Ingham, J. K., Rawson, P. F. (Eds.), *Atlas of Palaeogeography and Lithofacies*. Geol. Soc. London, pp. 1–4.
- Bott, M. H. P., Long, R. E., Green, A. S. P., Lewis, A. H. J., Sinha, M. C., Stevenson, D. L., 1985. Crustal structure south of the Iapetus suture beneath northern England. *Nature* 314, 724–727.
- Brewer, J. A., Matthews, D. H., Warner, M. R., Hall, J., Smythe, D., Whittington, R. J., 1983. BIRPS deep seismic reflection studies of the British Caledonides. *Nature* 305, 206–209.
- Brooks, M., Doody, J. J., Al-Rawi, F. R. J., 1984. Major crustal reflectors beneath SW England. *J. Geol. Soc. Lon.* 141, 97–103.
- Burdick, L. J., Langston, C. A., 1977. Modeling crustal structure through the use of converted phases in teleseismic body-wave forms. *Bull. Seis. Soc. Am.* 67 (3), 677–691.
- Busby, J. P., Smith, N. J. P., 2001. The nature of the Variscan basement in southeast England: evidence from integrated potential field modelling. *Geol. Mag.* 138 (6), 669–685.
- Cassidy, J. F., 1992. Numerical experiments in broadband receiver function analysis. *Bull. Seis. Soc. Am.* 82 (3), 1453–1474.
- Chadwick, R. A., Pharaoh, T. C., 1998. The seismic reflection Moho beneath the United Kingdom and adjacent areas. *Tectonophysics* 299, 255–279.
- Cheverot, S., van der Hilst, R. D., 2000. The Poisson's ratio of the Australian crust: geological and geophysical implications. *Earth Plan. Sci. Lett.* 183, 121–132.
- Cheverot, S., Vinnik, L., Montagner, J.-P., 1999. Global scale analysis of the mantle *Pds* phases. *J. Geophys. Res.* 104 (B9), 20203–20219.

- Christensen, N. I., 1995. Poisson's ratio and crustal seismology. *J. Geophys. Res.* 101 (B2), 3139–3156.
- Christensen, N. I., Mooney, W. D., 1995. Seismic velocity structure and composition of the continental crust: A global view. *J. Geophys. Res.* 100 (B7), 9761–9788.
- Clift, D., Turner, J., 1998. Paleogene igneous underplating and subsidence anomalies in the Rockall-Faeroe-Shetland area. *Mar. Petrol. Geol.* 15, 223–243.
- Conway, A., Dentith, M. C., Doody, J. J., Hall, J., 1987. Preliminary interpretation of upper crustal structure across the Midland Valley of Scotland from two East-West seismic refraction profiles. *J. Geol. Soc. Lon.* 144, 865–870.
- Davidson, K. A. S., Sola, M., Powell, W., Hall, J., 1984. Geophysical model for the Midland Valley of Scotland. *Trans. R. Soc. Edin. Earth Sci.* 75, 175–181.
- Dentith, M. C., Hall, J., 1989. MAVIS - an upper crustal seismic refraction experiment in the Midland Valley of Scotland. *Geophys. J. Int.* 99, 627–643.
- Deuss, A., Woodhouse, J. H., 2002. A systematic search for mantle discontinuities using SS-precursors. *Geophys. Res. Lett.* 29 (8), 90–1–90–4.
- Dueker, K. G., Sheehan, A. F., 1997. Mantle discontinuity structure from midpoint stacks of converted P to S waves across the Yellowstone hotspot track. *J. Geophys. Res.* 102 (B4), 8313–8327.
- Edwards, J. W. F., Blundell, D. J., 1984. Summary of seismic refraction experiments in the English Channel, Celtic Sea and St George's Channel. Marine Geophysics Report 144, British Geological Survey.
- Freeman, B., Klemperer, S. L., Hobbs, R. W., 1988. The deep structure of northern England and the Iapetus Suture zone from BIRPS deep seismic reflection profiles. *J. Geol. Soc. Lon.* 145, 727–740.
- Gaherty, J., Jordan, T. H., 1995. Lehmann discontinuity as the base of an anisotropic layer beneath continents. *Science* 268 (5126), 1468–1471.

- Gossler, J., Kind, R., Sobolev, S. V., Kämpf, H., Wylegalla, K., Stiller, M., Group, T. W., 1999. Major crustal features between the Harz Mountains and the Baltic Shield derived from receiver functions. *Tectonophysics* 314, 321–333.
- Grandjean, G., Guennoc, P., Recq, M., Andréo, P., 2001. Refraction/wide-angle reflection investigation of the Cadomian crust between northern Brittany and the Channel Islands. *Tectonophysics* 331, 45–64.
- Gu, Y. J., Dziewonski, A. M., Ekström, G., 2001. Preferential detection of the Lehmann discontinuity beneath continents. *Geophys. Res. Lett.* 28, 4655–4658.
- Hall, J., Brewer, J. A., Matthews, D. R., Warner, M. R., 1984. Crustal structure across the Caledonides from the "WINCH" seismic reflection profile: influences on the evolution of the Midland Valley of Scotland. *Trans. R. Soc. Edin. Earth Sci.* 75, 97–109.
- Holder, A. P., Bott, M. H. P., 1971. The crustal structure in the vicinity of south-west England. *Geophys. J. R. astr. Soc.* 23, 465–489.
- Jacob, A. W. B., Kaminski, W., Murphy, T., Phillips, W. E. A., Prodehl, C., 1985. A crustal model for a northeast-southwest profile through Ireland. *Tectonophysics* 113, 75–103.
- Jones, K. A., Warner, M. R., Morgan, R. P. L., Morgan, J. V., Barton, P. J., Price, C. E., 1996. Coincident normal-incidence and wide-angle reflections from the Moho: evidence for crustal seismic anisotropy. *Tectonophysics* 264, 205–217.
- Jones, S. M., White, N., Clarke, B. J., Rowley, E., Gallagher, K., 2002. Past and present influences of the Iceland Plume on sedimentation. In: Dore, A. (Ed.), *Uplift and Exhumation of the NE Atlantic margin*. Geol. Soc. Spec. Pub.
- Kennett, B. L. N., 1983. *Seismic Wave Propagation in Stratified Media*. Cambridge University Press.
- Kennett, B. L. N., Engdahl, E. R., 1991. Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.* 105, 429–465.
- Klemperer, S., Hobbs, R., 1991. *The BIRPS Atlas: Deep seismic reflection profiles around the British Isles*. Cambridge University Press.

- Knapmayer, M., Harjes, H., 2000. Imaging crustal discontinuities and the downgoing slab beneath western Crete. *Geophys. J. Int.* 143, 1–21.
- Kosarev, G., Kind, R., Sobolev, S., Yuan, X., Hanka, W., Oreshin, S., 1999. Seismic evidence for a detached Indian lithospheric mantle beneath Tibet. *Science* 283, 1306–1309.
- Langston, C. A., 1979. Structure under Mount Rainier, Washington, inferred from telesismic body waves. *J. Geophys. Res.* 84 (B9), 4749–4762.
- Lebedev, S., Cheverot, S., van der Hilst, R. D., 2003. Correlation between the shear-speed structure and thickness of the mantle transition zone. *Phys. Earth Planet. Int.* (in press).
- Lebedev, S., Chevrot, S., van der Hilst, R. D., 2002. Seismic evidence for olivine phase changes at the 410- and 660-kilometer discontinuities. *Science* 296, 1300–1302.
- Levin, V., Park, J., 1997. *P*-to-*SH* conversions in a flat-layered medium with anisotropy of arbitrary orientation. *Geophys. J. Int.* 131, 253–266.
- Levin, V., Park, J., 1998. *P*-to-*SH* conversions in layered media with hexagonally symmetric anisotropy: A cook book. *Pure App. Geophys.* 151, 669–697.
- Li, X., Kind, R., Yuan, X., 2003. Seismic study of upper mantle transition zone beneath hotspots. *Phys. Earth Planet. Int.* (in press).
- Li, X., Priestley, K., Sobolev, S., Tilmann, F., Yuan, X., 2000a. Mapping the Hawaiian plume conduit with converted seismic waves. *Nature* 405, 938–941.
- Li, X., Sobolev, S., Kind, R., Yuan, X., Estabrook, C., 2000b. A detailed receiver function image of the upper mantle discontinuities in the Japan subduction zone. *Earth Plan. Sci. Lett.* 183, 527–541.
- Maguire, P., Bermingham, P., Francis, D. J., 1981. A relative *P*-wave delay study between Eskdalemuir and Charnwood Forest. *Geophys. J. R. astr. Soc.* 65, 229–235.
- Maguire, P. K. H., Francis, D. J., Whitcombe, D. N., 1985. Determination of the three-dimensional seismic structure of the crust and upper mantle in the Central Midlands of England. *Geophys. J.R. astr. Soc.* 83, 347–362.
- Maroof, S. I., 1973. Geophysical investigations of the Carboniferous and Precambrian formations of the East Midlands of England. Ph.D. thesis, University of Leicester.

- Mason, F., Hauser, F., Jacob, A., 1999. The lithospheric trace of the Iapetus Suture in SW Ireland from teleseismic data. *Tectonophysics* 302, 83–98.
- McBride, J. H., Snyder, D. B., Tate, M. P., England, R. W., Hobbs, R. W., 1995. Upper mantle reflector structure and origin beneath the Scottish Caledonides. *Tectonics* 14 (5), 1351–1367.
- McGeary, S., 1989. Reflection seismic evidence of a Moho offset beneath the Walls Boundary strike-slip Fault. *J. Geol. Soc. Lon.* 146, 261–269.
- McGeary, S., Cheadle, M. J., Blundell, D. J., 1987. Crustal structure of the continental shelf around Britain derived from BIRPS deep seismic profiling. In: Brooks, J., Glennie, K. (Eds.), *Petroleum Geology of North West Europe*. pp. 33–41.
- McGeary, S., Warner, M. R., 1985. Seismic profiling of the continental lithosphere. *Nature* 317, 795–797.
- Meredith, J., Pearce, R., 1991. Seismic velocities beneath central and southern Britain from broadband surface waves. *Phys. Earth Planet. Inter.* 68, 230–249.
- Morgan, P. R. L., Barton, P. J., Warner, M., Morgan, J., Price, C., Jones, K., 2000. Lithospheric structure north of Scotland-I. P-wave modelling, deep reflection profiles and gravity. *Geophys. J. Int.* 142, 716–736.
- Nyblade, A., Knox, R. P., Gurrola, H., 2000. Mantle transition zone thickness beneath Afar: implications for the origin of the Afar hotspot. *Geophys. J. Int.* 142, 615–619.
- Parr, R. S., Hutton, V. R. S., 1993. Magnetotelluric studies in and adjacent to the Northumberland Basin, Northern England. *Phys. Earth Plan. Int.* 81, 43–66.
- Price, C., Morgan, J., 2000. Lithospheric structure north of Scotland-II. Poisson's ratios and waveform modelling. *Geophys. J. Int.* 142, 737–754.
- Ramesh, D. S., Kind, R., Yuan, X., 2002. Receiver function analysis of the North American crust and upper mantle. *Geophys. J. Int.* 150, 91–108.
- Sandvol, E., Seber, D., Calvert, A., Barazangi, M., 1998. Grid search modelling of receiver functions: Implications for crustal structure in the Middle East and North Africa. *J. Geophys. Res.* 103 (B11), 26899–26917.

- Sheehan, A. F., Abers, G. A., Jones, C. H., Lerner-Lam, A. L., 1995. Crustal thickness variation across the Colorado Rocky Mountains from teleseismic receiver functions. *J. Geophys. Res.* 100 (B10), 20391–20404.
- Shen, Y., Solomon, S., Bjarnason, I., Wolfe, C., 1998. Seismic evidence for a lower-mantle origin of the Iceland plume. *Nature* 395, 62–65.
- Shibutani, T., Sambridge, M., Kennet, B. N. L., 1996. Genetic algorithm inversion for receiver functions with application to the crust and uppermost mantle structure beneath eastern Australia. *Geophys. Res. Lett.* 23, 1829–1832.
- Smythe, D. E., Dobinson, A., McQuillin, R., Brewer, J. A., Blundell, D. J., Kelk, B., 1982. Deep structure of the Scottish Caledonides revealed by the Moiré reflection profile. *Nature* 299, 338–340.
- Snyder, D. B., Flack, C. A., 1990. A Caledonian age for the reflectors within the mantle lithosphere north and west of Scotland. *Tectonics* 9 (4), 903–922.
- Soper, N. J., England, R. W., Snyder, D. B., Ryan, P., 1992. The Iapetus suture zone in England, Scotland and eastern Ireland: a reconciliation of geological and deep seismic data. *J. Geol. Soc., Lon.* 149, 697–700.
- Tomlinson, J., Denton, P., Maguire, P., Evans, J., 2003. UK crustal structure close to the Iapetus Suture: A receiver function perspective. *Geophys. J. Int.* 154, 659–665.
- Ward, G., Warner, M., BIRPS, 1992. Lower crustal lithology from shear-wave seismic reflection data. In: *Continental Lithosphere: Deep Seismic Reflections*. Vol. 22 of AGU Geodynamic Series. pp. 343–349.
- Warner, M. R., McGearry, S., 1987. Absolute reflection co-efficients from deep seismic reflections. *Geophys. J.R. astr. Soc.* 89, 223–230.
- Whitcombe, D. N., Maguire, P. K. H., 1980. An analysis of the velocity structure of the Precambrian rocks of Charnwood Forest. *Geophys. J.R. astr. Soc.* 52, 405–416.
- Whittaker, A. (Ed.), 1985. *Atlas of onshore sedimentary basins in England and Wales: Post-Carboniferous tectonic stratigraphy*. British Geological Survey.
- Woodcock, N., Strachan, R., 2000. *Geological history of Britain and Ireland*. Blackwell Science.

- Yuan, X., Ni, J., Kind, R., Mechie, J., Sandvol, E., 1997. Lithospheric and upper mantle structure of southern Tibet from a seismological and passive source experiment. *J. Geophys. Res.* .
- Zandt, G., Ammon, C. J., 1995. Continental crust composition constrained by measurements of crustal Poisson's ratio. *Nature* 374, 152–154.
- Zelt, B. C., Ellis, R. M., 1999. Receiver–function studies in the Trans–Hudson Orogen, Saskatchewan. *Can. J. Earth Sci.* 36, 585–603.
- Zelt, C., Sain, K. S., Naumenko, J., Sawyer, D., 2003. Assessment of crustal velocity models using seismic refraction and reflection tomography. *Geophys. J. Int.* 153, 609–626.
- Zhu, L., Kanamori, H., 2000. Moho depth variation in southern California from teleseismic receiver functions. *J. Geophys. Res.* 105 (B2), 2969–2980.