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Seismological Modelling of Global Earth Structure

by

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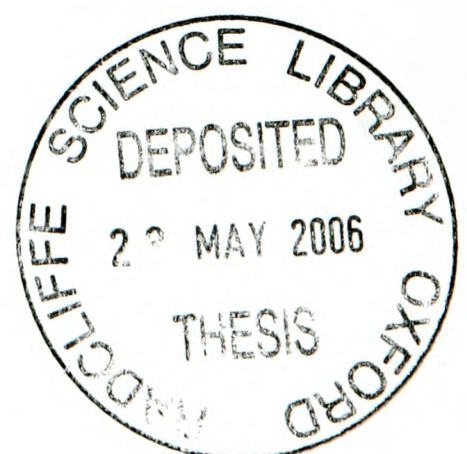
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Abstract

Upper mantle discontinuities from PP- and SS-precursors

The 410km discontinuity is examined using two global datasets of precursors to the *PP* and *SS* phases. The precursor amplitudes are used to constrain the impedance contrast, and examine lateral variations in the reflection coefficient of the discontinuity. P- and S-wave reflection amplitudes vary over different scale lengths, which could be due to the presence of melt, water or other chemical heterogeneities in the transition zone. Models for 410-topography are also derived, and the relationship between topography and transition zone seismic velocity anomalies examined. A moderate negative correlation exists between long wavelength 410-topography and transition zone velocity anomalies. However, when shorter wavelengths are included the relationship becomes more complex, and in some regions positive correlations are seen. This suggests that long wavelength variation of discontinuity topography and seismic velocity, is due to thermal effects. However, at shorter wavelengths the influence of chemical heterogeneities becomes important. Different spectra for thermal and chemical heterogeneity suggests that chemical anomalies can survive in convecting mantle.

Lower mantle reflectors and S-wave scattering

The lowermost mantle is investigated using a phase stripping technique and two migration methods: a backprojection, and a scheme with weights based on the Generalised Radon Transform. Resolution of the results is tested by migrating synthetic datasets. In some regions the results can be simulated using fairly simple distributions of point scatterers, but elsewhere the results require more complicated structures. The results identify several important properties of the D'' region including reflectors within the D'' region and a complex pattern of positive and negative scattering potentials near the core-mantle boundary. The results also show the presence of an intermittent D'' discontinuity which is not a continuous, nor a global, feature. This suggests that the D'' discontinuity is caused by either localised structures, such as thermal or chemical heterogeneity, or a global boundary with a variable impedance contrast.

Seismological Modelling of Global Earth Structure

Extended abstract

Upper mantle discontinuities from PP- and SS-precursors

A number of previous studies have examined lateral variations in upper mantle and transition zone discontinuities, by aligning large numbers of long-period seismograms on a surface reflection phase (termed *SS* and *PP*) before stacking. In particular there has been much emphasis on seismic discontinuities which occur near 410, 520 and 660km depth (Flanagan & Shearer (1998a); Gu *et al.* (1998); Flanagan & Shearer (1999); Deuss & Woodhouse (2002)), which result from phase transformations of olivine.

In the first part of this thesis I compile two global datasets of precursors to the *PP* and *SS* phases (termed *PdP* and *SdS* where *d* is reflection depth). The results show the presence of reflectors in the upper mantle and transition zone including, regionally variable discontinuities around 220 and 300km depth, and a 520km discontinuity. There is no evidence for a P-wave reflection from the 660km discontinuity. Although an *S660S* phase is seen in virtually all regions considered, which together with synthetic modelling suggests that the P-wave impedance change at 660km is overestimated in reference models such as PREM (Dziewonski & Anderson (1981)).

The majority of discussion is focussed on the *SS* and *PP* precursors from the 410km discontinuity, which are a major feature of both datasets. Previous global studies of precursors have been limited to describing the topography of the 410km discontinuity. Here analysis is extended to examination of the reflection coefficient of the 410km discontinuity. Using measurements of *P410P* and *S410S* amplitudes the impedance contrasts across the 410km discontinuity is constrained. Lateral variations in the P-wave impedance contrast at 410km are also found, which is typically low under North America and China, and higher beneath the North Pacific. The S-wave impedance contrast shows less variability on the regional scale. However, analysis of *P410P* – *PP*

and $S410S - SS$ amplitudes over smaller areas (by binning traces into spherical caps) shows that the S-wave reflection coefficient varies over much shorter scale lengths than that for P-waves. The different patterns of variation for $P410P$ and $S410S$ reflection amplitudes could be due to the presence of melt, water or other chemical heterogeneities in the transition zone. Other factors such as temperature or mantle olivine content variations could also influence precursor amplitudes, but would be expected to lead to correlated variations, and so cannot explain all the observed variation.

Topography of the 410km discontinuity is also discussed. Recent models suggest that the precursor derived topography of the 410km discontinuity is dominated by low degree spherical harmonics. However, there remain some discrepancies as to the details of the topography and its relation to mantle processes. Here a new model for 410-topography which is constrained using both $P410P$ and $S410S$ measurements is presented, and the relationship between 410-topography and seismic velocity anomalies in the transition zone is examined. A moderate negative correlation between discontinuity topography and transition zone seismic velocity anomalies exists at very low spherical harmonic degrees. This is consistent with the expected relationships between, the position of the $\alpha - \beta$ olivine phase boundary, seismic velocity anomalies, and mantle temperatures. However, when shorter wavelengths are included the relationship becomes more complex. In small regions beneath Northeast Asia and the Western Pacific positive correlations between seismic velocities and discontinuity topography are observed. This is interpreted as evidence for the effects of chemical heterogeneities on seismic velocities and the 410km discontinuity. The results suggest there is a long wavelength temperature pattern in the mantle, however, at shorter wavelengths the influence of chemical heterogeneities becomes visible in seismic velocity anomalies and mantle discontinuity topography. This separation of the spectra for thermal and chemical heterogeneities suggests that short wavelength chemical heterogeneities can survive in convecting mantle.

Lower mantle reflectors and S-wave scattering

Knowledge of structure in the lowermost mantle is of central importance to understanding mantle processes such as: the nature of the thermo-chemical boundary layer, mantle mixing and the distribution of compositional heterogeneity, plume formation, and core-mantle coupling. Despite this, the lower mantle is poorly understood. Recent seismological investigations have either concentrated on the use of travel time and amplitude data to retrieve detailed structure over small regions, or used global travel time datasets to constrain very long wavelength structure. This study, attempts to bridge this gap by applying migration processing to a global dataset of seismograms.

The data consists of 1075, high quality, transverse component, seismograms in the time window containing the S and ScS phases. The S and ScS pulses are removed from the dataset by applying the phase stripping technique of Kendall & Shearer (1994). This leaves more subtle arrivals which originate from lower mantle scattering. The structures responsible are located using two migration techniques, a simple backprojection and a more sophisticated migration with weights based on the Generalised Radon Transform (GRT). The backprojection is useful for the identification of strong scattering structures, however, the technique is limited by the use of deconvolved records and the lack of allowance for factors which affect the amplitude of scattered waves. For the GRT migration, a forward model for the scattered wavefield is developed from a distribution of point heterogeneities using the Born approximation and ray theoretical Green's functions. An inverse process for retrieving scattering potential as a function of position was developed by Miller *et al.* (1987) for acoustic media. Here this scheme is extended to S-to-S scattering in isotropic elastic media. The affects of biases in the source-receiver distribution on the migration results is tested by computing synthetic datasets using various distributions of point scatterers and the forward model for the scattered wavefield.

Migration results are presented for three regions of the lower mantle beneath Alaska, North Central Asia, and New Zealand. Positive scattering potentials are found at the base of the mantle. There is some suggestion that these high velocity zones are associated with high velocity anomalies in tomographic models, although disagreements between the two tomography models considered

here (S20RTS Ritsema *et al.* (1999), and SKS12WM13 Liu & Dziewonski (1994)) means that this is not conclusive. Beneath North Central Asia a complicated mix of positive and negative scattering potentials is seen at the base of the mantle.

Beneath Alaska, and North Central Asia a D'' reflector is seen. The structure beneath Alaska can be simulated with fairly simple flat discontinuity ~ 250 km above the core-mantle boundary. However, the exact dimensions of the scattering body are poorly constrained. Beneath North Central Asia the results have better resolution. Here the D'' reflector is more complicated, changing in depth between 250-400km above the CMB. The D'' reflector is not continuous beneath North Central Asia, and beneath southern New Zealand there is a region with no evidence for a D'' discontinuity in the range ~ 2800 - 2100 km depth. The results suggest that the D'' reflector is not globally present nor a continuous surface. This requires that the D'' discontinuity is caused by either localised structures, such as and thermal or chemical heterogeneity, or a global boundary with a variable impedance contrast.

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List of Abbreviations

‘410’	410km discontinuity
‘520’	520km discontinuity
‘660’	660km discontinuity
ORF	Ocean Response Function
CMB	Core-mantle Boundary
GRT	Generalised Radon Transform
TBL	Thermal boundary layer
CBL	Chemical boundary layer
TCBL	Thermo-chemical boundary layer
HVZ	High velocity zone
LVZ	Low velocity zone
ULVZ	Ultra-low velocity zone

Chapter 1

Introduction

Global seismology concerns the analysis of recordings from distant Earthquakes. The first teleseism was recorded in Potsdam Germany in 1889, which opened the door to the determination of the Earth's spherical structure throughout the first half of the 20th century. The first seismic discontinuity was identified by Andrija Mohorovičić, by analysis of seismic signals generated from the Croatia earthquake of 1909. The observation of a refraction from the base of the crust (subsequently termed Moho) was some of the first evidence for discontinuous jumps in seismic impedance within the Earth. Around the same time R. D. Oldham and Beno Gutenberg identified another major change in seismic properties associated with the boundary between the mantle and outer core (the CMB). The Earth's solid inner core was finally identified by Inge Lehmann in 1936. By the 1940's the major internal boundaries of the Earth had been established as shown in figure 1.1 (Watts & Tyler (1976)).

Throughout the second half of the 20th century finer scale delineation of Earth structure became possible with the development of global seismic networks. In the 1960's it became apparent that the mantle could be subdivided into upper, transition zone and lower mantle, by two discontinuities at around 410km and 660km depth. In the first part of this thesis upper mantle discontinuities are investigated with particular focus on the 410km discontinuity using precursors to the surface reflection phases *PP* and *SS*.

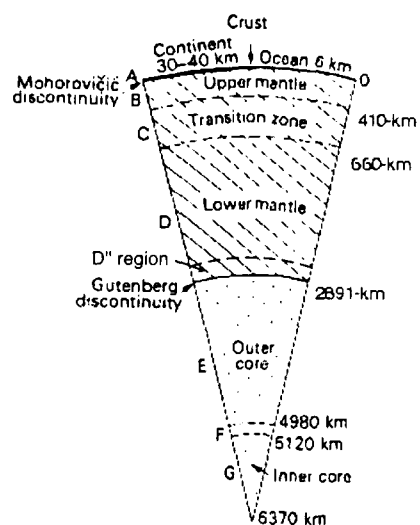


Figure 1.1: The major internal divisions of the Earth.

In chapters 3 and 4 I give a seismological and mineral physics background relevant to transition zone discontinuities, and outline how this information can be used to interpret the behaviour of the 410km discontinuity. In chapters 5 through 7 the techniques for assembly and analysis of two global datasets of *PP* and *SS* precursors are described. In chapter 8 variations in the amplitude of reflections from the 410km discontinuity are discussed and in chapter 9 a new global model for 410-topography is presented. The relationship between 410-depth and transition zone velocity anomalies is also examined. The results have implications for the distributions of chemical and thermal heterogeneities in the mantle.

Coincident with the discovery of upper mantle and transition discontinuities came the observation of structure in the lower mantle. In particular there has been much recent emphasis on the understanding of the lowermost ~ 400 km of the mantle (termed the D'' region), which is believed to play an important role in such processes as mantle convection and mixing, the generation of mantle plumes, the distribution of chemical heterogeneity in the mantle, and core-mantle coupling. However, the lowermost mantle still remains poorly understood and current seismological techniques either use amplitudes and travel times of high frequency waves to investigate structure over small isolated regions, or use just global travel time datasets to constrain very long wavelength structure. In the second part of this thesis I describe and apply a different approach which at-

tempts to bridge this gap by applying high resolution migration processing techniques to a global dataset.

In chapter 13 a phase stripping technique is described and applied to a dataset of 1047 high quality seismograms. The technique isolates scattered S-wave phases originating in the lower mantle. Two migration methods are used to find the positions of these scattering bodies in the lower mantle. The first (chapter 14), a simple backprojection, is useful for identifying positions of strong scattering. For the second migration method an expression for the scattered wavefield in isotropic media is developed using ray theory and the Born approximation (chapter 15). In chapter 16 a migration scheme for reconstructing seismic velocity anomaly within the Earth from the scattered wavefield is outlined. In particular the results for three regions of the lower mantle beneath Alaska (chapter 17), North Central Asia (chapter 18), and New Zealand (chapter 19) are discussed. The migration results reveal several features of the lowermost mantle, such as the presence of a D'' discontinuity with topography, and bodies with positive and negative scattering potentials near the CMB. The results from this part of the thesis have implications for our understanding of the structure and dynamics of the lowermost mantle.

Part I

Upper mantle discontinuities from PP and SS-precursors

Chapter 2

Introduction

2.1 Background

Seismic discontinuities represent abrupt changes in the Earth's elastic properties, and examples such as the Moho and CMB are due to major changes in the Earth's composition. Over the 100 years since the first identification of the Moho seismologists have refined techniques that allow identification of more subtle changes in seismic properties and have used these methods to identify a wide range of discontinuities. Some of these discontinuities are global in nature, while others are only regional features. Three global discontinuities feature in the upper mantle of the reference model PREM at 220, 410 and 660km depth (Dziewonski & Anderson (1981), figure 2.1). Observations of these and other discontinuities along with the knowledge of the associated changes in material properties, furthers our knowledge and understanding of the Earth's internal processes.

This part of the thesis focusses on the discussion of upper mantle discontinuities observed using underside reflections of P- and S-waves. In particular, reflections from the 410km depth discontinuity are studied. The behaviour of the 410km discontinuity is potentially a very important indicator of thermal and or chemical heterogeneity in the mantle.

First I give a brief discussion of the various techniques used by seismologists to identify upper

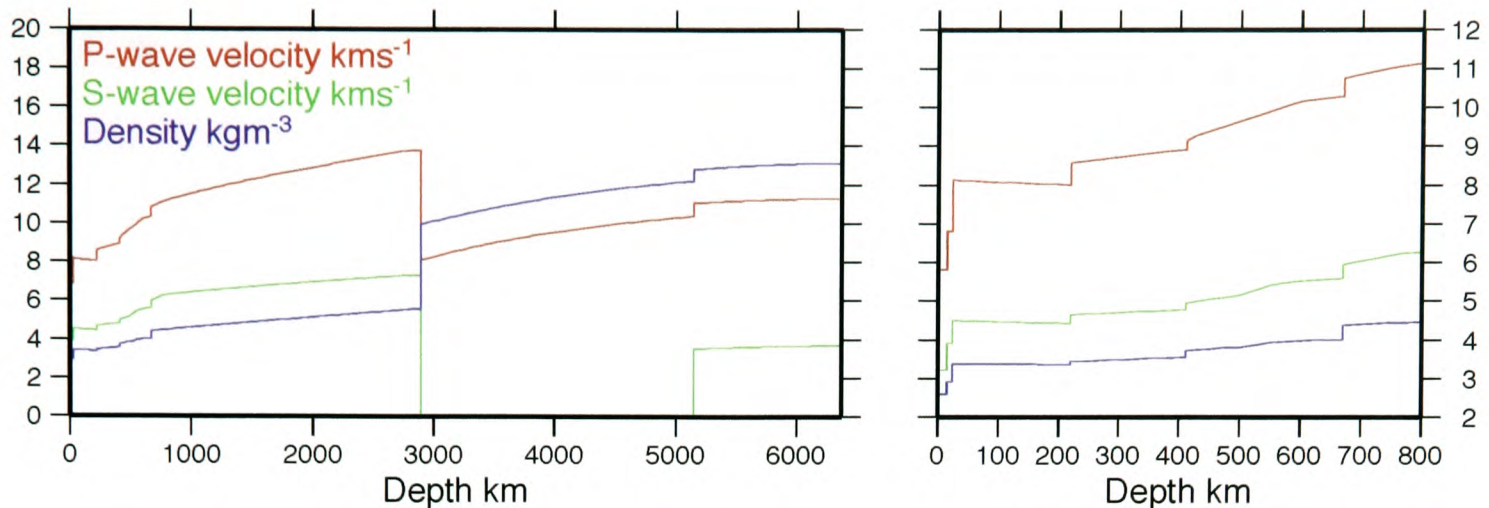


Figure 2.1: (left) The spherically symmetric earth model PREM (Dziewonski & Anderson (1981)). (right) Upper mantle and transition zone of the PREM model. It features 3 upper mantle seismic discontinuities (jumps in seismic velocity and or density) at 220, 410, and 660km depth.

mantle discontinuities. Then a background of the mineral physics relevant to transition zone discontinuities is presented. In the following chapters, this information is used to suggest alternative interpretations for the 410km discontinuity before describing the method and observations of this study. In particular variations in the reflection amplitudes from the discontinuity are analysed, and a new global model for topography of the 410km discontinuity is presented, The results have implications for the distributions of chemical and thermal heterogeneities in the mantle.

2.2 Seismological techniques for imaging discontinuities

Seismological techniques for the detection and observation of discontinuities can be divided into the identification of refracted phases and those using reflected or converted phases. The analysis of refracted phases generally requires a network of stations close to the source. The limited depth of penetration for active sources means that, in modern discontinuity seismology, refracted phases are only used for regional studies where there is a network of stations local to an area with several earthquakes (for example Song *et al.* (2004)).

Teleseismic studies of mantle discontinuities usually compare a phase which has interacted with a discontinuity by reflection or mode conversion with a simpler arrival which has travelled a similar path. The comparison of a scattered arrival to a reference phase allows the characteristics of the scattered phase to be attributed to the discontinuity rather than other structures encountered

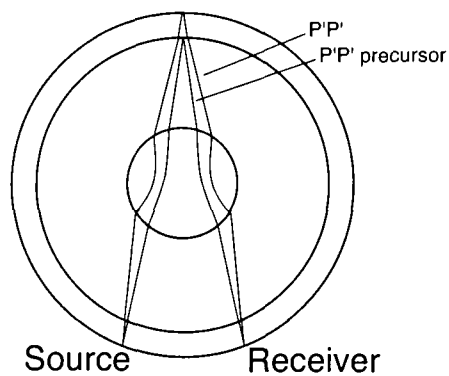


Figure 2.2: Rays paths for the $P'P'$ phase and a precursor which reflects at the surface, and a precursor phase which reflects off the underside of a mantle discontinuity.

between the source and the receiver. These should effect the scattered and reference arrivals equally. Methods studying reflected or converted phases can be further differentiated by the types of incident and scattered waves. In the remainder of this section the major types of incident and scattered waves will be discussed in turn. Throughout, it is important to remember that in all cases the amplitude of the scattered phase is very small. So to distinguish the scattered phases from the noise in teleseismic recordings seismologists employ averaging techniques (stacking) to enhance the signal to noise ratio and retrieve coherent information.

2.2.1 Precursors to $P'P'$

$P'P'$ is the shortened name of the core traversing phase PKPPKP. Figure 2.2 shows the $P'P'$ phase and a precursor which reflects off the underside of a mantle discontinuity. The phase is observed at high frequency and has a steep angle of incidence at the discontinuity. This makes it useful for determining the thickness of discontinuities. Benz & Vidale (1993) presented observations of a $P'P'$ precursor from the 410km discontinuity which suggested a discontinuity thickness of 4km or less.

The sensitivity of $P'P'$ precursors to a reflection point half way between the source and the receiver makes the techniques potentially useful for the global mapping of discontinuities. However, the requirement of short period array recordings and a narrow distance range for the observation of the $P'P'$ phase means that the technique is not as widely applied as the more frequently used

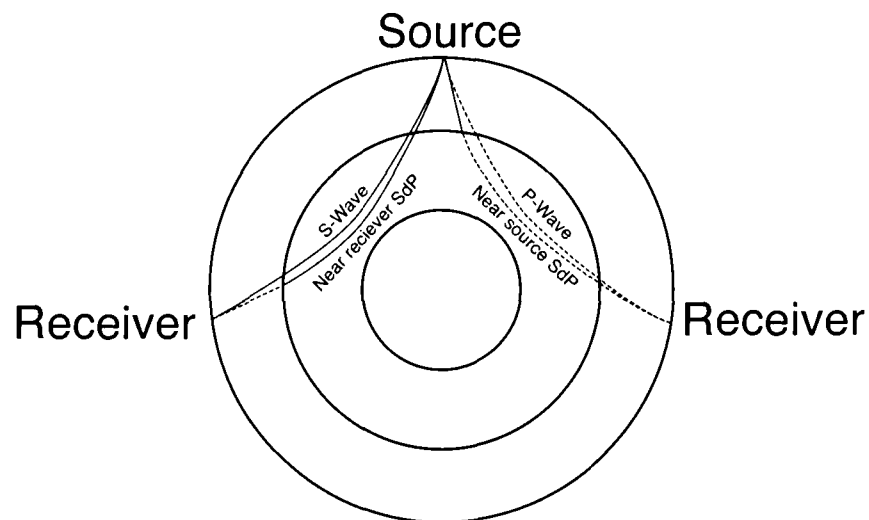


Figure 2.3: Ray paths for near source S to P conversion and a near receiver S to P conversion. The reference phases, the direct P and direct S, are also shown.

SS and *PP* precursors.

2.2.2 Converted phases

Figure 2.3 shows ray paths for direct P and S-waves, as well as phases which have been converted from P to S or vice versa at a mantle discontinuity. The most widely applied type of this data is in the form of receiver functions in which near receiver P to S conversions are separated from the coda following the P wave arrival.

Conversions from upper mantle and transition zone discontinuities are limited to regions near either the source or the receiver. For this reason the analysis of converted phases are of little use for the global mapping of upper mantle and transition zone discontinuities. However, converted phases have played an important part in the characterisation of discontinuities in regional settings near mantle plumes (Vinnik *et al.* (1997); Li *et al.* (2000b); Hooft *et al.* (2003)) and areas of subduction (Thirot *et al.* (1998); Li *et al.* (2000a); Lebdev *et al.* (2003)).

2.2.3 Near source underside reflections

Analysis of records from deep events have identified arrivals between the main *P* and the surface reflection phases *pP* and *sP* (figure 2.4). The requirement of a source in the transition zone

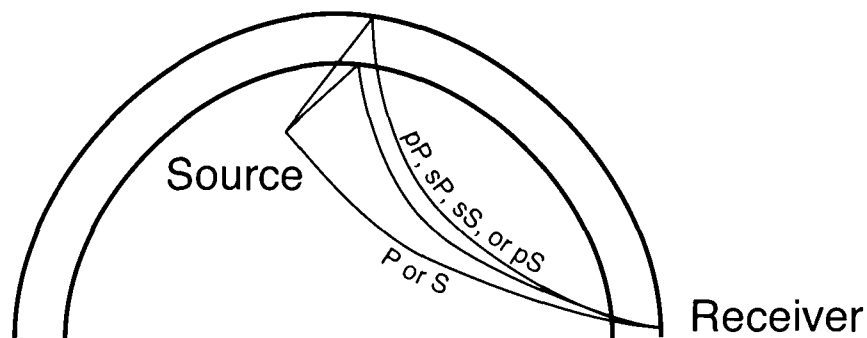


Figure 2.4: Rays for direct arrivals and near source surface and discontinuity reflections.

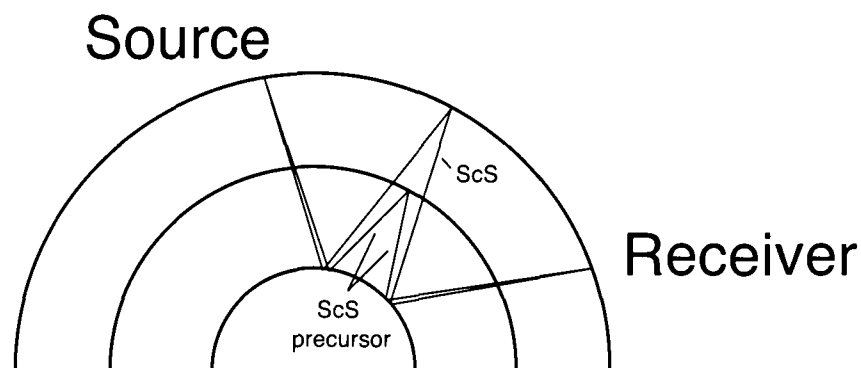


Figure 2.5: Rays for and ScS_2 reverberation (i.e. two ScS legs), and a precursor phase $ScSdScS$.

limits the usefulness of near source underside reflections to discontinuities in regions surrounding subducting slabs. However, the method potentially allows detailed mapping of upper mantle discontinuities in subduction zones and has been used to this effect by several authors (Vidale & Benz (1992); Collier & Helffrich (1997); Flanagan & Shearer (1998b)).

2.2.4 ScS reverberations

An ScS reverberation is an S-wave which reflects between the surface and the core-mantle boundary (figure 2.5). Arrivals in between the ScS reverberations are caused by reflections from the topside or underside of mantle discontinuities.

Revenaugh & Jordan (1991a,b,c,d) describe the details of this technique and its potential for mapping reflectors throughout the mantle. Revenaugh & Jordan (1991b) use the technique to describe a variety of upper mantle and transition zone reflectors by stacking ScS reverberations which fall within corridors between events occurring around the South and Western Pacific and

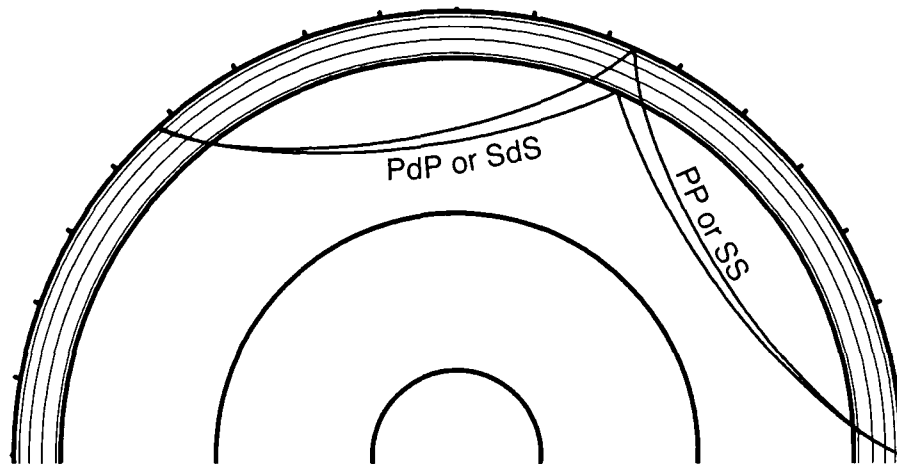


Figure 2.6: PP and SS phases with precursors PdP and SdS .

stations in Australia and South East Asia. Subsequent work by Revenaugh & Sipkin (1994a,b) applied the technique to corridors covering Northern China and Mongolia.

Although ScS reverberations are less spatially limited than previous techniques, the narrow distance interval ($10\text{-}50^\circ$) for which ScS reverberations are observed still limits their usefulness for the global mapping of discontinuities as it is impossible to retrieve structure beneath large ocean basins.

2.2.5 Precursors to SS and PP

Precursors to the SS and PP phases occur due to the underside reflections from mantle discontinuities (figure 2.6). These underside reflections (PdP and SdS , where d is reflection depth) from discontinuities provide, in principal, a powerful method for the study of upper mantle discontinuities on a global scale as the point of interaction of the seismic wave with the discontinuity is half way between the source and receiver (the bounce point). So the technique is less constrained, than other methods, by the uneven distribution of sources and receivers.

Neele *et al.* (1997) has shown that PdP and SdS phases have a complicated travel time surface near the bounce point. This can be seen in figure 2.7, where maps of the $S350S - SS$ travel time surface and the corresponding discontinuity depths are shown (figure adapted from Shearer *et al.* (1999)). The behaviour of PP precursors is similar. The mini-max behaviour of the SS phase produces a saddle shaped travel time surface where arrivals from off the great circle path can

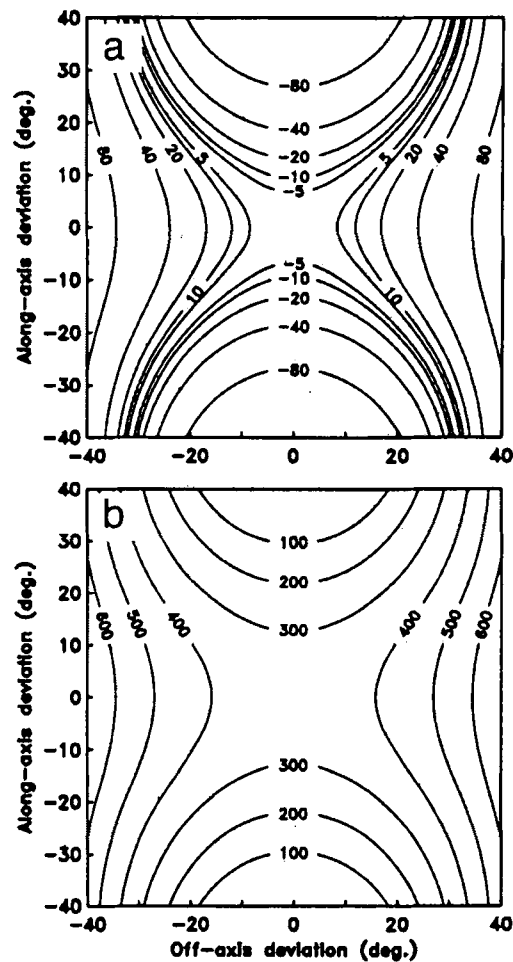


Figure 2.7: a) Contours of the travel time differences for $SdS - SS$. The centre (zero time) represents a symmetric $S350S - SS$ travel time at 140° . b) The corresponding depths for hypothetical point scatterers which can produce the $S350S - SS$ travel time at 140° . In each case the source receiver path runs vertically through the middle of the plot. Figure is adapted from Shearer *et al.* (1999).

arrive at the same time as a reflection from the mid-point. It can also be seen that the Fresnel zone, or maximum horizontal resolution (defined by the $\pm T/4$ contours), is about 10° wide for PP and SS waves at 10s and 20s respectively.

Neele *et al.* (1997) showed that mantle discontinuity topography narrower than the Fresnel zone can lead to errors in the determination of discontinuity depths. Zhao & Chevrot (2003) have also emphasised the sensitivity of precursor travel times to structure away from the bounce point. However, although a single observation of any $SdS - SS$ or $PdP - PP$ phase represents an average over the broad Fresnel zone, averaging of large numbers of records with similar bounce points reduces the contribution from structure away from the bounce points. So the technique still

remains the most effective method for large scale global mapping of upper mantle and transition zone discontinuities.

Much previous work has focussed on precursors to the SS phase (e.g. Shearer (1993); Gossler & Kind (1996); Flanagan & Shearer (1998a); Gu *et al.* (1998)). These studies have restricted discussion to long-wavelength transition zone discontinuity topography and its implications for lateral variations of transition zone temperatures. PP precursors have also been used to study the topography of the 410km discontinuity (Flanagan & Shearer (1999)), and constrain the properties of the 660km discontinuity (Estabrook & Kind (1996); Shearer & Flanagan (1999)).

Chapter 3

Summary of relevant mineral physics

3.1 Mantle mineralogy

There are two main petrological models for mantle composition, pyrolite and piclogite (shown in figure 3.1). Pyrolite was first proposed by Ringwood (1975) and is based on the composition of meteorites (CI Chondrites) and mantle xenoliths. The few potential minerals with elastic properties consistent with seismic data give a mantle which is dominated by olivine, pyroxene, and garnet. The pyrolite model features an upper mantle which is dominated by olivine compounds (65%). The alternative model (piclogite) was proposed by Anderson (1979). It is constrained by the comparisons of mineral physics data with seismic velocity profiles. It features a lower olivine content for the mantle, with larger amounts of garnet and clinopyroxene.

Of the two models the pyrolite mantle is more generally accepted. Although the piclogite model is constrained by seismological data it does not account for the contributions of minor elements and phases to seismic velocities and requires a high aluminium content and density for the mantle.

A major feature of both models is the polymorphic phase transformations of olivine, which

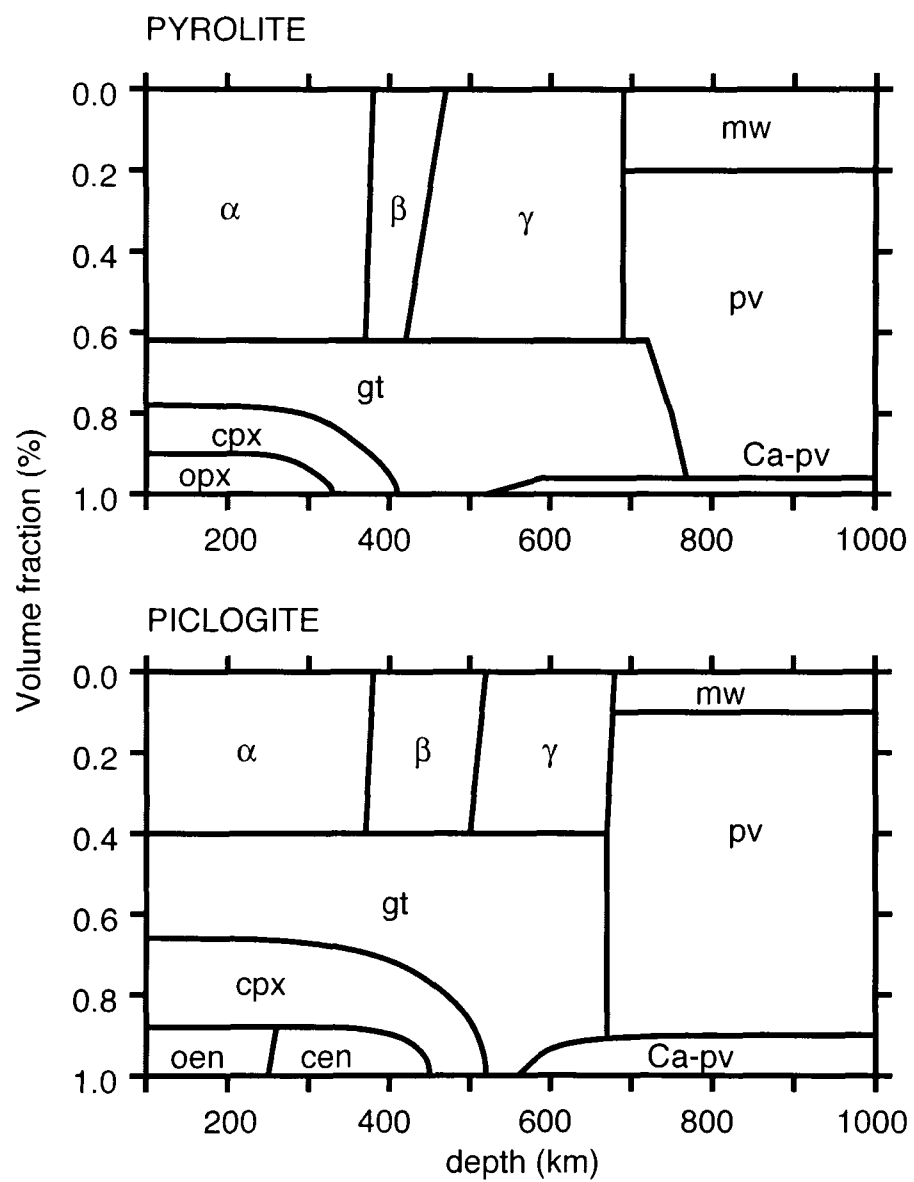


Figure 3.1: Pyrolite and piclogite compositions with depth in the mantle. The minerals are α -olivine, β -wadsleyite, γ -ringwoodite, mw-magnesianwusite, pv-perovskite, Ca-calcium perovskite, gt-garnet, cpx-clinopyroxene, opx-orthopyroxene, oen-ortho-enstatite, cen-clino-enstatite. Data is taken from Irifune & Ringwood (1987); Vacher *et al.* (1998); Fei & Bertka (1999).

occur as it changes to denser structures with increasing depth (figure 3.2). From the expected pressures of the olivine phase transformations at 1500-1600K, the depth of the transitions can be summarised as:

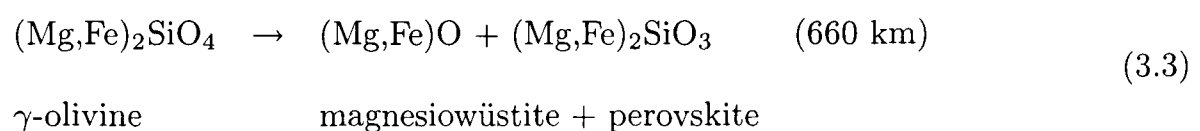
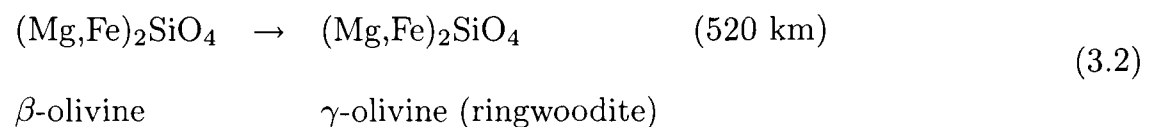
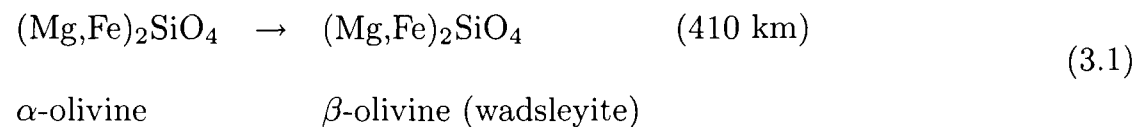


Figure 3.2 also shows some of the expected phase transitions in the non-olivine components (residuum mineralogy). Of particular importance is the formation of garnet from the pyroxene phase as this interacts with the olivine transitions via Fe-partitioning (for discussion see section 3.4 or Weidner & Wang (2000)).

3.2 Temperature dependence of the α – β olivine phase boundary

The pressure dependence of the olivine transformations would be expected to create a depth dependence of the discontinuities on mantle temperatures, which can be calculated from the Clapeyron slope (γ) of the phase transition and the mantle pressure gradient. The Clapeyron slope for a reaction is defined as:

$$\gamma = \frac{dP}{dT} = \frac{\Delta S}{\Delta V} \quad (3.4)$$

However, equation 3.4 does not formally hold for a solid solution (such as the Fe–Mg olivine system) as there is no single phase boundary but a region in which products and reactants can coexist (Helfrich (2000) and figure 3.3). Yet a two phase region will still produce reflections and conversions of seismic waves if the impedance gradient is steep enough. So, to characterise

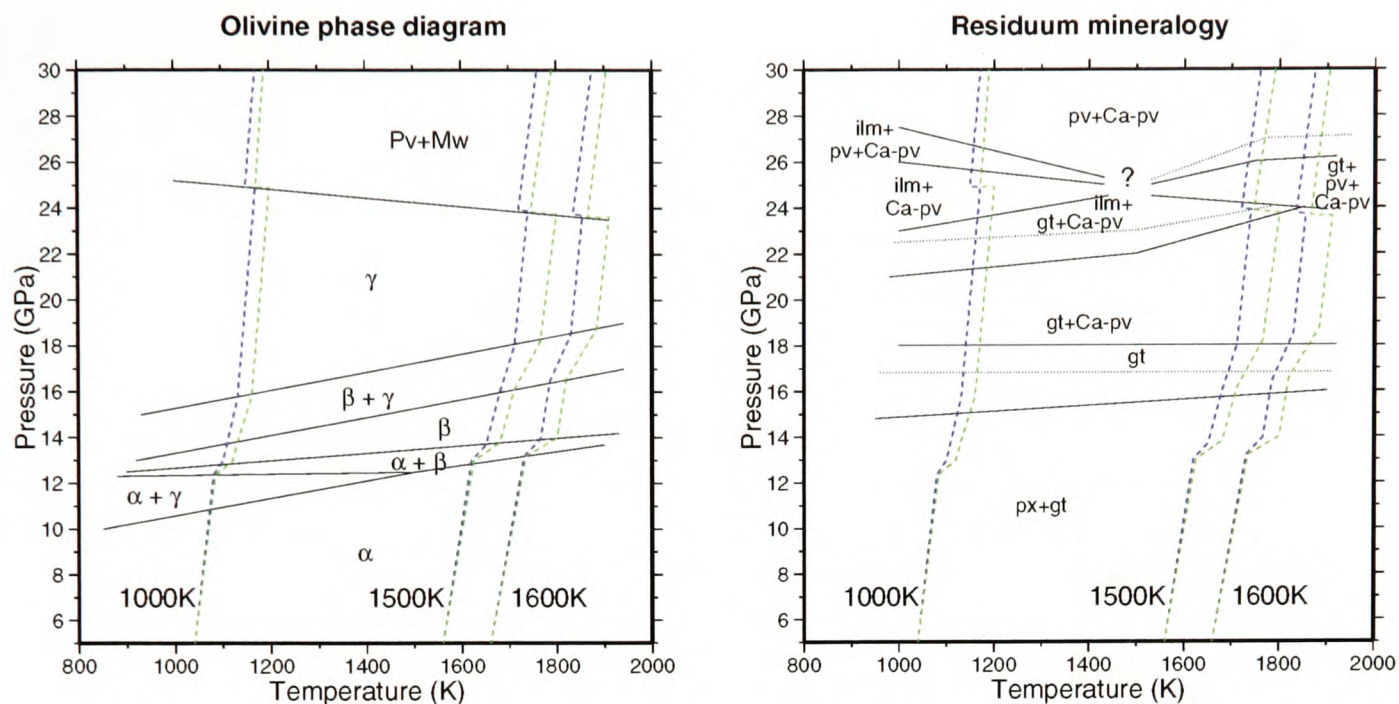


Figure 3.2: Phase diagrams of the olivine and residuum mineralogy (taken from Ita & Stixrude (1992) with alterations after Vacher *et al.* (1998)). Black lines represent phase boundaries. In the residuum mineralogy the solid lines are for pyrolite, the dotted lines are for a piclogite composition. The coloured lines show adiabats for 1000,1500, and 1600K with blue corresponding to piclogite and green to pyrolite. the mineral phases are labelled α - olivine, β -wadsleyite, γ -ringwoodite, mw-magnesiowustite, pv-perovskite, Ca-calcium perovskite, gt-garnet, px-pyroxene, ilm-Ilmenite.

	<i>PdP</i>	<i>SdS</i>
410km	3.15	3.21
520km	5.54	5.29
660km*	-2.0	-2.0

Table 3.1: Seismic Clapeyron slopes for the 410, 520km and 660km discontinuities. All slopes were taken from Helffrich (2000) and given in MPaK^{-1} , using waves with 20s period. *Note the 660km discontinuities sharpness means that the Clapeyron slope is the same for $P660P$ and $S660S$.

pressure dependence of an observable phase transition, Helffrich & Bina (1994) introduced a seismic Clapeyron slope that is dependent on the frequency and angle of the incident wave. Table 3.1 gives the seismic Clapeyron slopes for the olivine phase transformations observed using PdP and SdS waves with 20s period¹.

So, in principal, discontinuity topography provides seismologists with a tool for the measurement of lateral temperature variation in the mantle provided that the only cause for the impedance

¹For the frequency range used in this study the variation of seismic Clapeyron slope with incidence angle and frequency is not significant.

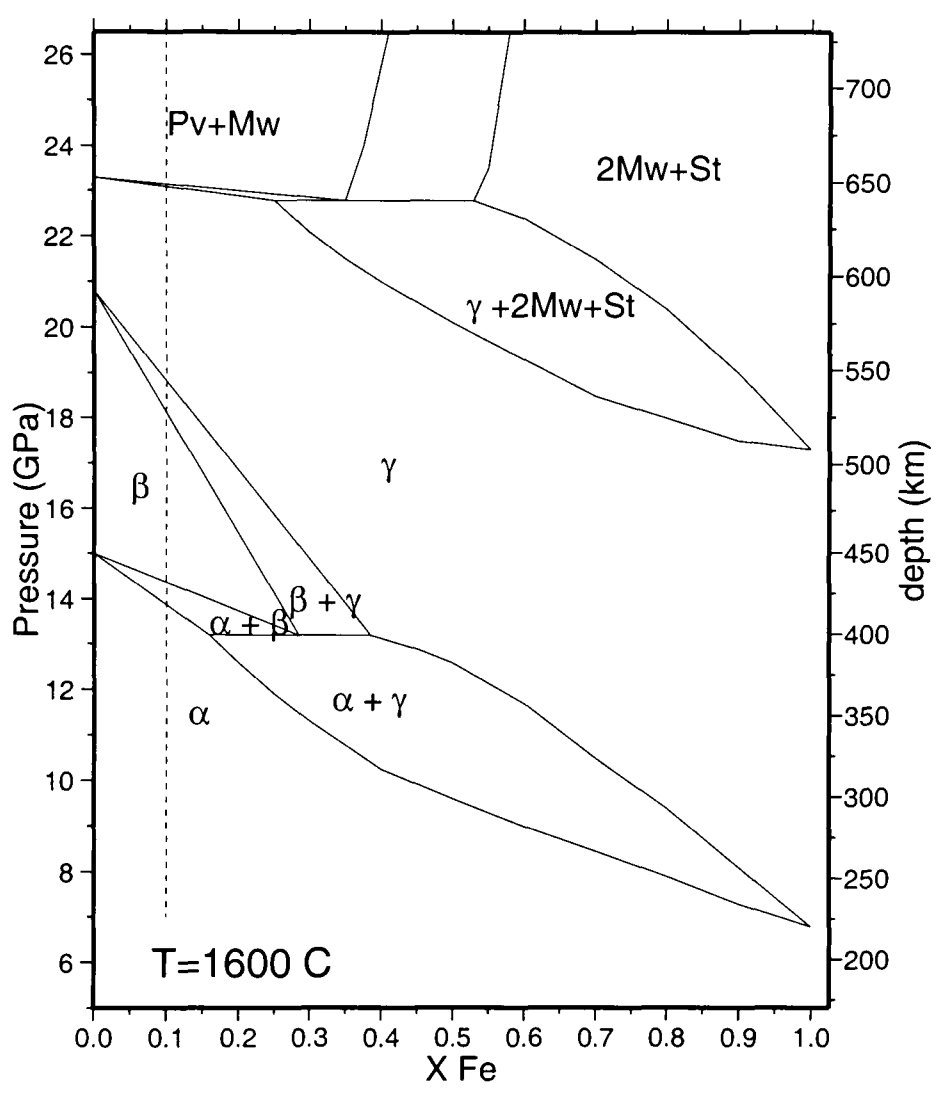


Figure 3.3: Phase diagram for olivine as a function of iron content (from Katsura & Ito (1989) and Fei & Bertka (1999)). The likely iron content of the mantle is around 10%, as shown by the dashed line. Phases are labelled α - olivine, β -wadsleyite, γ -ringwoodite, Mw-magnesianwüstite, Pv-perovskite, St-stishovite.

change is the $\alpha \rightarrow \beta$ -olivine phase transformation. However, this assumption could prove to be incorrect in regions of strong lateral heterogeneity in the mantle where mantle dynamics influence the phase boundary position.

The inference of mantle temperatures from discontinuity topography also assumes that the only factor in determining phases boundary deflections in an equilibrium mantle is the temperature. In fact recent results from mineral physics suggest that the minor components of mantle chemistry, particularly water and iron content, can also affect the phase boundary position.

3.3 Water and the α - β olivine phase boundary

The ability of transition zone minerals to hold significant amounts of water in their structures was first discussed by Smyth (1987). Wood (1995) also investigated the possibility that water preferentially fractionates into the wadsleyite compared to olivine by a factor of as much as 10:1. As a result, the stability field of β -wadsleyite should broaden in hydrous regions. This effect should broaden the phase transition region associated with the 410km discontinuity in hydrous regions. The observation of high frequency reflections from the 410km discontinuity (Vidale & Benz (1992)) limits the transition width to 4-10km, which in turn limits the maximum H₂O content of the mantle to about 200 parts per million by weight (Vidale & Benz (1992); Wood (1995)). However, this assumes that the width of the change in seismic properties is the same as the phase transition width. Helffrich & Wood (1996) have shown that if a non-linear gradient in seismic properties across the transition is taken into account a transition width of 10km can be seismically indistinguishable from a 5km wide phase transition. So it is possible that saturated β -wadsleyite could contain as much as 3% H₂O (by weight)² and the phase transition could widen to 40km (van der Meijde *et al.* (2003); Smyth & Frost (2002)).

Laboratory experiments have also shown that water can be accommodated into the γ -ringwoodite structure (Inoue *et al.* (1998); Smyth *et al.* (2004); Jacobsen *et al.* (2004)). So there is potential

²Compared with α -olivine which can incorporate 0.2% H₂O (by weight) into its structure

for the transition zone to be the largest reservoir of water in the Earth containing several times the volume of the worlds oceans.

The effects of variations in the mantle's water content have been investigated in numerous experimental studies. Yusa & Inoue (1997) examined compressibility of hydrous Mg-wadsleyite, finding that the addition of water can lower the bulk modulus by 5-11%. Smyth *et al.* (2004) and Jacobsen *et al.* (2004) have examined the seismic velocities and elastic moduli of iron bearing hydrous ringwoodite. They found that the effect of adding 1 wt.% H₂O to the ringwoodite structure reduced the bulk and shear moduli by 6 and 13% respectively. The resulting reductions in seismic impedance were 5.8% for P-waves and 7.6% for S-waves. There is currently a lack of experimental data for the effects of hydration on bulk moduli and seismic impedances of wadsleyite. However, assuming that the trend for wadsleyite is similar to that for ringwoodite, it might be expected that lateral variations in the mantle water content could be as important as temperature in determining seismic velocity in the mantle transition zone.

The presence of water in the mantle can also effect the pressures at which phase transformations take place (figure 3.4). Although Helffrich & Wood (1996) stress that the α - β phase transition pressure is more sensitive to temperature than water, in regions away from steep thermal gradients the effect of water on the phase boundary position could be significant. Smyth & Frost (2002) suggest that the addition of 3 wt.% water lowers the pressure of the α - β transformation by ~ 1 GPa.

3.4 Other chemical variations and the α - β olivine phase boundary

Variations in the amount of iron and aluminium in the mantle can also effect the width and pressure of transformation of the olivine phases. Weidner & Wang (2000) emphasise that the reactions which form wadsleyite from α -olivine, and garnet from pyroxene are coupled because iron preferentially partitions into a garnet phase rather than wadsleyite. So a mantle with a larger

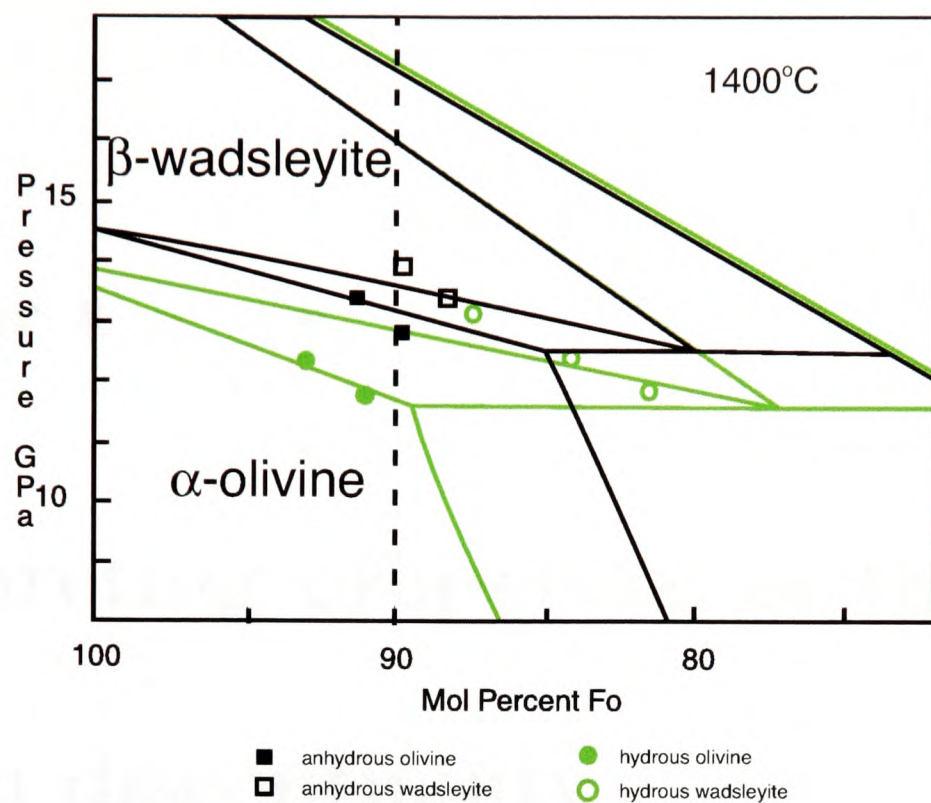


Figure 3.4: Effect of H_2O on the olivine phase diagram adapted from Smyth & Frost (2002). It shows that for a hydrous olivine (green) the α - β phase transition occurs at pressures 1.0 ± 0.5 GPa lower than for anhydrous olivine (black). The addition of H_2O also broadens the phase transition.

aluminium content will allow the formation of more garnet. This results in Fe-scavenging from the olivine-wadsleyite transformation, and a lower pressure narrower α - β transformation (3.3).

The effect of Fe-content on the densities and seismic velocities of transition zone minerals is not believed to be as significant as that of water (Jacobsen *et al.* (2004)). However, additional aluminium content does increase seismic velocities. This is mostly due to the formation of garnet over a range of depths and does not effect the changes in seismic properties across the 410km discontinuity (Weidner & Wang (2000)). The effect of Fe-content in the olivine transformations widths and pressures can be seen from figure 3.3.

Chapter 4

Interpreting properties of the 410km discontinuity

4.1 Motivation

In the previous sections I outlined the ways in which seismologists observe mantle discontinuities and the expected properties of the transition zone discontinuities. The seismologically observable characteristics of the 410km discontinuity are the amplitudes and arrival times of scattered waves, which in turn constrain the discontinuity's thickness or impedance contrast, and depth. Here, the purpose is to describe how seismologically derived information on the 410km discontinuity can be used with results from mineral physics to better constrain the properties of the discontinuity and gain insight into mantle processes.

4.2 Amplitude of 410 reflections: Thickness vs impedance contrast

Reflection amplitudes are influenced by many factors including, discontinuity thickness and impedance contrast, intrinsic attenuation, mantle structure, and anisotropy. The goal of any study of reflection amplitudes is to minimise the contribution of the latter three allowing interpretation in terms of discontinuity properties. Variations in reflection amplitudes observed with short period data have been primarily interpreted as changes in discontinuity thickness, while variations observed using long period data are more likely to be due to changes in impedance contrast.

Vidale & Benz (1992) observed a high frequency reflection from the 410km discontinuity as a precursor to the $P'P'$ phase. This restricted the possible width of the phase transition to ~ 4 km as reflected phases are only sensitive to impedance gradients shorter than one quarter wavelength. Rost & Weber (2002) studied transition zone discontinuities beneath the Western Pacific, finding the 410km discontinuity to be narrower than 6km with a minimum impedance contrast of 6.5%. Recently, van der Meijde *et al.* (2003) examined P to S converted phases (receiver functions) from the 410 under Europe finding the width of the discontinuity to be as much as 25-35km.

Measurements of discontinuity thickness are generally derived by comparison of the frequency dependence of observed amplitudes with those predicted using a linear impedance gradient. However, Helffrich & Wood (1996) have shown that 5km linear impedance gradients can yield a reflection coefficient the same as that for a 10km width transition that has a non-linear impedance gradient. Since it seems more likely that the impedance gradient is non-linear within the $\alpha - \beta$ olivine phase transition (Stixrude (1997)), this would suggest that the above estimates of discontinuity thickness are minima. The important point is that variation in reflected amplitudes observed in short-period data could be due to variation in discontinuity thickness, which is probably in the range 4-30km.

Variation in phase transition thickness is generally attributed to variations in mantle water content (Helffrich & Wood (1996); Helffrich (2000); van der Meijde *et al.* (2003)). As water

preferentially partitions into the wadsleyite phase by about 10:1 (Wood (1995)), the wadsleyite phase is stable at lower pressures in regions where there is more water. This results in a broadening of the region where both phases are stable in hydrous regions of the mantle.

Until now, the only long period observations of 410-reflection variability are those derived from *ScS* reverberations (Revenaugh & Jordan (1991b); Revenaugh & Sipkin (1994a); Gaherty *et al.* (1999)). These are in many ways similar to the *S410S* reflections studied here. However, observation is restricted to regional corridors between sources and receivers. Revenaugh & Sipkin (1994a,b) examined *ScS* reverberations between events in Japan and stations in China. They found the 410km reflection coefficient for S-waves to be increased under North Western China and Mongolia. These long period *ScS* studies utilised data with a cut off frequency of 25mHz, which equates to a minimum effective wavelength of ~ 194 km at 410km depth. This suggests that the reflection amplitude variations observed by Revenaugh & Sipkin (1994a) are more likely to be due to changes in the impedance contrast of the discontinuity, rather than variations in discontinuity thickness.

4.3 Topography of the 410: Chemical vs thermal heterogeneity

Traditionally the topography of the 410km discontinuity has been interpreted in terms of lateral variations in mantle temperature. Given the thermodynamic properties of the α - β olivine phase transition in section 3.2 a depressed discontinuity is expected in warmer regions of the mantle and an elevated discontinuity is expected in colder regions in accordance with a positive Clapeyron slope for the $\alpha - \beta$ -olivine transformation. So, in principle, topography of the 410km discontinuity provides seismologists with a tool for measuring lateral temperature variation in the mantle.

However, other factors effect the position of the $\alpha - \beta$ olivine phase boundary. Recent results from mineral physics (Wood (1995); Chen *et al.* (2002); Smyth & Frost (2002)) suggest that the presence of water in olivine-wadsleyite causes the transformation to occur at lower pressures

(section 3.3, figure 3.4). The presence of other chemical variations can also effect the depth of the 410km discontinuity. Weidner & Wang (2000) predict that a 14% decrease in Fe content would result in 3-4km change in discontinuity depth. The presence of aluminium effects the 410km discontinuity depth by allowing the formation of a garnet phase, into which iron preferentially partitions.

The contributions of chemical variations (water, aluminium, and iron) to the topography of the 410km discontinuity are expected to be secondary to that of temperature in regions of large temperature contrasts. For example under hydrous conditions the $\alpha - \beta$ olivine transformation occurs at pressures approximately 1 GPa lower than for anhydrous conditions (Smyth & Frost (2002)), which translates to a discontinuity deflection of around 30km. In contrast a 1000K temperature anomaly, consistent with a subducting slab, would be expected to uplift the discontinuity by up to 90km (Helffrich (2000)). Many previous studies have observed discontinuity deflection consistent with cold subducting slabs in the mantle (Vidale & Benz (1992); Collier & Helffrich (1997); Thiriot *et al.* (1998); Collier *et al.* (2001)). These elevations of the 410km discontinuity are up to 60km in magnitude but only 200km in width. However, the large Fresnel zones of $P410P$ and $S410S$ waves means that discontinuity topography this narrow will be averaged over > 1000 km (Neele *et al.* (1997); Shearer *et al.* (1999)). So the pattern of precursor topography will not retrieve the expected deflections associated with narrow zones of up-welling and down-welling expected in a convecting mantle. Instead, discontinuity topography derived from precursor phases reflects the long wavelength variations in mantle temperature and other chemical variations.

It is possible to examine the contributions of thermal and chemical variations to the topography of the discontinuity by examining the relationship between discontinuity topography and transition zone seismic velocities. If the major cause of precursor topography is lateral temperature variations then we would expect a negative correlation between transition zone seismic velocities, in accordance with the negative Clapeyron slope for the $\alpha - \beta$ transformation. However, the presence of water (or other chemical variations) in the transition zone lowers seismic velocities (Jacobsen *et al.* (2004); Weidner & Wang (2000)), and would be expected to result in an elevated disconti-

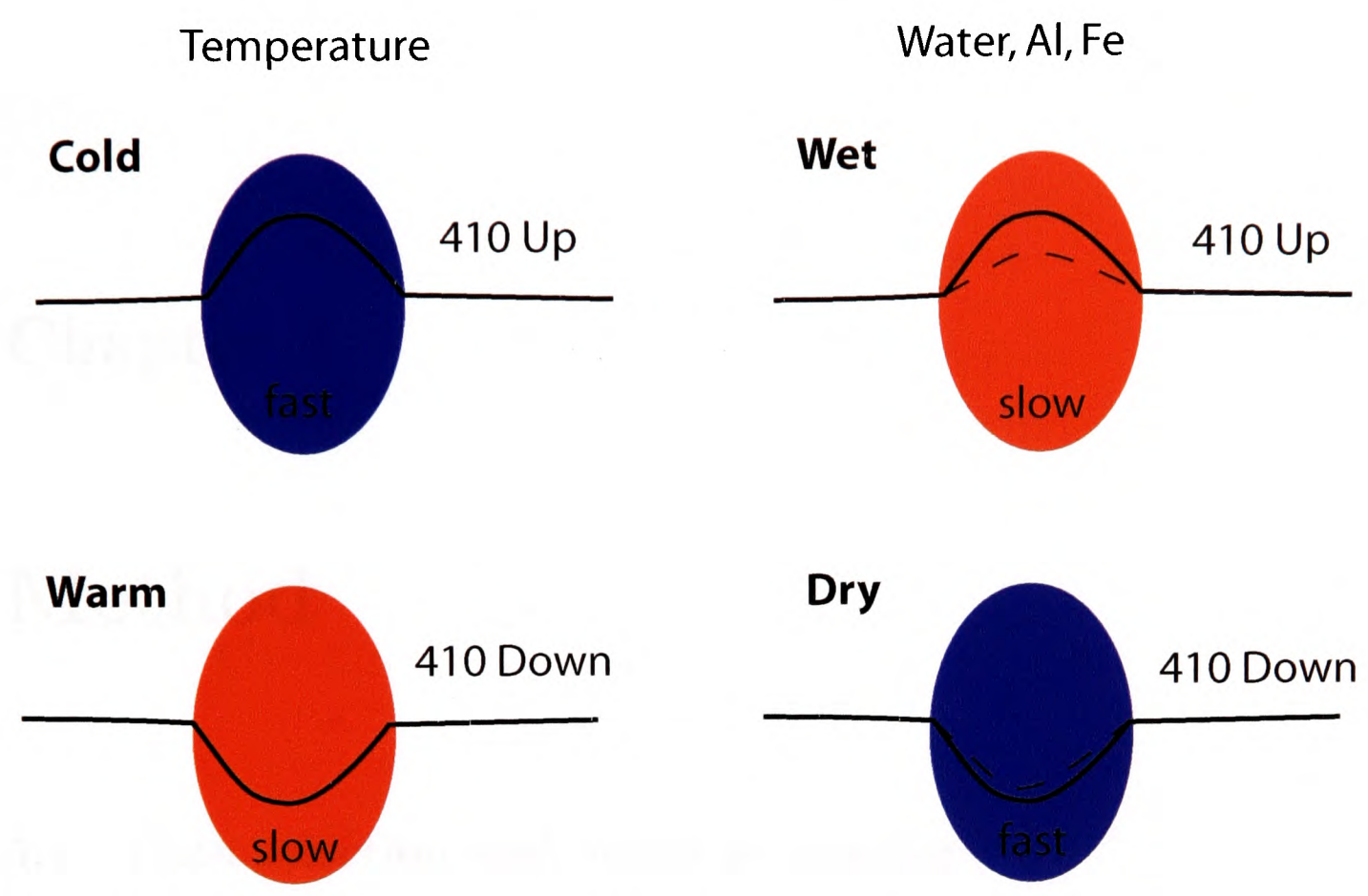


Figure 4.1: The expected response of the 410km discontinuity and seismic velocity perturbations for regions of chemical or thermal anomaly.

nuity (Smyth & Frost (2002)), resulting in a positive correlation between discontinuity topography and seismic velocity anomaly. Figure 4.1 illustrates how the effects of chemical heterogeneity and temperature on the position of the $\alpha - \beta$ olivine phase boundary and seismic velocities can be separated if both discontinuity topography and seismic velocity are taken into account.

Chapter 5

Method

5.1 Data selection and initial processing

This study used broad band data sampled at 1s from the IRIS/IDA/USGS network. The selection criteria is shown in table 5.1. The epicentral distance intervals were chosen to maximise underside reflections from the upper mantle and transition zone discontinuities while minimising the interference from other phases. Figure 5.1 shows the distribution of epicentral distances, azimuths (at the source-receiver midpoint), and event depths for our final *PP* and *SS* datasets. Figure 5.2 shows the global distribution of *PP* and *SS* source-receiver mid-points (bounce points). Stack quality, and hence reliability, is largely dependent on the number of traces in the stack. Figure 5.2 shows that data coverage is best for both datasets in the North Pacific and for the *PP* data this extends into China and Northern Asia. The *SS* dataset is less extensive than that for *PP*, largely due to the restriction of event size.

The data was band-pass filtered between 8s and 75s for *PP*, and 15s and 75s for the *SS*. In addition a cosine low pass filter was applied to cut off high frequency noise below 4s and 8s in *PP* and *SS* data respectively. The records were also deconvolved for long period instrument response. A Hilbert transform was applied to the *PP* data, and the *SS* data was rotated to the transverse component.

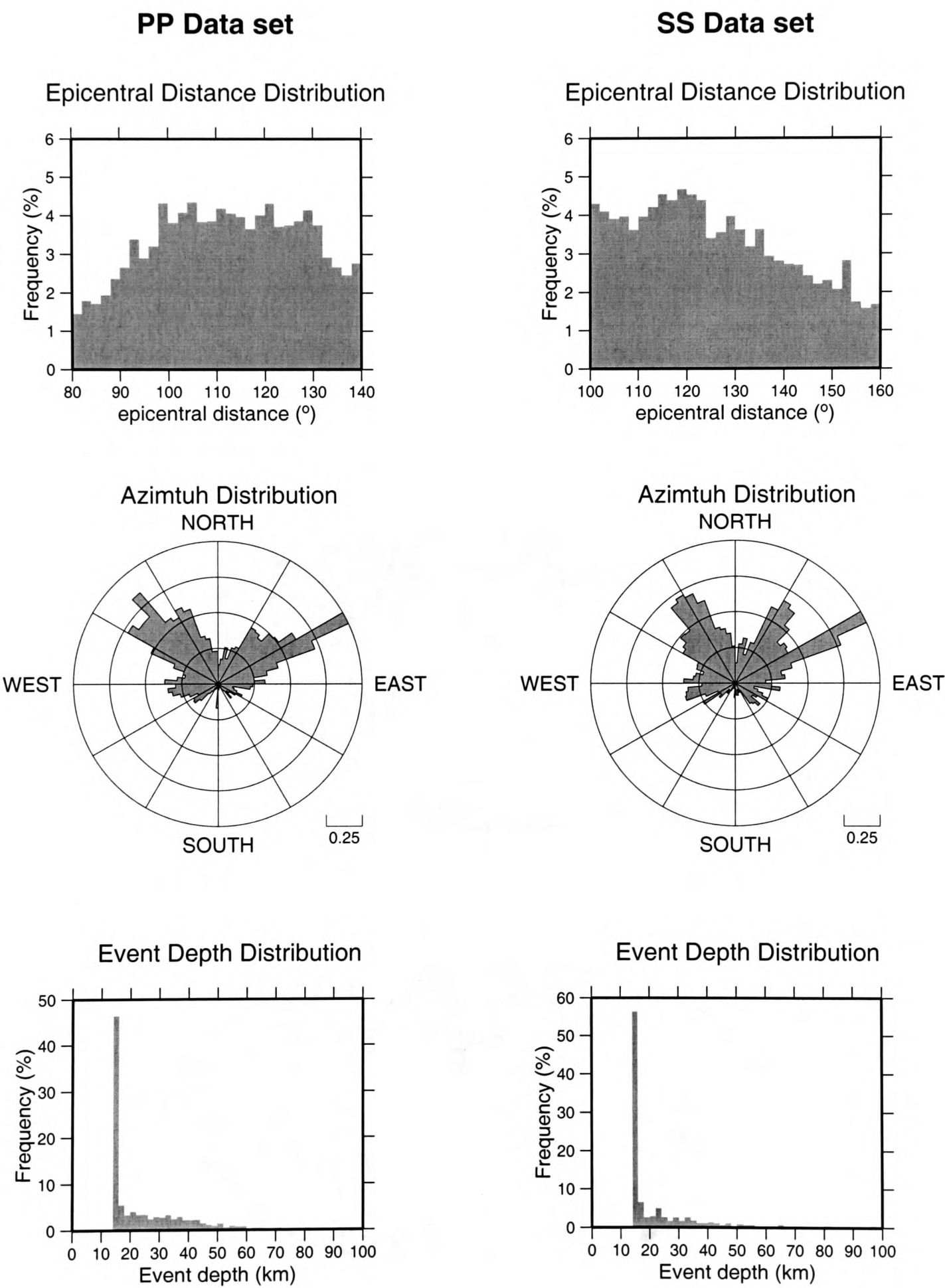


Figure 5.1: Epicentral distance, azimuth and event depth distributions for *PP* (left) and *SS* (right) datasets.

Dataset (reference phase)	PP	SS
Components	Vertical	Horizontal
Event date range	1990-1999	1990-1999
Event Size (m_w)	5.5-6.5	6.0-7.0
No. Events	2425	796
Start-finish time (s)	-270 to +90	-700 to +300
Epicentral distance ($^\circ$)	80-140	100-160

Table 5.1: Data selection parameters for the PP and SS datasets. The Start-finish were taken relative to the expected arrival times of the reference phase in PREM.

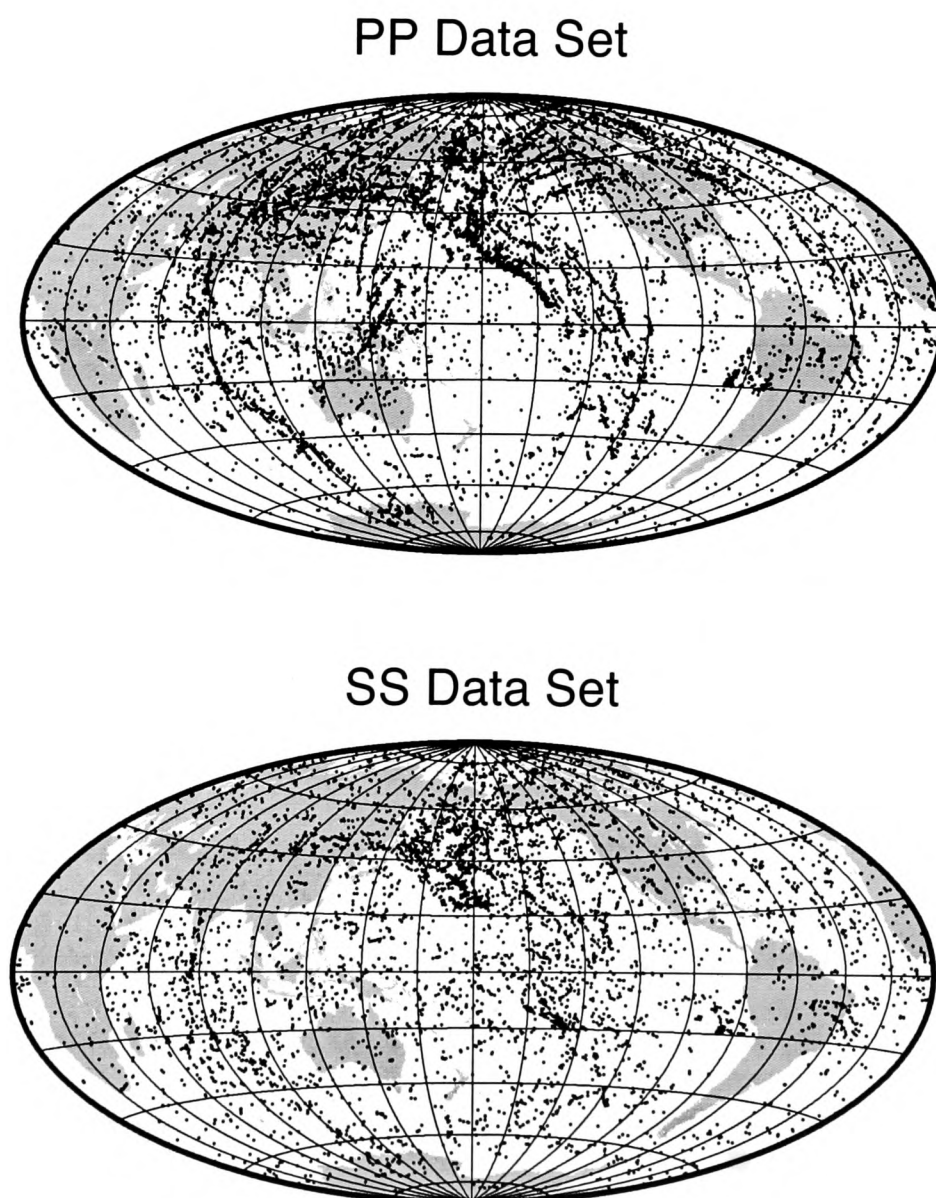


Figure 5.2: Locations of the bounce points (source-receiver mid-points) for the traces used in the PP and SS datasets.

5.2 Stacking procedure

Arrivals with amplitudes below the noise levels of individual records were enhanced by averaging large numbers of traces in the time-slowness domain. Except for the use of an automatic picking algorithm (section 5.3), and removal of ocean function (section 5.4), the initial (pre-deconvolution) stacking procedure is similar to previous precursor studies (Flanagan & Shearer (1998a); Gu *et al.* (1998); Flanagan & Shearer (1999); Deuss & Woodhouse (2002)). A summary of the stacking procedure is as follows:

1. Initial *PP* and *SS* reference pulses for use in the automatic picking algorithm were determined by stacking ~ 1000 hand picked data. The bounce point, epicentral distance, and event depth distributions for the traces used to make the reference pulses is shown in figure 5.3
2. Data were selected by cross-correlating the filtered seismograms with the average reference pulses (see following section). The working datasets were selected using only data with a correlation coefficient ≥ 0.6 , and signal to noise ratio less than 3 (section 5.3). The signal to noise ratio was measured as the ratio of energy in a time window (60 seconds for *PP*, 100 seconds for *SS*) around the surface reflection to that prior to the surface reflection. For a more detailed discussion of the automatic picking algorithm see section 5.3.
3. The arrival time of the surface reflection was measured for each remaining trace using the absolute maxima of the cross-correlation function within time windows (16 seconds for *PP*, 32 seconds for *SS*) around the expected arrival times from PREM.
4. Prior to stacking, the data are normalised by the energy of the surface reflection in order to stop large events dominating the stack. The stacks were weighted by the record's signal to noise ratio. The sign of the cross-correlation maximum was used to determine the traces polarity. Traces with negative polarity were 'flipped' prior to stacking.
5. Expected 410 reflection amplitudes were determined using WKBJ synthetic seismograms (Chapman (1978)), with PREM. The synthetics were filtered using the same functions as

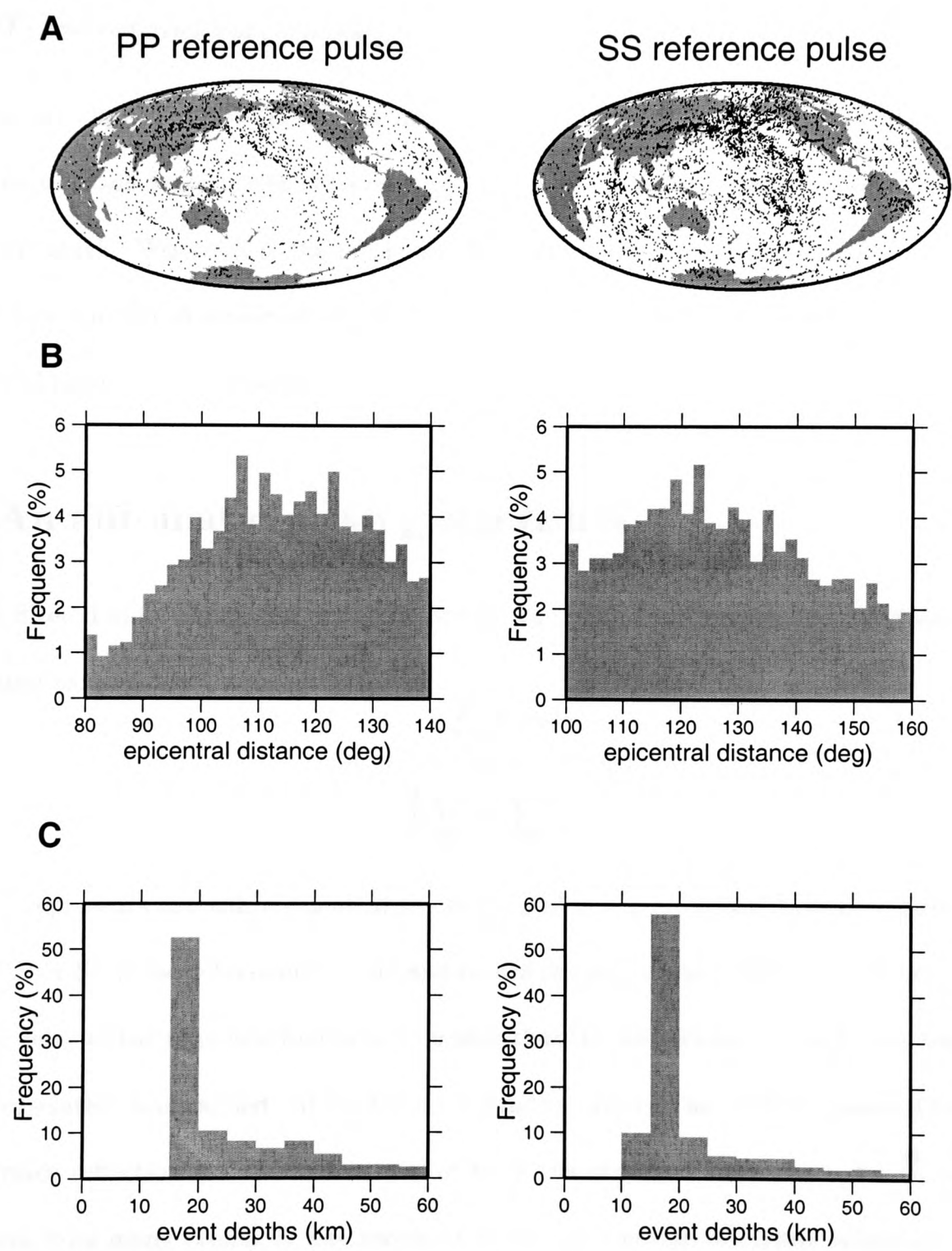


Figure 5.3: Bounce point, epicentral distance and event depth distributions for traces in *PP* and *SS* reference pulses.

the data. The expected 410 reflection amplitudes were used to correct the precursor section of each trace for epicentral distance effects prior to stacking.

6. For the *PP* data with oceanic bounce points the effect of the ocean (ocean response function, ORF) was removed from the reference phase section of the trace (see section 5.4).
7. Data are aligned on the arrival time of the surface reflection and stacked in the time slowness domain. Each trace in the stack is time shifted by $\Delta - \Delta_r$ times the slowness, and added to the stack. Where Δ is the epicentral distance, and Δ_r is a reference distance (110° for PP 130° for SS). A summary stacked trace is made by taking the profile along the expected arrival time and slowness path of precursor phases in PREM.

5.3 An automatic picking algorithm

From the filtered data usable traces were selected based on a cross-correlation function (r_i) which is calculated as:

$$r_i = \frac{\sum_n x_n s_{n+i}}{\left(\sum_n x_n^2 \sum_n s_n^2 \right)^{\frac{1}{2}}} \quad (5.1)$$

Where x_n represents the n th element of the data, and s is a reference function (in this case an average *PP* or *SS* pulse determined from hand picked records). The function r_i will be a maximum where the delay of the reference function (i) is such that the waveforms of s and x are best aligned.

The correlation was carried out for the data window around the PREM expected arrival time of the surface reflection and limited to within 8s of the expected time. In order to successfully stack traces from many events, it was necessary to change the sign of certain traces depending on the event mechanism. This was done where the maximum cross-correlation function between the data and reference pulse is negative.

The quality and quantity of the data used in the stack is varied by only using traces with a maximum cross-correlation greater than a certain threshold. In this study, unless otherwise stated, the threshold was taken as 0.6 which meant that 10627 and 6138 records were selected for the *PP*

Tolerance	PP	SS
Original No.	67872	12800
0.5	17030	8523
0.6	10627	6138
0.7	7441	4052
0.8	2538	1369
0.9	148	104

Table 5.2: Number of records selected from *PP* and *SS* datasets using the automatic picking algorithm for different cross correlation thresholds for the cross correlation. Only records with signal to noise ratio greater than 3 were used. The much larger number of traces in the *PP* dataset reflects the inclusion of data from smaller events.

and *SS* datasets respectively. Table 5.2 shows the number of records selected using the automatic picking algorithm for different cross correlation thresholds.

Signal to noise ratio is also incorporated into the into the picking algorithm. This is measured as the ratio of the energy of the trace in time windows around and prior to the surface reflection phase. No traces with a signal to noise ratio less than 3 were selected for stacking.

Figure 5.4 shows a stack made using the 2628 hand picked and aligned traces and a stack made using a similar number of traces selected using the automatic picking algorithm. The stack of the automatically picked data contains more coherent energy in the region between -80 and -30s where signals from upper mantle discontinuities would be expected. However, these differences are fairly minor.

5.4 Removal of the ocean function

The *PP* reference pulse is made up of reflections from the ocean bottom and ocean surface. This results in a broadening of the pulse relative to that for continents (figure 5.5). Accordingly we define an Ocean Response Function (ORF) which results in the broadening of the *PP* pulse.

In figure 5.5 an average ORF for our *PP* dataset is retrieved by deconvolving an average *PP* pulse obtained from traces with oceanic bounce points using a *PP* pulse obtained from a stack of traces with continental bounce points. The resulting ORF contains two main spikes, a main *PP* spike (amplitude ~ 0.95) due to a sea bed reflection followed by a smaller spike (amplitude ~ 0.25)

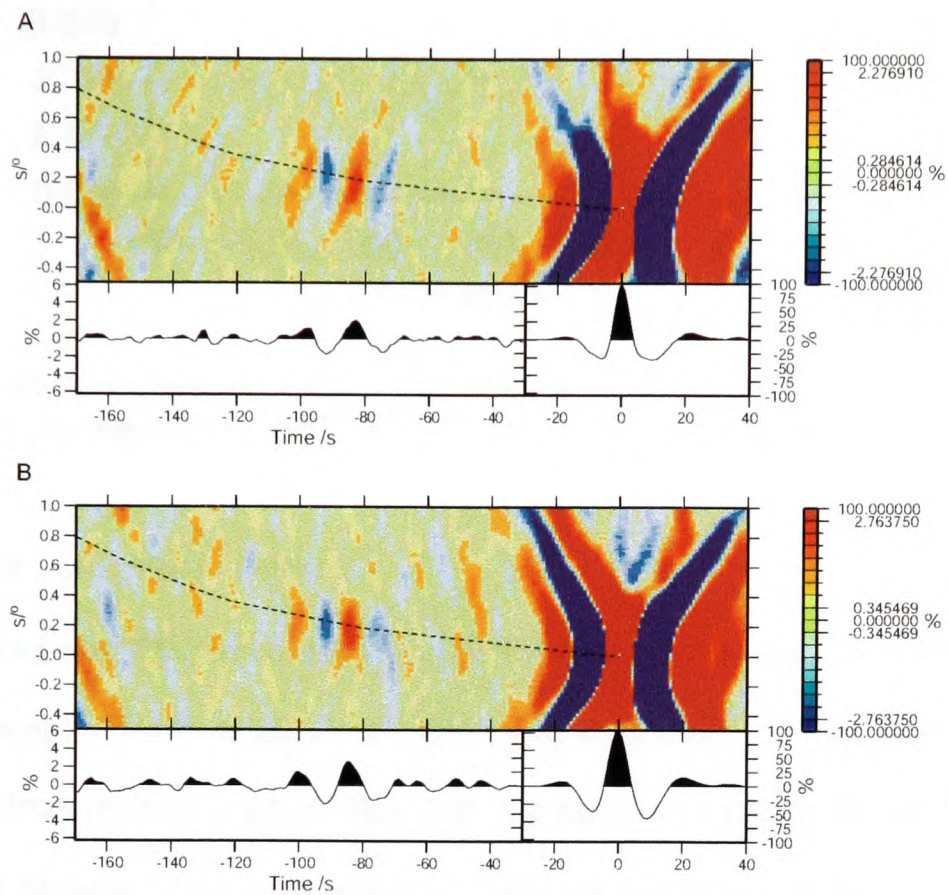


Figure 5.4: Comparison of stacks made using hand picked vs automatically picked records. **(A)** 2628 hand picked and aligned records. **(B)** 2772 records selected using the automatic picking algorithm (from a reduced PP dataset) with a tolerance of $r_{max} = 0.7$. Both stacks were made in the same way except that in the case of the hand picked data the data was aligned on the pick time and normalised to the maximum of the PP signal prior to stacking.

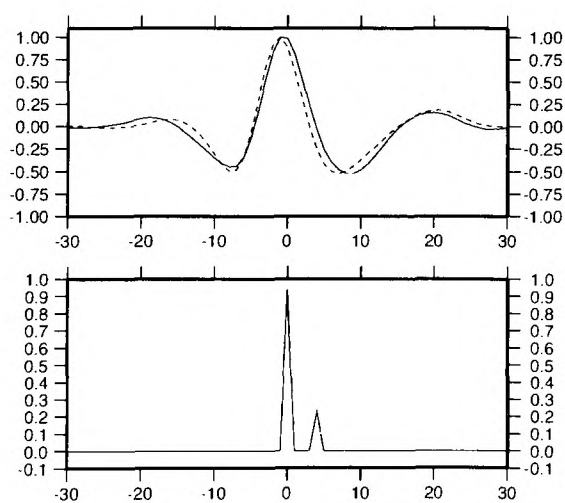


Figure 5.5: Top: Oceanic (solid) and continental (dashed) PP reference pulses. Bottom: The oceanic reference pulse has been deconvolved by the continental reference pulse (chapter 7 and appendix A).

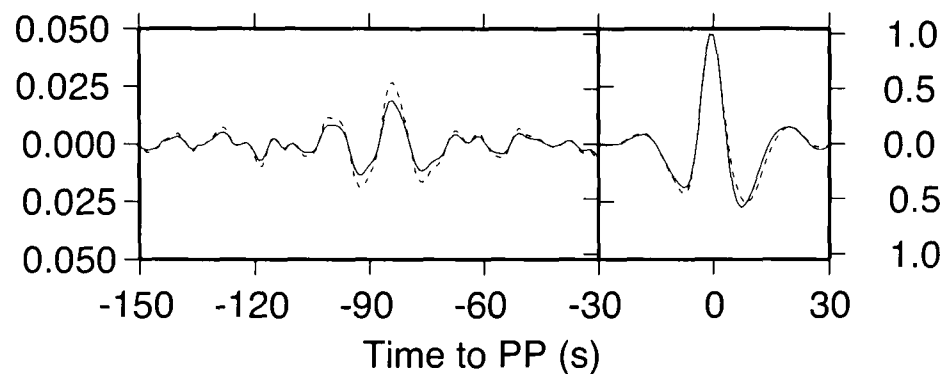


Figure 5.6: The effect of removal of the ocean function on a stack of traces with oceanic bounce points. Dashed = uncorrected for ORF, Solid = corrected for ORF.

4s later due to the sea surface reflection.

Prior to stacking the ORF is removed from the PP dataset. For each trace an ORF is estimated using the spike amplitudes from figure 5.5, and spike separation is determined using the ocean depth from ETOPO5 (NOAA (1988)). The ORF is then deconvolved from the PP section of the trace by spectral division. Figure 5.6 shows the effect of the deconvolution on a stack of traces with oceanic bounce points.

Crustal reverberations near to the bounce point also distort the PP and SS reference pulses. Figure 5.7 demonstrates this by showing two synthetic seismograms made using the WKBJ algorithm. One contains just PP and $P410P$ phases while the other also contains crustal phases. Because the trace is normalised to the amplitude of the PP pulse there is a slight difference in the $P410P$ wavelets. The bottom of figure 5.7 shows the traces deconvolved using the PP pulse (as per chapter 7). There is little or no difference in the retrieved amplitudes of the $P410P$ pulses ($\sim 0.2\%$). This suggests that the effect of the crust on the retrieval of $P410P$ and $S410S$ amplitudes is minor and so it is omitted from data processing.

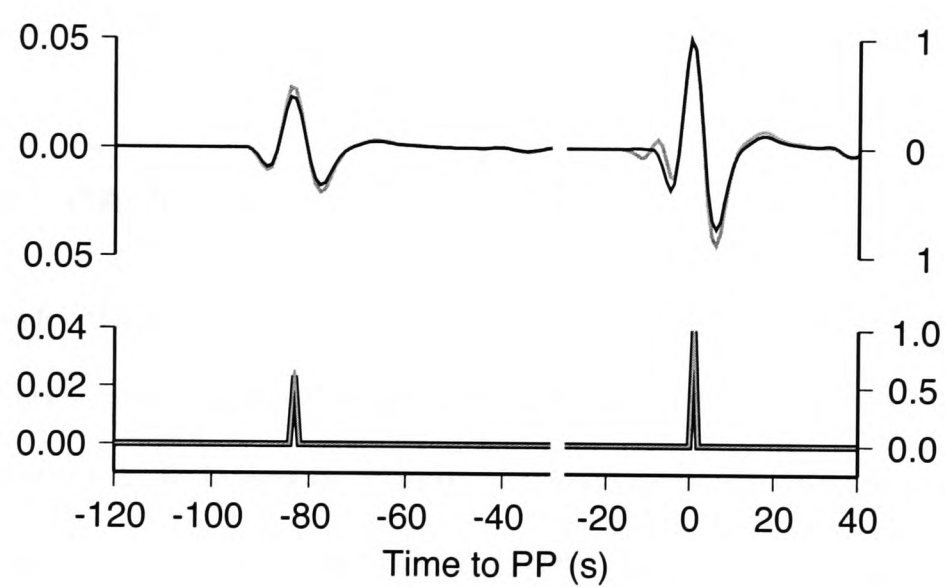


Figure 5.7: **Top** Effect of including the crust on a WKBJ synthetic stack trace containing PP and $P410P$ phases. The solid black trace includes only the PP and $P410P$ phases. The light grey trace also includes $P15P$, $P25P$ and $P80P$. **Bottom** results of deconvolution applied to the traces. There is little or no difference in the retrieved amplitudes of the $P410P$ pulses.

Chapter 6

Initial Stack Results

6.1 Global stacks

In figure 6.1 stacks for the global datasets are shown alongside stacks made using synthetic seismograms. The synthetics were computed using the WKB algorithm (Chapman (1978)) with Earth model PREM (Dziewonski & Anderson (1981)). The synthetics use the same source-receiver geometries as the original data but only use underside reflections from the surface and discontinuities at 220, 410 and 660km depth.

The 95% confidence limits for the stack traces are also given (dashed lines in figure 6.1). These were determined using bootstrap resampling, where by repeated stacks were made 500 times using randomly selected traces, selected with replacement.

The arrival 240s prior to the *SS* phases is consistent with a S660S reflection. However, no P660P phase is seen. The synthetic stack for P-wave reflections suggests that a P660P phase should be observable using the source-receiver geometries of our dataset.

The arrivals around 100s and 180s prior to the *PP* and *SS* phases are consistent with a global discontinuity at 520km depth. Another possible interpretation of these phases is side lobes of the larger arrivals at -80s and -160s. However, comparison with the reference pulse shows that this feature is much larger (nearly half the height of the main reflection) and so is unlikely to be a side

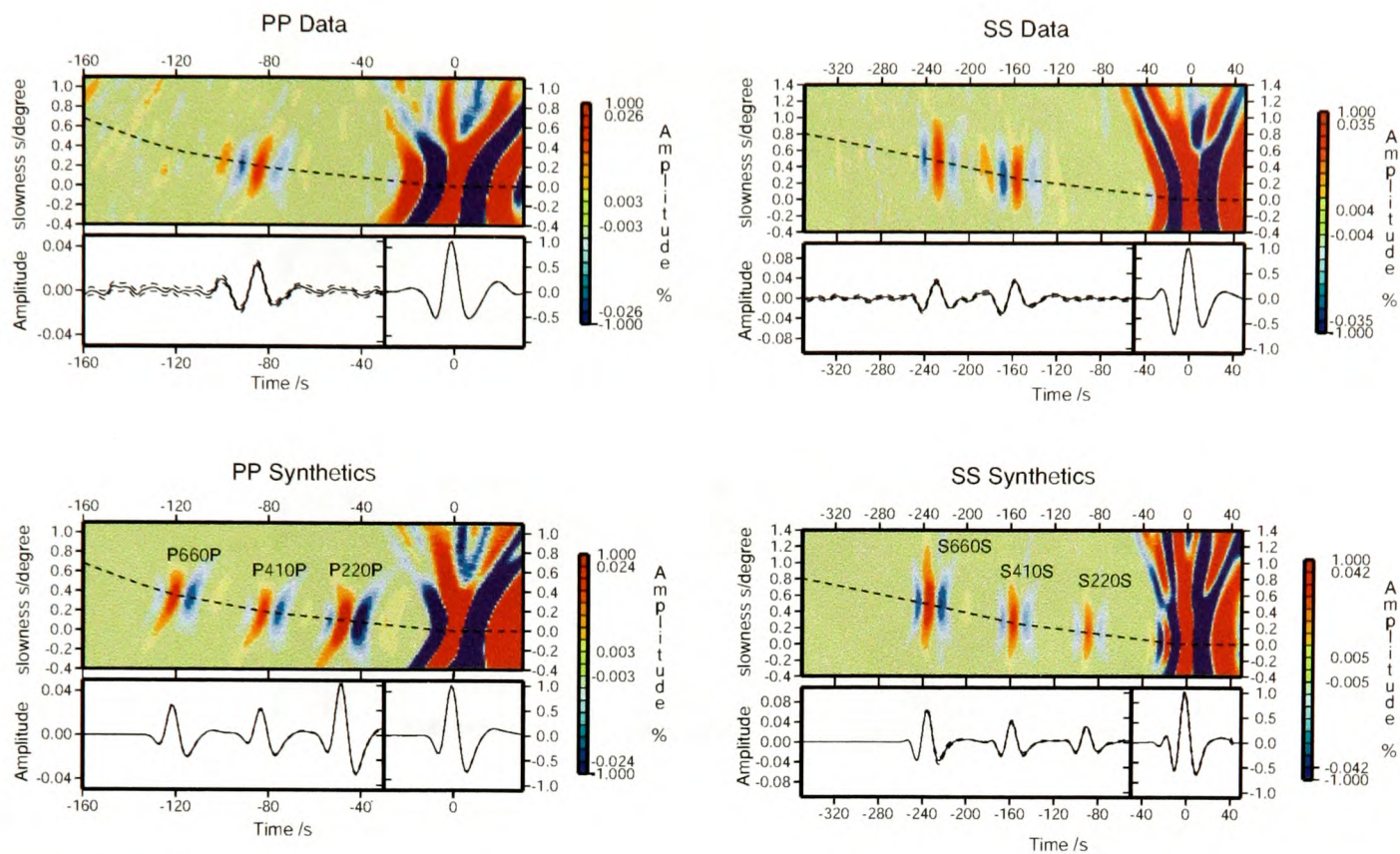


Figure 6.1: Stack results for our global PP and SS datasets. The colour contour plots show the stack as a function of time and slowness with respect to the reference phase (*PP* or *SS*). Stack amplitude is normalised to the magnitude of the surface reflection. The dashed line shows the expected arrival time and slowness of *PdP* and *SdS* phases in PREM. By taking a cross section along this line the time slowness stacks are converted to stack traces (shown beneath). The 95% confidence limits for the stacked traces computed using bootstrap resampling are also shown.

lobe. This suggests that a 520km discontinuity is indeed a global feature as suggested by Shearer (1996).

The major precursor arrivals visible in both datasets occur at -80s and -160s relative to *PP* and *SS*. These arrival times are consistent with reflections from 410km depth. Numerous studies have observed this discontinuity on a regional and global scale (e.g. Helffrich (2000)), and related it to the polymorphic $\alpha - \beta$ phase transition of olivine. Subsequent sections will focus on the lateral variations in topography and reflection amplitudes from this discontinuity.

The synthetic stacks also show large arrivals at -50s and -90s relative to *PP* and *SS*, which are due to reflections from a 220km discontinuity in PREM. Neither of the global stacks shows arrivals consistent with this reflector.

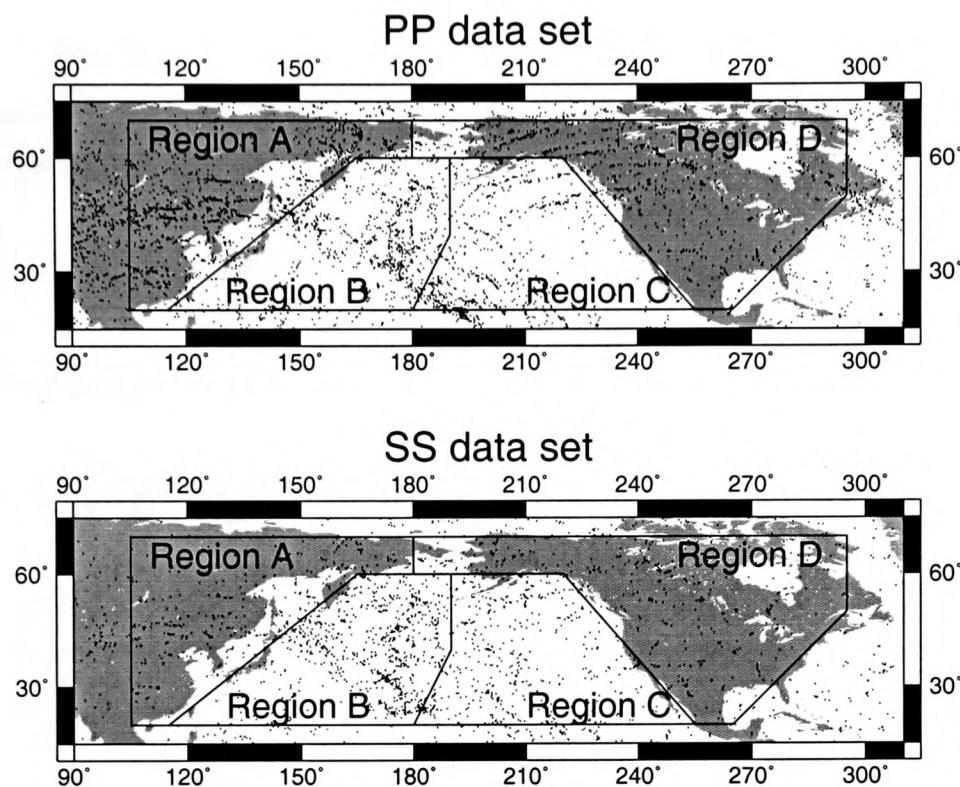


Figure 6.2: Locations of the bounce points together with outlines of the regions used to make the stacks in figure 6.3.

6.2 Regional stacks

In addition to global stacks, four stacks were made for the regions shown in figure 6.2. These regions were chosen to maximise the number of seismograms in the stack and hence stack trace quality without averaging out lateral variability. Figures 6.3 A and B show the resulting stack traces. As with the global stacks we also show stacks of the WKBJ synthetic seismograms.

In addition to the main mantle discontinuities observed in the global stack several of the regional stacks suggest the presence of weak upper mantle reflectors. There are possible, very low amplitude ($\sim 1\%$), precursor phases in regions D and B at around -50s and -100s in the *PP* and *SS* stacks. These could be due to reflections from a 220km discontinuity

Regions B and C for *PP*, and all of the regions for *SS*, suggest arrivals at 70s and 125s before *PP* and *SS* respectively. These would be consistent with a reflector at 300km depth. However, the arrivals could also be a side lobe of the large *P410P* and *S410S* phase which arrives ~ 80 s and ~ 150 s prior to the *PP* and *SS* pulses. A discussion of the lateral variations of the 410km discontinuity will be the focus of the remaining sections in this part of the thesis.

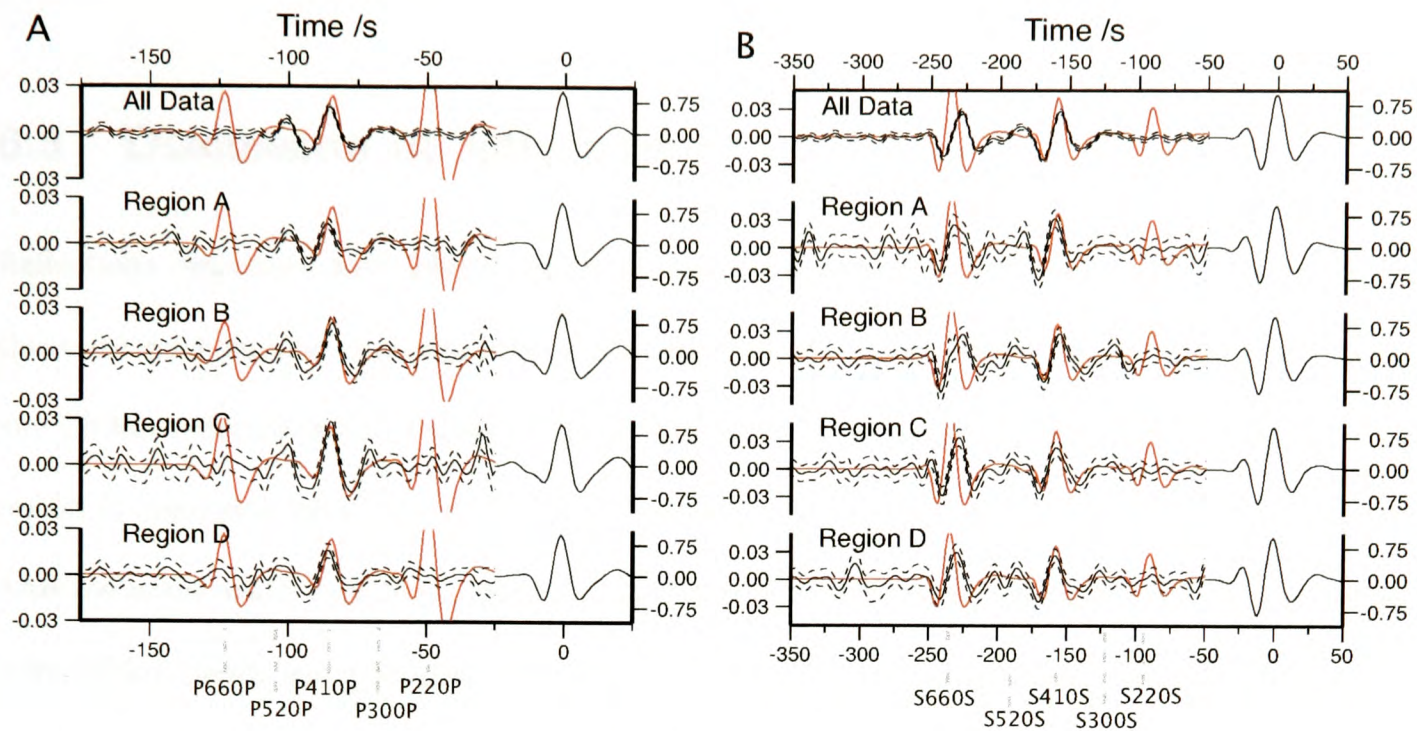


Figure 6.3: Stacks for the whole dataset and the 4 regions showing; **A** PP and **B** SS precursor phases. Black dashed lines show the 95% confidence limits for the stack determined by bootstrap resampling. The red line shows stacks of the synthetic seismograms which incorporate the reference phase and reflections from 220, 410, and 660km depth. The expected arrival positions of various PdP and SdS phases (where d refers to discontinuity depths) are shown at the bottom.

As in the global stacks an arrival prior to the $P410P$ and $S410S$ phases is seen in all the regional stacks. This arrival is consistent with a 520km discontinuity and shows considerable lateral variation. ‘Splitting’ of the 520km discontinuity is observed in the SS precursor stack for region D, and to a lesser extent in the other regions. This is consistent with the observations of Deuss & Woodhouse (2001). Although Deuss & Woodhouse (2001) also observe ‘splitting’ in region A, an area where this study does not.

A $S660S$ arrival is seen for all the regions as well as the global stack at ~ -230 s. Yet there is no evidence for a $P660P$ reflection in any of the regions. The stacks of synthetic seismograms suggest that a $P660P$ reflection should be observable in all the regions if the discontinuity has the same P-wave impedance contrast as that in the PREM model.

Prior to the $P660P$ and $S660S$ phases there are other potential arrivals from the lower mantle. In regions A and D a low amplitude PP precursor is observed at -140s and a large S-wave phase is seen at -300s. In region A this is accompanied by an arrival ~ 275 s prior to SS , and in region

D an arrival ~ 300 s prior to SS.

6.3 Discussion of reflectors

Reflections consistent with 220km depth are only observed in two of the regions (B and D), and these are much smaller than the magnitudes predicted by our synthetic stacks. This suggests that any 220km reflector is not a uniform global feature. The observation of a 220km reflector under region B contrasts with that of Gu *et al.* (2001) who suggested the discontinuity is primarily detected underneath continents. Deuss & Woodhouse (2002) have also observed robust reflections from 220km depth using SS precursor phases. Both these studies made stacks over much smaller regions than those used here. So it is likely that the quality of the 220 reflector in this study is degraded by incoherent stacking over these large regions. Karato (1992) has suggested a change in deformation mechanism from diffusion to dislocation creep as a cause for a 220km seismic discontinuity.

As mentioned previously the speculative arrivals from a 300km discontinuity could be due to side lobes of the much larger P410P and S410S phases. However, both Revenaugh & Sipkin (1994a) and Deuss & Woodhouse (2002) have also made observations of a discontinuity at this depth using long period S-waves. Deuss & Woodhouse (2002) have suggested a phase change in Ca-poor pyroxene as a possible cause for the impedance contrast at 300km depth.

In agreement with Shearer (1996) a clear 520km reflector is observed in all the regional and global stacks. This suggests that this discontinuity is a global feature and is unlikely to be caused by a side lobe of the P410P and S410S phases. The observation of ‘splitting’ of the 520km discontinuity in region D (and possibly C) agrees with Deuss & Woodhouse (2002). The lack of ‘splitting’ observations for the S520S phase in region A and in the PP dataset could be due to incoherent stacking over the broader regions used in this study. Another reason ‘splitting’ of the P520P phase is not as apparent as for the S520S could be that the phase transitions in the non-olivine mantle components, which are believed to cause the splitting, are more-transparent to

P-waves.

The observation of a clear *S660S* phase requires that the absence of the reflector in the *PP* dataset cannot be due to incoherent stacking discontinuity topography or focusing effects, because the two datasets have a similar effective wavelength and sampling (bounce point) distributions. This suggests that the P-wave impedance contrast across the 660km discontinuity is low enough as to make reflections undetectable in our stacks. This has been discussed by Estabrook & Kind (1996) and Shearer & Flanagan (1999), of which the former concluded that Lamé's constant (λ) and seismic parameter (ϕ) are continuous across the boundary making the discontinuity is only second order for shear modulus (μ) and density (ρ)

Both Shearer (1996) and Revenaugh & Jordan (1991b) have discussed the evidence for an S-wave reflector at around 720km depth. *P720P* and *S720S* phases would be expected around 140s and 275s prior to their reference phases. However, the *P720P* phase arrives at a similar time as the diffracted P-wave, and any P-wave reflection from the lower mantle is liable to be very weak due to the wide angle of incidence at the discontinuity. The existence of a reflector at 720km depth could be attributed to an illmenite–perovskite transformation. Deuss & Woodhouse (2002) have also observed a low amplitude lower mantle reflector at 1100km depth beneath America. This is consistent with the observation of an SS precursor arrival at -300s in region D.

Chapter 7

Deconvolution of stacked traces

7.1 Accurate measurement of precursor arrival time and amplitude

Arrivals in the stacked trace are separated out using an iterative deconvolution technique similar to that first used by Kikuchi & Kanamori (1982) and subsequently extended by Bona & Boatwright (1991). The details of this technique are presented in appendix A, here only a brief overview of the deconvolution algorithm is given.

Spike positions in the deconvolved trace are determined by cross correlating the reference (*PP* or *SS*) pulse with the trace. The spike position is placed at the maximum correlation, and its amplitude is calculated using a least squares fit. Then a model trace is derived by convolving the reference pulse with the spike sequence. This model trace is subtracted from the data, before subsequent cross-correlation with the reference pulse to determine the next spike position ... and so on. At each step all the reflection coefficients are solved for simultaneously and the fit of the model trace to the data is measured using an F-test on successive variances. The process continues until the addition of a spike in the reflectivity sequence fails to improve the fit of the model trace to the data by more than 0.1%.

7.2 Uncertainty in precursor amplitude and arrival time

Errors boundaries for the stacks were determined using a bootstrap resampling procedure (Efron & Tibshirani (1991)). For the pre-deconvolution data stacks the process is straightforward. A bootstrap distribution for each time sample of the stacked trace is derived by making repeated stacks using traces, randomly selected with replacement, from the original population. The error associated with the amplitude of each time sample is estimated from the standard deviation of the bootstrap distribution.

However, the application of the non-linear deconvolution operator complicates the procedure in the deconvolved stacks. When the deconvolution operator is applied to traces made from bootstrap samples, the position of an arrival will vary, leading to an overestimate of amplitude error at a given time sample if a straightforward bootstrap is applied. To obtain a true bootstrap distribution for the amplitude of an arrival it is necessary to sum the amplitudes over a time window. This is because the bootstrap statistic we require is no longer the amplitude for a given time sample, but an amplitude for a given arrival which varies in position between the bootstrap stacks. For example, in one random deconvolved stack an arrival may occur at i , and have zero amplitude at $i + 1$, but in another stack it may occur at $i + 1$ and has zero amplitude at i . Accordingly, the amplitude distribution for time sample i will not be the amplitude distribution for the arrival. This bootstrap distribution will overestimate the error for the amplitude of the arrival. The true bootstrap distribution for the arrival amplitude is obtained by summing the amplitudes at i and $i + 1$ in both the random deconvolved stacks.

The width of the window over which amplitudes are summed to derive the bootstrap distribution for the reflection amplitude is determined by the range for which a discontinuity is seen in the bootstrap samples. Similarly the mean and standard deviation of the arrival time can be calculated from the frequency distribution of the arrival times in the bootstrap data.

In cases where arrivals in the deconvolved trace are well constrained the technique works well. However, where the data is less robust the arrival positions in the bootstrap samples can vary greatly which can result in the arrival windows for adjacent reflections overlapping. This makes

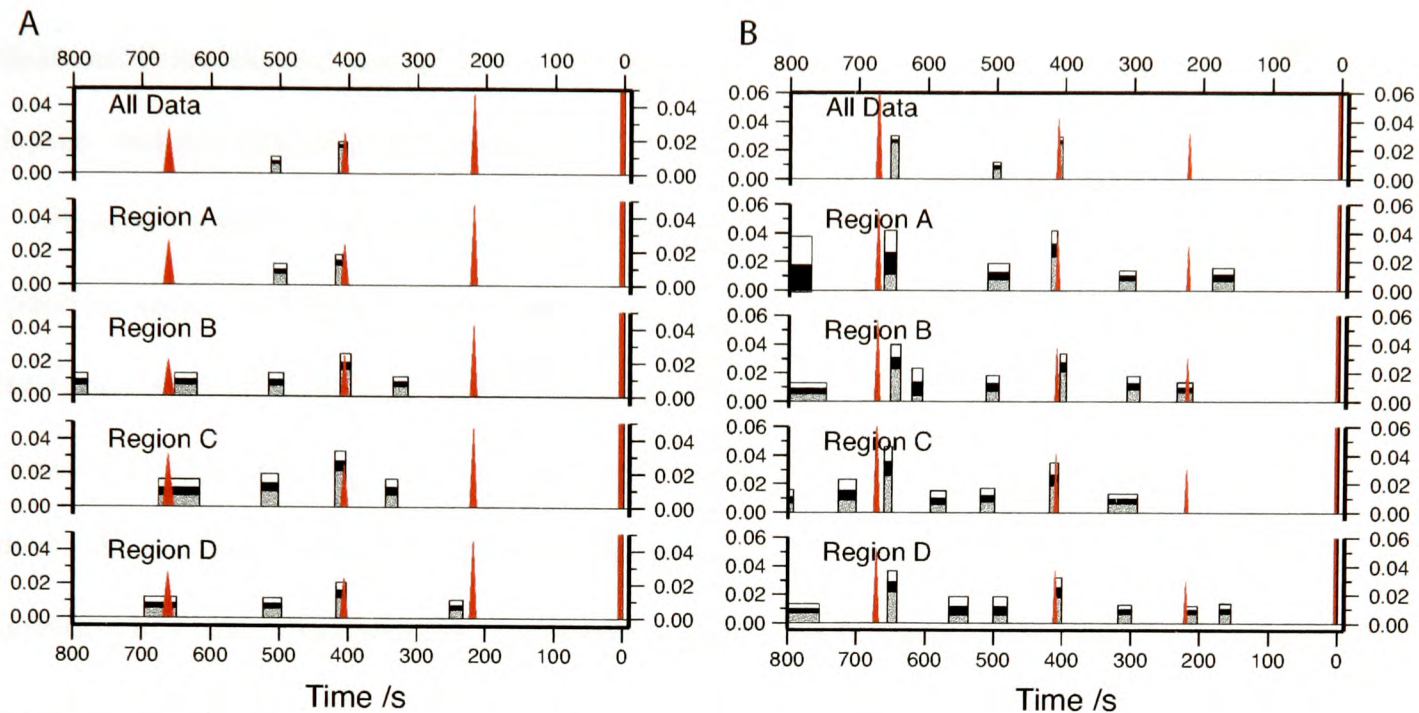


Figure 7.1: Results of the bootstrap analysis on the deconvolved stacks and depth mapping for PP and SS precursors of figure 6.3. The black bar shows the mean amplitude, while the grey and white represent the 95% confidence limits. Bar width represents the uncertainty in discontinuity depth. Red spikes show the deconvolution results for the WKBJ synthetics.

the method unsuitable for noisy/less reliable stacked traces; in these cases it is only possible to measure $P410P$ and $S410S$ amplitudes from the deconvolved mean trace.

7.3 Application of deconvolution to the stacked traces

In figure 7.1 the results of deconvolution, bootstrap analysis, and depth conversion, are shown for the global and regional PP and SS precursor stacks of figure 6.3. Many of the features noted in chapter 6 are made clearer by the deconvolution analysis.

In figure 7.1 it is easier to see the previously discussed upper mantle reflectors from 220 and 330km (section 6.3). It is also possible to observe lateral variations in reflection amplitudes from the 410km discontinuity, which is the subject of chapter 8. The deconvolution also separates out the reflections from 410 and 520km and shows up ‘splitting’ of the 520km discontinuity in the SS precursor stack for North America (region D), and to a lesser extent in the North Pacific. This is consistent with the observations of Deuss & Woodhouse (2001). A 660km reflector is consistently

observed in the *SS* data, while it is not observed in *PP*. Although the deconvolution algorithm picks out P-arrivals consistent with reflections from 600-700km depth these show a wide range of depths, and are inconsistently observed.

Figures 7.2 and 7.3 show the results of deconvolution on a selection of 10 stacks made for 10° spherical caps. In figure 7.2 the stacked traces with their bootstrap determined 95% standard errors are shown. The large negative swing prior to some of the *P410* and *S410S* arrivals, is presumably due to interference between reflections from 410km and 520km depth. Figure 7.3 shows the deconvolved and depth converted mean traces. These reflectivity sequences show lateral variation in 410km discontinuity topography as well as amplitude which will be the subject of the subsequent chapters.

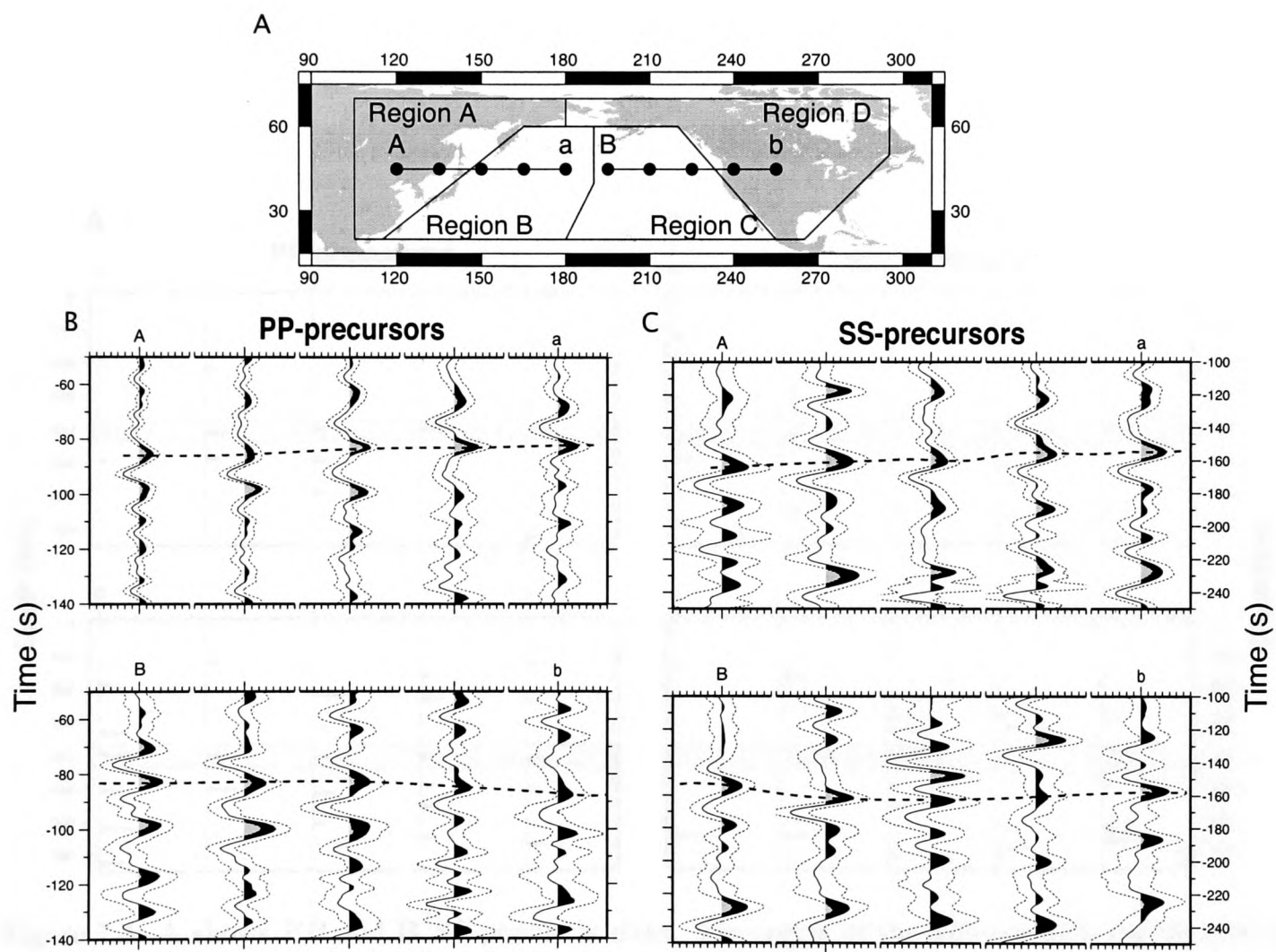


Figure 7.2: Profile through stacks made for spherical caps of 10° radius. **A** shows the location of the caps with respect to the study regions while **B** shows *PP* and **C** *SS* precursors. The position of the 410-arrival is marked with a thick dashed line while thin lines show the 95% confidence limits for the stacked traces. Stack amplitude is given relative to the reference phase with scale intervals of 0.01.

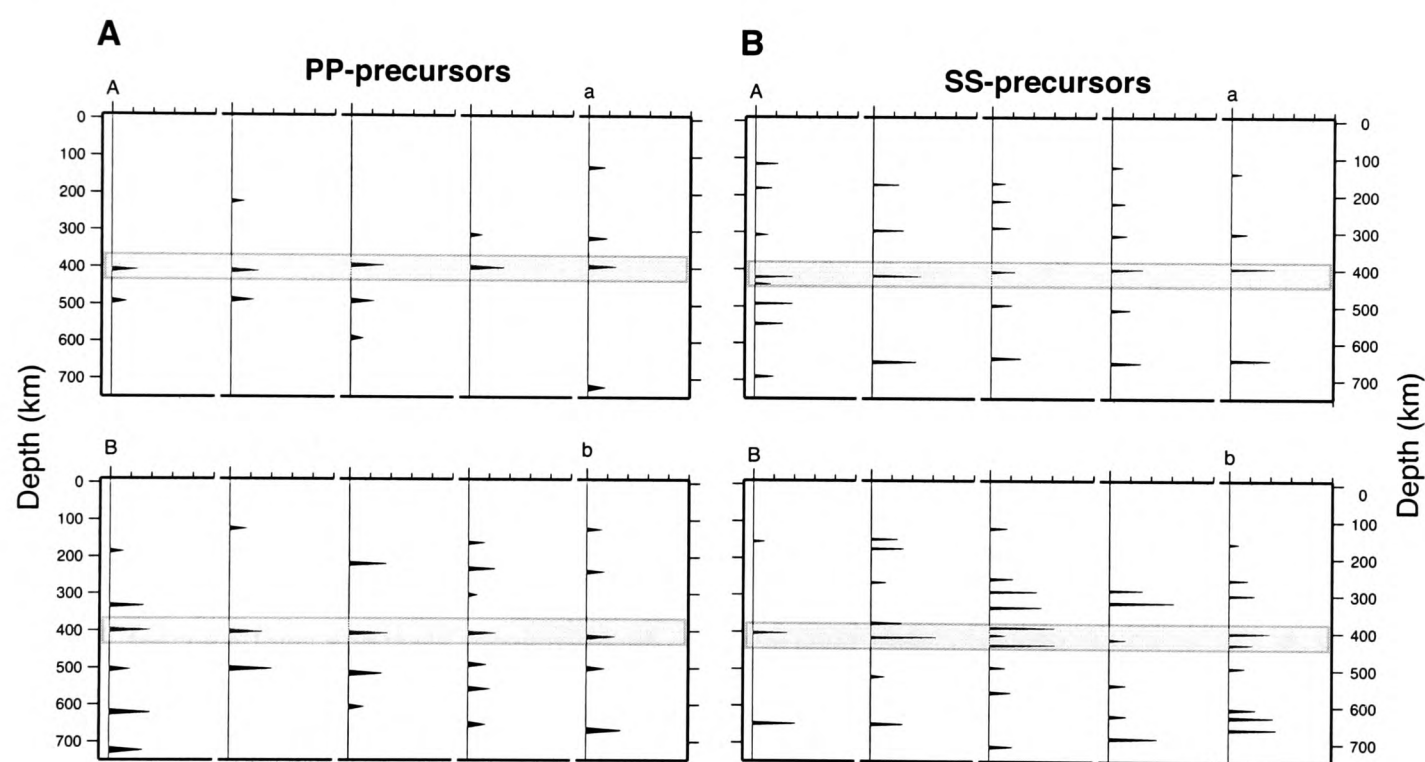


Figure 7.3: **A** shows *PP* and **B** *SS* precursor data. The region of the 410-arrival is highlighted in each cross section. Amplitude scale is the same as figure 7.2.

Chapter 8

The amplitude of reflections from the 410km discontinuity

8.1 Introduction

The application of the deconvolution algorithm (described in the last chapter and appendix A) to stack traces allows accurate measurement of the arrival time and amplitude of the $P410P$ and $S410S$ phases. This chapter is concerned with the observation and interpretation of the amplitudes of the $P410P$ and $S410S$ phases.

The amplitude of scattered waves from a discontinuity provides information on the discontinuity thickness and the contrast in elastic parameters across the boundary. However, current measurements of the reflection coefficient and impedance contrast of the 410km discontinuity are limited to local studies (Revenaugh & Jordan (1991b); Vidale & Benz (1992); Revenaugh & Sipkin (1994a); Rost & Weber (2002); van der Meijde *et al.* (2003)), or global average measurements (Shearer (1996); Shearer & Flanagan (1999)). Here the reflectivity of the 410km discontinuity is examined at both global and regional scales.

¹Material from this chapter was published in Chambers *et al.* (2005a)

Using measurements of $P410P$ and $S410S$ amplitudes the impedance contrasts across the 410km discontinuity are constrained. Lateral variations in the P-wave impedance contrast at 410km are also investigated. These are typically low under North America and China, and higher beneath the North Pacific. The S-wave impedance contrast shows less variability on the regional scale. However, analysis of $P410P$ and $S410S$ amplitudes over smaller areas (by binning traces into spherical caps) shows that the S-wave reflection coefficient varies over much shorter scale lengths than that for P-waves. The analysis of $P410P$ and $S410S$ reflection amplitudes provides a potential tool for investigating compositional and chemical mantle heterogeneity. Accordingly the results have implications for understanding mantle chemistry and processes.

8.2 Regional amplitude variations

To demonstrate the variability of the 410 reflection amplitudes in the regional stacks, frequency histograms for the $P410P$ and $S410S$ amplitudes measured in 1000 bootstrap samples are shown in figure 8.1. The PP dataset shows that the amplitude of the $P410P$ phase is slightly lower than the global stack when measured under the continental regions (A and D) and higher beneath oceanic regions (B and C). The SS dataset shows a different pattern. The mean amplitudes of regions B, C and D are similar for SS while region A shows a higher mean amplitude and larger variance.

The amplitudes of $P410P$ and $S410S$ in the global stacks are slightly lower than those of Shearer & Flanagan (1999). This could be due to differences in the way amplitudes are measured between the two studies. Shearer & Flanagan (1999) measure reflection amplitudes using the maximum of the precursor phase, which means that the presence of noise in the data could cause the reflection amplitudes to be biased towards larger values. In a previous study Shearer (1996) measured $S410S$ amplitudes consistent with this study using a waveform fitting technique similar to the deconvolution algorithm used here.

In order to interpret variations in the amplitudes of $P410P$ and $S410S$ in terms of changes in

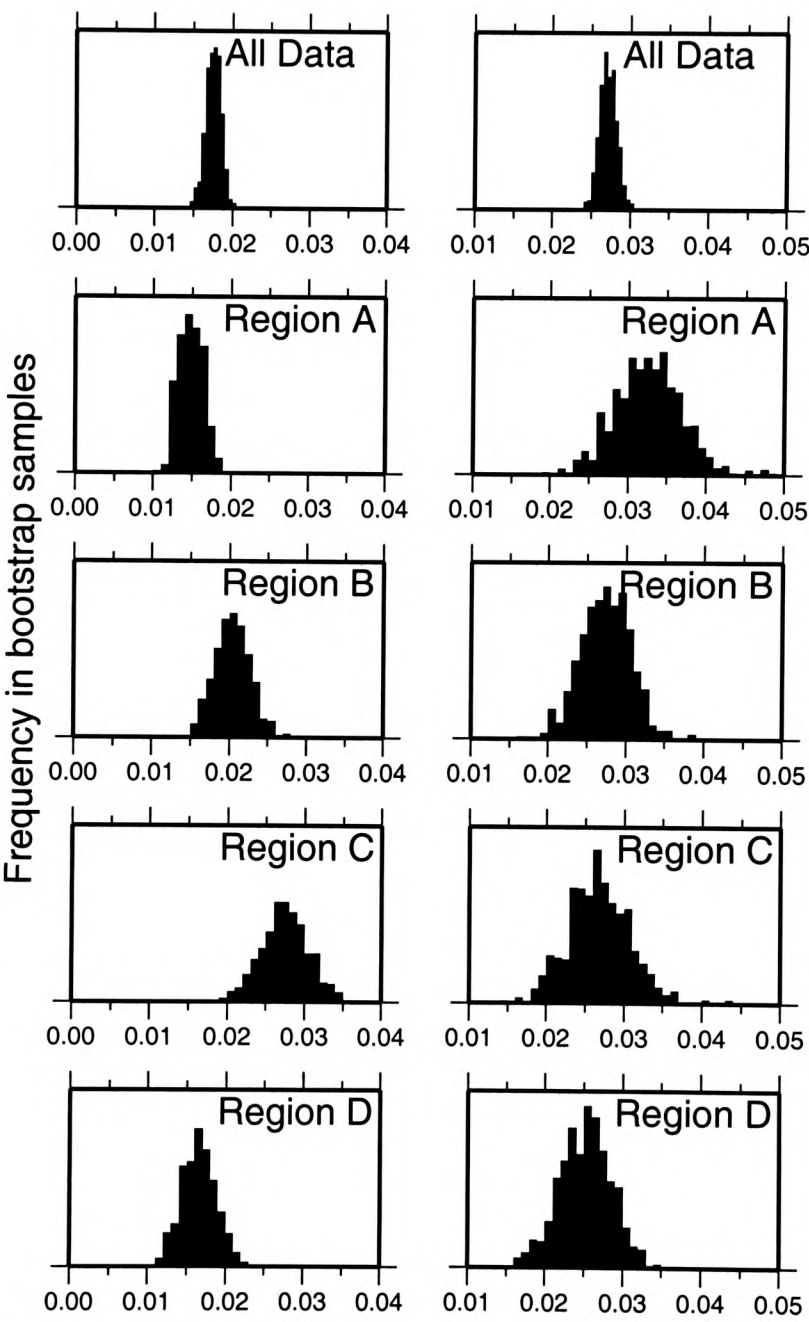


Figure 8.1: Distribution of $P410P$ (left) and $S410S$ (right) amplitudes measured in 1000 bootstrap samples made for each of the regional stacks.

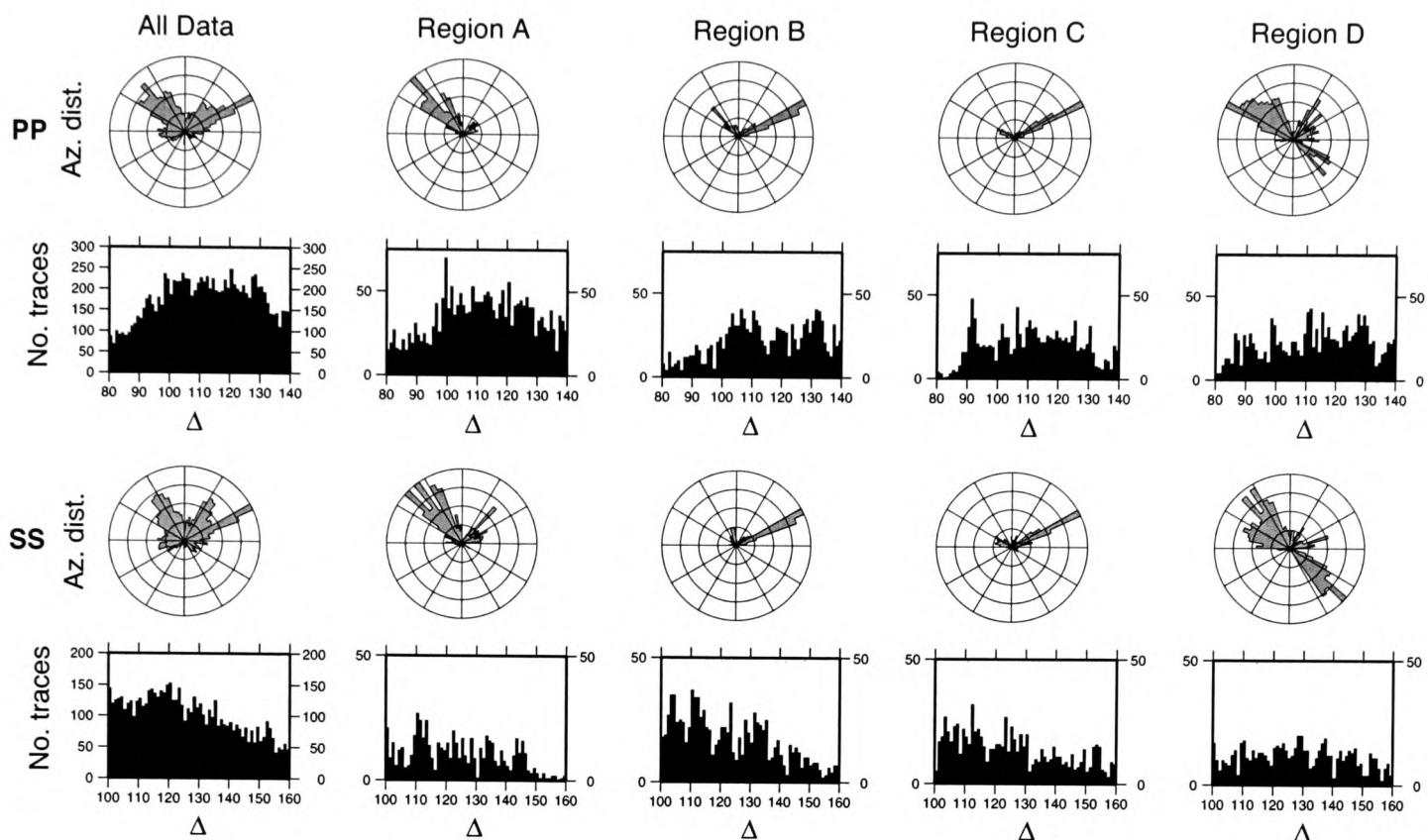


Figure 8.2: Histograms showing the range of epicentral distances (Δ) and azimuths that contribute to the stacks. Azimuths are measured at the bounce point.

the discontinuity properties it is necessary to exclude the possibility that the amplitudes are biased by the presence of other phases. Record sections such as those of Flanagan & Shearer (1998a, 1999), show that there are no interfering phases of significant amplitude that arrive coincident with $S410S$ at 130° . Similar sections for PP precursors show there is a possibility of interference from PKP at around 125° . However, the range of offset distances used (figure 8.2), and the procedure of stacking in the time-slowness domain, means that the contribution from other phases to the precursor amplitudes is minimal.

It is possible that the presence of velocity heterogeneity could focus or defocus rays and result in amplitude variations. However, heterogeneity located away from bounce point would effect precursor and surface reflection equally, and so would not change the amplitude of the precursor phases which is always measured relative to the surface reflection. Focusing due to structure between the discontinuity and the bounce point would be expected to effect the PP and SS datasets equally as they have the same effective wavelength.

A similar argument is true for the focusing due to discontinuity structure, which has been shown to affect precursor amplitudes in individual records by Neele & Snieder (1992). Incoherent stacking also reduces the potential amplitude of precursor phases (see the following section), however, this will also affect $P410P$ and $S410S$ equally due their similar wavelengths. Similar variation in $P410P$ and $S410S$ amplitudes is not seen, so the observed amplitude variations must be due the properties of the discontinuity rather than focusing effects.

The main factors which effect the $P410P$ and $S410S$ amplitudes are impedance contrast and discontinuity thickness. Previously the 410km discontinuity has been modelled by an impedance gradient ranging between 4 and 30km in length (Vidale & Benz (1992); Rost & Weber (2002); Helffrich & Wood (1996); van der Meijde *et al.* (2003)). These lengths are all shorter than one quarter of the peak wavelength of our data (~ 38 km). Accordingly, the long period data used here will not be sensitive to changes in discontinuity thickness. The variations in $P410P$ and $S410S$ amplitudes must relate to changes in the impedance contrast across the 410km discontinuity.

8.3 Impedance contrast of the 410km discontinuity

To investigate the lateral variations in $P410P$ and $S410S$ amplitudes a grid search method is used to establish the combination of S-wave and P-wave impedance contrasts that best suit the data. Because the precursor amplitudes are corrected for differences in epicentral distance, the amplitude for the 410 phases in the deconvolved trace can be considered a measure of the reflection coefficient at the stack reference distance. Formulae for calculating the reflection amplitude from contrasts in seismic velocity and density are given by Aki & Richards (1980). Using the PREM values for seismic velocity and density above the 410km discontinuity, with ray parameters corresponding to the reference distances, reflection coefficients for $P410P$ and $S410S$ phases were calculated for a range of density and seismic velocity jumps across the boundary (figure 8.3).

Strong trade-offs exist between density and seismic velocity changes when modelling reflection amplitudes in this way. So, it is more useful to interpret the results in terms of impedance

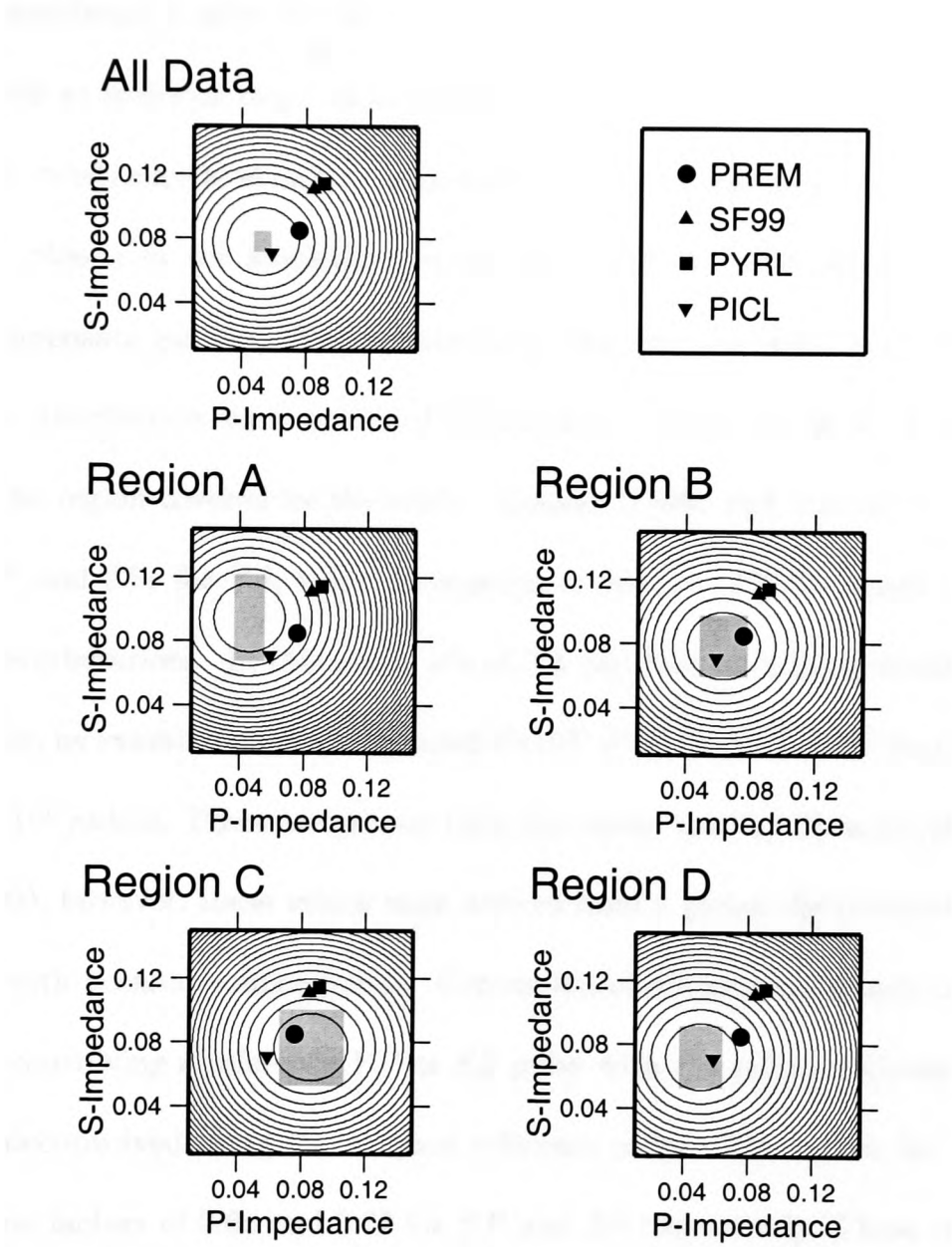


Figure 8.3: Grid search results for all data and the 4 regions. Contours show squared misfit to the measured 410 amplitudes. The shaded region refers to models which result in $P410P$ and $S410S$ amplitudes with in the range of our observations. For reference we also show the models of Shearer & Flanagan (1999) (SF99), and PREM (Dziewonski & Anderson (1981)). Expected impedance contrasts for the pyrolite (PYRL) and piclogite (PICL) models determined from Shearer & Flanagan (1999) are also shown.

contrasts. Here I refer to impedance contrast as the fractional change in impedance across the boundary $\frac{\Delta z}{z_1}$.

$$\frac{\Delta z}{z_1} = \frac{z_2 - z_1}{z_1} \quad (8.1)$$

Where seismic impedance is given by the product of seismic velocity and density ($z = \rho v$), and the subscripts refer to material properties on either side of the boundary.

Differences in crust/mantle structure and discontinuity topography can cause misalignment of the precursor phases in the stacking process and result in lower amplitudes. An attempt was made to compensate for this effect by assuming that the precursor arrival times in a stack have a Gaussian distribution with standard deviation σ . Determination of σ is difficult and will depend on the region covered by the stack. Shearer (1996) and Shearer & Flanagan (1999) used 1.7s for *PP* and 2.5s for *SS*, while Revenaugh & Jordan (1991b) found σ to be 3.0 ± 1.0 s for *ScS₆₆₀ScS* reverberations. In this study σ 's of 2.3 and 3.4s were determined for *P410P* and *S410S* respectively, by examining the uncorrected *P410P-PP* and *S410S-SS* lag times in stacks for spherical caps of 10° radius. These are greater than the values used by Shearer (1996) and Shearer & Flanagan (1999), however, these values were derived from a global distribution of stacks rather than just traces with oceanic bounce points. Conversion of σ 's into amplitude correction factors was achieved by convolving a reference *PP* or *SS* pulse with the relevant Gaussian distribution, which was then deconvolved using the original reference pulse. The values for σ for this study result in correction factors of 0.66 and 0.72 for *PP* and *SS* respectively. These correction factors were applied to the theoretical reflection coefficients before comparison with observed amplitudes.

The application of a single correction factor to the global and regional stacks assumes that discontinuity topography and velocity heterogeneity within the regions is similar to the global average. Given the broad regions used in this study this assumption seems satisfactory, however, application of the factors to smaller areas could lead to overestimates of the impedance contrast.

Using the grid search the P-wave and S-wave impedance contrasts across the 410km discontinuity were constrained for our global and regional stacks (figure 8.3). The best fitting values with their errors are given in table 8.1; for comparison we also give the results from previous studies in

Table 8.1: Optimum values for P-wave and S-wave impedance contrasts in the global stack and the 4 regions.

	P-wave Impedance	S-Wave Impedance
All Data	0.053 ± 0.005	0.078 ± 0.006
Region A	0.046 ± 0.010	0.095 ± 0.027
Region B	0.064 ± 0.015	0.079 ± 0.019
Region C	0.087 ± 0.020	0.077 ± 0.024
Region D	0.051 ± 0.013	0.072 ± 0.019

Table 8.2: Results of previous studies for P-wave and S-wave impedance contrasts at 410km. Expected impedance contrasts for pyrolite and piclogite mantles were adapted from Shearer & Flanagan (1999).

	P-wave Impedance	S-Wave Impedance
Revenaugh & Jordan (1991b)	-	0.046 ± 0.01
Shearer (1996)	-	0.067 ± 0.011
Shearer & Flanagan (1999) (optimum)	0.085	0.111
Gaherty <i>et al.</i> (1999)	-	0.092 ± 0.03
Rost & Weber (2002)	0.065	-
PREM (Dziewonski & Anderson (1981))	0.077	0.085
pyrolite ($\sim 65\%$ olivine)	0.0913	0.1138
piclogite ($\sim 40\%$ olivine)	0.0590	0.0705

table 8.2. The optimum model of Shearer & Flanagan (1999) has higher P- and S-wave impedance contrasts than we measure. This would suggest that the correction for the incoherent stacking was underestimated. However, as mentioned in section 8.2 the amplitudes of P_{410P} and S_{410S} measured by Shearer & Flanagan (1999) are biased upwards by the use of the maximum value as the precursor phase amplitude. Where a waveform fitting technique was used to determine the amplitude of oceanic precursor phases by Shearer (1996) the precursor phase amplitudes and the corresponding impedance contrast compares well with our global value.

Revenaugh & Sipkin (1994a) found normal incidence S wave reflection coefficients of 0.032 around Japan and South China and 0.066 in Northern China. The average of these values is consistent with this study in region A (0.044 ± 0.014). Gaherty *et al.* (1999) requires that the vertical incidence S-wave reflection coefficient at the 410km discontinuity lies between 0.03 and 0.055. The S-wave reflection coefficient determined from the global stack lies within this range at 0.038 ± 0.004 .

The measurements of global P-wave impedance contrast are lower than previous global studies such as Shearer & Flanagan (1999), although their 95% confidence limits include values as low as 0.06. The value of Rost & Weber (2002) for the Northwest Pacific (0.065) is within the range of the comparable regions B and C. There is a lack of observations of P-wave impedance contrast in regions A and D, to which the results can be compared.

8.4 Global amplitude variation

Figure 8.4 shows global maps for the amplitudes of $P410P$ and $S410S$ with a horizontal section through global S-wave velocity perturbation model S20RTS (Ritsema *et al.* (1999)) at 410km depth. The long-wavelength structure is emphasised, and the dependence of results on individual caps reduced, by expanding the amplitude surfaces using degree-12 spherical harmonics. The harmonic coefficients were determined using least squares inversion. Tests showed that stable results were achieved by adding 0.1% of the average diagonal value to the inner product matrix. In addition the data was weighted by the square root of the number of traces in each cap.

It is important to note that the individual amplitudes shown in figure 8.4 could be influenced by lateral variations in discontinuity topography and velocity heterogeneity. For this reason only the patterns of $P410P$ and $S410S$ amplitudes are compared, as the $P410P$ and $S410S$ amplitudes should be influenced equally by focusing and topography effects.

The amplitude measurements are also compared with local S-wave velocity perturbations in figure 8.5. There appears to be little relation between the global patterns of $P410P$ and $S410S$ amplitudes. Examination of the amplitude maps suggests that $P410P$ amplitudes are dominated by a long wavelength pattern: Eurasian and North American amplitudes are low and Pacific amplitudes are high. While $S410S$ amplitudes appear to be dominated by a higher degree pattern.

To demonstrate the difference in scale length between $P410P$ and $S410S$ anomalies power spectra for the two spherical harmonic expansions are shown in figure 8.6. The $P410P$ spectra is

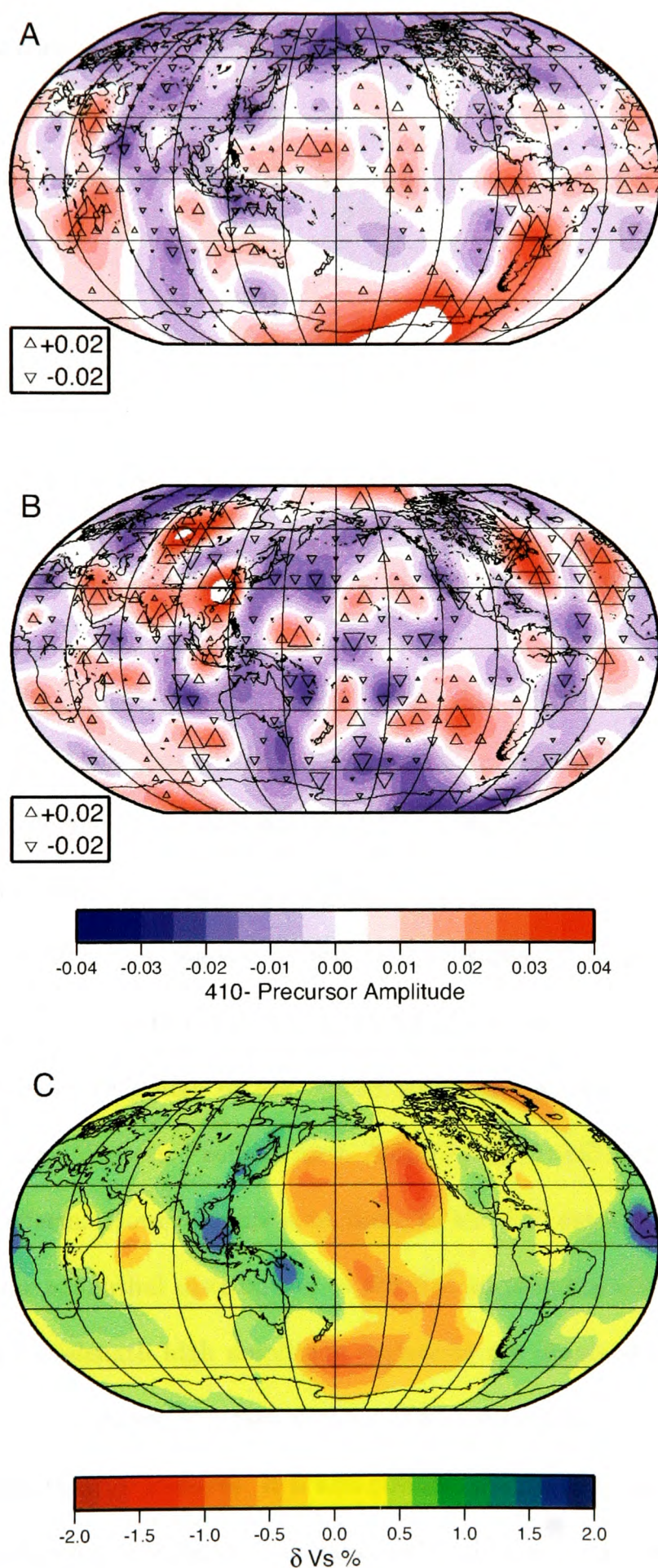


Figure 8.4: **A** P_{410P} and **B** S_{410S} reflection amplitudes. Triangles represent the amplitude difference from mean observed at each cap. Mean amplitudes are 0.032 for P_{410P} and 0.047 for S_{410S} . Coloured contours represent the amplitude surface expanded in spherical harmonics to degree-12 and evaluated at the cap locations. **C** Section through mantle tomography model S20RTS Ritsema *et al.* (1999) taken at 410km. For comparison with the amplitude variations velocity anomalies were averaged over a 10° circle around each cap location.

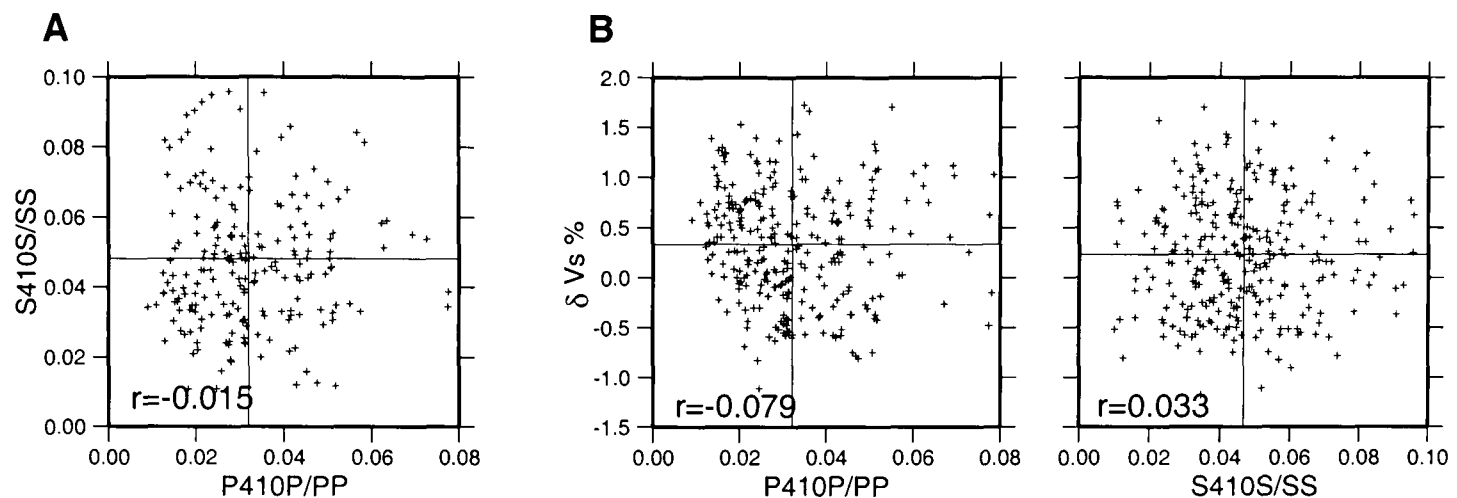


Figure 8.5: **A** Comparison of $P410P$ and $S410S$ amplitudes measured at each cap location. **B** Comparison of $P410P$ and $S410S$ amplitudes with velocity anomalies from tomography model S20RTS (Ritsema *et al.* (1999)) at 410km depth. As in figure 8.4 velocity anomalies were averaged over a 10° circle around the cap location. The product moment correlation coefficient (r) is shown for each pair of variables and the cross is centred on the mean value for each axis.

dominated by the degree-1 signature, while the $S410S$ has only a small contribution from degree-1. The contributions to the $P410P$ power spectra from the higher degrees are relatively uniform with a slight decrease after degree-10. In contrast the $S410S$ power spectra are dominated by higher degrees, the peak contributions being from degrees 4 and 10. There is no real decrease in the $S410S$ power spectrum at higher degrees suggesting that degree-12 spherical harmonics are not sufficient to represent the amount of variation in $S410S$ amplitudes observed here.

Many of the regions of very high or low $S410S$ amplitudes contain cap values which show considerable scatter (e.g. Western Pacific, North Atlantic) and could potentially be the results of localised focusing effects. However, this is unlikely, as the PP dataset has the same effective wavelength and similar azimuthal distribution to the SS , so focusing effects should effect both equally. Fewer short wavelength high amplitude variations are observed in $P410P$ amplitudes, suggesting that S-wave impedance contrast at the 410km discontinuity has a much shorter scale of lateral variation than that for P-waves. It is also possible that high amplitude short wavelength $S410S$ anomalies could be due to differences in data quality. However, the short wavelength $S410S$ amplitude structures persist in well sampled areas such as the Northern Pacific. Hence, we believe that the cause of the differences in the patterns of $S410S$ and $P410P$ amplitudes is related to

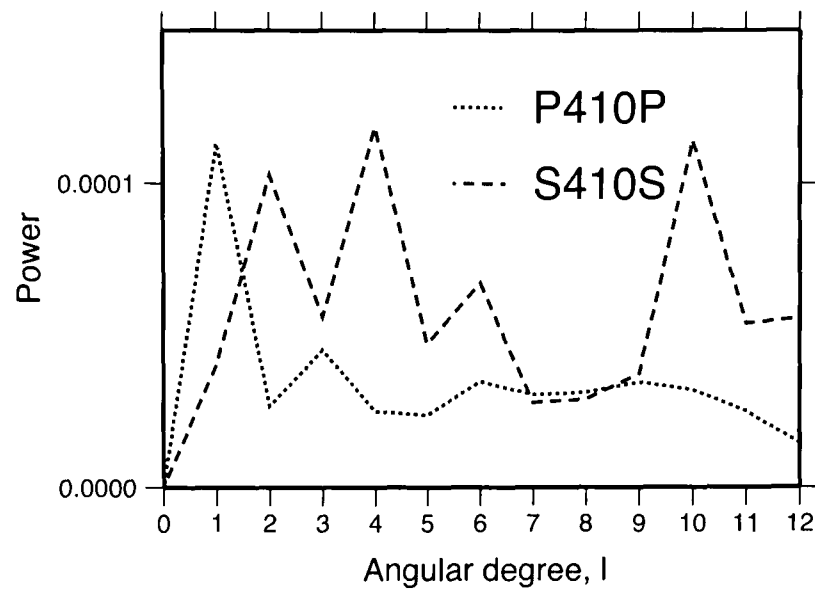


Figure 8.6: Amplitude spectra (average power per angular degree, l) for the spherical harmonic expansions of $P410P$ and $S410S$ amplitudes. In all cases the inner product of the spherical harmonics is normalised to 1. The $P410P$ spectra is dominated by a degree-1 signature while the $S410S$ is dominated by shorter wavelength signals at degrees 4 and 10.

the P and S impedance contrasts at the discontinuity. Accordingly, the spatial scale of variations in S impedance contrast must be shorter than the corresponding length for P wave impedance contrast.

Shorter scale lengths for S-wave impedance contrast at 410km could explain the lack of variation in the regional estimates of S-wave impedance contrast (section 8.3). Figure 8.4 shows that the lateral variations in $S410S$ amplitudes are considerably shorter than the dimensions of the regions used to make the stacks in figure 6.3. Yet, the P-wave impedance contrasts vary on similar length scales to the regions used in section 8.3. So, the lateral differences in $P410P$ amplitude became apparent in the regional stacks, while variations in $S410S$ amplitude were averaged out in the stacking process.

8.5 Discussion of $P410P$ and $S410S$ reflection amplitudes

Seismology, geodesy, geochemistry and mineral physics remain the only tools for constraining mantle composition and behaviour. In this study the contribution from seismology has been extended to include lateral variations in reflection coefficients and impedance contrasts for the

410km discontinuity on a global scale.

By making stacks which only include traces with bounce points within certain regions it was shown that variations in precursor amplitudes are observable. These differences cannot be attributed to interference from other phases, or focusing effects from discontinuity topography. Given the frequency range of the data used in this study, variation in discontinuity thickness is also rejected as a cause for our observations of precursor amplitude variation. Instead, these lateral variations in reflection amplitude are best modelled as lateral variations in impedance contrast.

The P-wave impedance contrast of the 410km discontinuity are significantly higher beneath the North Pacific compared to Asia and North America. This may suggest that the differences between oceanic and continental mantle extend to the top of the transition zone. However, when $P410P$ and $S410S$ amplitudes are measured in stacks for spherical caps there is no clear ocean continent difference. For example South America and the Middle East both show high $P410P$ amplitudes.

Differences in S-wave impedance are not as significant and show more scatter, which is suggestive of lateral variations in impedance contrast with shorter length scale. This is confirmed by examination of $P410P$ and $S410S$ amplitudes in stacks for spherical caps, which shows that the scale length of variations observed in P-wave reflectivity from the 410 is much greater than that for S-waves.

The low values for the P and S-wave impedance contrasts of the 410km discontinuity might appear to be more consistent with a low-olivine piclogite model for mantle composition (figure 8.3). However, the predicted impedance contrasts across the discontinuity in pyrolite and piclogite models assume an anhydrous wadsleyite phase. There is currently a lack of experimental information on the elastic properties of hydrated wadsleyite with realistic iron contents. Yusa & Inoue (1997) have shown that the addition of water to pure Mg-wadsleyite can reduce the bulk modulus by 5-11%, and Jacobsen *et al.* (2004) have shown that the addition of 1% wt of water to an Fo₉₀ ringwoodite can reduce P- and S-wave impedance by 5.8 and 7.6% respectively. So if

hydrated wadsleyite in the mantle behaves in a similar fashion then a pyrolite mantle could still be consistent with our measured impedance contrasts of the 410km discontinuity.

The lack of correlation between P and S-wave reflection amplitudes requires that mantle olivine content is not the major source of impedance variation for at least one of the data types. Other chemical variations such as the mantle's aluminium and iron content are more likely to effect the width and depth of the phase transition (Weidner & Wang (2000)), rather than the impedance contrast, and so are not likely to be the cause of the observed variations in precursor amplitude.

If lateral variations in mantle temperatures and their associated effects on elastic parameters were the cause of variations in 410 impedance contrast, we would expect the variations in P-wave and S-wave reflection amplitudes to correlate. So it is not surprising that no correlation is seen between the measurements of precursor amplitude and velocity anomalies in S20RTS, if these are taken as a proxy for lateral variations in temperature. Lateral differences in reflection coefficient will not be observed where a thermal anomaly is continuous across the discontinuity. As the $\alpha - \beta$ phase change and hence the 410km discontinuity is believed to have a positive Clapeyron slope this would aid the transfer of thermal anomalies across the boundary, making it more likely thermal anomalies are continuous across the discontinuity and hence less visible.

Lateral variation of intrinsic attenuation could also result in variation of the observed amplitudes of seismic phases. As the discontinuity reflection is always compared to the surface reflected phase only attenuation between the discontinuity and surface reflection point will contribute to variations in precursor reflection amplitudes. However, given the long wavelength nature of the data, attenuation of the surface reflections is not likely to be a significant factor in determining precursor amplitudes (for example see Neele & Snieder (1992))

The presence of an anisotropic material adjacent to the 410km discontinuity, with a varying orientation for the fast axis could also cause precursor amplitude variations. Anisotropic effects are generally more pronounced for S-waves than for P-waves, so this could explain the greater variability observed in *S410S* amplitudes compared to *P410P*. However, the details of reflection coefficients in anisotropic media are complex and a full discussion is beyond the scope of this study.

Bercovici & Karato (2003) have speculated that up-welling material passing through the 410km discontinuity could undergo dehydration induced partial melting. They suggest this as a mechanism by which whole mantle circulation can take place with the existence of isolated geochemical reservoirs. Both Revenaugh & Sipkin (1994b) and Song *et al.* (2004) have observed low velocity zones above the 410km discontinuity beneath China and the Western United States. These are interpreted as regions where partial melt exists atop the 410km discontinuity. If this occurs globally, then we would expect that regions of high reflection coefficient correspond to regions with a larger degree of partial melting above the discontinuity. However, there is a complication as the associated water content beneath the discontinuity would also lower the seismic velocities in wadsleyite. The presence of partial melt would be expected to effect S-wave velocities more than P-waves. Hence it may be that the short wavelengths in S-wave reflection coefficient we observe are due to short wavelength variations in a melt layer, while the P-wave variations are primarily sensitive to another process.

8.6 Conclusions

The amplitude of P and S-waves reflected from the 410km discontinuity was examined. $P410P$ amplitudes are regionally variable, being lower under North America and China, and higher beneath the North Pacific. The S-wave reflection amplitudes show less variability on a regional scale. However, when smaller areas were considered variation in $S410S$ amplitudes was found, but over shorter spatial lengths than for $P410P$ amplitudes.

The lack of correlation between our $P410P$ and $S410S$ reflection amplitudes could be due to the presence of melt, water or other chemical heterogeneities in the upper mantle and transition zone. Other factors such as temperature, mantle olivine content variations, or focussing from mantle structure would require $P410P$ and $S410S$ amplitudes to correlate, and so cannot explain all the variation.

The reflection amplitudes constrain the global average and regional impedance contrasts across

the 410km discontinuity. The values for the impedance contrast at 410km appear to be more consistent with a low-olivine piclogite model for mantle composition. However, this assumes an anhydrous wadsleyite phase. A high olivine pyrolite mantle could be consistent with the impedance contrasts of this study if hydration of the wadsleyite phase is taken into account.

Chapter 9

Topography of the 410km discontinuity

9.1 Introduction

In this chapter the time lags between precursors from the 410km discontinuity and the surface reflection phase ($P410P - PP$ and $S410S - SS$ lag times) are studied. This lag time is a function of the position of the $\alpha - \beta$ olivine phase change and velocity perturbations in the crust and upper mantle. For this reason the dependence of precursor lag time derived topography on the travel time corrections for crust and upper mantle structure is first examined. Then new models for 410-topography are presented, which are constrained using both $P410P - PP$ and $S410S - SS$ measurements.

Recent models suggest that the precursor derived topography of the 410km discontinuity is dominated by low degree spherical harmonics (Flanagan & Shearer (1998a); Gu *et al.* (1998); Flanagan & Shearer (1999)). However, there remain some discrepancies as to the details of the topography and its relation to mantle processes. For example, Gu *et al.* (1998) found that

¹Material from this chapter was published in Chambers *et al.* (2005b)

the discontinuity is often elevated beneath continents and depressed beneath oceans. This agreed with the earlier proposal of Gossler & Kind (1996) that ocean-continent differences extend into the transition zone. However, the correlation between discontinuity topography and surface features is not evident in the maps of Flanagan & Shearer (1998a) and Flanagan & Shearer (1999). The possibility of surface features influencing structure in the deep upper mantle and transition zone, has major implications for our understanding of mantle dynamics.

The relationship between topography of the 410km discontinuity and transition zone seismic velocity anomalies is also examined. A moderate negative correlation exists between discontinuity topography and transition zone seismic velocity anomalies at very low spherical harmonic degrees. This is consistent with the expected relationships between, the position of the $\alpha - \beta$ olivine phase boundary, seismic velocity anomalies, and mantle temperatures. However, when shorter wavelengths are included the relationship becomes more complex. In smaller regions beneath Northeast Asia and the Western Pacific positive correlations are observed, between seismic velocities and discontinuity topography, which are interpreted as evidence for the effects of chemical heterogeneities on seismic velocities and the 410km discontinuity. The pattern of temperature variation in the mantle is reflected in long wavelength discontinuity topography. However, at shorter wavelengths the influence of chemical heterogeneities becomes visible in seismic velocity anomalies and mantle discontinuity topography. This separation of the spectra for thermal and chemical heterogeneities suggests that chemical heterogeneities can survive in a convecting mantle.

9.2 Observations of 410km discontinuity topography

9.2.1 Precursor lag times and upper mantle corrections

In figure 9.1 the $P410P - PP$ and $S410S - SS$ lag times plotted at the cap locations are shown. There is a high degree of similarity between the maps of $P410P - PP$ and $S410S - SS$ differential times. Both maps show shorter lag times beneath Canada, North America and the Northwest Pacific, and greater lag times beneath the Eastern Pacific and Northern Asia. The regions where

the maps disagree are generally regions of sparser data coverage such as the Southern Pacific and Atlantic oceans.

The travel time corrections for the $S410S - SS$ differential times is also shown in figure 9.1. These can be considered as the expected $S410S - SS$ differential times for a flat discontinuity and crust and mantle structure corresponding to the S20RTS and CRUST5.1 models (Ritsema *et al.* (1999); Mooney *et al.* (1998)). These travel time corrections are virtually identical to those used by Flanagan & Shearer (1998a) for their study of $S410S - SS$ lag times. The travel time corrections for the $P410P$ phases were calculated in a similar manor to $S410S$ except that the upper mantle S-wave velocity perturbations from S20RTS were scaled by a factor of 0.5.

Figure 9.1 shows that there are strong similarities in the pattern of differential travel times and the travel time corrections. For example the lower than average precursor lag times beneath Canada, Western Australia and the Northwest Pacific are also regions with faster than average seismic velocities, and hence have a negative travel time correction.

The precursor phase-surface reflection lag time will be a function of the discontinuity depth and the seismic velocities of the crust and mantle between the discontinuity and the surface. So any discontinuity topography subsequently derived from precursor lag times will be influenced by the accuracy of the travel time corrections applied to compensate for crust and upper mantle structure.

Although the pattern of upper mantle velocity heterogeneity is well known from models such as S20RTS, the magnitude of these velocity perturbations may be inaccurate due to damping applied when deriving such models. Accordingly, there exists some uncertainty in the $P410P - PP$ and $S410S - SS$ correction times, which could lead to errors in the determination of discontinuity depths. This is investigated by calculating discontinuity topography with travel time corrections multiplied by factors of 1.0, 1.5, and 2.0 in figure 9.2. All of these topography maps retain the major pattern observed in the precursor lag time.

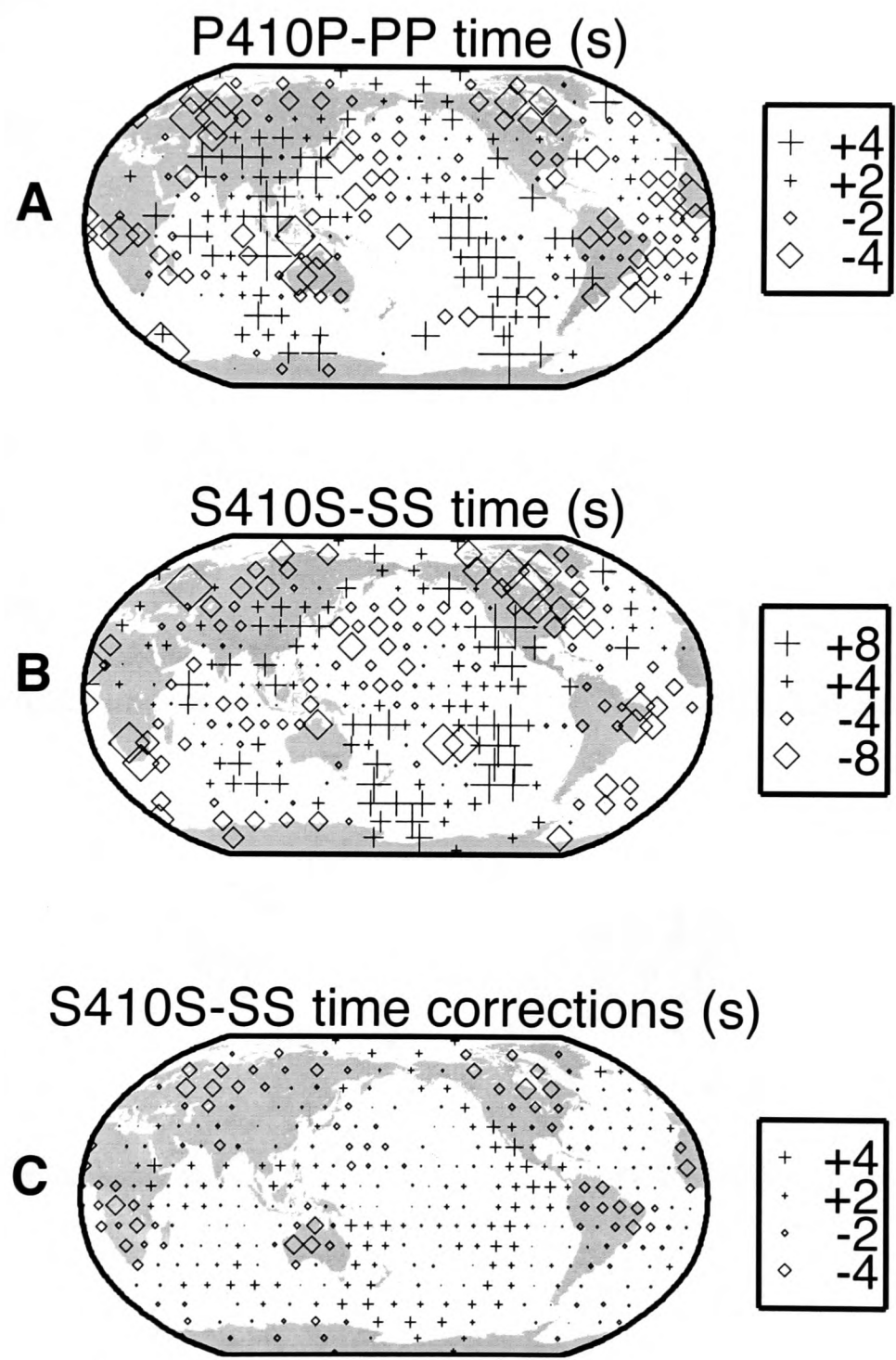


Figure 9.1: **A** and **B**, $P410P - PP$ and $S410S - SS$ lag times, measured in seconds and plotted at each cap location. **C** Travel time corrections for $S410S - SS$ phase at each of the caps.

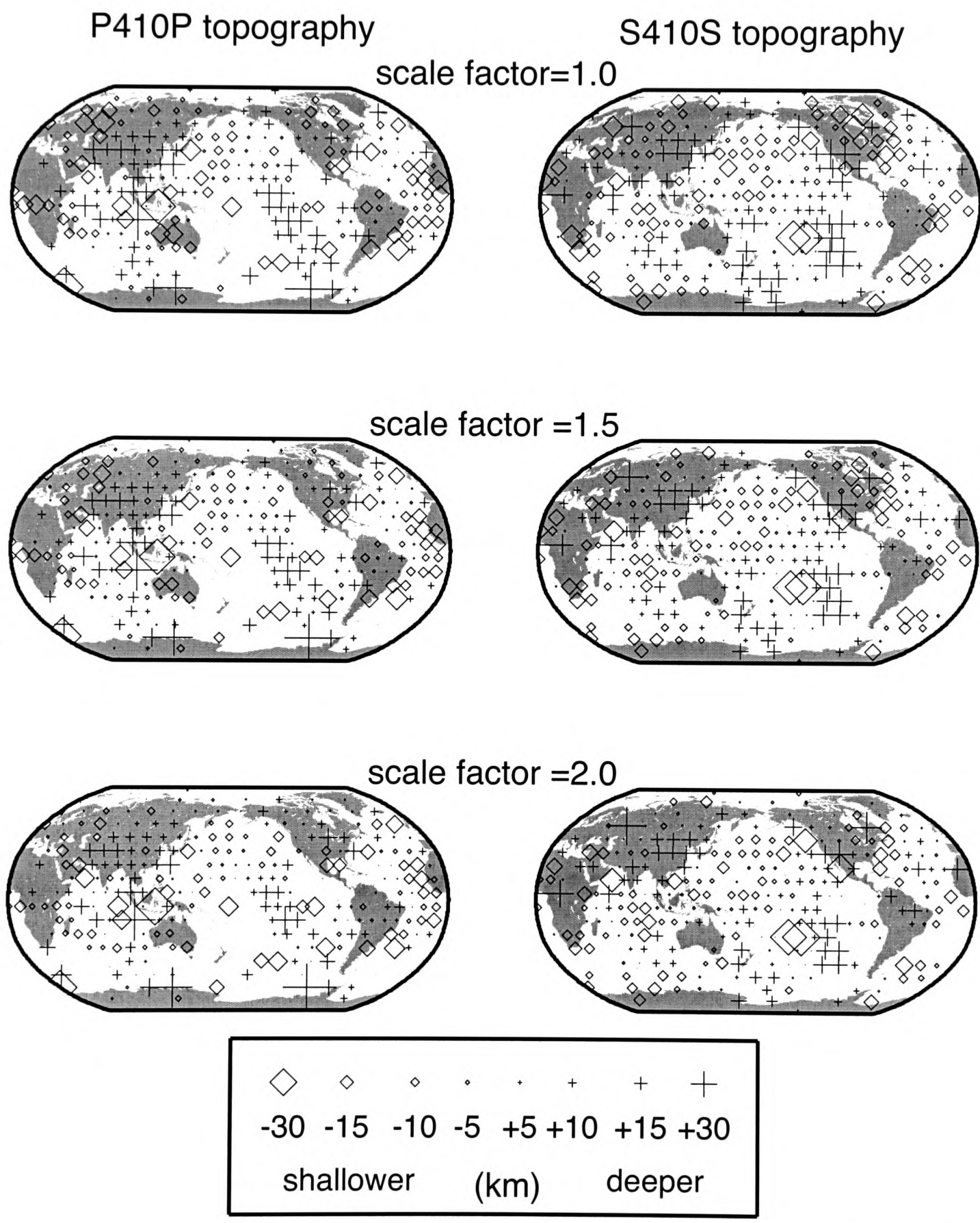


Figure 9.2: Lateral variations in discontinuity depths determined using the *P410P* and *S410S* datasets using different scaling factors for the travel time correction. The maps were derived by multiplying the normal (scale factor of 1.0) travel time corrections before applying these to the measured *P410P – PP* and *S410S – SS* lag times and converting these to discontinuity depths.

9.2.2 Models for discontinuity topography

Further examination of 410-topography is carried out using the original travel time corrections for crust and mantle structure from S20RTS and CRUST5.1 models (i.e. a scale factor of 1.0 in figure 9.2). Both the *PP* and *SS* datasets compare well in well sampled regions such as the North Pacific, North Asia, and North America. These show a depressed 410km discontinuity beneath East Asia and the West coast of North America, while the 410 is uplifted beneath the Northwest Pacific, Northeastern America, Australia, and Europe. Disagreement is restricted to more sparsely sampled regions such as around Tonga, and the Southern Atlantic and Indian Oceans.

In figure 9.3 smoothed maps of the topography for the *PP* and *SS* datasets are shown. To emphasise the long-wavelength structure and reduce the dependence of results on individual caps the topography surfaces were expanded using degree-12 spherical harmonics. The inverse problem for the spherical harmonic coefficients was regularised by adding 0.01% of the average diagonal value to the inner product matrix, and the data weighted by the square root of the number of traces in each cap.

The use of a damped inversion to determine spherical harmonic coefficients is preferred to previous techniques which first apply smoothing splines to fill in gaps in data coverage (Flanagan & Shearer (1998a, 1999)), the use of smoothing splines will bias the inversion for spherical harmonic coefficients away from regions of better data coverage. However, irregularities in the density and quality of sampling means that the topography models should be treated with caution in regions of sparse data coverage.

The preferred discontinuity topography model (figure 9.3 C) was derived by inverting for spherical harmonic coefficients using both datasets, with weighting by the square-root of the number of traces for each cap. This allows the two different distributions of bounce points to increase the resolution in regions where one or the other dataset is poorly represented. In this case data coverage is sufficient so that regularisation of the inverse problem is not required.

The 410km discontinuity topography maps (hereafter referred to by their relevant dataset as *P410P*, *S410S*, and Joint) have many features in common with previous maps (Flanagan &

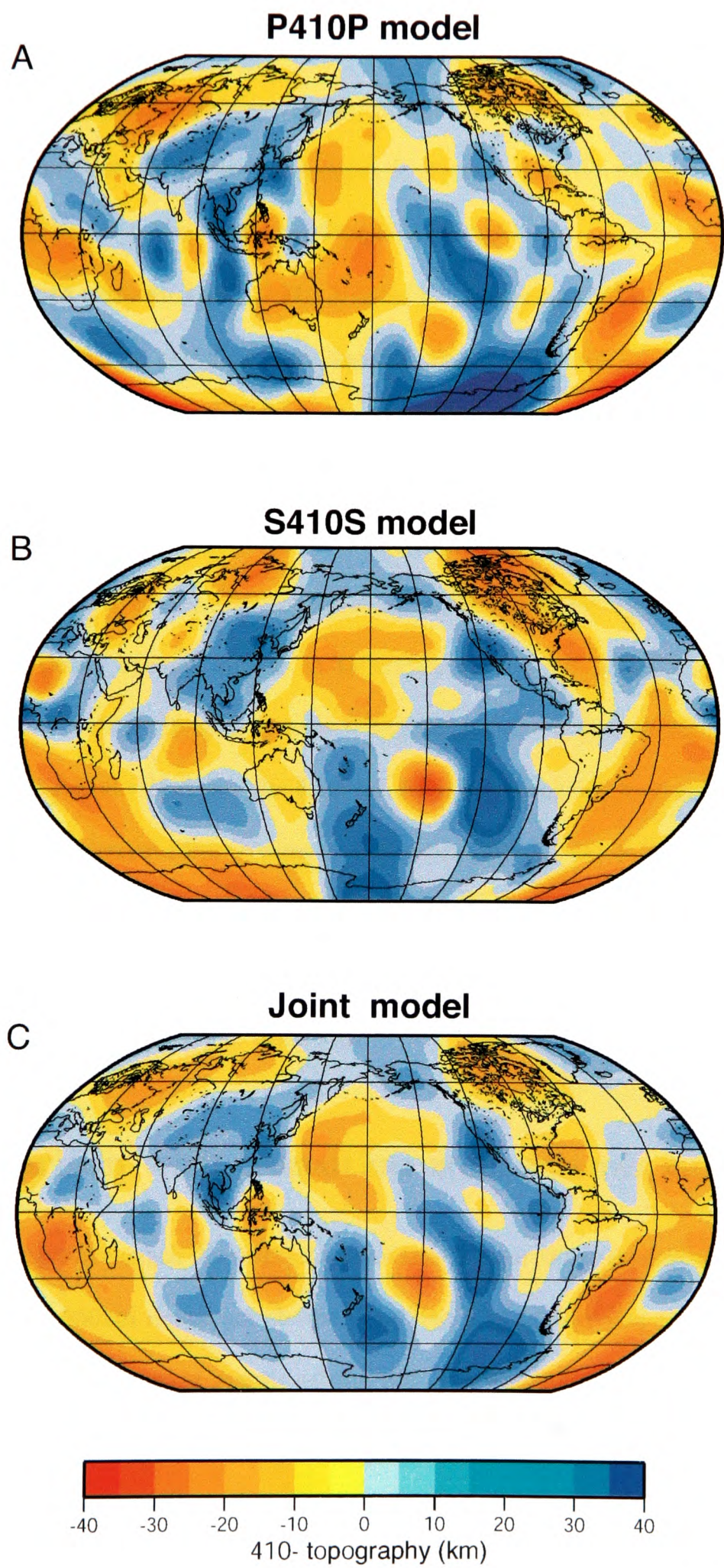


Figure 9.3: Topography of the 410km discontinuity expanded in degree-12 spherical harmonics. **A** *P410P* model, **B** *S410S* model, **C** Joint model. Topography is measured from the dataset mean these are 409.6 km, 408.6 km, and 409.2 km for the *P410P*, *S410S* and Joint models respectively. Negative topography indicates an elevated discontinuity.

Shearer (1998a); Gu *et al.* (1998)). Figure 9.3 shows highs beneath Indonesia, Northern Canada and Greenland. Beneath Asia, a region of discrepancy between previous studies, the topography models agree with Flanagan & Shearer (1998a) in having the 410km discontinuity deeper than average. However, in contrast to both Flanagan & Shearer (1998a) and Gu *et al.* (1998) the 410km discontinuity beneath the Northwest Pacific is elevated.

Although the maps show an uplifted 410 in some shield regions (Western Australia, Canada/Greenland, Southwest Africa and Northern Europe) and a depressed 410 beneath some oceans (Southeastern Pacific and Southern Indian Ocean) there are exceptions, notably the Northwest Pacific and Northern Asia. Thus, the results do not agree with the proposal of Gu *et al.* (1998) that 410-topography is strongly correlated with surface features.

In figure 9.4 the power spectra for the three spherical harmonic expansions are shown. In all the models the power spectra appear to be dominated by low spherical harmonic degrees. Both the *P410P* and *S410S* topography have a large degree-2 component and the *S410S* shows a moderate degree-1 signal. However, the Joint model shows large degree-1 and degree-2 signals. This contrasts with previous studies (Flanagan & Shearer (1998a); Gu *et al.* (1998)), which show topography dominated by a degree-1 signature with a relatively weak degree-2. The decrease in the power spectrum above degree-5 suggests that the degree-12 expansion of the topography surface is sufficiently representative of the data.

9.3 Comparison with transition zone seismic tomography

In figure 9.5 the long wavelength discontinuity structure for our three models and mantle seismic velocity anomalies at 550km depth are shown. The depth of 550km was chosen for comparison to minimise the intrusion of upper mantle seismic velocity anomalies. Furthermore the effects of chemical heterogeneities (particularly water) are less pronounced in upper mantle olivine compared to transition zone wadsleyite and ringwoodite (Jacobsen *et al.* (2004); Smyth *et al.* (2004)).

Figure 9.5 shows there is a moderate negative correlation between seismic velocities and to-

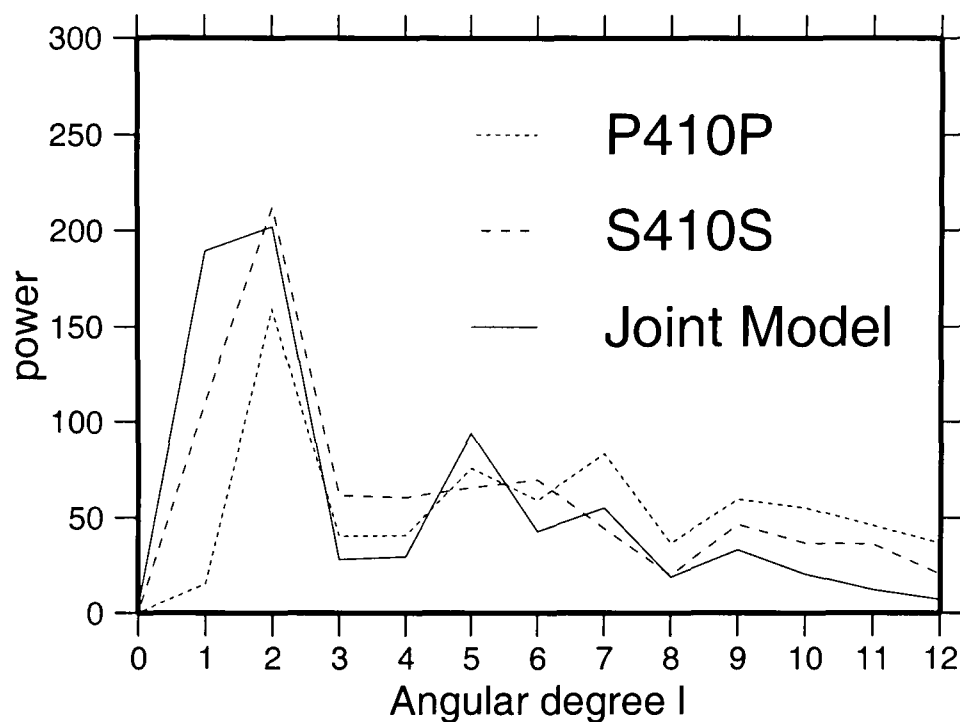


Figure 9.4: Amplitude spectra for the 3 spherical harmonic expansions of 410-topography. The power is given as an average for each angular degree.

pography of the 410km discontinuity at very low spherical harmonic degrees. The correlation coefficient of the spherical harmonic coefficients for the Joint and S20RTS models is -0.48, which is significant at the 81% confidence level. However, this correlation coefficient is probably an underestimate as the topography models are biased toward regions of better data coverage, such as the Northwest Pacific and Northeast Asia, where a topographic high matches a low velocity anomaly.

As higher degrees are included the negative correlation of topography and seismic velocity is reduced (-0.36 and -0.27 for degrees to $l = 4$ and $l = 6$ respectively). In the topography models, the discontinuity high moves into the North Pacific, while the matching fast anomaly in S20RTS stretches southwest to include the region near Tonga. The disappearance of the negative correlation could be due to competing influences from temperature and chemical heterogeneity on the topography of the 410km discontinuity and transition zone seismic velocities, when shorter wavelengths are considered.

In figure 9.6 the values of truncated spherical harmonic expansions for S20RTS and our Joint model interpolated at the locations of our caps are compared. As with figure 9.5 a negative corre-

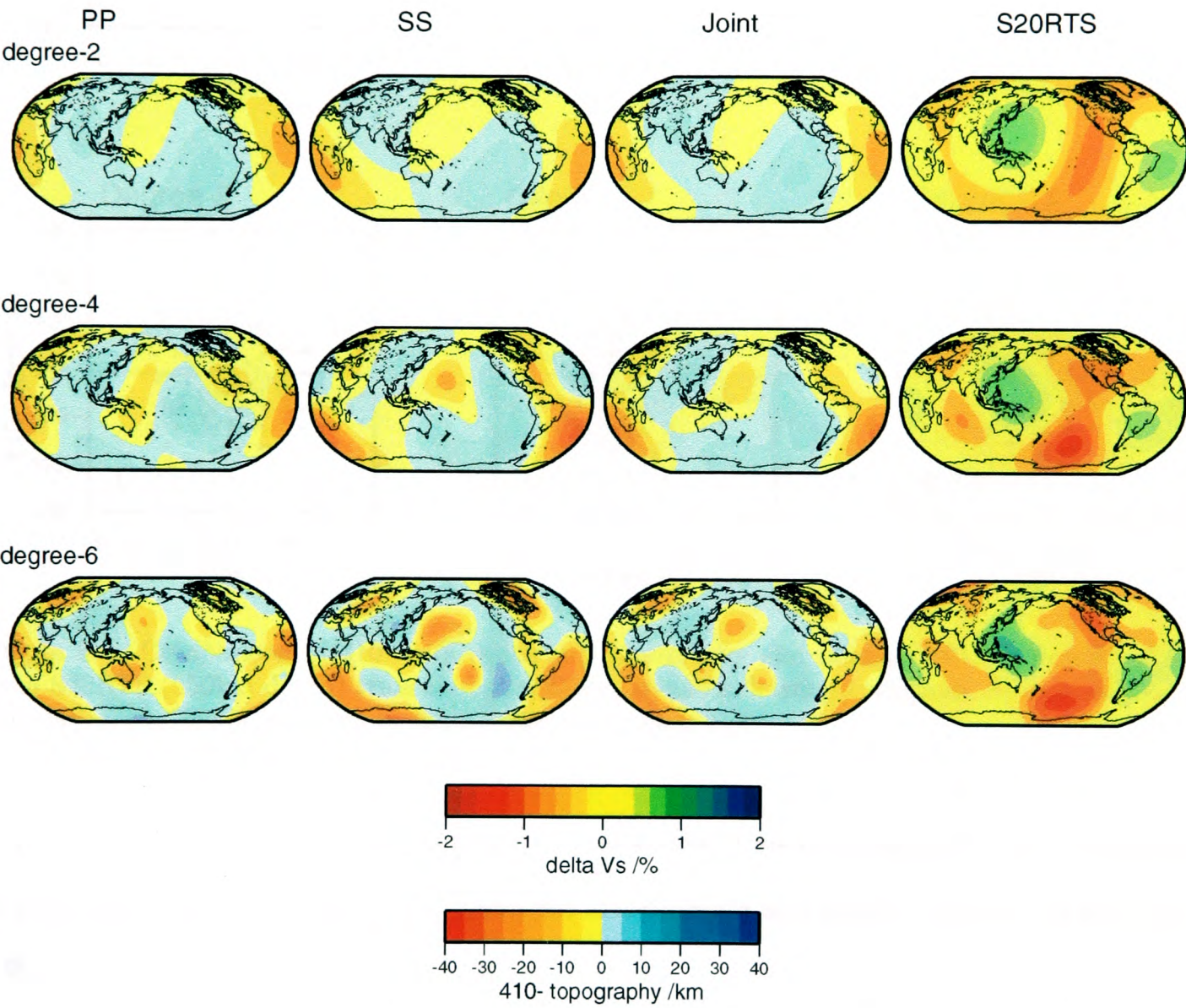


Figure 9.5: Topography of the 410km discontinuity as measured using the P410P, S410S, and joint inversion models expanded in degree-2, 4 and 6 spherical harmonics. We also show maps of S-wave tomography model S20RTS (Ritsema *et al.* (1999)) at 550km depth.

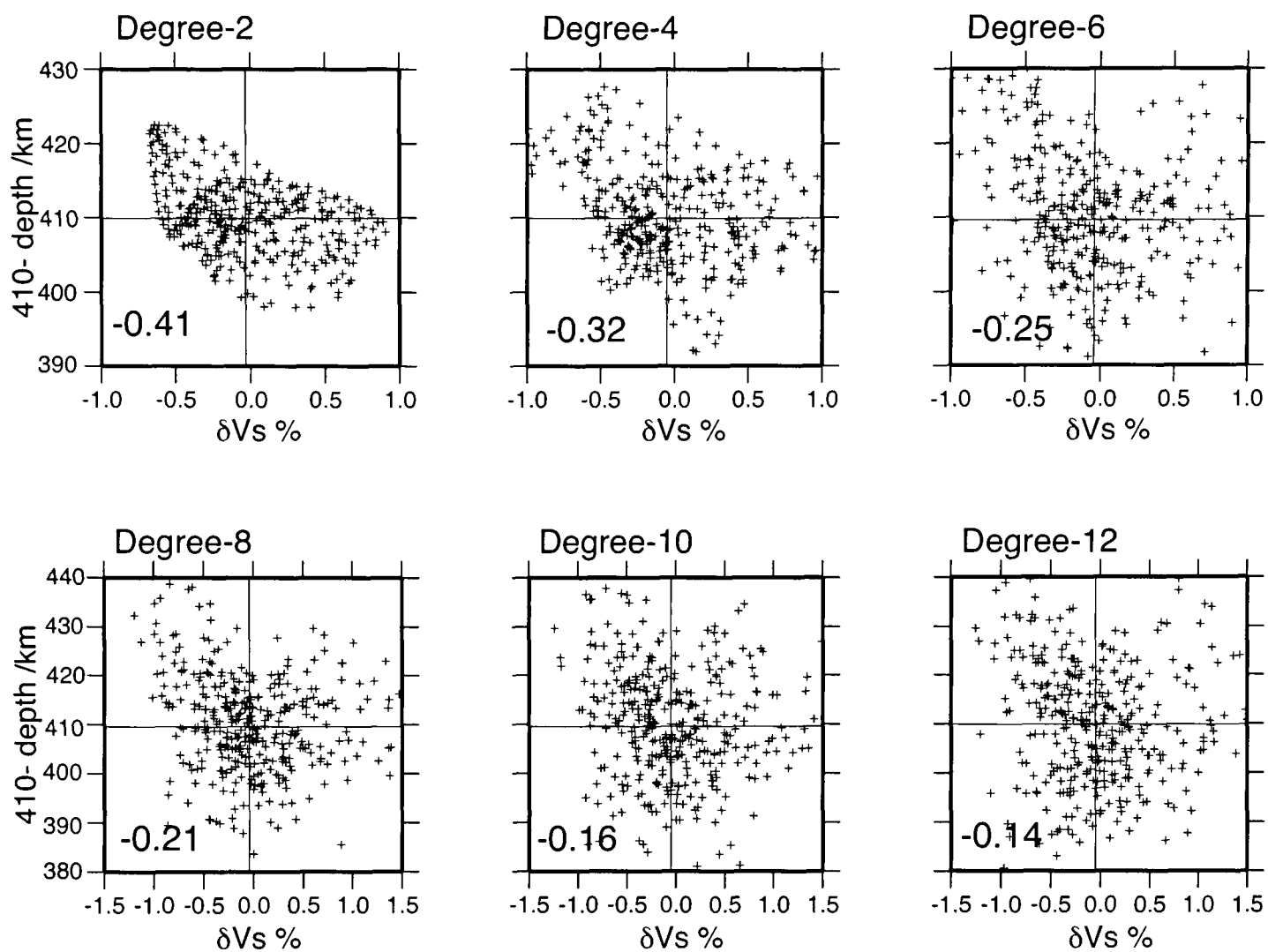


Figure 9.6: Comparison of 410km discontinuity topography and seismic velocity perturbations from S20RTS (550km depth) truncated at different spherical harmonic degrees. Spearman's rank correlation coefficient is shown in the bottom left of each plot.

lation exists between discontinuity topography and seismic velocity at degree-2, which is reduced as higher degrees are included in the expansions of topography and seismic velocity perturbations.

In figure 9.7A a comparison of 410-depth measurements sampled from the Joint model (degree-12) at the cap locations with seismic velocity anomaly at 550km are shown. The lack of correlation suggests there is no simple global correlation between the discontinuity depth from our Joint model and seismic velocity anomalies.

In figure 9.7B discontinuity depths determined from the $P410P - PP$ and $S410S - SS$ observations (with standard errors) are compared with velocity anomalies, for 4 groups of caps in the Western Pacific and Northeast Asia. The data are fit with Chi-squared minimising regression

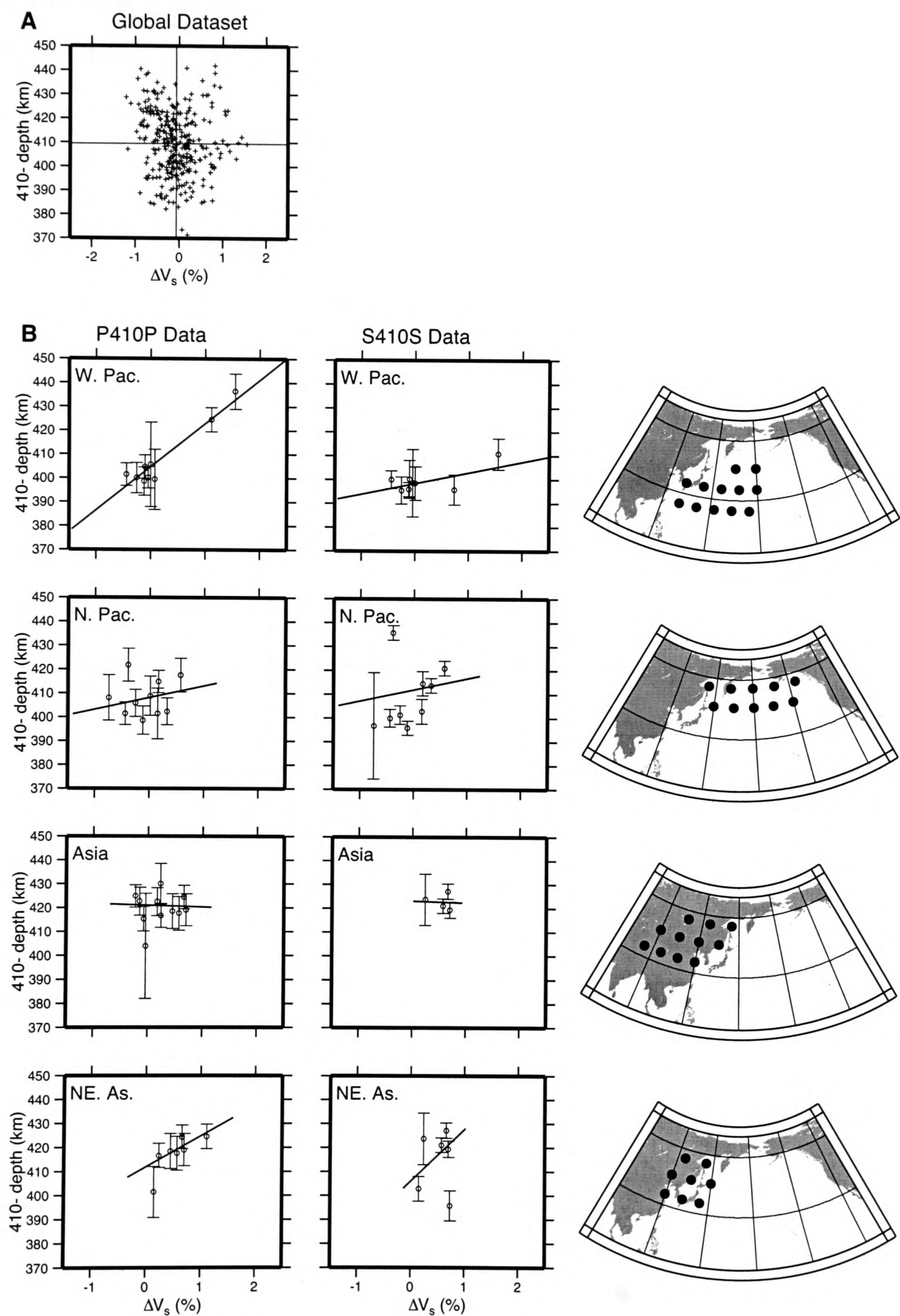


Figure 9.7: **A** Comparison of discontinuity depth in the joint model measured at each of the 410 caps with the corresponding seismic velocity anomaly from S20RTS at 550km depth. The cross is centred on the mean of both datasets. **B** Discontinuity depth as measured using *P410P* and *S410S* with seismic velocity anomaly at 550km depth for 4 overlapping groups of caps in Northeast Asia and the Western Pacific. The central cap points are shown on the right. Black lines mark the Chi-squared best fit regression lines for each pair of variables. For comparison with discontinuity depth measurements velocity anomalies were averaged over a radius of 10° about the cap centre.

lines. The slopes of the regression lines for three of the four groups (Western Pacific, Northern Pacific, Northeast Asia) suggest a positive relationship between discontinuity depth and seismic velocity anomaly. The remaining region (Asia) shows a best fit slope of near zero. These are the opposite of what would be expected from a discontinuity and seismic velocity controlled purely by temperature. This contrasts with the observation of a negative correlation between discontinuity topography and seismic velocity at longer wavelength (see above).

The magnitude of the slopes for discontinuity topography against seismic velocity are dependent on the magnitude of seismic velocity anomalies in S20RTS. As previously discussed it is likely that the seismic velocity anomalies in smooth tomography models such as S20RTS are under estimates. However, an increase in the magnitude of velocity anomalies in S20RTS would only change the magnitude of the best fit slopes not the sign.

9.4 Analysis of 410km discontinuity topography

9.4.1 Is discontinuity topography required to explain precursors?

The variations in $P410P - PP$ and $S410S - SS$ lag times show a considerable resemblance to the predicted upper mantle and crust two way travel times at the cap locations. However, the predicted upper mantle and crust travel times are not sufficient in magnitude to fully explain the amount of variation seen in the lag times. Such an explanation would require travel time contributions from the tomographic model to be approximately doubled, which would probably render S20RTS incompatible with the surface wave, overtone and arrival time data on which it is based.

One way to allow the travel time corrections for $S410S - SS$ and $P410P - PP$ to be doubled whilst preserving the velocity anomalies in S20RTS may be to allow for upper mantle seismic anisotropy. The primary constraint on upper mantle velocities in S20RTS are Rayleigh wave observations, which in turn are most sensitive to vertically polarised shear wave velocities (V_{SV}). Whereas the $S410S - SS$ and $P410P - PP$ are sensitive to horizontally polarised S-wave (V_{SH}),

and vertically polarised P-wave velocities (V_{PV}) respectively. So, a doubling of the ratio $\frac{\delta V_{SH}/V_{SH}}{\delta V_{SV}/V_{SV}}$ could explain the precursor lag times without any need for discontinuity topography.

The presence of global radial anisotropy (transverse anisotropy with a radial symmetry axis) in the upper mantle is well known and has been included in to several radially symmetric reference models. PREM allows for an average $\frac{\delta V_{SH}/V_{SH}}{\delta V_{SV}/V_{SV}}$ ratio of 1.03 above 220km which has been shown by Ekstrom & Dziewonski (1998) to be representative of most of the upper mantle. The major exception being the centre of the Pacific plate where there is a discrepancy between $\delta V_{SH}/V_{SH}$ and $\delta V_{SV}/V_{SV}$ of up to 7%. These differences fall well short of the factor of two required to explain precursor lag times. Although reference models such as PREM, and tomographic inversions like those of Ekstrom & Dziewonski (1998) are damped towards an minimum smooth model. There may be other models which allow for a doubling of $\frac{\delta V_{SH}/V_{SH}}{\delta V_{SV}/V_{SV}}$ while still satisfying the surface wave observations.

The affect of azimuthal anisotropy is a complicated function of relative orientations of, the precursor stack data, the data used to make S20RTS, and the direction and symmetry system of the anisotropy. Assuming a dataset with a relatively unbiased azimuthal distribution throughout most of the upper mantle, the effects of azimuthal anisotropy should be averaged out in tomographic models such as S20RTS. The same may also be true for the observations of precursor lag times, depending on the symmetry system and the source-receiver distribution. In this ideal scenario the required difference between $\delta V_{SH}/V_{SH}$ and $\delta V_{SV}/V_{SV}$ reduces to a requirement for radial anisotropy.

Even in the less ideal scenario of a biased azimuthal distribution, we would require the pattern of upper mantle anisotropy vectors to be such that, at each cap the $\delta V_{SH}/V_{SH}$ value in the average stack direction is approximately double the azimuthally averaged $\delta V_{SV}/V_{SV}$ value given by S20RTS. Whilst not impossible, this seems unlikely, as it requires the experimental geometry and global variation in azimuthal anisotropy to conspire against us at most if not all the 410 spherical caps. It also seems unlikely that a global model for azimuthal anisotropy which satisfies this criterion would be compatible with existing models for upper mantle azimuthal anisotropy (such as Debayle *et al.* (2005)), or the global average values for $\frac{\delta V_{SH}/V_{SH}}{\delta V_{SV}/V_{SV}}$ in PREM.

A thorough treatment of the influence of seismic anisotropy on precursor lag times and tomographic models is beyond the scope of the current study. However, from the discussion above it seems that it is difficult to explain $P410P - PP$ and $S410S - SS$ lag times with a flat discontinuity and variations in upper mantle seismic velocities. The magnitude of discontinuity topography observed is dependent on the magnitude of velocity variation in the upper mantle, but the pattern of topography will remain unchanged if a reasonable amount of error in the estimation of velocity perturbations in the upper mantle is allowed for. Accordingly, in subsequent sections only the patterns of discontinuity topography and seismic velocity anomalies are interpreted rather than their magnitudes.

9.4.2 Precursor derived discontinuity topography

Discontinuity depths produced using the individual datasets correlate well in regions of dense sampling. This allows production of a preferred model for the topography of the 410km discontinuity which is constrained using both $P410P$ and $S410S$ measurements. The joint inversion for spherical harmonic coefficients using two datasets allows the model to be better constrained in regions where sampling is poor in either the PP or SS dataset. As with previous studies the topography of the 410km discontinuity is found to be dominated by low spherical harmonic degrees. The larger contribution of degree-2 structure compared to previous models (Flanagan & Shearer (1998a); Gu *et al.* (1998)) is presumably due to the observation of an uplifted 410 beneath the Northwest Pacific, a feature which is not retrieved in these studies.

The topography models suggest that the 410 is deeper in the region south and west of Japan. This appears to contrast with previous studies of shorter period data (Collier & Helffrich (1997); Thiot *et al.* (1998)), which have found an uplifted 410 in this region. However, differences can probably be explained by the way long-period seismic data (such as that used here) ‘sees’ the mantle. Since the area any scattered phase ‘sees’ (is sensitive to) is a weighted average over its Fresnel zones. Short-period studies with their much smaller Fresnel zones will retrieve uplifted regions of smaller spatial extent, while the long-period precursor phases used here will retrieve

a more regional trend. Accordingly, the topography models reflect a broader regional trend of discontinuity topography, rather than the more localised uplift retrieved in studies of shorter period data.

Differences between observations of discontinuity topography presented here and those of previous precursor based studies could result from differences in the travel time corrections for crustal and mantle structure. However, the corrections to the $S410S - SS$ lag times are similar to those applied by Flanagan & Shearer (1998a), so it is unlikely that this is a major cause of discrepancies between the two models for 410-topography. It is more likely that differences are due to the method of determination of precursor arrival times.

The topography models presented here represent the long wavelength structure of the 410km discontinuity better than previous studies for two reasons. Firstly, the Joint model is constrained using both $P410P - PP$ and $S410S - SS$ differential times. Secondly, the arrival times of the $P410P$ and $S410S$ pulses were measured using a cross correlation with the surface reflection. Picking arrival times by wave shape is preferred to picking the maximum of the precursor pulse (Flanagan & Shearer (1998a, 1999)) which will be subject to errors due to the noise content of the stacked traces.

9.4.3 Discontinuity topography and seismic velocities

The models of 410km discontinuity topography were compared with transition zone seismic velocities at 550km depth, which is taken to be representative of the transition zone shear velocity. This assumes that any temperature or chemical heterogeneities are continuous from the discontinuity into the transition zone.

At very low spherical harmonic degrees there is a negative correlation between discontinuity topography and transition zone seismic velocities. However, as higher spherical harmonic degrees are included this simple relationship disappears. When the slopes of regression lines for discontinuity topography against transition zone seismic velocities were calculated for the 4 regions of figure 9.7 zero and positive correlations were seen.

The weak negative correlation between discontinuity topography and seismic velocity seen at long wavelength (degree-2) is interpreted as evidence for the influence of temperature on the phase boundary position. This could either represent a true long wavelength pattern in the mantle or it could constitute the long wavelength part of the expected signature near subducting slabs. The inferred long wavelength temperature pattern is consistent with cold down-wellings in the Western Pacific being fed by an up-welling in the Southeast Pacific. The less than perfect correlation (-0.48) between long wavelength discontinuity topography and seismic velocities could be due to the presence of chemical anomalies.

The disappearance of the negative correlation between discontinuity topography and seismic velocity when higher degrees are considered, and the positive correlations in the regions of figure 9.7, requires the presence of chemical heterogeneities (water, iron and aluminium) in the upper transition zone. However, not all the areas examined in figure 9.7 suggest there is such a simple positive relationship. It seems that the combined effects of thermal and chemical heterogeneities creates a complex pattern of discontinuity topography and seismic velocity anomalies at shorter wavelengths, which does not exist at longer wavelengths.

The domination of the topography spectra at different degrees by thermal or chemical effects requires that thermal and chemical anomalies in the mantle are described by different patterns. This conclusion is summarised in the cartoon in figure 9.8, where the spectrum of phase boundary topography is considered a sum of spectral contributions from patterns of thermal, and chemical heterogeneity. This in turn is viewed through the filter of precursor sensitivity; as a result the large contribution to discontinuity topography expected from the narrow limbs of mantle convection cells is not retrieved in our observed phase boundary topography. However, the precursor phases are able to image low degree discontinuity structure which is interpreted as being largely due to thermal effects, and higher degree features which have a greater contribution from chemical anomalies.

Different spectra for thermal and chemical heterogeneities in the mantle suggests that it may be possible to separate the two patterns. However, before this is achievable better knowledge of

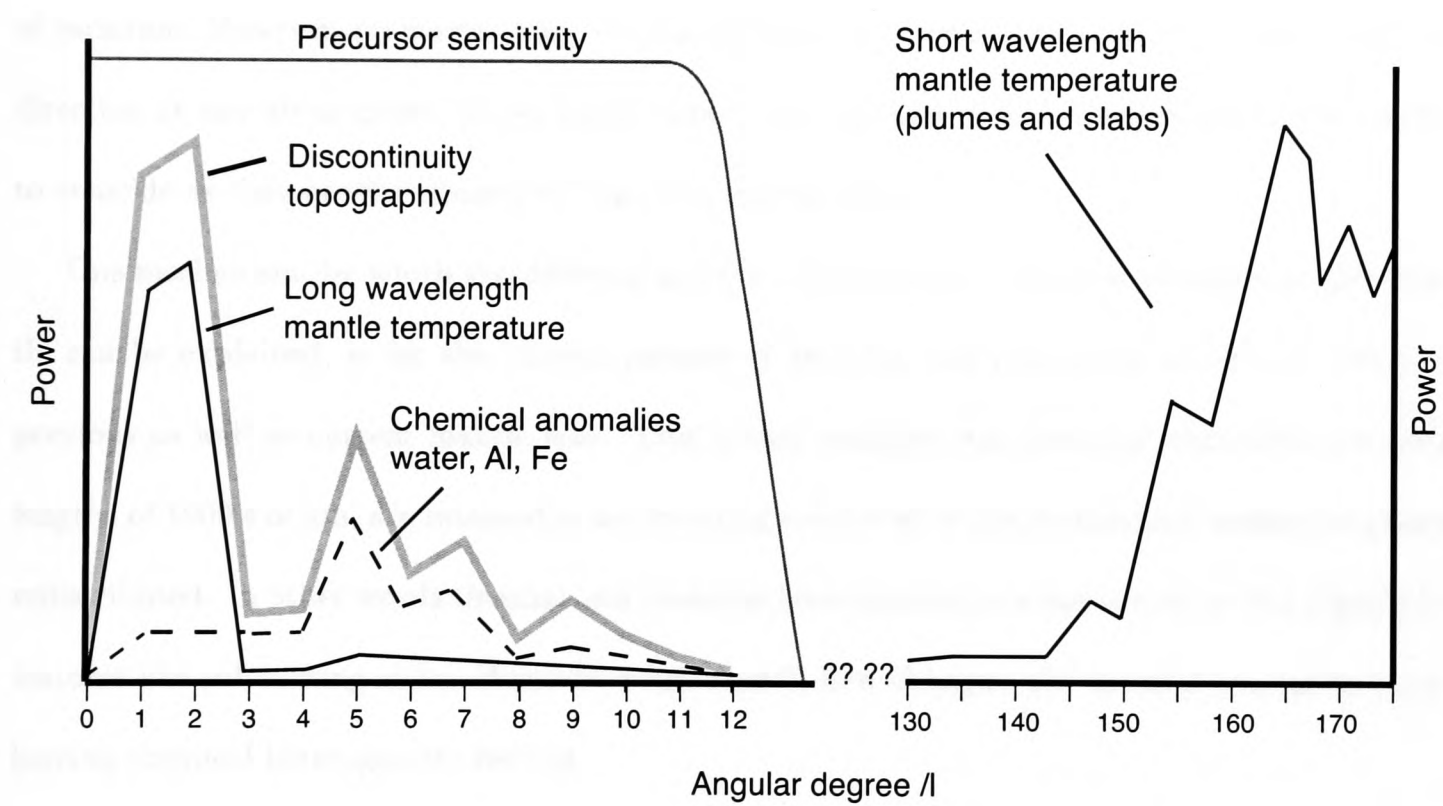


Figure 9.8: Sketch illustrating our interpretation of the discontinuity topography spectrum for the 410km discontinuity. Note that our results only require the discontinuity topography spectrum to be a function of temperature at long wavelengths where as at shorter wavelengths the contributions of chemical anomalies become more important. Details, such as the relative contributions of thermal and chemical anomalies to the discontinuity topography are speculative.

seismic velocity anomalies and discontinuity topography, as well as the relationships between the two in the presence of chemical heterogeneities is required.

9.4.4 Consequences for mantle circulation

The different spectra for mantle thermal and chemical anomalies has important consequences for the understanding of mantle processes. The results presented here require that mantle thermal and chemical anomalies vary over different scale lengths and consequently have different patterns of variation. However, no matter what the forces that create mantle flow are it can only go in one direction at any given point. So we might expect thermal and chemical anomalies in the mantle to coincide as they are distributed by the same mantle flow.

One mechanism, by which the different spectra of thermal and chemical variation in the mantle can be explained, is for the current pattern of chemical heterogeneities to be a function of previous as well as current mantle flow. This in turn requires that chemical anomalies, on scale lengths of 1000's of km, are retained in a convecting mantle after the thermal anomalies have been redistributed. In other words thermal and chemical heterogeneity are introduced to the mantle by features like subducting slabs. However, when mantle flow changes, the thermal anomalies decay leaving chemical heterogeneity behind.

Chapter 10

Concluding remarks part I

10.1 Introduction

In this part of the thesis the precursors to surface reflection phases (PP and SS) were studied. It was shown that by aligning and stacking large numbers of traces on the surface reflections a variety of upper mantle and transition zone discontinuities can be imaged. In particular, the application of a deconvolution algorithm allowed accurate measurement of the arrival times and amplitudes of reflections from the 410km discontinuity. As the preceding chapters show, the properties of the 410km discontinuity place important constraints on the nature of heterogeneity in the mantle.

10.2 Observations of upper mantle and transition zone reflectors

The stacks, made over broad regions in chapter 6, show upper mantle reflections from around 220 and 300km depth. These reflectors are not seen in a global stack, nor in all of the regional stacks. So neither of these reflectors are global features. It is also likely that the quality of these reflections is degraded by the stacking of traces over broad regions.

In the mid-transition zone a discontinuity at 520km is seen in the global and regional stacks.

This discontinuity is most likely related to the $\beta - \gamma$ olivine phase change. In certain regions ‘splitting’ of the 520km discontinuity is observed, mostly in the *SS* dataset. The separation of the 520km discontinuity is often interpreted as being due to variations in depth of a garnet to Ca-perovskite phase transition, which in some places occurs at the same depths as the $\beta - \gamma$ olivine phase change, while in others it produces ‘splitting’.

Splitting of the 520km discontinuity is not as apparent in the *PP* dataset. The observation of the *S520S* ‘splitting’ in similar regions using a dataset with similar wavelength suggests that this cannot be due to incoherent stacking effects. So it seems more likely that the absence of splitting in the *PP* dataset is due to the phase transitions in the non-olivine mantle components, which are believed to cause the ‘splitting’, being more-transparent to P-waves.

Clear S-wave reflections from a discontinuity at 660km depth are seen in every stack made using this dataset. Yet no *P660P* phase is reliably observed, despite the suggestion from stacks of synthetic seismograms (based on PREM) that it should be observable. The clear *S660S* phase requires that the absence of the *P660P* phase cannot be due to incoherent stacking discontinuity topography or focusing effects, because the two datasets have a similar effective wavelength and sampling (bounce point) distributions. This suggests that the P-wave impedance contrast across the 660km is low enough to make *P660P* reflections undetectable, and that the P-wave seismic discontinuity in PREM is overstated.

10.3 Properties of the 410km discontinuity

In chapter 8, it was shown that the amplitude of *P410P* phases varies throughout the northern Pacific. When the global pattern of *P410P* and *S410S* amplitude variations was examined, it was found that *S410S* amplitudes also show lateral variation but over shorter scale lengths than for *P410P*.

Reflection amplitudes are complicated functions of many factors including, the discontinuity thickness and impedance contrast, differing intrinsic attenuation between the reflected and refer-

ence phases, and focussing due to mantle structure. Since similar variations are not observed in *P410P* and *S410S* amplitudes, the variations in reflection amplitude cannot be fully explained using discontinuity thickness, differing intrinsic attenuation or focussing due to mantle structure. Some of the variation observed in precursor amplitudes must be due to differences in the impedance contrast of the 410km discontinuity, which in turn is a (complex) function of transition zone chemical and thermal heterogeneity.

Using measurements of *P410P* and *S410S* amplitudes regional average P- and S-wave impedance contrasts for the 410km discontinuity were calculated. These values are typically lower than those of previous studies, making them more consistent with a low-olivine piclogite model for mantle composition. However, a high olivine pyrolitic mantle could still be consistent with our results if hydration of transition zone wadsleyite is taken into account.

In chapter 9 the arrival times of *P410P* and *S410S* phases relative to the surface reflections was examined. It was shown that it is difficult to explain the variation in precursor lag times without some topography on the 410km discontinuity. Although the magnitude of discontinuity topography is dependent on the magnitude of velocity variation in the upper mantle, the pattern of topography will remain unchanged if a reasonable amount of error in the estimation of velocity perturbations in the upper mantle is allowed for.

Taking the velocity anomalies in the tomographic model S20RTS at face value three new maps of global topography on the 410km discontinuity were generated. The first two maps relate to the separate *P410P* and *S410S* datasets. These correlate well in regions of good coverage. The third map is the preferred model for discontinuity topography, which was generated using *P410P* and *S410S* depth measurements jointly to constrain global discontinuity topography.

To interpret global maps of 410-topography in terms of chemical and thermal heterogeneity, they were compared with maps of seismic velocity anomalies in the transition zone. At very long wavelengths (up to degree-2) seismic velocity correlates negatively with 410-depth. However, more detailed comparisons for some small regions show positive relationships between seismic velocity and 410-depth.

The relationship between long wavelength topography of the 410-discontinuity and seismic velocities suggests these are primarily due to thermal variations. This is consistent with the influence of temperature dominating 410-topography, as the power spectrum for discontinuity topography is dominated by degree-1 and 2 components. The presence of positive relationships between 410-topography and seismic velocities suggest that at short wavelengths mantle chemistry becomes more important in determination of 410-topography and seismic velocities.

10.4 What the 410km discontinuity can tell us about the mantle

To explain the results the pattern of 410-topography (or seismic velocity) is considered as a sum of responses due to thermal and chemical anomalies. So the topography map (figure 9.3) is the sum of two maps which reflect the distribution of thermal and chemical heterogeneities in the mantle. The same is also true for the variations in P_{410P} and S_{410S} reflection amplitudes. However, no clear relationship is seen between the reflection amplitude and seismic velocity anomalies, so the precursor amplitudes cannot be analysed in the same way as the discontinuity topography.

The domination of the topography spectra at different wavelengths by thermal or chemical effects requires that thermal and chemical anomalies in the mantle are described by different distributions. However, this seems at odds with the distribution of heterogeneities by mantle flow. Resolution for this probably requires that chemical heterogeneities are much longer lived than thermal heterogeneities in the transition zone. So, both thermal and chemical anomalies are distributed by the same patterns of mantle flow. However, the thermal anomalies would disappear over time (i.e. blend in with the rest of the mantle), while chemical anomalies are preserved due to their slower diffusion rate. Accordingly the pattern of thermal anomalies in the mantle would outline the current and recent patterns of mantle flow, whereas the pattern of chemical anomalies would also include contributions from long past mantle flow patterns, which gives rise to the different distributions of chemical and thermal heterogeneity.

Part II

Lower mantle reflectors and S-wave scattering

Chapter 11

An introduction to the lowermost mantle

11.1 Overview and motivation

Over the last 20 years there has been a growing body of seismological evidence that the base of the mantle (hereafter termed D'' in reference to Bullen's scheme) is unusual compared to the rest of the lower mantle. Numerous studies have indicated the presence of anomalous velocity gradients, ultra-low velocity zones, and scattering bodies in the bottom 200-400km of the mantle (e.g. Wright & Lyons (1980); Lay & Helmberger (1983); Lay & Young (1996); Bilek & Lay (1998); Revenaugh & Meyer (1997); Helmberger *et al.* (2000); Wen *et al.* (2001); Thomas *et al.* (2004b,a); Garnero *et al.* (2004); Braña & Helffrich (2004); Wallace & Thomas (2005)).

It might be expected that in a convecting system the most heterogeneous regions are those near the boundaries. Over a hundred years of geology and geophysics have characterised the processes which operate in the mantle's upper boundary layer (the lithosphere). However, the processes which operate at the base of the mantle have received less attention, despite their importance in the long term evolution of the mantle. For example the region is thought to contain (among other

things): a thermal boundary layer, a chemical boundary layer, a phase change, a source region for plumes, and a grave yard for subducting slabs.

One reason why the base of the mantle has received relatively little attention is that there is no direct way of observing it. Instead we can only gain information on the structure in the lower mantle by observing seismic waves which are sensitive to this region. Current seismological techniques either use amplitudes and travel times of high frequency waves to investigate structure over small regions, or use global travel time datasets to constrain global structure. Here a migration method, which is sensitive to short wavelength features but is applicable to large datasets on a global scale, is developed and applied. Such techniques are required in order to truly establish the features and processes in the lower mantle.

11.2 Seismological techniques for studying the lowermost mantle

Previous seismological research on the lowermost mantle has focussed around 3 main areas. The first, global seismic tomography, maps out variations in velocity structure using body wave travel times and normal mode observations. Velocity structure within the D'' layer is particularly constrained by core diffracted phases, however many models also constrain the lowermost mantle using core traversing phases such as SKS and SKKS (Liu & Dziewonski (1994, 1998); Castle *et al.* (2000)).

The second source of seismological information on the lower mantle comes from studies of seismic anisotropy. Although anisotropic parameters can be retrieved from the analysis of waveforms (e.g. Garnero *et al.* (2004)), the more usual technique is to measure shear-wave splitting which is then modelled as a fast orientation direction and magnitude (for example see Kendall & Silver (1996, 1998) and Kendall (2000)).

The third line of seismological investigation of structure in the lowermost mantle is the analysis of reflected or scattered waves. Figure 11.1 shows examples of ray path geometries including the

SdS phases seen in the S-wave coda, which is the main topic of this part of the thesis.

The observables of scattered/reflected phases are the arrival time and the amplitude. The arrival time together with a velocity model constrains the scatterer/reflector location, however, the amplitude of scattered phases is a more sensitive measure of discontinuities in elastic parameters. To minimise errors associated with upper mantle structure, earthquake location and timing, the arrival times and amplitudes of scattered phases are typically measured relative to a reference phase which travels a similar path in the upper mantle (figure 11.1).

Figure 11.1 A shows an example of a *PdP* or *SdS* reflection although it could equally apply to a conversion (*PdS* or *SdP*). *SdS* reflections at large epicentral distances ($\sim 60-75^\circ$) are often large enough to be observed in individual long-period records. Early attempts to image lower mantle structure using reflected phases, such as Gaherty & Lay (1992) and Kendall & Shearer (1994, 1995), assumed that the *SdS* phase is a reflection from a horizontal planar boundary. However, in the case of non-horizontal boundaries or scatterers, this will map incorrect depths back to the source-receiver mid point.

Several recent studies of *PdP* and *SdS* phases have used seismic migration techniques to image lower mantle structure. These studies either, use high frequency data from seismic arrays leading to a high resolution ($\sim 20\text{km}$) but a limited spatial coverage (for example see Thomas & Webber (1997); Kito & Kruger (2001); Thomas *et al.* (2004b)), or use lower frequency data leading to better spatial coverage with lower resolution ($\sim 500\text{km}$, Lay & Young (1996); Bilek & Lay (1998); Castle & Creager (1999)).

Figure 11.1 B shows the ray path geometries used for the interpretation of precursors to *ScS* reverberations. This technique has the advantage of being uniformly sensitive to discontinuities at all depths. However, from the figure it can be seen that the reflected phases have a steep angle of incidence, and will have a small reflection amplitude, so when analysing this type of data traces are grouped into corridors which have common sources and receivers and averaged to improve the signal to noise ratio. Due to this stacking process and the presence of *ScS* multiples lateral variations, in a discontinuity along a corridor will be averaged over. There is also a trade off

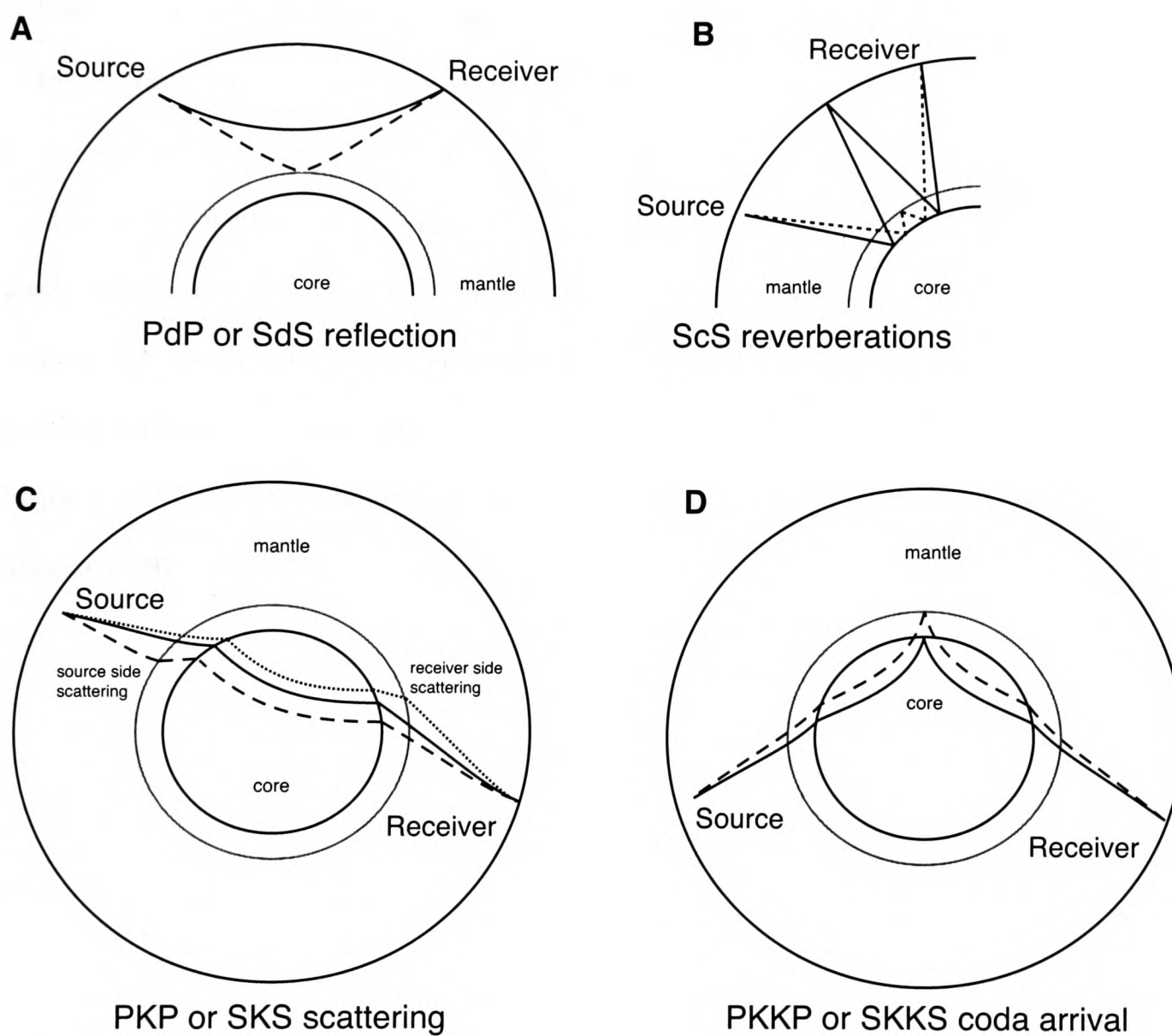


Figure 11.1: Techniques for analysis of phases scattered/reflected in the lower mantle. Scattered/reflected ray paths are shown in dashed or dotted lines. The reference phase is give as a solid line. **A** $PdP/SdS/SdP/PdS$ phases measured relative to the direct P or S phase. **B** $ScS_n d ScS_n$ measured relative to ScS_n , where n is the number of ScS legs. **C** Source side (dashed) and receiver side (dotted) scattering of the PKP or SKS phase. **D** Reflected/scattered arrivals following the $PKKP$ and $SKKS$ phases.

between the presence of upper and lower mantle discontinuities.

A ray diagram for the scattering of *PKP* and *SKS* phases is shown in figure 11.1 C. The scattered arrivals can be detected by the stacking of high frequency data usually from seismic arrays or regional networks (Thomas (1999); Hedlin *et al.* (1997); Vidale & Hedlin (1998)), or in lower frequency data by their interference with the *PKP* or *SKS* phase (Helmberger *et al.* (2000)). Note that in either case there exists a trade off between scattering structures on the source and receiver sides of the core.

The final technique for looking at the lowermost mantle considered here are precursors and postcursors to the inner core reflected phases *SKKS* and *PKKP* (figure 11.1 D). The *SKKS* and *PKKP* contain energy in a surrounding coda which is created by structure near the CMB and in the outermost layers of the core. As the *SKKS* and *PKKP* phases are observed at high frequencies ~ 1 Hz, the coda arrivals are modelled using statistical scattering techniques (Earle & Shearer (1997, 1998))

Chapter 12

The D'' region

12.1 Seismic properties of the D'' region

12.1.1 Global tomography in the D''

Anomalous amplitudes and travel times of seismic waves which sample the lower mantle suggest there is something distinct about this region. This can be seen in modern global tomography models such as S20RTS (Ritsema *et al.* (1999)) and SKS12WM13 (Liu & Dziewonski (1994)). Figure 12.1 shows the integrated power throughout the lower mantle for these two tomographic models, which is an indicator of the magnitude of velocity perturbations at that depth. Throughout most of the lower mantle total power is fairly constant. However, in the bottom 200-400km there is an increase, indicating a greater amount of heterogeneity at this depth. Similarly, radial velocity models such as PREM (Dziewonski & Anderson (1981)), show reduced gradients in the lowermost mantle suggesting a departure from adiabatic conditions.

Many separate tomographic studies have noted that the increased heterogeneity in the lower mantle is dominated by degree-2 structure (Liu & Dziewonski (1998); Garnero (2000); Lay & Garnero (2004)), with the slow regions under the Pacific and Africa surrounded by rings of high velocity. This is similar to the patterns of velocity heterogeneity seen in the upper mantle, where

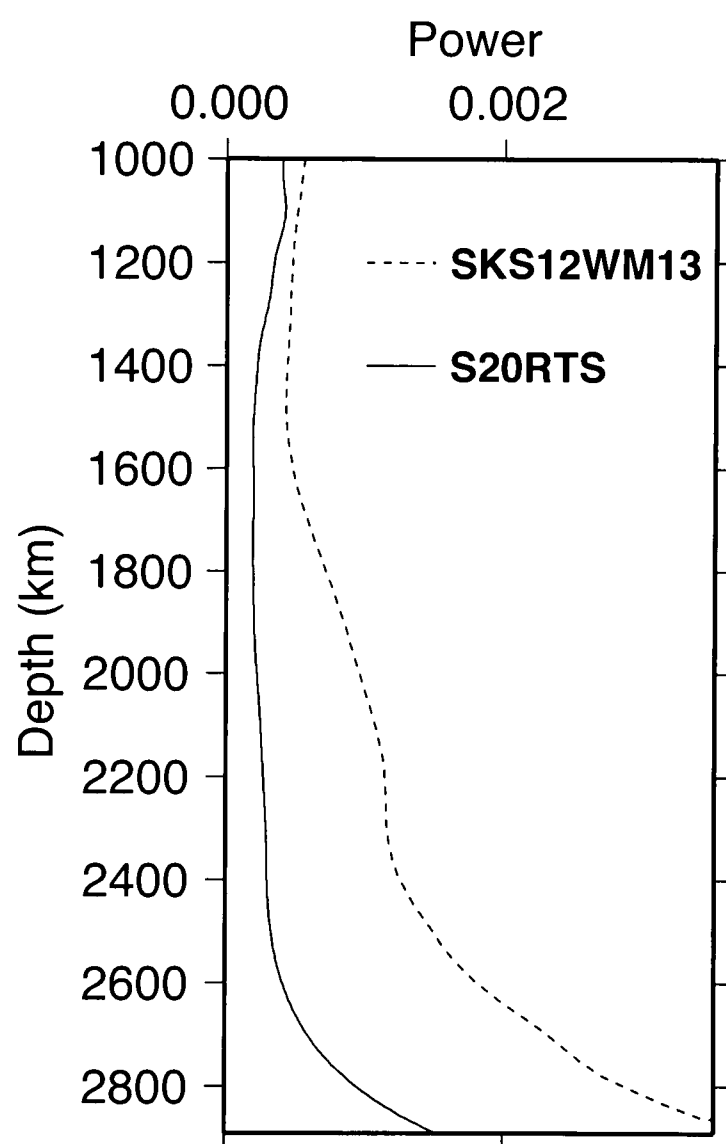


Figure 12.1: Integrated power against depth for tomography models S20RTS (Ritsema *et al.* (1999)) and SKS12W13 (Liu & Dziewonski (1994)) in the lower mantle. The base of the plot marks the CMB. Power was determined by summing the squares of the spherical harmonic coefficients.

Study	Region	Mean depth (km)	Data type
Thomas <i>et al.</i> (2002)	North Pacific	2650	PdP (SP)
Houard & Nataf (1993)	North Central Asia	2590	PdP (SP)
Kito & Kruger (2001)	Southwestern Pacific	2550 and 2650	PdP (SP)
Revenaugh & Jordan (1991d)	West Pacific and Australasia	2565	ScS (LP)
Lay & Young (1996)	Alaska	2600-2650	SdS (LP)
Bilek & Lay (1998)	North Central Asia	2605	SdS (LP)
Kendall & Shearer (1994)	Global (average)	2630	SdS (LP)
Kendall & Shearer (1994)	North Central Asia (average)	2620	SdS (LP)
Kendall & Shearer (1994)	Alaska (average)	2590	SdS (LP)
Kendall & Shearer (1994)	Australasia (average)	2645	SdS (LP)
Thomas <i>et al.</i> (2004b)	North Central Asia	2574-2684	SdS (SP)

Table 12.1: Some recent observations of a the D'' discontinuity. LP refers to long period data, SP short period. Kendall & Shearer (1994) values are averages for SdS observations within regions. For a more comprehensive list see Wyss *et al.* (1998).

velocity heterogeneity marks out the major tectonic features. Some authors regard the high velocities in D'' as due to the thermal effect of subducting slabs. However, the observed velocity heterogeneity could also be the result of chemical heterogeneity.

12.1.2 The D'' discontinuity

Some of the first authors to suggest that the top of the D'' region is marked by a first order increase in seismic velocity were Wright & Lyons (1980) (P-waves) and Lay & Helmberger (1983) (S-waves). More recent observations of a reflector on top of the D'' layer are given in table 12.1. Although a D'' reflector has been observed in many regions, it has not been observed everywhere (Revenaugh & Jordan (1991d); Kendall & Shearer (1994, 1995)). Nor is any D'' discontinuity included in global reference models such as PREM.

There is currently no consensus as to whether the D'' discontinuity is a globally present 'true surface', which is only intermittently observed due to focussing of scattered waves, or an interface created by 3-D volumetric heterogeneity (Wyss *et al.* (1998)). Many authors (for example Garnero (2004)) have noted that the D'' discontinuity is most evident beneath regions of past or present subduction.

A variety of 1D models have been created in order to explain scattered arrivals from a D''

Model	Region	Depth (km)	S-velocity Jump (%)
SYLO Lay & Young (1989)	Alaska	2648	2.75%
SLHE Lay & Helmberger (1983)	North Central Asia	2560	2.8%
SLHA Lay & Helmberger (1983)	Central Pacific	2640	2.8%
SYL1 Young & Lay (1987)	India	2610	2.8%
SGHE Garnero <i>et al.</i> (1988)	Central America	2640	0-2.8%

Table 12.2: Some of the existing models for a D'' discontinuity

discontinuity, and some of these are shown in table 12.2. These are characterised by a velocity jump of $\sim 2.8\%$, 250-300km above the CMB, and a negative or zero velocity gradient within the D'' . The majority of these models are based on the modelling of wave forms using a radial velocity profile, this type of analysis cannot retrieve boundary topography. Furthermore, the D'' is usually modelled as a first-order discontinuity when in reality it could have a thickness of up to 50km (Garnero (2000))

Several detailed array based studies have revealed significant topography on the D'' discontinuity. For example, Thomas *et al.* (2004b) has identified elevation changes of 150km over short distances (~ 700 km). Topography on the D'' discontinuity can lead to the mislocation of reflection positions, and incoherent stacking of scattered phases if a horizontal discontinuity is assumed during modelling. Also, Sidorin *et al.* (1999) have shown that a continuous reflector with extreme topography can explain many of the wave form anomalies associated with a D'' layer.

12.1.3 Scattering bodies and reflectors within the D''

As might be expected for a region as heterogeneous as the D'' there have been observations of first order discontinuities on virtually all length scales. Several studies of short period recordings have confirmed the existence of very small or point scatterers (dimensions < 100 km) within the D'' . Braña & Helffrich (2004), concluded the existence of chemical heterogeneity within the D'' based on the observation of a small/point scatterer with velocity contrasts of 0.4-0.7% and 1.4-6.6% for P- and S-waves respectively, at around 2720km depth under the North Atlantic. On a larger scale (~ 500 km) Bilek & Lay (1998) have suggested a sub D'' reflector at a similar depth (~ 2760 km)

beneath North Central Asia.

Recently there has been considerable attention paid to investigation of the thin layer (5-40km thick) adjacent to the CMB, and the Ultra-low Velocity Zone (ULVZ) it contains (Revenaugh & Meyer (1997); Garnero *et al.* (1998); Castle & van der Hilst (2000); Helmberger *et al.* (2000); Rost & Revenaugh (2003)). As the name suggests the ULVZ is characterised by reductions in seismic velocities, reductions of as much as 30% in S-wave velocity have been identified. Like the D'' reflector ULVZ's are not observed globally (Castle & van der Hilst (2000); Persh & Vidale (2004)). Although Revenaugh & Meyer (1997) have suggested a thin global low-velocity layer <15km thick could exist.

There is also a suggestion that the base of the D'' (the CMB) could display small scale topography (Earle & Shearer (1997); Rost & Revenaugh (2004)). Typical scale lengths for variation are of the order <10 km, and there are trade-offs between modelling CMB topography and overlying heterogeneity.

12.1.4 Anisotropy in D''

The top of the D'' marks a change from the largely isotropic lower mantle to a region of anisotropy. Most observations of D'' anisotropy from shear wave splitting or reflection modelling are consistent with polarisation anisotropy with a radial symmetry axis (Kendall (2000)), and typically $V_{SH} > V_{SV}$ (Karato (1998)). However, the structure beneath the Caribbean appears to be best modelled using a tilted symmetry axis (Garnero *et al.* (2004)).

Until recently, the favoured cause of the anisotropy is shape preferred orientations (SPO), such as horizontal layering or aligned inclusions (Kendall & Silver (1996, 1998)), possibly created by the presence of partial melt in subducted crust. This was favoured over the alternative, lattice preferred orientation (LPO), because perovskite which is the most abundant mineral at these depths fails to develop strain induced orientation. However, the discovery of a post-perovskite phase which could exist in the D'' has revived this possibility (see section 12.4).

12.2 D'': The mantle's other thermal boundary layer

The D'' region represents the lowermost layer of the convecting mantle, and as such it resides next to a temperature change of around 1000-1500K (Boehler (2000); Garnero (2000)). So it seems reasonable that the D'' is traditionally regarded as a thermal boundary layer for the mantle.

Many of the features of the D'' layer can be explained with reference to a thermal boundary layer. For example the presence of seismic anisotropy has been attributed to: horizontal flow within the boundary layer, structures created where subducting slabs settle near the base of the mantle, or aligned pockets of partial melt in the boundary layer (Kendall & Silver (1998); Kendall (2000); Lay & Garnero (2004)). While the complicated topography of D'', and the existence of low-velocity zones has been attributed to thermal gradients associated with downwellings (Thomas *et al.* (2004b)).

A thermal boundary layer at or within the D'' does not exclude the presence of chemical heterogeneities, and/or phase changes, in this region. After all the mantle's upper boundary layer (the lithosphere) contains a chemically heterogeneous layer (the crust). Chemical anomalies associated with subducted oceanic slab/crust material could exist within the D'' without requiring a chemical boundary layer.

12.3 D'': A layer of chemical heterogeneity

Differences in the isotopic compositions between ocean island basalts, with those from mid ocean ridges, suggest that the latter primarily sample a 'depleted' mantle which is complimentary to the extraction of the continental crust from the bulk silicate Earth. While the ocean island basalts also include regions of 'undepleted' mantle in their source regions (for example see Hofmann (1997) or Helffrich & Wood (2001)).

This requirement of separated, as well as a variety, of source regions for ocean island and mid-ocean ridge magmas requires a chemically inhomogenous mantle with isolated reservoirs of chemical heterogeneity. This has led to the suggestion of layered mantle circulation. Although

it should be noted that the geochemical evidence says nothing about the origin of the observed isotopic heterogeneity, and the evidence from geodynamics and seismic tomography suggests that whole mantle flow occurs (Davies & Richards (1992); Grand (1994); Helffrich & Wood (2001)).

One proposed solution is for a deep enriched mantle reservoir, which has extreme topography on its upper surface, and so the layer has remained seismically undetectable (Kellog *et al.* (1999)). The D'' discontinuity presents a suitable candidate for such an interface because it is intermittently observed, and has been found to vary in depth by 100km over short distances. Several authors have also reported evidence of chemical heterogeneity within the D'' region (Hedlin *et al.* (1997); Revenaugh & Meyer (1997); Braña & Helffrich (2004)).

12.4 D'': A phase transition

One hypothesis for the origin of the D'' reflector that has received much attention recently is the possibility of a phase change at this depth. Sidorin *et al.* (1999), investigated this possibility and found that for a phase change to satisfy their seismic data it would have to have a very strong pressure/ temperature gradient (a Clapeyron slope of 6 MPaK^{-1}). At the time no such phase change was known to exist. However, recently Oganov & Ono (2004) showed that at pressures and temperatures consistent with the top of the D'' region, MgSiO_3 transforms from perovskite into a layered post-perovskite phase.

The phase transition, and the new phase, together with the D'' as a thermal boundary layer can explain many of the properties of the D''. The phase transformation is predicted to have a steep Clapeyron slope $8.0\text{-}9.6 \text{ MPaK}^{-1}$, which could explain the intermittent observations of the D'' discontinuity. The post-perovskite phase develops strain induced LPO. So horizontal flow, which is expected within a boundary layer could explain the observations of seismic anisotropy. LVZ's can be explained by considering the intersections of curved geotherms associated with a thermal boundary layer. As shown in figure 12.2, the stability field of the post-perovskite phase has a base, where we would expect to see a velocity decrease. However, this does not explain

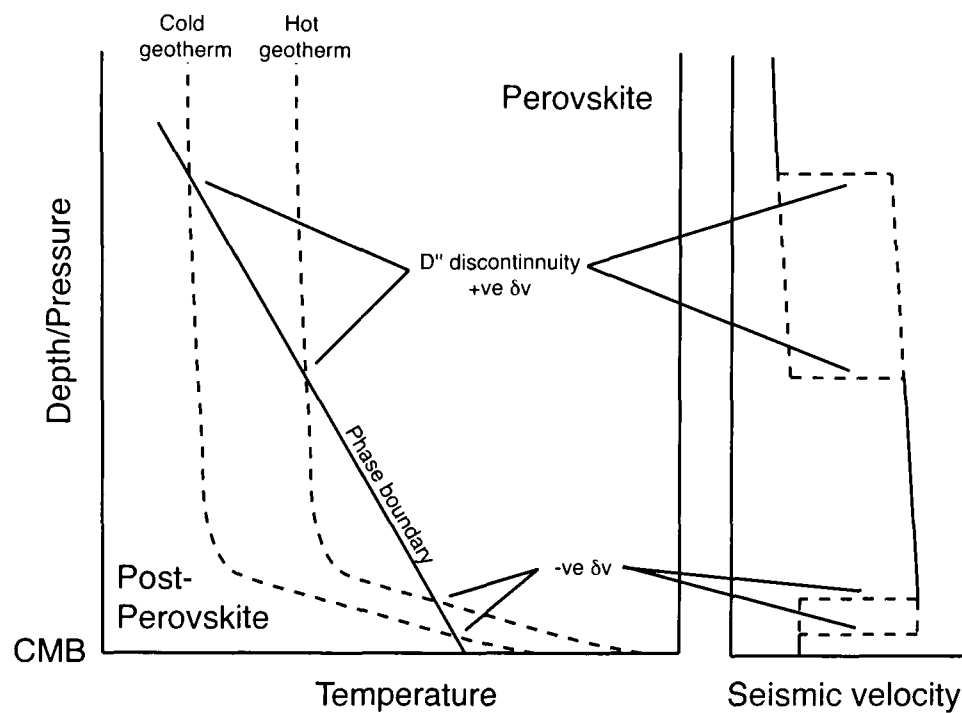


Figure 12.2: The effect of lower mantle temperature on post-perovskite phase change. The steep Clapeyron slope for the phase change means that the geotherm can cross the phase boundary twice resulting in a positive and a negative seismic discontinuity.

the magnitude of the velocity decreases observed in LVZ's being much greater than the velocity increase observed for a D'' discontinuity.

12.5 Integrated views of the D''

Clearly the processes that operate in the D'' (thermal, chemical, and phase variations) are not mutually exclusive. Recently, Lay & Garnero (2004) have summarised the current available evidence into two end-member conceptual models which describe a thermo-chemical boundary layer (TCBL) in the D'' (figure 12.3).

The first termed the 'hybrid TCBL' model describes D'' properties relating to whole mantle circulation. In this model subducting slabs descend to the D'' region, cooling surrounding material, and sweeping aside dense compositional heterogeneities. The authors suggest that this model requires an additional phase change or steep gradients in seismic properties to produce the D'' discontinuity.

The alternative model, termed the 'global TCBL', proposes that the D'' region is a chemically

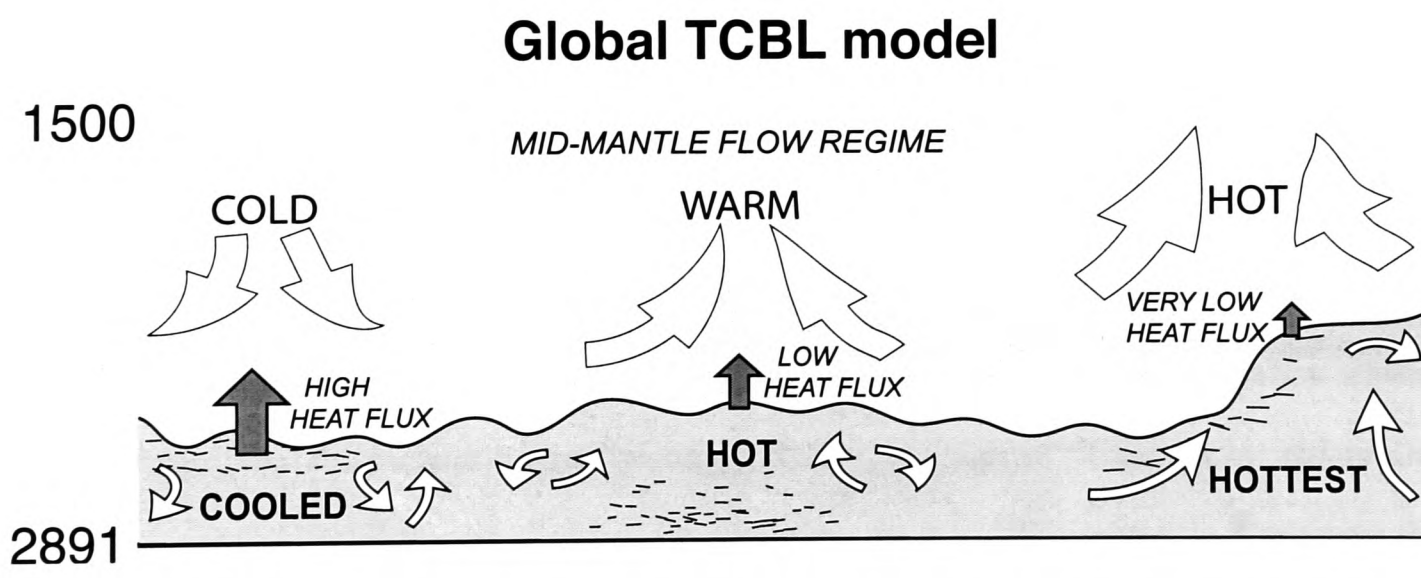
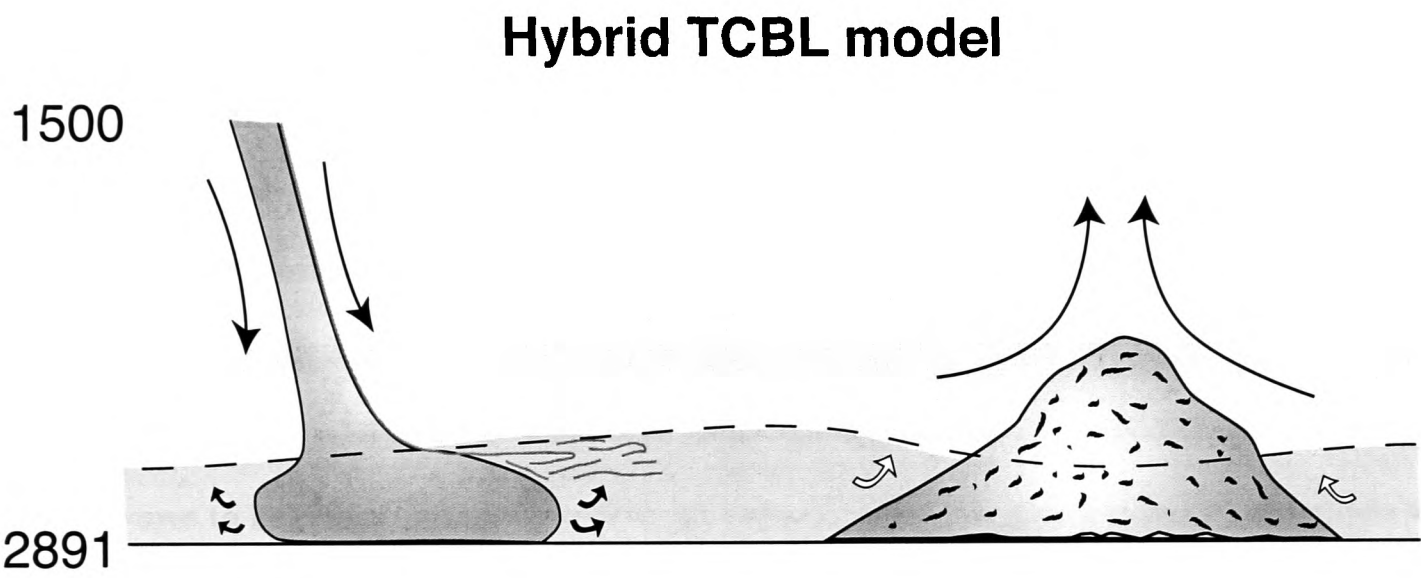


Figure 12.3: The two end-member models for the thermo-chemical boundary layer at the base of the mantle proposed by Lay & Garnero (2004). Figure is adapted from the same publication.

distinct from the rest of the mantle. This model is similar to the dense reservoir of enriched mantle proposed by Kellog *et al.* (1999). In this case the D'' discontinuity marks a chemical boundary. Scattering and anisotropy are attributed to thermal, and compositional variations within the separate layer.

12.6 Case for research

Despite the regions importance for our understanding of the processes such as, mantle convection and mixing, chemical heterogeneity in the mantle, and core-mantle coupling, the lowermost mantle remains poorly understood. The first step to understanding the lower mantle is better imaging techniques. Current seismological techniques either use amplitudes and travel times of high frequency waves to investigate structure over small regions, or use just global travel time datasets to constrain global structure. In the following sections a different approach is taken, which attempts to bridge this gap by applying high resolution migration processing techniques to a global dataset.

First the phase stripping technique of Kendall & Shearer (1994) is revisited, which is applied to a much larger set of high quality seismograms (1075 traces). The technique isolates scattered S-wave phases originating in the lower mantle. Then the positions of these scattering bodies are found using two migration methods.

The first migration technique is a simple backprojection of deconvolved records. This is useful for identifying positions of strong scattering. The second method is a weighted migration scheme developed using the Generalised Radon Transform (GRT). In chapters 15 and 16 ray theory and the Born approximation are used to obtain an expression for S-to-S scattering, and to adapt the solution of Miller *et al.* (1987), to isotropic elastic media. In particular, the results for three regions of the lower mantle beneath Alaska, North Central Asia, and New Zealand are discussed. The migration results reveal several features of the lowermost mantle, such as the presence of a D'' discontinuity with topography, and bodies with positive and negative scattering potentials near the CMB.

Migration procedures are potentially capable of resolving much more subtle structure than travel time data, as the amplitude of scattered phase as well as the arrival times are used. Such techniques are going to be required if we want to truly establish the features and processes in the lower mantle.

Chapter 13

Phase stripping

13.1 Method

13.1.1 Initial data selection and processing

Broad band data sampled at 1s from the IRIS/IDA/USGS network was used. The selection criteria is shown in table 13.1. The major restricting criteria are the use of data from either shallow ($<50\text{km}$) or deep events ($400\text{-}800\text{km}$), and the epicentral distance range ($60\text{-}75^\circ$). The first of these restrictions is necessary to minimise the interference from depth phases, such as sS , with lower mantle reflections. For the deep events depth phases will arrive outside the time window of interest, while for shallow events they are incorporated into the tail of the S and ScS pulses. The restricted epicentral distance range is necessary for the application of the phase stripping technique, which requires similar take-off angles for the S and ScS phases.

The data were filtered between 15 and 75 seconds using a Butterworth band pass, and a cosine taper was applied between 8s and 4s. The records were also deconvolved for long period instrument response and rotated to the transverse component.

Component	Horizontal (LHT)
Event date range	1990-2003
Event depths (km)	0-50 400-800
Event Size (m_w)	5.7-7.5
No. Events (initially)	1816
Start-finish time (s)	-60 +240
Epicentral distance ($^{\circ}$)	60-75

Table 13.1: Data selection parameters for the phase stripping dataset. The Start-finish times were taken relative to the expected arrival times of the reference phase in PREM.

13.1.2 The phase stripping method

Prior to phase stripping the traces are dominated by the S and ScS phases, which obscure the more subtle reflections from the lower mantle. In most previous studies reflections from the lower mantle have been distinguished from the S and ScS phases by the use of seismic arrays and higher frequency data.

However, Kendall & Shearer (1994) showed that by making a reference pulse from the first half of the S pulse and second half of the ScS pulse it is possible to make a reference trace which when subtracted from the original removes the large S and ScS phases as well as depth phases, because all depth and upper mantle scattered phases will follow closely behind both the direct S-wave and the ScS . This leaves a residual trace which consists of phases which have their origin in the lower mantle.

The technique is summarised as follows:

1. From the filtered records 1075 high quality traces were selected. For each trace a start and mid-point time for the S phase was picked, along with an end and mid-point for the ScS phase. Trace amplitudes were scaled to the absolute amplitude of the mid-point in the S-wave.
2. The picked portions of the traces are scaled and placed together to form a reference pulse for that trace. Hence the requirement for similar take-off angles at the source for the S and ScS phases, so that the phases have the same polarity.
3. A reference trace is made by placing the reference pulse at the position of the S and ScS

arrivals. The point of best alignment is determined, by cross correlating the reference pulse with the trace.

4. The reference trace is subtracted from the data to leave a residual trace which contains reflections and scattered phases from the lower mantle.

In figure 13.1 an example of the steps in the phase stripping procedure is shown. The figure shows that a good, but not exact fit to the trace can be achieved using the reference pulse aligned with the *S* and *ScS* phases. Some residual arrivals are isolated in the phase stripped trace (figure 13.2D).

In figure 13.2 the same procedure is applied to a synthetic seismogram generated using normal mode summation. The synthetic seismogram was generated using the same source-receiver geometry as the data and the PREM earth model (Dziewonski & Anderson (1981)). In this example the aligned reference pulse produces a better fit to the original trace, resulting in reduced amplitudes in the phase stripped trace.

As the synthetic was generated using normal model summation it includes all possible upper mantle and depth phases, but it does not include any reflected/scattered phases from the *D''* as there is no first order discontinuity at this depth in the PREM model. So, the lack of arrivals in the phase stripped trace, suggests the observed arrivals in the data have an origin in the lower mantle. As noise in the reference pulse will create spurious arrivals in the phase stripped trace the phase stripping technique only works with very high quality records. Accordingly most of the traces in the original dataset were rejected, leaving a working dataset of 1075 high quality phase stripped records. Figure 13.3 shows the source-receiver paths, and the source-receiver mid-points for the dataset. In subsequent chapters the origins for the arrivals in the phase stripped traces are located using migration techniques. From figure 13.3 it can be seen that coverage (and hence resolution) is best beneath North Central Asia, but other regions such as Alaska also contain a reasonable number of traces.

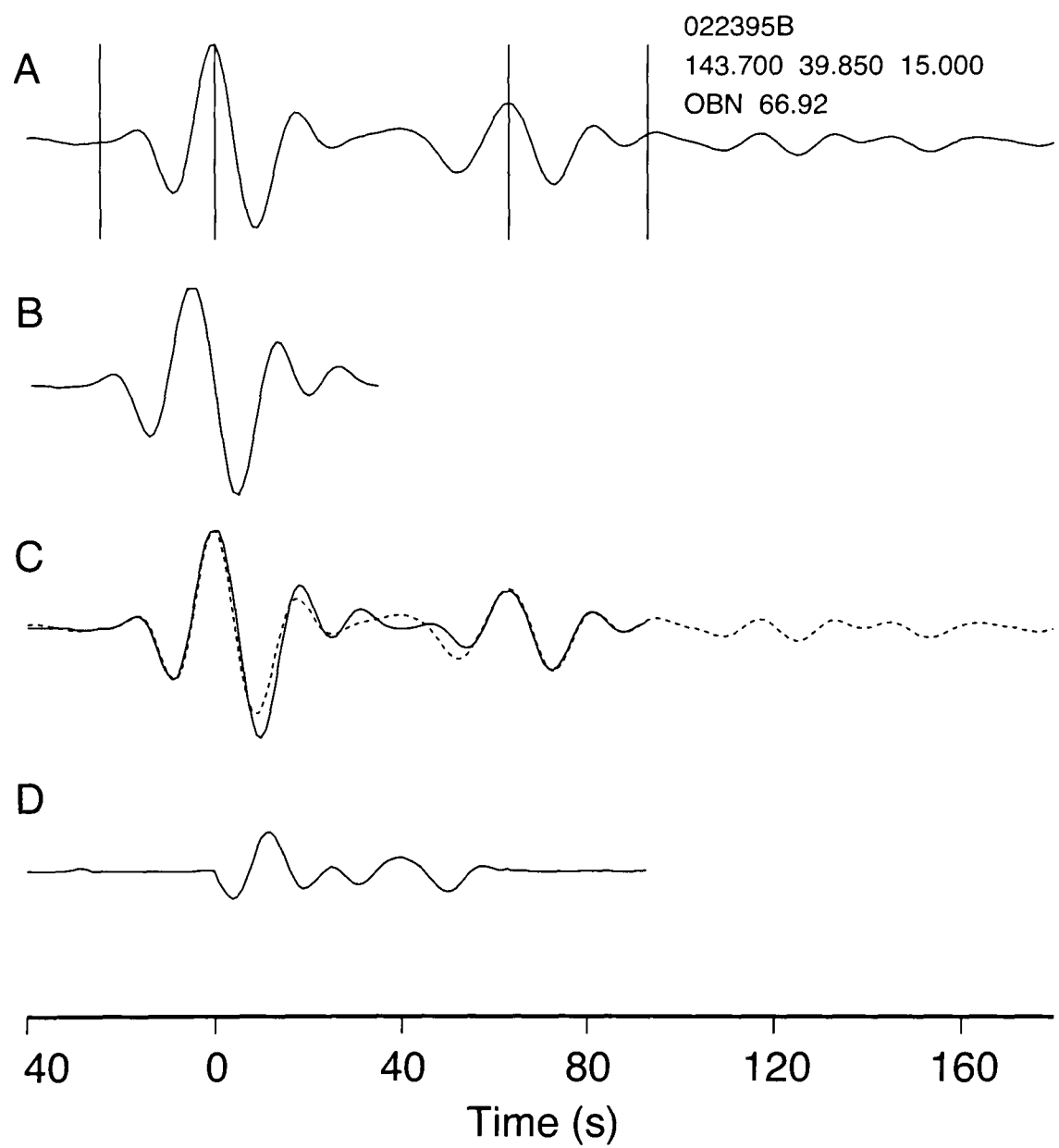


Figure 13.1: Example of the phase stripping procedure applied to data. **A** Original trace, solid lines mark the start at mid-point of the *S* phase and the mid-point and end of the *ScS* phase. Trace info (CMT event name, event co-ordinates, station name, and epicentral distance) are also shown. **B** The reference pulse made by joining the picked portions of the *S* and *ScS* phases. **C** Comparison of reference trace (solid line), made by aligning the reference pulse with the *S* and *ScS* phases, with the original data (dashed line). **D** The residual trace made by subtracting the reference trace from the original data (i.e dashed - solid).

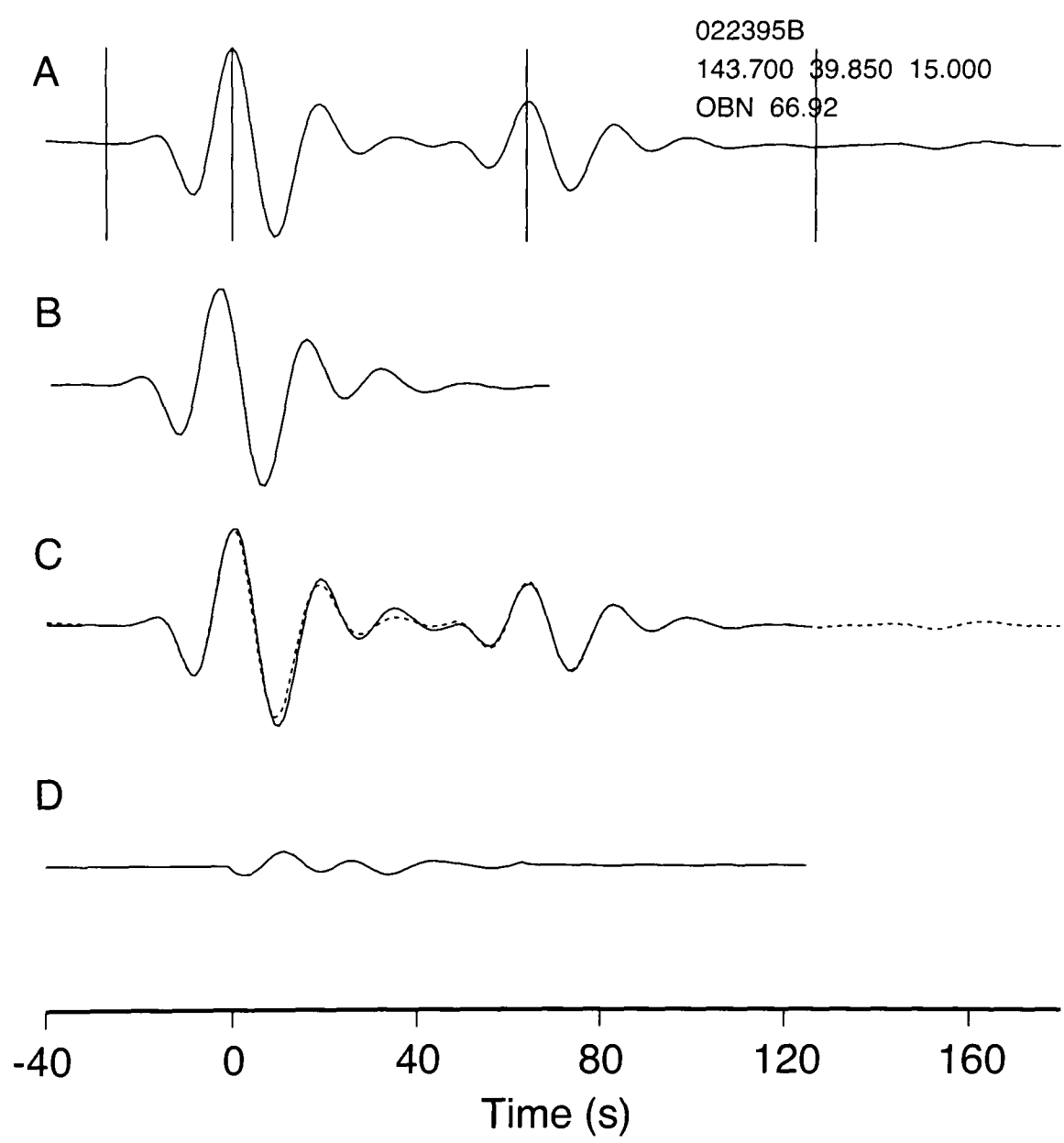


Figure 13.2: Example of the phase stripping procedure applied to a synthetic seismogram generated by normal mode summation and the earth model PREM . Plot lay out out is the same as figure 13.1.

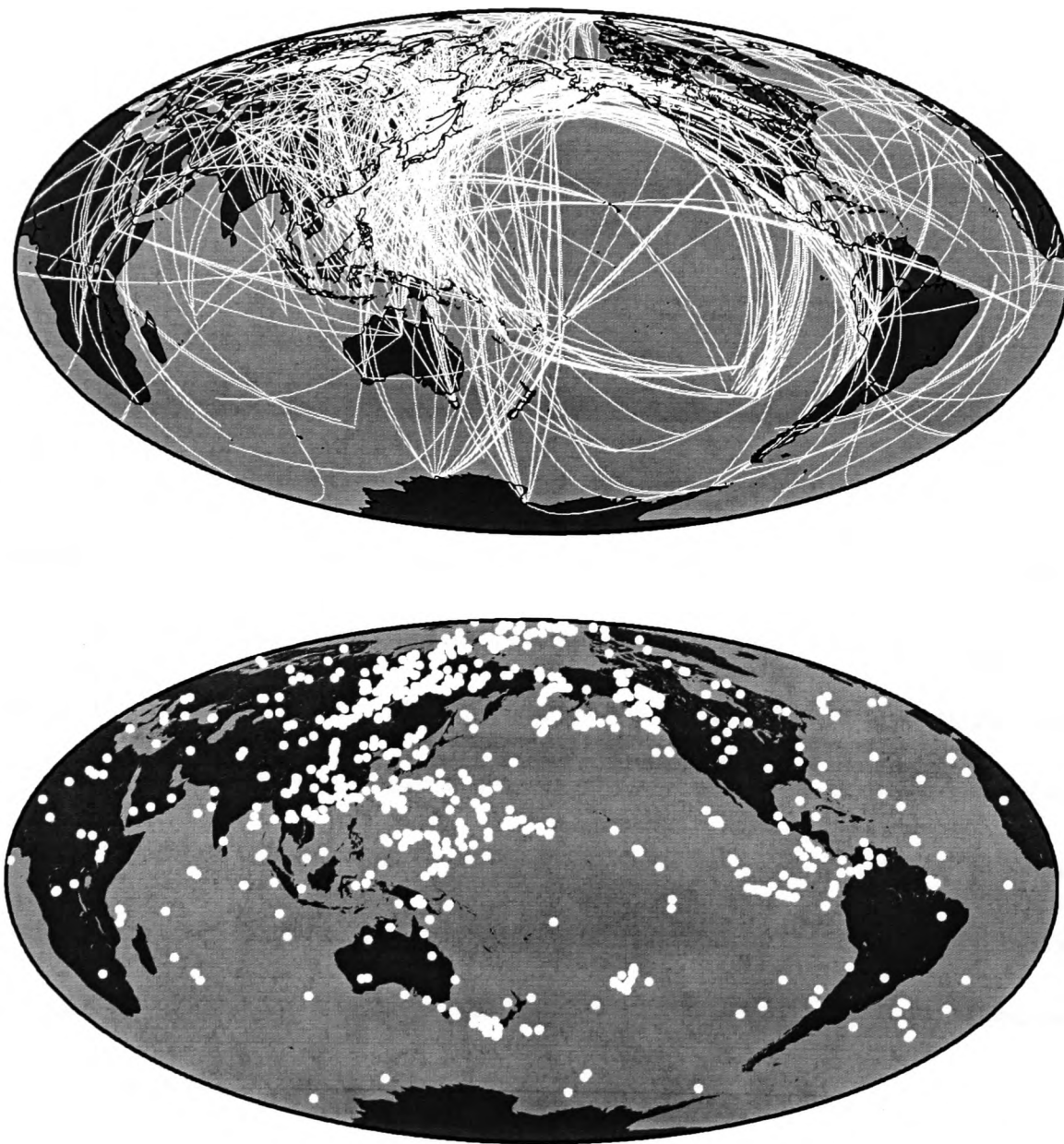


Figure 13.3: (top) Source receiver paths for the dataset of phase striped traces. (Bottom) global distribution of source-receiver mid-points.

13.1.3 Further deconvolution

For the purposes of initial modelling (section 13.2, and chapter 14) the phase stripped traces are simplified using an iterative deconvolution procedure. The deconvolution is similar to that applied to the precursor stacks in part I, and is described in detail in appendix A, so here only a brief summary is given.

The aim is to use a series of spikes (the deconvolved trace) which when convolved with the reference pulse explain 95% of the energy in the original trace. So, the process begins by checking the reference trace (the reference pulse aligned with S and ScS phases) against this criteria. If the trace passes, then the deconvolved trace will only consist of 2 spikes positioned at the arrival times of the S and ScS phases and the amplitudes of the spikes are determined by least squares inversion.

If the reference trace fails to explain 95% of the energy in the original, then in addition to the 2 spikes for the S and ScS phases spikes are progressively fitted to the phase stripped trace. This is done by cross-correlating the phase stripped trace with the reference pulse. A spike is placed at the maximum correlation, and its amplitude is solved for using least squares inversion. Then a reference trace is made by convolving the reference pulse with this spike sequence, which is subtracted from the original to make a residual trace. The process is then repeated with the residual trace in place of the original phase stripped record. At each iteration all the spike amplitudes are solved for simultaneously and continue until the 95% criterion is satisfied. Figure 13.4 shows an example of the deconvolution procedure.

13.2 Global record sections

The application of the iterative deconvolution from the last section to the phase stripped dataset finds ~ 1340 extra arrivals in these traces. In figure 13.5 these arrivals are shown in a record section format, with traces aligned on the S arrival. For reference the expected travel time curves for reflections from 2440, 2540, 2640, 2740km depth and the CMB are also given. In both of the plots

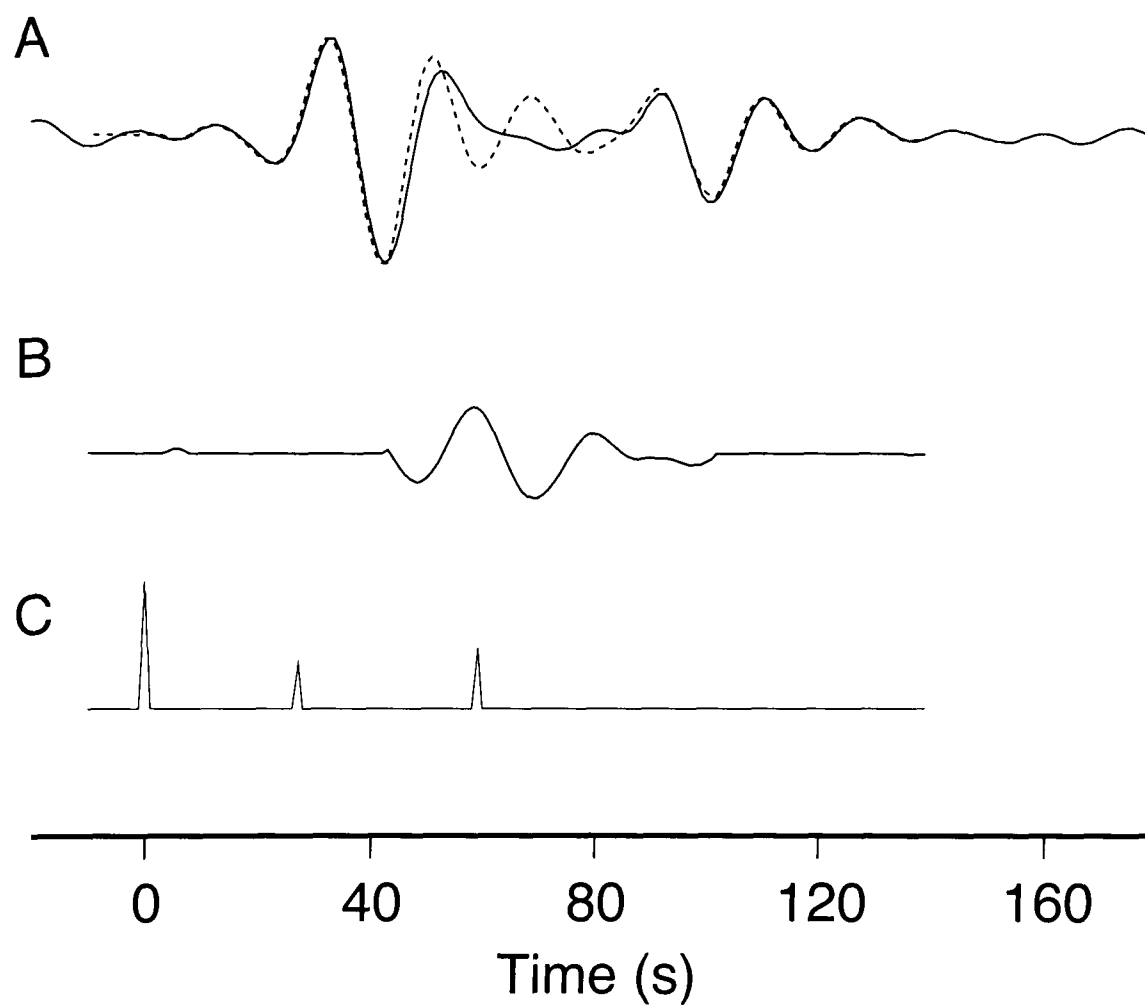


Figure 13.4: Example of the deconvolution procedure. **A** original (solid) and reference (dashed) traces. **B** Phase stripped trace (original - reference trace), and **C** deconvolved trace. The two largest spikes represent the S and ScS arrivals, the smaller spike in between is required to fit the phase stripped trace.

of figure 13.5 a clear move out pattern is seen for the ScS arrivals relative to S , but no other move out pattern is seen.

The lack of a clear global move out curve for the intermediate arrivals is in agreement with the analysis of Kendall & Shearer (1994) who found stacking techniques based on reflections from a horizontal boundary to be unsuccessful in enhancing these phases. Yet the lack of observation of these phases in the normal mode synthetics essentially requires a lower mantle origin for these phases. So, it seems that one (or more) flat global layers in the lower mantle cannot explain the distribution of arrival times observed in the deconvolved traces. It seems more likely that the lower mantle arrivals are from one or more interfaces with topography, or regions of heterogeneity. In the subsequent chapters seismic migrations are used to investigate this further.

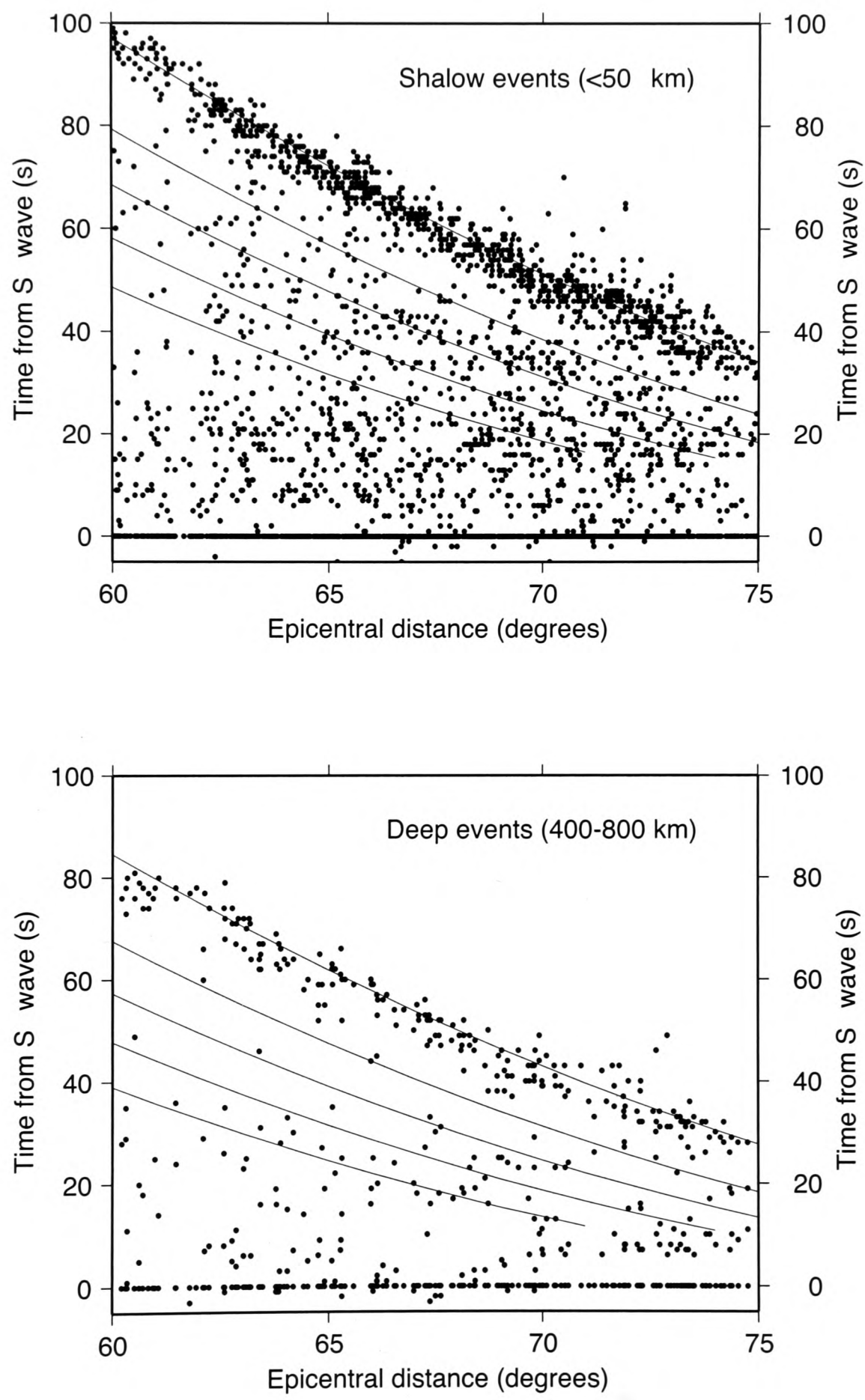


Figure 13.5: Record sections of deconvolved traces aligned on the S-wave arrival. Each circle denotes an arrival in the deconvolved traces. The solid lines mark the expected arrival times of reflections from 2440, 2540, 2640, 2740, and 2890 (the CMB) km depth. Top plot uses records for shallow events with travel times based on an event depth of 15km. Bottom plot uses records from deep events and travel times based on an event depth of 530km (the mean for deep events).

Chapter 14

Initial lower mantle backprojection

14.1 Method

Initially a simple backprojection of the deconvolved traces is used to find likely regions of scattering in the lower mantle. The backprojection method is useful for quickly establishing the likely positions of scatterers. More detailed modelling of seismic scattering is left to subsequent chapters.

Here and in the following chapters each arrival in the scattered trace is considered to be the sum of contributions from the surface of a ‘scattering sausage’. In other words the ‘scattering sausage’ marks the surface for which all the points have the same travel time. For S-to-S scattering this surface forms a hollow ellipsoid around the direct arrival for a given source-receiver pair. Figure 14.1 shows an example of an S-to-S scattering sausage in the lower mantle.

The technique here is similar to that used by Lay & Young (1996) and Bilek & Lay (1998), who both set out to find the most likely locations for the scattering bodies in the lower mantle, beneath Alaska and North Central Asia respectively. However, in the case of this study there was no initial intention to focus on a particular region using a specific source-receiver geometry.

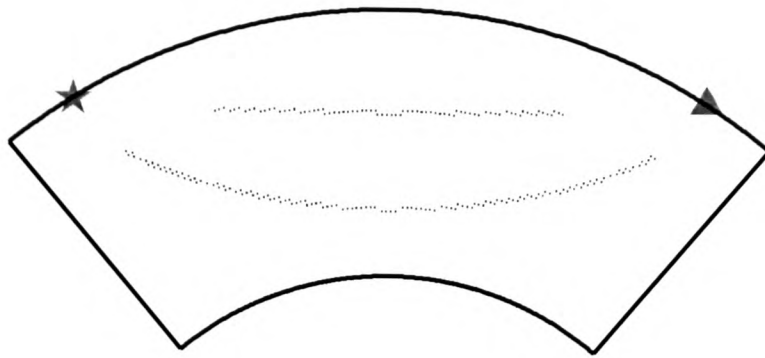


Figure 14.1: Cross-section, through the source-receiver plane, of a constant travel time surface for S-to-S scattering arriving 15 seconds after the direct arrival. The symbols mark the position of the source and receiver. Both are at the surface and separated by a 65° . The base of the plot is at the CMB. Note that the surface shown here stops at 780km depth, where as in reality it extends to the surface.

The aim is to find the most likely locations for the scattering bodies in the lower mantle which could produce arrivals in our deconvolved traces. Thus, a grid of equally spaced points within the Earth is considered. The points are spaced at intervals of 50km along lines of longitude, latitude and radius, between 2940 and 2040km depth. At each point the amplitude of the time samples from the deconvolved traces which could have come from that point are summed (the backprojection step). Accordingly, the largest amplitudes are expected, and hence the most likely scatterer positions, where multiple scattering ellipses intersect.

The shape of scattering ellipses is sensitive to the choice of background model. Here PREM was used for the calculation of the potential $S - SdS$ differential times. In order to calculate travel times for the lowermost layer of the model, which is below the CMB, a constant travel time gradient was assumed.

As the deconvolved traces are a series of spikes with widths of one time sample (1s), mantle heterogeneities will cause misalignment of the arrivals and reduce the stack quality. To prevent this each spike sequence was convolved with a Gaussian function prior to the backprojection. The standard deviation of the Gaussian was taken from the distribution of $S - ScS$ differential times (2.8s). The spikes corresponding to the S and ScS arrivals were removed prior to migration as the former does not correspond to a scattering surface and will produce erroneous results, while the latter being larger will dominate the stacking process and mask the smaller arrivals we wish

to identify.

14.2 Backprojection results

14.2.1 Initial results

The backprojection was performed for approximately 1.35 million image points in the lower mantle. Figure 14.2 shows map views of some of the results from the northern hemisphere, where the data set has best coverage and sampling. Additional results are given in appendix B. The ‘hit count’ at each image point is also shown in figure 14.2. This is the number of time samples in the dataset consistent with a particular point. Alternatively the ‘hit count’ gives the number of intersecting travel time surfaces at a point. Figure 14.2 suggests that coverage, and hence resolution, is best beneath China and North Central Asia. Although more than 100 time samples are summed at each point over most of North America as well.

Two anomalies stand out at the image depths in figure 14.2 beneath North Central Asia and Alaska. These features along with an anomaly beneath New Zealand will be the focus of discussion here and in the following chapters.

14.2.2 Alaska

The global results showed some of the strongest stack amplitudes beneath the Alaska/Canada border. Maps of the migration amplitude for 8 different depths are shown in figure 14.3. It can be seen that images are formed at all depths above 2690km, but the major positive pattern focusses down to a surface between 2590 and 2640km (250-300km above the CMB).

Lay & Young (1996) found evidence for a near horizontal reflector at roughly the same depth (2600km), using similar data and techniques. However, they suggest that the reflector extends much further to the west over all of Alaska (lateral dimensions of 1500km). Here this is not the case; figure 14.3 suggests that the reflector is limited to a region near the Alaska/Canada border.

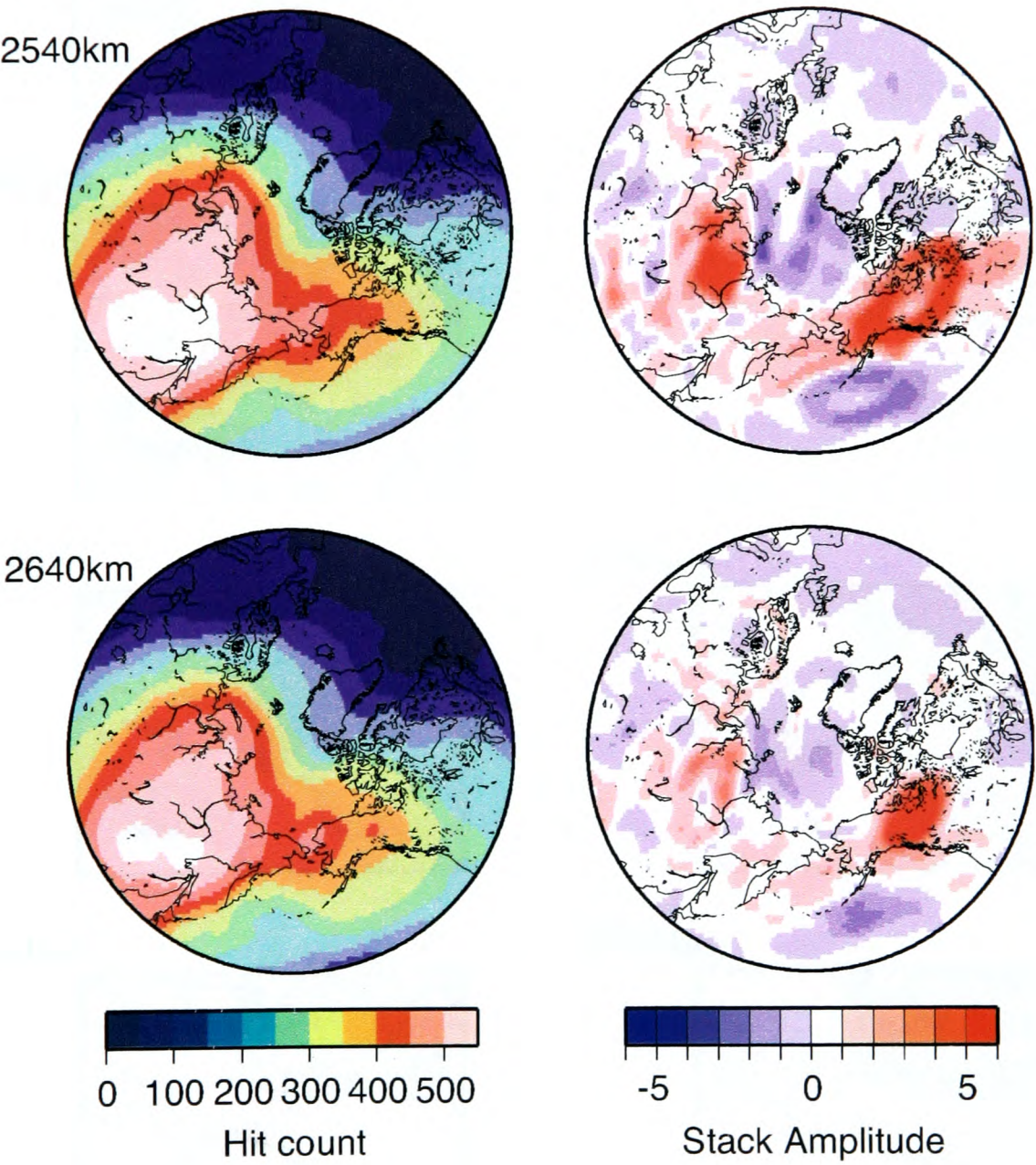


Figure 14.2: Backprojection results, 350 and 250km above the CMB, depths are shown on the left. (left) Map of the hit count at each image point, (right) Backprojection results. Maps are centred on the north pole.

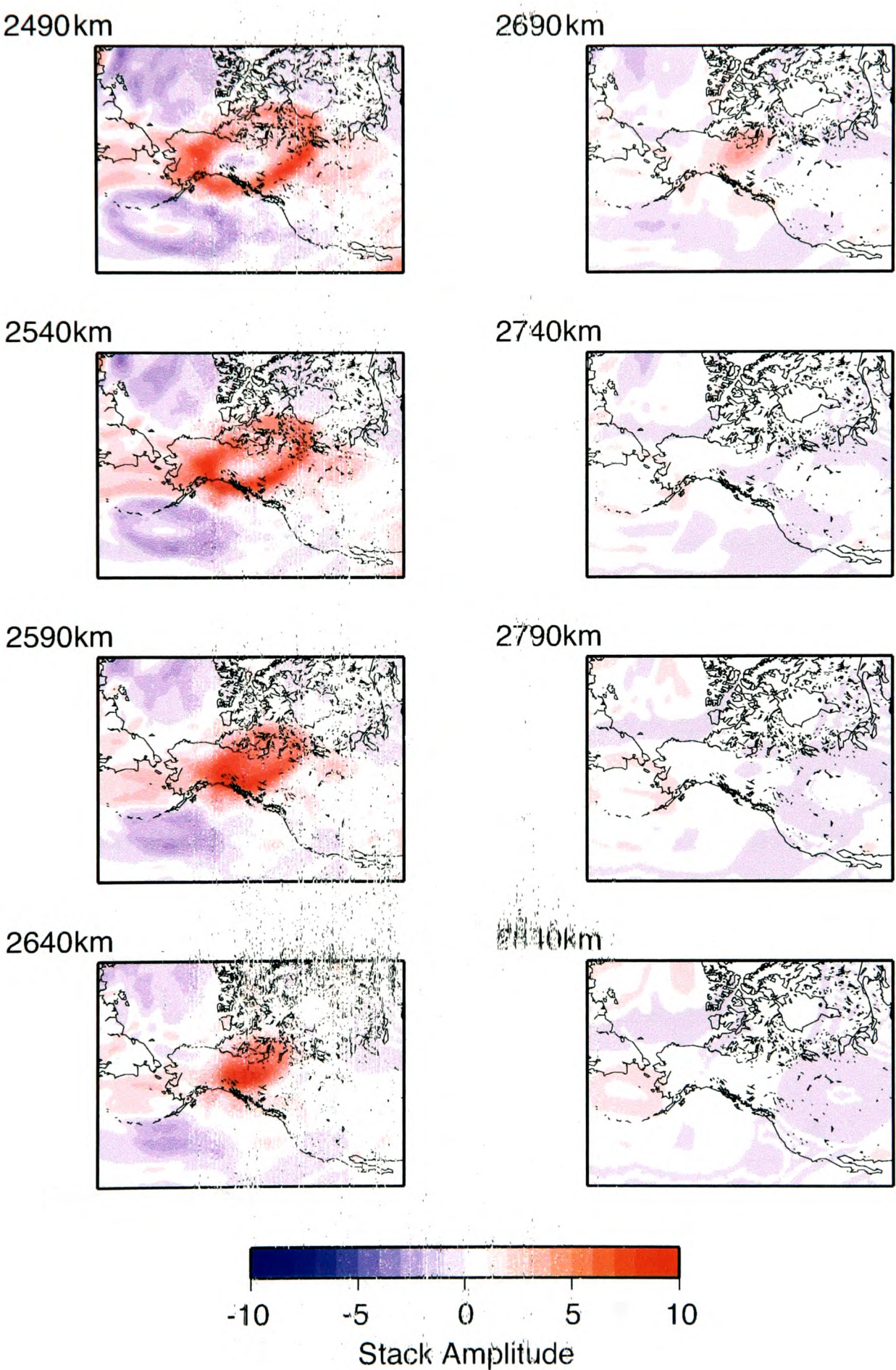


Figure 14.3: Results of backprojection using deconvolved records, beneath Alaska.

14.2.3 North Central Asia

In figure 14.4 maps for the mantle beneath Northern Central Asia are shown. The results are not as clear as those for Alaska. The largest amplitudes are limited to depths above 2590km (300km above the CMB), where the most prominent feature is a positive anomaly which appears to focus down to a body between 2540 and 2590km depth.

This region has been the focus of several previous studies Houard & Nataf (1993); Bilek & Lay (1998). Thomas *et al.* (2004b) have identified a reflector with topography between 2650-2550km depth within this region. The results shown here suggest the reflector is at the shallower end of this range. Thomas *et al.* (2004b) also presented evidence for a low velocity zone 55-85km above the CMB in this region. Using this technique negative reflections are not seen, although the layers thickness may be less than what is resolvable with the data and methods used here.

14.2.4 New Zealand

The region surrounding New Zealand and the South Pacific shown in figure 14.5, is more sparsely sampled than the previous two. Typical 'hit counts' are 50-100 in this region. The area contains several crossing ray paths from South Pacific events recorded in Antarctica (see figure 13.3). The major feature of figure 14.5 is a positive structure which comes into focus at 2840km depth. The migration suggests that the positive body is restricted to the region Southwest of New Zealand. Unlike the other regions there appears to be no indication of any reflectors consistent with the top of D'' (2540-2740km) in this region.

14.3 Brief analysis

In this chapter a backprojection of deconvolved phase stripped traces was used to find the most likely origins for arrivals in the lower mantle. Although this technique is useful for finding the locations of intermittent reflectors the amplitudes are not a direct measure of heterogeneity, nor are they useful as a comparison between regions, for two reasons. Firstly, no compensation was made

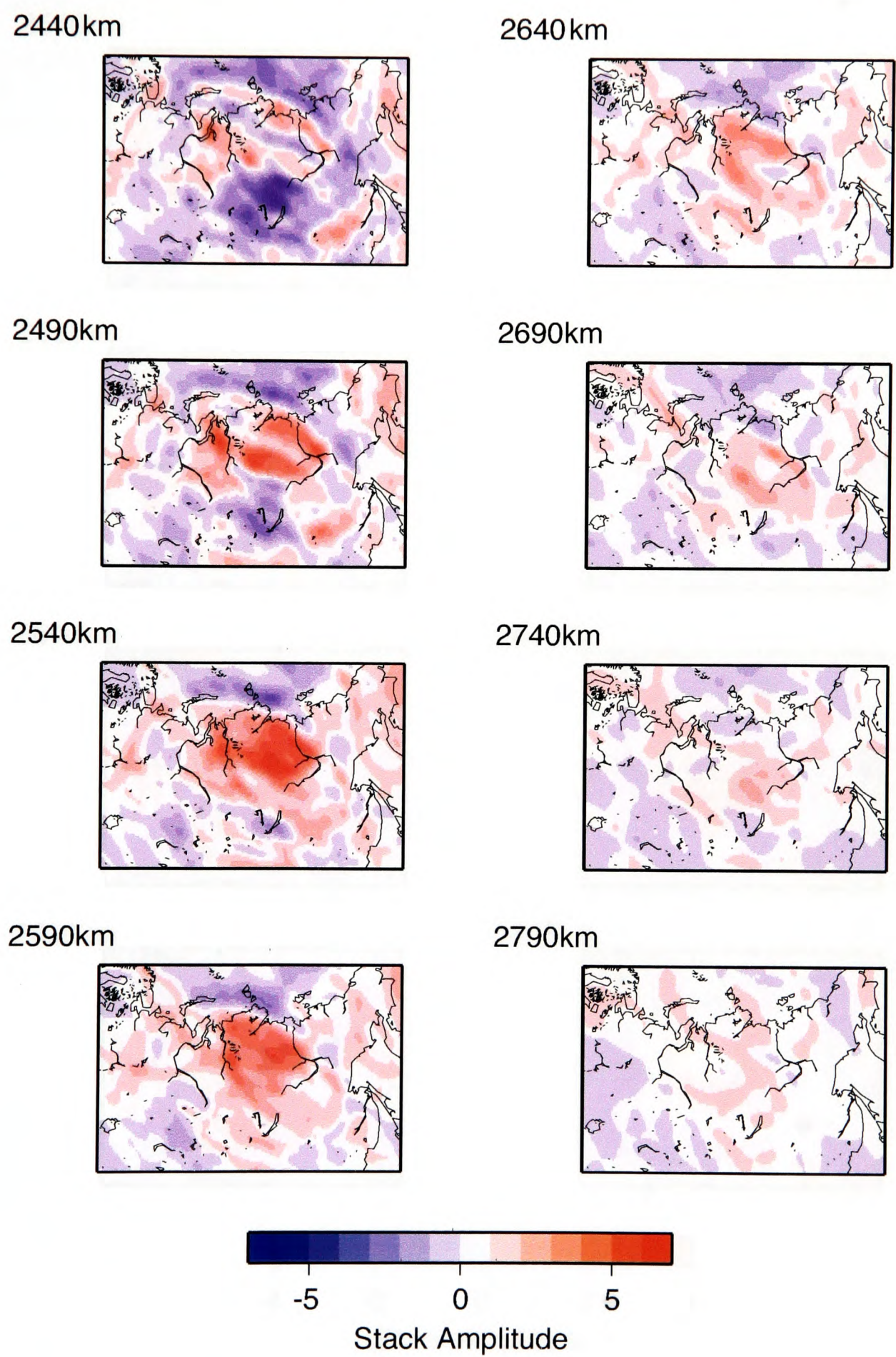


Figure 14.4: Results of backprojection using deconvolved traces beneath North Central Asia.

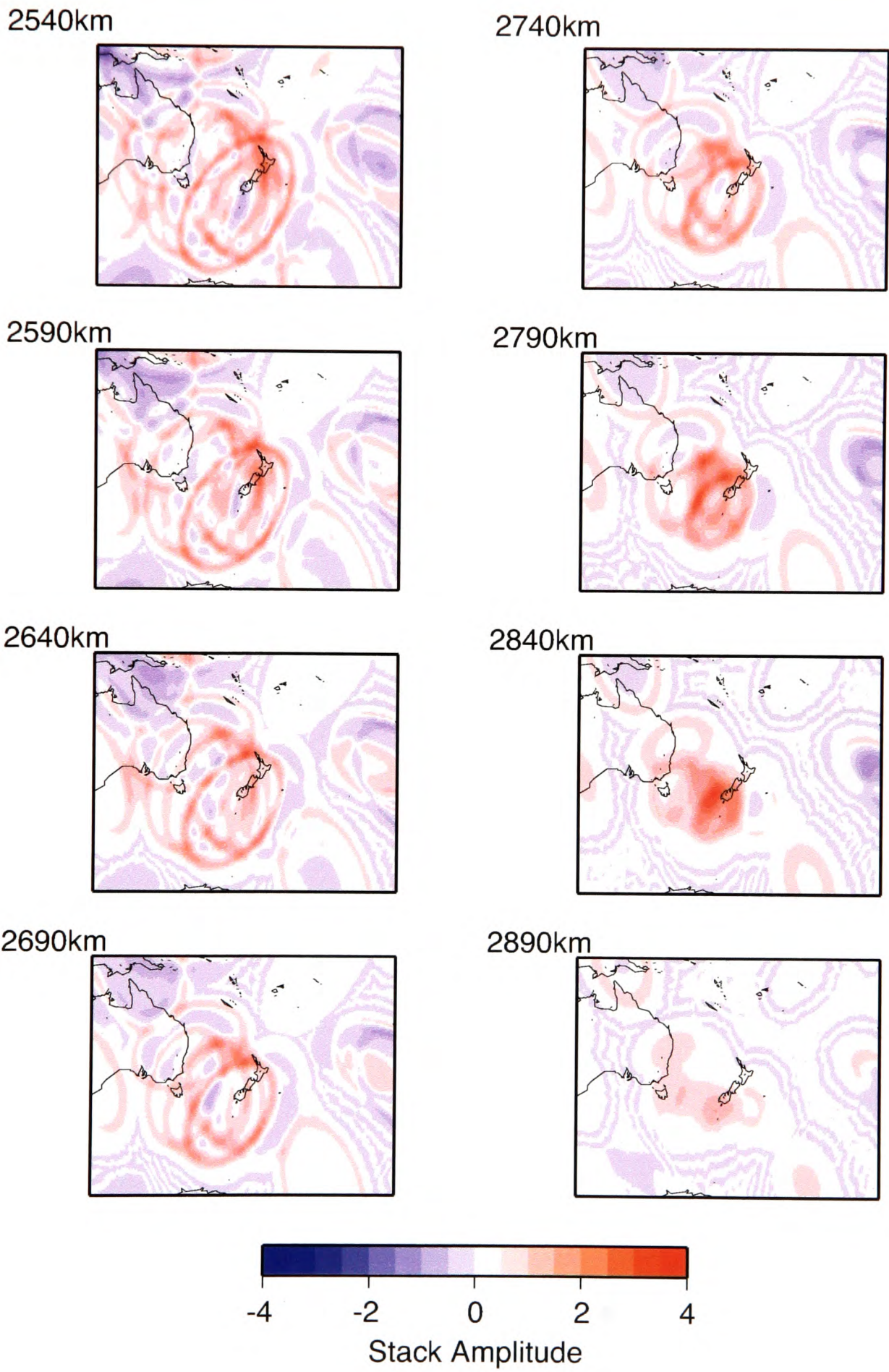


Figure 14.5: Results of backprojection using deconvolved traces beneath the New Zealand.

for the number of traces stacked at each point. Secondly, there was no compensation for factors which affect reflection amplitude such as reflection angle, or geometrical spreading. Furthermore the use of deconvolved traces assumes that the arrivals are reflected phases from a planar boundary. In subsequent sections these issues are addressed and a more sophisticated migration technique presented.

A dominant feature in two of the regions appears to be a reflector 250-350km above the CMB (2640-2540km depth). Previous studies have seen a similar reflector consistent with the top of the D'' layer beneath Alaska and North Central Asia. However, in the third area -New Zealand- no reflector is seen at this depth. Instead the backprojection showed a scattering body at 2840km depth (~ 50 km above the CMB). Structures so close to the CMB are at the limit of what is resolvable with the current dataset.

Although a D'' reflector is not seen at the depth we might expect beneath New Zealand, the observation of the reflector at 2840km suggests it should be resolvable in this region¹. So, it seems that either the D'' reflector observed beneath Alaska and North Central Asia is not a global feature and the reflector beneath New Zealand has a separate cause, or the D'' discontinuity varies between at least 2840km and 2540km in depth.

¹This is tested in chapter 19

Chapter 15

S-wave scattering theory

15.1 Introduction

In this chapter an expression for the scattered S-wave from a point heterogeneity in a homogenous medium is derived. This is a prerequisite for the development of the subsequent migration. It is also useful for the generation of synthetic seismograms which can be used to examine the influence of experiment geometry on migration results.

Similar derivations of expressions for the scattered wavefield can be found in a variety of other studies. Several authors (for example Bleistein (1987); Bostock (2002) have derived expressions for a scattered wavefield from irregular, but continuous interfaces (otherwise termed Kirchhoff scattering). Others (Beylkin & Burridge (1990); Frederiksen & Revenaugh (2000)) derive similar expressions to those given here for S-to-S scattering in isotropic elastic media at a point scatterer.

To begin with the discussion here follows closely to that of Aki & Richards (1980). We begin by describing separate wave equations for the incident and scattered wavefields. We then adopt the Born approximation (single scattering) to get a single perturbed wave equation for the scattered wavefield, where the perturbing force is a function of the inhomogeneity and the incident wavefield. This solution is extended using ray theoretical Green's functions to derive radiation patterns for a scattered wavefield. The scattered field contains two different radiation patterns which result

from the Greens function, and its derivatives. Finally, these two patterns are combined with a scaling factor κ which relates density to velocity anomalies, to obtain a simplified expression for the scattered wavefield in terms of geometrical spreading, source receiver geometry, and velocity heterogeneity. Throughout, particular emphasis is placed on the radiation pattern of the scattered wavefield, as these can be applied directly to more complicated media.

15.2 Wave equation for scattered wavefield

The general seismic wave equation (15.1, in indicial notation) describes the component of motion, u_i in a material with shear modulus μ , fluid incompressibility, λ , and density, ρ .

$$[\mu(u_{i,j} + u_{j,i}) + \lambda\delta_{ij}u_{k,k}]_{,j} = \rho\ddot{u}_i \quad (15.1)$$

We separate the medium into perturbed and unperturbed portions using, $\delta\mu$, $\delta\lambda$, $\delta\rho$, and μ_o , λ_o , ρ_o such that.

$$\begin{aligned} \lambda &= \lambda_o + \delta\lambda \\ \mu &= \mu_o + \delta\mu \\ \rho &= \rho_o + \delta\rho \end{aligned} \quad (15.2)$$

Substitution into equation 15.1 produces:

$$\begin{aligned} [\mu_o(u_{i,j} + u_{j,i}) + \lambda_o\delta_{ij}u_{k,k}]_{,j} &= \rho_o\ddot{u}_i \\ + [\delta\mu(u_{i,j} + u_{j,i}) + \delta\lambda\delta_{ij}u_{k,k}]_{,j} &+ \delta\rho\ddot{u}_i \end{aligned} \quad (15.3)$$

The first two terms on each side describe the wave in the unperturbed medium, while the second two terms describe the interaction with the perturbed medium. The interaction with heterogeneity separates the wavefield, $\mathbf{u}$, into components describing the perturbed, $\mathbf{u}^s$, and unperturbed $\mathbf{u}^o$ portions such that:

$$\mathbf{u} = \mathbf{u}^o + \mathbf{u}^s \quad (15.4)$$

Substituting equation 15.4 into 15.5 allows separation of the wavefield into scattered and incident/unperturbed parts. Accordingly we get:

$$\begin{aligned}
 & \left[\mu_o (u_{i,j}^o + u_{j,i}^o) + \lambda_o \delta_{ij} u_{k,k}^o \right]_{,j} - \rho_o \ddot{u}_i^o \\
 + & \left[\mu_o (u_{i,j}^s + u_{j,i}^s) + \lambda_o \delta_{ij} u_{k,k}^s \right]_{,j} - \rho_o \ddot{u}_i^s \\
 + & \left[\delta\mu (u_{i,j}^o + u_{j,i}^o) + \delta\lambda \delta_{ij} u_{k,k}^o \right]_{,j} - \delta\rho \ddot{u}_i^o \\
 + & \left[\delta\mu (u_{i,j}^s + u_{j,i}^s) + \delta\lambda \delta_{ij} u_{k,k}^s \right]_{,j} - \delta\rho \ddot{u}_i^s = 0
 \end{aligned} \tag{15.5}$$

A useful version of the equation for the scattered wavefield is obtained by applying the Born approximation to equation 15.5. First we note that, by definition, $\mathbf{u}^o$ satisfies the wave equation in the unperturbed medium. Accordingly, the terms in the first line sum to zero. Secondly, the interaction between the scattered wavefield and the perturbed medium is neglected (the 4th line is set to zero). This approximation is only valid if the strength of the heterogeneity is weak (i.e. $\delta\mu \ll \mu, \delta\lambda \ll \lambda, \delta\rho \ll \rho$).

Equation 15.5 now becomes the perturbed wave equation for the scattered wavefield in the homogenous background medium.

$$\left[\mu_o (u_{i,j}^s + u_{j,i}^s) + \lambda_o \delta_{ij} u_{k,k}^s \right]_{,j} - \rho_o \ddot{u}_i^s = f_i \tag{15.6}$$

Where the perturbing force, f_i is determined by the interaction of the incident wavefield with the heterogeneity.

$$f_i = \delta\rho \ddot{u}_i^o - \left[\delta\mu (u_{i,j}^o + u_{j,i}^o) + \delta\lambda \delta_{ij} u_{k,k}^o \right]_{,j} \tag{15.7}$$

15.3 Solution for the scattered wavefield

For the purposes of this thesis we are interested in S-to-S scattering from a point heterogeneity. Here the general equation for S-waves scattered by a point heterogeneity in a homogenous medium is derived. Later on this solution can be adapted to any source-receiver combination.

We assume an incident S-wave travelling in the x_1 direction, with particle motion in the x_2 direction, and a point scatterer located at the origin. The i 'th component of the scattered wavefield, $u_i(\mathbf{r}, t)$ at receiver, $(\mathbf{r})$, and time, t , can be obtained from a solution of equation 15.6 with an appropriate perturbation force term (equation 15.7). The general solution for a number of scatterers is given by the summation of the perturbation forces corresponding to each scatterer, $f_j(\mathbf{x}, \tau)$, with the appropriate Green's function, $G_{ij}(\mathbf{r}, \mathbf{x}, t)$.

$$u_i(\mathbf{r}, t) = \int G_{ij}(\mathbf{r}, \mathbf{x}, t) f_j(\mathbf{x}, \tau) d^3\mathbf{x} \quad (15.8)$$

Where, $\mathbf{x}$ refers to the position of scatterers, and τ refers to the arrival time of the incident wavefield at the given scatterer.

To simplify the problem we localise scattering to a small region around the origin and measure time from the arrival of the incident wavefield at the origin. Accordingly, differences in $\mathbf{x}$ and τ are likely to be small, and so these can be dropped from equation 15.8. As the scattering is assumed to come from a small heterogeneous volume, we temporarily change the integral to one over volume rather than point scatterers. Equation 15.8 now becomes:

$$\begin{aligned} u_i(\mathbf{r}, t) &= \int_v G_{ij}(\mathbf{r}, t) f_j dV \\ &= \int_v G_{ij} f_j dV \end{aligned} \quad (15.9)$$

The perturbing force is given by substituting the equation of motion for the incident S-wave into expression 15.7. The corresponding perturbation force acting at the origin in the frequency domain is:

$$f_j(t) = \begin{pmatrix} -i\frac{\omega}{\beta}(\delta\mu)_{,2} \\ 2\omega^2\rho\frac{\delta\beta}{\beta} - i\frac{\omega}{\beta}(\delta\mu)_{,1} \\ 0 \end{pmatrix} \quad (15.10)$$

Note that the subscript, $_0$, to describe the background medium has been dropped. Expression 15.10

is used with 15.9. To give expression 15.11 for the scattered wavefield.

$$u_i = \int_v \left(G_{i2} \left(2\omega^2 \rho \frac{\delta\beta}{\beta} \right) - i \frac{\omega}{\beta} [G_{i1}(\delta\mu)_{,2} + G_{i2}(\delta\mu)_{,1}] \right) dV \quad (15.11)$$

Equation 15.11 can be divided into three parts, of which the first represents the integral over velocity heterogeneity. The complex parts represent the integrals over gradients in shear modulus, $\delta\mu$. Integration by parts shows these to have non vanishing terms:

$$\begin{aligned} \int_v G_{i1}(\delta\mu)_{,2} dV &= - \int_v (G_{i1})_{,2}(\delta\mu) dV \\ \int_v G_{i2}(\delta\mu)_{,1} dV &= - \int_v (G_{i2})_{,1}(\delta\mu) dV \end{aligned}$$

Accordingly equation 15.11 becomes:

$$u_i = \int_v \left[G_{i2} \left(2\omega^2 \rho \frac{\delta\beta}{\beta} \right) + i \frac{\omega}{\beta} (\delta\mu) [(G_{i1})_{,2} + (G_{i2})_{,1}] \right] dV \quad (15.12)$$

15.4 Radiation patterns from a point scatterer

Equation 15.12 describes the i 'th component of the scattered S-wave from a point heterogeneity at the origin, and an incident S-wave travelling along the x_1 direction with particle motion in the x_2 direction. It contains two radiation patterns which are included in the G_{i2} and $[(G_{i2})_{,1} + (G_{i1})_{,2}]$ terms, which relate to heterogeneities in $\delta\beta$ and $\delta\mu$ respectively.

The G_{i2} term can be obtained from the far field S-wave Green's function for a homogenous medium [Aki & Richards (1980) p 74].

$$G_{ij} = \frac{1}{4\pi\rho r\beta^2} \left(\delta_{ij} - \frac{x_i x_j}{r^2} \right) \exp[-i\omega\tau] \quad (15.13)$$

Where, r , is the distance to the origin (scatterer location), and the phase delay τ is proportional to r/β . The radiation pattern from the point scatterer is controlled by the terms inside the

parentheses in equation 15.13. Accordingly the radiation of S-waves from a $\frac{\delta\beta}{\beta}$ heterogeneity is given by:

$$\mathbf{u} \propto \begin{pmatrix} \frac{-x_1 x_2}{r^2} \\ 1 - \frac{x_2^2}{r^2} \\ \frac{-x_3 x_2}{r^2} \end{pmatrix} = \begin{pmatrix} -\frac{1}{2} \sin^2(\theta) \sin(2\phi) \\ 1 - \sin^2(\theta) \sin^2(\phi) \\ -\frac{1}{2} \sin(2\theta) \sin(\phi) \end{pmatrix} \quad (15.14)$$

Where the spherical polar angles, θ and ϕ , are the angles of the scattered ray to the x_3 and x_1 axis respectively. Figure 15.1 shows the radiation patterns corresponding to equation 15.14. The radiation pattern for the imaginary part of equation 15.12 ($\delta\mu$ heterogeneity) is derived by first differentiating 15.13.

$$\frac{\partial G_{ij}}{\partial x_k} = \frac{-1}{4\pi\rho r\beta^2} \left[\frac{1}{r} \left(\frac{(\delta_{ik}x_j + \delta_{jk}x_i + \delta_{ij}x_k)}{r} - \frac{3x_i x_j x_k}{r^3} \right) + \frac{i\omega}{\beta} \frac{x_k}{r} \left(\delta_{ij} - \frac{x_i x_j}{r^2} \right) \right] \exp[-i\omega\tau] \quad (15.15)$$

Due to the $1/r$ factor, only the complex parts of equation 15.15 will effect scattered waves in the far-field. So the relevant parts of equation 15.15 reduces to:

$$\frac{\partial G_{ij}}{\partial x_k} \approx \frac{-i\omega}{\beta} \frac{1}{4\pi\rho r\beta^2} \frac{x_k}{r} \left(\delta_{ij} - \frac{x_i x_j}{r^2} \right) \exp[-i\omega\tau] \quad (15.16)$$

We can treat equation 15.16 in a similar way to equation 15.13, and derive a radiation pattern due to the $[(G_{i1})_{,2} + (G_{i2})_{,1}]$ term.

$$\mathbf{u} \propto \begin{pmatrix} \frac{x_2}{r} \left(1 - 2\frac{x_1^2}{r^2} \right) \\ \frac{x_1}{r} \left(1 - 2\frac{x_2^2}{r^2} \right) \\ -2\frac{x_1 x_2 x_3}{r^3} \end{pmatrix} = \sin(\theta) \begin{pmatrix} \sin(\phi) (1 - 2\sin^2(\theta) \cos^2(\phi)) \\ \cos(\phi) (1 - 2\sin^2(\theta) \sin^2(\phi)) \\ -\frac{1}{2} \sin(2\theta) \sin(2\phi) \end{pmatrix} \quad (15.17)$$

Figure 15.2 shows the radiation patterns corresponding to equation 15.17.

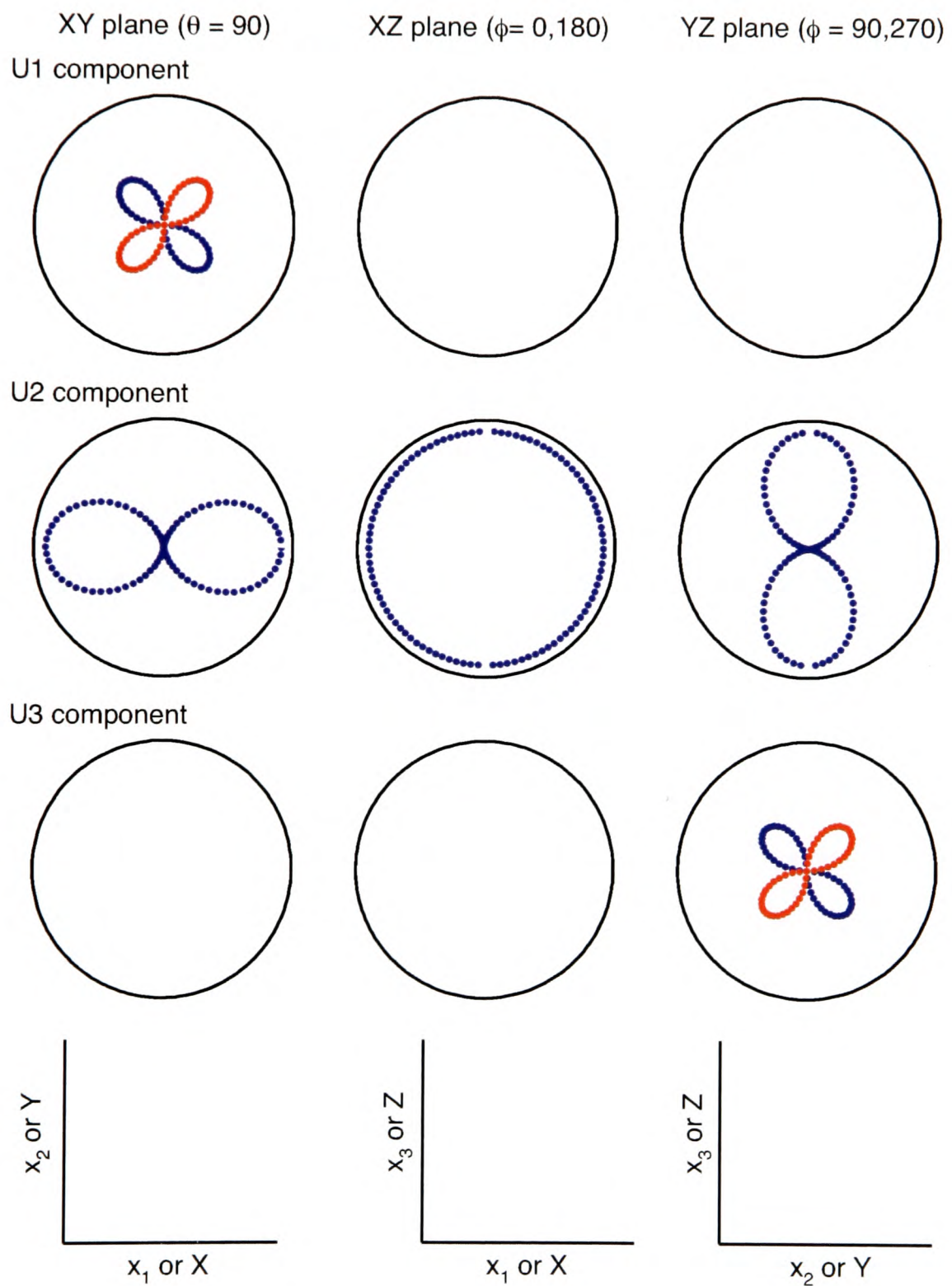


Figure 15.1: Radiation pattern from G_{i2} term of equation 15.12. U1, U2, and U3 refer to components in the x_1 , x_2 , and x_3 directions (ϕ and θ refer the spherical polar coordinates). The three columns refer to cross sections in the XY, XZ, and YZ planes, as shown by the diagrams at the bottom. Blue refers to a positive value, red is negative.

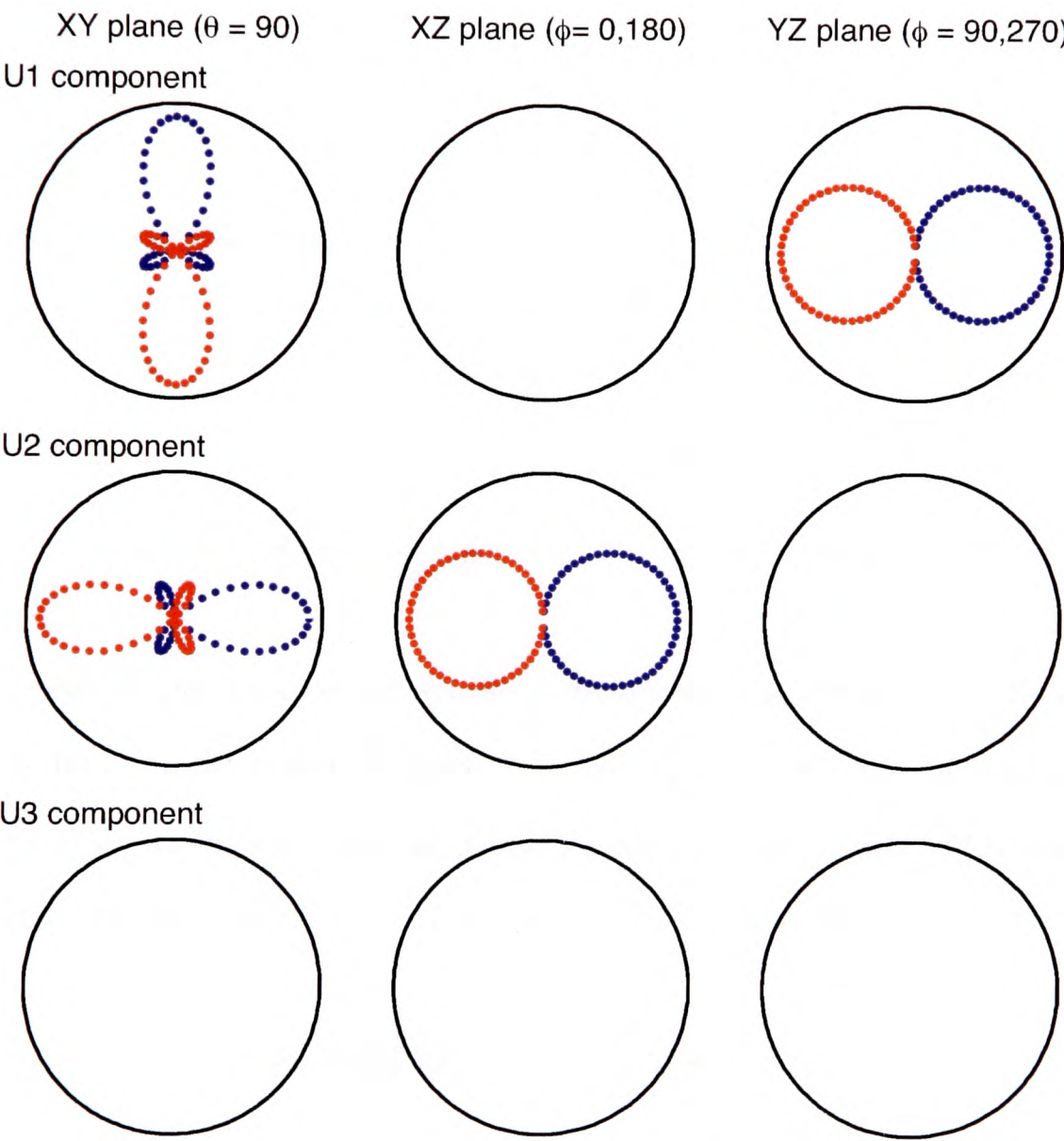


Figure 15.2: Radiation pattern from $[(G_{i1})_2 + (G_{i2})_1]$ term of equation 15.12. Layout is the same as for figure 15.1.

15.5 An expression for the scattered wavefield

The logical continuation of the discussion in the last section is to use the expressions for scattering radiation patterns in an expression for the scattered wavefield. We begin by substituting the appropriate Green's functions (equations 15.13 and 15.16) into 15.12 to obtain a solution for the scattered wavefield from a single point scatterer at the origin.

$$u_i = \frac{\omega^2}{4\pi r \beta^2 \rho} \left[\left(\delta_{i2} - \frac{x_i x_2}{r^2} \right) 2\rho \frac{\delta\beta}{\beta} + \left(\delta_{i1} \frac{x_2}{r} + \delta_{i2} \frac{x_1}{r} - \frac{2x_i x_1 x_2}{r^2} \right) \frac{\delta\mu}{\beta^2} \right] \times \exp[-i\omega\tau] \quad (15.18)$$

For clarity the following substitutions are made in equations 15.13, and 15.16:

$$\begin{aligned} A_i &= \left(\delta_{i2} - \frac{x_i x_2}{r^2} \right) \\ B_i &= \left(\delta_{i1} \frac{x_2}{r} + \delta_{i2} \frac{x_1}{r} - \frac{2x_i x_1 x_2}{r^2} \right) \\ R_s &= \frac{1}{4\pi r \beta^2} \end{aligned}$$

Where, A_i represents the radiation pattern due to $\delta\beta$ heterogeneity (figure 15.1), B_i is the radiation pattern due to $\delta\mu$ heterogeneity (figure 15.2), and the factor R_s represents the geometrical amplitude reduction due to wave front spreading in a homogenous medium. The expression for the scattered wavefield is now:

$$u_i = \omega^2 R_s \left(2A_i \frac{\delta\beta}{\beta} + B_i \frac{\delta\mu}{\rho\beta^2} \right) \exp[-i\omega\tau] \quad (15.19)$$

Using the substitution

$$\delta\mu = 2\rho\beta\delta\beta + \beta^2\delta\rho \quad (15.20)$$

We obtain:

$$u_i = \omega^2 R_s \left[2(A_i + B_i) \frac{\delta\beta}{\beta} + B_i \frac{\delta\rho}{\rho} \right] \exp[-i\omega\tau] \quad (15.21)$$

Equation 15.21 describes the scattered wavefield as a function of heterogeneities in S-wave velocity

and density. To gain a more useful solution (for the purposes of forward and inverse modelling) we assume that density anomalies are proportional to S-wave velocity anomalies (by a constant κ) in order to combine the two types of heterogeneity into a single scattering potential:

$$\begin{aligned}\frac{\delta\rho}{\rho} &= \kappa \frac{\delta\beta}{\beta} \\ \kappa &= \frac{d(\ln\rho)}{d(\ln\beta)}\end{aligned}$$

Estimates of κ vary between 0.2 and 0.4 (Ranalli (1996)). From this small range of possible values for κ , the uncertainty in its value is unlikely to affect the overall radiation pattern from a point scatterer significantly. Here, unless stated otherwise, we take $\kappa = 0.3$.

So, the final equation for the scattered S-wave from a point heterogeneity at the origin, with an incident S-wave travelling in the x_1 direction and polarised in the x_2 direction is:

$$u_i = \omega^2 R_s [2A_i + (2 + \kappa) B_i] \frac{\delta\beta}{\beta} \exp[-i\omega\tau] \quad (15.22)$$

Finally we consider the more realistic case of a incident S-wave polarised at an angle to our co-ordinate system. Until now we have only considered the radiation pattern from a point heterogeneity for a S-wave polarised in the x_2 direction. However, this is unlikely to be the case in reality. By swapping the axis in equations 15.14 and 15.17, it can be seen that the more general radiation patterns A'_i and B'_i will be given by:

$$A'_i = \left(\delta_{i2} - \frac{x_i x_2}{r^2} \right) \alpha_2 + \left(\delta_{i3} - \frac{x_i x_3}{r^2} \right) \alpha_3 \quad (15.23)$$

$$B'_i = \left(\delta_{i1} \frac{x_2}{r} + \delta_{i2} \frac{x_1}{r} - \frac{2x_i x_1 x_2}{r^3} \right) \alpha_2 + \left(\delta_{i1} \frac{x_3}{r} + \delta_{i3} \frac{x_1}{r} - \frac{2x_i x_1 x_3}{r^3} \right) \alpha_3 \quad (15.24)$$

Where α_2 and α_3 refer to the components of the incident S-wave in the x_2 and x_3 directions. Accordingly, our final solution for the i 'th component the scattered S-waves, at receiver $\mathbf{r}$ and

frequency ω , from a number of point heterogeneities, at positions $(\mathbf{x})$ will be given by:

$$u_i(\mathbf{r}, \mathbf{s}, \omega) = \omega^2 \int R_s(\mathbf{s}, \mathbf{x}, \mathbf{r}) (2A'_i(\mathbf{s}, \mathbf{x}, \mathbf{r}) + (2 + \kappa) B'_i(\mathbf{s}, \mathbf{x}, \mathbf{r})) \times \frac{\delta\beta}{\beta}(\mathbf{x}) \exp[-i\omega\tau(\mathbf{s}, \mathbf{x}, \mathbf{r})] d^3\mathbf{x} \quad (15.25)$$

Where $\mathbf{s}$ and $\mathbf{r}$ refer to a pair of source and receiver positions. Accordingly the quantities A_i , and B_i are functions of source-scatterer-receiver geometry, as is the generalised geometrical spreading $R_s(\mathbf{s}, \mathbf{x}, \mathbf{r})$. The strength of scattering at a given position is determined by scattering potential or $\delta\beta/\beta$.

15.6 Numerical Examples

In the last section an expression for the scattered wavefield as a function of S-wave velocity heterogeneity was derived using the Born approximation. Here equation 15.25 is used to produce examples of synthetic seismograms for S-to-S scattering from the lower mantle.

Equation 15.25 only gives the scattered wavefield due to a perturbation to a reference model (in this case PREM). Accordingly the formula is applied by specifying a series of point scatterers with strengths (or scattering potential) $\frac{\delta\beta}{\beta}$, and a source receiver geometry. The travel time for each scattered phase, τ , is determined by travel times in the reference model (PREM). While its amplitude is determined by the product of the scattering potential, the radiation patterns term and the wave front spreading factor, $R_s(\mathbf{s}, \mathbf{x}, \mathbf{r})$.

The radiation pattern terms are given by expressions 15.23 and 15.24, which give the scattered wave displacement in the co-ordinate system of the incident ray. This is then projected on to the component being modelled (in this case the transverse component). For the wave front spreading term, $R_s(\mathbf{s}, \mathbf{x}, \mathbf{r})$, we take the product of two spreading terms for propagation between source-scatterer and scatterer-receiver, $R_s(\mathbf{s}, \mathbf{x})R_s(\mathbf{x}, \mathbf{r})$.

The amplitude decay due to wave front spreading along a ray going from point $\mathbf{x}$ to point $\mathbf{y}$

in spherically symmetric media is given by (for example see Aki & Richards (1980)):

$$R_s(\mathbf{x}, \mathbf{y}) = \sqrt{\frac{v_{\mathbf{x}} \rho_{\mathbf{x}}}{v_{\mathbf{y}} \rho_{\mathbf{y}}}} \times \frac{1}{4\pi v_{\mathbf{x}}^2} \times \left| r_{\mathbf{x}} r_{\mathbf{y}} p \frac{v_{\mathbf{y}}}{v_{\mathbf{x}}} \cot i_{\mathbf{x}} \cot i_{\mathbf{y}} \sin \Delta \frac{d\Delta}{dp} \right|^{-\frac{1}{2}} \quad (15.26)$$

In the above equations r refers to radius, Δ epicentral distance, $i_{\mathbf{x}}$ is the angle between the ray and the radial vector at $\mathbf{x}$, and p is ray parameter which connects $\mathbf{x}$ and $\mathbf{y}$. All quantities in equation 15.26 can be determined by tracing rays through a reference model.

Note that all the terms in equation 15.25 (travel time, radiation pattern, and geometrical spreading) can be calculated relative to a reference phase. In which case the first term of equation 15.26 will be cancelled out. This also makes the small perturbations required when using the Born approximation more realistic.

In figure 15.3 two examples of seismograms and their amplitude spectra are shown. Example, A, shows the signal from a point scatterer located at 2690km depth half way between the source and the receiver. The amplitude spectra shows the contribution of the ω^2 frequency multiplier. In the time domain this produces a negative second time derivative. Effectively this means that the scattered wavefield is proportional to the acceleration of the incident wavefield, rather than its displacement.

In the second example scattered waves were calculated for a grid of point scatterers, with 10km spacing, at 2690km depth. Since the spacing between scatterers is only a small fraction of the wavelength, this can be regarded as equivalent to calculating synthetic reflections from an infinitely thin continuous reflector (Kirchhoff scattering). The amplitude spectrum and waveforms are consistent with the predictions of a $i\omega$ factor in the frequency spectrum for the scattered waves.

In the final example (figure 15.4) scattered arrivals from a layer of material extending from 2640km to 2840km depth, with $\frac{\delta\beta}{\beta}$ of 1.5%. WKBJ synthetic seismograms which included reflections from the top and the bottom of the layer are also shown. As opposed to the previous examples a causality constraint has been placed on the scattered arrivals by filtering the traces with a causal filter. While the Born approximate synthetics are not in exact agreement with

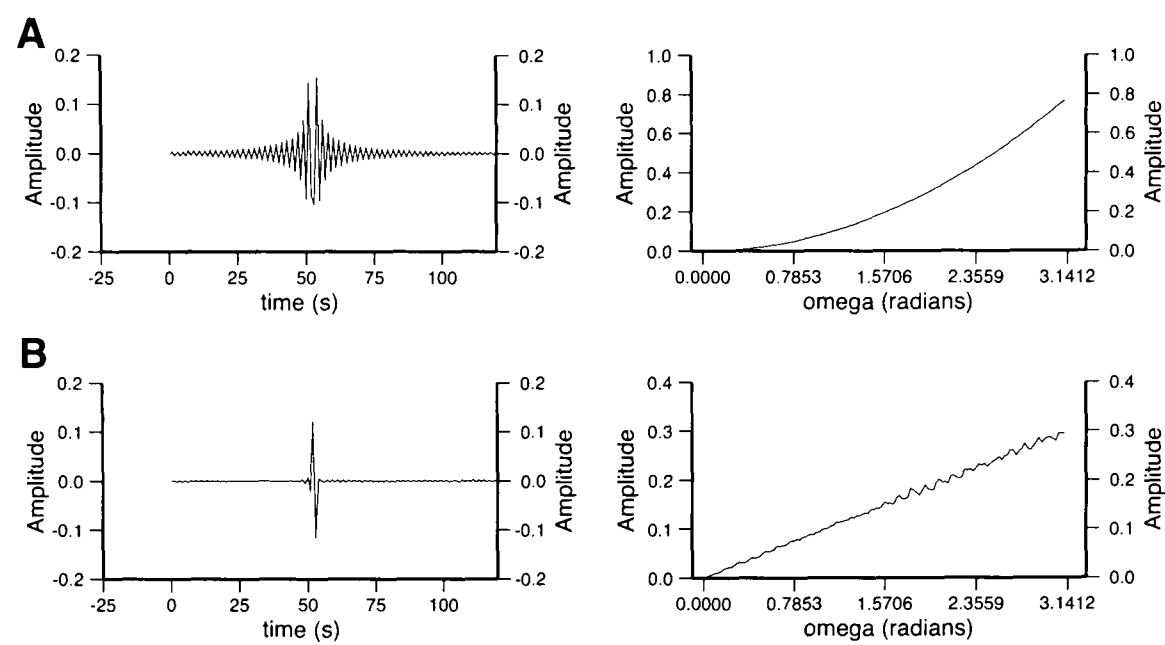


Figure 15.3: Scattered S-waves from **A** a point scatterer at 2690km depth, **B** Thin horizontal layer at 2690km depth. On the left is the synthetic seismograms on the right is the frequency spectrum. In each case the source receiver distance is 65° and an isotropic shear source is assumed.

the WKBJ calculations the general pulse shapes are similar. The contributions from the point scatterers add up to give a positive reflection consistent with the top of the layer followed a few seconds latter by a negative reflection consistent with the layer base.

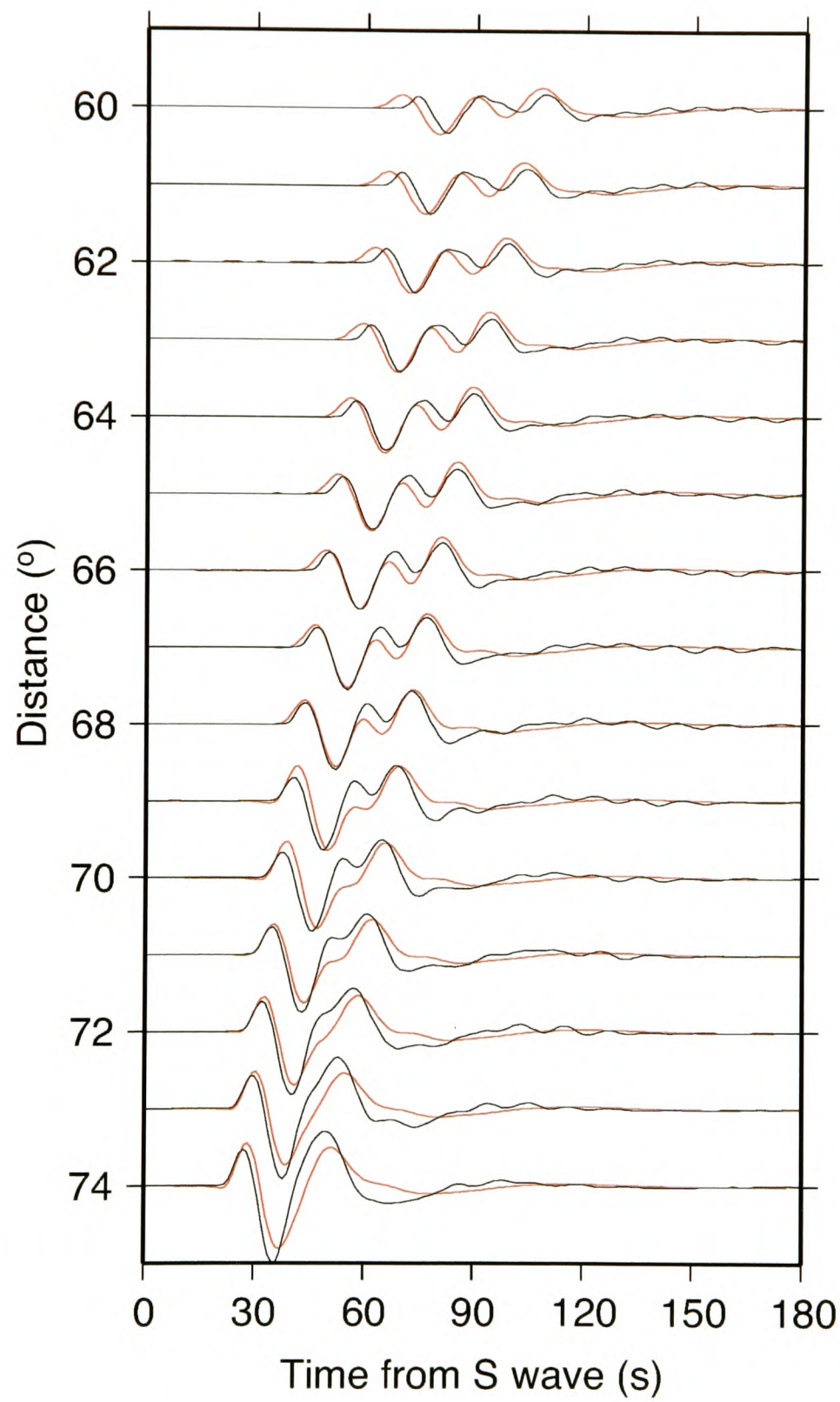


Figure 15.4: Record section for Born approximate (black) and WKBJ (red) synthetics for a slab between 2640 and 2840km depth with a velocity contrast of 1.5%. For the Born approximate synthetics the slab was simulated by placing point scatters at 50km intervals. The WKBJ synthetic traces were made using reflections from the top and base of the layer, and scaled to the peak amplitude of the direct S pulse. Records were filtered with a causal Butterworth filter (15-75s bandpass powers of 6 and 4, same as the data). Epicentral distances are shown on the left.

Chapter 16

The Generalised Radon

Transform: Theory and method

16.1 Introduction

The forward model for the scattered wavefield, as developed in chapter 15, relates the scattered wavefield to a sum of contributions from point heterogeneities (characterised by a velocity perturbation). The object of this chapter is to provide a theoretical framework for the inverse process of retrieving the medium heterogeneity from the observed scattered wavefield.

Under the Born approximation the scattered wavefield is a linear sum of contributions from point scatterers. As such it is possible to derive scattering quantities ($\frac{\delta\beta}{\beta}$, $\frac{\delta\rho}{\rho}$ etc.) by generalised matrix inversion. Bostock & Rodenay (1999) and Frederiksen & Revenaugh (2000) consider reconstructing scattering potential in the crust and shallow mantle beneath dense seismic networks this way. Both studies had to make approximations due to computational limitations and the size of the inverse problem they were presented with, such as inverting subsets of the data, or using 2.5-D approximations. The larger scattering volume being considered here (potentially the whole mantle) combined with the sparse and irregular spatial coverage of the global network dataset

means that any linear inversion will be very poorly posed.

Instead migration techniques developed for medical physics and exploration seismology are used to reconstruct scattering potential in the mantle. Such techniques retrieve scattering potential by performing weighted summations over all the points from which a scattered arrival could have originated. In an earlier chapter (chapter 14) this summation was performed using uniform weights to locate scatterers in the lower mantle. If the weights in the summation are the same as the row elements of an inverse matrix then the process is equivalent to solving the inverse problem outlined above. However, in practice incomplete spatial coverage will result in differences between the two procedures.

The process of how weights for a migration can be calculated using the Generalised Radon Transform (GRT) is described in this chapter. The theory has been developed extensively for acoustic scattering by Beylkin (1985); Miller *et al.* (1987); Beylkin & Burridge (1990) and Ikelé *et al.* (1992), for isotropic elastic media by Beylkin & Burridge (1990) and Ikelé *et al.* (1992), and for anisotropic elastic media by de Hoop & Bleistein (1997) and de Hoop *et al.* (1999). Here the aim is to show how the migration of seismic data can be viewed as a weighted inverse Radon transform, and hence obtain a useful migration formula for isotropic S-to-S scattering.

The theoretical framework here follows that of Miller *et al.* (1987), where a migration formula is derived by analogy to the Radon transform for the acoustic case. This is easily extended to the elastic case, however as we shall see, the nulls in the radiation patterns from sources and point scatterers require the damping of contributions from scatterers near these nodes.

16.2 The classical Radon Transform

The Radon transform of a 3-D function $f(\mathbf{x})$ at point $\mathbf{x}$ is given by the integral of the function over a plane passing through $\mathbf{x}$. The values of $\mathbf{x}$ on a plane with unit normal, $\boldsymbol{\xi}$ is given by $p - \boldsymbol{\xi} \cdot \mathbf{x}$, where p is the distance from the plane to the origin. So the Radon transform in 2 or more

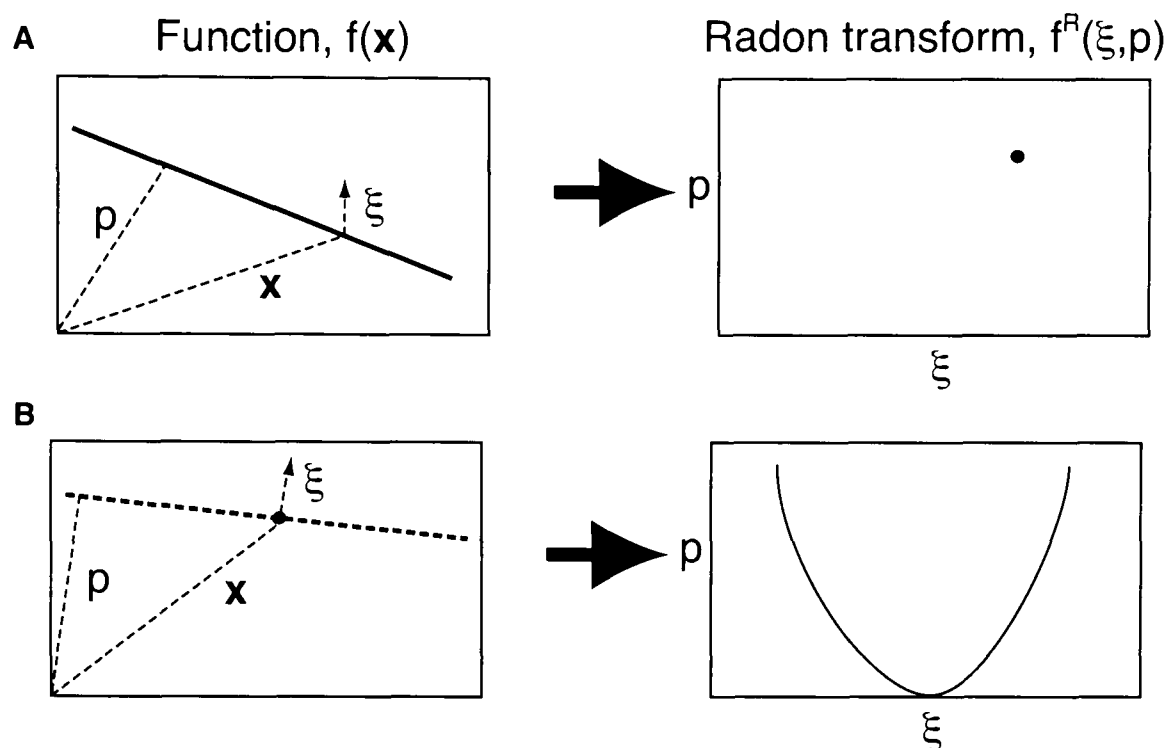


Figure 16.1: Demonstration of the Radon transform for a 2-D line or 3-D fence **A**, which is mapped to a point ξ, p where ξ will be a unit normal vector and p is the distance to the origin. **B** Radon transform of single point, which is spread out over a family of planes

dimensions is given by:

$$f^R(\xi, p) = \int \delta(p - \xi \cdot \mathbf{x}) f(\mathbf{x}) d^3 \mathbf{x} \quad (16.1)$$

Where $f^R(\xi, p)$ is the Radon transform of $f(\mathbf{x})$. Figure 16.1 demonstrates how under the Radon transform a 2-D line is mapped to a single point.

The inverse Radon transform (for a 3-D function) recovers the original function by integrating the second differential (with respect to p) of $f^R(\xi, p)$ over all possible planes ξ (Miller *et al.* (1987)).

$$f(\mathbf{x}) = -\frac{1}{8\pi^2} \int \frac{\partial^2}{\partial p^2} [f^R(\xi, p = \xi \cdot \mathbf{x})] d^2 \xi \quad (16.2)$$

Equation 16.2 is a filtered backprojection. A family of planes corresponding to the image point $\mathbf{x}$ (i.e. all the planes which pass through the image point) is selected, and the second differential with respect to p filters this function. The value at the image point is then regained by integrating the twice differentiated function with respect to the orientation of each plane in the family. This is done by letting the unit normal ξ vary over the surface of a sphere. This process is illustrated

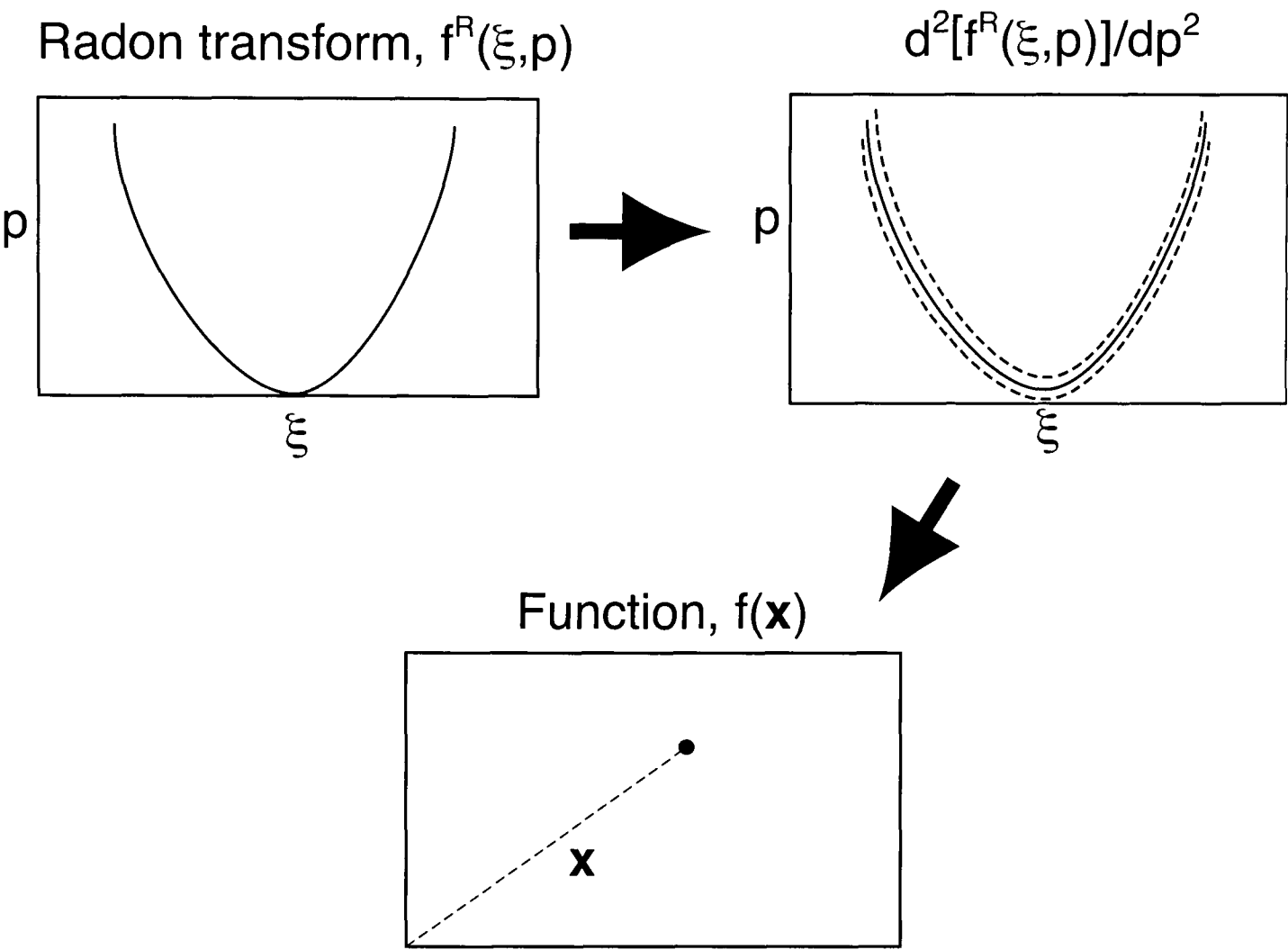


Figure 16.2: Demonstration of the Inverse Radon transform as a filtered backprojection, The value at $\mathbf{x}$ $f(\mathbf{x})$ is derived by taking the second derivative of the radon transform $f^R(\xi, p)$ with respect to distance from the origin p (filtering step), and then integrating over the region in Radon space which satisfies image point $\mathbf{x}$ (backprojection step).

in figure 16.2.

We can write the filtered function of equation 16.2 in terms of the forward transform:

$$\begin{aligned} \frac{\partial^2}{\partial p^2} f^R(\xi, p) &= \frac{\partial^2}{\partial p^2} \int \delta(p - \xi \cdot \mathbf{x}) f(\mathbf{x}) d^3 \mathbf{x} \\ &= \int \delta''(p - \xi \cdot \mathbf{x}) f(\mathbf{x}) d^3 \mathbf{x} \end{aligned}$$

Then combine forward and inverse transforms into a single equation. This allows us to derive an

estimate of the original function $\langle f(\mathbf{x}_0) \rangle$ at an image point, $\mathbf{x}_0$.

$$\begin{aligned}\langle f(\mathbf{x}_0) \rangle &= -\frac{1}{8\pi^2} \int d^2\xi \int \delta''(p - \xi \cdot \mathbf{x}) f(\mathbf{x}) d^3\mathbf{x} \\ &= -\frac{1}{8\pi^2} \int d^2\xi \int \delta''(\xi \cdot (\mathbf{x}_0 - \mathbf{x})) f(\mathbf{x}) d^3\mathbf{x}\end{aligned}\tag{16.3}$$

In the above equation the original function is estimated at an image point $\mathbf{x}_0$, from the second differential of the function summed over a particular plane. For example the first integral produces the second derivative (with respect to distance from the origin, p) for a family of planes specified by $p = \xi \cdot \mathbf{x}_0$. The outer integral sums this function over all possible unit normals ξ . Alternatively, given a function which is twice differentiated with respect to parallel planes, equation 16.3 shows the original function at a point can be reconstructed by integrating over all planes which pass through that point.

16.3 Relation between Radon Transform and migration

Before the application of expression 16.3 to scattered seismic waves the analogy between the reconstruction of scattering potential within the Earth (seismic migration) and the process of reconstructing a function from its second differential by integrating over planes is discussed. In section 15.5 the i 'th component of scattered wavefield, u_i , at receiver, r , for a given source, s , was described as a sum of contributions from point scatterers with scattering potential $\frac{\delta\beta}{\beta}(\mathbf{x}) = f(\mathbf{x})$ (equation 15.25):

$$u_i(\mathbf{r}, \mathbf{s}, \omega) = \omega^2 \int A_i(\mathbf{s}, \mathbf{x}, \mathbf{r}) f(\mathbf{x}) \exp[-i\omega\tau(\mathbf{s}, \mathbf{x}, \mathbf{r})] d^3\mathbf{x}\tag{16.4}$$

Where point scatterers are located at positions $\mathbf{x}$, and the amplitude and arrival time of a scattered wave are given by $A_i(\mathbf{s}, \mathbf{x}, \mathbf{r})$ and $\tau(\mathbf{s}, \mathbf{x}, \mathbf{r})$ respectively. In the time domain the shape of the scattered wave becomes a negative second differential of a delta function.

$$u_i(\mathbf{r}, \mathbf{s}, t) = - \int A_i(\mathbf{s}, \mathbf{x}, \mathbf{r}) f(\mathbf{x}) \delta''(t - \tau(\mathbf{s}, \mathbf{x}, \mathbf{r})) d^3\mathbf{x}\tag{16.5}$$

From equation 16.5 it can be seen that an amplitude at time, t , can be seen as the sum of weighted contributions from an isochron surface which is specified by $t = \tau(\mathbf{s}, \mathbf{x}, \mathbf{r})$. For a fixed source-receiver pair this will be an ellipse (or scattering sausage) with exact shape depending on the velocity of the material.

Comparison of equation 16.5 with the inner integral of equation 16.3, shows that time, t , in the former is comparable with distance from the origin, p , in the Radon transform. Meanwhile the source-receiver geometry will specify a plane or unit normal, ξ . Accordingly the solution for scattering potential at a point, $\mathbf{x}_0$, can be seen as a weighted average over a unit sphere of all the time samples which could result from a scatterer at $\mathbf{x}_0$. Figure 16.3 attempts to summarise this procedure.

16.4 Application of Radon Transform to migration

The last section hinted at a way of using the combined forward and inverse Radon transform equation 16.3 to derive scattering potential observations of the scattered wavefield. Here this analogy is made specific. By recasting the scattered wavefield at a given time, t , as a summation over normals to the isochron surface an analogy with equation 16.3 is obtained. What follows is a specific method for reconstructing scattering potential in the Earth.

We begin by shifting the origin of co-ordinates to an image point $\mathbf{x}_0$, such that $\mathbf{x} = \mathbf{x}_0 + \mathbf{y}$. Accordingly:

$$u_i(\mathbf{r}, \mathbf{s}, t) = - \int A_i(\mathbf{s}, \mathbf{x}_0 + \mathbf{y}, \mathbf{r}) f(\mathbf{x}_0 + \mathbf{y}) \delta''(t - \tau(\mathbf{s}, \mathbf{x}_0 + \mathbf{y}, \mathbf{r})) d^3\mathbf{y} \quad (16.6)$$

Assuming $|\mathbf{y}|$ is small allows us to make some simplifications. Firstly the travel time τ is replaced with its first order Taylor series.

$$\tau(\mathbf{s}, \mathbf{x}_0 + \mathbf{y}, \mathbf{r}) \approx \tau(\mathbf{s}, \mathbf{x}_0, \mathbf{r}) + \nabla \tau(\mathbf{s}, \mathbf{x}_0, \mathbf{r}) \cdot \mathbf{y} \quad (16.7)$$

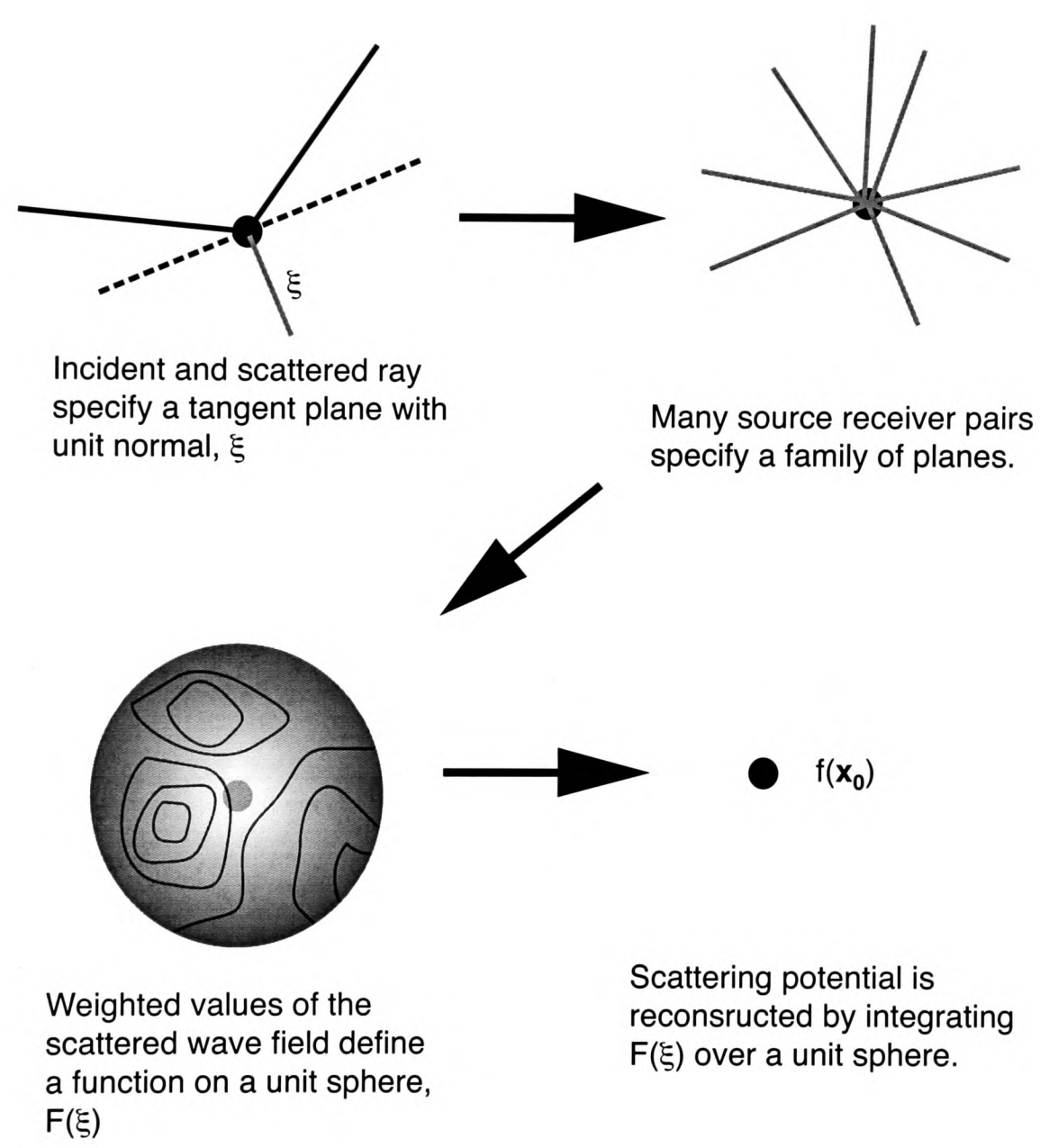


Figure 16.3: Outline of how the Inverse Radon transform is used to obtain scattering potential at a point $f(\mathbf{x}_0)$.

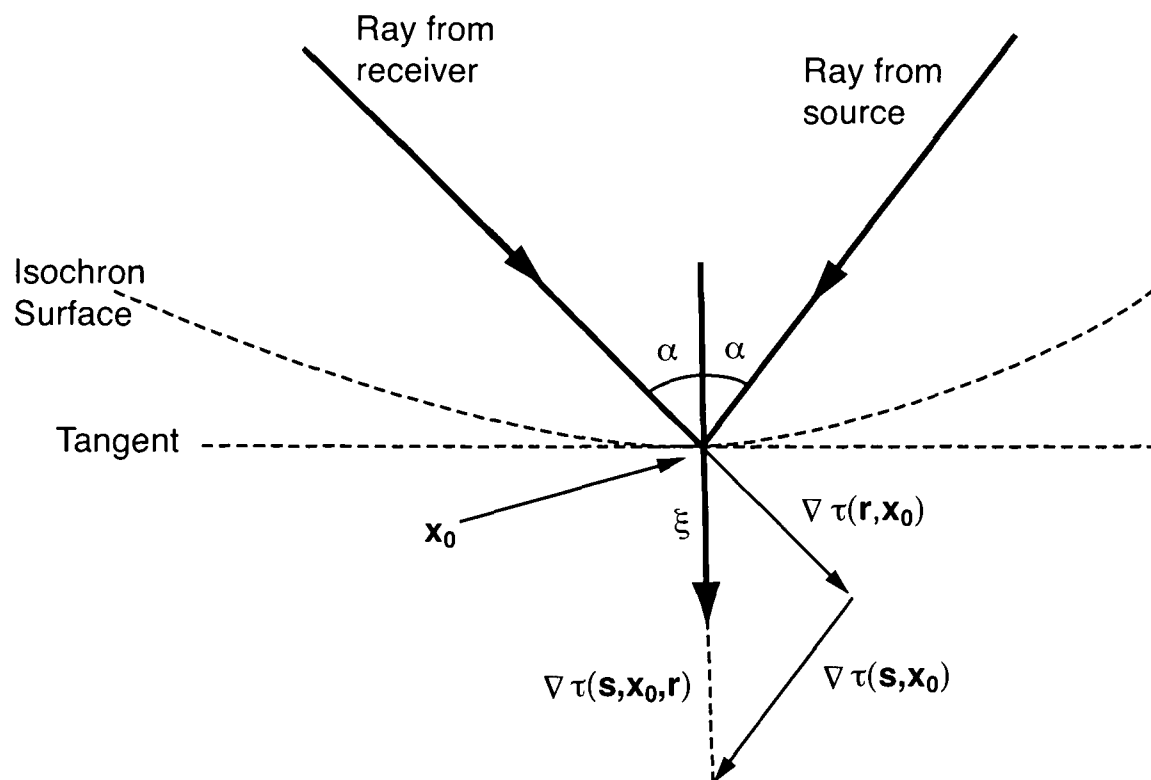


Figure 16.4: Ray geometry for a point scatterer at position $\mathbf{x}_0$. The source and receiver side travel time gradients will have length of $1/v(\mathbf{x}_0)$, where $v(\mathbf{x}_0)$ is the velocity at $\mathbf{x}_0$. Hence the total travel time gradient $\nabla\tau(\mathbf{s}, \mathbf{x}_0, \mathbf{r})$ for S-to-S or P-to-P scattering has direction $\boldsymbol{\xi}$ normal to the isochron surface and magnitude, $2|\cos \alpha|/v(\mathbf{x}_0)$. Figure adapted from Miller *et al.* (1987)

The gradient vector for the travel time $\nabla\tau(\mathbf{s}, \mathbf{x}_0, \mathbf{r})$ is given by the sum of the gradients for the incident and scattered ray paths (see figure 16.4 or Miller *et al.* (1987) figure 6).

$$\nabla\tau(\mathbf{s}, \mathbf{x}_0, \mathbf{r}) = \nabla\tau(\mathbf{s}, \mathbf{x}_0) + \nabla\tau(\mathbf{x}_0, \mathbf{r})$$

Figure 16.4 shows how the travel time gradient lies normal to the isochron surface, having unit vector $\boldsymbol{\xi}(\mathbf{r}, \mathbf{x}_0, \mathbf{s})$. It can be shown that for S-to-S and P-to-P scattering the travel time gradient has magnitude related to the velocity of the medium $v(\mathbf{x}_0)$ and the half angle between the incident and scattered rays, α . Accordingly:

$$\nabla\tau(\mathbf{s}, \mathbf{x}_0, \mathbf{r}) = \frac{2|\cos \alpha|}{v(\mathbf{x}_0)} \boldsymbol{\xi}(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) \quad (16.8)$$

The second approximation is to assume that the amplitude function $A_i(\mathbf{s}, \mathbf{x}_0 + \mathbf{y}, \mathbf{r})$ is ap-

proximately constant about $\mathbf{x}_0$, and so is moved outside the integral. Substituting equations 16.7 and 16.8 into 16.6 gives:

$$u_i(\mathbf{r}, \mathbf{s}, t) \approx -A_i(\mathbf{s}, \mathbf{x}_0, \mathbf{r}) \int \delta'' \left(t - \left(\tau_0 + \frac{2|\cos \alpha|}{v(\mathbf{x}_0)} \boldsymbol{\xi}(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) \cdot \mathbf{y} \right) \right) f(\mathbf{x}_0 + \mathbf{y}) d^3 \mathbf{y} \quad (16.9)$$

Where τ_0 is the travel time associated with image point $\mathbf{x}_0$. Next we select this particular image point by taking $t = \tau_0$.

$$u_i(\mathbf{r}, \mathbf{s}, t = \tau_0) \approx -A_i(\mathbf{s}, \mathbf{x}_0, \mathbf{r}) \int \delta'' \left(-\frac{2|\cos \alpha|}{v(\mathbf{x}_0)} \boldsymbol{\xi}(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) \cdot \mathbf{y} \right) f(\mathbf{x}_0 + \mathbf{y}) d^3 \mathbf{y} \quad (16.10)$$

Using the relation $\delta''(-ax) = \delta''(x)/|a|^3$ we get:

$$u_i(\mathbf{r}, \mathbf{s}, t = \tau_0) \approx -A_i(\mathbf{s}, \mathbf{x}_0, \mathbf{r}) \int -\frac{v(\mathbf{x}_0)^3}{8|\cos \alpha|^3} \delta''(\boldsymbol{\xi}(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) \cdot \mathbf{y}) f(\mathbf{x}_0 + \mathbf{y}) d^3 \mathbf{y} \quad (16.11)$$

Finally we replace the original variable $\mathbf{x} = \mathbf{x}_0 + \mathbf{y}$. Which gives:

$$u_i(\mathbf{r}, \mathbf{s}, t = \tau_0) \approx A_i(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) \int \frac{v(\mathbf{x}_0)^3}{8|\cos \alpha|^3} \delta''(\boldsymbol{\xi}(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) \cdot (\mathbf{x} - \mathbf{x}_0)) f(\mathbf{x}) d^3 \mathbf{x} \quad (16.12)$$

The right hand side of equation 16.12 is in a comparable format to the inner integral of equation 16.3. Accordingly we can substitute the scattered wavefield into 16.3 to obtain the migration formula for an estimate of the object function (in this case scattering potential) at $\langle f(\mathbf{x}_0) \rangle$ (Miller *et al.* (1987)):

$$\langle f(\mathbf{x}_0) \rangle = \frac{1}{\pi^2} \int \frac{|\cos \alpha(\mathbf{r}, \mathbf{x}_0, \mathbf{s})|^3}{v(\mathbf{x}_0)^3} \frac{1}{A_i(\mathbf{r}, \mathbf{x}_0, \mathbf{s})} u_i(\mathbf{r}, \mathbf{s}, t = \tau_0) d^2 \boldsymbol{\xi}(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) \quad (16.13)$$

16.5 Practical application of the Generalised Radon Transform

16.5.1 Damping the radiation pattern term

In an ideal scenario it is possible to apply equation 16.13 directly to a dataset. However, the presence of noise requires some minor alteration to the theoretical amplitude term, $A_i(\mathbf{r}, \mathbf{x}_0, \mathbf{s})$. In the acoustic case, examined by Miller *et al.* (1987), A_i is always non-zero. However, the introduction of the source and scatterer radiation patterns means that certain scatterer positions can result in zero amplitudes, which correspond to infinite weights in the integral of equation 16.13. In an ideal world this would not be a problem as these positions would also have a corresponding zero amplitude for the data. However, the world is not perfect and an undamped integral such as equation 16.13, amplifies noise at the expense of data.

Accordingly, an altered version of the equation 16.13 is used. By replacing the amplitude term with those which describe wave front spreading, $R_s(\mathbf{r}, \mathbf{x}_0, \mathbf{s})$, and a term for the damped radiation pattern, $\Gamma(\mathbf{r}, \mathbf{x}_0, \mathbf{s}, \sigma)$, the migration formula becomes:

$$\langle f(\mathbf{x}_0) \rangle = \frac{1}{\pi^2} \int \frac{|\cos \alpha(\mathbf{r}, \mathbf{x}_0, \mathbf{s})|^3}{v_0(\mathbf{x}_0)^3} \frac{\Gamma(\mathbf{r}, \mathbf{x}_0, \mathbf{s}, \sigma)}{R_s(\mathbf{r}, \mathbf{x}_0, \mathbf{s})} u(\mathbf{r}, \mathbf{s}, t = \tau_0) d^2 \boldsymbol{\xi}(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) \quad (16.14)$$

Where the damping of the radiation pattern is controlled by the value of σ using

$$\Gamma(\mathbf{r}, \mathbf{x}_0, \mathbf{s}, \sigma) = \begin{cases} \frac{1}{G(\mathbf{r}, \mathbf{x}_0, \mathbf{s})} & |G(\mathbf{r}, \mathbf{x}_0, \mathbf{s})| \geq \sigma \\ \frac{G(\mathbf{r}, \mathbf{x}_0, \mathbf{s})}{\sigma^2} \exp \left[1 - \left(\frac{G(\mathbf{r}, \mathbf{x}_0, \mathbf{s})}{\sigma} \right)^2 \right] & |G(\mathbf{r}, \mathbf{x}_0, \mathbf{s})| < \sigma \end{cases} \quad (16.15)$$

$$G(\mathbf{r}, \mathbf{x}_0, \mathbf{s}) = (2A_i + (2 + \kappa)B_i) \epsilon_i \quad (16.16)$$

A_i and B_i are the point scatterer radiation pattern terms, defined in equations 15.23 and 15.24, and the transverse polarisation vector is specified by the elements ϵ_i .

Figure 16.5 shows the effect of applying this damping curve to the weights in the migration

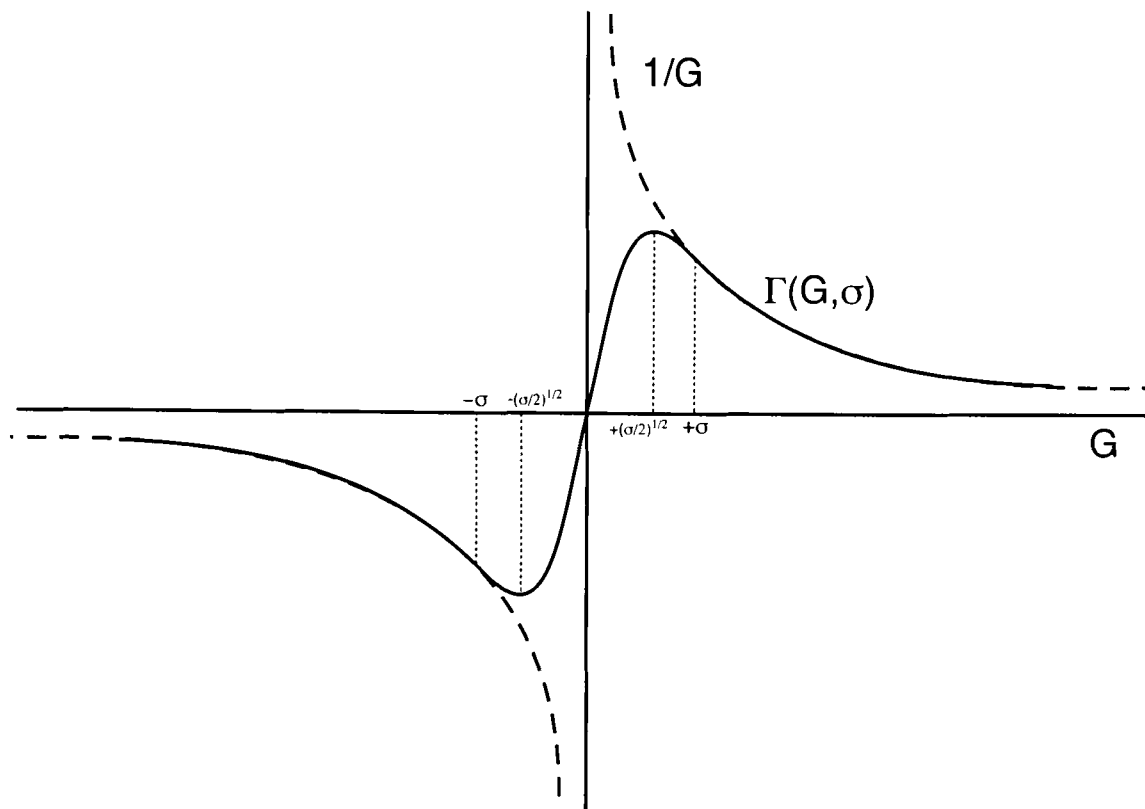


Figure 16.5: Illustration of the radiation pattern damping term $\Gamma(\mathbf{r}, \mathbf{x}_0, \mathbf{s}, \sigma)$, which is compared to an the undamped weight which will depend in $\frac{1}{G(\mathbf{r}, \mathbf{x}_0, \mathbf{s})}$. The weights from amplitudes higher than the damping parameter σ are left unaffected, but the contributions from lower amplitudes are tapered to zero.

integral. It can be seen from equations 16.15 and figure 16.5 that the weights will be left unchanged for amplitudes larger than σ , but the weights for lower amplitudes are reduced to zero. Testing showed that reasonable results were obtained using $\sigma=0.5$. In figure 16.6 the weights for different points in the Earth plotted along the equal travel time surface are shown. It can be seen that the major effect of the damping is to lower the weights for points lying along nodes in the source radiation pattern. The final weights used in the migration are shown in figure 16.7, as well as those relating to the radiation pattern, spreading, and travel time gradient terms, of equation 16.14.

16.5.2 Creating an acausal wavelet

The application of equation 16.13 or 16.14 assumes a zero phase acausal wavelet such as filtered versions of those shown in figure 15.3. However, the residual traces produced by phase stripping are causal, not zero phase, and still contain the source-time function. So before application of the migration equation the traces must be converted to a zero phase acausal form, and as much as

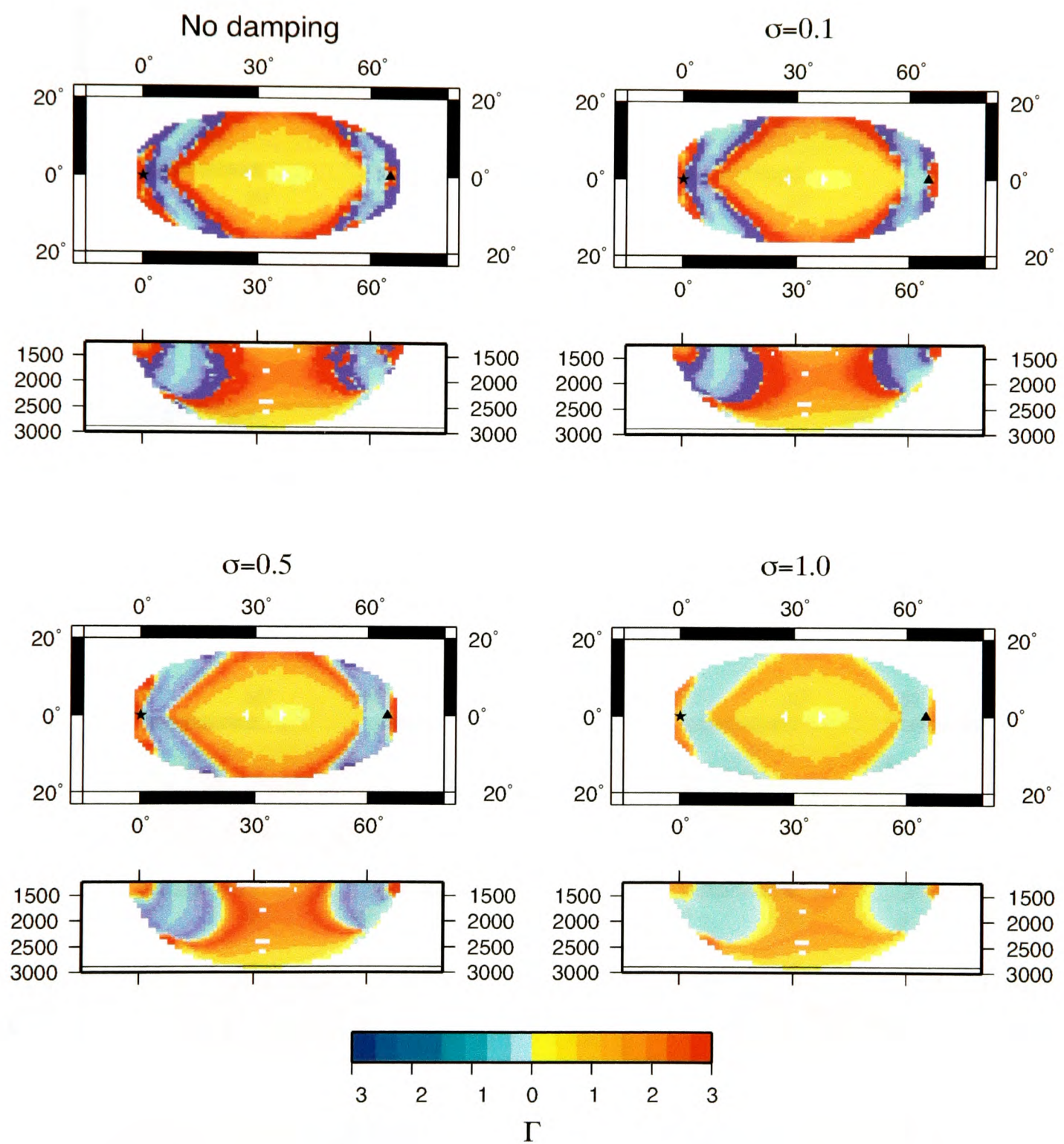


Figure 16.6: Plots of the radiation pattern damping term $\Gamma(\mathbf{r}, \mathbf{x}_0, \mathbf{s}, \sigma)$ plotted on travel time surface (viewed from above, and from the south), for different values of the damping parameter, σ (see text for details). The star marks the source position and triangle the receiver, in all cases a North-South polarised strike slip double couple source was used and amplitude is taken along the transverse component at the triangle. Note that the use of the damping eliminates singularities in the radiation pattern term.

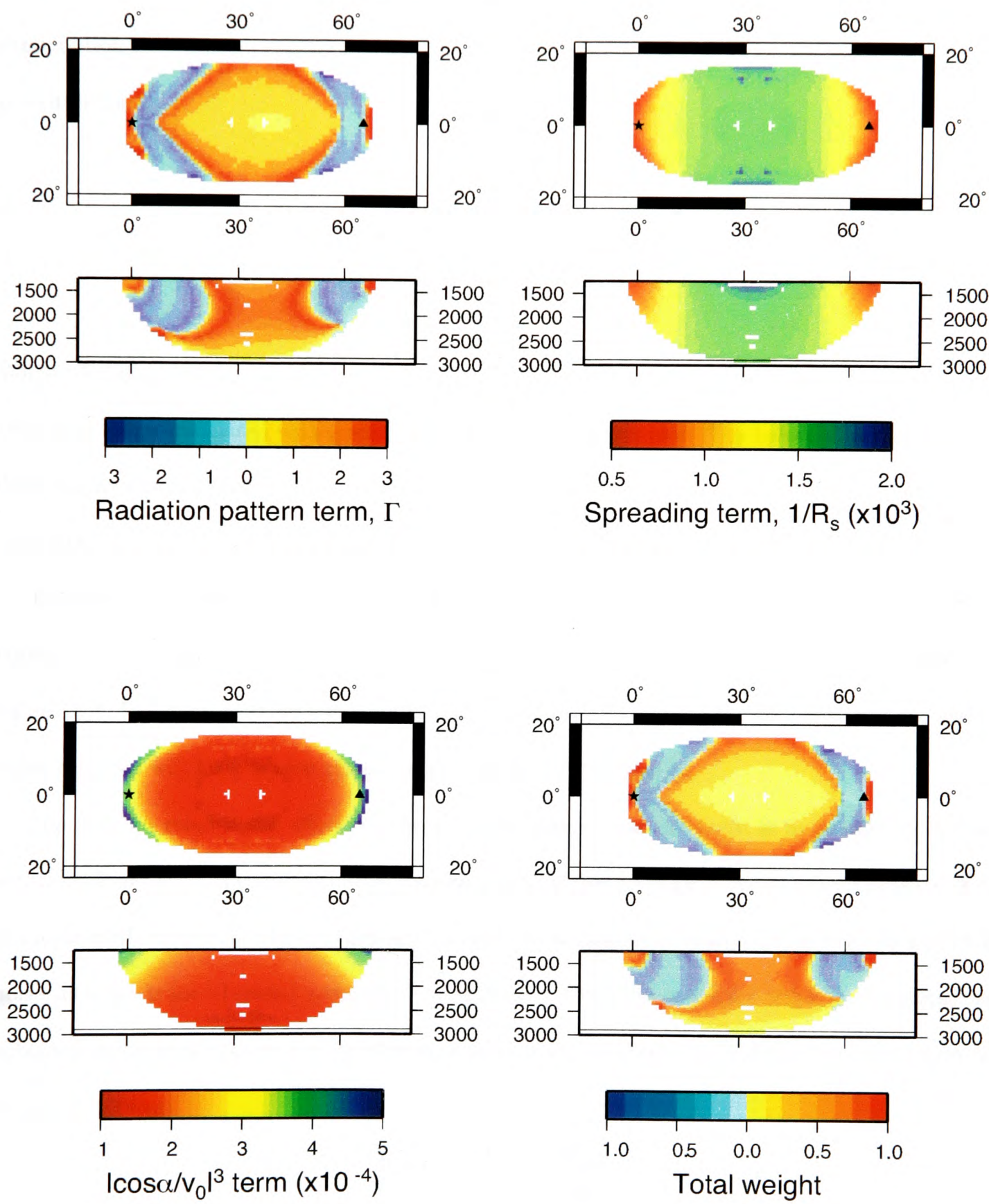


Figure 16.7: Total weights for the GRT migration integral and the contributions from damped ($\sigma = 0.5$) source-scatterer radiation patterns, wave front spreading, and the travel time gradient. As in figure 16.6 the weights are plotted along an equal travel time surface, which are viewed from above and the side. Source station geometry is the same as for figure 16.6

possible of the source-time function removed from the trace.

The residual trace (the data) produced by a single scatterer, u , can be considered as the contribution from a point scatterer convolved with a reference pulse. In the frequency domain this is written as:

$$u(\omega) = \omega^2 A \exp[-i\omega\tau] \times |R(\omega)| \exp[i\theta_R(\omega)] \quad (16.17)$$

$$= \omega^2 A |R(\omega)| \exp[i(\theta_R(\omega) - \omega\tau)] \quad (16.18)$$

Where A describes the source/scatterer radiation patterns and wave front spreading, while τ is the phase delay associated with a scatterer and source-receiver pair. The terms $|R(\omega)|$ and $\theta_R(\omega)$ describe the amplitude and phase spectra of the source respectively, which we wish to remove from the trace.

Equation 16.18 describes the wavelet from a point scatterer as a time delayed, twice differentiated with respect to time, and scaled version of the reference pulse. This effect is shown in figure 16.8. It follows that we can use the reference pulse to estimate the source signature, which when removed will leave a zero phase wavelet suitable for migration.

One way to remove the effects of the reference pulse from the trace is to perform a deconvolution. However, time domain least squares methods are unlikely to work since the wavelet is not necessarily minimum phase. Testing also showed that spectral division introduced long-period noise into the trace. Instead reduction of the impact from the reference pulse is attempted by subtracting the phase spectra of the reference pulse from the trace. Accordingly we define a phase subtraction filter:

$$F(\omega) = \frac{\exp(-i\theta_R(\omega))}{\sqrt{\int |R(\omega)|^2 d\omega}} \quad (16.19)$$

Figure 16.8 C, shows the application of the phase subtraction filter to a scattered wavelet. It can be seen that a zero phase wavelet is created with slightly lower peak amplitude and larger side lobes than the ideal scattered wave. This is because the phase subtraction filter does not correct for the shape of the source amplitude spectrum. This can also be seen by putting equations 16.18

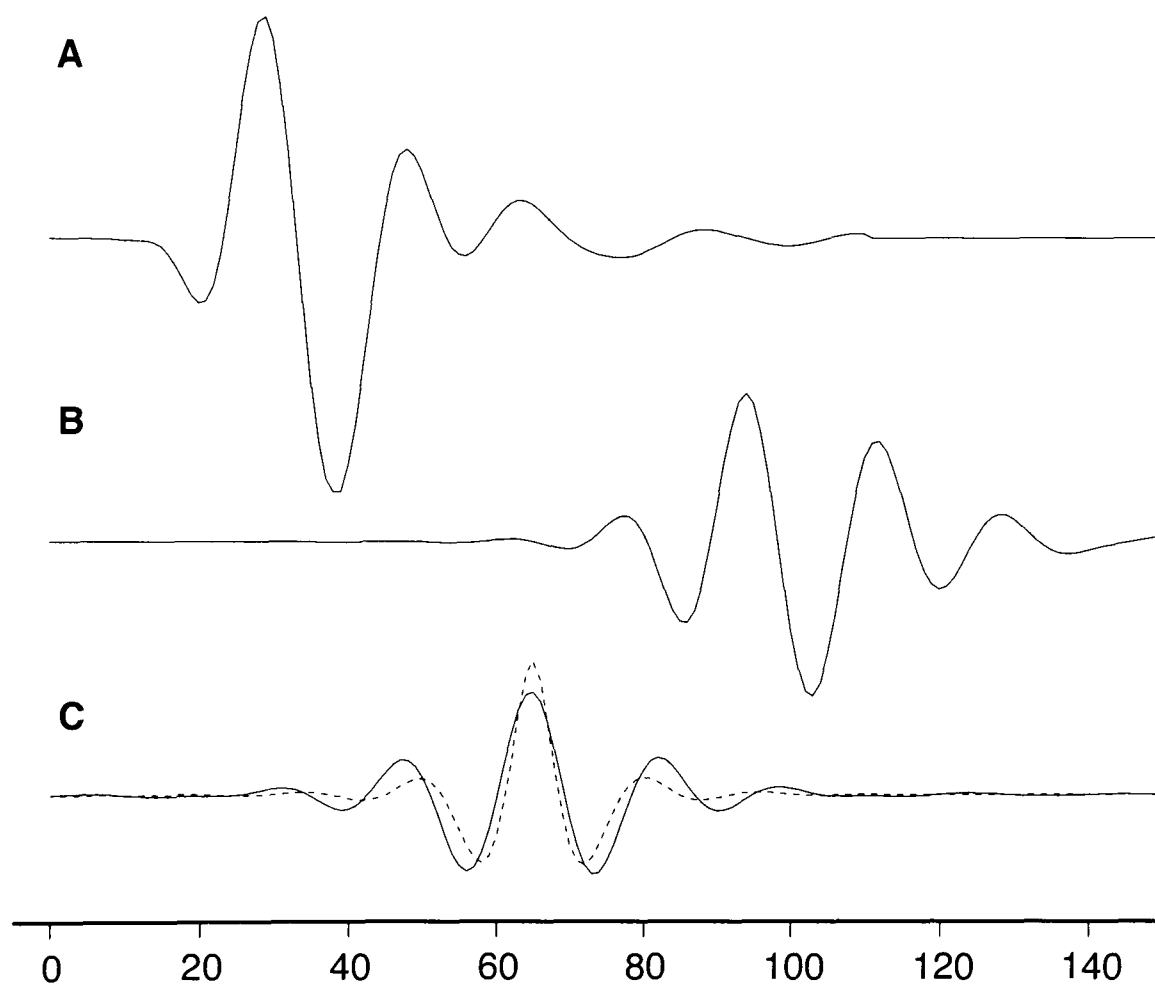


Figure 16.8: Example of application of phase subtraction filter. **A** Example reference pulse generated for phase stripping, $R(\omega) \exp[i\theta_R]$. **B** Example scattered trace, $\omega^2 A |R(\omega)| \exp[i(\theta_R - \omega\tau)]$. **C** (Solid) application of the phase subtraction filter to B. and (dashed) ideal scattered wavelet, $\omega^2 A \exp[-i\omega\tau]$.

and 16.19 together.

$$u'(\omega) = F(\omega)u(\omega) \quad (16.20)$$

$$= \frac{|R(\omega)|}{\sqrt{\int |R(\omega)|^2 d\omega}} A \exp[-i\omega\tau] \quad (16.21)$$

In other words, u' , is the ideal scattered wavelet, $A \exp[-i\omega\tau]$, modulated by the normalised amplitude spectrum of the reference wavelet. It is the normalised amplitude spectrum of the reference wavelet that creates the lower peak amplitudes and larger side lobes seen in figure 16.8C.

The result of equation 16.21 is similar to that obtained by cross-correlating the trace with a reference pulse, as is common in processing vibroseis data (Sheriff & Geldart (1995)). However, in this case the resulting wavelet is modulated by the power spectrum of the reference wavelet rather than the amplitude spectrum. So, for data with limited band width the phase subtraction is preferable as a cross-correlation produces larger side lobes due to greater distortion of the frequency spectra.

Chapter 17

Migration results: Alaska

17.1 Initial results

In chapter 16 a migration scheme, which retrieves scattering potential from observations of scattered S-waves (equation 16.14), was developed. Here and in the following two chapters the migration is applied to the dataset of phase stripped traces. In this chapter a region of lower mantle beneath Alaska and western Canada is considered. This area showed one of the most prominent reflections in our global diffraction stack migration of deconvolved traces (chapter 14), at about 2590–2640km depth.

The damped version of the GRT migration equation (16.14) was used with the 1075 phase stripped traces to generate a scattering potential model for the lower mantle. In addition to the damping of the source radiation pattern term, weights in the migration integral are limited to a factor of 3 to stop the stack being dominated by just a few records. The migration integral was performed for approximately 1.35 million points, spaced at radial and tangential increments of 50km between 2090 and 2940km depth. The raw results contained large amounts of short wavelength noise and are subject to spatial aliasing. Accordingly neighbouring points were averaged using Gaussian weights, which corresponded to a normalised symmetric distribution with a standard deviation of 50km.

Unlike the amplitudes of the diffraction stack migration the amplitudes retrieved in the GRT migration can be related to elastic properties. Formally the migration retrieves scattering potential. However, given the approximation in relating the scatterer radiation patterns from density heterogeneity to seismic velocity heterogeneity, the amplitudes can be considered as a measure of the jump in isotropic seismic velocity compared to the background model (PREM), $\delta\beta/\beta$.

Figure 17.1 shows depth slices (maps) through the scattering model at 8 different depths. While figure 17.2 shows a selection for cross-sections through the same region. The slices for shallower depths (2490 to 2590km) show little coherent structure. The positive anomalies beneath Alaska and Northern Canada appear to focus into a positive anomaly around 2640km depth, in the same position as the feature seen in the backprojection results. However, here the anomaly appears to extend to 2690km depth. As in the previous backprojection the anomaly appears to be of limited spatial extent.

Beneath the 2640km positive reflector the migration gives a negative anomaly, of roughly the same spatial extent at 2740 to 2790km depth. The cross-sections (figure 17.2) show this negative layer to be sandwiched between two positive layers.

Near the base of the model, (2840km depth) a group of positive scatterers is seen in the centre of the study region, and beneath the Northwest Pacific. Due to the scatterers depth (50-100km above the CMB) it might be expected that this is a reflection from the core-mantle boundary. This is not the case as the *ScS* reflection was removed from the traces prior to the migration, and if by some error a few *ScS* phases escaped the phase stripping process we would expect them to produce much larger amplitudes than the small values seen in figure 17.1 and 17.2.

17.2 Synthetic tests

17.2.1 Simple tests

The results of the migration could be biased by the sparse data coverage. This is tested by migrating synthetic datasets made using the forward model for the scattered wavefield developed

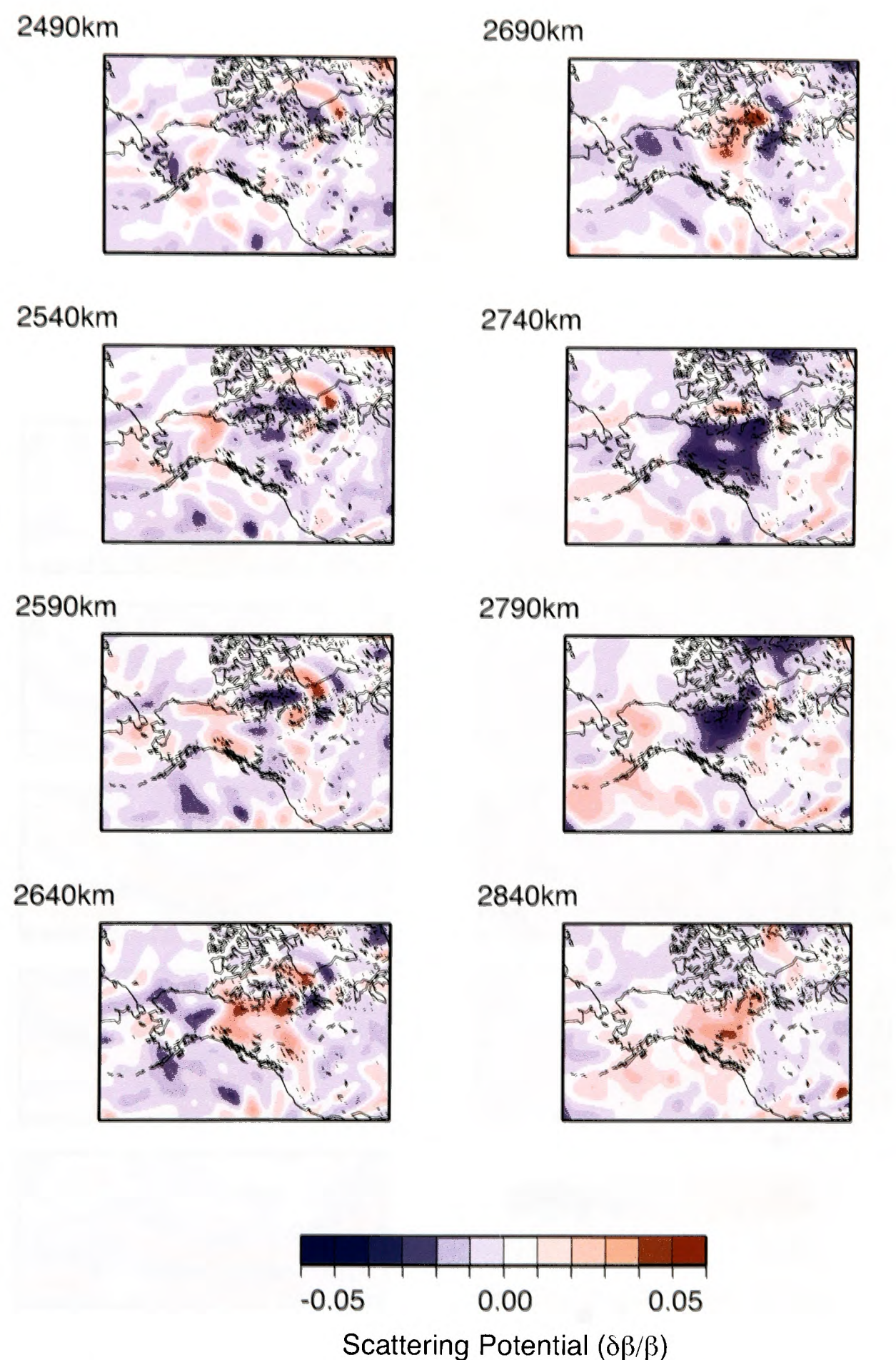


Figure 17.1: Depth slices through the migration results beneath Alaska. Colour scale is saturated above/below ± 0.05

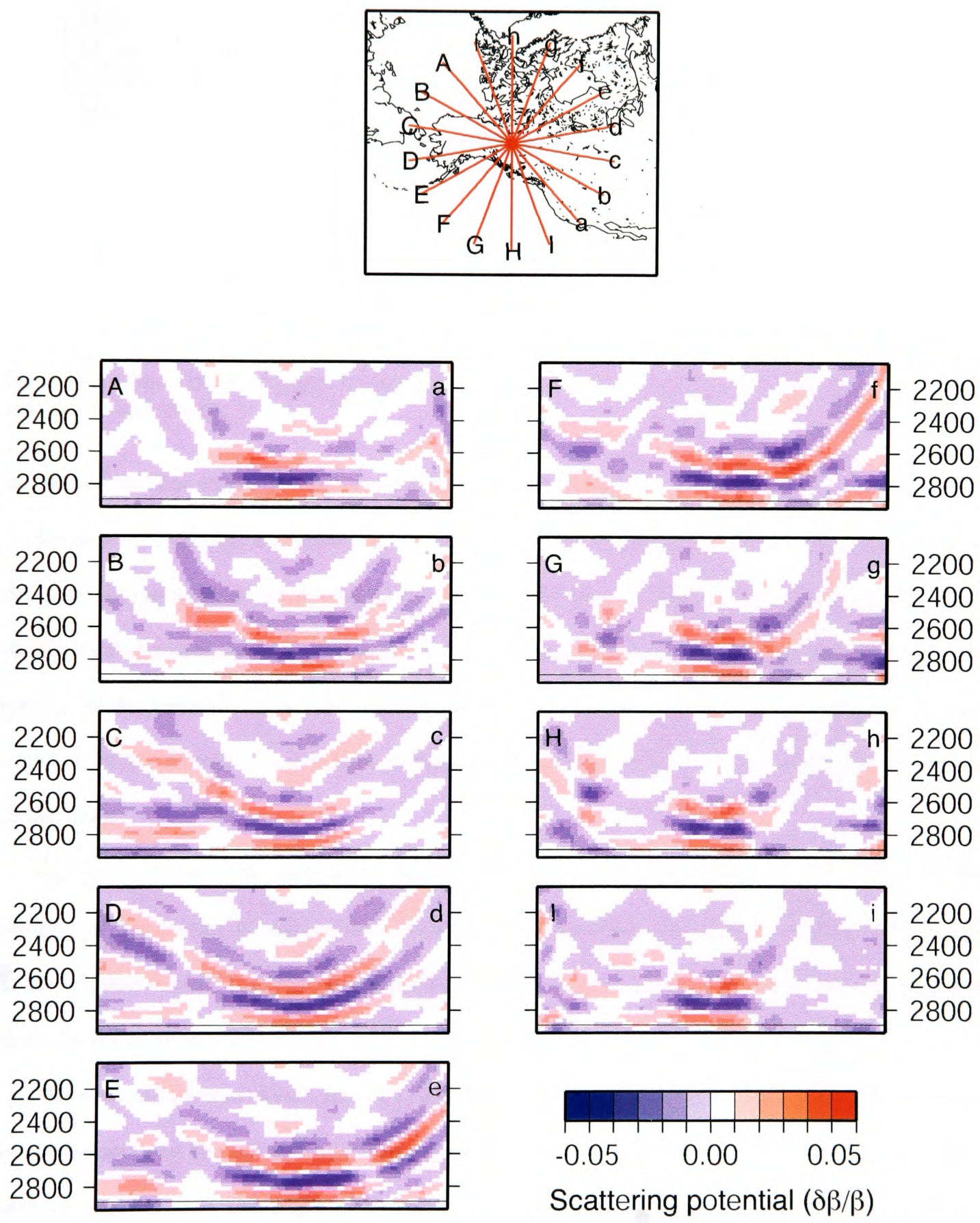


Figure 17.2: Cross-sections through the GRT migration/inversion results. Colour scale is saturated above/below ± 0.05

Test	Body position	Scattering potential ($\delta\beta/\beta$)	Figure No.
Model 1	Single point 220/62 2640km	0.05	17.3
Model 2	Disc radius=250km 235/60 2640km	0.05	17.4 17.5
Model 3	Disc radius=500km 235/60 2640km	0.035	17.6 17.7
Model 4	Disc radius=1000km 235/60 2640km	0.035	17.8 17.9
Model 5	Disc radius=2250km 235/60 2640km topo. 200km, 1500km wavelength	0.035	17.10
Model 6	Disc radius=250km 235/60 2640km Disc radius=250km 235/60 2690km	0.035 0.035	17.11
Model 7	Disc radius=350km 235/60 2740km	-0.04	17.12
Model 8	Disc radius=500km 235/60 2640km Disc radius=500km 195/52 2790km Disc radius=350km 235/60 2840km	0.03 0.02 0.02	17.13

Table 17.1: Description of input bodies for the synthetic tests, co-ordinates for discs refer to the disc centre and depth.

in chapter 15. The range of input models is described in table 17.1.

Given a distribution, and strength, for some scatterers synthetic datasets were generated using the same source-receiver geometries as the data. In the cases of bodies made up of multiple point scatterers the scatterers are positioned at equally spaced points. The effect of the amplitude spectra of the reference pulse (or source time-function) is also included in the synthetics as per equation 16.21. Finally the synthetic dataset is then subjected to the same migration procedure as the data.

Figure 17.3 shows the migration results for input model 1 (a single point scatterer). The position of the scatter is well resolved, however, the migration also creates artifacts not present in the original model. It is also worth noting that the retrieved amplitude of ~ 0.01 is much lower than the input amplitude of 0.05.

The next model considered, model 2, is a small disc of 250km radius at 2640km depth. Maps and cross-sections though the migration results are shown in figures 17.4 and 17.5. The cross-sections suggest that the edges of the body are well resolved in most directions but in other directions the body is smeared out along travel time surfaces. The map at 2690km also shows positive contamination from above, which is a similar shape to that seen in the data. None of the cross-sections show an absence of the reflector where it should be.

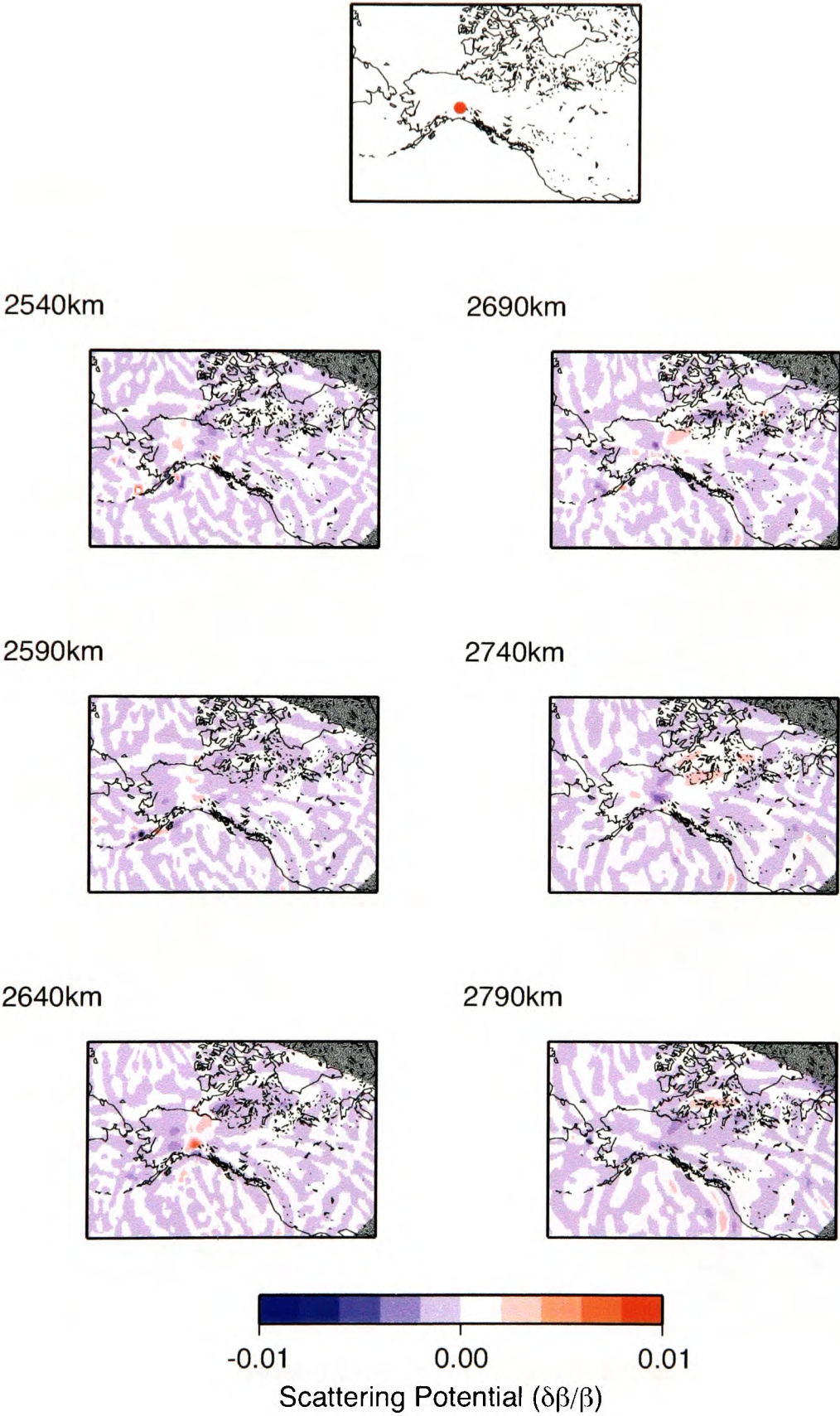


Figure 17.3: Results of GRT migration applied to Born approximate synthetics for a single scatterer. Scatterer position is shown in the top map and lies at 2640km depth, and has a strength of 0.05. In this case no horizontal smoothing was applied to the migration results.

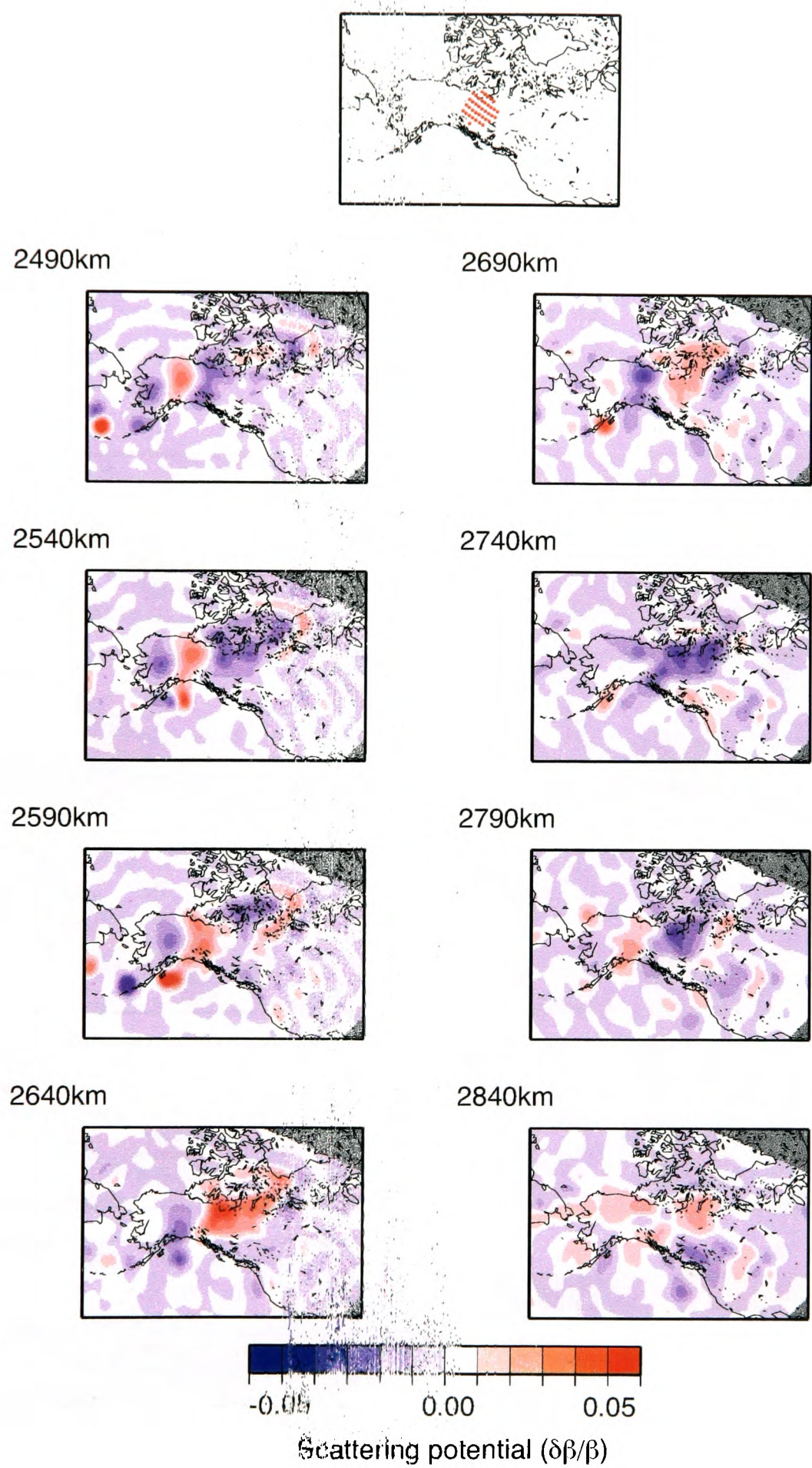


Figure 17.4: Results of migration applied to Born approximate synthetics generated a disc of radius 250km at 2640km depth (model 2). Scatterer position is shown in the top map and each scatterer has a strength of 0.05.

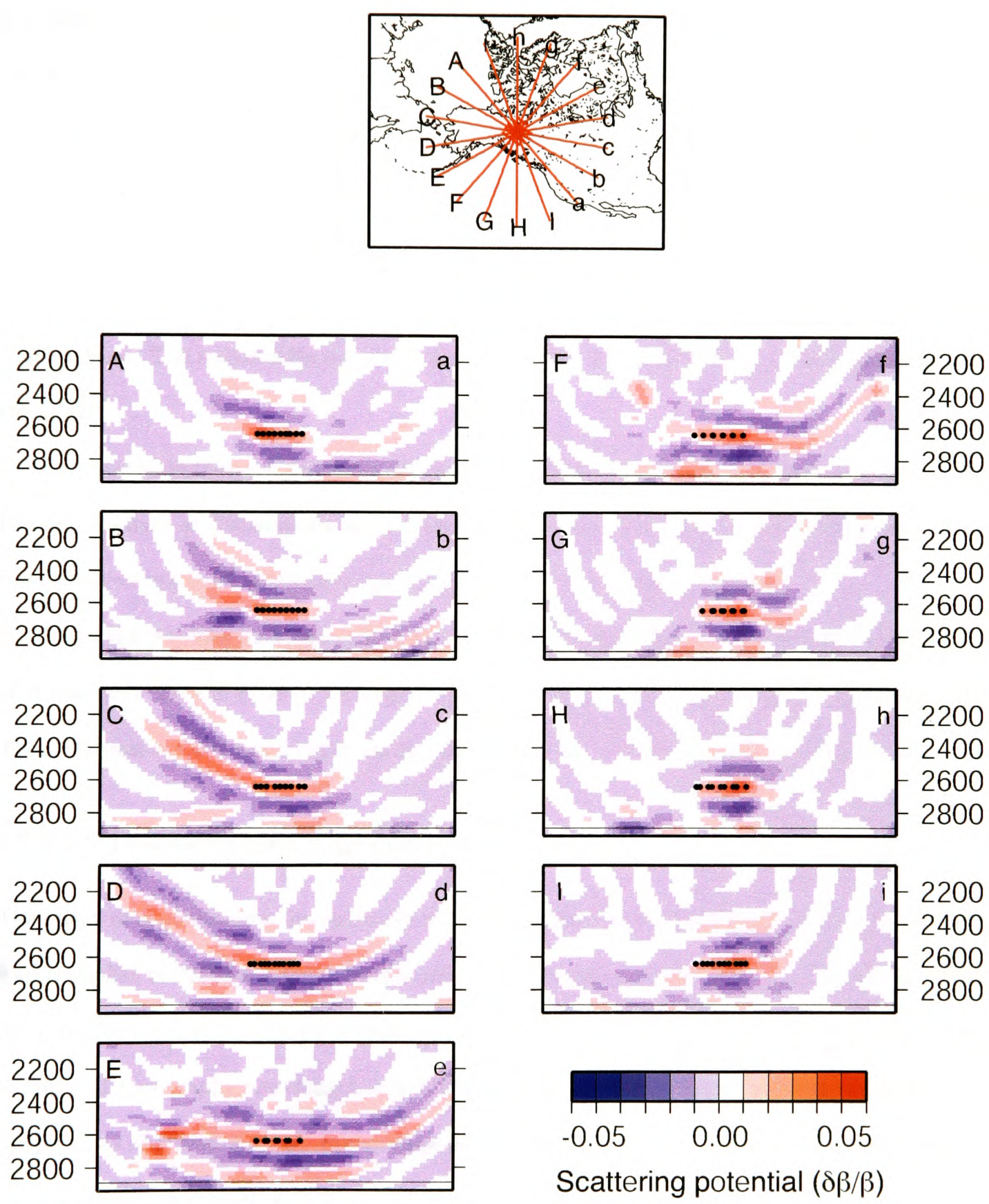


Figure 17.5: Cross-sections for synthetic results generated using model 2. The position of point scatterers in the input model are indicated by the black dots.

The cross-sections of figure 17.5 also show the creation of a positive just above the CMB, similar to that seen in the data. However, the map views (figure 17.4) show that this anomaly is weaker than the comparable anomaly seen in the data. It is also limited to the region directly beneath the scatterers, where as in the data this anomaly appears to extend into the North Pacific.

In the third synthetic experiment, model 3, a slightly larger disc of scatterers (500km radius) at the same depth (2640km) is considered. The results (figures 17.6 and 17.7) show many similarities to those obtained using a small disc. For example, the maps at 2540km depth show positive anomalies beneath eastern Alaska and Western Hudson Bay, created by smearing along travel time surfaces. The major difference from the smaller disc is the lack of any positive anomaly near the CMB.

The fourth synthetic experiment (model 4, figures 17.8 and 17.9) consists of a larger disc (1000km radius) at 2640km depth. It can be seen that the edges of the anomaly are poorly resolved in virtually all directions, and no artifacts are placed near the CMB.

Unlike the previous tests, using model 4, small positive amplitudes are retrieved beneath much of Alaska and there is little or no negative anomaly beneath western Canada at 2740km depth. These features are recreated better by the more localised distributions of scatterers.

17.2.2 Discontinuity topography

The effect of a continuous discontinuity with topography is examined using model 5. Model 5 features a disc of 2250km radius. Discontinuity topography is generated within the disc by setting the scatterer depth (km) using $d = 2640 + 100 \cos(\frac{2\pi\Delta}{\lambda})$, where Δ is the distance to the disc centre, in km, and topography wavelength λ is 1500km. This amount of topography is consistent with the previous observation of D'' topography (for example Thomas *et al.* (2004b)). Each scatterer has a scattering potential of 0.035.

Cross-sections through the migration results are shown in figure 17.10. Unlike the previous models a single continuous horizon is not seen. Instead, a series of broken up anomalies is retrieved with many of the largest anomalies at the topographic turning points. However, much of the

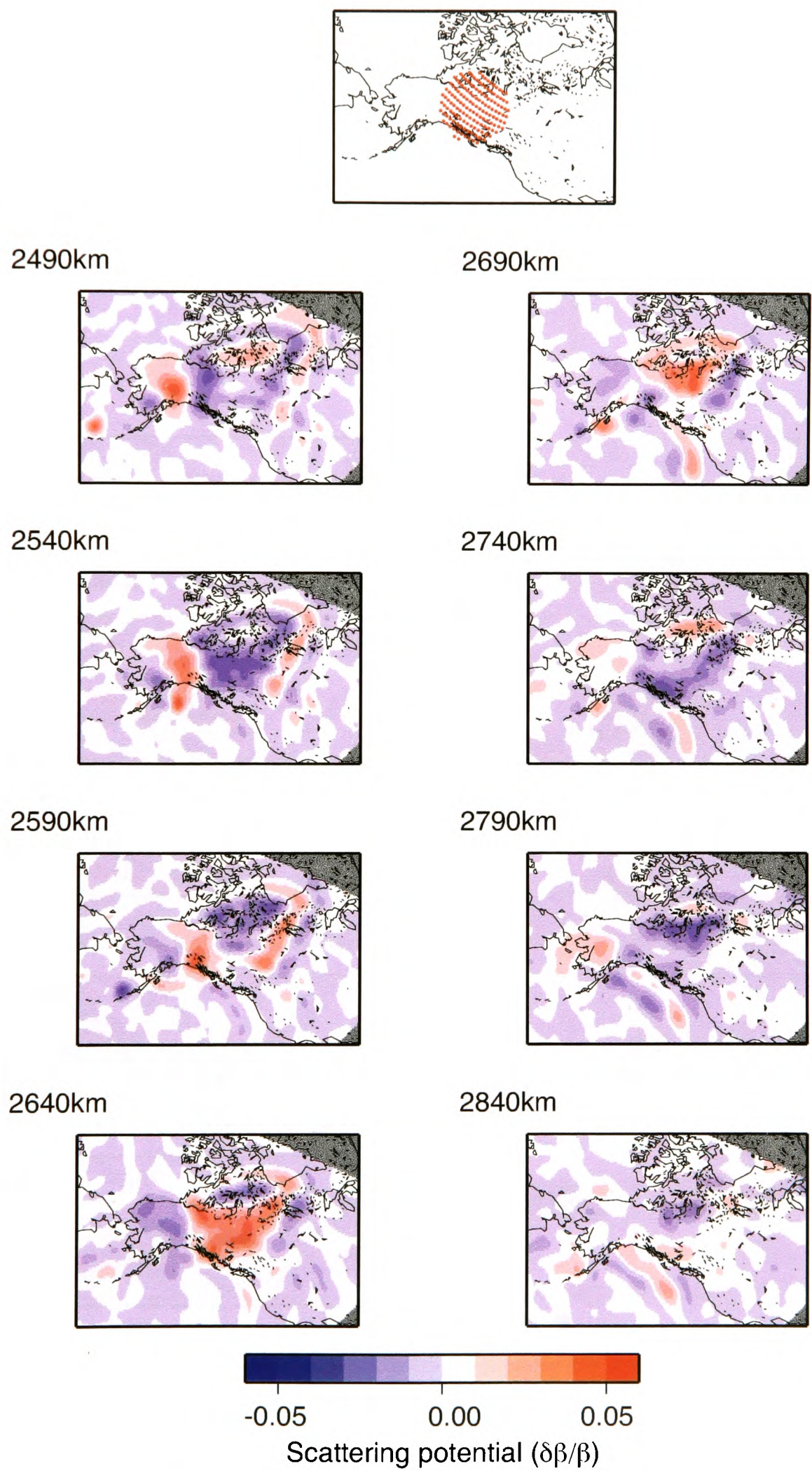


Figure 17.6: Results of migration applied to Born approximate synthetics generated a disc of radius 500km at 2640km depth (model 3). Scatterer position is shown in the top map an each scatterer has a strength of 0.035.

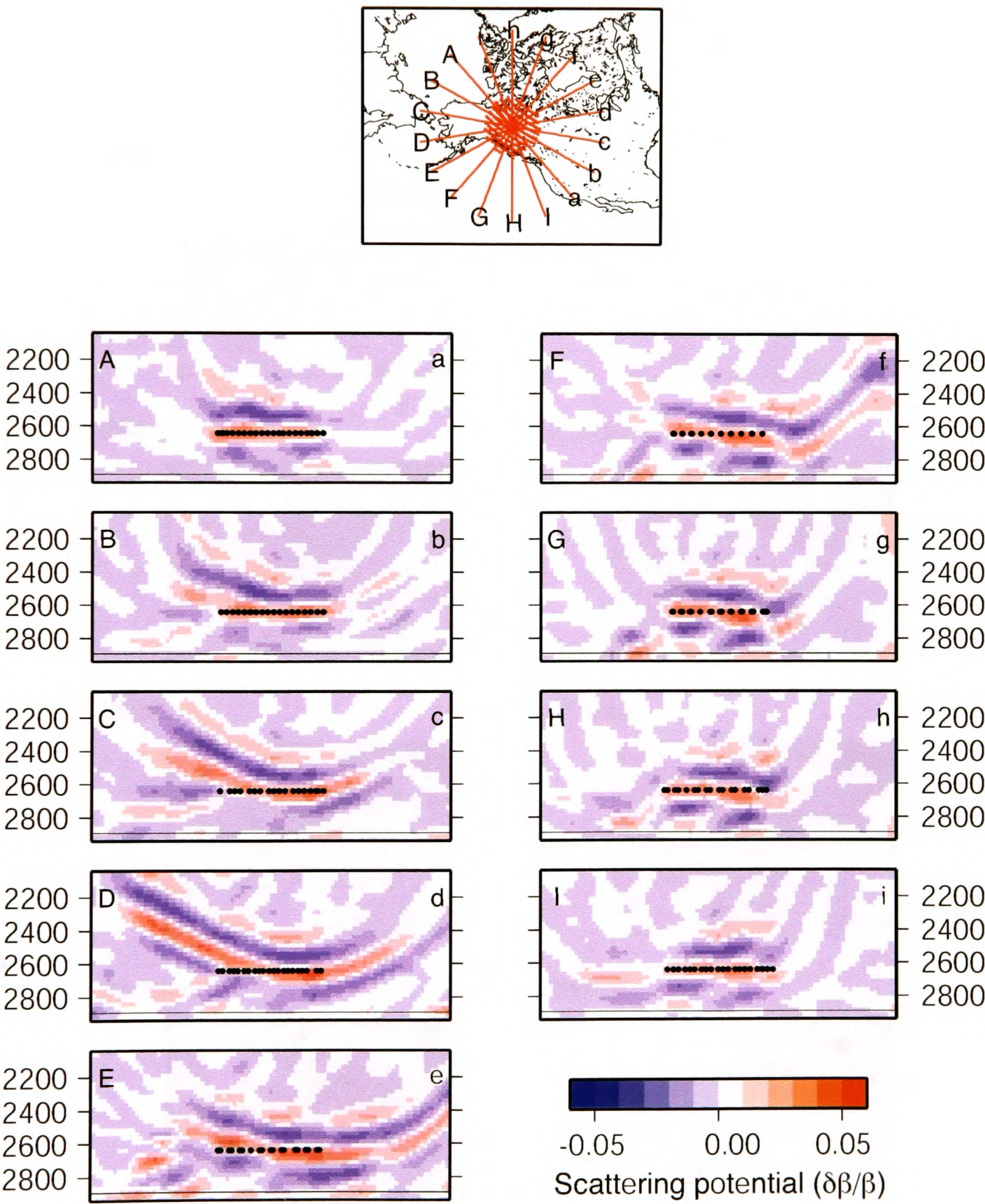


Figure 17.7: Cross-sections for synthetic results generated using model 3.

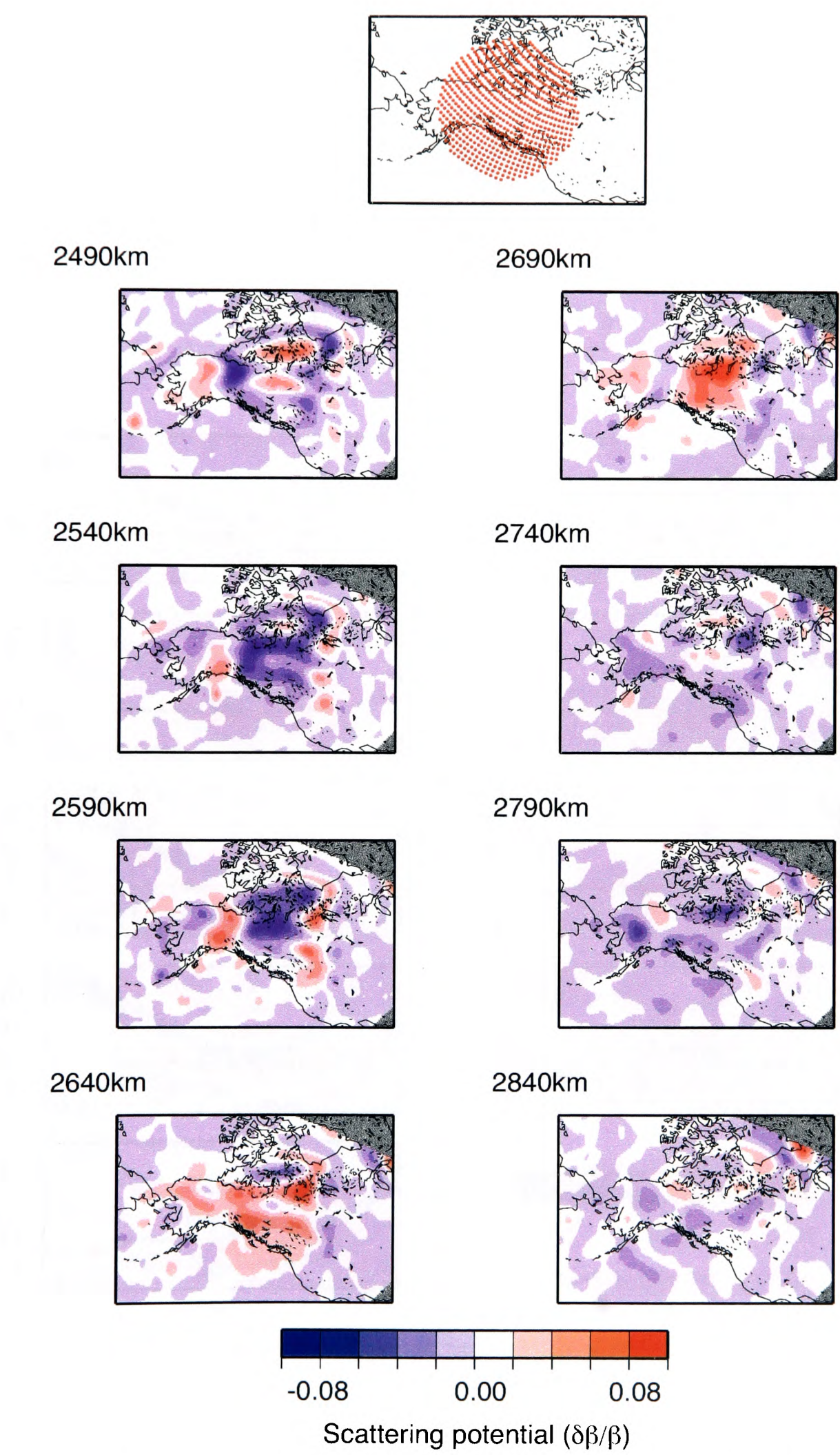


Figure 17.8: Results of migration applied to Born approximate synthetics generated a disc of radius 1000km at 2640km depth (model 4). Scatterer position is shown in the top map an each scatterer has a strength of 0.035.

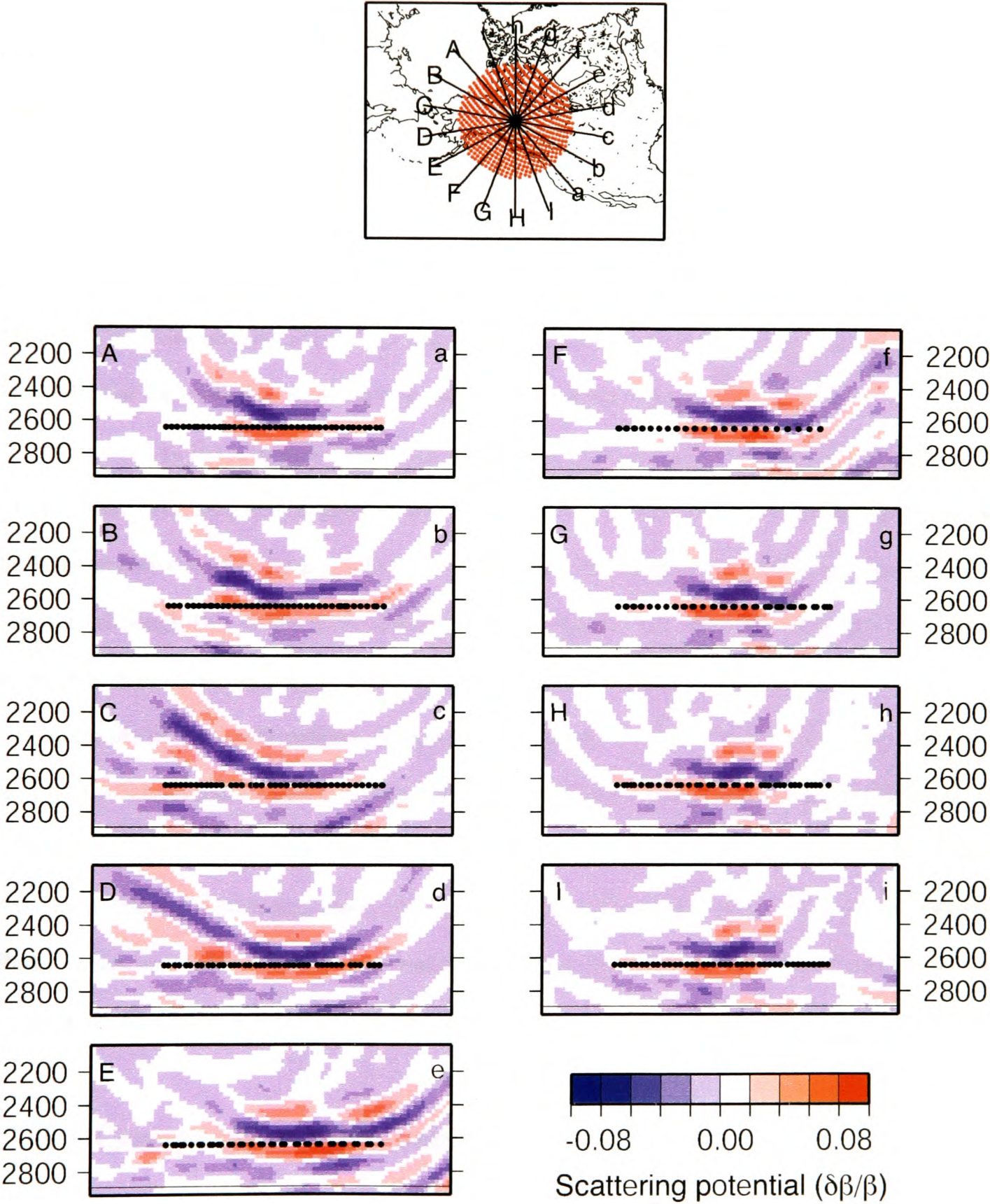


Figure 17.9: Cross-sections for synthetic results generated using model 4.

reflector is retrieved beneath Alaska (section C-c), although with some added artifacts.

17.2.3 Other models

In the next synthetic test (figures 17.11), two (250km radii) discs of scatterers at 2640 and 2690km depth (model 6) are used. The retrieved scattering potentials are only slightly less than those for a single layer, yet input scattering potential is half that used for model 2. The top layer (2640km depth) is retrieved with lower amplitudes than the bottom. This is the opposite to what is seen in the data where the central anomaly at 2640km depth is of greater spatial extent than that at 2690km. The cross-sections show that, like model 2, model 5 produces positive amplitudes near the CMB, but like model 2, they follow a different pattern to those seen in the data.

In figures 17.12 the migration results for input model 7 are shown. The input model consists of a deeper layer (2790km) with negative scattering potential. This experiment successfully replicates the large negative scattering potentials, as well as positive amplitudes at 2640km and 2840km. However, although this model gives similar anomaly outlines, the positive amplitudes at 2640km are weaker than those near the CMB. This is the opposite of what is seen in the data.

The results of the final synthetic example (model 8) are shown in figures 17.13 and 17.14. In this case the input model consists of 3 bodies at depths of 2640, 2790, and 2840km (see figure 17.13 for details). It can be seen that this model is closest to the data, with most of the major features of the data replicated including much of the structure near the CMB. Although structure beneath Alaska is not simulated exactly. For example the migration places a negative beneath eastern Alaska at 2840km depth, which is not seen in the data.

17.3 Comparison with seismic tomography

In figure 17.15 the migration results are compared with the bottom 800km of seismic tomography model S20RTS (Ritsema *et al.* (1999)). Figure 17.16 is the same but uses SKS12WM13 (Liu & Dziewonski (1994)). Both the tomography and the migration results presented here are sensitive to variations in the same elastic parameters (shear modulus and density in isotropic media), from

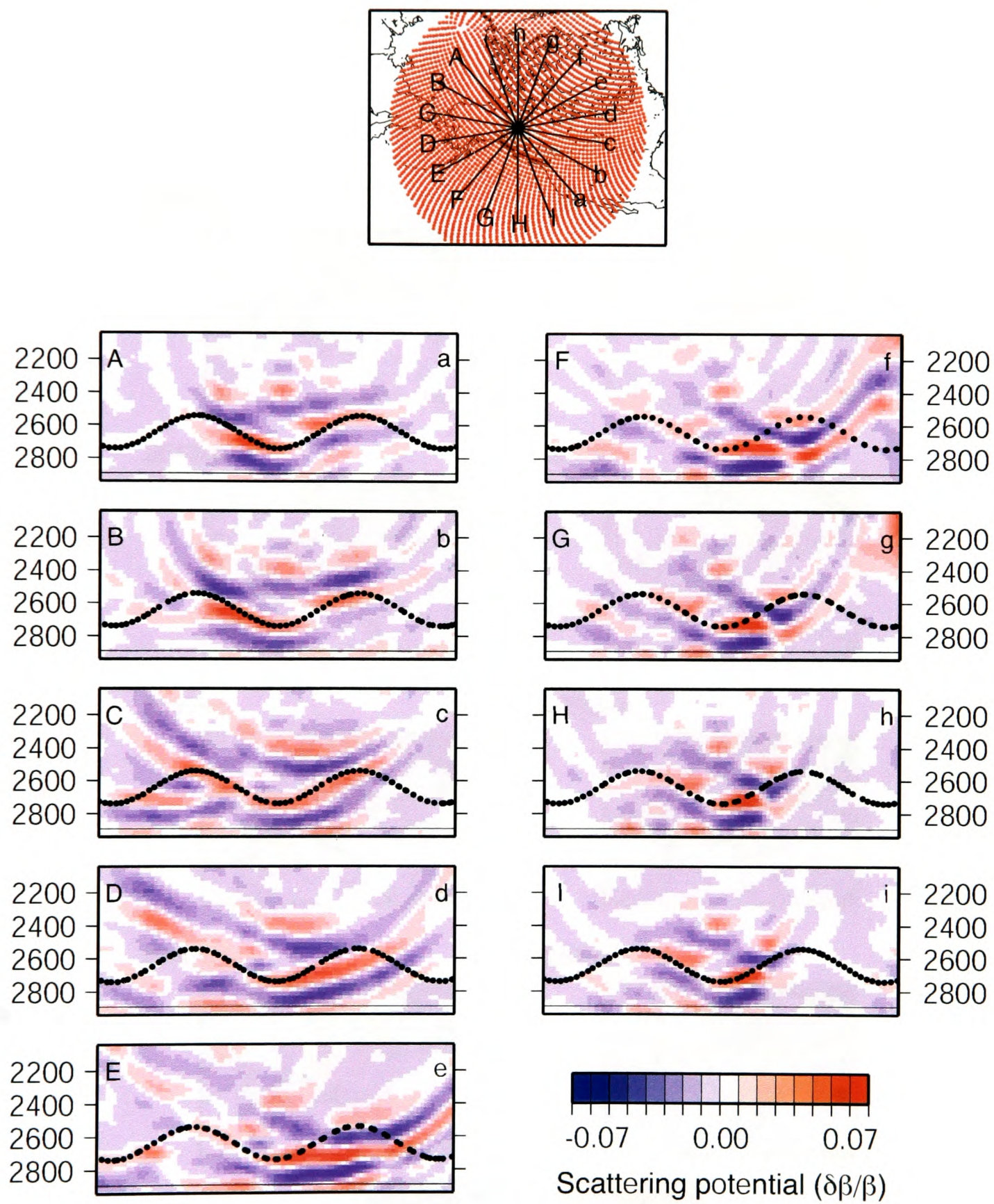


Figure 17.10: Cross-sections for synthetic results generated using model 5. A disc of 2250km radius. Scatterer depth is given by $2640 + 100 \cos(\frac{2\pi\Delta}{\lambda})$, where Δ is the distance to the disc centre, and topography wavelength λ is 1500km. Black dots shown the scatterer positions projected from 100km either side of the cross-section. Each scatterer has a scattering potential of 0.035. The colour scale is saturated at ± 0.07 .

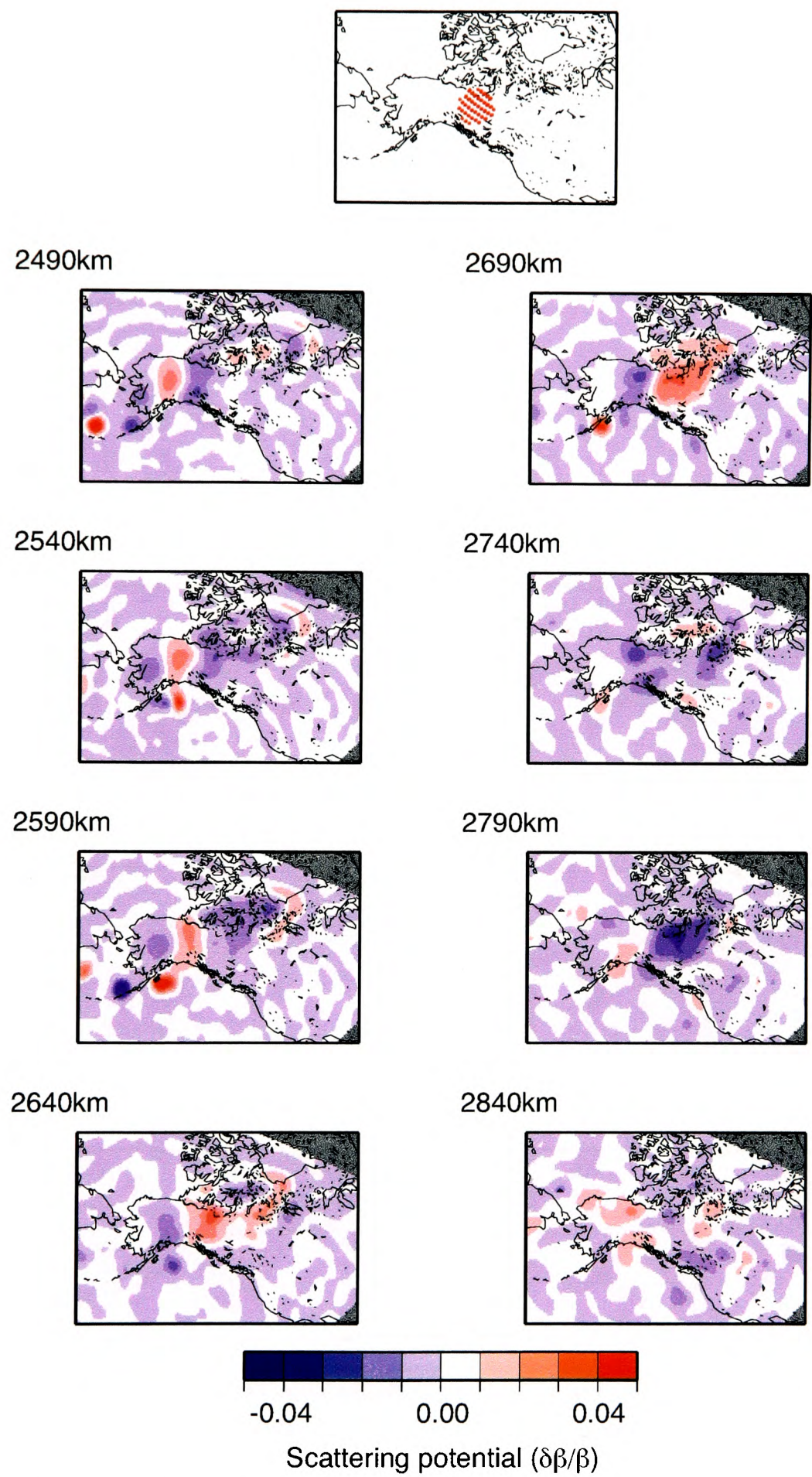


Figure 17.11: Results of migration applied to Born approximate synthetics generated a 2 discs of radius 250km at 2640 and 2690km depth (model 5). Scatterer position is shown in the top map an each scatterer has a strength of 0.035. Colour scale saturated at ± 0.04 .

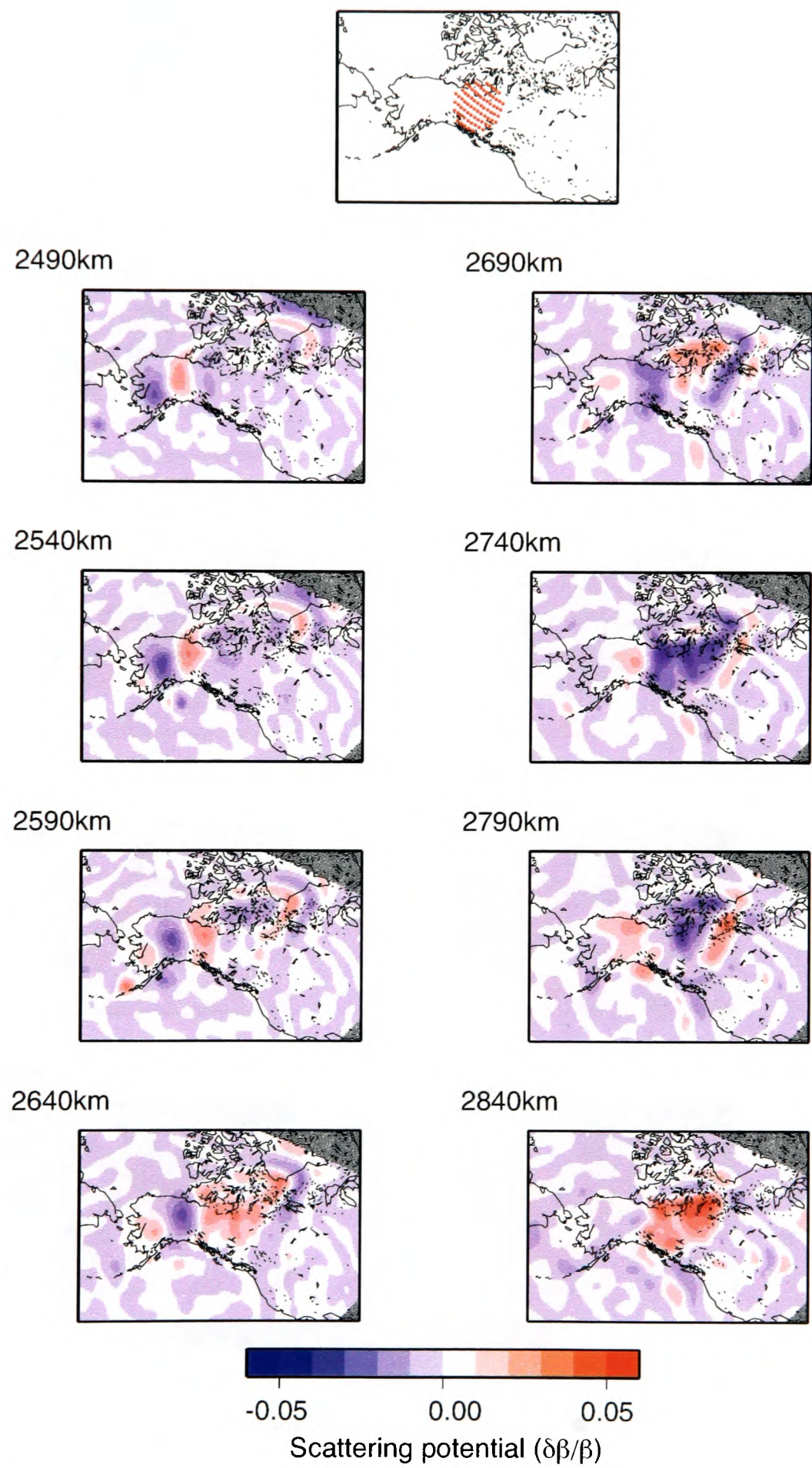


Figure 17.12: Results of migration applied to Born approximate synthetics generated a disc of radius 350km at 2740km depth (model 7). Each point scatter has a scattering potential of -0.04.

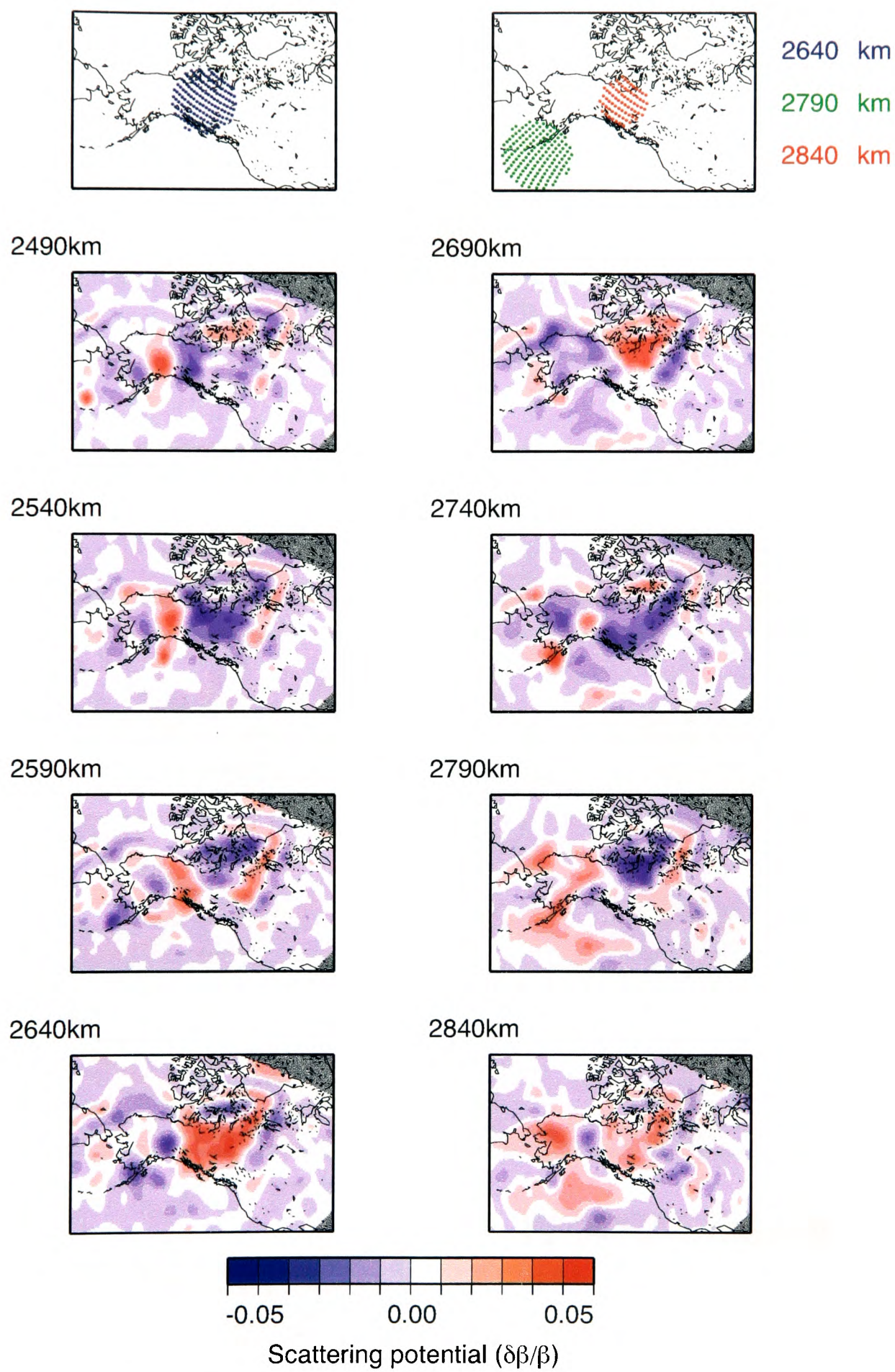


Figure 17.13: Results of GRT migration applied to Born approximate synthetics for model 8. Input model consists of 3 separate reflectors at 2640, 2790, and 2840km depth. The body at 2640km has a scattering potential of 0.03 and a radius of 500km. While the other bodies at 2790 and 2840 have a scattering potentials of 0.02 and radii of 500 and 350km respectively.

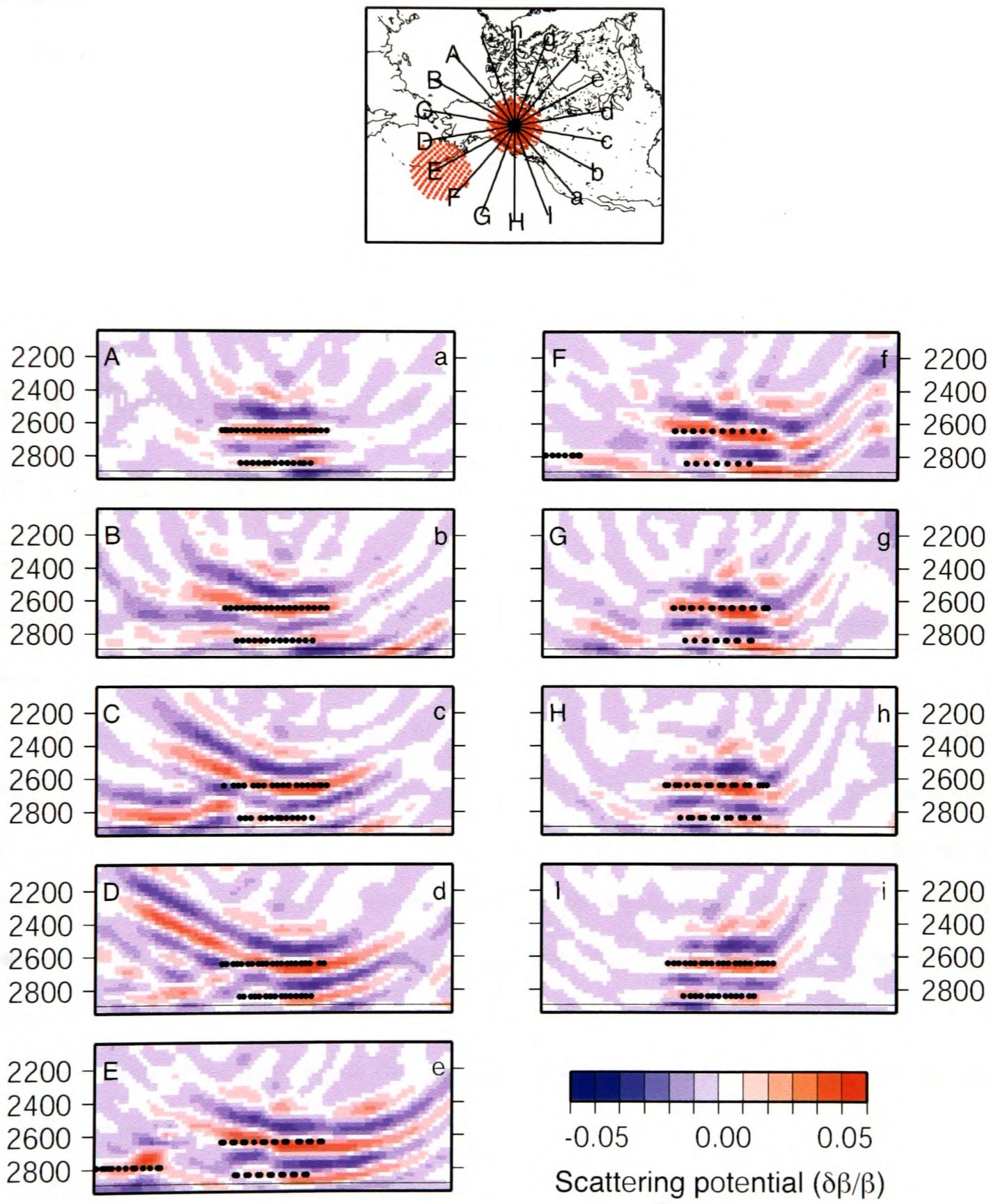


Figure 17.14: Cross-sections through migration results for model 8.

the background velocity model PREM. However, seismic tomography averages variations along ray paths and so will smooth out abrupt changes in velocity, whereas the migration results are sensitive only to abrupt changes in elastic properties, and will not image velocity gradients.

Despite the differences in imaging technique, figure 17.15 shows that S20RTS has several features in common with the migration results in the Alaska region. In the centre of the cross-sections, the positive reflector at 2640km roughly coincides with an area of steeper gradients in seismic velocity. The near CMB reflector also agrees with a positive velocity anomaly in the centre of the sections.

17.4 Discussion and Conclusions

17.4.1 The D'' reflector

Although in general the synthetic tests of the migration show that the precise scatterer geometries is poorly constrained, the synthetic examples do show that the majority of the features from the data can be produced by a few relatively simple structures.

The results require the presence of a body with positive scattering potential at around 2640km depth (~ 250 km above the CMB) near the centre of the study region (235° longitude, 60° latitude). This reflector is consistent with a D'' discontinuity. A deeper negative reflector can be made to give similar anomaly shapes, however, it produces much larger amplitudes near the CMB than were observed. Furthermore the presence of a negative reflector is not easily reconciled with the presence of positive velocity anomalies in S20RTS (figure 17.15).

In the centre of the study region the body appears well resolved, however, tests with different size discs show that the exact dimensions of the body cannot be resolved. The images produced by a disc of 250km radius are very similar to those for a disc of 500km radius. However, those produced by a disc of 1000km radius do not produce the negative migration amplitudes seen at 2640km under much of Alaska and the North Pacific. These are more successfully replicated using the smaller scattering bodies (i.e. the 250 and 500km radii discs).

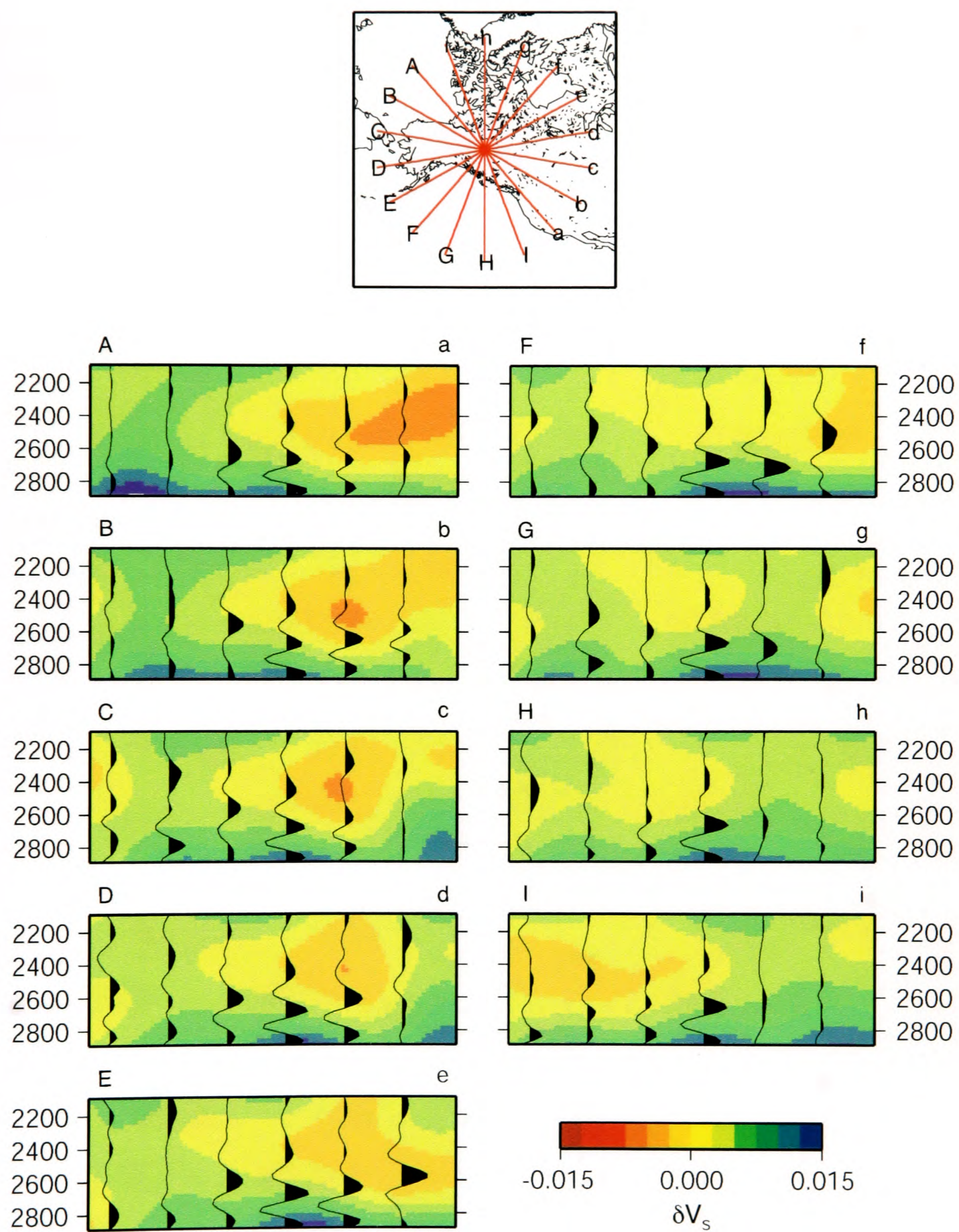


Figure 17.15: Comparison of GRT results for Alaska region with seismic tomography model S20RTS (Ritsema *et al.* (1999)). Colour backgrounds show S-wave velocity perturbation for bottom 800km of the mantle, the superimposed traces are depth profiles through the GRT migration results plotted at their location along the section profile.

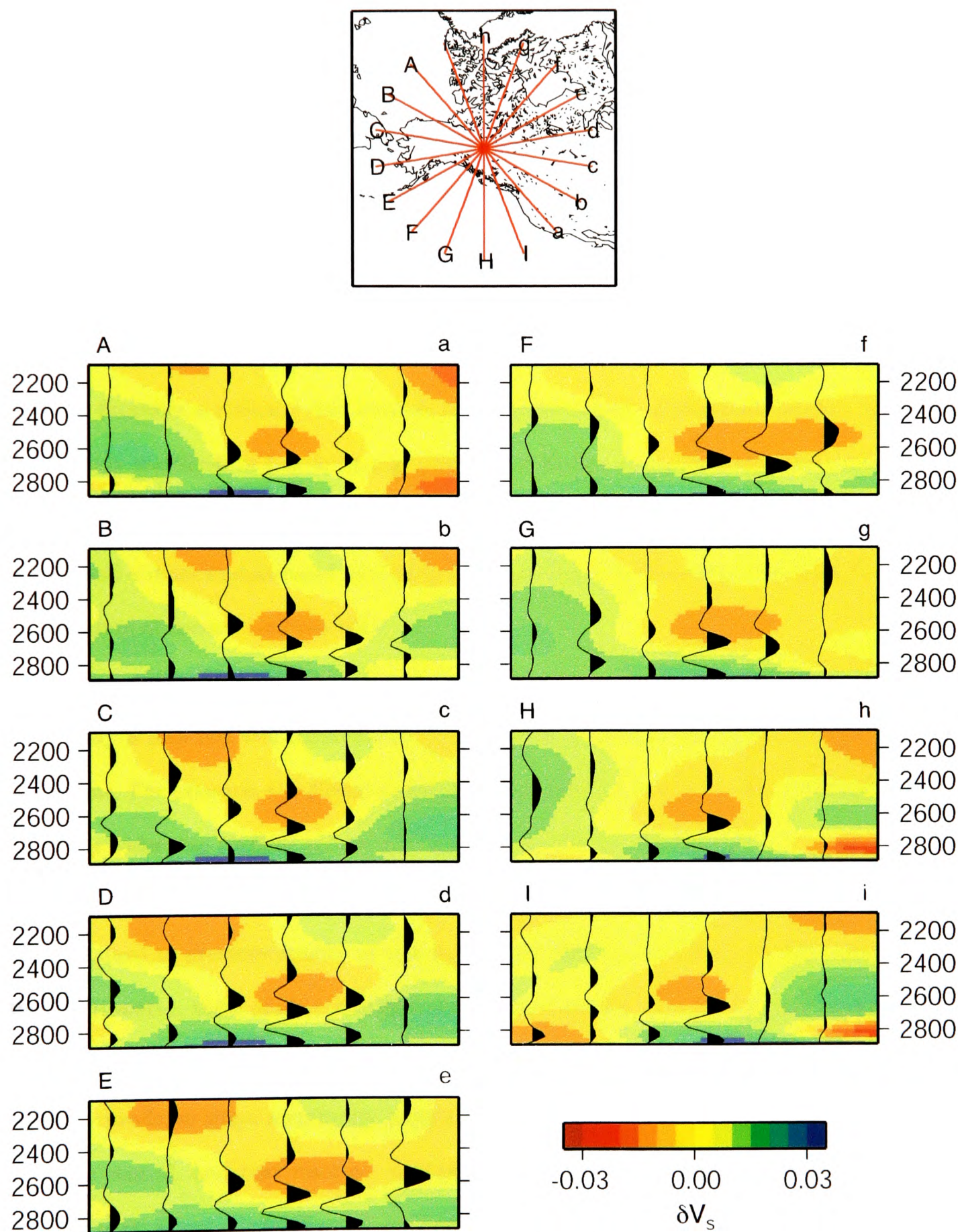


Figure 17.16: Comparison of GRT results for Alaska region with seismic tomography model SKS12WM13 (Liu & Dziewonski (1994)). Colour backgrounds show S-wave velocity perturbation for bottom 800km of the mantle, the superimposed traces are depth profiles through the GRT migration results plotted at their location along the section profile.

The test for discontinuity topography shows that a continuous reflector with significant topography would be observed in the centre of the study region as broken up series of near horizontal bodies. This is not seen in the data, and hence the results cannot be explained using a continuous reflector with large amounts of discontinuity topography in this region. Instead the anomaly is best explained by a near horizontal disc at around 2640km depth with radius $\sim 250\text{-}500\text{km}$ near the Alaska/Canada Border.

This contrasts with the findings of Lay & Young (1996), who also identified a flat scattering surface near 2600km depth beneath Alaska that has dimensions exceeding 1500km. Their interpretation proposes a reflector that extends right across Alaska. Differences between the results of this study and those of Lay & Young (1996) could be due to the different data and modelling approaches used. The interpretation here should be preferred, because of the use of a larger dataset, and a more sophisticated modelling technique based on the entire waveform, rather than the migration of deconvolved records.

The uncertainty in the dimensions of the scattering body means that the scattering potential of the body is also not well constrained, as there exists a trade-off between scattering potential and volume. For example, the 250km radius disc with velocity contrast of 5%, produces a similar result to a wider disc (500km radius) with a lower velocity contrast (3.5%).

The scattering potentials, and the corresponding velocity anomalies, are slightly larger than those obtained in previous studies of this region ($\sim 2.75\%$, Lay & Young (1989); Young & Lay (1990)). However, these studies modelled the reflector using a 1-dimensional model (infinite reflector), so given the trade off between scatterer volume and amplitude it is not unexpected that this study produces larger amplitudes using a smaller body.

In the central, and best resolved, part of the study area the observation of a D'' reflector coincides with a region of enhanced gradient in the seismic tomography models S20RTS and SKS12WM13 (Ritsema *et al.* (1999); Liu & Dziewonski (1994)). Given that the migration procedure used here is not sensitive to velocity gradients, it seems likely that the region contains a first order discontinuity in seismic velocity, which has been smoothed out in the tomography models.

However, the discrepancy between the two seismic tomography models used here means that this should be treated with caution.

Kendall & Shearer (1994) also examined the region beneath Alaska, finding reflections from an average depth of 2590km. Although this figure is biased by some high values observed under the Aleutian chain. Their values ranged between 2730 and 2520km with deeper values corresponding to the region where the reflector is observed in this study. Kendall & Shearer (1994) also observed apparent D'' reflections beneath Western Alaska and the Aleutians, a region where no D'' discontinuity is seen in our results. However, their modelling approach of mapping $S - SdS$ time to depth under the source-receiver midpoint, assumes a horizontal reflector and an absence of other structure. Accordingly the presence of complicated scattering bodies, could lead to errors in the derived depths and locations using this method.

17.4.2 The near CMB reflector

Artifacts consistent with a deeper reflector at 2840km (50-100km above the CMB), as seen in the data, can be reproduced using a shallower body at 2640km (e.g. model 1). However, this anomaly is much stronger in the data, and is not limited to the region beneath the Alaska/Canada border but extends into the North Pacific. Alternatively, this scattering body could be due to multiples from the CMB and D'' which would be misplaced by the migration. However, multiples from the CMB would arrive within the tail of the ScS phase which was removed in the phase stripping procedure. Accordingly, the results require a second scattering body within the D'' to satisfy these arrivals, although the testing here has not been exhaustive.

This scattering body was not seen using the backprojection of deconvolved records (section 14.2). This is probably because the scattered phases from this body are of small amplitude, and so were not picked out in the iterative deconvolution process. Even in cases where the phase was picked out by the deconvolution it may have been masked out by the larger reflections from the overlying D'' reflector.

The near CMB reflector appears to coincide with seismic velocity anomalies in both S20RTS

and SKS12WM13. It also appears to exist over wider regions than the D'' reflector. In the final synthetic test two separate scattering bodies at 2790, and 2840km were used, which when the migration was applied were separated by a negative region. This negative region is not observed in the data, so it seems likely that this near CMB reflector stretches from the region beneath Alaska into the North Pacific.

None of the previous studies of the D'' region beneath Alaska region show a reflector within the D'' (Kendall & Shearer (1994); Lay & Young (1996)). In fact Castle & van der Hilst (2003), suggested that the region beneath the Gulf of Alaska contains no sharp impedance contrasts between 1800km depth and the CMB. However this was based on stacks for four sources, which assume a planar wavefront and a horizontal reflector. The study of Castle & van der Hilst (2003) also used higher frequency data (2-5 seconds) than that used here, so a diffuse scattering boundary may also explain the discrepancy.

Positive scattering potentials have been observed in other parts of the D'' region. Bilek & Lay (1998) identified a group of scatterers at depths of 2750 to 2770km beneath North Central Asia. While the region identified beneath the North Atlantic by Braña & Helffrich (2004) at 2720km depth, was found to contain small scale heterogeneities (~ 700 km is size) with S-wave velocity contrasts of 1.4-6.6%. This range is consistent with the scattering potentials found here. Braña & Helffrich (2004) suggested that the scattering structures they observed were chemical heterogeneity based on the scatterers size, positive velocity anomaly, and association with the reconstructed position of ancient subducted material.

Chapter 18

Migration results: North Central Asia

18.1 Initial results

In figures 18.1 and 18.2 the results of the migration for the mantle beneath North Central Asia are shown. The results are different to those for the Alaska region (section 17.1). The retrieved amplitudes are much lower and the maps show a more complicated structure. The most coherent anomalies are the positive structures at around 2490-2540km depth (350-400km above CMB), and at around 2840km. The depths in between and above appear to contain little coherent structure.

The complex structure of the study region can also be seen from the cross-sections (figure 18.2). In all the sections some kind of reflector is seen in the range 2700-2400km depth. Although the behaviour of this body varies with azimuth. For example, in sections A-a and B-b the reflector appears to be continuous with an increase in depth to the North and South of more than 100km. In the 'c' direction the reflector appears almost flat. However, sections H-h, G-g, F-f suggest a collection of scattering bodies rather than a continuous reflector. To the east and northeast the reflector appears to stop near the centre of the cross-sections.

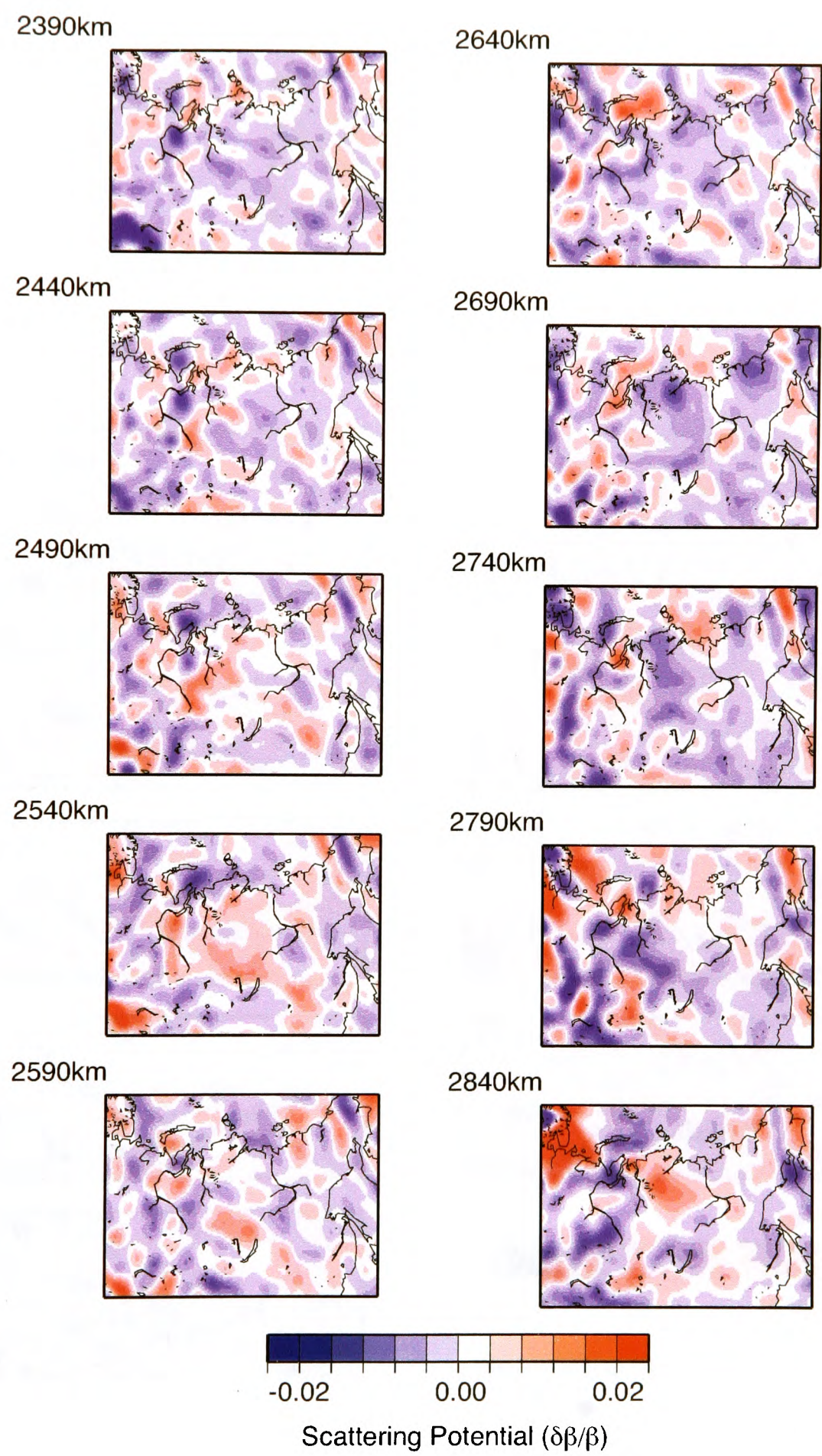


Figure 18.1: Depth slices through the migration results beneath Northern Central Asia. The colour scale is saturated at ± 0.02 .

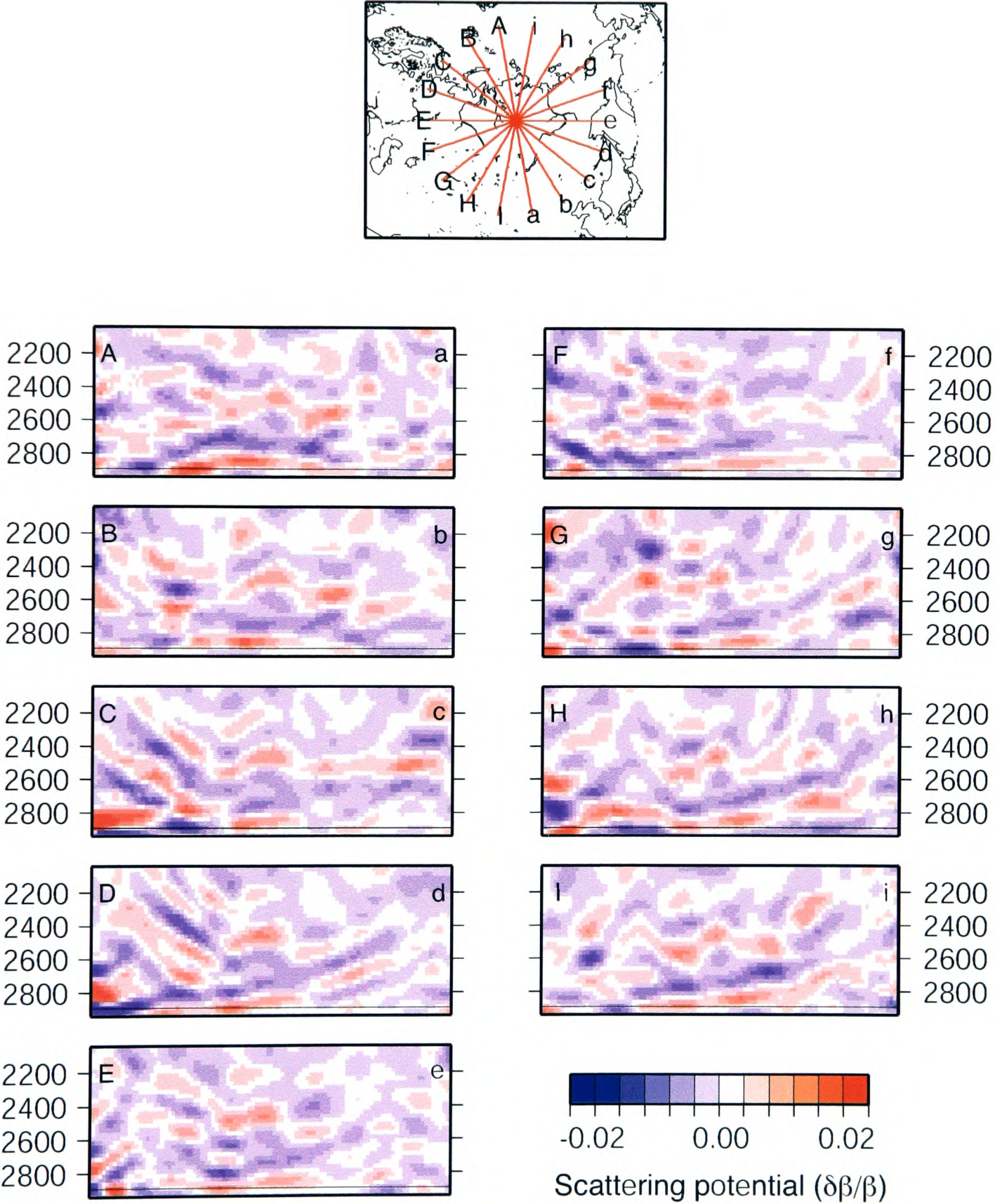


Figure 18.2: Cross-sections through the migration results for Northern Central Asia. The colour scale is saturated at ± 0.02 .

Test	Body position	Scattering potential ($\delta\beta/\beta$) (km^{-3})	Figure No.
Model 1	Disc radius=800km 100/65 2540km	+0.035	18.3 18.4
Model 2	Disc radius=1600km 100/65 2540km	+0.03	18.5
Model 3	Disc radius=1500km 100/65 2540km 200km peak to peak, 1500km wavelength topo.	+0.03	18.6
Model 4	Disc radius=400km 105/65 2840km Disc radius=500km 105/65 2840km	+0.02 -0.05	18.7
Model 5	Various discs 100-300km radius 2490-2640km Disc radius 250km 100/70 2840km	+0.04 +0.02	18.8 18.9
Model 6	Complex layer 2490-2640km depth Disc radius 250km 100/70 2840km Disc radius=250km 78/63 2840km	+0.04 +0.02 -0.05	18.10 18.11

Table 18.1: Description of input bodies for the synthetic tests, co-ordinates for discs refer to the disc centre.

In section F-f two reflectors are seen: one is the D'' reflector at around 2500km depth, the second around 2740-2790km depth. The deeper reflector is also visible -but not as distinctive- in section G-g, where it looks as though it could be connected to either the upper (D'') or structure near the CMB.

At the depths around 2790-2840km there is a complex pattern of positive and negative anomalies. The major feature is a positive reflector in the centre of the study region. This reflector is just as bright as the top D'' reflector, and appears to display some topography (section H-h). The central positive anomaly appears to be bounded by negative smaller broken up regions of both positive and negative anomalies, although at this stage it is not clear which elements of these are artifacts of other anomalies and which are real.

18.2 Synthetic tests

18.2.1 D'' structure

As with the Alaska region the influence of the source-receiver geometry on the retrieved images is tested. The test input models are described in table 18.1.

Model 1 is a simple disc of 800km radius at 2540km depth, and the migration results are shown

in figures 18.3 and 18.4. From the maps, it can be seen that a large central anomaly is retrieved, and this introduces very little structure below a depth of 2640km. However, the maps for the synthetic test show a much simpler structure than that seen in the data. This can also be seen from the cross-sections of figure 18.4, where -for the most part- the migration reconstructs a flat discontinuity. Toward the northwest ('f' 'g' 'h' and 'i' directions) the reflector is visible in the synthetic test but not in the data. It is also worth noting that some small near CMB artifacts are created but these are of very low amplitude compared to the main reflector.

Figure 18.5 shows that the results of a synthetic test using a much larger disc (radius 1600km) are virtually identical to those for the smaller disc. The only difference worth noting is the slightly larger amplitudes retrieved using this model.

Using model 3 the affect of reflector topography, on the migration results, is tested. In a similar manner to test 5 in section 17.2, a disc with scatterer depth (km) set using $d = 2540 + 100 \cos(\frac{2\pi\Delta}{\lambda})$ was used, where Δ is the distance to the disc centre, and topography wavelength λ is 1500km. A series of horizons is seen rather than a continuous interface (for example section E-e, figure 18.6). In sections A-a, B-b, C-c, D-d, and E-e the reflector is still visible along most if not all its length. Although some of this could be attributed to the anomaly being fortuitously spread out in the same direction as discontinuity topography. In the 'e' and 'f' directions the reflector is still visible to the east of centre, beyond the region where it exists in the data.

The reflector topography also introduces artifacts into the well resolved regions. For example, the central upward pointing cusp in sections E-e through G-g is imaged as a downward pointing cusp. Also, in the 'B', 'C', and 'D' directions the downward pointing cusp creates an apparent double layer or split discontinuity.

In model 4 the ability to resolve features at the base of the mantle is tested. Model 4 consists of 2 adjacent discs with positive (+0.02) and negative (-0.05) scattering potentials. From the results in figure 18.7 it can be seen that both anomalies are well resolved, and artifacts are limited to the deeper layers (below 2640km depth). However, this simple model does produce the short scale variation at 2840km depth as seen in the data

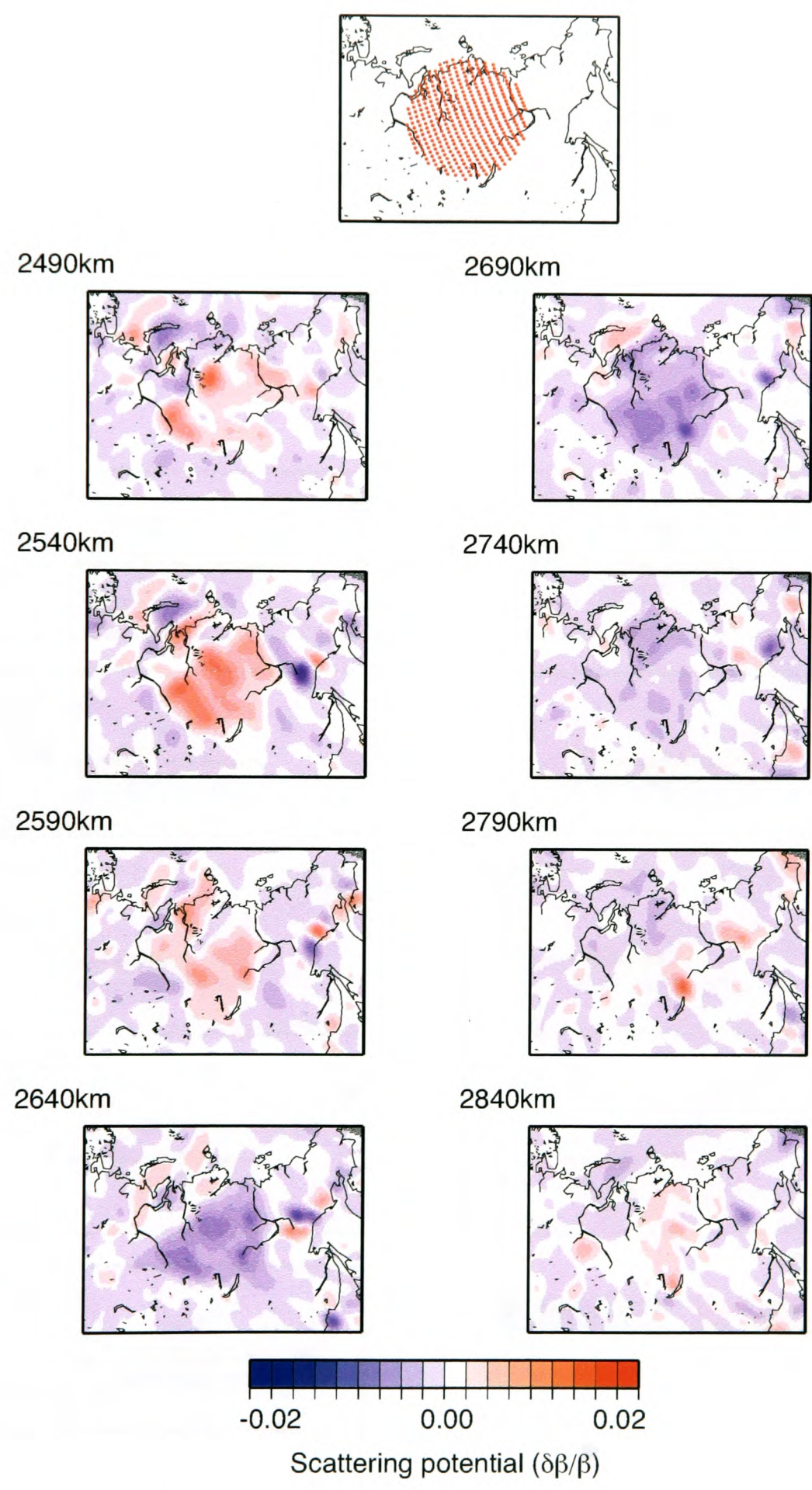


Figure 18.3: Results of migration applied to synthetics generated a disc of radius 800km at 2540km depth (model 1). Scatterer position is shown in the top map an each scatterer has a strength of 0.035.

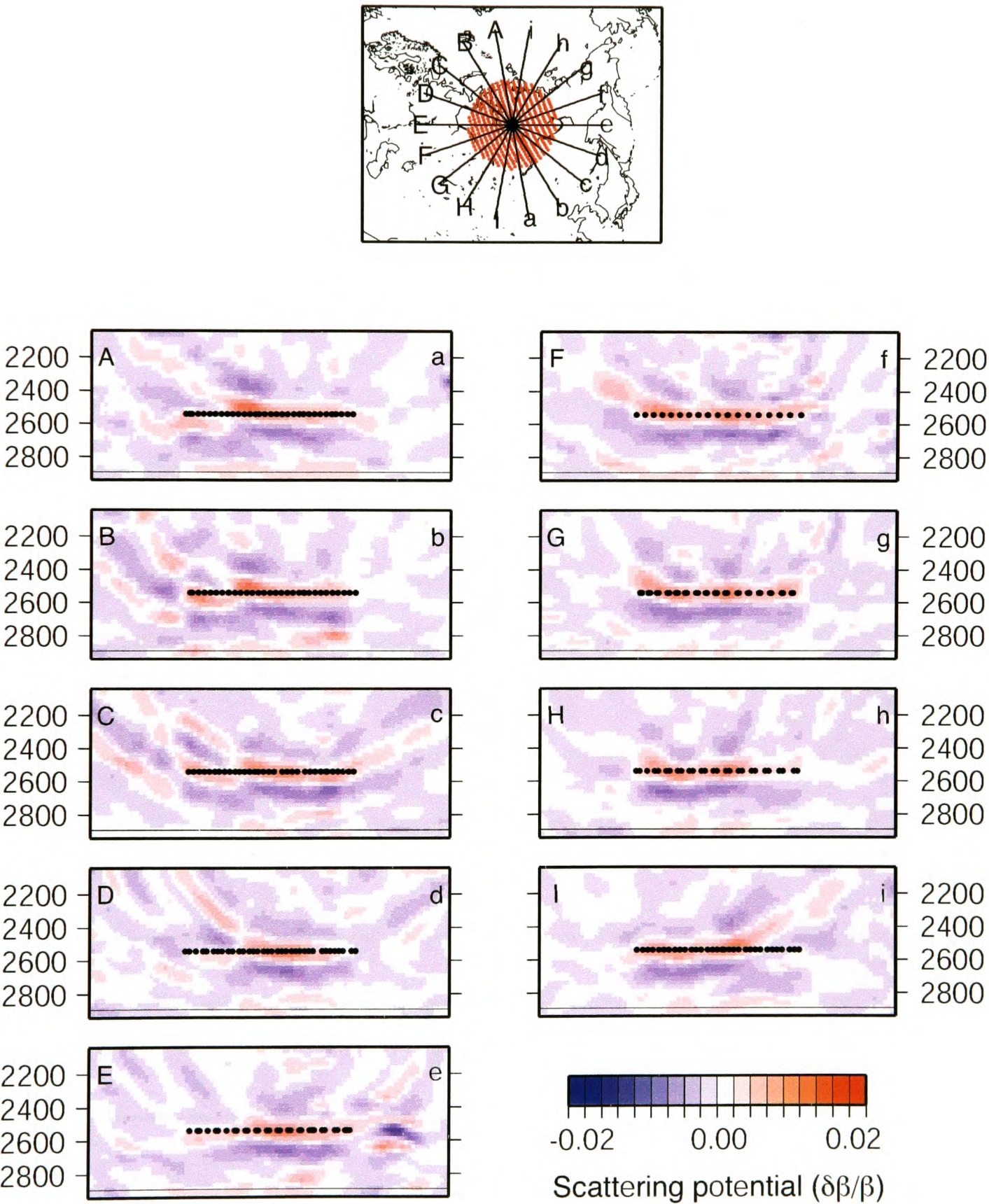


Figure 18.4: Cross-sections for synthetic results generated using model 1.

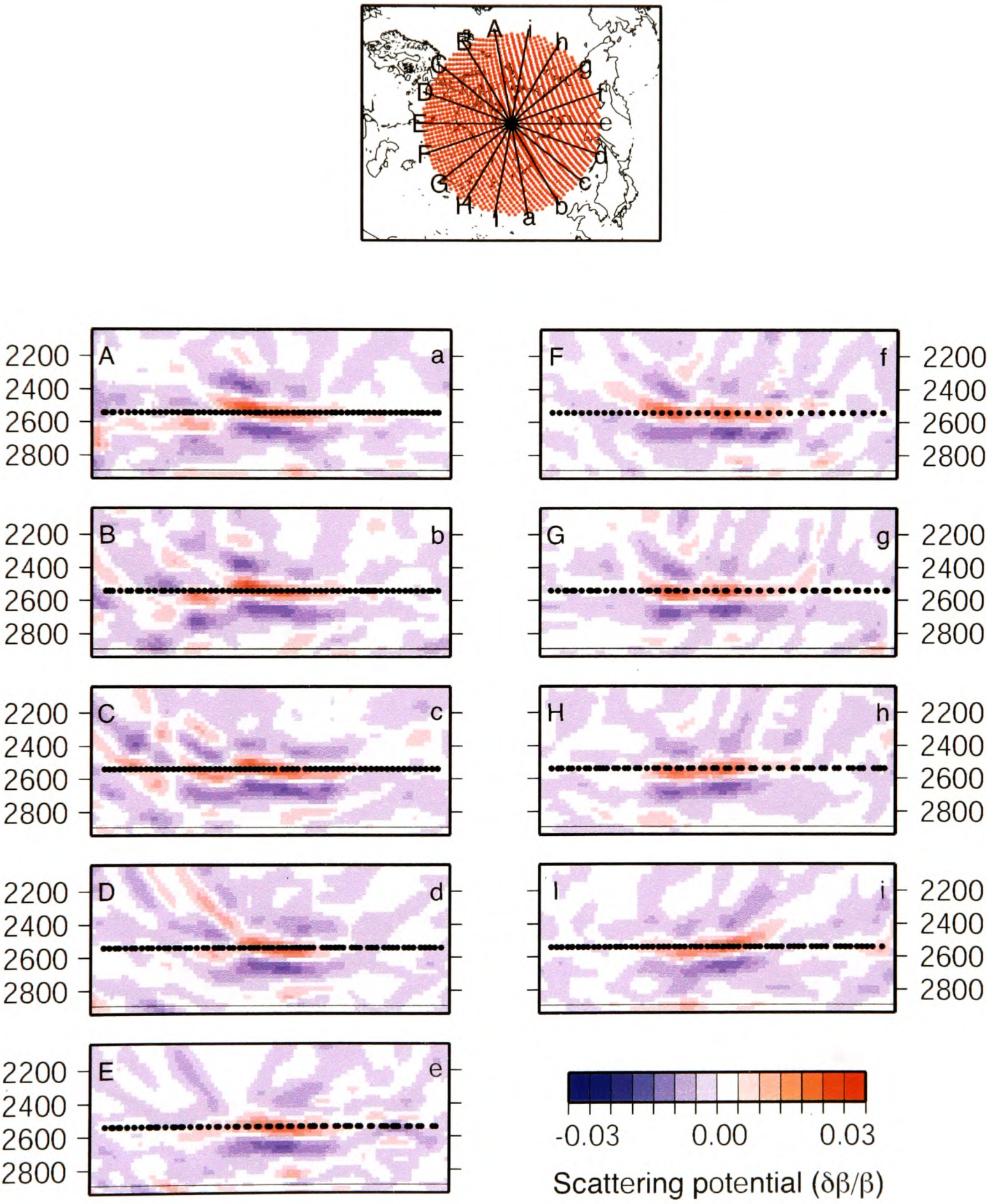


Figure 18.5: Cross-sections for synthetic results generated using model 2 a disc of radius 1600km at 2540km depth. Each scatterer has a strength of 0.03.

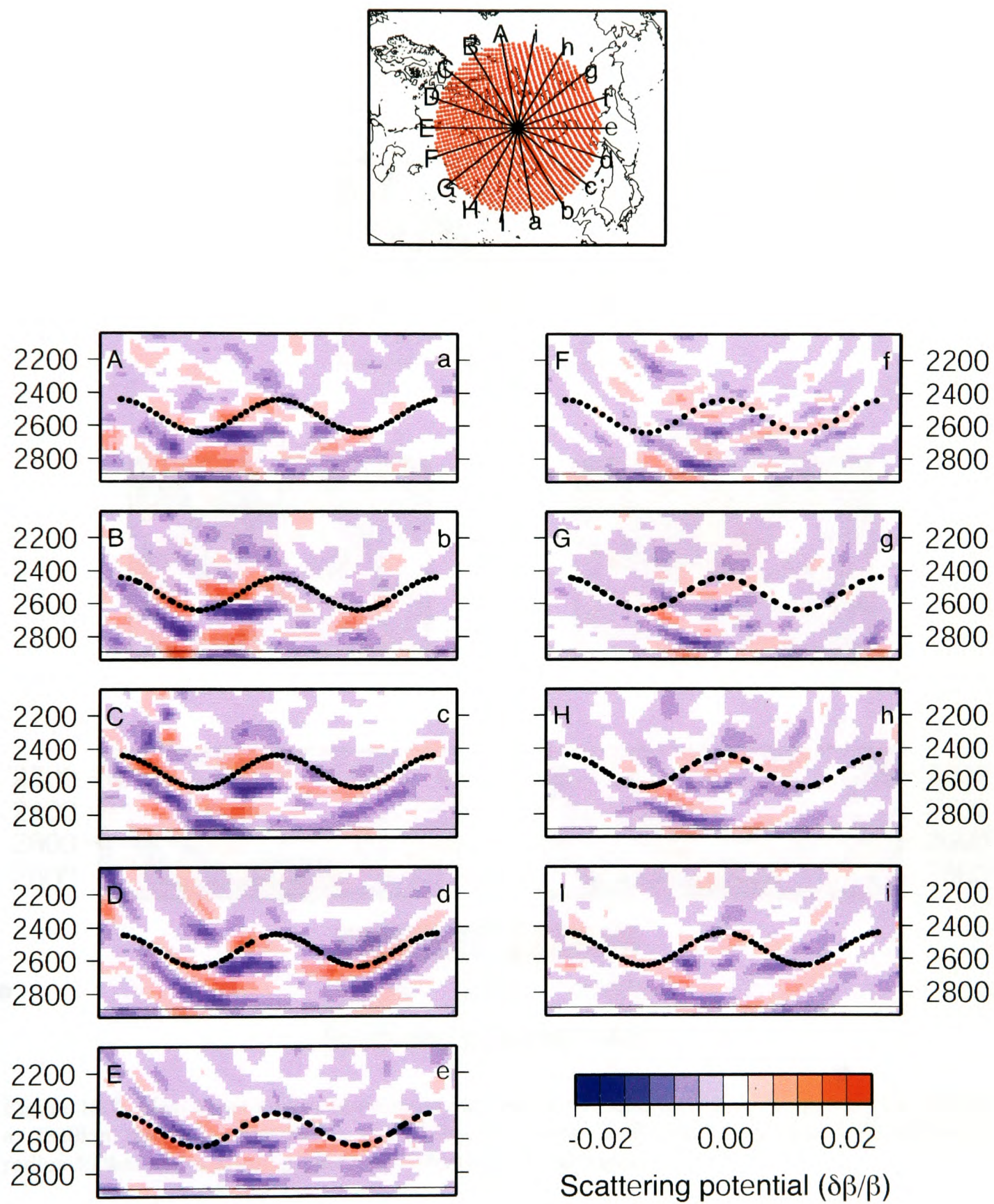


Figure 18.6: Cross-sections for synthetic results generated using model 3 a disc of radius 1500km at 2540km depth. Scatterer depth is given by $2540 + 100 \cos(\frac{2\pi\Delta}{\lambda})$, where Δ is the distance to the disc centre, and topography wavelength λ is 1500km. Each scatterer has a strength of 0.035.

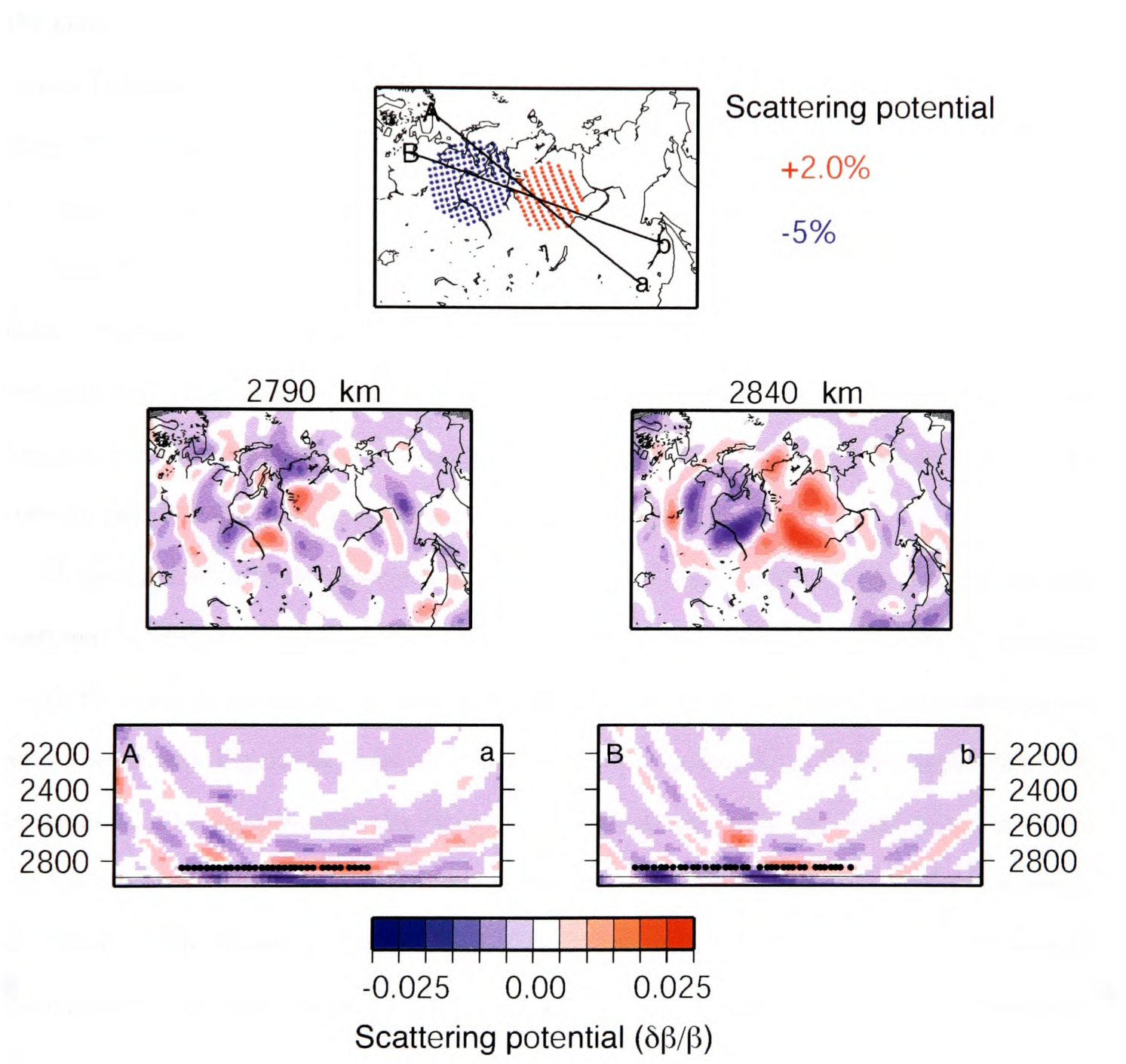


Figure 18.7: Map views and cross-sections for synthetic results generated using model 4; 2 discs at 2840km, one has radius 400km and scattering potential 0.02 (red in top figure), the other has radius 500km and scattering potential -0.05 (blue in top figure).

18.2.2 Complex Bodies

In figures 18.8 and 18.9 simulation of the migration results is attempted using a distribution of small circular heterogeneities (model 5). This distribution of heterogeneities was chosen to reflect the major features seen in the data. The map sections show that the central region has many similar features to the data. Although the exact shapes are not retrieved, and the cross-sections show less reflector topography than that seen in the data (for example see section A-a). Also note that apparent gaps in the reflector are retrieved in sections C-c, D-d, and E-e.

Away from the centre of the cross-sections anomalies are often weaker than those seen in the data. For example the anomalies toward the southeast (for example in the 'c' direction) are not well retrieved. The model recreates the positive anomaly at 2840km, however, figure 18.8 shows less negative structure at 2790km than that observed in the data. It is also worth noting that the reflector near 2700km is not created as an artifact using this model.

Model 6 is similar to the previous synthetic test except that a more continuous distribution of scatterers is used, to simulate an irregular layer (figures 18.10 and 18.11). As with the previous model the centre of the region appears well resolved, but features extending to the Southeast are not retrieved well. Figures 18.10 and 18.11 also show lower amplitudes at 2640km depth than seen in the data. The cross-sections also show that the topography from the model is well resolved, but less reflector topography is seen than for the data. The introduction of the negative anomaly at 2790km depth appears to recreate some of the deeper structure, however, lowermost layers of the migrated data appear to have a more complicated structure than that in the model presented here.

18.3 Comparison with seismic tomography

As with the results for Alaska, (section 17.3) the migration results are compared with the seismic tomography models S20RTS and SKS12WM13 (Ritsema *et al.* (1999); Liu & Dziewonski (1994)), in figures 18.12 and 18.13 respectively. In the cross-sections the D'' reflector, which lies around

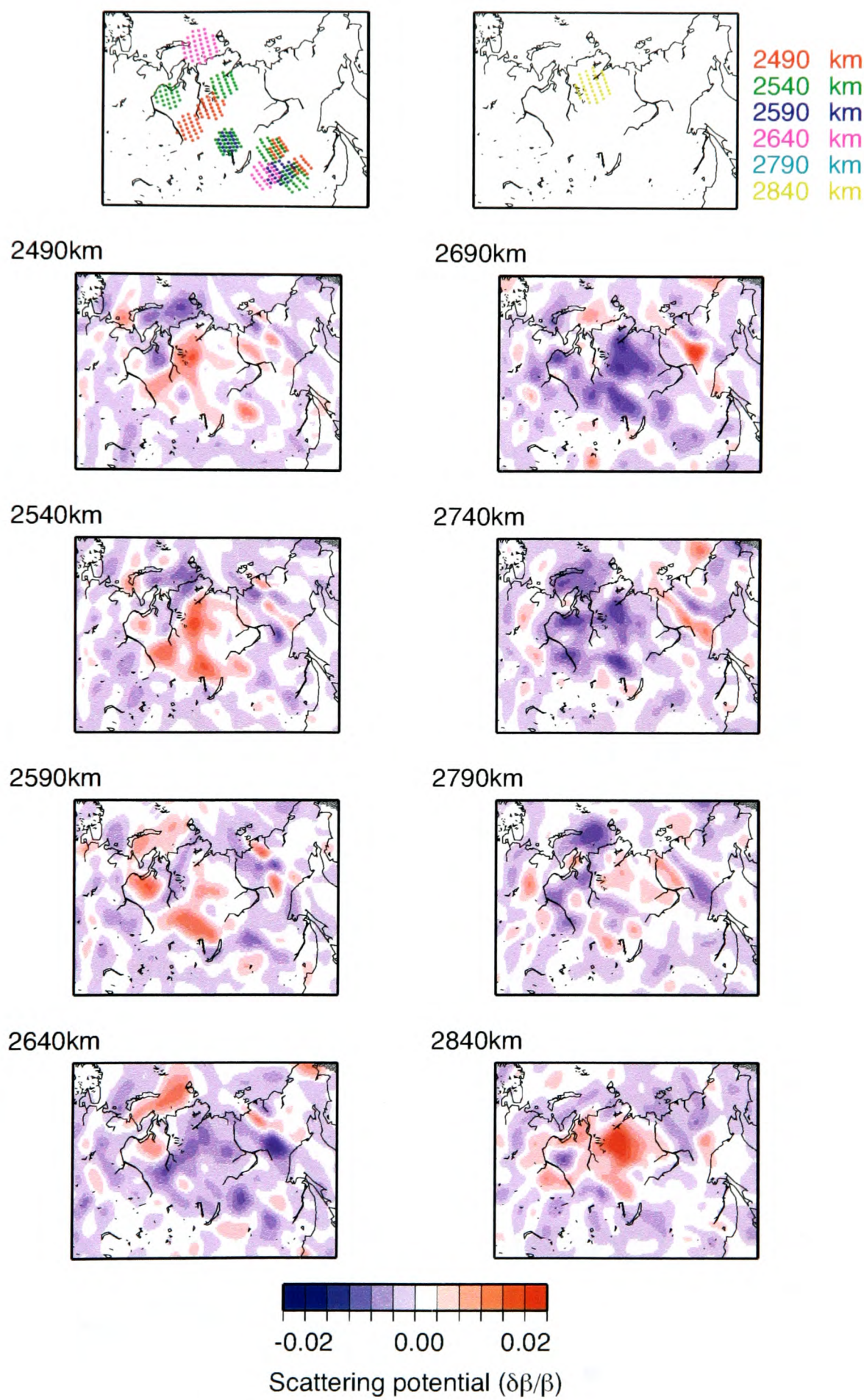
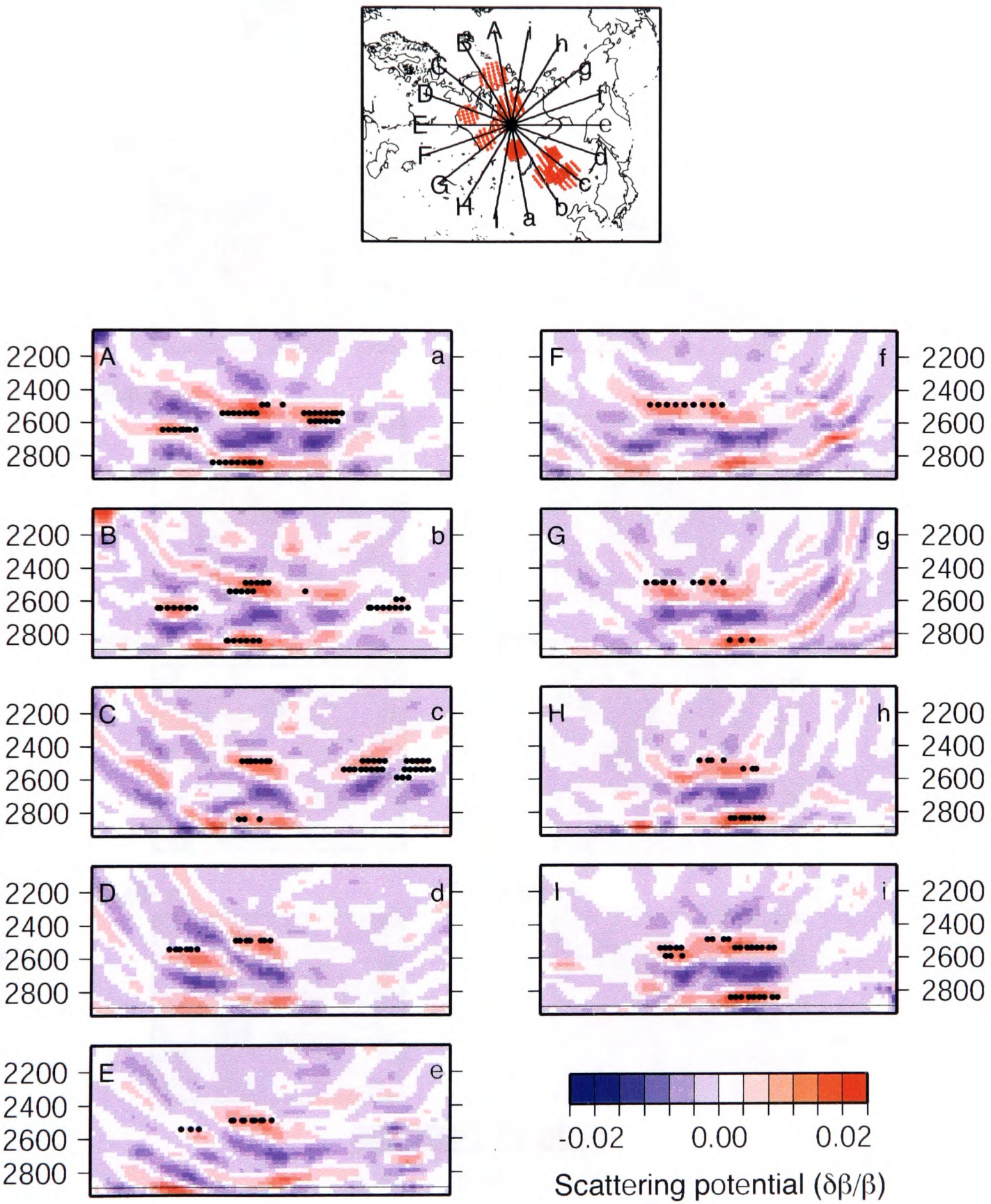


Figure 18.8: Results of migration applied to synthetics generated using model 5. The point scatterer distribution is shown in the top maps, all scatterers have a strength of +0.035, except those at 2840km depth which use +0.02.



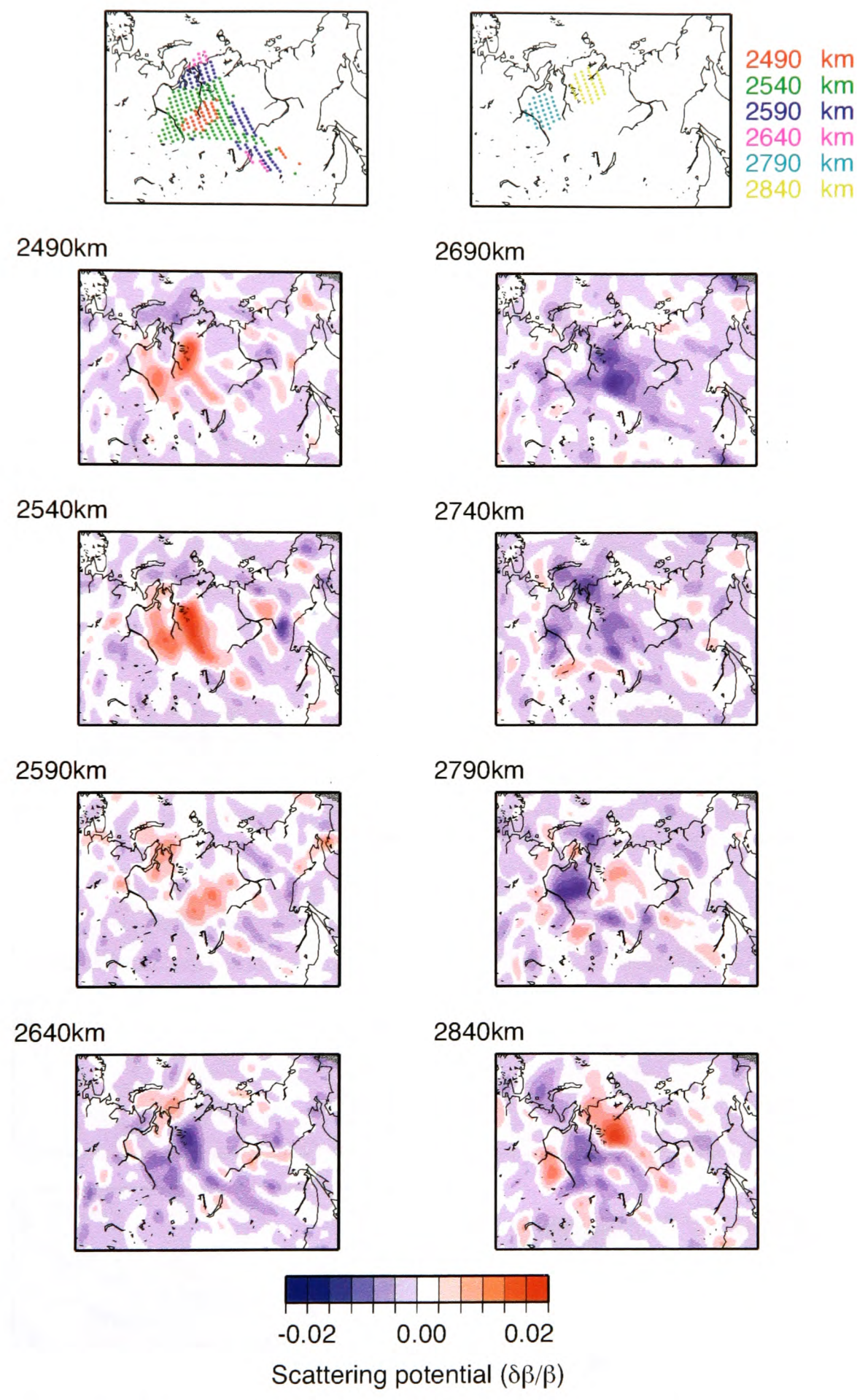


Figure 18.10: Results of migration applied to synthetics generated using model 6. The point scatterer distribution is shown in the top maps, all scatterers have a strength of +0.035, except those at 2840km depth which use 0.02, and 2790km which use -0.05.

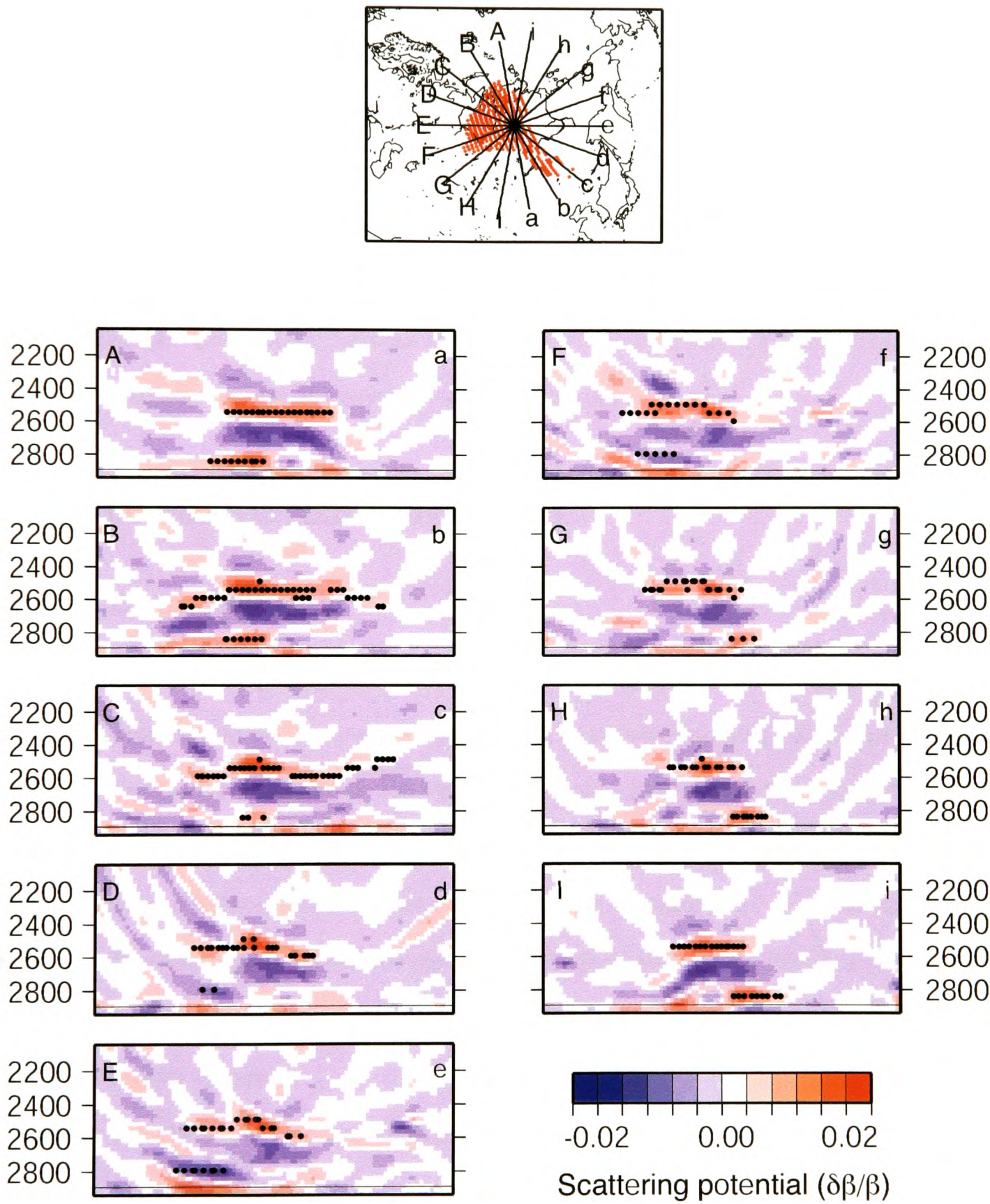


Figure 18.11: Cross-sections for synthetic results generated using model 6.

2600-2400km depth does not appear to be associated with any particular velocity gradients.

Figure 18.12 suggests that the D'' reflector exists in regions of faster seismic velocity in S20RTS. For example in transects E-e, F-f and G-g, where the reflector appears to be restricted to the southwest half of the section, the place where the reflector stops coincides with the edge the fast velocity anomaly. However, comparison with SKS12WM13 does not support this, sections A-a, B-b, and C-c in figure 18.13 show the reflector coinciding with regions of near zero or positive velocity anomalies.

The near CMB reflector often coincides with fast velocity anomalies. Although as with the Alaska region the disagreement between the tomographic models makes it difficult to make any definite comparison between the reflectors and local velocity anomalies.

18.4 Discussion and Conclusions

18.4.1 D'' discontinuity

The D'' discontinuity in this region has been extensively studied over the last 15 years. Gaherty & Lay (1992) proposed the one dimensional model SGLE for this region which features a 2.75% jump in shear wave velocity 290km above the CMB. The results of Kendall & Shearer (1994) suggest that the discontinuity slopes down to the north and east, and ranges in depth over 2750-2520km with an average of 2620km. However, both of these studies interpreted $S - SdS$ differential times based on reflections from horizontal boundaries. So, a more complicated interface (as suggested by the results of this study) could generate a wide range of measurements for the discontinuity depth and velocity change.

Bilek & Lay (1998) used a migration approach which assumes reflections from a planar but not necessarily horizontal boundary. They suggested a reflector which varies between 2605-2620km depth. However, recently the short period study of Thomas *et al.* (2004b) identified topography for the D'' reflector in this region ranging between 2545-2705km depth over a small region (just to the west of the centre of the study region). Differences between the results presented here and

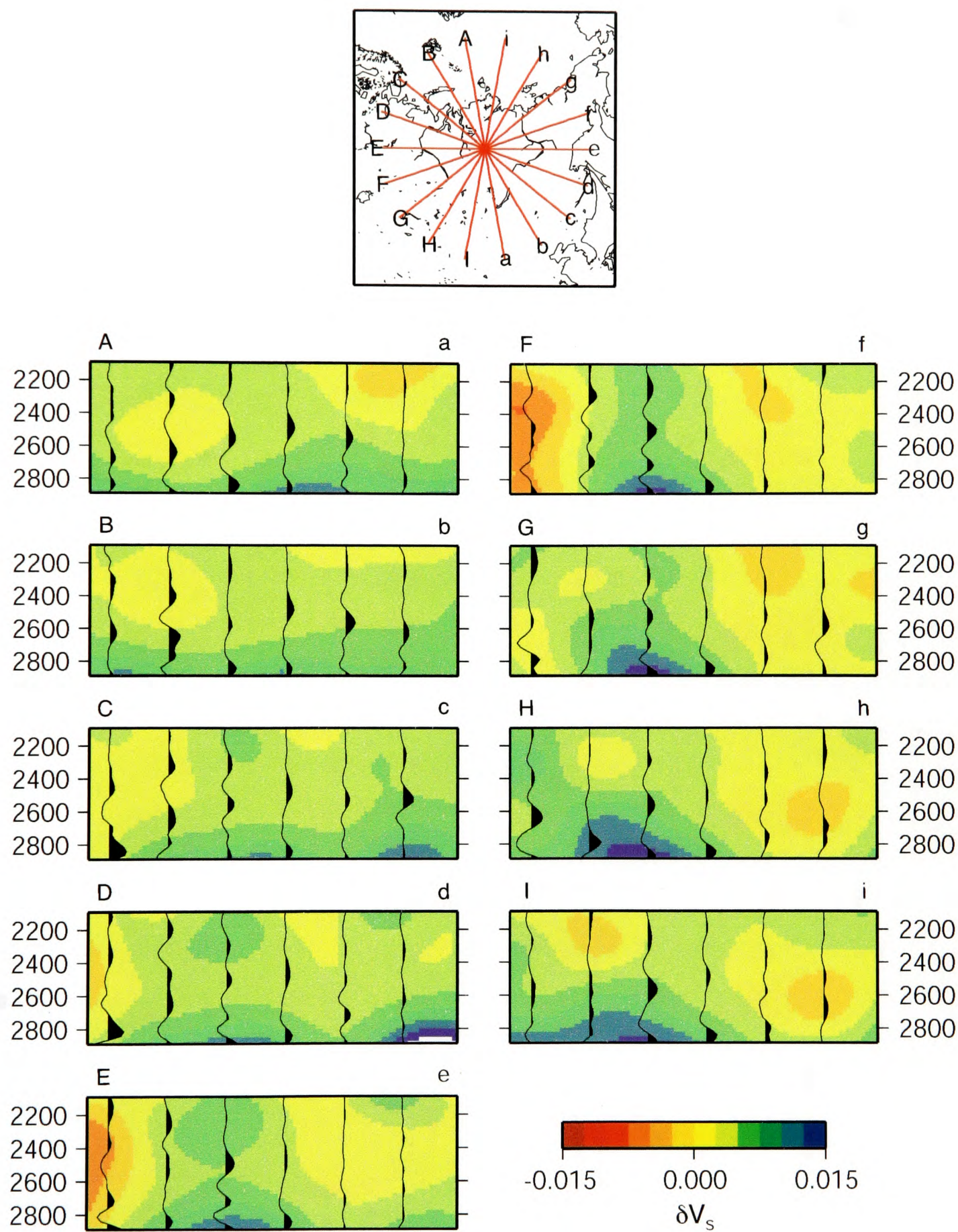


Figure 18.12: Comparison of migration results for Northern Central Asia region with seismic tomography model S20RTS (Ritsema *et al.* (1999)). Colour backgrounds show S-wave velocity perturbation for bottom 800km of the mantle, the superimposed traces are depth profiles through the migration results plotted at their location along the section profile.

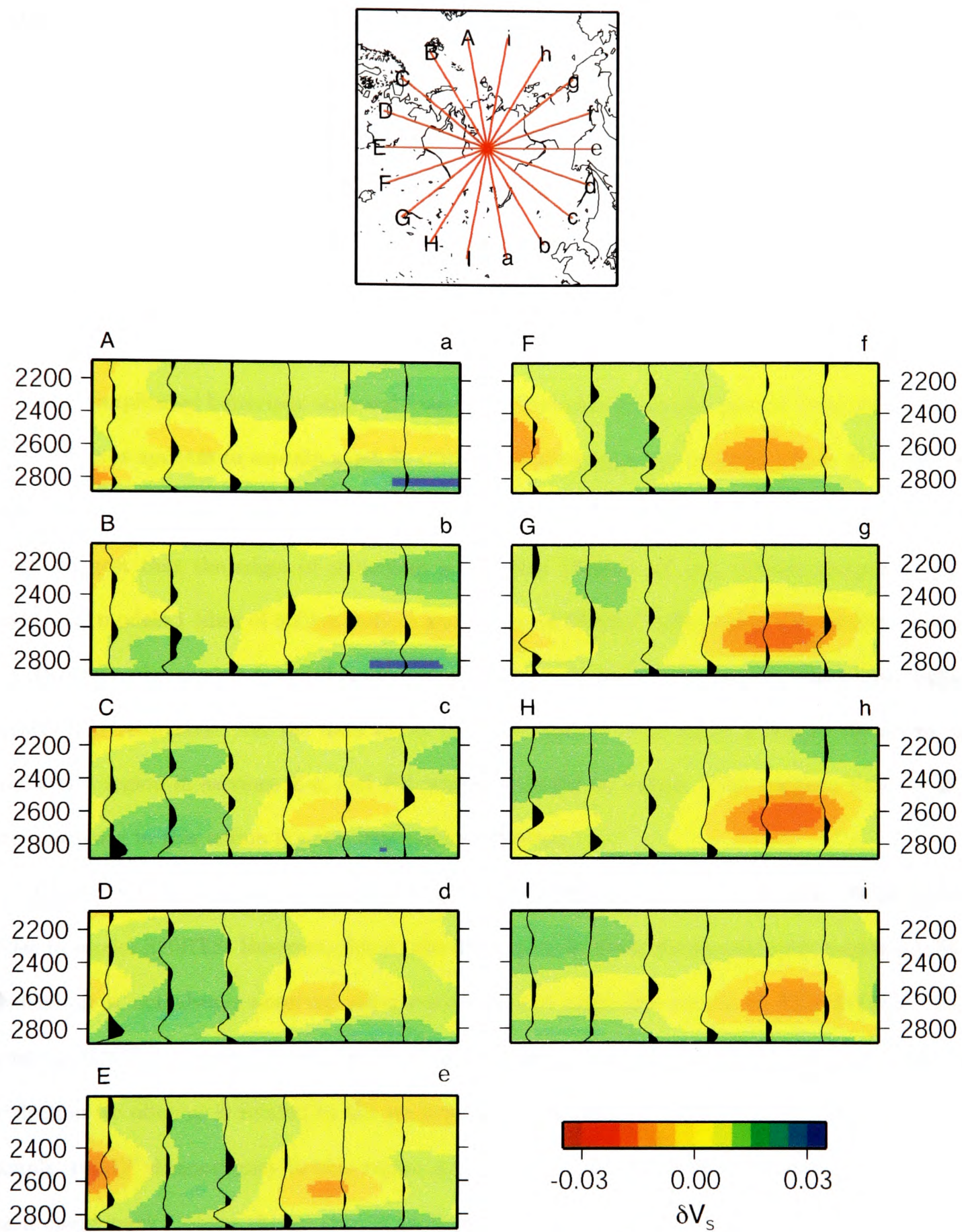


Figure 18.13: Comparison of migration results for Northern Central Asia region with seismic tomography model SKS12WM13 (Liu & Dziewonski (1994)). Colour backgrounds show S-wave velocity perturbation for bottom 800km of the mantle, the superimposed traces are depth profiles through the migration results plotted at their location along the section profile.

these previous studies could be due to the choice of reference phase and background model used for the migration.

The migration results suggest a complicated D'' discontinuity beneath North Central Asia with the depth of the discontinuity at the shallow end of the spectrum suggested by previous studies, mostly lying in the depth range 2640-2490km (250-400km above the CMB). However, the topography appears complicated, and given the different frequency content of the data, grid spacing, and the use of a Gaussian smoothing filter the most recent results of Thomas *et al.* (2004b) are probably not irreconcilable with those found here.

The complicated behaviour observed from the D'' reflector is not just limited to its topography. The reflector appears to contain gaps. In many cases these gaps are well away from the centre of the study region where discontinuity structure is not reliably retrieved. The first two resolution tests suggest that the edges of structures larger than 1600km are not reliably imaged, as the synthetic models 1 (disc of 800km radius) and 2 (disc of 1600km radius) give very similar results.

However, the models 1 and 3 show that if the layer is present just to the east of the region centre it is resolvable. Yet the data shows that the reflector stops more or less in the centre of the study region in sections E-e, and F-f, where it should be visible. This suggests that the D'' discontinuity in this region is not a continuous reflector.

Figure 18.12 shows that the truncation of the D'' reflector restricts it to regions of high velocities in model S20RTS. However, due to the differences between the tomography models similar behaviour is not always observed when comparing our migration results to SKS12WM13. Assuming S20RTS is correct in this region, it is tempting to speculate that the cause of the D'' reflection we observe is related to the presence of a fossil slab in the lowermost mantle. For example, the D'' discontinuity in this region could simply mark a region of thermal heterogeneity, or it could mark a phase change that is made possible by the presence of subducted material near the CMB. However, this speculation is premature as further study is required to decide if the D'' discontinuity is indeed limited to a region of fast velocity in this region.

Our inability to fully reconstruct the D'' reflector in this region may be due to a variations

in the scattering potential along the reflector. The synthetic modelling suggests the D'' reflection amplitudes can be approximately simulated with velocity anomalies of 3-3.5%. This is only slightly greater than the results of previous studies (Gaherty & Lay (1992); Thomas *et al.* (2004b)), which find a +2.75-3% velocity increase in this region.

18.4.2 Reflectors within the D'' layer

In addition to reflections from the top of the D'' layer the migration results suggest the presence of first order heterogeneities within the D'' layer. At around 2740-2790km depth a second reflector is seen, which is visible as a separate body from the D'' reflector in section F-f. In section G-g it appears to be connected to either the D'' reflector or the near CMB structure. Bilek & Lay (1998) also found evidence for a reflector between depths of 2750 to 2770km underneath North Central Asia. They suggest that this reflector is of smaller spatial extent than the D'' reflector. However, the synthetic tests using Model 3 showed that it is possible to create an apparent second layer using a D'' discontinuity with topography. So it is not possible to be certain that this reflector is real without further study.

If the reflector is genuine the results suggest that the scattering body consists of a planar layer that is limited spatially. The reflector could mark the surface of a thermal or chemical heterogeneity, although it's proximity to a low velocity zone (see next section) might favour the latter. However, Thomas *et al.* (2004b) showed that complicated D'' structure can be created by thermal gradients.

18.4.3 Structure at the base of the mantle

The lowermost layers of the migration results have a complicated structure. However, the synthetic tests which use structure at the base of the mantle (models 4, 5, and 6) show it to be generally well resolved, suggesting that this region contains a complicated pattern of bodies with both positive and negative scattering potentials. Although it is worth remembering that velocity anomaly as retrieved by the migration is an average over a volume. As such anomalies in this region should

be considered averages of the bottom ~ 100 km of the mantle.

A high resolution study was undertaken, within this region, by Thomas *et al.* (2004b) who found velocity decreases 55-85km above the CMB. The results here show a negative anomaly at around 2790km depth in the region examined by Thomas *et al.* (2004b), which could be the same low velocity feature. However, the results over the whole study region suggest that negative scattering potentials are localised features, and are mixed in with positive velocity anomalies.

The results of this study might appear to contrast with the possibility of a global ULVZ at the base of the mantle (Revenaugh & Meyer (1997); Garnero *et al.* (1998)). However, the thickness of this layer is mostly believed to be of the order of 10-20km. Such thin layers are beyond the resolution obtainable here. One possible reason for the patchy low-velocity zones in the lowermost layers of the migration image is that these structures can only be resolved where they are thickest (~ 50 -100km).

Chapter 19

Migration results: New Zealand

19.1 Initial results

The migration results for the New Zealand region are shown in figures 19.1 and 19.2. As with the results of the backprojection (figure 14.5) an anomaly at 2840km to the southwest of New Zealand is retrieved, but here it extends off to the east. In addition the migration also retrieves some much larger positive features to the north and east of New Zealand (for example see section F-f).

In contrast to the previous regions a D'' discontinuity is not a major feature of the migration results for this region. The maps suggest a collection of positive scattering anomalies exists around 2400km depth, near the regions centre. These can also be seen in some of the cross-sections ('A', 'B', 'i' and 'F' directions). While other cross-sections show a collection of scattering bodies at this depth (C-c, D-d, G-g, and H-h). To the south of New Zealand no structure is seen above the anomaly at 2840km (directions H, I, and a).

19.2 Synthetic tests

To assess the impact of the source-receiver distribution on the migration results a series of tests using synthetics were performed (as in chapters 17 and 18). Table 19.1 describes the distributions

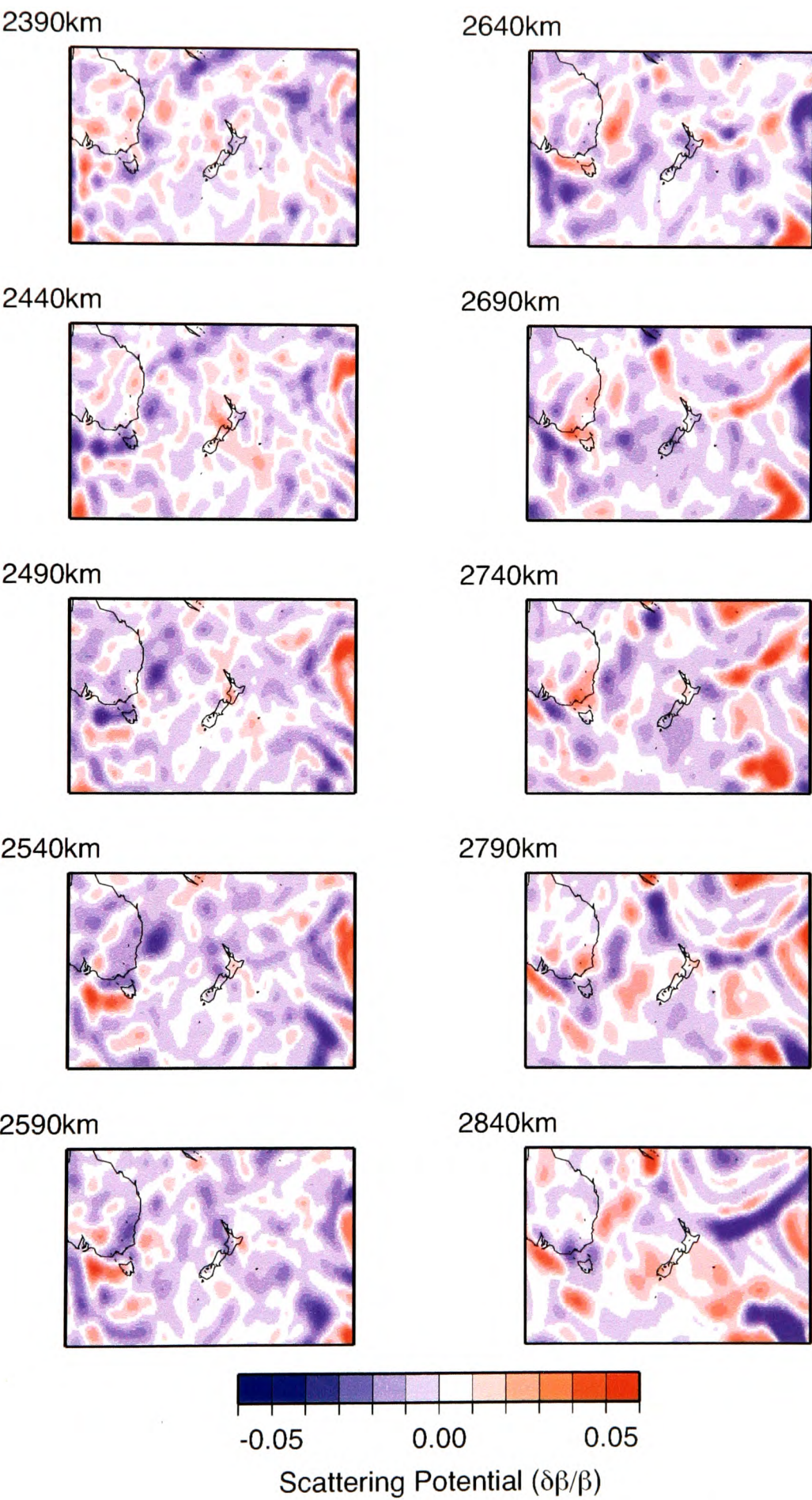


Figure 19.1: Depth slices through the migration results beneath New Zealand.

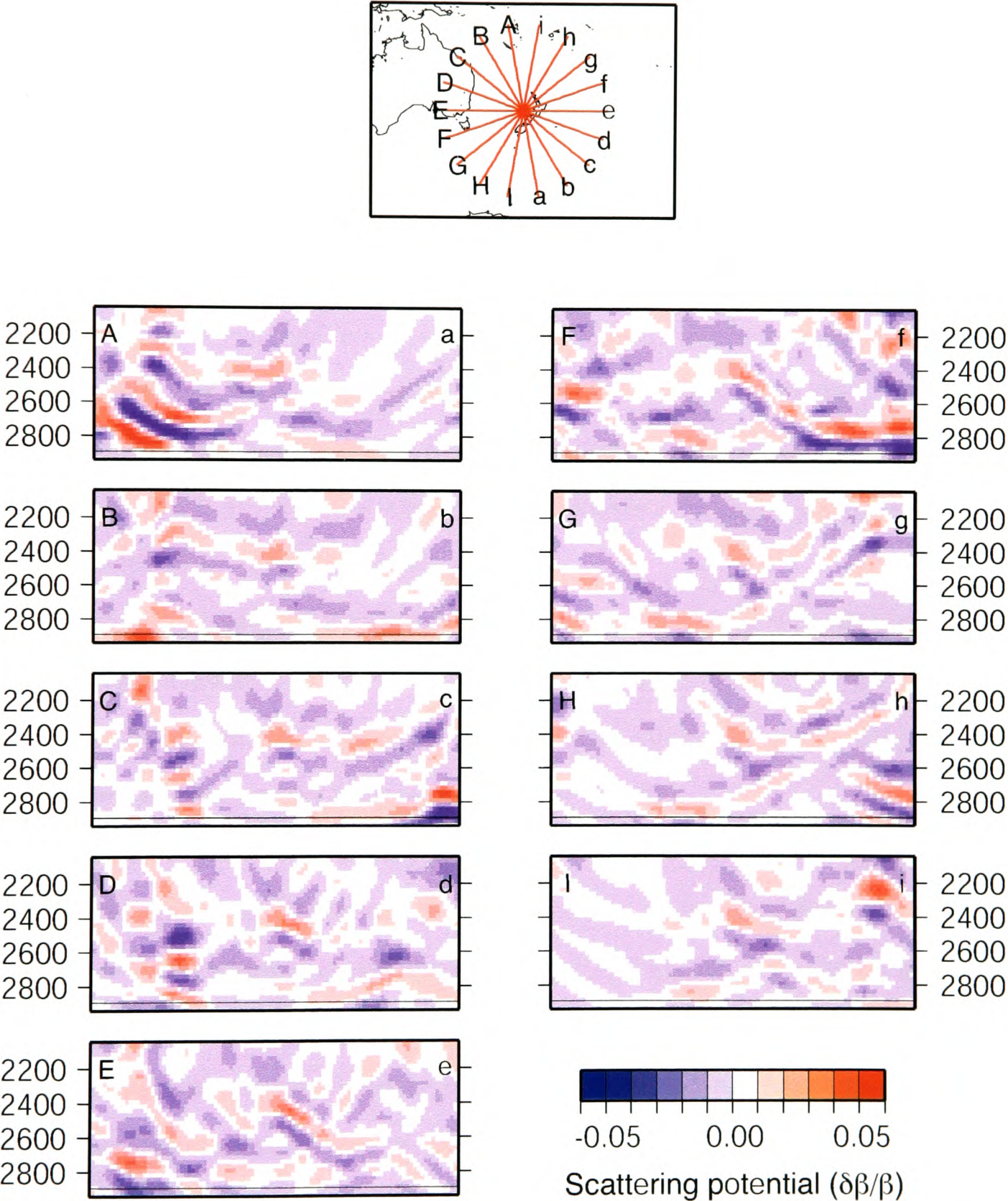


Figure 19.2: cross-sections through the migration results for New Zealand.

Test	Body position	Velocity contrast ($\delta\beta/\beta$)	Figure No.
Model 1	Disc radius=300km 163/-47 2840km	0.025	19.3
Model 2	Disc radius=800km 190/-45 2840km	0.025	19.4
	Disc radius=500km 157/-43 2840km	0.025	
Model 3	Disc radius=1200km 170/-40 2540km	0.035	19.5
Model 4	Disc radius=2250km 175/-30 2540km 200km peak to peak, 1500km wavelength topo.	0.035	19.6

Table 19.1: Description of input bodies for the synthetic tests, co-ordinates for discs refer to the disc centre.

of scatterers used for each test.

Model 1 consists of a small disc of scatterers at 2840km similar to the positive feature seen to the South of New Zealand at this depth in the data. The results (figure 19.3) show that the anomaly at 2840km is well retrieved, with only slightly higher amplitudes than observed in the data. Some artifacts are created at shallower depths. However, nothing is placed to the east or northeast where other anomalies are seen in the data.

For model 2 the response of the migration for two large discs of positive scattering potential at 2840km depth is examined. The results are shown in figure 19.4. It can be seen that the scatterers in the west Tasman sea, and those in the east of the study region are not retrieved. Nor is the gap between the two scattering bodies retrieved by the migration.

In model 3 the visibility of a D'' reflector in the study region is investigated. The input model consists of a circular disc (1200km radius) at 2540km depth. The results in figure 19.5 show that the reflector is only reliably retrieved beneath southern New Zealand. Artifacts are also created in layers above and below the reflector.

For the final synthetic test (model 4) the possibility of D'' discontinuity topography beneath New Zealand and the South Pacific in figure 19.6 is investigated. Model 4 features a disc of 1500km radius. Discontinuity topography is generated within the disc by setting the scatterer depth (km) using $d = 2540 - 100 \cos(\frac{2\pi\Delta}{\lambda})$, where Δ is the distance to the disc centre, and topography wavelength λ is 1500km. Each scatterer has a scattering potential of 0.035.

The migration most accurately retrieves structure in the Southwest of New Zealand ('H' and

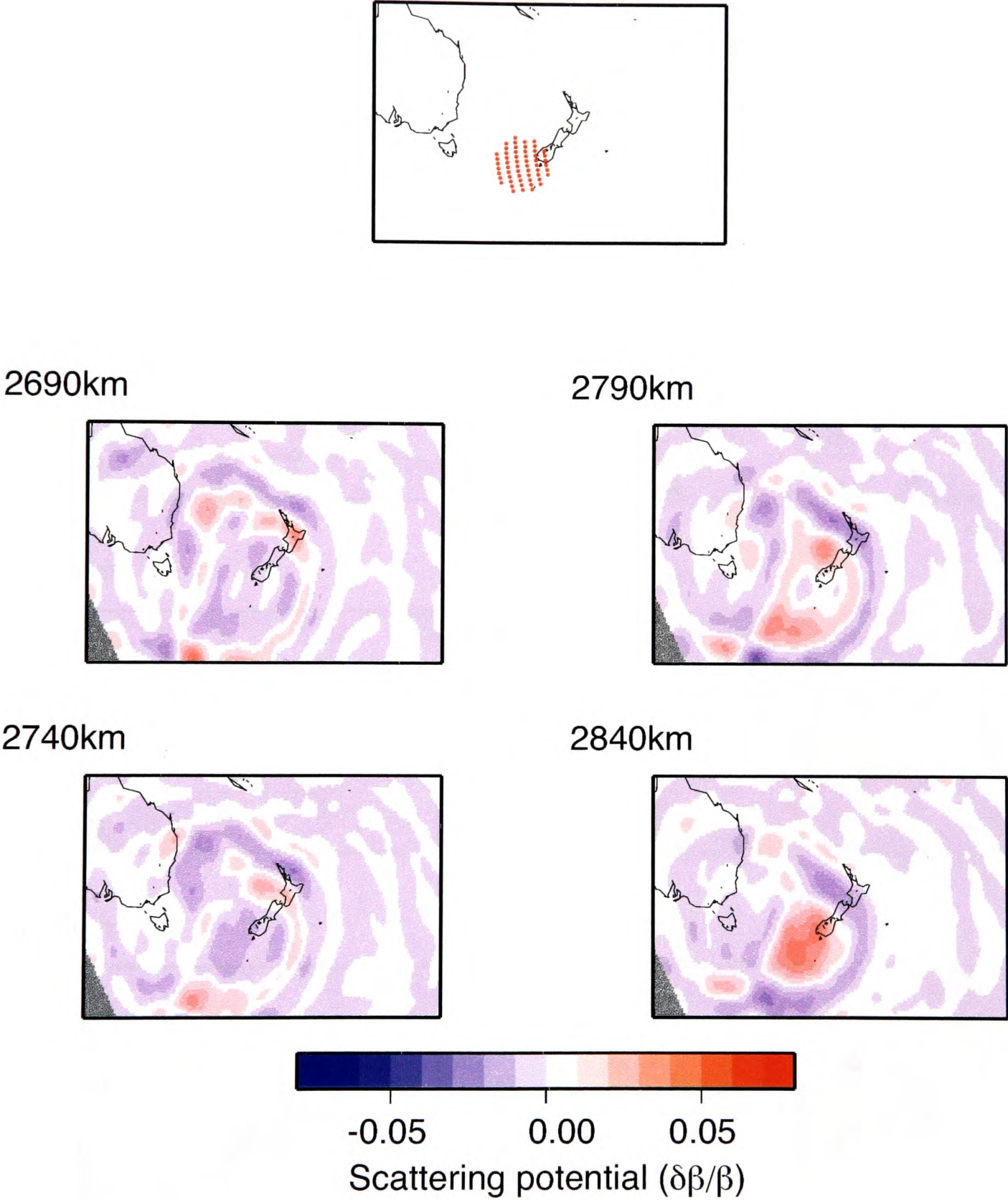


Figure 19.3: Results of migration applied to synthetics generated using model 1 (see table 19.1 for details). Scatterer position is shown in the top map, each scatterer has a strength of 0.025.

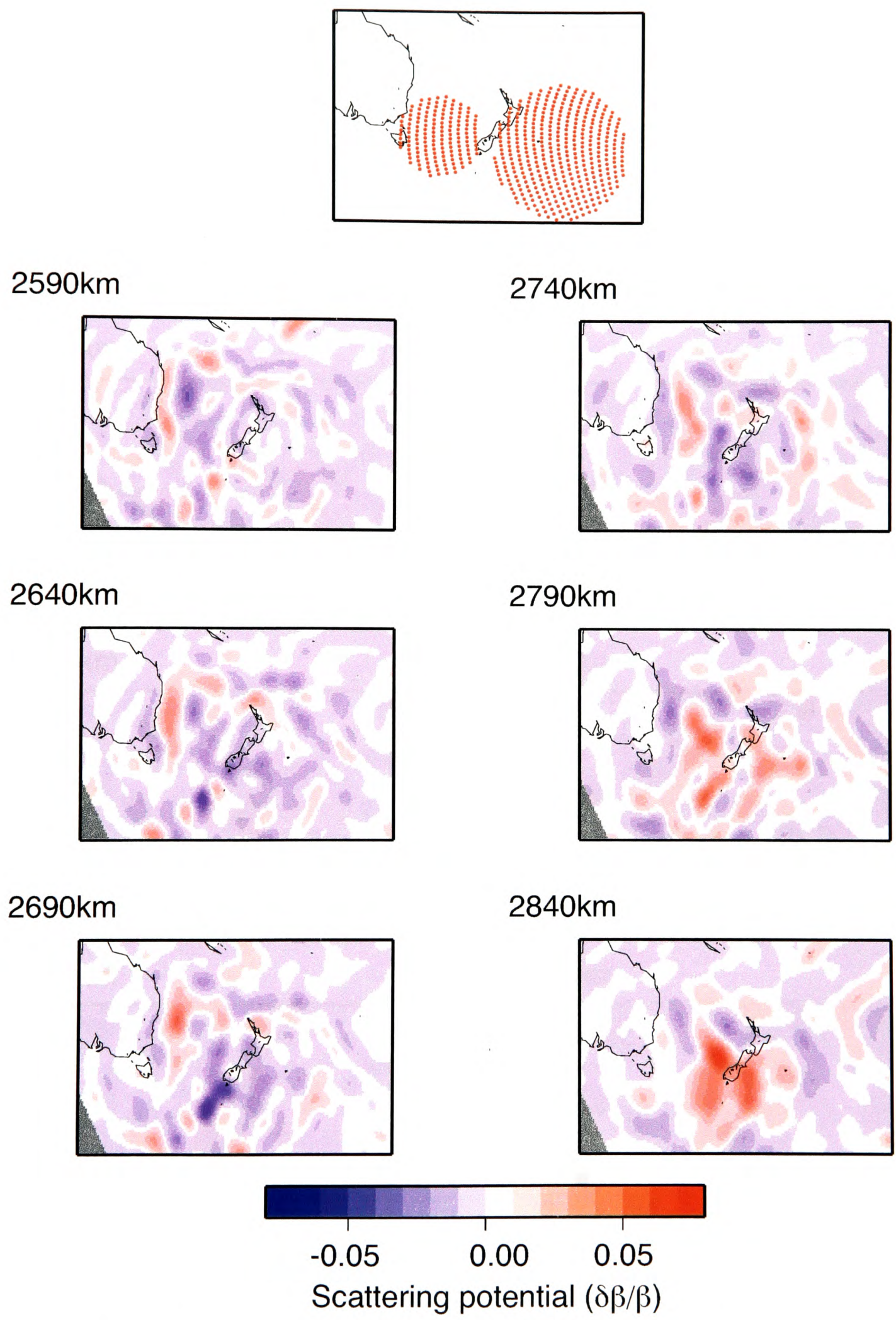


Figure 19.4: Results of migration applied to synthetics generated using model 2. Scatterer position is shown in the top map, each scatterer has a strength of 0.025.

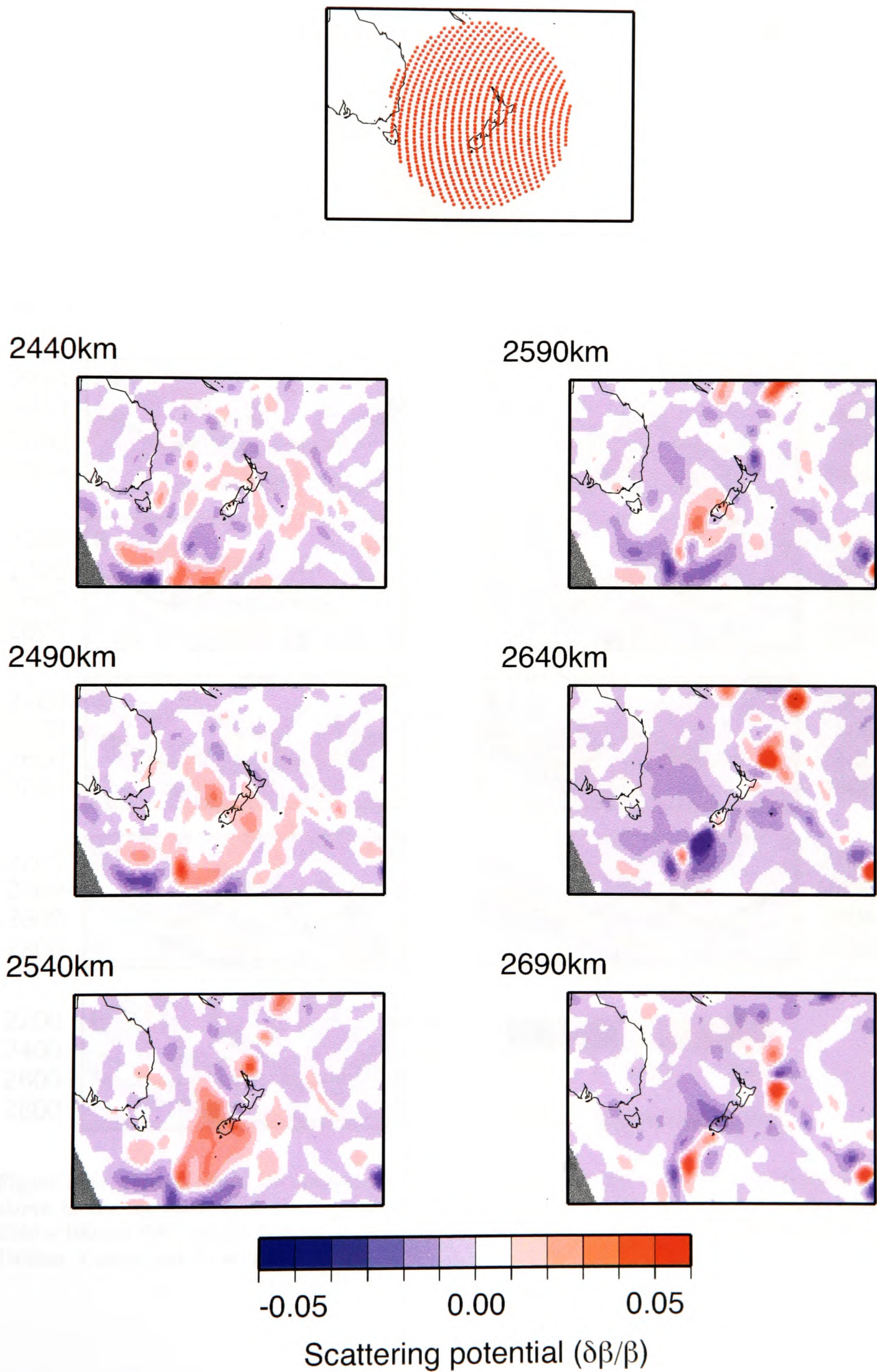


Figure 19.5: Results of migration applied to synthetics generated using model 3. Scatterer position is shown in the top map, each scatterer has a strength of 0.035. Colour scale is saturated at ± 0.05 .

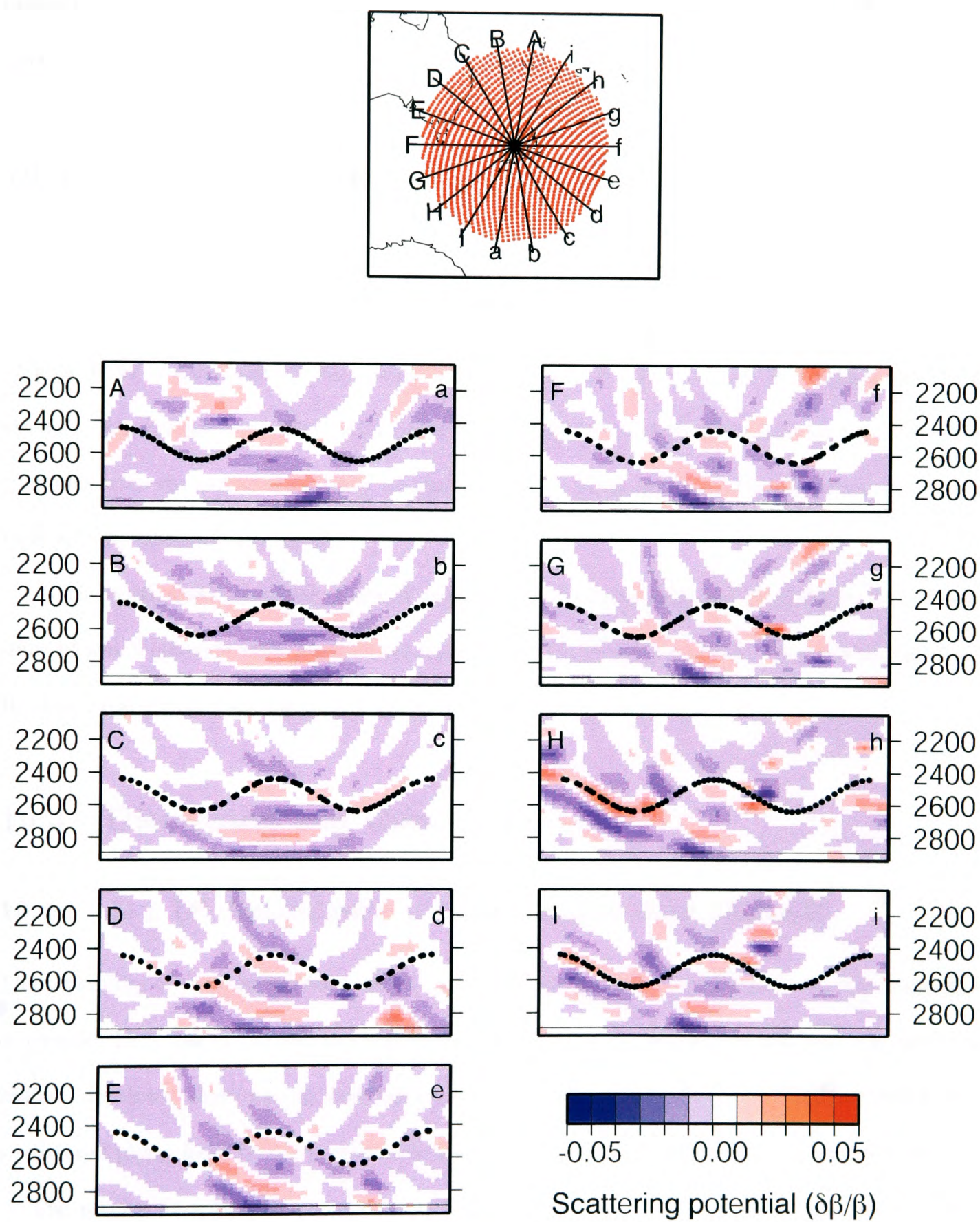


Figure 19.6: Cross-sections for synthetic results generated using model 4. Scatterer position is shown in the top map, each scatterer has a strength of 0.035. Scatterer elevation is given by $2540 + 100 \cos(\frac{2\pi\Delta}{\lambda})$, where Δ is the distance to the disc centre, and topography wavelength λ is 1500km. Colour scale is saturated at ± 0.05 .

'I' directions). In other directions the reflector is not resolved. In the centre of the region the upward cusp of the reflector creates two reflectors in the migration results, this is not seen in the data.

19.3 Comparison with seismic tomography

In figures 19.7 and 19.8 the migration results are compared with the seismic tomography models S20RTS (Ritsema *et al.* (1999)) and SKS12WM13 (Liu & Dziewonski (1994)). Unlike previous regions the D'' region beneath New Zealand and the South Pacific is mostly slower than the global average. An exception to this is the region to the southwest of New Zealand in S20RTS which coincides with the positive anomaly at 2840km depth. However, SKS12WM13 does not show a high velocity anomaly in this area.

To the north and east of New Zealand seismic velocities in S20RTS and SKS12WM13 are slower, some of which coincide with positive anomalies in the migration results (for example section A-a). However, resolution tests show this area to be poorly resolved.

19.4 Discussion and Conclusions

19.4.1 Is a D'' discontinuity present beneath New Zealand?

The possibility of the non-existence of a D'' discontinuity beneath New Zealand was first suggested in chapter 14.3, based on the analysis of the backprojection results. A reflector was found at 2840km depth but there was no evidence for a reflector 200-450km above the CMB consistent with a scattered phase from D'' . The migration results do nothing to contradict this interpretation.

The results from synthetic tests 3 and 4 show that if a D'' discontinuity is present beneath New Zealand and the South Tasman sea it should be retrieved by the migration procedure. Even if discontinuity topography were more extreme than that in model 4 the migration should retrieve some artifacts from reflections off this boundary. However, no such artifacts were observed. The results suggest that for a D'' discontinuity to be present in this area between ~ 100 -850 km above

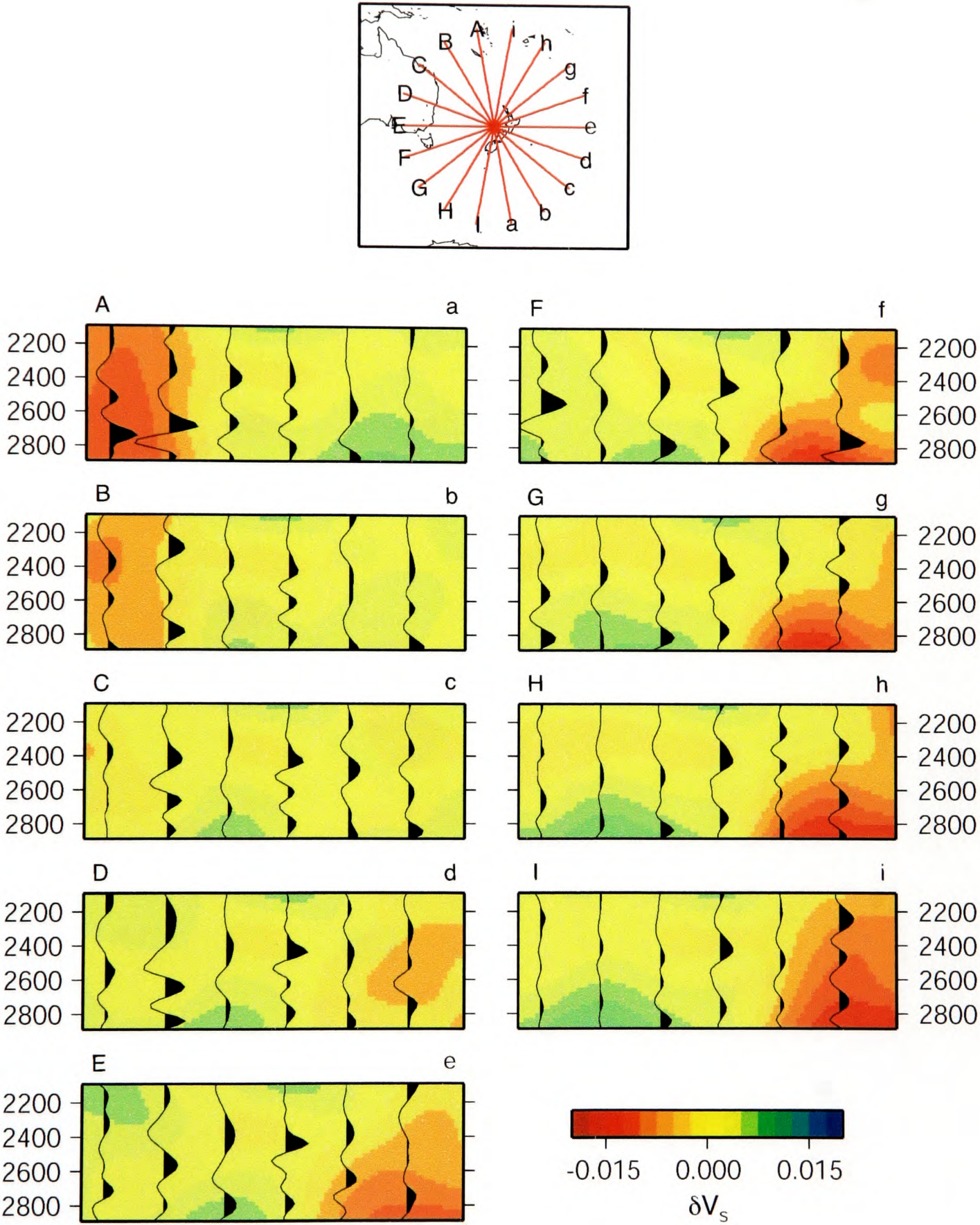


Figure 19.7: Comparison of migration results for New Zealand region with seismic tomography model S20RTS (Ritsema *et al.* (1999)). Colour backgrounds show S-wave velocity perturbation for bottom 800km of the mantle, the superimposed traces are depth profiles through the migration results plotted at their location along the section profile.

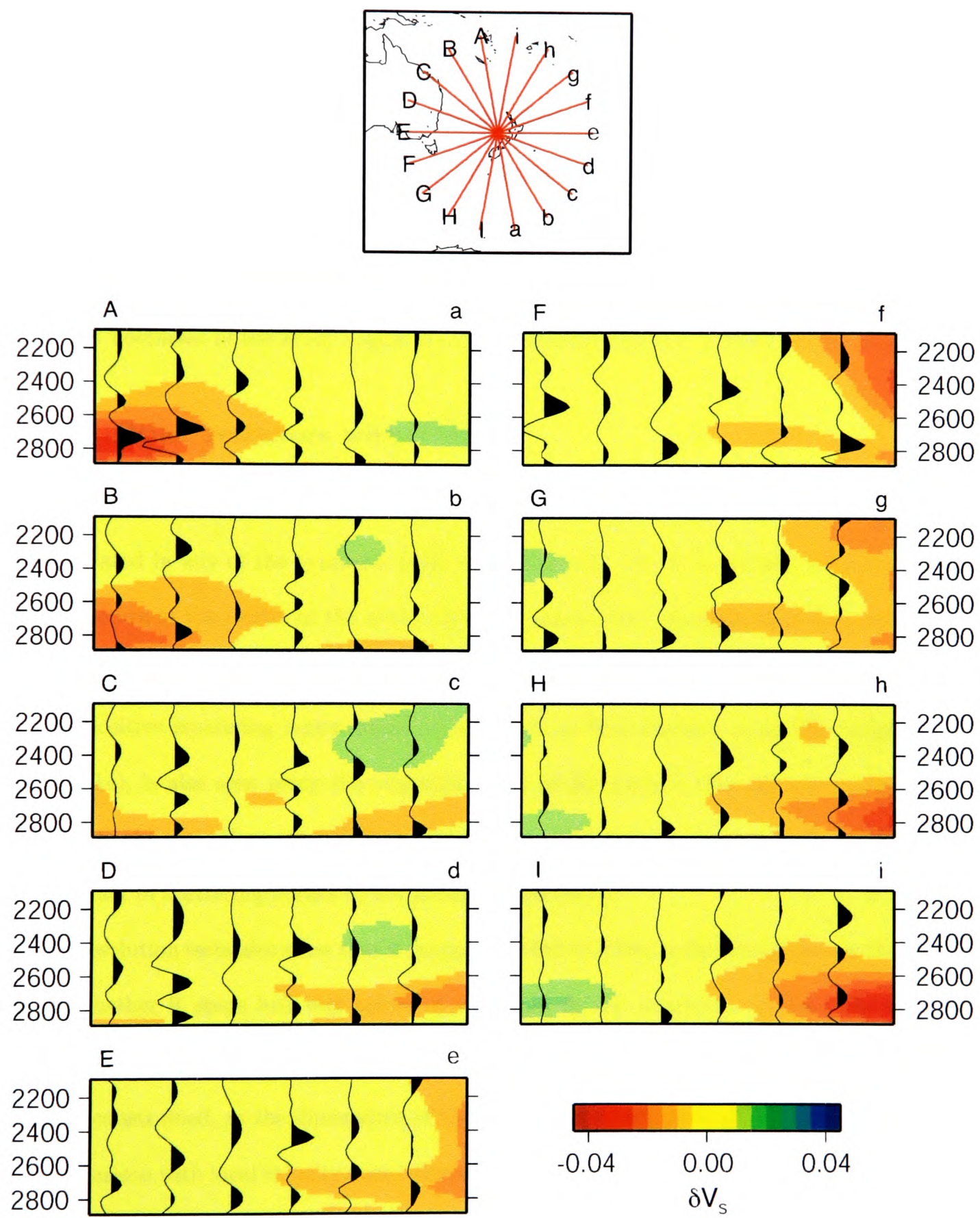


Figure 19.8: Comparison of migration results for New Zealand region with seismic tomography model SKS12WM13 (Liu & Dziewonski (1994)). Colour backgrounds show S-wave velocity perturbation for bottom 800km of the mantle, the superimposed traces are depth profiles through the migration results plotted at their location along the section profile.

the CMB (2790-2090km depth), to the south of New Zealand, the discontinuity must be much weaker than the scattering potential in model 4 (3.5%).

In the region where a D'' reflector is definitely not seen (south and southwest of New Zealand) seismic velocities are near constant and close to the global average in both S20RTS and SKS12WM13. Other parts of the study region are not so well resolved and the results do not preclude the existence of a D'' reflector over the entire area. The migration results show positive scattering anomalies around 2400km depth beneath Cook Strait and the North Island. Scatterers in the north and northeast of the study region are not necessarily resolved in their correct location.

19.4.2 Other reflectors within the D''

The eastern side of the study region shows large positive and negative amplitudes. These were not replicated in any of the synthetic tests considered here. So, it is unlikely they are artifacts from structure in the centre of the study region. It seems more plausible that they are artifacts from less well resolved regions to the north and east.

The positive scattering region found to the south of New Zealand using the backprojection (chapter 14), is also seen using the migration. The results suggest that the scattering region is extended to the east. However, this is an area of lower resolution and it may be that this structure is an artifact of scattering bodies to the north and northeast.

The resolution tests also show that it cannot be established if this body extends across the Tasman sea, or whether it stops half way across as suggested by our migration results. The migration results and synthetic tests suggest the body has a velocity anomaly of around 2.5%, although this is poorly constrained, as the dimensions of the body are not well resolved.

Comparison with local velocity anomalies in S20RTS suggest that this scattering body is related to the presence of a fast velocity anomaly. However, comparison with the tomography model SKS12WM13 does not support this.

Chapter 20

Concluding remarks part II

20.1 General comments on the backprojection

In this part of the thesis a dataset of phase striped seismograms was produced by removing the S and ScS pulses from 1075 high quality traces. In the subsequent sections backprojection and migration procedures were developed and applied in order to map arrivals in the phase stripped traces to the corresponding scattering positions in the mantle.

Initially a simple diffraction stack/backprojection was applied to arrivals in the deconvolved phase stripped traces. This technique has the advantage that the processing steps are all fairly simple, and the results can be easy to interpret. This makes the technique useful as a ‘first look’ for the presence of scattering bodies.

However, the use of deconvolved records has a few drawbacks such as, the assumption of reflections from planar boundaries, and the use of a reduced/summarised dataset (i.e. deconvolved records). The latter means that features can be missed out. For example the reflectors near the CMB beneath Alaska and North Central Asia were not picked up using this technique, presumably because only the D'' reflection was required to fit 95% of the energy in the trace.

In addition the technique does not take into account factors which affect the amplitude of scattered phases such as source mechanisms, reflection angles, geometrical spreading, or even the

number of traces stacked at each image point. This means that the backprojection amplitudes are not comparable between regions, and are not a direct measure of heterogeneity.

20.2 General comments on the GRT migration

The migration produces an estimate of S-wave velocity anomaly, through an assumption of proportionality between density and seismic velocity perturbations. Errors in the retrieved amplitudes can result from wave propagation effects not allowed for in the very simplistic forward model. Effects such as attenuation, anisotropy, 3-D heterogeneity, and errors in source positions and mechanisms, can all effect the observed scattered amplitudes but were not included in the migration procedure.

More fundamentally the Born approximation (in the form used here) does not allow for frequency dependent wave propagation effects, or multiple scattering events. These factors alter the magnitude and shape of the retrieved anomalies either directly, by producing scattered arrivals of different amplitudes, or by producing variations in the arrival times of scattered phases and leading to errors due to incoherent stacking of scattered arrivals.

In addition to amplitude errors due to factors not included in the modelling, the retrieved amplitudes for the synthetic experiments also show that the migration amplitudes are often lower than those used as an input. This is partly an artifact of the Gaussian smoothing applied to the migration results. Other factors include, approximations made in the migration procedure such as the damping of the radiation pattern term or the limiting of weights in the integral to a factor of 3. However, these are unlikely to effect amplitudes away from the nulls in source and scatterer radiation patterns. The reduced amplitudes could also be due to the incomplete formation of the spherical integral surface, which in turn is caused by the limited range of source receiver geometries. For this reason the migration amplitudes only have meaning when compared to amplitudes produced using synthetics for the same source-receiver geometries.

The data for the synthetic tests were produced using the Born approximation along with ray-

theoretical Green's functions, and the migration is just the inverse of this process. Accordingly, the synthetic amplitudes do not include any of the propagation effects that are not included in the inverse process. So, they cannot be regarded as test of the validity of the migration algorithm. The synthetic tests only expose the influence of source receiver geometry on the recovered migration amplitudes.

It can also be seen from the migration amplitudes produced using different size discs in the synthetic tests that different size discs can produce the same amplitudes, if the scattering potential changes. This highlights a trade-off between scattering potential and volume. However, this trade-off will only exist where the scattering volume is under sampled, and is smaller than the Fresnel volume.

20.3 Types of structure in D''

In this section the aim is to give a brief summary of the types of structure in the three study regions. Additional results of the migration can be found in appendix B, although these results are only preliminary as synthetic resolution tests have not been performed. To simplify the descriptions the scattering bodies are grouped into 3 types of structure, although, this grouping is rather subjective.

20.3.1 The D'' discontinuity

The major structure beneath Alaska and North Central Asia is a reflector with a typical depth of 2640-2490km (250-400km above the CMB). These depths are consistent with the upper boundary of the D'' layer. Beneath Alaska it is possible to simulate the scattering body with a flat disc at 2640km. However, the body beneath North Central Asia requires considerable topography (~ 150 km peak to peak).

In the Alaska and North Central Asia regions the D'' reflection amplitudes are consistent with scattering potentials of 3%. These amplitudes are a little higher than those from previous studies (for example see table 12.2). However, these mostly assume reflections from infinite, and

sometimes planar, layers. This assumption leads to an underestimation of the changes in velocity and density if the reflector is not infinite.

Perhaps one of the most important observations of the D'' reflector is evidence that it is not continuous beneath North Central Asia, and is not present in other areas (e.g. to the south of New Zealand). So, the D'' discontinuity is not a global feature. Previous authors have suggested this could be due to topography and focussing affects (for example see Wyession *et al.* (1998)). However, the synthetic tests allow this possibility to be discounted. Variations in reflection amplitudes can also be created by variations in discontinuity width. However, most estimates of the width for the D'' reflector are a maximum of 50-75km (Wyession *et al.* (1998)). This is of the same order as the grid spacing used in the migration. Accordingly we believe that the presence/absence of the D'' discontinuity is due to the reflector being created by localised structures, or a global boundary with a variable impedance contrast.

In chapter 14 it was speculated that the scattering body at 2840km depth south of New Zealand may be the D'' discontinuity. However, the migration results showed that a D'' reflector coexists with a scattering body at this depth in other regions (e.g. Alaska). So it seems more likely that the scattering body seen beneath New Zealand and the Tasman sea is a different class of reflector to the D'' discontinuity (see the following section).

There is some suggestion based on the migration results beneath North Central Asia that the D'' reflector is limited to regions of fast seismic velocity. This is consistent with the reflectors absence beneath southern New Zealand where the local velocity anomaly is near zero or negative.

20.3.2 Positive scatterers near the CMB

All the regions discussed in this part of the thesis have shown the presence of positive velocity anomalies in the lowermost 100km of the mantle. Beneath Alaska and North Central Asia these scattering bodies were found beneath a D'' discontinuity, but are not limited to regions where a D'' discontinuity exists. Beneath New Zealand and the Tasman sea there is no evidence for a D'' discontinuity, but a positive near CMB reflector still exists in this region.

The positive scattering anomalies near the CMB are only intermittently present and show little or no topography at the scale lengths investigated here. However, the reflectors often coincide with high velocity anomalies in one or both of the tomographic models used here (S20RTS and SKS12WM13).

The velocity anomalies near the CMB scatterers are of the same order as those for the D'' discontinuity. Beneath Alaska and North Central Asia the reflection amplitudes are consistent with a velocity jumps of $\sim 2\%$.

20.3.3 Negative scatterers near the CMB

Negative scattering potentials are seen in the lowermost mantle beneath North Central Asia at around (2790-2840km depth). As with the other near CMB anomalies these observations of LVZ's should be interpreted as an average of the bottom 100km of the mantle, given the data frequency, migration grid spacing, and image processing steps.

The scattering amplitudes are consistent with a scattering potential/velocity anomaly of around -5%. However, this figure could be inaccurate if the scattering anomaly is produced by compositional changes rather than a thermal heterogeneity.

Importantly, in chapter 18, it was shown that these regions of negative scattering potential are not laterally extensive. Instead they form part of a complicated pattern of both positive and negative anomalies near the CMB in the North Central Asia region.

20.4 Interpretation of structures

Our results show that the D'' discontinuity is not globally present nor a continuous surface. This may be because the reflector is caused by localised thermal gradients or compositional heterogeneity. However, compositional and thermal heterogeneity may still create lateral variations in the impedance contrast across a continuous phase change.

The results also suggest the possibility that the D'' discontinuity is associated with the presence of high velocity anomalies. If this is the case, it might favour the thermal origin for the D''

boundary. However, chemical and thermal anomalies may be associated in the lower mantle. For example chemical anomalies associated with relatively young subducting slabs, would coexist with thermal anomalies.

In the three regions considered positive scattering potentials were found within the D'' region. Like the D'' reflector the positive structures near the CMB often coincide with regions of higher seismic velocity. Previous studies of positive scattering structures at these depths in other regions have interpreted them as chemical heterogeneity associated with subducting slabs (for example Braña & Helffrich (2004)). Other potential explanations could include thermal gradients, and or phase changes in this region.

The migration results show negative scattering potentials at the base of the mantle beneath North Central Asia. Their location is roughly consistent with previous studies of low-velocity zones in this region (Thomas *et al.* (2004b)). The results require that these features are not present everywhere but, along with positive anomalies, make up a complicated pattern in the bottom 100km of the mantle.

Low velocity zones could be created by the presence of partial melt (Revenaugh & Meyer (1997); Garnero *et al.* (1998); Vidale & Hedlin (1998)), alternatively they could mark the base of a high velocity region (Thomas *et al.* (2004b)). Another possibility is that positive and or negative scattering structures are created by chemical/compositional variations. Variations in the portions of: major mineral components (such as Magnesio-wustite, Mw, and perovskite, Pv), minor elements (for example Ca, and Al), and Fe-partitioning between mineral phases and partial melt, all effect elastic parameters (Trampert *et al.* (2001)). There is also the possibility of regions of metal-silicate reactions at the CMB. These possibilities are interrelated. For example the Fe-partitioning coefficient between Mw and Pv, is temperature sensitive which results in a likely trade off between thermal and compositional velocity anomalies.

20.5 Consequences for D'' and mantle processes

In section 12.5 two end member models for the dynamical behaviour of the D'' region (Lay & Garnero (2004)) were discussed. The first, the 'hybrid TCBL' model proposes that chemical and thermal heterogeneity in the D'' is provided by subducting slabs which sink to the CMB. The alternative end-member model is the 'global TCBL' where the D'' region is regarded as chemically distinct from the rest of the mantle.

Both of these models describe the range of positive and negative anomalies seen within the D'', as either thermal, or compositional heterogeneity. However, as they stand, both of these models assume the presence of a global discontinuity at the top of the D'' region. The 'hybrid TCBL' model assumes the existence of a global phase change or a radial gradient in structural fabric. The 'global TBCL' model requires the existence of a global discontinuity which separates a chemically distinct layer from the rest of the mantle.

In the preceding chapters it was shown that the D'' reflector is not continuous nor a global feature. Both of the end member models for the D'' region have to account for this observation. In the hybrid TCBL model thermal and compositional variations could change the impedance contrast across a phase boundary. Alternatively a D'' discontinuity which is caused by localised thermal gradients, changes in structural fabric, or chemical heterogeneity, could be incorporated into this model.

The global TCBL model requires a global boundary across which there is heat flux and hence a thermal gradient. However, variations in amount of partial melt in the D'' could be used to explain variations in the reflection amplitudes from the discontinuity. For example, we would expect the strongest positive increases in velocity in cold regions beneath downwellings, where the down-welling cools the D'', leading to an absence of partial melt. Beneath upwellings the D'' layer should be warmer and there should be higher incidence of partial melting, leading to a lower (or even negative) velocity jump across the D'' boundary (Lay & Garnero (2004)). So, the transition from a region of downwelling to a region of upwelling could explain the abrupt termination of the D'' reflector we observe. However, this does suggest that the flow regime switches over distances

of the order of a few hundred kilometres.

So could there be a globally stratified mantle? Yes, but if the boundary is the D'' reflector then it seems the results here require a very complicated structure to produce variations in scattering potential of the order of 100% over short distances. Alternatively, a boundary could exist outside the region explored by the migration used here. van der Hilst & Karason (1999) have argued for a much shallower boundary at around 1000km depth. The migration used here will not image a boundary at this depth.

Although the results from this study can help bridge some of the existing gaps in our knowledge of geodynamics in the D'' layer several issues still remain. Answers to questions such as: where are the regions of chemically enriched mantle? and what exactly causes the D'' discontinuity and other structures? still remain key to our understanding of deep Earth processes and mantle dynamics.

Chapter 21

Conclusions and future work

21.1 Conclusions

In this thesis different regions of the Earth's mantle were studied using different techniques. It is by integrating studies using different techniques and the results they produce that progress is made in our understanding of the mantle.

The precursors to surface reflection phases provide a vast amount of information on the upper mantle and transition zone discontinuities. Comparison of the seismological observables of discontinuities with results from mineral physics and geodynamics place important constraints on the nature of heterogeneity in the mantle. Using the precursors from the 410km discontinuity the impedance contrasts across the discontinuity was constrained, and lateral variations in the P-wave reflection coefficient identified. The observations show that the S-wave reflection coefficient varies over much shorter scale lengths than that for P-waves. This could be due to the presence of melt, water, or other chemical heterogeneities in the transition zone. Other factors such as varying temperature or mantle olivine content could also influence precursor amplitudes, but would be expected to lead to correlated variations, and so cannot explain all the observed amplitude variance.

The topography of the 410km discontinuity also suggests the presence of non-thermal hetero-

geneity in the transition zone. At very long wavelengths a moderate negative correlation exists between discontinuity topography and transition zone seismic velocity anomalies. This is consistent with the expected relationships between, the position of the $\alpha - \beta$ olivine phase boundary, seismic velocity anomalies, and mantle temperatures. However, as shorter wavelengths are included this simple relationship breaks down, and positive correlations are seen in some areas. This is interpreted as evidence for the effects of chemical heterogeneities on seismic velocities and the 410km discontinuity. The conflicting results are reconciled by suggesting that long wavelength discontinuity topography and seismic velocities reflect temperature variations, but shorter wavelength features reflect a mix of thermal and chemical heterogeneity. This means that the patterns of thermal and chemical heterogeneity must be different in the transition zone, and suggests chemical heterogeneities (on scale lengths of 1000's of km) can survive in a convecting mantle.

Our knowledge of the nature and structure of heterogeneity in the lowermost mantle is still in its infancy. It is through the application of modelling techniques, such as seismic migration, which assume the minimum about the structure in the D'' that we can begin to study this region. Here two such methods have been developed and applied to a limited but high quality dataset.

The backprojection is useful for the identification of strong scattering structures. However, the technique is limited by the use of deconvolved records and the lack of allowance for factors which affect the amplitude of scattered waves. The weighted migration does not have this problem, and together with the synthetic tests is a powerful technique for imaging structure inside the Earth.

In some areas positive reflectors are found at the base of the mantle which may be associated with high velocity anomalies in tomographic models, although disagreements between the two tomography models considered here (S20RTS, and SKS12WM13) means that this is not conclusive. In other regions a complicated mix of positive and negative scattering potentials is seen at the base of the mantle.

A D'' discontinuity is observed, but not always, and it is not a continuous reflector. In some areas structure in the D'' can be simulated with a fairly simple flat discontinuity, while in other areas the reflector changes in depth by up to 150km. The discontinuous nature of the D'' discontinuity

could be explained if the boundary is the result of discontinuous features such as localised thermal or compositional heterogeneity. Alternatively, it may be possible to explain the presence/absence of the reflector using variations in impedance contrast on a continuous feature such as a phase change or global compositional boundary.

21.2 Future work

The work in this thesis could be extended in the following ways:

The techniques developed for studying the 410km discontinuity could be applied to other upper mantle and transition zone features such as the 250 and 660km discontinuities. Although for the latter the lack of a P-wave reflection may prove an obstacle. It should also be possible to examine the 410km discontinuity topography and reflectivity in more detail using migration techniques. Future advances in discontinuity studies, mineral physics and seismic tomography may make it possible to separate the patterns of thermal and chemical heterogeneity in the transition zone at some stage. There is also room for further investigation of the effects of upper mantle anisotropy, on both reflection amplitudes and topography of mantle discontinuities.

The application of seismic migration to image structure in the lower mantle is in its early stages. Weighted migration schemes, such as the GRT based method developed here, could potentially provide detailed pictures of the entire mantle. Resolution could be greatly increased by the use of a larger dataset of 3 component seismograms. The modelling could be further enhanced by the inclusion of S-to-P, P-to-S, and P-to-P scattering. Future studies may even be able to separate variations in velocity, density, or anisotropic parameters using information from the scattered wavefield.

Appendix A

The Deconvolution Algorithm

In several sections of this thesis an iterative deconvolution was used to separate and measure minor phases more accurately. Here the method is described in more detail. The technique is similar to that first developed by Kikuchi & Kanamori (1982) and subsequently extended by Bona & Boatwright (1991), who both applied it to retrieving sub-event time histories from teleseismic recordings. Unlike the traditional methods of Wiener filtering or spectral division the technique is not limited by assumptions of a minimum phase source wavelet, and does not introduce long-period noise into the deconvolved trace.

We define the digital time series for the trace to be deconvolved as d_m , a reference pulse, g_m , and a reflectivity sequence, r_m .

$$(d_1, d_2, d_3 \dots d_{m+n-1}), (g_1, g_2, g_3 \dots g_n), (r_1, r_2, r_3 \dots r_m) \quad (\text{A.1})$$

The trace, d_m , is modelled as a convolution of the surface reflection (or reference pulse), g_m , with a reflectivity sequence (or deconvolved trace), r_n .

$$d = g * r \quad (\text{A.2})$$

$$d_m = \sum_n g_{m-n} r_n \quad (\text{A.3})$$

The objective of any deconvolution procedure is to obtain the reflectivity sequence given a time series and an estimate of the reference pulse.

The convolution (equation A.3) can be represented in matrix form:

$$\begin{pmatrix} d_1 \\ d_2 \\ d_3 \\ \vdots \\ \vdots \\ \vdots \\ \vdots \\ d_{m+n-1} \end{pmatrix} = \begin{pmatrix} g_1 & 0 & 0 & \dots & 0 & 0 \\ g_2 & g_1 & & \dots & 0 & 0 \\ g_3 & g_2 & g_1 & \dots & 0 & 0 \\ \vdots & \vdots & \vdots & & \vdots & \vdots \\ g_m & g_{m-1} & g_{m-2} & \dots & g_2 & g_1 \\ \vdots & \vdots & \vdots & & \vdots & \vdots \\ 0 & 0 & 0 & \dots & g_{n-1} & g_n \\ 0 & 0 & 0 & \dots & 0 & g_n \end{pmatrix} \begin{pmatrix} r_1 \\ r_2 \\ r_3 \\ \vdots \\ \vdots \\ \vdots \\ r_m \end{pmatrix}$$

$$\mathbf{d} = G\mathbf{r}$$

Where the trace vector $\mathbf{d}$ is obtained by multiplying the reflectivity sequence vector $\mathbf{r}$ by the matrix G . The columns of G are the reference pulse with a delay appropriate to the corresponding element of $\mathbf{r}$. In principal, it is possible to obtain a deconvolved reflectivity sequence directly by multiplying the data trace by the generalised inverse operator G^{-g} .

$$\mathbf{r} = G^{-g}\mathbf{d} \quad (\text{A.4})$$

However, in practice best results are obtained by applying equation A.4, with a limited number of positions for reflection coefficients and hence a limited number of columns in the matrix, G , rather than parameterise each time sample as a potential reflector. To choose the most suitable placing and determining spike amplitudes we use an iterative procedure.

The process begins by cross correlating the reference pulse with the trace. The spike position is placed at the maximum for the cross correlation function. We then form the matrix G , which

for the first iteration only consists of one column containing the delayed reference pulse. We then solve the rather trivial inverse problem for the spike amplitude, and place the spike in the deconvolved trace.

The fit of our deconvolved trace to the original data is tested by forming a model trace. The model trace is the convolution of the deconvolved signal and the reference pulse, $s = g * r$. If the fit of the model trace is insufficient then the model trace is subtracted from the original data, and this residual trace is used as input for the next iteration. So, the reference pulse is cross correlated with the residual trace, and a spike position is placed at the maxima of the cross correlation function. We now solve for two unknown spike amplitudes, and place these in the deconvolved trace. The process of forming a model trace and testing fit to the data is repeated and so on ...until sufficient fit is achieved.

At each iteration all the spike amplitudes are solved for simultaneously so the fit of the model trace to the data will always improve. However, larger numbers of spikes may not represent a meaningful result so it is necessary to introduce a set of conditions into the algorithm to stop the iterations.

The specifics of how to stop the deconvolution are tailored to the individual process. In part I the model trace to the data was measured using an F-test on successive variances. The iterations were stopped when the addition of a spike in the reflectivity sequence failed to improve the fit of the model trace to the data by more than 0.1%. This is more suitable for the application to a large batch of traces with varying noise levels. As the decision to stop only depends on the amount of improvement from an additional spike. In part II the decision to stop was based on the error between the data trace and the model trace. This method was chosen because in this particular situation our goal was to explain 95% of the energy in trace using the deconvolution, rather than to retrieve all the arrivals present in a trace.

Appendix B

Extra backprojection and migration results

In this appendix I give additional figures for the back projection and migration results. The figures B.1 and B.2 show global plots of the back projection results, along with the maps of the hit count (see chapter 14 for details). Figures B.3 through B.6 show results of the weighted migration in some areas not covered in part II. These results should be treated with caution as no resolution tests have been performed to assess the influence of source-receiver geometry.

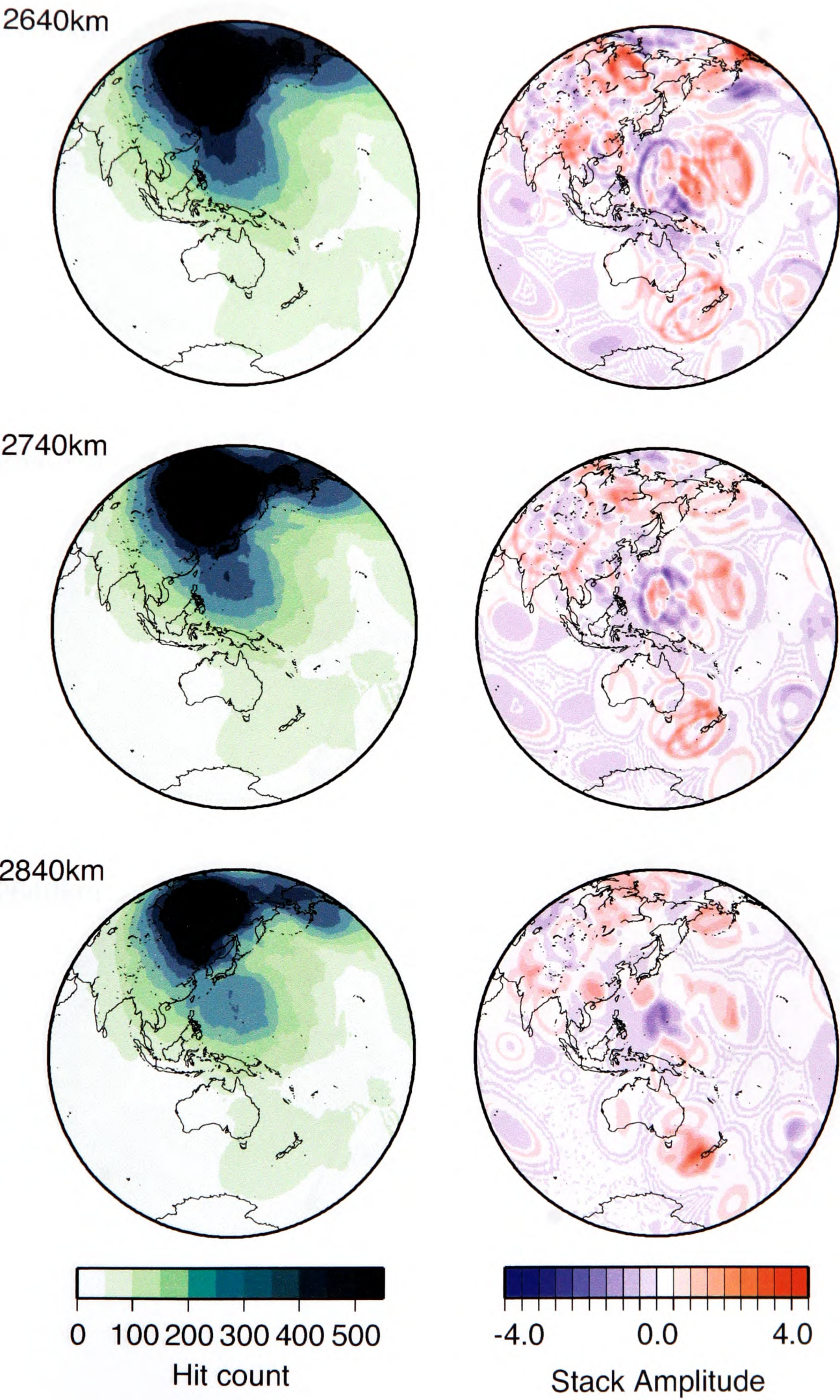


Figure B.1: Backprojection results for the eastern hemisphere.

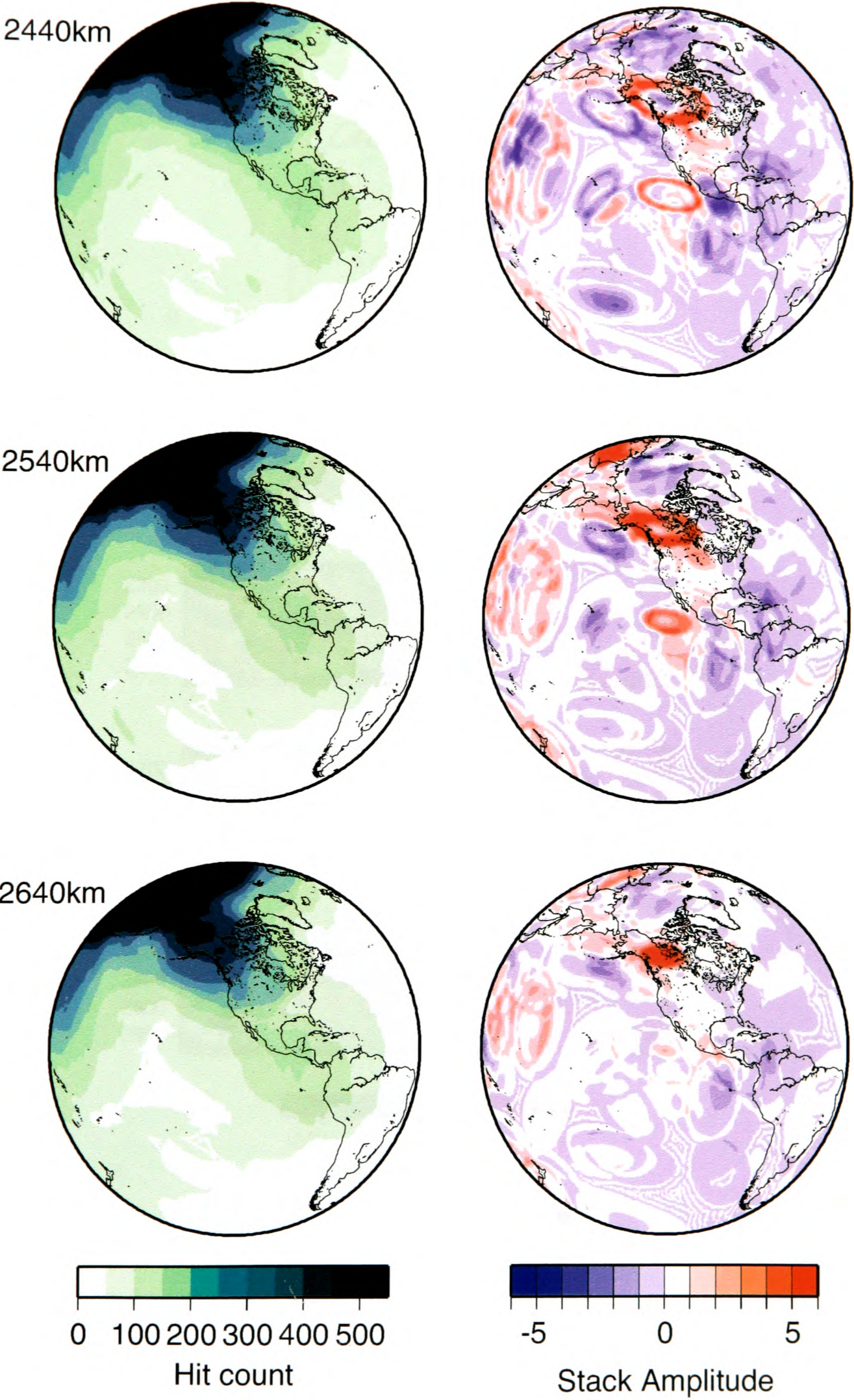


Figure B.2: Backprojection results for the western hemisphere.

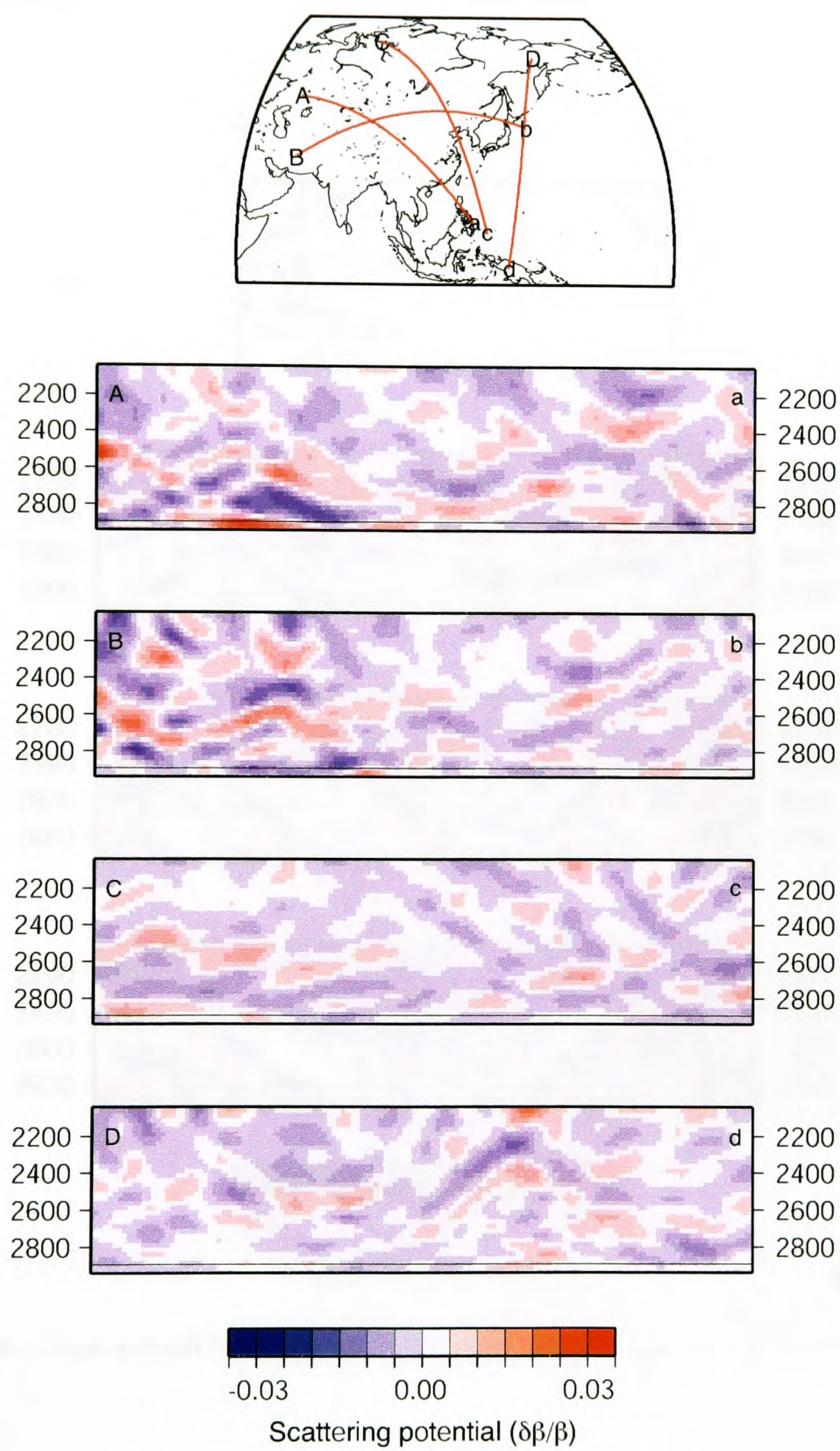


Figure B.3: Cross sections through the migration results beneath Asia.

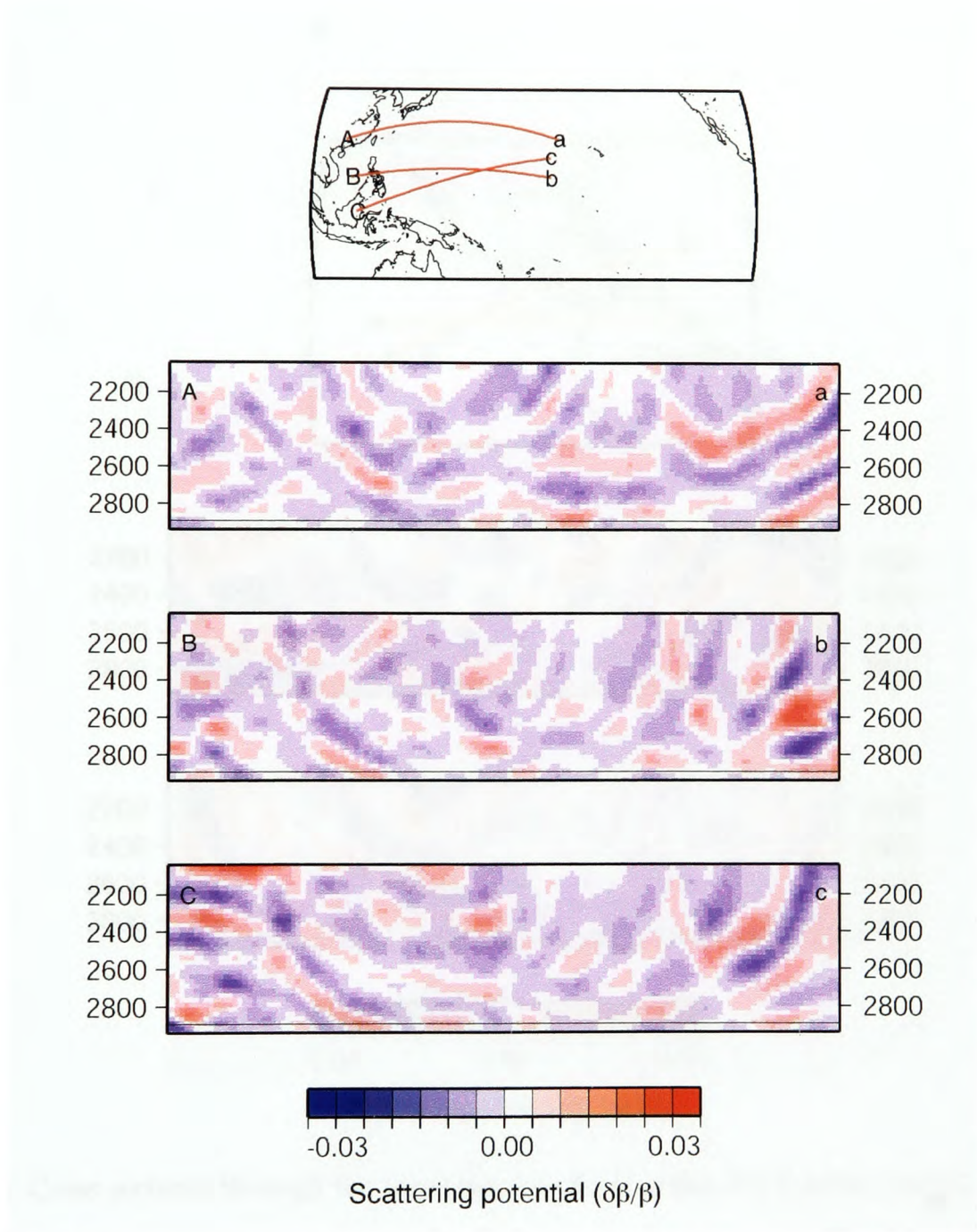


Figure B.4: Cross sections through the migration results beneath Southwest Pacific.

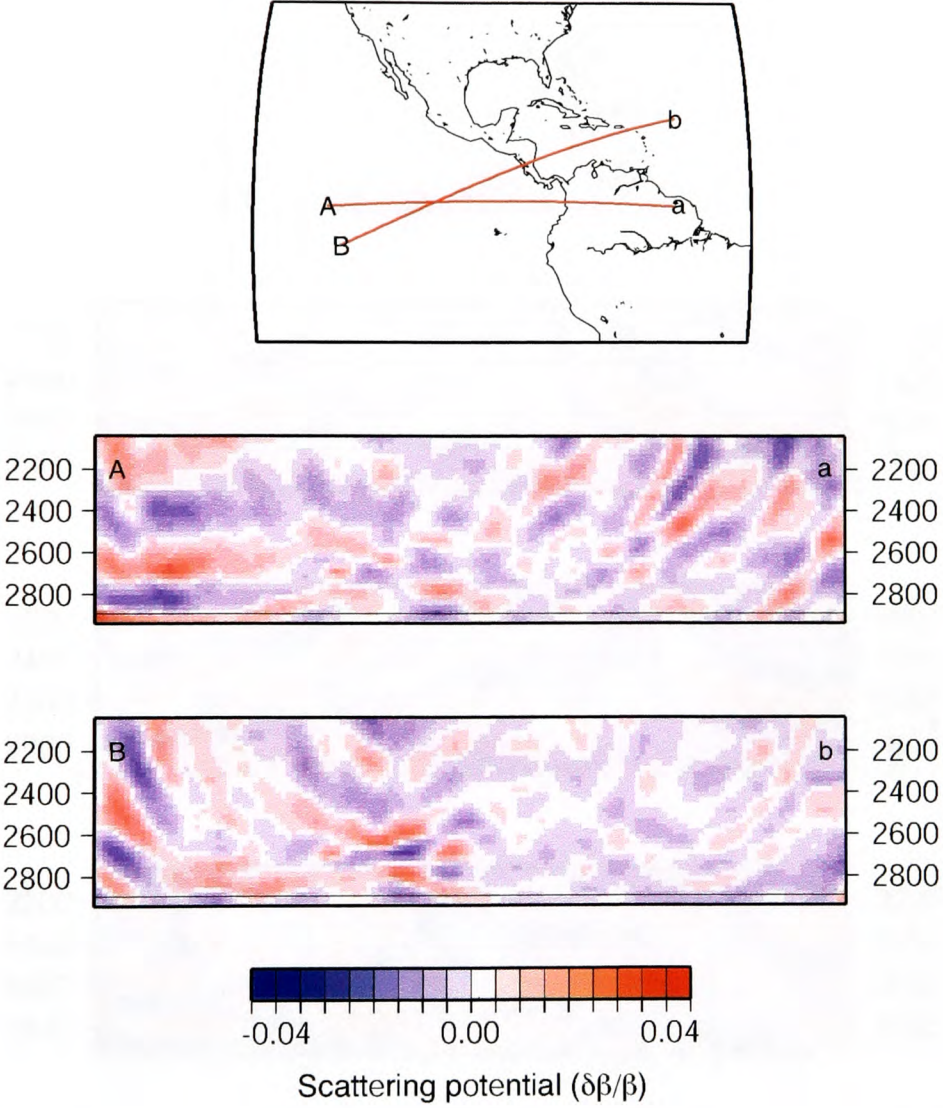


Figure B.5: Cross sections through the migration results beneath the Eastern Pacific and Central America.

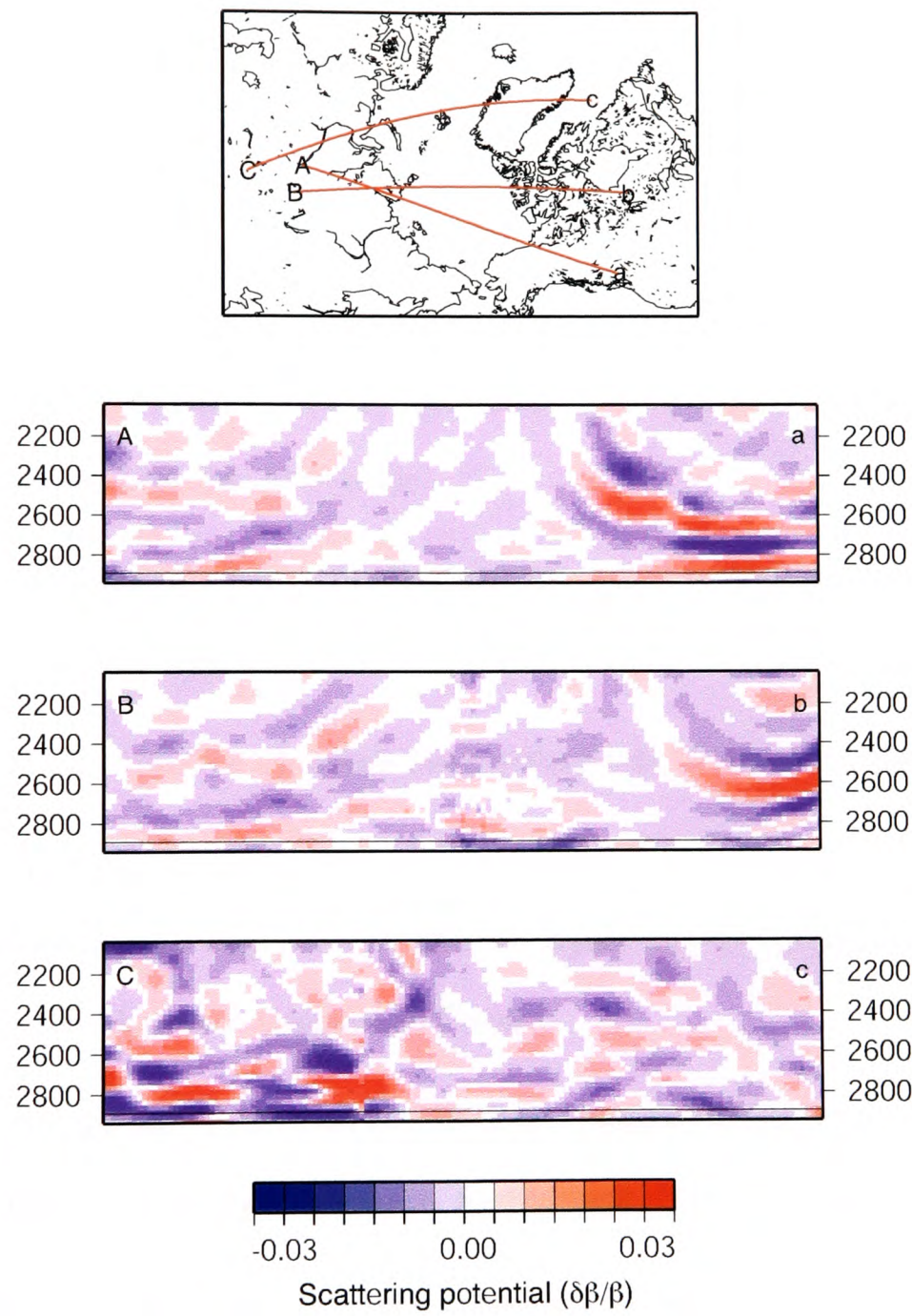


Figure B.6: Cross sections through the migration results beneath the Arctic.

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