STUDIES OF SEISMIC TOMOGRAPHY ON REGIONAL AND GLOBAL SCALE

The work described in this thesis was undertaken while I was a full-time student at the Research School of Earth Sciences, at the Australian National University, between February 1994 and June 1997. Except where mentioned in the text, the research described here is my own. No part of this thesis has been submitted for any other university or similar institution.

Sri Widiyantoro

A thesis submitted for the degree of
Doctor of Philosophy
of the
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June 1997
STATEMENT

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Sri Widiyantoro
June 1997
To my wife and children

Lika, Adri and Rena
A merry heart doeth good *like* a medicine:
but a broken spirit drieth the bones.

(Proverbs 17: 22)
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ABSTRACT

Knowledge of detailed structure of the Earth's interior is of particular importance in understanding the dynamics of the Earth and internal processes such as mantle convection. This thesis describes the delineation of three-dimensional structure in the Earth's mantle on regional and global scales. The work involves an establishment of a tomographic imaging technique, the application of the technique to the probing of regional structure of subduction zones worldwide and global deep slab structure, and an attempt to interpret the resulting tomographic images. The last part of this thesis is concerned with the extraction of bulk-sound speed for which a joint tomographic inversion procedure has been developed to invert global $P$ and $S$ travel-time data for bulk-sound speed and shear wavespeed heterogeneity models representing one of the major new results of this research.

An extensively reprocessed global travel-time data set with high-resolution earthquake locations has been used. About 7 million travel-time residuals of body waves such as direct $P$ and incorporated depth phases $pP$ and $pwP$ have been employed to improve sampling of upper-mantle structure, in particular beneath back-arc regions. In the regional inversions, the mapping of distant aspherical mantle structure was minimized by combining a high-resolution regional and a low-resolution global inversion. To bridge the gap in resolution of the images resulting from the combined inversion, global inversions of $P$ travel-time data have been performed by employing a relatively fine parameterization using uniform $2^\circ \times 2^\circ$ cells throughout the mantle. A similar global inversion was also conducted using $S$ data in which $SKS$ phases were incorporated to improve the model for the lowermost mantle by improving ray path coverage. The joint inversion for bulk-sound speed and shear wavespeed was accomplished by an iterative partitioned scheme which turned out to be efficient and suitable for a powerful workstation. All of the inversions were linearized about the $akl35$ reference model, and a combination of minimum norm and gradient damping was applied to constrain the solution. The reliability of the tomographic images were assessed carefully through test inversions of synthetic data and inversions of different subsets of data.
The results of the regional studies provide examples of the variety of styles of processes related to subducting lithosphere in detail, for example: slab penetration just into the upper mantle, slab deflection in the transition zone, slab penetration into the lower mantle with and without a change of slope, a double subduction zone with slabs subducting in opposite directions, and a slab detachment. The images of the resulting global compressional and shear mantle models depict a snapshot of convection in the mantle which provides evidence for mantle-wide convective flow. One of the important observations is that long, narrow features in the mid mantle are detected which partially can be traced to the present-day or recent subduction zones in the upper mantle. The connection between these linear structures and the long-wavelength heterogeneity on top of the core-mantle boundary remains unclear, but the global images suggest that some slabs can sink into the lowermost mantle. From the results of the joint inversion, the different sensitivity of the shear and bulk moduli to temperature provides a new insight into deep structures in the mantle; thermal heterogeneity seems to be more marked for shear wavespeed than bulk-sound speed. The bulk-sound speed model could be of particular significance in characterizing mineralogy.
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Part one: Introduction

1 General introduction

1.1 BACKGROUND

Tomographic imaging represents the reconstruction of an object from observations of physical quantities which represent the effect of the passage of some form of radiation through the object, where each quantity represents the integrated effect of interaction along a slice through the object. A major development has been in the field of medical research, where this approach forms the basis for Computer Tomographic (CT) scanning which has been successfully applied to retrieve high resolution tomograms (slice pictures) of a human body allowing e.g. the size and shape of a tumour to be measured. In medical tomographic imaging, the three-dimensional structure of the body of a patient is usually imaged in terms of spatial absorption anomalies of the intensity of X-ray energy.

The word tomography has been adopted in seismology as a synonym for methods of extracting two- or three-dimensional images of structure. Seismic tomography has been developed mostly over the last two decades. A block inversion for three-dimensional structure using seismic travel times was proposed by Aki & Lee (1976), and Aki, Christoffersson & Husebye (1977). Subsequently a variety of styles of seismic tomography have been used to produce three-dimensional images of the Earth’s interior, and have provided a major tool for improving our knowledge and understanding of the structure of the Earth. Seismic tomography has been of particular significance in retrieving information on three-dimensional structures in the upper mantle and transition zone and provides a major complement to inferences from seismicity patterns in
subduction zones, but also has a major role in studies of the lower mantle down to the core-mantle boundary.

In seismic tomographic studies the Earth's interior structure is investigated through the inversion of either seismic-wave travel-time residuals (delay times) or full waveforms to determine seismic velocities. In this thesis, a method of inversion using delay-time data has been carried out in order to investigate regional and global mantle structures. Delay-time tomography is particularly suitable for application to the delineation of mantle structure beneath regions, which contain or lie close to distributed seismicity such as subduction zones. The tomographic method comprises several major steps i.e. data preparation, model parameterization, formulation and calculation of quantities relating to the model parameters, inversion, and assessment of the solution quality (resolution tests).

The technique relies upon observations of the arrival times of seismic waves, which are determined by seismological station operators worldwide who report their arrival-time readings to the International Seismological Centre (ISC). Our ability to produce a reliable image is not only dependent on the availability of arrival-time observations for a large number of source-receiver paths which span the region of interest over a wide range of angles of incidence and azimuths, but also the data quality and accuracy of the estimates of event location. Recently, Engdahl, Van der Hilst & Buland (1997) have made an effort to improve the data quality by re-processing data reported to the ISC and the National Earthquake International Center (NEIC) from January 1964 to December 1995. The re-processing procedure includes non-linear hypocenter relocation and phase re-identification using initially the *iasp91* velocity model (Kennett & Engdahl, 1991) and subsequently the updated *ak135* model with improved S and core-phase times (Kennett *et al.* 1995). The revised data set has been used in the work in this thesis and the data refinement seems to have been crucial to allowing us to produce images of the Earth's mantle with high resolution. In contrast to conventional models employing a layered medium with uniform material properties within each layer, we have utilized a model parameterization using a local basis function in terms of non-overlapping cell...
anomaly functions, which can accommodate relatively complex geological structures. Following the pioneering work by Aki et al. (1977), we have applied a modeling approach starting with a one-dimensional Earth model consisting of homogeneous layers with fixed average velocity. By inverting the local and teleseismic seismic-wave travel-time data from seismological stations worldwide, the velocity perturbation for each cell in the model parameterization is calculated and the results are assessed by performing resolution tests.

Since the work by Aki et al. (1977), there have been a number of developments in the delay-time tomographic methods e.g. Hirahara (1977), Nolet (1985), Spakman & Nolet (1988), and Spakman (1988). In this decade alone, there are at least two distinct developments that have been made i.e. (i) Van der Hilst (1990) has further developed the method of Spakman & Nolet (1988) by incorporating $pP$ and $PP$ phases in the inversion, and (ii) Fukao et al. (1992) introduced a method of regional inversion by including an inversion for global structure to replace station corrections usually applied in regional tomographic studies. In this thesis, we to some extent attempt to combine the techniques introduced by Van der Hilst (1990) and Fukao et al. (1992). For the regional inversion we will include depth phases such as $pP$ and $pwP$ to improve sampling in the upper mantle in particular beneath back-arc regions and to get a better constraint on focal depth, and also incorporate a global inversion to reduce contaminations that may result from structure outside the study region.

The objectives of the research described in this thesis are: (1) A further development of the tomographic method of Van der Hilst (1990) by the inclusion of an inversion for global structure using a procedure similar to what for the first time applied by Fukao et al. (1992). This has involved the development of a new computer code to implement the tomographic algorithm. (2) The application of the algorithm to the probing of the structure of subduction zones worldwide with Indonesia as the focus of our study region. For Indonesia, we invert not only $P$-wave data but $S$-wave data as well. (3) The modification of the algorithm to be applied to the probing of the Earth's mantle structure on global scale using $P$ and $S$ data. (4) The extraction of bulk-sound speeds from
knowledge of the resulting global $P$ and $S$ models. Further, we have developed a joint inversion procedure to invert $P$ and $S$ travel-time data in order to construct global bulk-sound and shear velocity heterogeneity models which provide a different way to look at deep structures in the Earth's mantle.

From the results of this study, we aim to improve our knowledge of the Earth's interior, in particular we attempt to interpret our results to describe regional and global mantle structure, and the tectonic implications of the tomographic images. We also hope to get a better understanding of the regional evolution of lithospheric slabs e.g. beneath Indonesia and global mantle dynamics from the knowledge provided by the images.

1.2 THESIS SCOPE

In this thesis, our initial target is to invert delay-time data for subduction zone structure beneath the Indonesian island arc. We use the resulting tomographic images to examine the complex structure and evolution of the deep slab beneath the study region and to discuss the tectonic implications of the images. We have not only inverted $P$-wave travel-time residuals, but also the $S$-wave travel-time data in the first attempt to conduct $S$-wave delay-time tomography for the mantle structure beneath Indonesia. Despite the higher noise level in the observations of $S$ travel times, the shear-wave model is in good agreement with the images based on the $P$-wave data.

The inversion procedure has been established for travel-time data for Indonesia exploiting the reprocessed global data and has produced good results for the complex slab structure beneath the Indonesian archipelago. Following these encouraging results, we have applied a similar procedure of inversion of $P$ delay-time data for mantle structure beneath nine different subduction zones and have produced tomographic images of the subduction zones worldwide with about the same resolution level. Here our purpose is focused on the delineation of the structure of subducted slabs; detailed interpretations of all the resulting tomographic images are beyond the scope of this thesis.
In each of the regional inversions, we have incorporated a global inversion in which we have discretized the whole mantle with small cells (1° x 1°) inside the study area and bigger cells (5° x 5°) in the mantle elsewhere. As a by-product of each of these regional studies, we have obtained a global P velocity model with lower resolution than the regional model as a result of the nature of the model parameterization employed.

To provide a bridge in resolution between the regional and the coarse global images, we have performed a full mantle inversion of P delay-time data using 2° x 2° cells throughout the global model which represents a useful attempt to apply such a fine parameterization to produce a high-resolution whole mantle P velocity model. The reprocessed global data set (Engdahl et al. 1997) contains not only P but also S data, and with the ak135 model (Kennett et al. 1995) a comparable level of representation of the travel times is achieved for each wave type. We have therefore applied a similar procedure of whole mantle inversion method to that employed to produce the high-resolution global P wavespeed model to construct a model of shear velocity heterogeneity in the entire mantle. We have tried to restrict attention to the first arriving "S" arrival and have therefore incorporated data for the core phase SKS in the inversion in order to improve our S model for the lowermost mantle.

The last part of this research is concerned with the development of an algorithm for a joint inversion of P and S travel-time data to produce global bulk-sound and shear velocity models. We have jointly inverted P and S delay-time data using more or less the same P and S ray coverage. In this way, we hope to achieve about the same resolution of the resulting models although we lose some of the ray sampling. To provide a comparison, we also have directly extracted a global bulk-sound velocity model from the knowledge of our new global compressional and shear wavespeed models. The resulting bulk-sound velocity models may be of particular significance in characterizing mineralogy and geodynamic processes.
1.3 THESIS STRUCTURE

In the second chapter, we give a brief description of delay times in seismic tomography and the inversion technique used in this study, and address the establishment of a technique for inverting regional and global mantle structure simultaneously which follows the technique introduced by Fukao et al. (1992). We also describe the ray tracing technique employed to construct matrix elements to be used in the inversion.

In chapter 3, we present the results of an application of the established method to the delineation of mantle structure beneath Indonesia. We discuss the structure and evolution of lithospheric slab beneath the entire Indonesian region in detail. The material in this chapter has been published as "Structure and Evolution of Lithospheric Slab Beneath the Sunda Arc, Indonesia" in Science (271, 1566-1570) and "Mantle Structure Beneath Indonesia Inferred from High-Resolution Tomographic Imaging" (Geophys. J. Int., in press). We also demonstrate the effectiveness of the use of depth phases such as *pP* and *pwP* in sampling the uppermost mantle beneath the Indonesian back-arc regions.

In chapter 4, we present the highlights of results of the application of the method to retrieve the regional structure of nine different subduction zones worldwide. These zones include Europe, India including Himalaya, northwest Pacific, Tonga-Kermadec, the Aleutian region, north America, central America including Caribbean, south America and the Scotia plate.

The results of global imaging are presented in chapter 5 through to chapter 8. Chapter 5 contains a presentation of an average global *P* velocity model produced by averaging ten different low-resolution global *P* models resulting from the incorporation of a global inversion in each of the regional inversions presented in chapters 3 and 4. We present the high-resolution global model for compressional (*P*) velocity model in chapter 6 and the high-resolution global model for shear (*S*) velocity in chapter 7. We will show some cross sections for *P* and *S* from the same locations in order to investigate the similarities and differences between the *P* and *S* models. Parts of the global *P* model presented in chapter 6 have appeared in Nature (386, 578-584) and GSA Today (7, No.
4. 1-7), and most of chapter 7 was submitted to *Physics of the Earth and Planetary Interiors* in April 1997. The global bulk-sound velocity models are presented in chapter 8, where we describe the development of an algorithm for a joint inversion of $P$ and $S$ travel-time data to produce bulk-sound and shear velocity heterogeneity models (see Appendix 8A). In this chapter, we also make a comparison between a bulk-sound model directly extracted from the $P$ and $S$ models presented in the two previous chapters and that derived from the joint inversion of $P$ and $S$ data using more or less the same $P$ and $S$ ray coverage.

In the last chapter, we globally synthesize the results of the regional studies and summarize the global inversion results. Finally, we address some aspects arising from this research that suggest further work.

2.1 DELAY TIME

In a delay-time tomographic inversion, delay times are the difference between observed and calculated travel times. They are inverted for velocity variations in a three-dimensional volume relative to a radially stratified reference or velocity model. Delay times are also referred to as "departure" times in ray-tracing and the inversion technique used.

Following Spetzler & Geller (1976), we can write the travel time $T$ between event $i$ and receiver $j$ as

$$T_{ij} = f(x_{ij}, d_{ij})$$

where $f$ is the travel-time function, $x_{ij}$ is a function of position and $d_{ij}$ is the ray-segment length of the integration path $ij$, which depends on the ray-trace earthquake location and the three-dimensional Earth structure. The problem is non-linear.
Part two: Imaging technique

2 Seismic delay-time tomography

In this chapter, we present the tomographic method used in this study including an overview of delay times as the input data in seismic tomography, model parameterization, the system of tomographic equations, ray tracing and the inversion technique used.

2.1 DELAY TIME

In a delay-time tomographic inversion, delay times (the difference between observed and calculated arrival times), are inverted for aspherical variations in seismic-wave speed relative to a radially stratified reference velocity model. Delay times are also referred to as either travel-time residuals or residual times.

Following Spencer & Gubbins (1980), we can write the travel time \(T_{ij}\) between event i and station j as follows:

\[
T_{ij} = \int_{l_{ij}(3D)} s(r) \, dl
\]

(2.1)

where slowness \(s(r)\) is the reciprocal of velocity as a function of position and \(dl\) is the ray-segment length of the integration path \(L_{ij(3D)}\) which depends on the unknown earthquake location and the three-dimensional Earth structure. Thus the problem contained in (2.1) is non-linear.
For each $T_{ij}$ we have an observed travel time ($O_{ij}$) used to calculate a delay time ($\delta t_{ij}$), that is

$$\delta t_{ij} = O_{ij} - T_{ij} = \int_{L_{ij}(0)} s_0(r) \, dl_0 - \int_{L_{ij}(D)} s(r) \, dl$$

(2.2)

where $L_{ij}(0)$ is the ray path computed in the reference Earth model and the subscript $o$ denotes the corresponding reference model quantities. In seismic tomography, the delay time can be interpreted as

$$\delta t_{ij} = \int L_{ij}(0) + dT_{ij} + \epsilon$$

(2.3)

where $\Delta s(r)$ is the slowness perturbation, $dT_{ij}$ is the small change in the calculated travel time due to a small change in hypocentre location and the $\epsilon$ term contains errors in the data, and the second and higher order terms of approximations (Spakman 1988).

In linearized tomography, we assume that the reference Earth model is close enough to the real Earth. Therefore, we may apply Fermat’s principle which states that for a fixed velocity field and fixed end points changes in travel time caused by changes in ray path are negligible to a first order approximation (due to the stationarity of the ray path). In the linearization we neglect the second and higher order terms, and substitute the unknown ray path $L_{ij}(D)$ in the real Earth by $L_{ij}(0)$.

The small change $dT_{ij}$ containing the partial derivatives of the travel time $T_{ij}$ can be expressed in the form (Gubbins 1990):

$$dT_{ij} = (\partial T_{ij} / \partial t_{0,i}) dt_{0,i} + (\partial T_{ij} / \partial h_i) dh_i + (\partial T_{ij} / \partial \theta_i) d\theta_i + (\partial T_{ij} / \partial \phi_i) d\phi_i$$

(2.4)

where,

$$\partial T_{ij} / \partial t_{0,i} = 1.0$$

(2.4.1)

$$\partial T_{ij} / \partial h_i = - (\eta_i^2 - p_{ij}^2)^{1/2} / r_i$$

(2.4.2)
\[
\frac{\partial T_j}{\partial \theta_i} = -p_j \cos Z_j
\]
(2.4.3)

\[
\frac{\partial T_j}{\partial \phi_i} = p_j \sin Z_j \sin \theta_i
\]
(2.4.4)

with \( t_{0,i} \) the origin time, \( h_i \) the focal depth, \( (\theta_i, \phi_i) \) the colatitude and longitude of the epicentre, \( \eta_i = r_i / v(r_i) \), \( r_i \) the radius from the Earth's centre to the hypocentre and \( v(r_i) \) the velocity as a function of radius, \( p_j \) the ray parameter, and \( Z_j \) the azimuth. The velocity \( v(r_i) \) is taken from the reference velocity model used, while \( p_j, Z_j \) and \( \theta_i \) are provided along with the phase data used in the inversions (Engdahl, Van der Hilst & Buland 1997).

As described by Spakman (1991), delay times generally contain errors which come from event mislocations and origin time error caused by: (i) reading and phase mispicking errors, (ii) systematic errors arising from single station or instrument effects, (iii) non-uniform and/or insufficient azimuthal distribution of seismological stations, (iv) the Earth’s lateral heterogeneity, (v) systematic deviations of the laterally averaged velocity structure of the Earth from the reference model and (vi) bias which may arise from the event location procedure caused by a priori assumptions regarding the statistical properties of delay times.

Following Spakman (1988), the slowness perturbations \( \Delta s \) can be converted into velocity perturbations \( \Delta v \) using the following formula

\[
\Delta s = s - s_0 = 1 / (v_0 + \Delta v) - 1 / v_0
\]
(2.5)

where \( s \) is the true slowness and the subscript \( o \) denotes the corresponding reference model quantities. For small deviations in \( \Delta v \), (2.4) can be rewritten as

\[
\Delta v = -\Delta s \quad v_0^2
\]
(2.6)
We will present and plot our results in terms of the velocity perturbations.

2.2 MODEL PARAMETERIZATION

Following Aki et al. (1977), we have employed a model parameterization using non-overlapping cells. We discretized the whole mantle into a large number of cells with different sizes. In regional studies, we employed a subdivision of the mantle volume inside the study area with finer cells than outside.

As a result of the parameterization, we dealt with various cell sizes where bigger cells tend to absorb bigger anomalies than the smaller ones because of the larger total path length. To avoid this problem, we have followed Nolet (1987a) and used a volume scaling for each cell. Assuming that the total ray-segment length in a cell is proportional to the cell volume, ray-segment lengths both inside and outside the study area are scaled by the square root of the cell volume. This scaling is also applied during the relocation of events during inversion, in which the relocation coefficients are scaled by the square root of the average volume. The solution vector resulting from the inversion then does not contain the slowness perturbations ($\Delta s$) represented in (2.4) anymore, but $\Delta \xi$ which relates to $\Delta s$ as follows

$$\Delta s_j = \Delta \xi_j / V_j^{1/2}$$

where $V$ is the volume and subscript $j$ denotes the $j$-th cell (see Spakman 1988).

Besides the above volume scaling, we also have included the application of a weighting scheme to the ray-segment lengths and relocation coefficients in order to tune the inversion to get a clear image.

2.3 TOMOGRAPHIC SYSTEM OF EQUATIONS
For the linearized tomographic problem, we set up a system of equation for least squares inversion in matrix notation as follows:

\[ Ax = \delta t \]  

(2.8)

where \( x \) is the solution vector containing the unknown and \( \delta t \) is the delay-time data vector. The matrix \( A \) projecting the model space to the data space usually contains e.g. ray-segment lengths and hypocenter relocation coefficients.

In tomographic studies, (2.8) is usually over-determined in the sense that there are more data used in constructing the linear equations than unknowns. Thus matrix \( A \) is not square and in order to get a square matrix, matrix \( A \) is multiplied by \( A^T \) leading to a set of equations

\[ (A^T A)x = (A^T \delta t) \]  

(2.9)

which is called the normal equations of the linear least-squares problem. Note that \( A^T \) in (2.9) denotes the transpose of the matrix \( A \). At the same time (2.8) and (2.9) are under-determined, because many rays sample more or less the same mantle region. Thus the associated equations are not independent and the matrix \( A \) is usually sparse. In consequence not all unknowns can be resolved. Also, as mentioned in section 2.1, since there are errors in the data, we do not expect that the equations will be consistent. Therefore, there is no unique solution and we seek a minimum norm solution by minimising \( \| x - (A^T A)^{-1} A^T \delta t \|_2^2 \). It is convenient to take the Euclidean norm (\( \xi=2 \)), because then we deal with the well known least squares problem.

In regional tomographic studies, we usually consider only a fraction of the whole Earth's mantle volume and apply station corrections in order to suppress contaminating signals from outside the mantle volume under study. In this case, we can rewrite (2.8) as
where matrix $A$ contains $L_{in}$, $H$ and $R$, which are ray-segment lengths in cells in the study region, station-correction and event relocation coefficients, respectively. The $x$ vector contains the solution vectors of $L_{in}$, $H$ and $R$ i.e. $s_{in}$, $h$ and $r$.

Following Fukao et al. (1992), we have developed a method to incorporate an inversion for global structure at the same time as the regional inversion. In this method, we replace $H$ by $L_{out}$ which contains ray-segment lengths elsewhere in the mantle:

$$
(L_{in}/H/R) \begin{pmatrix}
     s_{in} \\
     h \\
     r
\end{pmatrix} = \delta t
$$

(2.10)

Here, $s_{in}$ and $s_{out}$ are the slowness perturbations inside and outside the volume under study, respectively. Equation (2.11) shows that the inversion allows for the simultaneous determination of velocity perturbations in the study area and the whole mantle, and also event relocation upon inversion. We have tested the use of station corrections and it has been confirmed that the two methods give qualitatively the same result. The replacement of $H$ by $L_{out}$, however, gives better internal consistency and produces a global model as well. In computing the relocation coefficients ($R$), we used the travel time partial derivatives given in (2.6).

### 2.4 RAY TRACING

In order to compute $L_{in}$ and $L_{out}$ in (2.11) for the construction of matrix $A$, we have traced seismic rays for given hypocentre-station pairs. In this study, we have used a standard ray tracing scheme described by Bullen & Bolt (1985) which utilizes Mohorovicic's law

$$
v = ad^b
$$

(2.12)
Figure 2.1. The left panel is the \textit{ak135} reference velocity model (Kennett \textit{et al.} 1995) used in this study. The right panel is a plot of \textit{P} ray paths displaying the depth intervals used in the ray tracing i.e. 1 km close to the ray-bottoming point and 10 km elsewhere.

where \(a\) and \(b\) are constants, \(d\) is the distance between the Earth's centre to any point \(R\) in the model and \(v\) is the wave speed at \(R\). We assume that the transmission speed of seismic body waves in the Earth's interior only depends on \(d\) and is symmetric about the centre of the Earth.

The Mohorovicic law is adequate to provide an approximation to the actual velocity variation over many depth ranges in the Earth. We use a depth interval of 10 km which is smaller than the thinnest layer used in the model parameterization. Near the bottoming point, however, the rays propagate almost horizontally, and to increase the
accuracy we take a finer depth interval of 1 km (see Fig. 2.1). The left panel in Fig. 2.1 is the velocity model used in this study, that is the *akl35* model derived by Kennett *et al.* (1995).

![Figure 2.1](image1.png)

**Figure 2.1.** The left panel shows the velocity model (*akl35*) used in this study. The right panel illustrates the ray paths for both *P* and *sP* waves, highlighting the paths from various sources to receivers.

For the next step, we linearly interpolate between each pair of two successive points along the ray to find the intersections of the ray path with cell boundaries and to compute the ray-segment length in each cell. This linear interpolation application is
acceptable, because in our application of seismic tomography we assume that all the
properties of the Earth within a cell is homogeneous.

The important parameter in ray tracing is called the ray parameter \( p \) which is
constant along the entire ray.

\[
p = d \sin \theta_o / v(d)
\]  

(2.13)

where \( \theta_o \) is the incident angle. It is convenient that we do not need to compute the ray
parameters, since they are provided in the data set. We prefer to trace the seismic ray
starting from a station to a hypocentre by calculating its back azimuth first. This has the
advantage that we will never miss the known point, that is the station coordinate, since
the hypocentre is actually unknown. An example of a plot of several \( P \) and \( pP \) ray paths
is given in Fig. 2.2. These ray paths already include linear interpolation.

2.5 INVERSION TECHNIQUE: LSQR

To solve the inversion problem to extract the vector \( x \), we have used an iterative
Conjugate Gradient (CG) method that is the LSQR method by Paige & Saunders (1982)
introduced to seismic tomography by Nolet (1985). The LSQR method is quite similar to
the various inversion techniques based on singular value decomposition (SVD). In an
SVD direct inversion method, the solution is built up in a \( p \)-dimensional subspace of the
model space, which is spanned by the eigenvectors of \( A^T A \) belonging to the largest
eigenvalues \( \lambda_1, \ldots, \lambda_p \). In tomography with systems of large size, it is not practical to
compute the eigenvectors of \( A^T A \) in a reasonable amount of CPU time. Paige &
Saunders (1982) introduced an LSQR method in which the explicit use of \( A^T A \) is
avoided. A Ratfor version of their algorithm given by Nolet (1987a) is used in this
study. The LSQR method seeks the \( x \) that minimizes \( |Ax - \delta t| \) with subject to
minimizing \( x^T x \).
Upon inversion we used two kinds of explicit damping i.e. minimum norm and gradient damping:

(i) The first damping (minimum norm) biases toward zero slowness perturbations from an average model and suppresses the amplitude in regions of poor sampling (Spakman & Nolet, 1988). In the present of this damping, the LSQR algorithm iteratively approximates the minimum norm least-squares solution of (2.8) by solving $x$ from

$$\min [(Ax - \delta t)^T (Ax - \delta t) + \lambda^2 x^T x]$$

where $\lambda^2$ is a damping parameter.

(ii) The application of the second (gradient) damping, which biases toward a smooth model, was conducted by using a roughening matrix described by Nolet (1987a). The roughening operator takes the difference of a cell with all its neighbours. Imposing the condition

$$\min \left[ \sum_{i=1}^{M} \frac{1}{2N_i} \sum_{j=1}^{N_i} (x_i - x_j)^2 \right]$$

where $N_i$ is the number of neighbours of cell $i$ and $M$ is the number of cells, this adds to the linear system $M$ more equations

$$x_i - \sum_{j=1}^{N_i} x_j = 0 \quad (i = 1, \ldots, M)$$

(See Nolet, 1987a, for a detailed description). Our purpose of applying the gradient damping is to get a relatively smooth model so that it can be used for other studies e.g. studies on the lateral strength of slabs (see Moresi & Gumis 1996).
3 Structure and evolution of lithospheric slab beneath Indonesia

3.1 INTRODUCTION

The tectonic setting of the Indonesian archipelago in southeastern Eurasia is determined by the complex interaction of several major and minor plates (Fig. 3.1). The junction of island arcs includes the Sunda arc, the Banda arc, the Sangihe arc to the west of the Molucca Collision Zone (MCZ) and the Halmahera arc to the east. The Sunda arc, from northwestern Sumatera to Flores, marks the subduction of the Indo-Australian plate beneath the Eurasian plate. The Banda arc to the east of the Sunda arc displays a strong curve in map view. The strike of the Banda arc used to be east-west, just like the eastern part of the Sunda arc, but was twisted counter clockwise in the Pliocene due to the combination of the collision with the northward moving Australian continent, the counterclockwise rotation of New Guinea, and the westward thrust along the Sorong fault system. These tectonic events interrupted the regular growth of the eastern Indonesian arc (Katili 1975). The Sorong fault, striking nearly east-west from the Bird’s Head of Irian Jaya to the east of central Sulawesi, marks the boundary between Banda and Halmahera. The active collision between Halmahera and Sangihe, the only known example of two colliding arcs, has caused the two opposing subduction directions of the Molucca Sea plate (Hamilton 1979; Cardwell et al. 1980; and Hall 1987).

The convergence rate of the Indo-Australian and Eurasian plates increases from about 60 mm/yr near Sumatera to 78 mm/yr in the easternmost part of the Sunda arc, with
an azimuth N20°E predicted by the RM2 plate model (Minster & Jordan 1978). A recent study based on the first geodetic measurement of convergence across the Java trench (Tregoning et al. 1994) suggests that convergence of $67\pm7$ mm/yr is measured between Christmas Island, southwest of Java, and west Java in a direction of N11°E±4°. This is similar to the relative plate velocity of $71\pm2$ mm/yr between Australia and Eurasia according to NUVEL-1 (DeMets et al. 1990). In the east, the slip rate increases from $\sim80$ mm/yr near the Philippines to $\sim100$ mm/yr near Halmahera (Ranken et al. 1984). The age of the subducted oceanic lithosphere increases eastward from about 50 - 90 million years ago (Ma) below Sumatera to 140 - 160 Ma near Flores. The age of the Molucca Sea lithosphere, however, is unknown.

Given this tectonic setting, it can be expected that the junction of island arcs overlies a strongly heterogeneous mantle. This structural complexity is partly evident from regional seismicity. The character of subduction related seismicity changes abruptly from Sumatera to Java. With the exception of some small events in the southeast, seismicity does not exceed a depth of about 300 km beneath Sumatera, but earthquakes occur at depths of up to 670 km below Java, Banda, and Sangihe. A pronounced seismic gap exists below Java in a depth interval of about 300-500 km (e.g. Newcomb & McCann 1987; Wortel & Vlaar 1988). The Wadati-Benioff zone is steep below Java and dips rather gently beneath the easternmost Banda arc. Beneath the MCZ, Wadati-Benioff zones dip to the west and to the east below Sangihe and Halmahera, respectively (Cardwell et al. 1980).

The aim of this study is to explore the deep structure of slabs beneath the region by means of tomographic imaging. Until the late 1980's investigations of mantle structure beneath the Indonesian region had mostly been based on volcanism, marine geophysical data, and seismicity (e.g. Katili 1975; Cardwell & Isacks 1978; Hamilton 1979; and Hall 1987), and lower mantle structure had not yet been revealed. Previous seismic tomographic studies of mantle structure below the region are due to Fukao et al. (1992) and Puspito et al. (1993). Fukao and co-workers (1992) argued that the slab beneath the eastern Sunda arc penetrates directly into the lower mantle, where it deflects
in the uppermost lower mantle and sinks vertically to a depth of at least 1200 km. Puspito et al. (1993) also detected slab penetration to the lower mantle beneath Java and, in addition, presented evidence for two opposing subducted slabs of the Molucca Sea plate. These tomographic studies did not use depth phases such as pP and pwP. In this study we included those depth phases in order to improve the sampling of structure, especially in the upper mantle beneath the back-arc regions, and to constrain earthquake focal depth (Van der Hilst et al. 1991; Van der Hilst & Engdahl 1991 & 1992). Following Fukao et al. (1992), we incorporated a global inversion in this regional study to minimize the mapping of signal from structure outside the study region. Our results are generally in good agreement with the results of previous works, but reveal the structural complexity in more detail, in particular in the upper mantle and transition zone.

The tomographic images confirm previous evidence for regional slab penetration into the upper part of the lower-mantle but suggest a more complex trajectory of mantle flow beneath Indonesia. The implied lateral variation in the shape of the upper mantle slab is generally in good agreement with inferences from seismicity. The data reveal a seismic anomaly beneath the Sunda arc, which is detected to at least 1500 km depth and forms the eastern end of a deep anomaly associated with the past subduction of lithosphere of the Mesozoic Tethys ocean. Below Java the lithospheric slab is probably continuous from the Earth's surface to the lower mantle but beneath Sumatera the deep slab seems to be detached from the seismogenic slab. The high-resolution images suggest that the subducted slab is deflected in the transition zone beneath the Banda arc and that the descending slabs form a spoon-shaped structure. North of the Banda Sea we detect the two opposing subducted slabs of the Molucca Sea plate; this is consistent with previous studies but our images suggest that the westward subduction has a significantly steeper dip than the eastward subduction. Slab structure beneath Banda and Molucca Sea is explained in the context of continent-arc collision between Australia and the southeast Indonesian Island arc, and the arc-arc collision between Sangihe and Halmahera, respectively.
We also inverted travel-time residuals of S phases, a novelty for this region, to investigate the differences and similarities with structure as revealed by the inversion of P data, which can help to provide a better understanding of the physical origin of the wavespeed anomalies. Routinely processed S data are generally regarded to be more noisy than P data. For P and also for S we used the carefully re-processed data set by Engdahl, Van der Hilst & Buland (1997). The S residuals have larger uncertainties than the P data but thanks to the re-processing they do not have some of the problems inherent in ISC S data. In contrast to ISC procedures, Engdahl et al. (1997) used S phases in the initial source location and a reference velocity model (ak135 by Kennett et al. 1995) that is more appropriate for S than the reference model used by the ISC. Results of the inversion of travel-time residuals of direct S phases largely confirm the inferences from the P-wave data even though less detail is resolved.

Most parts of this chapter are already published in Widiyantoro & Van der Hilst (1996) and in Widiyantoro & Van der Hilst (in press). In the first paper we focus on the northward subduction of the complex Indo-Australian plate along the Sunda arc, while in the second paper we present results of the resolution tests used to substantiate the arguments in the first paper. In addition, we discuss the structure beneath the Banda, MCZ, and Philippine Island regions as inferred from the tomographic images, and present some results of the inversion of S data.
Figure 3.1. Map of the study region. Dashed lines indicate major plate boundaries after NUVEL-1 by DeMets et al. (1990). Solid line to the North of Banda Sea depicts the Sorong fault. Arrows indicate the direction of plate motion relative to the Eurasian plate. Abbreviations used: MCZ, Molucca Collision Zone; PNG, Papua New Guinea; SR, Sorong; Hlm, Halmahera; and Sgh, Sangihe.
3.2 METHOD

In this section we describe the tomographic method used, including the model parameterization used for regional and global inversions, the system of linear equations, and the resolution tests used to assess image quality.

3.2.1 Model parameterization

We discretized the entire mantle by means of local basis functions in the form of a uniform grid of constant velocity cells of $5^\circ \times 5^\circ$, but in the study area we used a finer grid of $1^\circ \times 1^\circ$ in order to allow the resolution of relatively small-scale structure (Fig. 3.2a). The cell-layer divisions inside and outside the volume under study are displayed in Fig. 3.2 (b). We discretized the mantle volume under study, to a depth of 1600 km, into 19 layers with layer thickness ranging from 35 km in the uppermost layer to 200 km for the layer centred at 1500 km. For the global inversion, we used 16 layers with an average-layer thickness of about 180 km (Table 3.1). The number of small and big cells is 43,225 and 41,472, respectively.

Upon inversion we also accounted for the effect on the travel-time residuals of source mislocation due to three-dimensional structure. Instead of solving for four relocation parameters for each earthquake we used the same partial derivatives for all earthquakes located within an event cluster. This reduced the number of unknowns considerably. We used 5916 event-clusters with a dimension of $1^\circ \times 1^\circ \times 50$ km, adding almost 23,664 parameters [see equation (2.4)] to the 84,697 unknowns associated with the constant slowness blocks. The total number of unknowns to be solved for by the tomographic inversion is thus almost 110,000.

For stations and/or events located outside the mantle volume under study we combined ray paths from event to station clusters into a single ray, commonly referred to as the summary ray, in order to reduce the uneven sampling of mantle structure by ray paths, the dimension of the matrices involved in the inversions (and thus computer memory requirements), and the CPU time for ray tracing. The datum (residual time)
assigned to the summary ray was the median of all data considered for that summary ray. The number of rays that contributed to the summary ray was not restricted. For stations and events inside the study area, however, we used individual rays to optimize the sampling.

Figure 3.2. Model Parameterization. (a) The grid of non-overlapping blocks of horizontal dimension of 1° × 1° and 5° × 5° depicts the parameterization used inside and outside the study region, respectively. (b) Cell layer divisions inside and outside the study region; the y-axis is exaggerated seven times.
Table 3.1. Information of the cell layer division *inside* and *outside* the volume under study, and the corresponding layer-average velocity of *ak135* model (Kennett *et al.*, 1995).

<table>
<thead>
<tr>
<th>Layer no:</th>
<th>Depth range (km):</th>
<th>Average velocity (km/s):</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inside:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.</td>
<td>0-35</td>
<td>6.10</td>
</tr>
<tr>
<td>2.</td>
<td>35-70</td>
<td>8.04</td>
</tr>
<tr>
<td>3.</td>
<td>70-110</td>
<td>8.05</td>
</tr>
<tr>
<td>4.</td>
<td>110-160</td>
<td>8.10</td>
</tr>
<tr>
<td>5.</td>
<td>160-220</td>
<td>8.25</td>
</tr>
<tr>
<td>6.</td>
<td>220-280</td>
<td>8.45</td>
</tr>
<tr>
<td>7.</td>
<td>280-340</td>
<td>8.67</td>
</tr>
<tr>
<td>8.</td>
<td>340-410</td>
<td>8.90</td>
</tr>
<tr>
<td>9.</td>
<td>410-490</td>
<td>9.49</td>
</tr>
<tr>
<td>10.</td>
<td>490-570</td>
<td>9.76</td>
</tr>
<tr>
<td>11.</td>
<td>570-660</td>
<td>10.05</td>
</tr>
<tr>
<td>12.</td>
<td>660-750</td>
<td>10.91</td>
</tr>
<tr>
<td>13.</td>
<td>750-840</td>
<td>11.12</td>
</tr>
<tr>
<td>14.</td>
<td>840-930</td>
<td>11.28</td>
</tr>
<tr>
<td>15.</td>
<td>930-1020</td>
<td>11.42</td>
</tr>
<tr>
<td>16.</td>
<td>1020-1130</td>
<td>11.58</td>
</tr>
<tr>
<td>17.</td>
<td>1130-1250</td>
<td>11.76</td>
</tr>
<tr>
<td>18.</td>
<td>1250-1400</td>
<td>11.95</td>
</tr>
<tr>
<td>19.</td>
<td>1400-1600</td>
<td>12.18</td>
</tr>
</tbody>
</table>

| Outside:  |                   |                          |
| 1.        | 0-110             | 7.43                     |
| 2.        | 110-280           | 8.27                     |
| 3.        | 280-410           | 8.80                     |
| 4.        | 410-660           | 9.78                     |
| 5.        | 660-840           | 11.01                    |
| 6.        | 840-1020          | 11.35                    |
| 7.        | 1020-1250         | 11.67                    |
| 8.        | 1250-1400         | 11.95                    |
| 9.        | 1400-1600         | 12.18                    |
| 10.       | 1600-1850         | 12.47                    |
| 11.       | 1850-2050         | 12.74                    |
| 12.       | 2050-2250         | 12.97                    |
| 13.       | 2250-2450         | 13.20                    |
| 14.       | 2450-2600         | 13.40                    |
| 15.       | 2600-2750         | 13.58                    |
| 16.       | 2750-2889         | 13.67                    |
3.2.2 Tomographic system of equations

In regional tomographic studies one usually considers only a fraction of the whole Earth's mantle volume and sets up a system of linear equations for least squares inversion as explained in (2.10) of the previous chapter. In this regional study we incorporated a global inversion and employed (2.11). Test inversions based on either (2.10) or (2.11) suggest that the two methods give similar results.

To compute $L_{in}$ and $L_{out}$ for the construction of matrix $A$ in (2.11) we traced seismic rays through a radially stratified medium represented by Mohorovicic's law (cf. Bullen & Bolt, 1985), which provides a good approximation to the actual velocity variation over many depth ranges in the Earth (see section 2.4). For source relocation we used the partial derivatives given by Gubbins (1990) as described in the previous chapter.

To solve (2.11) for slowness perturbations (inside and outside the volume under study) and for the event relocation solution vector we used the iterative $LSQR$ method by Paige & Saunders (1982), a conjugate gradient technique first applied to seismic tomography by Nolet (1985); see section 2.5 in the previous chapter. Upon inversion we used two kinds of explicit damping i.e. the minimum norm and the gradient damping described in (2.14) and (2.15). Application of the gradient damping results in a relatively smooth model that facilitates its use as input for quantitative geodynamical studies e.g. Moresi & Gurnis (1996). The gradient damping was applied separately to the blocks inside and outside the study area to avoid smoothing from the low- into the high-resolution part of the model.

We tuned the inversion carefully in order to obtain good results both inside and outside the study area since better constraints on anomalies elsewhere minimize smearing into the study region. Therefore, we actually performed the global inversion with all available data and thus also considered paths that did not transect the mantle volume under study. In Fig. 3.3, we present an example of velocity-perturbation map for both the study region and the rest of the globe for the upper mantle (layer 5 inside and layer 2 outside). The image reveals a pronounced low $P$-wavespeed region around the Pacific,
which is consistent with results of previous global studies e.g. Inoue et al. (1990), and regions of high P-wavespeed associated with the cratonic parts of the continents. For instance, the global image displays the sharp contrast between high and low wavespeed across the Tornquist-Teissyre Zone in Europe (Zielhuis & Nolet 1994), and there is a suggestion that this contrast extends southeastward all the way to India. Figure 3.3 also displays the significant difference between the image resolution inside and outside the study region. For example, the high-wavespeed anomaly, in the right-bottom corner of the study region, associated with the oceanic lithosphere of the Coral Sea and the slab along the Indonesian Island arc are well imaged; these structures would have been smoothed out in models based on a coarser parameterization.
Figure 3.3. Layer anomaly map depicting results of the inversion for upper mantle structure. Contour scales are -1.5% to +1.5% and -0.5% to +0.5% relative to ak135 for solution inside and outside the study area, respectively.
3.2.3 Resolution tests

We assessed the tomographic images by conducting inversions of synthetic data computed from artificial but known 3-D models using the same ray coverage and linearized theory used in the real data inversion. We also used different subsets of data, e.g. different time periods and focal depth intervals, to test the sensitivity of the model to differences in data coverage. Among the techniques most often used are sensitivity tests (Spakman & Nolet 1988; Humphreys & Clayton 1988) and checker-board tests (Inoue et al. 1990; Fukao et al. 1992). The techniques, however, have limitations (Van der Hilst et al. 1993; Léveque et al. 1993) that may lead to incorrect assessment of the reliability of tomographic images. Léveque et al. (1993) argued that the inferences pertinent to image quality depend on the spatial characteristics of the input model used to compute the synthetic data. In addition, we realize that the tests with synthetic data do not simulate accurately the effect of source mislocation and errors in the data and that they are conducted with the same approximations (e.g. model parameterization) and theoretical assumption (e.g. linearization and the neglect of anisotropy) as in the inversion of the real data phase. Also, the neglect of the effects of the fast slab on wave velocity and ray path position can result in the overestimation of the thickness of the slab (Engdahl & Gubbins 1987) but does probably not produce an effect on the large scale structures discussed here that is significant enough to invalidate our conclusions.

In assessing the solution quality, we have conducted some checker-board and target anomaly tests. In this study, however, our main purpose is to assess the resolution of major (large-scale) slab structure beneath the study region. Therefore, we choose to present the results of hypothesis testing using input models which have similar spatial characteristics as the slab structure inferred from the real data inversion. One of the advantages of this approach is that the effects of regularization on the images are similar to that in the inversion of the reported data. We computed the synthetic data by taking the difference between travel times calculated in a 3-D input model (e.g. a slab model or a checker-board pattern) and in the 1-D reference model used. To simulate errors in the data, we added random artificial errors to the synthetic data between -1.0 and +1.0
seconds for $P$, and between -1.5 and +1.5 seconds for $pP$ and $pwP$. We inverted the noisy synthetic data using damping parameters used in the real data inversion to see how well the shape of the input model and the amplitude are recovered. Judging from the recovery, we can tune the inversion by exploring the damping parameters which give the optimum trade off between bias (i.e. the amplitude recovery and smoothness in the model) and the variance reduction of the data. In section 3.4 we present in horizontal and vertical slices the results of hypothesis testing along with the presentation of the real data inversion.

3.3 DATA

The data used in this study are travel-time residuals produced by event relocation and phase re-identification based on arrival times reported to the International Seismological Centre (ISC) from 1964 to 1992, and to the National Earthquake Information Center (NEIC) from 1993 to 1995 (Engdahl, Van der Hilst & Buland 1997). This data set includes data from about 40 portable seismographs deployed in the Australian the SKIPPY project (Van der Hilst et al. 1994). Engdahl et al. (1997) used an improved global travel-time model ak135 (Kennett et al. 1995) and arrival times of first-arriving regional and teleseismic $P$ phases, regional $S$ phases, depth phases ($pP$, $pwP$, and $sP$), and the $PKPdf$ branch to relocate all teleseismically well-constrained earthquakes that have occurred during the period 1964 - 1995. Their procedure insures that depth errors and the mapping of source heterogeneity into mislocation are minimized, and it creates a powerful uncontaminated database of $P$, $pP$ and $pwP$ residuals for use in tomographic inversions. Hereafter, the re-processed data used in this study are referred to as EVB data. We used almost 17,000 (35,000) events inside (outside) the volume under study, recorded at a subset of almost 3000 stations worldwide (Figs 3.4 a & b). All epicentral distances are considered. We used more than $6.0 \times 10^6$ data for the construction of summary rays (see sub-section 2.1). For the inversion we only kept the summary rays
for which the absolute travel-time residual has a magnitude less than 5.0 seconds. If both the source and receiver are located with the study region we used individual ray paths to optimize the sampling. The total number of paths and travel-time residuals constituting the linear system is nearly 550,000.

Figure 3.4. Distribution of stations and epicentres outside the study area. (a) Open triangles depict stations. (b) Small dots depict epicentres of earthquakes.

Figure 3.5 depicts the distribution of the stations and earthquake epicentres used in this study and the $pP$ and $pwP$ surface-reflection points inside the study area. The region between Java and Borneo with only few reflection points (Fig. 3.5c) reflects the pronounced seismic gap between depths of 300 and 500 km beneath the eastern Sunda arc. Comparison of the sampling by either $P$-wave paths or the combination of $P$ and $PwP$ paths (Fig. 3.6) depicts that for regions away from the trench e.g. south
Borneo, and the area between Timor and Australia, sampling is improved through the incorporation of depth phases.

Figure 3.5. Distribution of stations, epicentres and $pP$ & $pwP$ surface reflection points within the study area. (a) Solid triangles depict Skippy stations. (b) Epicentre distribution of earthquakes occurred between 1964 and 1995 in the study region. Locations after ISC (EVB locations). Small dots, crosses and diamonds depict epicentres of shallow ($0 < z \leq 70$ km), intermediate ($70 < z \leq 300$ km) and deep ($z > 300$ km) events; $z$ is the hypocentre depth. (c) Small dots depict surface reflection points of $pP$ and $pwP$ waves. The reflection points of $pP$ and $pwP$ are at land-air or land-water and water-air interfaces, respectively.
Figure 3.6. Sampling by seismic rays in the top layer inside the study region. Hereafter, sampling is plotted in logarithmic scale of the total ray length in each block defined by the model parameterization.

(a) Sampling by $P$ ray only. (b) Sampling by $P$, $pP$ and $pwP$ rays.
3.4 PRESENTATION OF TOMOGRAPHIC IMAGES

We interpret $P$-wave travel-time residuals in terms of velocity perturbations relative to the *ak135* reference velocity model (Kennett *et al.* 1995). In this section we present the images (achieved after conducting 50 iterations with the *LSQR* algorithm) for the entire Indonesian region. The variance reduction upon inversion is 31%. Notice that we used the EVB data set which has a variance that is already significantly smaller than that of published ISC $P$ residuals (Van der Hilst *et al.* 1991; Van der Hilst & Engdahl 1992; and Engdahl *et al.* 1997). We present three anomaly maps for depth intervals representing the upper mantle, transition zone and lower mantle.

From our inversions, we infer that the subducted slab is defined by a laterally continuous region of higher-than-average $P$-wave velocity in the upper mantle, transition zone, and lower mantle (Figs 3.7 a-c). Most parts of the region of interest, in particular along the island arcs, are sufficiently sampled by seismic rays (Figs 3.7 d-f). The image of the slab in the upper mantle (Fig. 3.7a) resembles the present-day Java trench, Timor trough, the curved Banda arc, and the Molucca Collision Zone. It also reveals fast $P$-wave propagation beneath northern Australia, consistent with a thick continental lithosphere (Zielhuis & Van der Hilst 1996). In the transition zone (Fig. 3.7b) a high-wavespeed slab is detected beneath the Java and Banda arcs, but not beneath central Sumatera. In the lower mantle (Fig. 3.7c) the observed prominent and robust feature is the elongated structure of faster-than-average $P$-wave propagation trending east-west over a distance of about 2500 km from below the northwestern tip of Sumatera to below the southern Philippines. Resolution tests suggest that these large-scale anomalies are well resolved. In particular, inversions of synthetic data computed from a specifically designed slab model suggest that the shape of major structure is reliably imaged, but the amplitude of the anomalies is underestimated significantly (Figs 3.7 g-i). The observed deep slab beneath Indonesia (Fig. 3.7c) is confirmed by the result of the inversion of $S$ travel-time residuals (Fig. 3.8a) and is also resolved by the $S$ data (Fig. 3.8b).
Figure 3.7. Results of tomographic inversions. (a-c) Solutions representing upper mantle, transition zone and lower mantle structures with contour scales -2.0% to +2.0%, -1.5% to +1.5% and -1.0% to +1.0% relative to ak135, respectively. (d-f) Sampling by P, Pp and PpwP rays at the corresponding depth intervals. (g-i) Recovery of a slab model designed from the inversion result in (a-c) plotted in percentage of amplitude of the input anomaly; see inset in the upper-right hand corners for input at each depth interval. Velocity perturbations in the input model are set to +5.0%, +4.0% and +3.0% relative to ak135 for the slab in the upper mantle, transition zone, and lower mantle, respectively, and zero elsewhere.
Figure 3.8. (a) A layer anomaly map representing lower mantle structure from the inversion of S data. (b) Recovery of a slab model designed from the inversion result in (a); see inset in the lower-left hand corner for input. Velocity perturbations in the input model are set to +2.0% relative to ak135 in the slab model and zero elsewhere.
In the upper mantle and transition zone the dimension of the smallest feature that is resolved is about 150-200 km. In the lower mantle the resolution length is of the order of about 300-400 km based on P-wave imaging. Notice that resolution generally degrades with increasing distance away from the slab due to irregular or poor sampling. With this and the shortcomings of the resolution tests in mind we only interpret the large-scale structures. In the following we focus on the three regions indicated by three boxes shown in Fig. 3.9.

3.4.1 The Sunda arc

In accord with previous studies, the lithospheric slab is imaged as a feature stretching from the surface to the lower mantle below the eastern Sunda arc with a local deflection where the slab continues into the lower mantle. However, our images suggest that the deep slab is detached from the seismogenic slab beneath Sumatera and perhaps beneath Java as well.

In the upper mantle, the anomaly map depicts the slab as a region of higher-than-average P-wave speed, that parallels the present-day Sunda arc (Fig. 3.10a). The region along the arc is well sampled by seismic rays (Fig. 3.10b). Fig. 3.10 (c) depicts the result of a resolution test suggesting that the large-scale slab is adequately resolved. Three vertical sections across the Sunda arc are presented to illustrate the inferred lateral variation in slab morphology along the arc (Figs 3.11 a-c). The imaged positive velocity anomaly in the lower mantle beneath Sumatera (Fig. 3.11a) has the appearance that the deep part of the slab is detached from the seismogenic slab. The images in Figs 3.11 (b&c) suggest that the Indo-Australian plate dips steeply beneath the Java arc and is only partially outlined by a seismic zone. In the seismic gap, between 350 and 500 km in depth, we detected higher-than-average seismic velocities, but with small amplitudes. The amplitude reduction in the seismic gap may suggest a thinning or "necking" of the Java slab. The Java slab seems to deflect in the uppermost lower mantle and the continuation of flow to even larger depths is off set to the north. We address the resolution of the inferred slab detachment beneath Sumatera, the kink in the slab and the
"necking" of the slab beneath the eastern Sunda arc using semi realistic 3-D slab models. We have used two slab models; one with a continuous and one with a detached slab. The recovery of the continuous slab suggests that our inversion would be able to resolve such a continuous feature. Here, we only present the model of the detached slab (Figs 3.11 d-f). We have added low-wavespeed anomalies adjacent to the model slab in order to investigate the possibility that horizontal smearing of such low-velocity anomalies give the appearance of slab detachment. The recovery of the synthetic anomalies is presented in Figs 3.11 (g-i). Notice, in particular, that the low-velocity anomalies do not interact to conceal the fast slab. On the contrary some low amplitude smearing occurs in the down-dip direction, which may indicate that the gap in which the slab seems absent is even larger than inferred from the images. As for the Java slab, the resolution test results indicate that sampling is possibly sufficient to discriminate between a continuous and a detached slab but the recovery does reveal significant smearing in the down-dip direction (Figs 3.11 h & i). The images suggest the presence of a thinner slab (perhaps due to "necking") across the pronounced seismic gap (Figs 3.11 b & c) but this interpretation is necessarily tentative for such detail is probably at the edge of our current resolution. The tests suggest that the kink in the slab is well resolved.

The summary of our observations and interpretation is given in Fig. 3.12 (see also Widiyantoro & Van der Hilst 1996). Examination of anomaly maps for different depth intervals indicates that the depth range, in which the fast slab is absent, intermittently increases northwestward from zero near 105°E to about 300 km in depth beneath the northwestern tip of Sumatera. The depth to the leading edge of the shallow slab decreases in the same direction, but the depth to the top of the deep slab does not change much. This deep structure seems to sink almost vertically. The observation of the deep slab beneath the Sunda arc is in good agreement with a long history of subduction in the region (Katili 1975).
Figure 3.9. Sub-area division to focus the presentation of tomographic images. Solid lines depict the position of cross sections; 11 a-c, 14 a & b, and 17 a & b are beneath the Sunda, Banda, and MCZ and Philippine Island regions, respectively.
Figure 3.10. Close up of inversion results representing the structure in the upper mantle beneath the Sunda arc. (a) P-wave velocity relative to ak135. (b) Sampling by P, pP and pwP rays. (c) Recovery of a slab model designed from the inversion result in (a); see inset in the upper-right hand corners for input. Velocity perturbation in the model slab is set to +5.0% relative to ak135 and zero elsewhere.
Figure 3.11. Vertical cross sections of tomographic images across the Sunda arc. (a-c) Cross sections 11 a-c are beneath West Sunda (Sumatera), Central Sunda (Java) and East Sunda (between Flores and Java), respectively; contour scale: -1.0% to +1.0%. Sections are plotted from the back-arc in the north (left) to the fore-arc region in the south (right). Hereafter, open dots depict earthquake hypocentres of magnitude ≥5.5 on the Richter scale, projected from a distance of up to 50 km on both sides of the plane of section. (d-f) Model detached slabs synthesised from the slab structure displayed in (a-c); contour scale: -5.0% to +5.0%. We added low-wavespeed anomalies just outside the model slab to assess the effect of the low-wavespeed anomalies observed in (a-c). In the slab and in the added low-wavespeed anomalies, P-wave velocity is +5.0%, +4.0% and +3.0% faster- and slower-than-average for the upper mantle, transition zone and lower mantle, respectively, and zero elsewhere. (g-i) Recovery of the model detached slabs in (d-f); contour scale: -5.0% to +5.0%. Notice the loss of amplitude.
Figure 3.12. Cartoon summarizing our preferred interpretation of the images of inferred slab structure beneath the Sunda arc. The shallow slab dips in a north-east direction at an angle of ~40° below Sumatera and in a north direction at an angle of ~60° below Java; the deep slab sinks almost vertically into the lower mantle. The tectonic stress perpendicular to the trench as derived by Cloetingh & Wortel (1986) is also displayed. Positive and negative stress is associated with tension and compression, respectively.
3.4.2 The Banda arc

Despite numerous studies of the Banda region, e.g. by Cardwell & Isacks (1978), Hamilton (1979), Von der Borch (1979), and McCaffrey (1988), there is no consensus on the complicated structure beneath the Banda region. Cardwell & Isacks (1978) investigated the geometry of the subducted lithosphere in the upper mantle and transition zone beneath the Banda Sea from seismicity and fault plane solutions. Their conclusions are confirmed by our current study.

The inversion reveals the twisting of the slab in the upper mantle, which parallels the present-day curved Banda arc (Fig. 3.13a). Resolution tests indicate that sampling by seismic rays is sufficient to recover slab structure beneath the Banda region, in particular along the arc (Figs 3.13 b & c). At this depth interval (110 - 160 km), the twisted slab seems to connect with the top of the two opposing subducted slabs associated with the Molucca Sea plate. The lateral variation in slab structure beneath the region is further illustrated by vertical sections across the arc (Figs 3.14 a & b). Resolution tests (Figs 3.14 c-f) reveal that the northward and westward subducting slabs are resolved. The observed anomaly that stretches in east-west direction just beneath the 660 km discontinuity (Fig. 3.14b) is just within the resolving power of the data used. Beneath the westernmost part of the Banda region the subducted slab dips steeply in the upper mantle and is deflected in the transition zone, in accord with the observed seismicity (Fig. 3.14a). This deflection, however, is not very well recovered by the inversion (Fig. 3.14e) and one should thus be cautious with interpretation from the images alone. The east-west trending cross section shows that the seismically fast slab dips rather gently, which is also in good agreement with seismicity in parts of the slab (Fig. 3.14b). Also here the dip of the slab decreases abruptly in the transition zone. Although not resolved in detail, our images suggest that the slab beneath Banda forms a spoon-shaped structure in accord with inferences from earthquake locations (Cardwell & Isacks 1978).

The structure of the slab beneath the Banda arc, summarized in Fig. 3.15, can be explained in the context of recent northward movement of the Australian continent. In the Miocene, the small islands along the present-day curved Banda arc were striking east-
west. In the Pliocene, a dramatic event occurred where the Banda arc was twisted counter clockwise due to a combination of the northward drift of the Australian continent and the counter-clockwise rotation of New Guinea (Katili 1973). The deflection of the slab in the transition zone is likely to pre-date the rotation.
Figure 3.13. Close up of inversion results representing the structure in the upper mantle beneath the Banda arc. (a) $P$-wave velocity relative to $ak135$. (b) Sampling by $P$, $pP$ and $pvP$ rays. (c) Recovery of a slab model designed from the inversion result in (a) plotted in percentage of amplitude of the input anomaly; see inset in the upper-right hand corners for input. Velocity perturbation in the model slab is set to $+5.0\%$ relative to $ak135$ and zero elsewhere.
Figure 3.14. Vertical cross sections of tomographic images across the Banda arc. (a & b) Cross sections 14a and 14b are plotted from the inner-arc in the north (left) to the outer-arc region in the south (right) and from the inner-arc in the west (left) to the outer-arc region in the east (right), respectively; contour scale: -1.0% to +1.0%. (c & d) Model slabs synthesised from the slab structure displayed in (a & b); contour scale: -5.0% to +5.0%. In the slab, $P$-wave velocity is +5.0%, +4.0% and +3.0% faster-than-average for the upper mantle, transition zone and lower mantle, respectively, and zero elsewhere. (e & f) Recovery of the model slabs in (c & d); contour scale: -5.0% to +5.0%. Notice the loss of amplitude.
Figure 3.15. Block diagram illustrating the geometry of the subducted lithospheric slab beneath the Banda arc (modified from Cardwell & Isacks, 1978, based on the tomographic images). Modification was made to show clearly that the structure beneath Banda forms a spoon-shaped feature of the descending slab with a left lateral fault in the eastern part of the Banda arc as inferred from seismicity (e.g. Cardwell & Isacks 1978). Notice the existence of the southward subducted slab beneath the northern part of Banda.
3.4.3 The MCZ and Philippine arc

The configuration of the Molucca Sea and Philippine Sea plates has been discussed by e.g. Silver & Moore (1978), Hamilton (1979), Cardwell et al. (1980), Hall (1987). From seismicity, they infer two oppositely dipping Benioff zones. With P-wave imaging Puspito et al. (1993) detected two opposing subducted slabs of the Molucca Sea plate and also the westward subducted slab of the Philippine Sea plate beneath the Philippine Islands to the northeast of the MCZ. Our results are in good agreement with these observations, but suggest that the westward subducting slab dips more steeply than the eastward one.

Fig. 3.16 (a) the image shows an offset between the top of the Molucca and Philippine Sea plates at about 5° N, with the Molucca Sea plate located at the southwest corner of the Philippine Sea plate. The upper mantle beneath the MCZ and the Philippine arc are well sampled by seismic rays and test inversions suggest that the major structures, including the observed offset between the Molucca Sea and the Philippine Sea plate, are reliably imaged (Figs 3.16 b & c). We present two arc-perpendicular, vertical sections across the MCZ and the southern Philippines (Figs 3.17 a & b). The cross section in Fig. 3.17 (a) shows the two opposing subducted slab of the Molucca Sea plate, with the dip of the westward subducting slab steeper than the one of the eastward dipping slab. The image suggests that the westward subducting slab penetrates into the lower mantle, but one has to be aware of possible 3-D effects since this intersects the deep part of the Java slab. In the west-dipping limb seismicity is deeper than in the east seismic zone (see dots in Fig. 3.17a), which may imply that the westward subduction was either active longer or that it has a faster rate of subduction (Silver & Moore 1978). To the northeast of the Molucca Sea, the image reveals the subducted lithospheric slab that may be associated with the westward subduction of the Philippine Sea plate beneath the Philippines (Fig. 3.17b). Results of resolution tests suggest that the two opposing subducted slabs beneath the MCZ and the slab beneath the southern Philippines are resolved (Figs 3.17 c-f). We simulated the inferred sub-horizontal slab in the uppermost lower mantle beneath the reversed U-shape of the Molucca Sea plate (Fig. 3.17c). Its
recovery (Fig. 3.17e) suggests that we have sufficient data coverage to resolve this intriguing and unexpected structure.

Our inferences from the new P-wave images in general compare favourably with previous results. An important new observation is that the dips of the eastward and westward subducting slabs of the Molucca Sea plate are different. We postulate that this difference in dip angle is due to the shear caused by a combination of the westward thrust of the Pacific plate combined with movement on the left-lateral Sorong fault to the south of the Molucca Sea and viscous drag in the mantle. Hall (1987) reported that the westward shift of the Halmahera arc occurred in the Pleistocene. We summarize our preferred interpretation of the images using the block diagram given in Fig. 3.18, which is a slight modification of the geometry inferred by Cardwell et al. (1980).
Figure 3.16. Close up of inversion results representing the structure in the upper mantle beneath the MCZ and Philippine arc. (a) $P$-wave velocity relative to $ak135$. (b) Sampling by $P$, $pP$ and $pwP$ rays. (c) Recovery of a slab model designed from the inversion result in (a) plotted in percentage of amplitude of the input anomaly; see inset in the upper-right hand corners for input. Velocity perturbation in the model slab is set to $+5.0\%$ relative to $ak135$ and zero elsewhere.
Figure 3.17. Vertical cross sections of tomographic images across the MCZ and Philippine arc. (a & b) Cross sections 17a and 17b are plotted from the inner-arc in the northwest-west (left) to the outer-arc region in the southeast-east (right) and from the inner-arc in the southwest-west (left) to the outer-arc region in the northeast-east (right), respectively; contour scale: -1.0% to +1.0%. (c & d) Model slabs synthesised from the slab structure displayed in (a & b); contour scale: -5.0% to +5.0%. In the slab, P-wave velocity is +5.0%, +4.0% and +3.0% faster-than-average for the upper mantle, transition zone and lower mantle, respectively, and zero elsewhere. (e & f) Recovery of the model slabs in (c & d); contour scale: -5.0% to +5.0%. Notice the loss of amplitude.
Figure 3.18. Configuration of the Molucca Sea and Philippine Sea Plates (modified from Cardwell et al., 1980, based on the tomographic images). An important modification is that the westward subduction of the Molucca Sea plate is steeper than the eastward subduction i.e. - 60° and - 40°, respectively. See text for discussion.
**3.5 DISCUSSION**

Simultaneous inversion for regional and global-scale structure has successfully imaged major structures outside the study area and is likely to have minimized the mapping of distant structure into the study region. The inclusion of depth phases added significantly to the sampling and to focal depth control, which improves resolution of slab structure. The overall improvements of the resulting images are probably due to the combination of the high quality of the re-processed (EVB) data and the *ak135* reference velocity model, the incorporation of depth phases, the simultaneous inversion for global structure, and the use of gradient damping.

The slab detachment beneath Sumatera was probably triggered by the Early Tertiary arrival of the oceanic spreading center at the western Sunda trench. The buoyancy of the young lithosphere and the change to oblique subduction due to the rotation of Sumatera, which reduced flow in the vertical direction (Hilde *et al.* 1977), may have combined to temporarily cease subduction, while the deep, old slab continued to sink to larger depths. The along-strike variation in slab morphology beneath the Sunda arc appears similar to that beneath the Tonga-Kermadec arc (compare, for instance, Figs 3.11 a-c with Fig. 3 of Van der Hilst (1995), but we argue that the deep Sunda slab evolved differently. Kinks in the subduction trajectory beneath southwestern Pacific island arcs have been explained by retrograde motion, or "roll back", of the slabs in a dynamically self-consistent way, causing oceanward trench migration and concurrent back arc spreading. Lateral variation in the shape of slab was explained by trench rotation accompanying changes in plate motion (e.g. Van der Hilst & Seno 1993). Due to the rotation, the trench has remained orthogonal to the direction of subduction. In contrast, the structural trends in Sumatera are oblique to the present-day direction of plate motion; from the difference in the strike of the high velocity anomalies in the upper (Fig. 3.7a) and lower mantle (Fig. 3.7c) we infer a clockwise rotation of the shallow relative to the deep slab. The inference that strike of the lower mantle slab is consistent with the present-day direction of subduction of the Indo-Australian plate beneath Eurasia (Minster
suggests that the rotation can not be explained by a change in the
direction of relative plate motion of the major plates. Instead, we postulate that the
rotation of Sumatera relative to the deep slab may have been caused by the Early Tertiary
northward movement of India and the collision of a mid-oceanic ridge complex with the
volcanic arcs of western Indonesia (see Widiyantoro & Van der Hilst 1996 and the
references therein). The northward collision in the west and the southward expansion of
the overriding plate in the east may have formed a torque that caused rotation of the Sunda
trench. This torque may have been enforced by tensional stresses along the Java trench
(Fig. 3.12) owing to the large negative buoyancy of the old oceanic lithosphere
(Cloetingh & Wortel 1986), which may have facilitated the oceanward migration of the
eastern part of the Sunda arc and, consequently, the kinking of the slab. The deep slab
beneath the region as revealed by P and S data inversions (Figs 3.7c & 3.8a) is confirmed
by the results of our inversions of P- and S- travel-time residuals for global structure
shown in Figs 3.19 (a & b). Our global images suggest that the deep structure beneath
Indonesia forms the eastern end of a deep anomaly beneath southern Eurasia that is
associated with the past subduction of the Mesozoic Tethys ocean floor (see chapters 6, 7
& 8 and Van der Hilst, Widiyantoro & Engdahl 1997).

Near the boundary between the Sunda and Banda arcs, between Flores and Timor
(Puspito & Shimazaki 1995), the images suggest an abrupt change in the depth of slab
penetration into the mantle (see Fig.3.12). Beneath Banda the slab does not directly
penetrate into the lower mantle but seems to deflect in the transition zone. The curved
Banda arc can perhaps be explained in the context of continent-island arc collision
between the recent, about 5 million years ago, northward movement of the Australian
continent and the Banda arc. The collision may have caused the shortening of the upper
plate in the convergence direction beneath the forearc and backarc basin over the area
between the Timor trough and northern Banda (McCaffrey 1988). One of the most
striking aspects of the Banda arc is the apparent counter clockwise rotation of almost
180°. The rotation may have formed the narrow-deep region in the Banda Sea, named as
the Weber Deep, which is assumed to be oceanic crust trapped by the surrounding
younger arcs (Katili 1975). The curious high-wavespeed anomaly in the uppermost lower mantle (Fig. 3.14b) may represent excess slab materials that results from the distortion of the deflected slab. Cardwell & Isacks (1978) interpret that the subduction zone to the north of the Banda Sea is not a continuation of the one along the southern and eastern parts of the arc. Our inversions cannot resolve such detail. The complete exploration of the lateral variation in the shape of the subducted slab (e.g. Figs 3.14 a & b) and inferences from seismicity suggest that the large-scale structure of the slab in the upper mantle and transition zone beneath Banda forms a spoon-shaped feature (Fig. 3.15).

The structure of subducted slab beneath the Molucca Collision Zone is more complex than beneath other convergent margins. The MCZ is underlain by two oppositely dipping slabs associated with the subducted Molucca Sea plate. The westward thrust of the Pacific plate combined with the large left-lateral movement of the Sorong fault to the south of the Molucca Sea may have shifted the Molucca Sea plate and its surrounding micro plates toward the Eurasian continent. This shift may have caused the offset between the Philippine Sea and the Molucca Sea plate (Fig. 3.16a) as well as the observed different dips of the opposing subducted slab (see Fig. 3.17a). Puspito et al. (1993) suggested that the westward dipping slab beneath the MCZ penetrates into the lower mantle. Due to the complex geological setting in Indonesia, however, we must be cautious to interpret the origin of deep anomalies beneath the region. Not all deep anomalies can be traced back unambiguously to slab structure in the shallower mantle; in complex regions the use of 2-D cross section can be misleading. We argue that the deep high-wavespeed anomaly west of the MCZ most likely represents the deep Indo-Australian slab subducted towards the north along the Sunda arc (Fig. 3.11c). Our argument is based both on a careful examination of the three dimensional model and on the size of the plate. The Molucca Sea plate is relatively small and is unlikely to have formed such a large, deep anomaly. Furthermore, the long subduction history of the Indo-Australian plate supports our argument, whereas the subduction of the Molucca Sea plate is only recent. A confusing feature, which is within our resolution, is the sub-
horizontal anomaly in the uppermost lower mantle beneath the Molucca Sea plate (Fig. 3.17a). We do not know its origin. One may speculate that it may be remnant slab of past subduction of the Pacific plate. To the northeast of the Molucca Sea, the image (Fig. 3.17b) shows that the subducted lithospheric slab associated with the westward subducting Philippine Sea plate seems also to penetrate into the uppermost lower mantle beneath the southern Philippine arc. Based on inferences from the complete images and from the seismicity, however, we prefer to interpret that the shallow part of the slab is associated with the subducted Philippine Sea plate while the deeper part may correspond to the westward dipping slab of the Molucca Sea plate. This preferred interpretation suggests that the deeper part of the westward dipping slab of the Molucca Sea plate extends to the north up to the mantle beneath the southern Philippines. The interpretation of the northward extension of the deep slab is supported by the corresponding deep seismicity (Fig. 3.5b) and the fast $P$-wavespeed slab trending almost north-south from the Philippines to north Sulawesi (Fig. 3.7b). It is also in accord with the present strike of the Sangihe and Halmahera arcs i.e. NNE-SSW and with the WNW direction of the Pacific plate motion.
Figure 3.19. (a) A layer anomaly map in the lower mantle from an inversion of $P$ data for global structure. We used more than $7.0 \times 10^6$ data upon the inversion and fine parameterization with horizontal cell size of $2^\circ \times 2^\circ$ for whole globe (see chapter 6). (b) The same with (a), but inferred from $S$ data (see chapter 7). Notice the long, positive anomaly stretching from the Mediterranean to Indonesian region.
3.6 CONCLUDING REMARKS

The new tomographic images reveal mantle structure beneath Indonesia in more detail than previous studies. The improvements are likely to be due to the combination of the use of better data (extensively processed EVB data instead of ISC), the use of the ak135 reference velocity model, the inclusion of a global inversion in the regional study, the use of depth phases, and the use of gradient damping. The results of resolution tests suggest that the large-scale slab structure beneath the Indonesian archipelago is resolved by either P or S data. In spite of relatively noisy data, the results of S data inversion are in general in good agreement with those from P data.

In the top of the lower mantle beneath the eastern part of the Sunda arc the subducted slab forms a kink, which gradually disappears further west. Beneath Sumatera the image suggests that the slab is detached in the transition zone. However, the discussion of the slab detachment beneath Sumatera is still tentative. As discussed in the previous section, at this stage, we can only speculate on the evolution of the slab detachment. A further investigation of this intriguing structural feature is required. The interpreted “necking” of the slab beneath Java coincides with the pronounced existing seismic gap. We related this complex slab structure to the Tertiary evolution of southeastern Asia and the Indian ocean region.

Further east, beneath the curved Banda arc, the twisting of the slab in the upper mantle is well imaged. The vertical section across the arc along with the anomaly maps show that the subducted slab is deflected in the transition zone. A counter clockwise rotation of the arc has resulted in a spoon-shaped structure of the slab complex beneath the Banda Sea. This rotation is probably related to the Pliocene continent-arc collision between Australia and Banda.

Beneath the Molucca Collision Zone (MCZ) the images reveal two opposing subducted slabs of the Molucca Sea plate, which is related to the active arc-arc collision between Halmahera and Sangihe. The dips of the two slab segments are significantly different, which may be due to the westward shift of the slab complex in a viscous
mantle. Although the wavespeed anomaly associated with the westward dipping subduction seems to continue into the uppermost part of the lower mantle, this deep structure is more likely associated with the deep subduction of the Indo-Australia plate along the Sunda arc. The use of 2-D cross sections in such complex structures can thus be misleading. The westward dipping slab beneath the MCZ may extend further north to beneath the southern Philippines. Anomaly maps in the upper mantle reveal the offset of the Molucca Sea and Philippine Sea plates at about 5°N, which may indicate the boundary between those two plates.
4 Structure of subduction zones worldwide

4.1 INTRODUCTION

In the previous chapter, we have established a procedure to determine the seismic structure in the Indonesian region based on the inversion of P-wave travel-time residuals. We now extend this approach to the probing of three-dimensional structure to consider nine other regions where current subduction is occurring.

The ten regions are specified in Table 4.1 and Fig. 4.1; region 3, Indonesia has been treated in the previous chapter. We will consider the other regions here, with special attention to the delineation of subduction zones worldwide. Detailed interpretations of all the resulting tomographic images is beyond the scope of this thesis.

Table 4.1: Regional P-wave travel-time tomography

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<tr>
<th>Number</th>
<th>Region</th>
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<tbody>
<tr>
<td>1.</td>
<td>Europe</td>
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<td>2.</td>
<td>Iran, India and Himalaya</td>
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<td>3.</td>
<td>Indonesia</td>
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<td>4.</td>
<td>Northwest Pacific</td>
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<td>Tonga-Kermadec</td>
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<td>6.</td>
<td>Aleutian region</td>
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<td>7.</td>
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<td>8.</td>
<td>Central America including Caribbean</td>
</tr>
<tr>
<td>9.</td>
<td>South America</td>
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<tr>
<td>10.</td>
<td>Scotia plate</td>
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We have used the re-processed data set of global arrival times described in the previous chapter and have once again minimized the mapping of aspherical mantle structure from outside the volume under study by incorporating a low-resolution global inversion in each regional inversion. As in the study of mantle structure beneath Indonesia, we have parameterized each region using $1^\circ \times 1^\circ$ cells and the rest of the mantle using $5^\circ \times 5^\circ$ cells. The cell layer division is the same as described in Table 3.1. The size of the matrices required for the inversion for each region is comparable to that used in the inversion for mantle structure below Indonesia (see chapter 3).

In order to test the resolution of mantle structure beneath each study region, we have conducted resolution tests using a regular pattern of synthetic anomalies as displayed in the insets in Figs 4.2 (d-f). The anomalies were regularly spaced and were assigned to alternating layers to help assess vertical resolution. Spakman et al. (1993) remarked on the disadvantages of the use of "checkerboard" or harmonic tests. They noted the difficulty in interpreting the results of such a test, where lack of resolution causes averaging of amplitudes between or among cells in the test result. Therefore, we intentionally use "spike" or impulsive anomalies in the input model in order to detect more easily the directions in which there may be poor resolution. In general, the results of our resolution tests imply that the major observed structures beneath each region are within the resolution of the data used in the inversions; the exception is underneath the Scotia plate (region 10). In particular, a pattern of input anomalies on the scale of the order of two cells ($2^\circ \times 2^\circ$) or more are reasonably recovered. However, the amplitude of the recovered anomalies are suppressed as a result of the use of damping and incomplete convergence of the iterative LSQR inversion method described in the previous chapter (also see Spakman & Nolet 1988).

The results of the inversions are generally in good agreement with the results of previous tomographic studies, but include more detail on some features than before. As examples: we detect (i) an intriguing lithospheric slab structure beneath the Himalaya front close to the Pamir-Hindu Kush region in which a cluster of intermediate earthquakes
exists (Roecker et al. 1980), (ii) a subducted oceanic lithospheric slab deflected in the transition zone beneath the central Aleutian islands, and (iii) a sub-horizontal subducted slab beneath north America. The tomographic images confirm structures reported in earlier studies e.g. the subducted slabs beneath the Mediterranean (Spakman et al. 1993), northwest Pacific (Kamiya et al. 1988; Van der Hilst et al. 1991; and Fukao et al. 1992), Tonga (Van der Hilst 1995), the Caribbean region (Van der Hil & Spakman 1989; Van der Hilst 1990) and south America (Engdahl et al. 1995). The images also provide strong evidence for regional slab penetration into the lower mantle.

In this chapter, we present an overview of the different classes of subduction zone structures by presenting anomaly maps at three depths chosen to best display the features of each region:

(a) in the upper mantle:

- 70 - 110 km, for regions 2, 4, 6, 7 and 9;
- 220 - 280 km, for regions 1, 8 and 10; and
- 280 - 340 km, for region 5.

(b) in the transition zone:

- 490 - 570 km, for regions 1, 2, 4, 5, 7, 8, and 9; and
- 570 - 660 km, for regions 6 and 10.

(c) in the lower mantle:

- 930 - 1020 km, for all the nine regions.

For each layer depicted, the appropriate resolution tests are displayed in terms of the amplitude recovery of the test pattern. At least one vertical cross-section is displayed for each of the nine regions. The global models which are obtained as a by-product of the recovery of regional structure are presented in the next chapter.
Figure 4.1. Study regions. Boxes labelled by the numbers [1-10] indicate regions under study. They were intentionally chosen to represent subduction zones worldwide. Note: results of region 3 have been presented in the previous chapter.
4.2 REGIONAL TOMOGRAPHIC IMAGES

In the following, we present the tomographic images obtained after conducting nine different simultaneous high-resolution regional and low-resolution global inversions. The presentation follows the order of numbers shown in Fig. 4.1 starting from Europe, around the Pacific, to the Scotia region.

5.2.1 Region 1: Europe

Over the past few years, several tomographic studies have been performed to delineate mantle structures beneath Europe and the Mediterranean region. Spakman et al. (1993) used ISC travel-time data to produce a three-dimensional P wave-speed structure; they mapped a slab anomaly below the Aegean down to 800 km depth. Zielhuis & Nolet (1994) detected a sharp contrast between high and low shear-wave velocities along the Tornquist-Teisseyre Zone (TTZ) by waveform inversion of surface waves. More recently, Marquering et al. (1996) have produced a new shear-wave velocity model by using an improved waveform inversion method. Our results confirm the striking features observed in the previous studies.

Our model shown in Fig. 4.2 (a) depicts the boundary between high and low wave-speed anomalies along the TTZ in the upper mantle. In the transition zone (Fig. 4.2b), we have detected a high wave-speed structure beneath central Europe and a low wave-speed anomaly below eastern Europe. These observations support the shear-wave models by Zielhuis & Nolet (1994) and Marquering et al. (1996). We observed a deep positive anomaly (Fig. 4.2c), which may be correlated with the deep subducted Aegean slab previously reported by Spakman et al. (1993). The results of resolution tests displayed in Fig. 4.2 (d-f) suggest that (i) the sharp boundary detected in the upper mantle is within the resolved mantle region, (ii) the fast wave-speed feature in the transition zone beneath central Europe is well resolved, (iii) the deep positive anomaly is also resolved with possible smearing in east-west direction. This smearing also exists in the model by Spakman et al. (1993).
Figure 4.2. Layer anomaly maps depicting results of the inversion for region 1. (a) Solution representing upper mantle structure with contour scale: -2.0% to +2.0% relative to ak135. Line A1 depicts the position of cross section displayed in Fig. 4.3. (b) Transition zone (scale: -1.5% to +1.5%). (c) Lower mantle (scale: -1.0% to +1.0%). (d-f) Amplitude recovery of an input model i.e. regularly spaced positive anomalies as shown in the insets. The input anomalies (see insets) have horizontal dimensions of 2° x 2°, 3° x 3° and 4° x 4° in the upper mantle, transition zone and lower mantle, respectively. Velocity perturbations inside the input anomalies are set to +3.0% relative to ak135 and 0.0% outside. The same input anomaly patterns in the upper mantle, transition zone and lower mantle are used for regions 2, 4 through to 10. Hereafter fat dashed lines indicate major plate boundaries after NUVEL-1 DeMets et al. (1990).
A detailed examination of our model gives some indication that there is slab detachment below the Mediterranean region as discussed in detail by e.g. Wortel & Spakman (1992) but the image is not as convincing as the image of detachment beneath Sumatra presented in the previous chapter. An intriguing feature observed in the upper mantle is a low wave-speed spot interrupting the large fast wave-speed anomaly beneath the Precambrian craton in eastern Europe, but the result of the resolution test suggests that this feature is not well resolved (see Figs 4.2 a & d).

A vertical mantle section taken across the observed deep anomaly is shown in Fig. 4.3. The image suggests that the subducted Aegean slab penetrates into the lower mantle clearly, which is in excellent agreement with the observation by Spakman et al. (1993). The slab seems to reach a depth of about 1000 km. In chapter 6, we test and discuss penetration depths of subducted slabs, including the Aegean slab in detail.

Figure 4.3. Vertical section of the tomographic image for region 1 plotted from southwest to northeast (left to right). Hereafter open dots indicate earthquake hypocenters from a distance of up to 50 km on both sides of the section plane; the model parameterization in the radial direction is illustrated in the column superimposed at the right-hand side of the cross section; the number plotted above the top-left corner of the vertical section indicates the length of the cross section at the Earth's surface.
4.2.2 Region 2: Iran, India and Himalaya

Our present study region shown in Fig. 4.4 includes the Himalaya, a region which may have resulted from continent-continent collision of the Indian and Eurasian plates (e.g. Molnar & Tapponnier 1975). It also includes the Pamir-Hindu Kush region in which a cluster of intermediate depth seismicity exists (Roecker et al. 1980). Many studies for this region have been made especially by Soviet seismologists, since the region (although it does not have an island arc structure) is one of the most active sources of earthquakes felt within the CIS formerly named as the USSR. Roecker (1982) conducted a tomographic study using P and S travel-time data to produce three-dimensional velocity structures below the Pamir-Hindu Kush region. His results provide possible evidence for subducted continental rather than only oceanic lithosphere to depths of at least 150 km. Our results strongly confirm this observation and also reveal a subducted slab down to the transition zone beneath the Makran subduction zone located between Saudi Arabia and India.

The upper mantle section in Figure 4.4 (a) displays a large area of fast wave propagation in the upper mantle from beneath eastern Saudi Arabia to beneath India. While low wave-speed anomalies are imaged beneath Iran, Afghanistan and intriguingly underneath a spot in northwest India. The amplitude recovery shown in Fig. 4.4 (d) suggests that resolution is good for structures beneath the continent, but poor for those beneath the ocean, largely because of the absence of stations. In the transition zone (Fig. 4.4b), structures seem to be dominated by low-velocity anomalies spreading beneath the entire study region. However, a positive anomaly is imaged below a short subduction zone between eastern Saudi Arabia and India (the Makran subduction zone). Again the resolution seems to be better below the continent than below the ocean because of lack of sampling beneath the ocean domain (see Fig. 4.4e). A layer anomaly map representing the lower mantle (Fig. 4.4c) reveals the existence of deep, positive anomaly enveloped by low-velocity features beneath India. This deep structure is reasonably resolved although smearing may have occurred (Fig. 4.4f).
Figure 4.4. (a-c) Layer anomaly maps depicting structures in the upper mantle, transition zone and lower mantle, respectively. (d-f) Amplitude recovery of input anomalies with horizontal dimensions of 2° x 2°, 3° x 3° and 4° x 4° in the upper mantle, transition zone and lower mantle, respectively (as displayed in the insets in Figs 4.2 d-f). Lines A2, B2 & C2 in (d) depict the position of cross sections displayed in Fig. 4.5. See caption of Fig. 4.2 for more explanation.
Three mantle sections are displayed in Fig. 4.5 and are plotted across the deep slab beneath the region, which may be an indication of remnant subduction beneath the Indian region. The image shown in cross section A2 implies that a high-velocity anomaly exists beneath eastern Saudi Arabia, which may be correlated with part of the Arabian shield. This anomaly seems to be significantly thicker than that imaged below the region further north (see the fast wave-speed anomaly in the top-right corner of cross section A2). Cross section B2 provides an image of subducted slab down to the transition zone beneath the short Makran subduction zone in the northern part of the Gulf of Oman. Vertical section C2 is taken across a region close to the Pamir-Hindu Kush region in which a cluster of intermediate depth earthquakes exists. The image provides evidence for a subducted slab down to the lowermost upper mantle, which is consistent with its seismicity. The subducted slab seems to be enveloped by low-velocity anomalies which is in excellent agreement with the observation by Roecker (1982). As suggested by Roecker (1982), this may be evidence for subducted crust with lower velocity than that in the slab (see a thin layer of slow wave propagation just above the subducted slab). Notice that despite the limited resolution of structures beneath the ocean, the images shown in cross sections B2 and C2 depict thinner high-velocity lids just below the ocean (upper-left sides) than those below the continent (upper-right sides).
Figure 4.5. Vertical sections of the tomographic images for region 2 plotted from southern to northern sides (left to right). The model parameterization in the radial direction is illustrated in the column superimposed at the right-hand side of the lowermost panel. See caption of Fig. 4.3 for more explanation.
4.2.3 Region 4: Northwest Pacific

Seismic tomographic studies for mantle structure beneath Japan and other island arcs of the northwest Pacific have been carried out by many Japanese investigators e.g. Hirahara (1977 & 1981), Kamiya et al. (1988) and Fukao et al. (1992) as well as non-Japanese researchers e.g. Spakman et al. (1989) and Van der Hilst et al. (1991 & 1993). In accord with previous studies, our images indicate that subducted slabs beneath southern Kuril, Japan and central Izu Bonin are deflected in the mantle transition zone, but sink almost vertically into the lower mantle beneath the Marianas. Our results also clearly reveal a subducted slab beneath the Ryukyu trench to the west of the Izu Bonin trench.

Lateral variations in velocity below the region are shown in Figs 4.6 (a-c). A long zone of fast wave propagation in the upper mantle approximately marks the eastern plate boundary and the Ryukyu trench (Fig. 4.6a). Some low-velocity features are imaged beneath the back-arc regions. In the transition zone (Fig. 4.6b), the image is almost in contrast to the structure in the upper mantle. High-velocity anomalies are observed largely beneath the back-arc regions, but the eastern plate boundary is now marked by a long slow wave-speed anomaly. Figure 4.6 (c) depicts that high-velocity anomalies are imaged in the lower mantle beneath some regions including northern Kuril and Mariana. The results of the resolution analyses displayed in Figs 4.6 (d-f) suggest that the structure in the upper mantle is rather poorly resolved, but deeper structures beneath regions of interest (subduction zones) are much better resolved. In the upper mantle, smearing occurs rather severely but anomalies beneath Honshu which has a high station density are well recovered.
Figure 4.6. (a-c) Layer anomaly maps depicting structures in the upper mantle, transition zone and lower mantle, respectively. (d-f) Amplitude recovery of input anomalies with horizontal dimensions of 2° x 2°, 3° x 3° and 4° x 4° in the upper mantle, transition zone and lower mantle, respectively (as displayed in the insets in Figs 4.2 d-f). Lines A4, B4, C4 & D4 in (d) depict the position of cross sections displayed in Fig. 4.7. See caption of Fig. 4.2 for more explanation.
Four vertical sections across the region are displayed in Fig. 4.7. Cross sections A4, B4 and C4 clearly display that the subducted slabs are stagnant in the transition zone, which is in excellent agreement with the observation by e.g. Fukao et al. (1992). However, in chapters 6 and 7 we will demonstrate that the slab beneath the northwestward extension of cross section B4 may have sunk to the core-mantle boundary (CMB). The image in vertical section C4 reveals a subducted slab below the Ryukyu trench as well as below Izu Bonin and suggests that there is a deep, high-velocity anomaly just beneath the Izu Bonin trench which is consistent with an observation from an independent study by Castle and Creager (personal communication, 1996). In good agreement with the result by Van der Hilst et al. (1991), cross section D4 indicates that the slab beneath Mariana penetrates into the lower mantle. The low amplitude in the slab may be due to lack of ray coverage, because the slab is almost vertical. Note that a vertical slab is not well sampled by currently used seismic phases i.e. $P$, $pP$ and $pwP$ as they leave the slab quickly. Phases with steeper take-off angles such as $PKP$ would have better sampled such a vertical slab (see also Okano & Suetsugu 1992).
Figure 4.7. Vertical sections of the tomographic images beneath northwest Pacific plotted from western to eastern sides (left to right). See caption of Fig. 4.5 for more explanation.
4.2.4 Region 5: Tonga-Kermadec

Previous tomographic investigations of the subducted lithosphere below the Tonga trench using $P$ wave travel-time data were carried out by e.g. Zhou (1990) and Van der Hilst (1995). They observed that the slab beneath the Tonga and Kermadec regions penetrate into the lower mantle in accord with results of residual sphere analyses of travel times by Fischer et al. (1991) and Fischer & Jordan (1991). On the other hand, other studies e.g. by Giardini & Woodhouse (1984 & 1986) provide evidence for local horizontal flow in the transition zone. The results of tomographic imaging by Van der Hilst (1995) depict that the slab penetrates into the lower mantle beneath the Kermadec trench without any kinking although beneath the Tonga trench there does appear to be a kink in the slab. This reconciles the evidence for positive anomalies below and local horizontal flow in the transition zone from previous studies. Our results are in excellent agreement with the observation by Van der Hilst (1995).

Three layer anomaly maps are displayed in Figs 4.8 (a-c). In the upper mantle (Fig. 4.8a), the slab (high wave-speed anomaly) seems to parallel the present-day Tonga and Kermadec trenches. Some short slab segments are also imaged below the Vanuatu trench and the Lau basin to the west of the northern Tonga trench. Figure 4.8 (b) depicts a positive anomaly beneath the back arc region, which may associate with the previously inferred local horizontal flow in the transition zone beneath north Tonga. The inversion reveals a deep anomaly trending northwest (Fig. 4.8c). This could imply that the Tonga trench has rotated clockwise as suggested previously by Van der Hilst (1995). The results of resolution tests for the corresponding depth intervals (Figs 4.8 d-f) suggest that structures below the back-arc region ranging from the upper mantle to the lower mantle are resolved. In particular, they imply that the observed large-scale slabs are well within the resolution of the data.
Figure 4.8. (a-c) Layer anomaly maps depicting structures in the upper mantle, transition zone and lower mantle, respectively. (d-f) Amplitude recovery of input anomalies with horizontal dimensions of 2° x 2°, 3° x 3° and 4° x 4° in the upper mantle, transition zone and lower mantle, respectively (as displayed in the insets in Figs 4.2 d-f). Lines A5 and B5 in (a) depict the position of cross sections displayed in Fig. 4.9. See caption of Fig. 4.2 for more explanation.
In Fig. 4.9, we show two vertical sections (A5 and B5) across the Tonga and Kermadec trenches. As implied by the images in the map views (Figs 4.8 a-c), we observe a kink in the subducted slab depicted in cross section A5. This section also displays subducted slabs in an opposite direction ("eastward") beneath the Vanuatu trench and Lau basin (Fiji), which are consistent with seismicity inferences (see open dots superimposed). Also as expected from the layer anomaly maps, we detect that the slab beneath Kermadec penetrates into the lower mantle without a kink and without being trapped in the transition zone (Fig. 4.9b). In good agreement with the seismicity inferences, the dip of the slab beneath the Kermadec trench is significantly steeper than that below the Tonga trench.

**Figure 4.9.** Vertical sections of tomographic images beneath Tonga and Kermadec plotted from western to eastern sides (left to right). Notice the kinking in the slab below northern Tonga (upper panel), but not below Kermadec (lower panel). See caption of Fig. 4.3 for more explanation.
4.2.5 Region 6: Aleutian region

Previous tomographic studies for upper mantle structure beneath parts of the Aleutian islands have been conducted by e.g. Engdahl & Gubbins (1987) and Abers (1994). Engdahl & Gubbins (1987) indicated that subducted slab below the central Aleutian islands extends well below the deepest seismicity and reaches a depth of about 400 km. In this thesis, a three-dimensional $P$-velocity model has been produced for mantle structure beneath the entire Aleutian islands down to 1600-km depth. Our results confirm the observation by Engdahl & Gubbins (1987), but also reveal that the subducted slab is deflected in the transition zone implying that a roll back or retrograde motion of the slab has occurred causing oceanward trench migration and possible concurrent back-arc spreading.

In the upper mantle, the image of slab parallels the present-day plate boundary (Fig. 4.10a). The result of the spike tests suggests that structure along the plate boundary is reasonably resolved (Fig. 4.10d). In the transition zone, an approximately east-west trending positive anomaly is imaged (Fig. 4.10b). The southern part of this anomaly is well resolved (Fig. 4.10e). Figure 4.10 (c) displays images of a low-velocity zone beneath the back-arc region and a high-wavespeed anomaly beneath the fore-arc region. The result of the resolution tests shown in Fig. 4.10 (f) suggests that the deep, low-velocity zone is well resolved, but the high-wavespeed anomaly is not.
Figure 4.10. (a-c) Layer anomaly maps depicting structures in the upper mantle, transition zone and lower mantle, respectively. (d-f) Amplitude recovery of input anomalies with horizontal dimensions of 2°×2°, 3°×3° and 4°×4° in the upper mantle, transition zone and lower mantle, respectively (as displayed in the insets in Figs 4.2 d-f). Line A6 in (d) depicts the position of cross section displayed in Fig. 4.11. See caption of Fig. 4.2 for more explanation.
A vertical section across the convex side of the plate boundary i.e. toward the Pacific Ocean is displayed in Fig. 4.11. The image shows clear indications of a deflection of the subducted slab in the transition zone beneath the central Aleutian Islands. The slab may not penetrate into the lower mantle, but seems to buckle just beneath the 660-km discontinuity. The positive anomaly in the lower mantle located in the lower-right side of the cross section is not well resolved (cf. Figs 4.10 c & f).

Figure 4.11. Vertical section of the tomographic image beneath the central Aleutian region plotted from north to south (left to right). See caption of Fig. 4.3 for more explanation.
4.2.6 Region 7: North America

To the best of our knowledge, a high-resolution $P$-wavespeed model for mantle structure beneath north America is not yet available. Our images provide strong evidence for an existing sub-horizontal subducted slab in the upper mantle beneath the study region, which has been discussed previously by e.g Bird (1984) and Humphreys (1995).

A layer anomaly map in the upper mantle (Fig. 4.12a) reveals a northwest striking sharp boundary between low- and high-wavespeed anomalies beneath the Rocky Mountains and the northeastern part of the modeling area, respectively. The resolution test result for the corresponding depth interval (Fig. 4.12d) suggests that most of the low-velocity anomaly is resolved by the data, but the fast anomaly associated with part of the north America shield is not well resolved due to lack of data coverage. In the transition zone (Fig. 4.12b), a positive anomaly trending in a northwest direction is observed below the central part of the study region. A similar feature is imaged in the lower mantle (Fig. 4.12c), but the location is shifted to northeast relative to the anomaly in the transition zone. These features, especially the slab in the transition zone, seem to be well resolved (see Figs 4.12 e & f). The recovery shown in Fig. 4.12 (f) implies that smearing in a northwest direction occurs in the location of the deep slab, but with a significantly low amplitude.
Figure 4.12. (a-c) Layer anomaly maps depicting structures in the upper mantle, transition zone and lower mantle, respectively. (d-f) Amplitude recovery of input anomalies with horizontal dimensions of $2' \times 2'$, $3' \times 3'$ and $4' \times 4'$ in the upper mantle, transition zone and lower mantle, respectively (as displayed in the insets in Figs 4.2 d-f). Line A7 in (d) depicts the position of cross section displayed in Fig. 4.13. See caption of Fig. 4.2 for more explanation.
A vertical section normal to the observed deep, positive anomaly is displayed in Fig. 4.13. The image clearly reveals the existence of a sub-horizontal slab underlying the Rocky Mountain Foreland and Great Plains of the western United States and suggests that the sub-horizontal slab is laid down on top of the 410-km discontinuity. This slab appears to turn down beneath the Rocky Mountains, where it then penetrates into the lower mantle. This sub-horizontal slab may be associated with Farallon lithosphere, for which the former Farallon plate has been moving northeast (Bird 1984).

![Diagram of P-wave speed relative to ak135](image)

**Figure 4.13.** Vertical section of the tomographic image beneath North America plotted from southwest to northeast (left to right). See caption of Fig. 4.3 for more explanation.
4.2.7 Region 8: Central America including Caribbean

The three-dimensional mantle structure beneath this study region has been investigated in detail by Van der Hilst (1990). He used not only $P$, $pP$ and $pwP$ phases as used in this study but $PP$ phases as well. Therefore, he may obtain a better seismic ray illumination but not necessarily better resolution, because of the limitations in the quality of $PP$ picks. He demonstrated the existence of a high-wavespeed anomaly associated with subduction at the Mid American trench and beneath the eastern part of the Caribbean plate (see also Van der Hilst & Spakman 1989). Our inversion results also reveal these structures and indicate that the upper mantle slab, which may correspond to the subduction of the Cocos plate has not been clearly imaged as a higher-than-average wavespeed structure.

Structures in the upper mantle seems to be dominated by a negative anomaly underlying the Caribbean plate (Fig. 4.14a). High-wavespeed features in this depth interval beneath the northern part of the study region is not resolved, since the corresponding recoveries of impulsive anomalies are poor (Fig. 4.14d). Van der Hilst (1990) also noted that the upper mantle structure beneath central America is not well resolved. In the transition zone (Fig. 4.14b), our inversion reveals fast-wavespeed anomalies below the western and eastern parts of the Caribbean plate; these structures are reasonably resolved (see Fig. 4.14e). Our image shown in Fig. 4.14 (c) strongly suggests the existence of a deep, positive anomaly striking north-south beneath the middle of the modeling area. The result of the resolution tests (Fig. 4.14f) suggests that this deep anomaly is well resolved. This structure is in excellent agreement with an $S$-wave velocity model produced by Grand (1994) and may form a part of the deep Farallon slab, which stretches beneath north and central America. A further discussion on this deep anomaly is given in chapter 6 along with the presentation of our new high-resolution global $P$-wave model.
In this sub-section, we present two vertical sections A8 and B8 which are approximately perpendicular to the western and eastern plate boundaries of the Caribbean plate, respectively (Fig. 4.15). These two cross sections confirm the observations by Van der Hilst & Spakman (1989) in which a deep, positive anomaly is clearly imaged beneath the Mid America trench and a subducted slab trapped in the transition zone is observed beneath the eastern part of the Caribbean plate. The results reveal that upper mantle anomalies beneath the Mid America trench are not imaged as higher-than-average velocity structures, but are still faster than the surrounding anomalies. However, this
character of this image depends strongly on the used of the \textit{ak135} model as a reference because the quantity plotted is the velocity perturbation relative to the reference model. Consequently, if a slower reference model were used, the image would have enhanced the fast slab associated with the subducting Cocos plate through its faster velocities than its surroundings.

Figure 4.15. Vertical sections of the tomographic images beneath central America including the Caribbean region plotted from western to eastern sides (left to right). See caption of Fig. 4.3 for more explanation.
4.2.8 Region 9: South America

For the first time, Engdahl et al. (1995) produced high-resolution P-wave tomographic images for the subducted slab structure below western south America. As expected our results are in agreement with theirs, since we have used the same data set prepared by Engdahl et al. (1997). However, the present study which has incorporated global inversion combined with the use of gradient damping (which was not used by Engdahl et al. 1995) seems to have successfully minimized the mapping of signal from outside the volume under study and have produced a smoother model. This may help investigators interpret our model somewhat more easily.

Layer anomaly maps presented in Figs 4.16 (a-c) reveal slabs indicated by areas of higher-velocity material. In the upper mantle (Fig. 4.16a), the slab approximately parallels the west coast line of south America. In Fig. 4.16 (b), the fast-wavespeed structure in the transition zone is shifted to the east relative to the upper mantle slab implying that the slab associated with the Nazca plate has been subducted in a northeast direction. The broadening in map view of the slab in this depth interval below the southern part of the study region is consistent with the result by Engdahl et al. (1995). Figure 4.16 (c) provides evidence for the existence of a northwest striking, deep anomaly. Results of the tests using synthetic data (Figs 4.16 d-f) suggest that input anomalies are recovered in most regions for which fast-wavespeed features are imaged. However, anomalies below some regions away from the trench are not resolved due to the decrease in sampling. Engdahl et al. (1995) observed some east-west smearing along ray paths to stations in eastern America. This smearing is likely to also occur in this study, but with low amplitude.
Figure 4.16. (a-c) Layer anomaly maps depicting structures in the upper mantle, transition zone and lower mantle, respectively. (d-f) Amplitude recovery of input anomalies with horizontal dimensions of 2° x 2°, 3° x 3° and 4° x 4° in the upper mantle, transition zone and lower mantle, respectively (as displayed in the insets in Figs 4.2 d-f). Lines A9 and B9 in (a) depict the position of cross sections displayed in Fig. 4.17. See caption of Fig. 4.2 for more explanation.
Two vertical sections are displayed in Fig. 4.17; cross sections A9 and B9 are positioned close to cross sections 3d and 3e in Engdahl et al. (1995). In cross section A9, the slab seems to break through the 660-km discontinuity and is deflected horizontally in the uppermost lower mantle in good agreement with the image by Engdahl et al. (1995). While cross section B9 clearly displays that most of the subducted slab is trapped in the transition zone before penetrating into the lower mantle. The image suggests that the subducted slab is continuous, even in the region with a gap in seismicity (see lower panel in Fig. 4.17).

**Figure 4.17.** Vertical sections of the tomographic images for region 9 plotted from western to eastern sides (left to right). See caption of Fig. 4.3 for more explanation.
4.2.9 Region 10: Scotia plate

The 3-D mantle structure beneath the Scotia plate located between the south American and Antarctic plates appears to not have been investigated in detail previously. Pelayo & Wiens (1989) have conducted source parameter inversions of $P$ and $SH$ waveforms and amplitudes to derive focal mechanisms and produce a tectonic map of the Scotia region. Their results show e.g. a westward subduction along the Sandwich trench in the eastern part of the region and a spreading ridge trending north-south to the west of the trench.

An earlier study by Minster & Jordan (1978) indicated an absolute eastward motion of the eastern Scotia plate with a rate of 55 mm yr\(^{-1}\). In this study, we have made an attempt to retrieve the 3-D slab structure beneath the modeling space but unfortunately the data set used in the inversion does not have a good resolving power.

Layer anomaly maps given in Figs 4.18 (a-c) reveal some structures, but results of the resolution tests (Figs 4.18 d-f) suggest that these structures are poorly resolved. This is mainly caused by the lack of seismic ray coverage provided by the present available data set. Positive anomalies imaged in the transition zone and lower mantle (Figs 4.18 b & c) may have smeared northwest. Anomalies beneath the eastern part of the Scotia plate may have the highest resolution among features in this model as $P$-ray illuminations along subducted slabs are generally good.
Figure 4.18. (a-c) Layer anomaly maps depicting structures in the upper mantle, transition zone and lower mantle, respectively. (d-f) Amplitude recovery of input anomalies with horizontal dimensions of 2° x 2°, 3° x 3° and 4° x 4° in the upper mantle, transition zone and lower mantle, respectively (as displayed in the insets in Figs 4.2 d-f). Line A10 in (a) depicts the position of cross section displayed in Fig. 4.19. See caption of Fig. 4.2 for more explanation.
Surprisingly the inversion results can still image a westward subducted slab below the Sandwich trench (see Fig. 4.19). This suggests that the southernmost part of the Atlantic plate has been subducted beneath the eastern Scotia plate. The slab seems to be "fingering" where a part of the slab is deflected in the transition zone and uppermost lower mantle, and another part appears to sink almost vertically into greater depths at 30° W. The observed feature is in agreement with a laboratory investigation result by Griffiths et al. (1995) which suggests that a deflected (horizontal) slab may eventually descend into the lower mantle as it is gravitationally unstable.

Figure 4.19. Vertical section of the tomographic image beneath the Scotia plate plotted from west to east (left to right). See caption of Fig. 4.3 for more explanation.
4.3 REMARKS

We have concisely presented nine regional solutions resulting from nine different simultaneous high-resolution regional and low-resolution global inversions. Following are some remarks on the results of each region.

The sharp contrast in seismic velocities observed below Europe (Fig 4.2a) marks the TTZ that is a boundary between the Precambrian craton in eastern Europe and Phanerozoic Europe to the west of the TTZ. Our model supports the existence of negative anomaly in the transition zone beneath the Precambrian craton revealed in the shear-wave models by Zielhuis & Nolet (1994) and Marquering et al. (1996), in which the resolution of deep mantle structure produced by their waveform inversions usually begins degrading. The observation of deep anomaly beneath the Mediterranean region (Fig. 4.2c) may be associated with the subducted Aegean slab, which is most likely to penetrate into the lower mantle (Spakman et al. 1993).

The inversion results for India reveal a subducted slab, which may have caused intermediate depth earthquakes in the Hindu Kush region, and a subducted slab beneath the Gulf of Oman. The images shown in Figs 4.4 (c) and 4.5 also suggest that there is a pronounced, positive anomaly in the lower mantle beneath India. This may well be part of a long, narrow slab stretching from below the Mediterranean to below the Indonesian region, which may be associated with the past subduction of the plate underlying the Mesozoic Tethys ocean. This long, narrow, deep structure (which is also mentioned in the previous chapter) will be discussed in detail in chapter 6.

Vertical mantle sections for region 4 shown in Fig. 4.7 clearly display deflected slabs in the transition zone below south Kuril, Japan and Izu Bonin, but a sinking slab into the lower mantle beneath Mariana. These slab deflections imply that the 660-km discontinuity from beneath south Kuril to beneath Izu Bonin forms a strong barrier for slab penetration into the lower mantle causing slab stagnancy as discussed by Fukao et al. (1992). Van der Hilst & Seno (1993) have attempted to explain the geometry of subducted slabs beneath the Izu-Bonin and Mariana island arcs in term of the amount and
rate of trench migration. From a multi-disciplinary analysis, they argued that differences in migration of these two trenches may have caused the deflection of subducted slab in the transition zone below Izu Bonin, but penetration into the lower mantle beneath Mariana (cf. cross sections C4 and D4 in Fig. 4.7). The total amount and rate of the Izu-Bonin oceanward trench migration was reported to be approximately 3 times larger than those for the Mariana trench (Griffiths et al. 1995).

In good agreement with a previous study by Van der Hilst (1995), the tomographic images for mantle structure below Tonga provide evidence for local deflection of slab in the transition zone beneath the Tongan back-arc region and slab penetration into the lower mantle. The observed pronounced, positive anomaly below this region is consistent with the old subducted oceanic lithosphere (> 120 Ma) and the fast rate of subduction (> 100 mm yr⁻¹); after Van der Hilst (1995). Glancing at the clockwise rotation of the Tonga trench, it looks like the rotation of the Sunda arc presented in chapter 3. However, a detailed investigation suggests that the rotation of the Java trench resulted from a torque produced by the northward movement of India combined with the southward drift of southeast Asia including the eastern part of the Sunda arc which has been colliding with the Australian plate (Widiyantoro & Van der Hilst 1996). While the rotation of the Tonga trench may arise from a "free" oceanward rollback mechanism (Van der Hilst 1995). From a laboratory investigation, Griffiths et al. (1995) suggest that if trench migration is fast compared to the sinking rate of a slab, the slab is temporarily laid down atop the more viscous layer before flowing to greater depth. On the contrary, the slab buckles without deflection if slab descent is fast compared trench migration (e.g. Gurnis & Hager 1988). The above two situations may have applied to northern Tonga and Kermadec (cf. cross sections A5 and B5 in Fig. 4.9); Van der Hilst (1995).

Inspired by the oceanward trench migration processes for Izu Bonin and northern Tonga, we speculate that the structure of subducted slab displayed in Fig. 4.11 implies that the central Aleutian trench has migrated southward forming the present-day plate boundary which is convex toward the Pacific. This speculation is supported by the
approximately east-west trending slab observed in the transition zone beneath the back-arc region (see Fig. 4.10b) in which the position is shifted to the north relative to the upper mantle slab. Judging from the shape of the subducted slab and results from a laboratory experiment by Griffiths et al. (1995), we interpret that the rate of trench migration for the Aleutian islands is significantly faster than the slab descent.

The tomographic images for mantle structures beneath north America strongly confirm the existence of a sub-horizontal slab beneath the western United States reported by previous studies. The cross section displayed in Fig. 4.13 is in excellent agreement with a schematic cross section produced by Bird (1984). Such a sub-horizontal slab geometry may have resulted from the dynamics of rapid subduction of a relatively young oceanic plate (Engebretson et al. 1985; Jarrard 1986). It can also be attributed to slab buoyancy associated with hypothesized thick oceanic crust (e.g. Livaccari et al. 1981), but Tovish et al. (1978) suggest that a low-angle subduction rapidly becomes horizontal owing to viscous stresses even if the slab is not buoyant.

The inversion results for central America suggest that subducted slabs clearly penetrate into the lower mantle beneath the Mid America trench (see Fig. 4.15). However, the inversion did not image the upper mantle anomalies associated with the subducting Cocos plate as higher-than-average velocity structures. This could imply that the reference velocity model used is rather too fast for this particular upper mantle region. The penetration depths of slabs beneath this region will be assessed in detail by using our new high-resolution global compressional velocity model presented in chapter 6. In the next two chapters, we will demonstrate that the leading edge of the deep slab displayed in cross section A8 (Fig. 4.15) may have reached the D" layer.

Our model for lithospheric slab structures beneath south America supports the conclusion by Engdahl et al. (1995) that the subducted slab is continuous over regions with gaps in seismicity (e.g. see cross section B9 in Fig. 4.17). The images provide evidence for locally deflected slabs in the transition zone and in the uppermost lower mantle, and slab penetration into the lower mantle in accord with the results by Engdahl et al. (1995). A shear-wave model by Grand (1994) also suggests that high-velocity
structures exist in the lower mantle beneath the region down to depths of about 1250 km.

The image of the subducted slab beneath the easternmost part of the Scotia plate shown in Fig. 4.19, although it may not be well within the resolution of the data, intriguingly depicts a slab laid down almost horizontally on the interface between transition zone and lower mantle. This suggests that the Sandwich trench has migrated to the east, which is in good agreement with the direction of absolute plate motion in this area predicted by Minster & Jordan (1978). It seems that the envisaged trench migration has been accompanied by back-arc spreading with the spreading ridge trending north-south just west of the Sandwich trench as indicated by Pelayo & Wiens (1989).
Part Four: Global compressional, shear, and bulk-sound velocity models

5 Global P velocity model: a by-product of the imaging of regional structure

5.1 INTRODUCTION

In the previous two chapters, we have presented the results of regional tomographic imaging for ten different areas. We have incorporated an inversion for global P velocity structure in each regional inversion, which has minimized the mapping of signal from distant aspherical mantle structures outside the different mantle volumes under study. Furthermore, the incorporation of the global inversions has produced ten global models revealing major global structures. The tomographic images from one global model to another do not vary significantly. Therefore, in this chapter we only present the average model derived from these ten solutions. By taking the average of the ten global solutions, we hope to optimize the use of sampling; because the global ray coverage used in each of the simultaneous inversions for regional and global structure was not precisely the same. This averaging procedure should also have minimized the artefacts in the final global model.

In the global model parameterization, we have used non-overlapping 5° x 5° cells and have divided the whole mantle into 16 layers (see Table 3.1 for further details). The technique for simultaneous regional and global inversions has been described in chapter 2, in which the global solution (i.e. $s_{out}$ in 2.11) was obtained as a spin-off of the simultaneous inversion. As events and stations inside as well as outside the regional
modeling space have been employed in the inversions, this provides good regional and
also global ray coverage. As a result, we have not only produced the high-resolution
regional models but global models with a reasonably good quality as well.

An assessment of the reliability of global images was also performed at the same
time as the resolution tests for regional structures presented in the previous chapter. We
have used as an input model regularly spaced synthetic anomalies as depicted in the inset
shown in Fig. 5.1 (h). As in the regional model parameterization presented in the
previous chapter, anomalies were also assigned to alternating layers. The results for the
resolution tests presented in this chapter also result from averaging ten different solutions
from inversions of synthetic data.

In general the tomographic images of the average model are consistent with the
results of previous studies e.g. by Inoue et al. (1990) who performed a whole-mantle
inversion using \( P \) wave travel-time data for the first time. Our results clearly reveal long
wave-length structures such as fast anomalies in the uppermost mantle beneath continental
shields and slow anomalies beneath active tectonic regions. However, the images of
subducted slabs seem to be broadened considerably which is likely to arise from the use
of rather coarse cells in the global model parameterization coupled with the application of
gradient damping. The results also provide strong evidence for the existence of long
wave-length features in the D\(^{\prime}\) layer i.e. fast anomalies beneath the regions around the
Pacific and slow anomalies beneath the southern Pacific.

5.2 PRESENTATION OF AN AVERAGE GLOBAL \( P \) VELOCITY
STRUCTURE

In this section, a global model resulting from averaging ten global solutions is presented.
Four horizontal slices together with the corresponding results of resolution tests and five
vertical mantle sections are described in the following two sub-sections.
5.2.1 Horizontal sections

Figures 5.1 (a-d) are layer anomaly maps representing structure in the upper mantle, the transition zone, the mid mantle and the lowermost mantle. The recovery of synthetic anomalies displayed in the inset of Fig. 5.1 (h) for the corresponding four depth intervals are given in Figs 5.1 (e-h).

In the depth range of 110 - 280 km, the layer anomaly map shown in Fig. 5.1 (a) seems to strongly relate to the surface tectonics. The image reveals high-velocity anomalies associated with shields beneath Eurasia, north America, the central and western parts of Australia, and parts of Africa and south America. Pronounced low-velocity features are imaged, in particular beneath the back-arc regions around the Pacific. These structures generally extend down to a depth of about 400 km. However, this may be due partially to some smearing resulting from lack of resolution in the radial direction in this model caused by the coarse depth parameterization used (16 layers). Low-velocity anomalies with small amplitudes imaged beneath the Pacific resulted from smearing of the slow anomalies beneath the back-arc regions and beneath Hawaii arising from the use of gradient damping.

Figure 5.1 (b) depicts structures in P velocity for the depth interval representing the transition zone between the upper and lower mantle. In this depth interval (410 - 660 km), there are some structural changes relative to upper mantle features. In particular, we observe a slow anomaly beneath western Eurasia and some high-wavespeed structures beneath back-arc regions e.g. the Aleutian islands, Japan and eastern Indonesia. The existence of these fast anomalies is consistent with evidence for subducted cold materials beneath the regions. Large-scale fast anomalies are imaged beneath eastern Asia and northeast America. On the other hand, clear slow anomalies are detected below the south Pacific, Tasman Sea and southern Eurasia. These features seem to be in good agreement with the whole-mantle P-wave model by Inoue et al. (1990) and with surface wave studies e.g. by Woodhouse & Dziewonski (1984). These P structures also appear to have some correlation with the \( l = 2 \) anomaly pattern in 3-D structures inferred from free oscillation spectra by Masters et al. (1982).
The image for the mid mantle between the depths of 1250 and 1400 km (Fig. 5.1c) shows two long, continuous fast wave-speed anomalies below southern Eurasia, and north and central America. These two prominent features may be associated with the past subduction of the Izangi and Farallon plates (see e.g. Widiyantoro & Van der Hilst 1996; Grand 1994). Some high-velocity segments are intermittently imaged beneath the western Pacific ranging from Tonga to China, in which slab penetration into the lower mantle was observed in the regional models presented in the previous chapter. Low-velocity anomalies are detected, in particular beneath the central and south Pacific.

The layer anomaly map representing the upper part of the D" layer (2600 - 2750 km) structure is displayed in Fig. 5.1 (d). The image provides evidence for the existence of long wave-length anomalies such as pronounced low-velocity anomalies beneath the southwest Pacific and East Pacific Rise surrounded by large fast-wavespeed anomalies. These results are in good agreement with the results of previous studies e.g. by Dziewonski (1984), Inoue et al. (1990) and Wysession (1996). Lateral variation in amplitude, in particular for the pronounced fast anomaly beneath east Asia and the less distinct positive anomaly below central and south America may be due to the difference in sampling intensity in which lack of sampling tends to give less pronounced amplitudes.

The results of the resolution tests for the four depth intervals in Figs 5.1 (a-d) are displayed in Figs 5.1 (e-h), and suggest that good resolution is only obtained beneath some limited regions. For the upper mantle (110 - 280 km), Fig. 5.1 (e) depicts that structures shown in Fig. 5.1 (a) have good resolution only beneath south and southeast Eurasia, and beneath the United States where seismic station coverage is densest. The pattern of the input anomalies is also recovered below Africa, Tonga and some places along the mid-oceanic ridges, but with weak amplitudes. In the transition zone (410 - 660 km), resolution is comparable to the resolution in the upper mantle (see Fig. 5.1f). However, the resolved areas become larger in the mid mantle (1250 - 1400 km) which represents the best resolved depth interval in the entire model (see Fig. 5.1g); the input anomaly pattern is almost recovered across the entire world. The amplitude recovery for each anomaly varies and is likely to indicate the intensity of sampling. Smearing of the
anomalies occurs beneath some places, in particular below the north and south polar regions. In the D" layer, structures are resolved reasonably well beneath the northern hemisphere but resolution degrades into the southern hemisphere owing to lack of sampling (Fig. 5.1h). As expected, resolution below the polar regions in all depth intervals is rather poor. This is mainly caused by the application of gradient damping weighed to give heavy smoothing in the longitude direction around the polar regions. More comprehensive hypothesis testing for global structure will be presented in the next chapter.

Figure 5.1. Tomographic inversion results for 4 layers. (a-d) Solutions representing upper mantle, transition zone, mid mantle and D' layer structures, respectively.
Figure 5.1. (continued)
Figure 5.1. (e-h) Resolution maps for the corresponding depth intervals shown in (a-d) displayed in terms of amplitude recovery of the input anomalies shown in the inset in the lower-left corner of (h). These anomalies have a 20° x 20° horizontal dimension. Velocity perturbations are set to +3.0% inside the synthetic anomalies and 0.0% outside.
Figure 5.1. (continued)
5.2.2 Vertical sections

The locations of five vertical sections through the mantle across different plate margins are indicated in Fig. 5.2 and the corresponding sections are displayed in Fig. 5.3. The resolution of the tomographic images in Fig. 5.3 will be assessed in detail in the next chapter along with a comparison of these images with results of the high-resolution global $P$-velocity model presented in chapter 6.

![Figure 5.2](image)

*Figure 5.2.* Solid thick lines depict the position of cross sections A-E displayed in Fig. 5.3. Dashed lines indicate major plate boundaries after NUVEL-1 by DeMets *et al.* (1990).

Vertical section A is across the Aegean region and the position is close to the cross section of the high-resolution image shown in Fig. 4.3 of the previous chapter. The global inversion images a deep anomaly associated with the subducted Aegean slab, but could not image the upper part of the slab. This may be caused by the cell size used in the global model parameterization which seems to be rather too coarse, since the size of the
upper mantle slab appears to be significantly smaller than the slab in the lower mantle (cf. the slab in the upper and lower mantle displayed in Fig. 4.3). The upper mantle slab may have been smoothed out by low-velocity structures outside the slab. The high-velocity lid displayed in the upper-right side of the cross section is probably correlated with the Precambrian craton in eastern Europe.

The image displayed in cross section B reveals a high-wavespeed feature associated with the subducted slab beneath Tonga, which is most likely to be deflected in the transition zone. This low-resolution image is somewhat consistent with the high-resolution image shown in vertical section A5 (see Fig. 4.9), but loses the detailed or small-scale structures. In particular, upper mantle structures such as the subducted slabs beneath Vanuatu and Fiji could not be imaged by the incorporated global inversions. Besides the use of the coarse parameterization, this seems to also arise from lack of sampling. In the regional studies, we could get better resolution because we have employed not only finer cells but more sampling as well. This arises from the use of single rays from stations and events inside the study area. Notice that we only have used summary rays in the incorporated global inversion.

Cross section C is for the Mid American region and is located near the position of cross section A8 displayed in Fig. 4.15. The image shows clearly that the slab, which may be associated with the subducted Farallon plate has sunk down to the D'' layer. As in the high-resolution image (see cross section A8), the upper mantle slab seems to be heavily smoothed out by the surrounding low-velocity structures.

Cross section D beneath central Japan also displays a fast-wavespeed velocity from the Earth's surface down to the D'' layer. The continuous slab seems to be interrupted by a negative anomaly in the uppermost lower mantle, but this may not be well within the resolution of the data. The upper mantle slab beneath central Japan is well imaged, because the ray coverage beneath this region is very good. The slab dips gently in the upper mantle and appears to be consistent with the dip of the Wadati-Benioff zone. The image in this particular cross section somewhat depicts a whole mantle circulation of fast down-going slab and low-velocity up-welling. The dip of the fast material in the
lower mantle appears to be inconsistent with the dip of the fast slab in the upper mantle. This is intriguing and will be discussed in the next chapter.

Cross section E taken just east of Tonga-Kermadec is dominated by a low-velocity feature. In the right-hand side of this section, we detect a plume-like structure that is a rather pronounced negative anomaly stretching almost vertically from the D" layer up to the upper mantle. This is in good agreement with the results of previous studies that reveal a prominent slow anomaly beneath the south Pacific region (e.g. Dziewonski, 1984; Inoue et al. 1990; and Fukao et al. 1994).
Figure 5.3. Vertical mantle sections through the global compressional model, from the Earth's surface to the D" layer. (a) Vertical section A, across Aegean, is plotted from southwest (left) to northeast (right). (b) Vertical section B, across Tonga, is plotted from the inner-arc in the west (left) to the outer-arc region in the east (right). (c) Vertical section C, across the convergent margin in central America, is plotted from southwest (left) to northeast (right). (d) Vertical section D, across central Japan, is plotted from northwest (left) to southeast (right). (e) Vertical section E, which parallels the present-day Tonga and Kermadec trenches is plotted from southwest (left) to northeast (right). Open dots indicate earthquake hypocenters from a distance of up to 50 km on both sides of the section plane; the model parameterization in the radial direction is illustrated in the column superimposed at the right-hand side of cross section E; the number plotted above the top-left corner of each panel indicates the length of the cross section at the Earth's surface.
5.3 CONCLUDING REMARK

We have presented the average global $P$-velocity model; the results indicate that the incorporated global inversions have imaged large-scale global structures reasonably well and are comparable to the results of previous studies. The advantage of averaging the ten solutions is that this procedure may have cancelled out or minimized the artefact produced by errors contained in the data. The size of cells used in the global model parameterization is about the same as the cell size used by e.g. Inoue et al. (1990). Therefore, the resolution of the average model should be about the same as the resolution of the model by Inoue et al. (1990). The current resolution seems to be adequate for long wave-length structures, but small-scale structures could not be delineated in detail. For example: slabs are likely to be significantly broadened and lose their detailed morphology.

In the next chapter, we attempt to bridge the gaps in resolution between the regional images presented in the previous chapters and the images described in this chapter by performing a global inversion using smaller cells i.e. $2^\circ \times 2^\circ$ throughout the entire mantle. We will also demonstrate results of a more comprehensive resolution test for global structures.
6 High-resolution global compressional wavespeed structure

6.1 INTRODUCTION

Motivated by the encouraging results of the incorporated global inversions in the regional studies (chapter 5), we have conducted inversions of P-wave travel-time data for the entire mantle structure using finer cells than were used in the previous chapter i.e. $2^\circ \times 2^\circ$ throughout the global model. This represents a useful attempt to apply such a fine parameterization to produce a whole mantle compressional wavespeed structure and provides an important bridge in resolution between the high-resolution regional models presented in chapters 3 and 4, and the low-resolution global model described in chapter 5. Previous studies of global P-wave travel-time tomography using the International Seismological Centre (ISC) data e.g. by Inoue et al. (1990) and Vasco et al. (1993) have used larger cell sizes i.e. $5.625^\circ \times 5.625^\circ$ and $6^\circ \times 6^\circ$, respectively.

We have used a data set re-processed by Engdahl, Van der Hilst & Buland (1997) based on non-linear relocation of more than 80,000 events and phase re-identification of over $10^7$ readings, which is much more self-consistent than the raw ISC data, to produce a global model of the 3-D variation in P wavespeed. The re-processed data set seems to have been crucial in improving the tomographic images. Based on the improved images, we aim to probe global slab structure in detail, and also attempt to resolve the penetration depth of slabs.

In this chapter, we present the results of comprehensive resolution tests to substantiate our interpretations and focus on the structure of global slabs based on the
tomographic images. We discuss our results in correlation with the locations of subducted lithosphere for the past about 100 million years ago (Ma), and investigate the fate of slabs and its implication for mantle convection. In general, the images confirm results of previous global tomographic studies but with greater resolution of mantle structure, particularly beneath subduction zones and in the lower mantle. The superior resolution arises from the improved data set along with the model parameterization and velocity reference model used.

In the uppermost mantle, the images show low-wavespeed anomalies around the Pacific and indicate that shields on major continents have higher than average $P$ velocities. The most pronounced structure in the transition zone is the high-wavespeed anomaly beneath northwest Pacific, in which the corresponding vertical cross sections show that the slab is deflected or stagnant in the transition zone (see Fukao et al. 1992). The results reveal two long, narrow high-wavespeed anomalies in the lower mantle at about 1200 km depth beneath the southern margin of Eurasia, and below central and north America. They are referred as to the Tethys and Farallon anomaly, respectively (see also Van der Hilst, Widiyantoro & Engdahl 1997). We observe large scale high-wavespeed anomalies in the D'' layer, but the connection of the slab in the upper part of the lower mantle to the structure on top of the core-mantle boundary (CMB) is not clear. The images, however, suggest that slabs can sink to the D'' layer e.g. beneath central America and central Japan. On the contrary, in accord with the result by Fukao et al. (1994), we observe an intriguing feature, which is still within the resolution of the data used, that is a slow-wavespeed anomaly that extends from the top of CMB to the Earth's surface just east of Tonga. The resulting tomographic images also provide evidence for deep convective mantle circulation.
6.2 METHODOLOGY AND DATA

In this section, we briefly describe the tomographic method employed to produce the global model and the data set used.

6.2.1 Global model parameterization

In the current global tomographic imaging, for the full mantle down to the CMB, we have used 18 layers with a layer thickness ranging from 100 km in the upper mantle to 200 km in the lower mantle. We have parameterized the whole mantle using uniform 2° x 2° cells with the layer division given in Table 6.1. The average velocity of each layer has been calculated from the ak135 velocity model (Kennett et al. 1995). The number of cells in the model is 180 x 90 x 18. Upon inversion, we have relocated only a single event of an event cluster in order to reduce the number of unknowns as described in chapter 3. The event-cluster size is 1° x 1° x 50 km giving a total of 7297 clusters. This adds 7297 x 4 parameters (see equations 2.4.1 to 2.4.4) leading to a total of 320,788 unknowns to be solved. We have combined ray paths from earthquakes in the same source region to closely spaced stations into a single ray, commonly referred to as a summary ray (see chapters 3 & 4). In this way, we have suppressed noise in the arrival time data set and reduced the uneven sampling of mantle structure by ray paths. The datum (residual time) assigned to the summary ray is the median of all data considered for that summary ray. There is no restriction of the number of rays that contribute to a summary ray, but the minimum number of rays was set to 4 and 3 for rays bottoming above and below a depth of 2000 km, respectively. By using the summary rays, we have reduced the amount of data used in ray tracing considerably.
Table 6.1. Information of the cell layer division and corresponding layer-average velocity of *ak135* model.

<table>
<thead>
<tr>
<th>Layer no.</th>
<th>Depth range (km)</th>
<th>Average velocity (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>0-100</td>
<td>7.35</td>
</tr>
<tr>
<td>2.</td>
<td>100-200</td>
<td>8.14</td>
</tr>
<tr>
<td>3.</td>
<td>200-300</td>
<td>8.45</td>
</tr>
<tr>
<td>4.</td>
<td>300-410</td>
<td>8.84</td>
</tr>
<tr>
<td>5.</td>
<td>410-520</td>
<td>9.55</td>
</tr>
<tr>
<td>6.</td>
<td>520-660</td>
<td>9.98</td>
</tr>
<tr>
<td>7.</td>
<td>660-820</td>
<td>10.99</td>
</tr>
<tr>
<td>8.</td>
<td>820-1000</td>
<td>11.31</td>
</tr>
<tr>
<td>9.</td>
<td>1000-1200</td>
<td>11.62</td>
</tr>
<tr>
<td>10.</td>
<td>1200-1400</td>
<td>11.91</td>
</tr>
<tr>
<td>11.</td>
<td>1400-1600</td>
<td>12.19</td>
</tr>
<tr>
<td>12.</td>
<td>1600-1800</td>
<td>12.44</td>
</tr>
<tr>
<td>13.</td>
<td>1800-2000</td>
<td>12.68</td>
</tr>
<tr>
<td>14.</td>
<td>2000-2200</td>
<td>12.91</td>
</tr>
<tr>
<td>15.</td>
<td>2200-2400</td>
<td>13.14</td>
</tr>
<tr>
<td>16.</td>
<td>2400-2600</td>
<td>13.37</td>
</tr>
<tr>
<td>17.</td>
<td>2600-2750</td>
<td>13.57</td>
</tr>
<tr>
<td>18.</td>
<td>2750-2889</td>
<td>13.66</td>
</tr>
</tbody>
</table>

where α and γ are weights that determine the trade-off between data misfit, model and gradient terms, and B is the gradient damping matrix. Following Hager (1987), we have applied the above minimization problem to a linear system.
6.2.2 Tomographic system of equations

In this global study, as we only employ uniform cells and do not include a regional inversion, the set of equations may be represented schematically in a form which is somewhat simpler than (2.10)

\[
Ax = (L / R) \begin{pmatrix} s \\ r \end{pmatrix} = \delta t
\]

(6.1)

Here, \(L\) and \(R\) are ray-segment lengths in cells defined by the parameterization used and event relocation coefficients, respectively. The slowness perturbations within the entire mantle are given by \(s\), while \(r\) is the corresponding solution vector of \(R\). The inversion allows for the simultaneous determination of velocity perturbations and event relocation upon inversion. To compute \(L\) and \(R\) in order to construct the matrix \(A\), we used the technique presented in chapter 2. The data vector (\(\delta t\)) contains the travel-time residuals.

As described in chapter 2, we have used a conjugate gradient technique to solve the inversion problem i.e. the iterative LSQR method by Paige & Saunders (1982) applied to seismic tomography by Nolet (1985). During the inversion we have imposed two kinds of explicit damping: the first damping (minimum norm) biases toward zero slowness perturbations from a reference model and suppresses the amplitude in regions of poor sampling (Spakman & Nolet 1988). The application of the second (first order or gradient) damping, which biases toward a smooth model, is conducted in spherical coordinate following e.g. Zielhuis (1992).

Applying the two damping factors in the inversion, we search the model \(x\) that minimizes a weighted sum of data misfit, model norm and gradient norm of the solution:

\[
\min [(Ax - \delta t)^{T}(Ax - \delta t) + \alpha x^{T} x + \gamma(Gx)^{T}(Gx)]
\]

(6.2)

where \(\alpha\) and \(\gamma\) are weights that determine the trade-off between data misfit, model and gradient norms, and \(G\) is the gradient damping matrix. Following Nolet (1987b), we have applied the above minimization problem to a linear system
\[
\begin{pmatrix}
A \\
\alpha I \\
\gamma G
\end{pmatrix}
\begin{pmatrix}
x \\
\delta t \\
0
\end{pmatrix}
= 
\begin{pmatrix}
0 \\
0
\end{pmatrix}
\tag{6.3}
\]

where \( I \) is the identity matrix and \( G \) consists of three sub-matrices

\[
G = \begin{pmatrix}
\gamma_v G_r \\
\gamma_h G_\phi \\
\gamma_h G_\theta
\end{pmatrix}
\tag{6.4}
\]

The subscripts \( h \) and \( v \) denote the horizontal and vertical directions, respectively. Here, we can allow different weights for the horizontal and vertical directions by simply assigning different values for \( \gamma_h \) and \( \gamma_v \). The three matrices yield the first derivative with respect to the directions of \( r \), \( \theta \) and \( \phi \), in which the gradient in the spherical coordinate of a quantity \( \psi \) is expressed as:

\[
\nabla \psi = e_1 \frac{\partial \psi}{\partial r} + e_2 \frac{1}{r} \frac{\partial \psi}{\partial \theta} + e_3 \frac{1}{r \sin \theta} \frac{\partial \psi}{\partial \phi}
\tag{6.5}
\]

where \( e_1 \), \( e_2 \) and \( e_3 \) are orthogonal unit vectors associated with the corresponding coordinate directions. Following Zielhuis (1992), we have imposed the most significant gradient damping in the inversions in the horizontal directions and set a small value for \( \gamma_v \). In this way, we reduce the bias of the final model in the radial direction. We did not apply gradient damping to \( R \) (see 6.1) and assigned a very small value for the coefficient \( \alpha \) that weights \( R \), in order not to damp the solution of event relocation severely.

Choosing the values for \( \alpha \), \( \gamma_h \) and \( \gamma_v \) is a crucial step. Generally the choice of damping is \textit{ad hoc}. One needs to exercise some discretion before deciding a certain level of value for the weight that is assigned to a particular damping class. To do this, we have conducted a series of inversions of synthetic data and have chosen the damping parameters (weights) that suppress noise considerably, but do not suppress the amplitude of signal too much. Furthermore, we also take into account the variance reduction upon inversion produced by the use of different damping parameters.
6.2.3 Data

The data used in this study are travel-time residuals derived from observed arrival times following event relocation and phase re-identification (Engdahl, Van der Hilst & Buland 1997). The arrival times used were those reported to ISC from 1964 to 1992, and to the National Earthquake Information Center (NEIC) from 1993 to 1995. This data set has been augmented by the available data from the Australian SKIPPY project (Van der Hilst et al. 1994). Engdahl et al. (1997) have used robust statistics, an improved global travel-time model \textit{ak135} (Kennett et al. 1995), which is also used as the reference model in this study, and the arrival times of first-arriving regional and teleseismic \textit{P} phases, regional \textit{S} phases, depth phases ($pP$, $pwP$, and $sP$) and the \textit{PKPdf} branch to relocate all teleseismically well-constrained earthquakes. Their procedure is iterative and non-linear, and minimizes depth errors and the mapping of source heterogeneity into mislocation, thereby creating a powerful database of \textit{P}, $pP$ and $pwP$ residuals with less contamination than other data sets which is very suitable for use in tomographic inversions.

We have used 68,134 events recorded at a subset of 3541 stations worldwide (Fig. 6.1). We selected only those data for which the absolute travel-time residual for \textit{P} waves is less than or equal to 5.0 seconds; all epicentral distances are considered. We have used more than $7.0 \times 10^6$ data to construct summary rays (see sub-section 6.2.1). The number of travel-time residuals leading to the linear system of equations is 491,137. The use of the two kinds of damping upon inversion, however, adds many more equations. The norm damping adds a further 320,788 equations and gradient damping an additional set of equations, with 3 equations for each cell i.e. 291,600 equations, in all. Thus the total number of equations to be solved is 1,686,725; which can still be accomplished within a powerful work station such as a Sun Ultra Sparc machine with 512 Mbyte RAM.
Figure 6.1. Distribution of stations and epicentres worldwide used in this study. (a) Open triangles depict stations. (b) Open dots depict epicentres of earthquakes occurred between 1964 and 1995. Locations after ISC (EVB locations).
6.3 IMAGE ASSESSMENT

One of the most crucial steps in tomographic imaging is the assessment of the reliability of the image. As mentioned in previous chapters, we have assessed the reliability of the tomographic images by conducting inversions of synthetic data computed from artificial but known 3-D models using the same ray coverage and linearized theory as used in the real data inversion. Among the techniques most often used are the sensitivity tests (Spakman & Nolet 1988; Humphreys & Clayton 1988) and checkerboard tests (Inoue et al. 1990; Fukao et al. 1992). In assessing the solution quality, we have conducted some checkerboard and target anomaly tests. In this chapter, however, our main purpose is to know how well the major (large-scale) slab structure can be resolved. Therefore, we choose to present the results of hypothesis testing, using input models which have similar spatial characteristics to the slab structure inferred from the real data inversion. We have used the global model produced by Su et al. (1994) as an input model; assuming that the real Earth's structure bears some resemblance to their model, we attempt to resolve it using the data coverage used in the real data inversion. The choice of an existing global model as the input in the hypothesis testing provides a realistic 3-D model of the Earth, which is important, since as indicated by Léveque et al. (1993) the inferences pertinent to image quality depend on the spatial characteristics of the input model used to compute the synthetic data. Despite the limitations of hypothesis tests discussed by e.g. Van der Hilst et al. (1993) and Léveque et al. (1993), the results of resolution tests do provide an important basis for interpreting the final images. We only discuss in detail well resolved structures, and will not speculate on interpreting structural features that are beyond the resolution of the data employed.

We have computed synthetic travel-time residuals by taking the difference between travel times calculated for a 3-D input model (e.g. a slab model or a global model) and for the ak135 1-D reference model. To simulate the effect of errors in the data, we have added random artificial errors to the synthetic data from a uniform distribution between -1.0 and +1.0 seconds for $P$, and between -1.5 and +1.5 seconds for $pP$ and $pwP$. We
then inverted the noisy synthetic data using the same damping parameters as used in the real data inversion to see how well the shape of the input model and the amplitude are recovered. Judging from the amplitude recovery, we can tune the inversion by exploring those damping parameters which give the optimum trade off between bias (i.e. the amplitude recovery and smoothness in the model) and the variance reduction of the data.

In order to test the effect of random noise in the inversion, we have followed a procedure conducted by Grand (1994) that is to invert synthetic data that simulate random travel-time residuals, with a gaussian distribution with an average error of 0.0 second and a variance in the data of 1.0 second. The corresponding inversion results e.g. for layers in the upper mantle and transition zone (Figs 6.2 a & b, respectively) show that the amplitude recovery is quite small relative to the amplitude of the signal produced by the real data inversion presented in the next section. Therefore, we have a strong suggestion that random errors contained in the data do not add severe artefacts to the results of the real data inversion.

The results of inversion of synthetic data calculated using the assumption that the real Earth looks like the model by Su et al. 1994 (see Fig. 6.3) give some important insights for the interpretation of the final model presented in the following section. Figures 6.3 [1-5]a contain the input model representing the structure in the upper mantle down to the lowermost mantle. The corresponding results after inversion given in Figs 6.3 [1-5]b show that major anomalies can be recovered well from the data, but the amplitude is suppressed significantly. In general, the inversion can provide better resolution for structures beneath the northern hemisphere than those below the southern hemisphere. Since we use only body wave data, most of the structures in the upper mantle beneath the ocean domain cannot be recovered. The domain in which structures can be resolved, does however, become larger at greater depths because the sampling of the mantle is improved.
Figure 6.2. Results of inversion of random noise having a gaussian distribution with 1.0 second predicted error in the data. (a) & (b) represent depth slices in the upper mantle and transition zone i.e. 150 km and 500 km, respectively.
Figure 6.3[1-5]. Resolution test. The numbers [1-5] indicate different layers ranging from the upper mantle to the lowermost mantle. (a) Input model i.e. the model by Su et al. (1994). (b) The corresponding recoveries plotted in the same scale of percentage variation in velocity in (a). Notice the loss of amplitude.
Slice at 500 km depth

Figure 6.3[2]. (continued)
Slice at 1300 km depth

Figure 6.3[3]. (continued)
Figure 6.3[4]. (continued)
Figure 6.3[5]. (continued)
6.4 LAYER ANOMALY MAPS

In this section, we present tomographic images obtained after conducting 50 iterations in the inversion for the re-processed real data. The variance reduction achieved in the inversion is 47%. We display anomaly maps from the inversion in Fig. 6.4 for 5 layers corresponding to those depth intervals presented in the hypothesis testing (Fig. 6.3). For each anomaly image in Fig. 6.4 we display the sampling diagram in order to provide a direct comparison between recovered anomalies and the sampling pattern.

In the upper mantle (Fig. 6.4 [1]), high-wavespeed anomalies are imaged beneath the major continents i.e. parts of Africa, Eurasia, Australia and most parts of north America. Regions of low P-wave speed are observed around the Pacific, along mid-ocean ridges, and beneath Hawaii. The Hawaii anomaly may correspond to the existing hot spot underneath the island chain. The sampling is very poor beneath the oceans, because of the limited station distribution and since surface wave data are not included in the inversion. Whereas regions beneath plate boundaries are sampled well by seismic rays allowing an investigation of upper mantle structure beneath convergent margins in fine detail.

In the transition zone, the highest positive-velocity anomaly is observed beneath northwest Pacific (Fig. 6.4 [2]). Complete examination of layer anomaly maps suggests that this anomaly extends down to below the 660-km discontinuity. This occurs in the region where the results of Shearer & Masters (1992) depict the largest depression of the boundary between the upper and lower mantle. A larger region is sampled by seismic rays in the transition zone than in the upper mantle, but sampling beneath the oceans is still poor. As in the upper mantle, almost the entire mantle regions beneath the oceans that have adequate sampling appear to be slow-wavespeed features.

In the lower mantle (Figs 6.4 [3-5]), the global sampling increases significantly. The most prominent features in the depth interval around 1200 km are the Tethys anomaly i.e. the elongated high-wavespeed structure stretching from the Mediterranean to Indonesia, and the north-south trending high-wavespeed feature beneath north and central
America which has been termed the Farallon anomaly. The image also shows a fast-wavespeed anomaly beneath Tonga and part of northwest Pacific. The vertical extension of such deep slabs will be illustrated by presenting several vertical mantle cross sections in the next section. In the mid lower mantle, at around 2100 km depth, we do not observe pronounced fast-wavespeed structures associated with slabs. Higher than average wavespeed anomalies seem to be scattered in and around the Pacific Ocean domain. In the D'' layer sampling decreases, in particular beneath the southern hemisphere. In accord with the long wavelength structure recognized for the first time by Dziewonski (1984), we observe a large structure of high wavespeed around the Pacific and pronounced slow-wavespeed anomalies in south Pacific (see also e.g. Inoue et al. 1990; Fukao 1992). These two anomalies are well resolved by the inversion (see Fig. 6.3 [5]).

The resolution of major slab structures and the penetration depth of slabs are presented in the following section along with the display of vertical cross sections. The intriguing phenomenon, in which the continuation of slabs from the upper part of the lower-mantle to the lowermost mantle is likely to be interrupted in the mid lower mantle, will be discussed in the last section.
Figure 6.4[1-5]. Tomographic inversion results. Again, the numbers [1-5] depict different layers. (a) Solutions representing mantle structures with contour scales -1.0% to +1.0% and -0.75% to +0.75% relative to *ak135* in the upper mantle and transition zone, and lower mantle, respectively. Again, notice the loss of amplitude. (b) Sampling by *P*, *pP* and *pwP* rays for the corresponding layer. Hereafter, sampling is plotted in logarithmic scale of the total ray length in each block defined by the model parameterization; contour scale: 1 km to $10^6$ km.
Slice at 500 km depth

Figure 6.4[2]. (continued)
Slice at 1300 km depth

Figure 6.4[3]. (continued)
Slice at 2100 km depth

Figure 6.4[4]. (continued)
Figure 6.4[5]. (continued)
6.5 VERTICAL MANTLE CROSS SECTIONS

In this section, we present vertical mantle sections across some convergent margins and a "plume" marked by slow-wavespeed anomaly that originates from the D'' layer and extends almost vertically to the Earth's surface east of Tonga, and describe their resolution. Such vertical sections, extending from the Earth's surface to the CMB, are useful to illustrate the complexity of down-going slabs and upwelling features and may provide and indication of implications for mantle convection. The positions of the cross sections presented are depicted by thick lines (A-E) in Fig. 5.2 of the previous chapter.

A vertical section across the Mediterranean (Fig. 6.5a) shows that the Aegean slab penetrates into the uppermost lower mantle. The penetration depth was assessed by two kinds of slab model (Figs 6.5 c & d) i.e. a slab model as deep as inferred by the result of real data inversion and a deep slab model down to the CMB. The recovery (Fig. 6.5e) demonstrates that the depth of slab penetration is well resolved. Smearing with low amplitude, however, occurs in the down-dip direction following the pattern of sampling (Fig. 6.5b). The recovery of the model of a very deep slab (Fig. 6.5f) suggests that if the real slab has gone down to the D'' layer, the inversion should have been able to resolve it. Note the loss of amplitude just above the CMB; this is most likely due to the uneven sampling, in which rays of seismic phases used leave the modelled slab before sampling the deepest part of the high-wavespeed structure associated with the slab. In such case, where the slab sinks to the lowermost mantle with a steep dip, inclusion of PKP data would be useful.

Figure 6.6 (a) shows a cross section for Tonga, which indicates that the subducted slab is deflected in the transition zone in a way that is in good agreement with the result of regional study by Van der Hilst (1995) and chapter 4. The deflected slab seems to sink to at least 1600 km depth. The depth of penetrating slab was tested in the same way as the assessment of the penetration depth of the Aegean slab. The two different model slabs are displayed in Figs 6.6 (c) & (d). The inversion result (Fig. 6.6e) for synthetic data computed for the model slabs suggests that the penetration depth of the
slab (Fig. 6.6a) is within the resolution of the available data. If the Tonga slab has sunk to the CMB, however, the inversion would have not been able to image the lowermost part of the slab owing to the poor sampling in the lowermost mantle, in the region where the model slab is located (Fig. 6.6b). This seems to be a general problem in the southern hemisphere, where we do not have good data coverage (see Fig. 6.4 [5]b).

Vertical mantle sections C and D in Figs 6.7 (a) and 6.8 (a) lie across central America and central Honshu-Japan, respectively. These two images provide evidence that slabs can sink to the lowermost mantle. As they are located in the northern hemisphere, where we have sufficient sampling, the resolving power of the data is good. Figure 6.7 (a) shows that the slab associated with the subducted Farallon plate has reached the D” layer. The slab in the upper mantle, however, seems to be smoothed out by the pronounced slow-wavespeed anomaly around the seismically active slab. Our regional inversion results using 1° x 1° cells presented in chapter 4 do not clearly reveal a higher than average velocity upper mantle slab either, which suggests that the anomaly in the upper mantle connected with the Farallon slab is not prominent (see chapter 4, the upper panel in Fig. 4.15). The results of hypothesis testing shown in Figs 6.7 (e) & (f) indicate that the deep penetration of the Farallon slab is well resolved. More than that, the thickening of the slab in the mid mantle and the thinning of the slab approaching the D” layer seem to be also resolved. This may provide an answer to the curiosity of those who may envisage that slabs are always broadened when they penetrate into the lower mantle.

Cross section D (Fig. 6.8a) demonstrates that the subducted slab beneath central Japan seems to penetrate into the lower mantle and further sink to the CMB. The image suggests that the slab is deflected in the transition zone before penetrating into the lower mantle and shows that the slab is buckled in the lowermost mantle, which may be due to the increase of mantle viscosity with depth (Bunge et al. 1996). Alternatively, the buckling of the slab may occur after the slab has hit the boundary between the outer core and the lower mantle while flow from above continues (personal communication with G. Davies, 1996). The image of the deep slab beneath the region is in good agreement with the result by Su et al. (1994); see Fig. 6.3[3] (a). The slab like feature in the upper
mantle dipping in the opposite direction against the seismogenic slab (see Fig. 6.8a) is likely to arise from the reported data from a subset of stations in China. The results of the hypothesis testing indicate that the fast-wavespeed feature extending from the Earth’s surface to the D” layer beneath central Honshu is well recovered (Fig. 6.8f).

Cross section E in Fig. 6.9 (a) is intentionally drawn parallel to the Tonga-Kermadec arc in order to intersect a pronounced low-velocity anomaly located to the east of Tonga. Fukao et al. (1994) reported an obvious upwelling in the south Pacific. The cross section displayed in Fig. 6.9 (a) confirms their observation, in which an up-going slow-wavespeed zone extending from the lowermost mantle to the upper mantle is clearly imaged. The inversion also reveals a fast-wavespeed lid associated with part of the westward subducted Pacific plate beneath Tonga. We have conducted resolution tests (see Fig. 6.9c for input model) and the result suggests that the inversion may only have imaged part of the upwelling or “plume” (Fig. 6.9d). Again, this is caused by the poor sampling in the uppermost mantle beneath the Pacific ocean (right-top side of the cross section in Fig. 6.9b).
Figure 6.5. Vertical mantle section through our global model, from the Earth's surface to the CMB, across Aegean. (a) Cross section A is plotted from the south (left) to the north (right); contour scale: -1.0% to +1.0%. Hereafter, open dots superimposed in vertical cross sections depict earthquake hypocentres of magnitude ≥5.5 on the Richter scale, projected from a distance of up to 50 km on both sides of the plane of section. (b) Sampling by $P$, $pP$ and $pwP$ rays; contour scale: 1 km to 10⁶ km. (c) Model slab synthesized from the slab structure displayed in (a); contour scale: -5.0% to +5.0%. (d) Modified from (c) for a slab that sinks to the D" layer. (c & f) Recovery of the model slabs in (c & d); contour scale: -5.0% to +5.0%. Note that the recovery of the deepest part of the model slab, see (f), is reasonably good due to the better sampling beneath the northern hemisphere than that below the southern one. (cf. Fig. 6.6f)
Figure 6.6. Vertical mantle section B, across Tonga, is plotted from the inner-arc in the west (left) to the outer-arc region in the east (right); contour scale: -1.0% to +1.0%. Input models and their recoveries are plotted in the scale between -5.0% and +5.0%, and -5.0% and +5.0%, respectively. Note the poor recovery of the deepest part of the model slab in, see (f), due to the poor sampling displayed in (b). See caption of Fig. 6.5 for more information.
Figure 6.7. Vertical mantle section C, across Central America, is plotted from southwest (left) to northeast (right); contour scale: -0.75% to +0.75%. Input models and their recoveries are plotted in the scale between -5.0% and +5.0%. Note the excellent recovery of the deepest part of the model slab in, see (f), due to the adequate sampling displayed in (b). See caption of Fig. 6.5 for more information.
Figure 6.8. Vertical mantle section D, across Central Japan, is plotted from the inner-arc in the northwest (left) to the outer-arc region in the southeast (right); contour scale: -0.75% to +0.75%. Input models and their recoveries are plotted in the scale between -5.0% and +5.0%. Notice the good recovery along the slab from the Earth's surface down to the D^" layer (f). See caption of Fig. 6.5 for more information.
Figure 6.9. Vertical mantle section east of Tonga. (a) Cross sections E is plotted from south (left) to north (right); contour scale: -0.6% to +0.6%. (b) Sampling by $P$, $pP$ and $PnP$ rays; contour scale: 1 km to $10^6$ km. (c) Model plume synthesized from the plume like structure displayed in (a); contour scale: -5.0% to +5.0%. (d) Recovery of the model plume in (c); contour scale: -5.0% to +5.0%. Note the poor recovery of the structure on the right-top side and the bottom of the cross section owing to the poor sampling in the corresponding locations.
6.6 DISCUSSION

We have successfully imaged global slab structure using the improved quality of EVB data (Engdahl, Van der Hilst & Buland 1997). The current parameterization, however, may not be the most efficient in terms of reducing the dimension of matrices involved in the inversion and the CPU time, but has been derived with only minor modification from the parameterizations used in the simultaneous inversion (chapters 3 and 4).

The results of the resolution tests indicate that the wavespeed images for those regions well sampled by seismic rays have, in general, good resolution. The resolution becomes poorer with increasing distance away from convergent margins due to lack of sampling or irregularity in sampling associated with the worldwide station distribution. The hypothesis test results displayed in Fig. 6.3 show that mantle structure beneath the northern hemisphere is better resolved than that beneath the southern hemisphere. The limited resolution in the South arises from relatively poor station coverage (see Fig. 6.1a) leading to poor sampling beneath the southern part of the world. Any efforts such as the Australian SKIPPY project will be useful to overcome the problem in filling the gaps in sampling.

We have attempted to assess the resolution of depth penetration of slabs carefully. In the previous section (see Figs 6.5 and 6.6), we performed a specific hypothesis test for depth penetration of the Aegean and Tonga slab representing subducted slabs in the northern and southern hemisphere by designing different styles of input models where, firstly the model slab is as deep as imaged and secondly, the model slab goes down to the CMB. The results demonstrate that the penetration depth of the Aegean slab is well resolved (Fig. 6.5). On the other hand, the depth of slab beneath Tonga is not well constrained owing to the poor sampling beneath the deepest part of the imaged slab (Fig. 6.6). In general, the results of the resolution tests suggest that the recovery of the deepest part of steeply sinking slabs, just above the CMB, are rather poor. This is mainly due to the lack of sampling by $P$, $Pp$ and $pwP$ phase data. This problem can be handled by
incorporating seismic phases with steeper take-off angle than that of the phases currently used. For this purpose an inclusion of phases such as PKP would be useful.

Our images suggest that in some places, e.g., beneath central America and central Japan (Figs 6.7a and 6.8a) slabs can sink down to the top of CMB in accord with suggestions by e.g., Davies & Richards (1992) and Wysession (1996). On the other hand, we also observe a slow-wavespeed anomaly that extends from the D" layer to the Earth's surface just east of Tonga in agreement with the result by Fukao et al. (1994). We, however, observe a major gap in high-wavespeed anomalies which might represent slabs in the mid lower mantle at about 2100 km depth. As indicated in Van der Hilst, Widiyantoro & Engdahl (1997), the connection between slabs in the upper part lower-mantle and the long wavelength high-wavespeed anomaly observed in the D" layer is still enigmatic. The gap in the mid lower mantle may relate to a break in subduction (Grand & Engebretson 1992). A phenomenon referred to as wavefront healing (Claerbout 1985; Nolet & Moser 1993) may also play an important role, since which a loss of information about time contained in the initial front of a wave field can occur during long passage through a heterogeneous structure (Gudmundsson 1996). On the other hand, the result of the resolution test (Fig. 6.3 [4]) suggests that the data used in the inversion have a good resolving power, in particular beneath the northern hemisphere. If the real structure in the mid-mantle were to correspond to the large-scale structure in the input model by Su et al. (1994), it would have been resolved by our inversion. This implies that the failure of the inversion to produce long wavelength structure in the mid lower mantle (at 2100 km ± 300 km depth) may represent true information. The continuation of slabs from the upper part of the lower mantle through to the D" layer is likely to occur only in narrow segments (leaky places) for instance beneath central America and central Japan (Figs 6.7a and 6.8a).

The map of past locations of convergent margins based on the subduction history model taken from Lithgow-Bertelloni & Richards (1997) plotted in a Mollweide projection is displayed in Fig. 6.10. Our results, especially the Tethys and Farallon anomalies (Fig. 6.4 [3]a) correlate well with locations of past subduction. Large
discrepancies, however, occur beneath the Pacific ring in the southern hemisphere and the northwestern part of the Pacific ocean domain, where we do not have enough data coverage. Therefore, our discussion is necessarily limited. Beneath the northeast Eurasian region, the images suggest that deep subduction is evident but not as a laterally continuous sheet. Deep subduction is only imaged beneath some short segments beneath the region (see Fig. 6.8a). The imaged long, narrow regions of fast P-wavespeed propagation in the mid mantle represent an important observation and we have tested the reliability of the images using a specifically designed artificial slab as displayed in Fig. 6.11 (a). The recovery shown in Fig. 6.11 (b) suggests that such narrow feature in the lower mantle can be resolved by the data, especially beneath the northern hemisphere.

Figure 6.10. Map depicting the location of subduction zones in a hotspot reference frame during the last 110 Ma; the model of subduction history was taken from Lithgow-Bertelloni & Richards (1997). (Figure: courtesy of Bernhard Steinberger 1996; after Grand et al. 1997)
Figure 6.11. (a) Artificial long, narrow slab structure in the mid mantle with a peak anomaly of +3.0% relative to ak135. Perturbations are set to 0.0% outside the slab. (b) The recovery; notice that the width of the synthetic anomaly is not significantly overestimated and that if the deep structure beneath eastern Eurasia were a laterally continuous sheet, it would have been resolved by the data.
One of the most frequent discussed subjects in seismic tomography concerns the fate of slabs and the consequent implications for mantle convection. Our images indicate that in some places, e.g. Aegean and central America, slabs directly penetrate into the lower mantle. There are two major viewpoints, the concept of whole-mantle convection is opposed to layered-mantle convection which has been argued by those who claim that the 660-km discontinuity is a sharp boundary (e.g. Lees et al. 1983). This disputation has been reconciled by Ringwood & Irifune (1988), who for the first time proposed a hybrid model. They suggested that subducted oceanic plates are trapped in the depth around the 660-km discontinuity, but when mature then they may break through the barrier. The images we have obtained also display that slabs are deflected in the mantle transition zone before penetrating into the lower mantle e.g. central Japan and Tonga. The image of the slab penetration into the lower mantle beneath Tonga may provide good evidence for the hybrid convection model. The slab deflected in the transition zone beneath Tonga seems to have formed a megalith, which is probably a large melange (Ringwood & Irifune 1988) and may have become entrained in the convective circulation of the lower mantle.

Another long debate has been whether the penetration of slabs into the lower mantle has resulted from heat transfer or mass flux. Our results suggest that the leading edge of subducted slabs can reach the D" layer (see Figs 6.7a and 6.8a). Therefore, the distance from the upper mantle is too far to be caused by the heat-transfer process when we take account of the slow rate of thermal diffusion. The proposed sinking rate of slabs is significantly higher than the rate of thermal diffusion which further weakens the dynamic significance of thermal coupling. The deflection of slabs in the transition zone may only be a local, short-lived phenomenon that seems to relate to present-day dynamic processes such as a rapid trench migration (Van der Hilst 1995; Griffiths et al. 1995). Such a deflection on a long time scale, probably does not prevent the mass flux or flow of material across the boundary between the upper and lower mantle. The evidence of deep
slab penetration described above simply requires flow across the boundary between the upper and lower mantle (Van der Hilst, Widiyantoro & Engdahl 1997; Masters 1997).

For future work, which is beyond the scope of this thesis, we believe that it would be desirable to apply the parameterization using natural neighbours proposed by Sambridge et al. (1995). This method known as natural-neighbour interpolation has some useful properties e.g. the ability to represent large variations in the scale-lengths of the interpolated function. This seems to be very suitable as the basis of parameterization for the style of heterogeneity revealed in our global tomographic inversion.
7 High-resolution global shear wavespeed models

7.1 INTRODUCTION

Models of shear-wave velocity heterogeneity of the Earth's mantle on both regional and global scales have been derived from several different approaches. An example of regional studies is available from the work of Zielhuis (1992) who produced S-wave velocity models for the mantle region beneath Europe from delay-time and surface waveform inversions. On the global scale, there have been several models produced using spherical harmonics, recent examples include e.g. Su et al. (1994) and Li & Romanowicz (1996). Another area of imaging of mantle shear velocity structure involves the use of the splitting of free oscillation spectra e.g. Giardini et al. (1987) and Masters (1989). Inversions of S-wave travel-time data for mantle shear structure has been performed for example by Grand (1994) and Vasco et al. (1994). Grand (1994) has made a very detailed study of the S wave distribution in the entire mantle through carefully analysing direct as well as multiple reflected shear waves and ScS waves, and has continually updated his model with more data.

The re-processing of the ISC phase data conducted by Engdahl et al. (1997), in which event relocation and phase association for S and SKS waves have been performed based on the ak135 reference model of Kennett et al. (1995), has enabled us to produce a global shear velocity model which compares favourably with previous results. Although in general S data are relatively more noisy than P data, our results of S data inversion reveal most of prominent mantle structures observed in the inversion of P data presented in the previous chapter. The results depict that subduction zone structures, especially in
the upper mantle and transition zone, are revealed in some detail. The long, narrow structures in the top of the lower mantle beneath southern Eurasia and the Americas revealed by the inversion of $P$ data are also well imaged. The $S$ wavespeed images suggest that these structures extend intermittently from the depth of 900 km to 1500 km or so. Results of the resolution tests indicate that these structures are within the resolution of the $S$ data we have used.

In this chapter, we present results of global inversions of $S$-wave travel-time residuals from different subsets of data i.e. with $S$ phases only and also with SKS phases included in the inversions. The results show that the inclusion of the SKS information significantly improves the model for the lowermost mantle by improving ray path coverage. In this chapter, we will display horizontal and vertical mantle sections from the same depths and locations as used in the presentation of the $P$ model in chapter 6 to provide direct comparisons between the $P$ and $S$ model. The results reveal a high level of correlation between the $P$ and $S$ structure, which strengthens the reliability of the images. For example, the inferred subducted slab expanding from the Earth's surface to the D" layer beneath central Japan (see Fig. 6.8 a) is also clearly observed in the $S$ model. Some differences are observed most likely in regions poorly sampled by seismic rays of one wave type or the other.

### 7.2 DATA AND METHOD

In the inversions we have utilized the reprocessed $S$-wave travel-time data set of Engdahl, Van der Hilst & Buland (1997), and the ak135 velocity reference model of Kennett et al. (1995) as a starting model. As mentioned in the introduction, Engdahl et al. (1997) have conducted event relocation and phase association for $S$ waves and SKS waves which are also based on the ak135 reference model. We note that for the ak135 model the construction of the $S$ velocity model in the lowermost mantle is based on the combination of SKS arrivals and $P$ velocity information in the core.
Engdahl and co-workers (1997) have improved the association of the $S$ phase, particularly at the larger distances near the cross-over with the $SKS$ branch. They have also separated $S$ and $SKS$ by excluding the region of overlap (personal communication with E.R. Engdahl, 1997). For detailed description of their data reprocessing procedure we refer to Engdahl et al. (1997).

For this chapter, three separate inversions have been performed for $S$-wave travel-time data using different data sets as follows:

(i) Only epicentral distances ($\Delta \sigma$) less than 82° for phases associated as $S$ are used. As a result we are restricted to $S$ waves with a turning point above about 2100 km depth, and in consequence we do not have any sampling for the lowermost mantle region.

(ii) All $\Delta \sigma$ are included for phases designated as $S$. For the re-processed data Engdahl and co-workers (1997) have separated $S$ and $SKS$ by excluding the region of overlap near 83°, leading to a zone of limited sampling by turning $S$ rays.

(iii) We follow the first arrival with an $S$-wave character by using $S$-wave arrivals for all $\Delta \sigma$ as (ii) and include epicentral distances less than 105° for phases associated as $SKSac$. As noted above this approach is consistent with the way in which the $akl35$ reference model has been determined. The inclusion of $SKS$ data turns out to improve our shear model particularly for the structure in the $D^\prime$ layer in which the amplitude of velocity deviations are enhanced significantly.

Hereinafter the above three $S$ models are referred to as $S_{FLT}$, $S_{ALL}$ and $S_{SKS105}$, respectively. As the $akl35$ reference velocity model was derived partly using $SKSac$ data from a distance range of 91° - 123° (Kennett et al. 1995) we have also performed an inversion similar to (iii) but have extended the use of $SKSac$ information by including epicentral distances up to 120°. The results of this inversion referred as to $S_{SKS120}$ are given in Appendix 7.

There are 306,144 summary rays for the construction of $S_{FLT}$, 320,752 for $S_{ALL}$, and 347,980 for $S_{SKS105}$. This means that there are 27,228 $SKSac$ summary rays involved. To construct the $S_{SKS120}$ model we have used 351,224 summary rays. In the $S$ data inversions, we have excluded data with absolute travel-time residuals greater
than 7.5 seconds. Although $S$ travel-time residuals can be twice as large as the $P$ travel-time residuals, we restrict the range of the $S$ residuals since we deal with a linearized inversion scheme. The $S$ travel-time residuals can easily reach -10.0 seconds for $Sn$ ray paths travelling through a fast craton, but such a magnitude of travel-time residuals would be rejected as outliers in the linearized inversion.

For the inversion of $S$ data we have used the same inversion procedure employed in the $P$ data inversion presented in the previous chapter including the iterative $LSQR$ algorithm, the same model parameterization (see Table 6.1), and the application of two kinds of damping i.e. minimum norm and gradient damping in a similar way to that described in section 6.2.2. However, we have preferred to damp the event relocation coefficients upon inversion heavily in order to force the $S$-wave travel-time residuals to be more likely to be interpreted in terms of velocity perturbations. As a result of this restriction on relocation and the likelihood that $S$ data contain a larger measurement error component than $P$ data, we have only achieved about 33% variance reduction after conducting 50 iterations for each of the inversions described above, which is smaller than the variance reduction from the $P$ data inversion i.e. nearly 50%.

In the construction of the $S_{SKS105}$ and $S_{SKS120}$ models we assume that the Earth's core looks like the $ak135$ reference model which is constructed to match empirical travel-time observations. Therefore, we do not perturb the solution for the $P$ velocities in the core.

### 7.3 ASPHERICAL MANTLE SHEAR STRUCTURE

In Fig. 7.1 we present the three $S$ models i.e. $S_{FLT}$, $S_{ALL}$ and $S_{SKS105}$ for several depth slices representing structure in the upper mantle, transition zone and lower mantle (Figs 7.1[1]-[5]). Comparing the three models allows us to investigate possible bias that may arise from the cross-over between $S$ and $SKS$ around $83^\circ$, and any improvement in
the model, in particular for the structure in the lowermost mantle, from the inclusion of the core-phase data.

The results of the three inversions show similar structure from the Earth's surface down to the upper part of the lower mantle (Figs 7.1[1]-[3]), except for structure beneath the central Pacific for which the inclusion of SKS reveals regions of slightly faster shear wave propagation than the average. From the mid lower mantle to greater depths we start seeing some significant differences in the models. For example, the low-velocity anomalies beneath south America observed in the S_SKS105 model do not appear in the other two models (cf. Fig. 7.1[4]c with Figs 7.1[4] a & b). Rather large differences in the three models occur in the lowermost mantle as we would expect. In the S_FLT model, as expected, we do not see any structure in the D'' layer as we only used Δs less than 82°. Some anomalies imaged with very low amplitudes may result from seismic rays from very deep earthquakes that sample the cells in the second lowermost layer in the model parameterization and/or from the effects of the smoothing applied to the inversion (Fig. 7.1[5]a). By comparing the images in the S_ALL and S_SKS105 models (Figs 7.1[5] b & c) we can clearly see that the amplitude of velocity deviations is significantly enhanced by the inclusion of the SKS phases.

In the layer centred at 150 km depth (Fig. 7.1[1]) the images from the three models look like the results of P data inversion presented in the previous chapter (cf. Fig. 6.4[1]a). They depict low wavespeeds characterizing major marginal basins and tectonic continental regions, and fast wavespeeds outlining stable continental shields. The S data inversions reveal low-velocity anomalies along mid-oceanic ridges more clearly than the inversion of P data presented in chapter 6.

In the upper part of the transition zone (Fig. 7.1[2]) the inferred structure is somewhat similar with the structure in the upper mantle (cf. Fig. 7.1[1]). This may indicate some degree of vertical smearing in the model particularly for the structure in the upper mantle and transition zone. For example, if viewed in a vertical cross section, the low-velocity anomaly underneath Hawaii will seem to form a plume-shaped structure but this is not well constrained by the data.
The structure in the layer centred at 1300 km depth is dominated by two pronounced, long, narrow zones of fast wave propagation beneath southern Eurasia and the Americas (Fig. 7.1[3]). These features are also prominent in the P model presented in chapter 6 (see Fig. 6.4[3]a). As in our P model, a broader inspection of the models suggest that these two high-velocity anomalies extend intermittently about 400 km above and below the 1300 km depth, and in some places they connect to the sites of present-day plate convergence at the Earth's surface. For example, beneath Indonesia (Widiyantoro & Van der Hilst 1996) and beneath the Mediterranean (Spakman et al. 1993). On the other hand, the long, narrow anomaly beneath the Americas locally connects to a long wavelength structure in the lowermost mantle (see also Grand, Van der Hilst & Widiyantoro 1997; Van der Hilst, Widiyantoro & Engdahl 1997).

In the mid lower mantle (Fig. 7.1[4]) the images show areas of fast wavespeeds beneath the regions around the Pacific. The inclusion of SKS phases begins to pick up low-velocity anomalies beneath Africa, south America and the region east of Tonga (cf. Fig. 7.1[4]c with Figs 7.1[4] a & b). The lowered wavespeeds in these regions are consistent with the recent global shear heterogeneity model of Grand (presented partially in Grand et al. 1997). Note that the long, narrow structures do not appear anymore, and also notice the increment in the amplitude of velocity perturbations from (a) to (c) which may arise from the increasing sampling.

As we would expect, the images in Fig. 7.1[5]a show that the structure in the lowermost mantle cannot be inferred by the inversion of S data with only epicentral distances less than 82°. The results of inversion in which all epicentral distances are included are generally in good agreement with our results from P data inversion (cf. Fig. 7.1[5]b with Fig. 6.4[5]a). The inclusion of SKS information increases the perturbation amplitude of structure in the D" layer significantly (see Fig. 7.1[5]c). The long wavelength high-velocity anomaly beneath the regions around the Pacific and low-velocity feature beneath mid-west Pacific and Africa are consistent with previous global models.
Figure 7.1[1-5]. Five layer anomaly maps representing mantle structures with contour scales -1.0% to +1.0% relative to ak135 throughout the mantle. (a) The S_FLT model. (b) The S_ALL model. (c) The S_SKS105 model. The numbers [1-5] depict different layers in the mantle.
Slice at 500 km depth

Figure 7.1[2]. (continued)
Slice at 1300 km depth

Figure 7.1[3]. (continued)
Slice at 2100 km depth

Figure 7.1[4]. (continued)
Slice at 2700 km depth

Figure 7.1[5]. (continued)
Five vertical mantle sections from the same location as used in the previous chapters (see Fig. 5.2 for the location of the cross sections) are displayed in Fig. 7.2[1] to provide a direct comparison between our $P$ and $S$ models. We prefer to present and discuss cross sections from the $S_{SKS105}$ model as this model has the best sampling among the three model displayed in Figs 7.1[1-5]. We also display the corresponding cross sections from the $S_{ALL}$ model in Fig. 7.2[2] in order to explore any additional structure revealed by the use of the core-phase data. These cross sections depict the complexity of the cold down-welling in the mantle.

The vertical section A (Fig. 7.2[1-2] a), across the Mediterranean region, shows that the subducted slab associated with the subducted Aegean plate penetrates into the upper part of the lower mantle in good agreement with e.g. the result of $S$ delay time regional inversion by Zielhuis (1992) and the image based on $P$ data (cf. Fig. 6.5 a). Intriguingly the $S_{SKS105}$ model as well as the $S_{SKS120}$ presented in Fig. 7A.2 (Appendix 7) depict a fast anomaly in the lowermost mantle, which does not appear in the $S_{ALL}$ model nor the $P$ model (cf. Figs 7.2[2]a and 6.5a). Results of the hypothesis testing using the same ray coverage as used to produce the $S_{SKS105}$ or $S_{SKS120}$ model indicate that this fast feature is resolvable and is not likely caused by the common down-dip smearing along the slab. Thus it is clear that the fast anomaly is imaged because of the inclusion of the core-phase data.

Cross section B beneath Tonga depicts a similar result to that from $P$ data inversion i.e. a kink in the slab (cf. Fig. 7.2[1-2] b with Fig. 6.6 a). However, there is a suggestion from the $S$ models that the deflected slab occurs in the uppermost lower mantle and not in the transition zone. This does not seem to be a sampling problem and results of resolution tests for $P$ and $S$ data indicate that the structure is within the resolution of both data. Therefore, it requires further study in order to be able to explain the difference.

Cross sections C and D beneath central America and central Japan are in excellent agreement with the results derived from $P$ data and strongly support the idea that some slabs can sink to the lowermost mantle (Figs 7.2[1-2] c & d). The high-wavespeed
anomaly dipping through the whole lower mantle displayed in cross section C may represent the slab of the Farallon plate that has subducted over the last 100 Ma or so (cf. Grand et al. 1997). Interestingly, the re-processed S data set starts revealing a fast anomaly in the lower part of the upper mantle and the transition zone associated with the subducted slab beneath central America (see Figs 7.2[1-2] c). Note that the results of P data inversion do not show a fast slab in the upper mantle (cf. Fig. 6.7 a). The image displayed in Fig. 7.2[1-2]d provides evidence that there is a leaky region beneath northwest Pacific which allows a down-going slab to reach the deepest part of the mantle (see Van der Hilst, Widiantoro & Engdahl 1997 for discussion).

The hot upwelling observed beneath the region east of Tonga in the P model is also imaged in the S_SKS105 model but the SKSac information also reveals a fast anomaly with a low amplitude in the lowermost mantle, which perhaps would not be resolved by the inversion of P data nor S data only (cf. Fig. 7.2[1]e with Figs 6.9a & 7.2[2]a). Cross sections E (Figs 7.2[1-2] e) reveal a high-wavespeed lid associated with part of the subducted Pacific plate beneath Tonga consistent with the image based on the P data inversion (cf. Fig. 6.9 a).
Figure 7.2[1]. Vertical mantle section through our global $S_{SKS105}$ model from the Earth’s surface to the D" layer with contour scales -1.0% to +1.0% relative to $akl35$. See Fig. 5.2 for the position of the cross sections. (a) Section A, across Aegean, is plotted from the south (left) to the north (right). (b) Section B, across Tonga, is plotted from the inner-arc in the west (left) to the outer-arc region in the east (right). (c) Section C, across central America, is plotted from southwest (left) to northeast (right). (d) Section D, across central Japan, is plotted from the inner-arc in the northwest (left) to the outer-arc region in the southeast (right). (e) Section E is plotted from south (left) to north (right). Hereafter, open dots superimposed in vertical cross sections depict earthquake hypocentres of magnitude ≥5.5 on the Richter scale, projected from a distance of up to 50 km on both sides of the plane of section.
Figure 7.2[2]. The same with 7.2[1] but for the S_ALL model. See text for more information.
7.4 HYPOTHESIS TESTING

We have conducted resolution tests in two ways: (i) we used the model SH13_WM13 from the Harvard group (Su et al. 1994) as an input model, and (ii) we produced synthetic slab models for hypothesis testing for the penetration depth of a slab into the lower mantle as also performed in the previous chapter. We have added gaussian noise to the calculated travel-time residuals based on the above two models. As $S$ data are usually contaminated more severely than $P$ data we used gaussian noise with the expected error in the data of 1.5 seconds, which is higher than used in chapter 6 i.e. 1.0 second. In this chapter we only present some of the results of the resolution tests but prefer to display and discuss layer anomaly maps at common depths and vertical cross sections from the same locations as presented in the previous chapter. This provides a direct comparison which may represent a more rigorous way to test the reliability of the images. A good correlation between totally different models ($P$ and $S$) can in any case build confidence in the interpretation of the tomographic images (see Grand et al. 1997 for discussion).

Figure 7.3 displays results of the hypothesis testing using the same ray coverage used to construct the $S_{ALL}$ and $S_{SKSJ05}$ models. Here we demonstrate the effects of the inclusion of the SKS data in improving the recovery in the lowermost mantle. The images in Figs 7.3 (b) and (c) show that if the structure in the lowermost mantle looks like that inferred by Su et al. (1994) the inversion can resolve much of the feature, especially when we include the core-phase data. In particular, we note that the long-wavelength anomalies i.e. the slow-wavespeed feature beneath the Pacific and the surrounding positive anomalies are reasonably resolved. Also notice that the amplitude of the input model is much better recovered by the inclusion of the core-phase data (cf. Fig. 7.3b with Fig. 7.3c). The resolutions attainable with the ray coverage employed to produce the $S_{ALL}$ and $S_{SKSJ05}$ models for the upper layers (above the $D''$ layer) appear to be very similar.
Figure 7.3. Results of hypothesis testing. (a) Input model i.e. the model by Su et al. (1994) plotted with contour scales -1.0% to +1.0% relative to ak135. (b) The recovery derived using the ray coverage used to produce the S_ALL model plotted in the same scale of percentage velocity deviation in (a). (c) The same with (b) but derived using the ray coverage used to produce the S_SKS105 model. Notice the significant increase of the amplitude recovery in (c) relative to (b).
The images in Fig. 7.4 show that if the Tonga slab looks similar to the inferred structure (Fig. 7.4a) or has sunk to the lowermost mantle (Fig. 7.4b), the inclusion of the core-phase data would be able to resolve such feature. The recovery shown in Fig. 7.4 (c) indicates that common vertical smearing may not have severely occurred since if the slab is not stretching down to the lowermost mantle the ray coverage used to produce the S_SKS105 model will not produce such vertical smearing. The good recovery of the deep slab model displayed in Fig. 7.4 (d) implies that the incorporation of the SKS data has increased the sampling in the deepest part of the mantle beneath the region, as these data have a steeper take-off angle than that of the direct phases. An inversion using S data only would not be able to resolve the deepest part of the modelled slab. Also, note that the P data inversion presented in the previous chapter cannot resolve the deep modelled slab beneath Tonga (cf. Figs 6.6 d & f).
Figure 7.4. (a) Model slab synthesized from the slab structure displayed in Figs 7.2[1-2] (b); contour scale: -5.0% to +5.0%. Velocity perturbations in the input model are set to +5.0%, +4.0% and +3.0% relative to *ak135* for the slab in the upper mantle, transition zone, and lower mantle, respectively. Low-velocity anomalies outside the slab are set to -5.0%, -4.0% and -3.0% in the upper mantle, transition zone, and lower mantle, respectively. (b) Modified from (a) for a slab that has sunk to the lowermost mantle (c & d) Recovery of the model slabs in (a) and (b), respectively, derived using the ray coverage used to produce the *S_SKS105* model; contour scale: -5.0% to +5.0%. Note that the low-velocity features in the upper mantle do not smear horizontally and that the recovery of the deepest part of the model slab is reasonably good due to the inclusion of *SKS* phases. Also, notice the loss of amplitude.
7.5 DISCUSSION

We have presented three global S models for 3-D structure in the mantle, derived from S data only, and from S and SKSac data. We remark that the fast anomalies associated with the subducted slabs from the upper mantle down to the upper part of the lower mantle are well imaged in all models. The inclusion of the core-phase data has turned out to increase the amplitude of the 3-D structure in particular in the lowermost mantle. Figure 7.5 displays some comparisons of the maximum and minimum velocity deviations plotted against depths as well as the r.m.s. perturbations for each layer from the three S models. These plots clearly depict that the amplitude and r.m.s. of the velocity deviations in the lower part of the lower mantle increases from (a) to (c) i.e. from the $S_{FLT}$, $S_{ALL}$ to the $S_{SKS105}$ models.

From the comparison of the above three models there is no suggestion that there is a problem with the cross-over between S and SKS near 83°. This confirms that Engdahl and co-workers (1997) have successfully conducted the association of the S phase particularly at the larger distances near the cross-over with the SKS branch in which they have partially separated S and SKS by excluding the region of overlap (see also Engdahl et al. 1997).

Comparisons between our S models presented in this chapter and the P model in the previous chapter indicate a high level of correlation for many large-scale and smaller-scale structures in the mantle, in particular down to depths of 2000 km or so. Both models reveal that the structure in the mid mantle is dominated by the pronounced, long, narrow regions of high-velocity features beneath southern Eurasia and the Americas. As in the case of the P model, from the global S-wave travel-time inversion we cannot explain the nature of the connection of the long, narrow slabs in the mid mantle and the long wavelength high-wavespeed anomaly on top of the core-mantle boundary. Van der Hilst, Widiyantoro & Engdahl (1997) indicate the possibility of the existence of a lower
mantle transition zone in the depth interval of around 1800-2300 km. The observed broadening in the fast-wavespeed regions in the lowermost mantle is probably caused by the significant increase in viscosity as suggested by e.g. Bunge & Richards (1996). Comparisons with other studies based on spherical harmonic expansion, e.g. Li & Romanowicz (1996) and Su et al. (1994), and the study using the differential times of diffracted $S$ and $SKS$ by Kuo & Wu (1997) suggest that the structure in the lowermost mantle of the S_SKS105 model is somewhat consistent with these results.

The inferred two long, narrow structures at a depth of around 1300 km represent an important observation. Although similar structures are also revealed by previous studies e.g. Su et al. (1994) from spherical harmonic studies, our images are significantly narrower than previous results. In map view the two long, narrow structures have a width of about three to four times of the cell size and a length of around ten times of the width. The block size used, i.e. $2^\circ \times 2^\circ$, does not seem to control the shape of the long, narrow slabs. Our model is in good agreement with the latest result of Grand who has carefully incorporated travel-time data he has measured himself from multiple reflected phases to get better sampling in particular beneath the ocean regions. Grand’s result plotted in the same style as ours (Fig. 7.6) suggests that the long, narrow structure beneath the Americas is more focused (thinner) than ours but he infers a broader structure for the long, narrow feature beneath eastern Europe to Indonesia (cf. Fig. 7.1[3]). Also, we find a similarity in terms of the increase in the velocity deviation in the lowermost mantle between the $S_{SKS105}$ model and Grand’s model (partially presented in Grand et al. 1997) but some disagreements as well, which may arise from the use of $SH$ arrivals by Grand, whereas much of the reported $S$ arrivals are likely to be $SV$, as is $SKS$ (Widiyantoro, Kennett & Van der Hilst 1997). Other differences most likely result from the use of different wave type leading to different sampling for the two data sets employed.

The global images provide a snapshot of the whole convection system in the mantle; this allows the ultimate nature of descending slabs and their effect on the entire convection pattern of the lower mantle to be addressed in more detail than the inference
from the results of regional tomographic studies or any regional seismic study (see Grand
et al. 1997 for further discussion). From the images there is a suggestion that beneath
many convergent margins fast anomalies are continuous across the 660-km discontinuity
(e.g. cross sections A and D in Figs 7.2[1-2] and also the regional inversion results
presented in chapters 3 and 4). The observed continuity in the slabs across the 660-km
discontinuity may require a flow across the seismic discontinuity, which argues against
the worldwide mantle stratification by a flow impeding interface at 660 km. Slab
deflection observed locally e.g. by Fukao et al. (1992), Van der Hilst (1995), and
Widiyantoro & Van der Hilst (1997) does not seem to be wide-spread phenomena in our
planet. The descending slabs can reach a depth of at least 1600 km and in some places
even much deeper than that. For example, cross sections C and D in Fig. 7.2 indicate
that the fast anomalies sink through the entire lower mantle and reach the core-mantle
boundary. These observations which argue against slab stagnation above 1100 km depth
(Wen & Anderson 1995) would be well explained by a flow in the deep mantle, since the
distance to the upper mantle is too far for them to result from conductive cooling alone;
the discussion of the evidence for deep mantle circulation is presented in detail in Van der

The results of S data inversion as well as P data indicate that the data can resolve
cold down-going slabs remarkably well. However, the inversions cannot reveal hot
upwelling (plumes) because they are generally located in regions which are unlikely to be
well sampled by seismic rays. The inclusion of multiple reflected phases by Grand
(1994) seems to be still not sufficient to allow the construction of images from which it
might be possible to explain the nature of upward return flows from the lower mantle to
the upper mantle in some detail. Installing seismic stations in places where upwelling
may occur as in the Japanese initiative, the 'South Pacific Broad-band Seismic Network'
(SPANET) project, is a development to be encouraged.
Figure 7.5. (a-c) Plots of maximum and minimum amplitude of velocity deviations of the S\_FLT, S\_ALL and S\_SKS105 models, respectively. (d-f) Plots of the r.m.s. velocity deviations of the three models. Note that all values are plotted at the central of each layer defined by the model parameterization (see Table 6.1 in the previous chapter).
Figure 7.6. Global S model by Grand plotted in same style as earlier plots (Model: courtesy of S. Grand 1997). Layer anomaly map depicting structure in the mid mantle; contour scales -1.0% to +1.0%. Notice the existence of fast slabs beneath the southern margin of Eurasia and the Americas (see text for discussion).
The presentation of the global S models in this chapter reveals a good correlation between major structures based on the P data (presented in the previous chapter) and S data inversion, especially in the mantle down to depths of the order of 2000 km, which helps to build further confidence in the reliability of the images. However, there are some significant differences in the structures in the lowermost mantle revealed in our P and S models which will need further investigation.

7.6 REMARK ON THE INCLUSION OF THE CORE-PHASE DATA

From the comparison between the S_ALL and S_SKS105 models through displaying the horizontal and vertical sections through the mantle, there is a clear suggestion that the incorporation of the core phases has enhanced the amplitude of velocity anomalies and has revealed some more structures arising from improved sampling (cf. Fig. 7.1[5]b with 7.1[5]c, and the lower parts of the cross sections in Fig. 7.2[1] with those in Fig. 7.2[2]). However, we need to be cautious in interpreting the S_SKS105 model especially for the structure at the base of the model for the inclusion of the core-phase data may produce some degree of vertical smearing. This sort of smearing could be potentially overcome by incorporating further information such as from the diffracted S waves at the core-mantle boundary (see Kuo & Wu 1997 for a detailed description).

The inclusion of SKSac data up to 120° as shown in Appendix 7 (the S_SKS120 model) generates very similar structures throughout the mantle to those revealed by the S_SKS105 model. This is likely to be caused by the fact that there are not many summary rays added between 105° and 120° i.e. only 3244 rays. It also suggests that there has not been any significant bias introduced into the S wavespeed inversion in the mantle from our limited knowledge of the P wave structure at the top of the outer core. Note that the rays associated with the construction of the S_SKS120 model sample the outer core more deeply than those of the S_SKS105 model.
Appendix 7

As mentioned in section 7.2 the ak135 reference velocity model was derived partly using SKSac data from a distance range of 91° - 123° (Kennett et al. 1995). We have performed an inversion using the SKSac information up to 120°. The results of this inversion referred as to S_SKS120 are given below. This model is meant to be compared with the S_SKS105 model presented in this chapter.
Figure 7A.1. Five layer anomaly maps from the $S_{SKS/20}$ model for depth intervals presented in Fig. 7.1 with contour scales -1.0% to +1.0% relative to $akl_{35}$ throughout the mantle. The letters (a-e) depict different layers in the mantle i.e. at depths of 150, 500, 1300, 2100 and 2700 km.
Figure 7A.2. Vertical mantle section through the global S_SKS120 model from the Earth's surface to the D" layer with contour scales -1.0% to +1.0% relative to ak135. See Fig. 5.2 for the position of the cross sections. (a) Section A, across Aegean, is plotted from the south (left) to the north (right). (b) Section B, across Tonga, is plotted from the inner-arc in the west (left) to the outer-arc region in the east (right). (c) Section C, across central America, is plotted from southwest (left) to northeast (right). (d) Section D, across central Japan, is plotted from the inner-arc in the northwest (left) to the outer-arc region in the southeast (right). (e) Section E is plotted from south (left) to north (right). Hereafter, open dots superimposed in vertical cross sections depict earthquake hypocentres of magnitude ≥5.5 on the Richter scale, projected from a distance of up to 50 km on both sides of the plane of section.
Figure 7A.3. (a) Plot of maximum and minimum amplitude of velocity deviations of the $S_{SKS120}$ model. (b) Plot of the r.m.s. velocity deviations of the same model. Note that the maximum and minimum amplitude of velocity deviations, and the r.m.s. velocity deviations are almost the same with those of the $S_{SKS105}$ model (see Fig. 7.5).
8 Global bulk-sound and shear velocity models from joint inversion of $P$ and $S$ data

8.1 INTRODUCTION

Attempts to produce global compressional and shear models have been made since the mid-1980s, for instance by Dziewonski (1984) and Woodhouse & Dziewonski (1984 and 1986). Early suggestions that $P$ and $S$ velocity variations can be postulated to be similar were based on comparison of $P$ and $S$ models derived from seismic data sets with different periods (e.g. see Dziewonski & Woodhouse 1987). As a result of the different sampling by $P$ and $S$ paths, the derived models are reliable in different mantle regions (see also Robertson & Woodhouse 1996).

In recent global studies, some investigators have tried to solve the above problem by jointly inverting $P$ and $S$ arrival-time data from similar $P$ and $S$ data coverage. For example, Vasco et al. (1994) construct mantle compressional and shear models using the International Seismological Centre (ISC) arrival-time data set. A relatively high level of correlation between their $P$ and $S$ models is encouraging. More recently, Robertson & Woodhouse (1995 and 1996) jointly invert $P$ and $S$ data sets of the same period. They also use the ISC catalogue of short-period arrival times. The similarity of the resolutions of the resulting models arising from the use of approximately the same sampling by $P$ and $S$ seismic rays has enabled them to compare structures in terms of both geographical location and amplitude. Robertson & Woodhouse (1996) provide evidence of proportionality between $P$ and $S$ models in a low-order spherical harmonic expansion, which gives physical significance to the values of parameter representing the ratio of
relative $S$ to $P$ velocity heterogeneity. They indicate that this parameter can be used to constrain the physics of deep mantle materials.

For this chapter, we have established a procedure to jointly invert $P$ and $S$ travel-time residuals to generate full-mantle models for bulk-sound speed and shear wavespeed using an iterative partitioned scheme which is suitable for a powerful workstation. We have used $P$ and $S$ data from more or less the same ray coverage and although some important sampling may have been sacrificed, we hope to get better bulk-sound velocity models in the sense that they are derived from $P$ and $S$ models with about the same resolution.

Encouraged by the results of the inclusion of core phases presented in chapter 7, we have included information of $SKS_{ac}$ with epicentral distances less than $105^\circ$. We have then carried the following steps using the same $P$ and $S$ travel-time data sets:

(i) Inversion of $P$ data only using an iterative $LSQR$ algorithm employed in the previous chapters (hereinafter referred to as $P_{LSQR}$ model).

(ii) Inversion of $S$ and $SKS_{ac}$ data also using the iterative $LSQR$ algorithm ($S_{SKS_{ac}}_{LSQR}$ model).

(iii) Direct extraction of bulk-sound speeds from models in (i) and (ii), referred as to the $Q_{SKS_{DIRECT}}$ model.

(iv) Joint inversion of $P$ and $S + SKS_{ac}$ data to produce bulk-sound and shear velocity models, referred as to $Q_{SKS_{JOINT}}$ and $S_{SKS_{JOINT}}$ models, respectively.

We conduct some comparisons e.g. between the $S_{SKS_{LSQR}}$ and $S_{SKS_{JOINT}}$ models, and between the $Q_{SKS_{DIRECT}}$ and $Q_{SKS_{JOINT}}$ models. The result of these comparisons suggests that the $S_{SKS_{LSQR}}$ and $S_{SKS_{JOINT}}$ models appear to be similar, which may indicate a high level of the reliability of the images since they were produced using totally different inversion techniques. We also find a good correlation between the $Q_{SKS_{DIRECT}}$ and $Q_{SKS_{JOINT}}$ models but detailed comparison of these models suggests that the joint inversion gives more stable estimate of the amplitude of velocity deviations from the reference model as a function of depth in the $Q_{SKS_{JOINT}}$ model than that in the $Q_{SKS_{DIRECT}}$ model.
The results of the joint inversion, in particular the bulk-sound speed model could be of particular significance in characterizing mineralogy and may allow us to investigate the agreements and differences between $P$ and $S$ models in more detail. Much of the experimentation at high pressures provides results which can be directly related to the bulk-sound speed.

8.2 METHODOLOGY

For the first two inversions to produce the $P_{LSQR}$ and $S_{SKS_{LSQR}}$ models noted above we have just employed the iterative $LSQR$ inversion method described in the two previous chapters. To construct bulk-sound velocity heterogeneity models we extract bulk-sound speeds in two different ways:

(i) We have made use of the $P_{LSQR}$ and $S_{SKS_{LSQR}}$ models, and have computed the bulk-sound velocity perturbations using a direct extraction procedure described in 8.2.1.

(ii) We have established and performed an iterative partitioned scheme for a joint inversion of $P$ and $S + SKSac$ data using a technique presented in 8.2.2.

We note that in all of the inversions performed for this chapter we make the assumption that the hypocenters of events are sufficiently well-known so we can work directly with the $P$ and $S$ travel-time data. In this way, we do not have to conduct hypocenter relocation in the inversions and thus reduce the number of unknowns to be solved. Also, notice that we use the global data set which has been extensively reprocessed by Engdahl, Van der Hilst & Buland (1997) in which the hypocenters are already significantly improved from the original catalogues.

8.2.1 Direct extraction of bulk-sound speeds

From knowledge of the compressional and shear wavespeeds we can directly extract bulk-sound speeds, which can be expressed in terms of $P$ and $S$ velocities as follows:
In a homogeneous isotropic body the velocity of compressional elastic waves ($\alpha$) and elastic shear waves ($\beta$) is equal to

\[ \alpha = \sqrt{(\kappa + \frac{4}{3}\mu)/\rho} \]  
(8.1)

\[ \beta = \sqrt{\mu/\rho} \]  
(8.2)

where $\kappa$, $\mu$, and $\rho$ are bulk modulus, rigidity and density, respectively. The bulk-sound velocity ($v_\phi$) relates to $\kappa$ and $\rho$ as

\[ v_\phi = \sqrt{\kappa/\rho} \]  
(8.3)

Therefore from (8.1) to (8.3) we can rewrite

\[ v_\phi = \sqrt{\alpha^2 - \frac{4}{3}\beta^2} \]  
(8.4)

For a reference model with $P$ and $S$ velocities ($\alpha_0$, $\beta_0$) the reference bulk-sound velocity is $v_{\phi_0}^2 = \alpha_0^2 - \frac{4}{3}\beta_0^2$, and then we can construct the 3-D bulk-sound velocity by

\[ v_\phi^2 = \alpha_0^2 \left(1 + \frac{\delta \alpha}{\alpha_0}\right)^2 - \frac{4}{3}\beta_0^2 \left(1 + \frac{\delta \beta}{\beta_0}\right)^2 \]  
(8.5)

and the deviations from the reference model are

\[ \frac{\Delta v_\phi}{v_{\phi_0}} = \frac{v_\phi - v_{\phi_0}}{v_{\phi_0}} \]  
(8.6)

Note that here $\alpha_0$ and $\beta_0$ are taken from the ak135 reference model for $P$ and $S$ (Kennett et al. 1995).

Using (8.5) and (8.6) we have extracted bulk-sound speeds from the velocity deviations of the $P_{LSQR}$ and $S_{SKS_{LSQR}}$ models, in which the resolution of the two
models is likely to be similar due to the use of approximately the same \( P \) and \( S \) ray coverage (see sub-section 8.2.4).

8.2.2 Joint inversion of \( P \) and \( S \) data

For this chapter we have made an attempt to develop an algorithm to jointly invert \( P \) and \( S \) travel-time data to produce bulk-sound and shear velocity models. We have established an iterative partitioned scheme which turned out to be suitable for a powerful workstation e.g. a Sun Ultra Sparc machine. The algorithm we have developed can also be constructed using a partitioned form of Jacobi iteration (see e.g. Conte & De Boor 1980 for discussion). We present the detailed mathematical derivation of the iterative partitioned scheme in Appendix 8A.

We have accomplished the joint inversion using the subroutine \textit{LINBCG}, an iterative biconjugate gradient method from \textit{Numerical Recipes} (Press et al. 1992). This routine solves equations of the form \( Ax = b \) for \( x \), where \( b \) is given, and \( A \) is a square matrix. This method needs an initial guess for the solution which is of particular importance in solving a type of equation that we set up for the joint inversion. As described in Appendix 8A in detail, we deal with a type of equation for which the critical parts can be expressed in the form \( G^\top Gx = b \). We start with the current estimate for the vector \( x \) and then perform the following multiplications: \( Gx = x' \) followed by \( G^\top x' \), so that we in effect consider \( G^\top x' = b \). Notice that we never actually perform the \( G^\top G \) multiplication which takes a large memory space, particularly when we use matrices with a large order as have been employed in this study. An other requirement of the \textit{LINBCG} is a preconditioner matrix, and after some experimentation we have just used the identity matrix, which seems to work well. For further discussion on the iterative biconjugate gradient method we refer to Press \textit{et al.} (1992) and the references therein.

In the joint inversion, we employ a parameterization in terms of bulk-sound and shear velocity and represent the \( P \) wave slowness in terms of the \( S \) wave slowness and the bulk-sound slowness. The merit of using slowness is that the Fréchet derivatives of travel times take a much simpler form than when velocity is used.
8.2.3 Data selection

In order to achieve a similarity in the \( P \) and \( S \) ray coverage we have selected from the global data set assembled by Engdahl and co-workers (1997) only those \( P \) and \( S \) ray paths from the same event and station. With this restriction we end up with a smaller number of \( P \) and \( S \) summary rays than were used in chapters 6 and 7. For the four steps (i) to (iv) mentioned above, we have employed a total of 312,549 summary rays for each wave type to solve for 291,600 unknowns for each wavespeed considered, so that there are 583,200 unknowns in the joint inversion for bulk-sound speed and shear wavespeed. (There are 9453 \( SKS \) summary rays involved which are significantly less than those used in chapter 7 i.e. 27,228 \( SKS \) data).

8.2.4 Resolution from the common \( P \) and \( S \) data set

In this chapter, the potential resolution that can be achieved by the common \( P \) and \( S \) data set has been assessed through test inversions of synthetic data generated based on a periodic model of high-velocity blocks (see inset in Fig. 8.1[1a]). As in the previous chapter, we have added gaussian noise to the calculated residual times with the expected errors of 1.0 second and 1.5 seconds in the \( P \) and \( S \) data, respectively. We display the recovery of the input model for three different depth intervals in the lower mantle in Figs 8.1[1-3]. In general, the resolution is good in most parts of the regions beneath the northern hemisphere. The recovery in oceanic areas at shallower depths is unfortunately rather poor and parts of the regions beneath the southern hemisphere (e.g. south Africa) remain undersampled by seismic rays at all depths. However, notice the relatively high level of similarity of the recovered feature and amplitude attainable with the common \( P \) and \( S \) data set (cf. panels (a) and (b) in Figs 8.1[1-3]).
Figure 8.1[1-3]. Results of the resolution tests. (a) Recovery of the input model displayed in the inset in (a) derived using the ray coverage used to produce the $P_{LSQR}$ model; plotted in percentage of amplitude of the input anomaly. High wavespeeds in the anomalous blocks are set to $+3.0\%$ relative to $akl35$ and $0.0\%$ elsewhere. These anomalies have a $10'' \times 10''$ horizontal dimension and are assigned to alternating layers. (b) The same with (a) but derived using the ray coverage used to produce the $S_{SKS_{LSQR}}$ model. The numbers (1-3) depict different layers in the lower mantle i.e. at depths of 1300, 2100 and 2700 km.
Figure 8.1[2]. (continued)
Figure 8.1[3]. (continued)
8.3 COMPARISON OF THE RESULTING MODELS

In this section, we present and compare results of the LSQR and joint inversions. For the joint inversions, we have conducted five iterations of the partitioned scheme [see *) and **) in Appendix 8A] in which in general the inversions become stable after the third iteration.

We compare the results from the different classes of inversion procedures [the four steps (i) to (iv)] mentioned in the introduction of this chapter. We present horizontal slices through the resulting models for the five layers which have previously been used in chapters 6 and 7, which enables us to provide direct comparisons with the P and S models presented in these two chapters as well. In the following two sub-sections we compare the $Q_{SKS\_DIRECT}$ and $Q_{SKS\_JOINT}$ models, and the $S_{SKS\_LSQR}$ and $S_{SKS\_JOINT}$ models. We also compare the properties of these models in terms of both maximum and minimum wavespeed perturbations in each layer as well as the r.m.s. perturbations in each layer.

8.3.1 Comparison of bulk-sound velocity models

In this sub-section, we compare the bulk-sound model resulting from the direct extraction procedure and from the joint inversion i.e. the $Q_{SKS\_DIRECT}$ and $Q_{SKS\_JOINT}$ models (Figs 8.2[1-5] a and b), respectively. From this comparison there is a clear indication that the amplitude of velocity deviations for both models differs significantly throughout the mantle regions. However, the geographic patterns of the imaged anomalies are very similar. The images for the $Q_{SKS\_JOINT}$ model have smaller amplitude and appear to be smoother than that for the $Q_{SKS\_DIRECT}$ model. This may be due to somewhat different effects of damping in the different inversion schemes, that seems to have suppressed the amplitude of velocity anomalies more in the joint inversion than when using the LSQR algorithm, and is possibly the result of different convergence rates for the two styles of iterative equation solvers (Kennett, Widiyantoro & Van der Hilst; in preparation).
The difference in the amplitude between these two bulk-sound velocity models can be seen even more clearly in Fig. 8.3 which represents a comparison of the properties of the models derived using different approaches. The plot of minimum and maximum velocity perturbations against depth shown in Fig. 8.3 (a) indicates some degree of "ringing" in the amplitude, in particular for the minimum velocity deviations in the layers in the lower mantle. Notice that this problem arises when we extract the bulk-sound speeds directly from the $P_{LSQR}$ model presented in Appendix 8B and the $S_{SKS\_LSQR}$ model presented in the following sub-section. One of the problems with the direct extraction is that subtle differences in ray paths for $P$ and $S$, even though the endpoints are the same, can lead to differences in the amplitude ascribed to cells at the edge of coverage. If there is an imbalance between $P$ and $S$ it will show as either a high or low apparent perturbation in bulk-sound speed depending on whether $P$ or $S$ dominates.
Figure 8.2[1-5]. Bulk-sound velocity anomaly maps from the direct extraction and joint inversion. (a) The Q_SKS_DIRECT model. (b) The Q_SKS_JOINT model. The numbers (1-5) depict different layers in the upper mantle, transition zone and lower mantle.
Slice at 500 km depth

Figure 8.2[2]. (continued)
Figure 8.2[4]. (continued)
Figure 8.2[5]. (continued)
Figure 8.3. (a-b) Plots of maximum and minimum amplitude of the bulk-sound speed perturbations for the $Q_{SKS\_DIRECT}$ and $Q_{SKS\_JOINT}$ models. (c-d) Plots of the r.m.s. velocity perturbations for each of the two models. The values are plotted at the centre of each of 18 layers in the model parameterization.
8.3.2 Comparison of shear velocity models

In Fig. 8.4 we present a comparison between the global shear model resulting from the \textit{LSQR} inversion (the \textit{S\_SKS\_LSQR} model) and from the joint inversion (the \textit{S\_SKS\_JOINT} model). The results of this comparison is very similar to those presented in the previous sub-section, once again the geographic patterns of the two models are nearly the same but the amplitude of velocity deviations for the model resulting from the joint inversion is smaller. There is a suggestion that, compared with the \textit{S\_SKS\_LSQR} model, the \textit{S\_SKS\_JOINT} model looks more like the \textit{P\_LSQR} model particularly in terms of the level of the perturbation amplitude. Again, the spectrum of heterogeneity displayed in Fig. 8.5 shows clearly the difference in the amplitude and the r.m.s. velocity deviations.

From the above comparisons there is a clear suggestion that although we have used the same values for the damping parameters in the \textit{LSQR} and joint inversions, the effects of these parameters turn out to be somewhat different when they act in different inversion schemes. Also, as noted in sub-section 8.3.1, this difference may arise from the result of different rates of convergence for the \textit{LSQR} and \textit{LINBCG} methods.
Figure 8.4[1-5]. Shear velocity anomaly maps from the LSQR and joint inversions. (a) The $S_{SKS\_LSQR}$ model. (b) The $S_{SKS\_JOINT}$ model. The numbers (1-5) depict different layers in the upper mantle, transition zone and lower mantle.
Figure 8.4(2). (continued)
Slice at 1300 km depth

Figure 8.4[3]. (continued)
Figure 8.4[4]. (continued)
Slice at 2700 km depth

Figure 8.4[5]. (continued)
Figure 8.5. (a-b) Plots of maximum and minimum amplitude of the shear wavespeed deviations from the ak135 model for the S_SKS_LSQR and S_SKS_JOINT models. (c-d) Plots of the r.m.s. velocity deviations for each of the two models. The values are plotted at the centre of each of 18 layers in the model parameterization.
8.3.3 Comparison with results of joint inversion without SKS data

To provide further comparison and examine effects or bias that may have been produced by the core-phase data we also have repeated steps (ii) to (iv) but with the exclusion of the core-phase data. The results are referred to as $S_{LSQR}$, $Q_{DIRECT}$, $Q_{JOINT}$ and $S_{JOINT}$ models, respectively, and are displayed in Appendix 8C.

In general, the images and heterogeneity spectra of both results of the LSQR and joint inversions in which the SKS information are included are very close to those for the inversions without the incorporation of the core-phase data (cf. Figs 8.2 to 8.5 with Figs 8C.1 to 8C.4). This similarity may arise from the addition of only less than 10,000 SKS summary rays used in this chapter (see sub-section 8.2.3). From the spectrum of heterogeneity we could observe a slight increase in the amplitude of velocity perturbations for the results of inversion using SKS data, in particular for the shear structure in the deep mantle (cf. Fig. 8.5 with Fig. 8C.4). However, the geographic patterns of the velocity perturbations differ only slightly in the lowermost mantle.

8.4 DISCUSSION

We have performed inversions using different methods i.e. LSQR and LINBCG, and have found that we achieve a better estimated data variance reduction from the use of the LSQR method. The variance reductions of $P$ and $S$ travel-time residual data from the LSQR inversions are nearly 50% and 35%, respectively. However, we note that the "true" variance reduction computed using 3-D ray tracing through the resulting 3-D model may be worse than the above values of linearized estimations (see Sambridge 1990 for discussion). The variance reductions before and after the joint inversions for bulk-sound and shear speeds are almost the same for both $P$ and $S$ data i.e. about 30%. All the variance reduction is accomplished with 3-D structure since we have not included any hypocentre relocation parameters. The similarity in the variance reductions upon the joint inversions suggests that the coupling terms which contain $P$ and $S$ information seem to
have worked well (see Appendix 8A). Notice that as the right-hand side of the iterative partitioned scheme does not explicitly contain the travel-time data, see *) and **) in Appendix 8A, therefore the calculation of the data variance is accomplished using the Fréchet derivative matrix as described in Note 2 in Appendix 8A.

There is an indication that the bulk-sound velocity deviations derived by the joint inversion procedure are more stable than those based on the direct extraction method. By jointly inverting $P$ and $S$ data and thereby including the shear modulus contribution to the $P$ wavespeed we may have better constrained the resulting $S$ model as well. However, the comparison that shows that the $S_{SKS\_LSQR}$ and $S_{SKS\_JOINT}$ models appear to be very similar in their geographic patterns is encouraging because both models were constructed by means of totally different inversion methods, as mentioned above (see sub-section 8.3.2). Thus, this builds confidence in the reliability of these models.

In the previous chapter, we have discussed that there is generally a high level of correlation between the $P$- and $S$-wave images. The results of the joint inversion to some extent can facilitate the investigation of such similarities or discrepancies between the two images, for such features are likely to be dependent on the pattern of the bulk-sound speed heterogeneity model (Kennett, Widiyantoro & Van der Hilst; in preparation).

Glancing at the results of the joint inversion i.e. the $Q_{SKS\_JOINT}$ and $S_{SKS\_JOINT}$ models presented in Figs 8.2[1-5]b and 8.4[1-5]b, we find that in general the two models look similar. However, when we look at them carefully there are some pronounced discrepancies. For examples: the $S_{SKS\_JOINT}$ model clearly depicts slow wavespeed anomalies beneath oceanic mid ridges, but the $Q_{SKS\_JOINT}$ model does not (cf. Figs 8.2[1]b and 8.4[1]b). Reversed anomalies are also observed e.g. we see a slow wavespeed anomaly in the transition zone beneath north America in the $S_{SKS\_JOINT}$ model, but the $Q_{SKS\_JOINT}$ model reveals a fast wavespeed anomaly (cf. Figs 8.2[2]b and 8.4[2]b). These discrepancies most likely indicate that shear and bulk-sound moduli have different sensitivity, in which thermal heterogeneity seems to be more marked for shear wavespeed than bulk-sound speed. In the mid mantle, we observe that the images of the long, narrow features beneath southern Eurasia and the Americas seem to be rather
diffused in the bulk-sound model but more focused in the shear model. In the D" layer, good agreement between the two models are evident for the large-scale anomalies i.e. pronounced fast and slow wavespeed structures beneath the northwest Pacific and eastern Indonesia, respectively.

One question that is often raised concerns with the size of the true bulk-sound velocity perturbations in the real Earth. From this study alone, we cannot give a satisfactory answer to this question because we have worked with damped inversions, for which it is likely that the resulting perturbations have been suppressed quite significantly. Therefore, we may envisage that the real perturbations are actually larger than what result from the inversions. However, the relative size of the bulk-sound speed deviations and the shear wavespeed deviations should be reliable since they are derived using more or less the same sampling by P and S ray paths in the same joint inversion.

The work of Robertson & Woodhouse (1995) provides evidence for proportionality of P and S heterogeneity in the lower mantle. From the joint inversion we would expect strict proportionality of the P and S velocity deviations if either there was no bulk-sound speed variations or the spatial variation of bulk-sound speed was the same as for shear wavespeed. Robertson & Woodhouse (1996) suggests that for large scale structure (angular order $l < 8$), the ratio of relative S to P velocity heterogeneity varies between 1.7 at the top of the lower mantle and 2.6 at 2000 km depth. The ratio of the r.m.s. S to r.m.s. P velocity deviations [Figs 8.5 (c) and 8B.2 (b)] at the same depth interval varies between 1.25 and 1.54. This ratio is in agreement with the ratio variations calculated from mineral physics by e.g. Isaak et al. (1992) and Karato (1993).

We have linearized all the above inversions about the $akl35$ reference velocity model. The model $akl35$ is a good fit to the empirical travel times constructed from averages of phase times from source and station pairs across the globe for a wide range of phases including $P$, $S$ and $SKS$ (Kennett et al. 1995). Therefore the use of $akl35$ as a base model for travel-time tomography is very sensible and should have the effect of minimizing the potential residuals due to 3-D structure, which allows a linearized inversion to be undertaken. The presence of zones of large scale systematic deviation
from the reference model e.g. the old continental shields in the uppermost mantle, however, indicates that we are reaching the limits of linearized analysis. For future work, we should employ 3-D ray tracing on the updated models, which would represent a major effort even for the number of summary rays in the order of about 300,000 as used in this study.

8.5 CONCLUSION

We have successfully demonstrated that the joint inversion of $P$ and $S$ travel-time residual data works well. The resulting shear models are in agreement with those from the iterative $LSQR$ inversion method. The global $S$ models from the joint inversions may even provide a better representation since as noted above they are not constrained by $S$ data only but by $P$ data as well. The bulk-sound velocity models resulting from the joint inversion are also encouraging. Comparisons with the models extracted directly from $P$ and $S$ models derived separately using the $LSQR$ method suggest that although the joint inversion produces smaller amplitude of velocity deviations, these deviations are more stable and the geographic patterns of the imaged anomalies are relatively well maintained. The smaller amplitude in the images resulting from the joint inversion compared with the $LSQR$ inversion is likely due to the different set up of damping terms in both methods [cf. (6.3) with Note 1 in Appendix 8A].

We have used a certain amount of $SKS$ information so that we need to be cautious about the possibility of a discrepancy in the baseline between $S$ and $SKS$ which would result in a mismatch of the estimates of the travel-time residuals for each phase. However, the comparison presented in sub-section 8.3.3 between the results for $S$ inversions with and without the inclusion of $SKS$ information, suggests that any such bias is small. The results from the combined data set (the $S_{SKS\_LSQR}$ and $S_{SKS\_JOINT}$ models) are very consistent with those parts of the $S_{LSQR}$ and $S_{JOINT}$ models where $S$ data have enough coverage, and the geographic patterns are almost precisely the same.
with slightly larger amplitude of velocity deviations, in particular for structure in the lowermost mantle.

From the model comparisons conducted in this chapter we gain insights that thermal effects would be more marked for shear wavespeed than bulk-sound speed. A pronounced anomaly in bulk-sound may then be more indicative of chemical heterogeneity. Such insights represent the merit of working with the images of the bulk-sound and shear wavespeeds since the sensitivity of shear and bulk moduli to temperature seems to be different. We realize that in-depth interpretations of the bulk-sound models are still desirable and will require further study which will necessitate the incorporation of information from mineral physics. Nevertheless the results of the joint inversion have already provided us with a different way to look at deep structures in the mantle.
Joint inversion for bulk sound and shear velocities

Seismic wavespeeds

For the global studies we have observations of $P$ and $S$ arrival times, and for assumed locations we can derive travel times $t_p$ and $t_s$. From ray theory, the equivalent theoretical times are

$$ t_{p,i} = \int_{\text{ray}_i} \frac{ds}{\alpha}, \text{ and } t_{s,j} = \int_{\text{ray}_j} \frac{ds}{\beta} $$

where the $P$ wave velocity

$$ \alpha = \sqrt{\frac{\kappa + \frac{4}{3} \mu}{\rho}} $$

in terms of bulk modulus $\kappa$, shear modulus $\mu$ and density $\rho$;

the $S$ wave velocity

$$ \beta = \sqrt{\frac{\mu}{\rho}} $$

From knowledge of $\alpha$ and $\beta$ we can extract the bulk sound speed

$$ v_o = \sqrt{\frac{\kappa}{\rho}} = \sqrt{\alpha^2 - \frac{4}{3} \beta^2} = \alpha \sqrt{1 - \frac{4 \beta^2}{3 \alpha^2}} $$

which is of particular significance in characterizing mineralogy, since it is accessible from laboratory experiments at high temperatures and pressures.

Joint tomography for bulk-sound and shear wavespeeds

As a measure of data fit for the tomography problem consider
\[ P_d = (d - d_o) \mathbb{C}_d (d - d_o) \]

where \( d = (t_p, t_s) \) are the travel times and \( d_o = (t_{po}, t_{so}) \) are the predictions from the reference model, and we partition the data covariance matrix

\[
\mathbb{C}_d^{-1} = \begin{pmatrix} C^{-1}_{pp} & C^{-1}_{ps} \\ C^{-1}_{sp} & C^{-1}_{ss} \end{pmatrix}
\]

and make the plausible assumption that \( C^{-1}_{sp} = C^{-1}_{ps} = 0 \) (i.e. \( P \) and \( S \) travel-time data are uncorrelated).

Using a parameterization in terms of bulk sound and shear velocity, the \( P \) wave velocity \( \alpha = \alpha(v, \beta) \); and we need to consider a partitioned model parameter space. Impose a smoothness constraint

\[ P_s = \nu(m - m_o)^T D(m - m_o) \]

where \( D \) is a first or second order difference operator, and

\[
D = \begin{pmatrix} D_{\phi \phi} & 0 \\ 0 & D_{\beta \beta} \end{pmatrix}
\]

with \( D_{\phi \phi} \) and \( D_{\beta \beta} \) are narrow banded matrices. We can include possible a-priori cross coupling through a term

\[ P_m = (m - m_o)^T C_m^{-1} (m - m_o) \]

with \( C_m \) a model covariance, and

\[
C_m^{-1} = \begin{pmatrix} C^{-1}_{\phi \phi} & C^{-1}_{\phi \beta} \\ C^{-1}_{\beta \phi} & C^{-1}_{\beta \beta} \end{pmatrix}
\]
To extract 3-D structure we then seek to minimize a combination of the data fit and regularization terms

\[ P = P_d + P_s + P_m = (d - d_o)^T C_d^{-1} (d - d_o) + \]
\[ \nu (m - m_o)^T D (m - m_o) + (m - m_o)^T C_m^{-1} (m - m_o) \]

where the data are regarded as functionals of the model parameters and \( \nu \) is a variable weighting for smoothing.

We expand \( P \) to quadratic terms in the model perturbation \( \Delta m = m - m_o \)

\[ P = P(m_o) + \theta^T \Delta m + \frac{1}{2} \Delta m^T H \Delta m + ... \]

The gradient with respect to the model parameters

\[ \theta = G^T C_d^{-1} \Delta d \] where \( G = \frac{\partial d}{\partial m} \bigg|_{m_o} \)

and the Hessian matrix

\[ H = G^T C_d^{-1} G + \frac{\partial G}{\partial m} \bigg|_{m_o} \Delta d + \nu D + C_m^{-1} \]

For a minimum with respect to \( \Delta m \) we require

\[ \theta + H \Delta m = 0 \]

where, formally, \( \Delta m = -H^{-1} \theta \) in terms of the generalized inverse \( H^{-1} \).

In terms of the data partitions
\[ \theta = G^T C_d^{-1} \Delta d = G^T \begin{pmatrix} C_{pp}^{-1} & 0 \\ 0 & C_{ss}^{-1} \end{pmatrix} \begin{pmatrix} \Delta t_p \\ \Delta t_s \end{pmatrix} \]

and the Fréchet derivative matrix

\[ G = \begin{pmatrix} \partial(\Delta t_p) / \partial \phi & \partial(\Delta t_p) / \partial \beta \\ 0 & \partial(\Delta t_s) / \partial \beta \end{pmatrix} = \begin{pmatrix} G_{p\phi} & G_{p\beta} \\ 0 & G_{s\beta} \end{pmatrix} \]

so that the gradient vector

\[ \theta = \begin{pmatrix} G^T_{p\phi} \\ 0 \\ G^T_{s\beta} \end{pmatrix} \begin{pmatrix} C_{pp}^{-1} \Delta t_p \\ C_{ss}^{-1} \Delta t_s \end{pmatrix} = \begin{pmatrix} G^T_{p\phi} C_{pp}^{-1} \Delta t_p \\ G^T_{s\beta} C_{ss}^{-1} \Delta t_s \end{pmatrix} + \begin{pmatrix} vD_{p\phi} + C_{p\phi}^{-1} \\ G^T_{p\beta} C_{pp}^{-1} G_{p\beta} + vD_{p\beta} + C_{p\beta}^{-1} \end{pmatrix} \]

and the action of the Hessian on the model perturbation (neglecting the second order derivatives of the data) is

\[ \begin{pmatrix} G^T_{p\phi} C_{pp}^{-1} G_{p\phi} + vD_{p\phi} + C_{p\phi}^{-1} \\ G^T_{p\beta} C_{pp}^{-1} G_{p\beta} + vD_{p\beta} + C_{p\beta}^{-1} \end{pmatrix} \begin{pmatrix} \Delta \phi \\ \Delta \beta \end{pmatrix} \]

The "least-squares" solution therefore brings in cross-coupling between the two data sets irrespective of the model covariance terms.

**Note 1:**

Because of the nature of the equations to be solved, the application of smoothing upon inversion is implemented by using the gradient damping scheme presented in chapter 6 such that

\[ vD = \gamma_h^2 G^T \delta G + \gamma_h^2 G^T \theta G + \gamma_h^2 G^T \phi G \]

[cf. (6.5) in chapter 6].
INVERSION OF A COUPLED SYSTEM

We can write the equations to be solved in the form

\[
\begin{pmatrix}
H_{\phi\phi} & H_{\phi\beta} \\
H_{\beta\phi} & H_{\beta\beta}
\end{pmatrix}
\begin{pmatrix}
\phi \\
\beta
\end{pmatrix}
=
\begin{pmatrix}
\theta_{\phi} \\
\theta_{\beta}
\end{pmatrix}
\]

where \( \theta_{\phi} = G_{\rho\rho}^T C_{\rho\rho}^{-1} \Delta t_{\rho} \) and \( \theta_{\beta} = G_{\rho\rho}^T C_{\rho\rho}^{-1} \Delta t_{\rho} + G_{\tau\tau}^T C_{\tau\tau}^{-1} \Delta t_{\tau} \) are the components of the gradient vector \( \theta \) projected onto the bulk-sound and shear speeds. The vectors \( \phi \) and \( \beta \) represent the vectors of the cellular variations of the bulk-sound and shear wavespeeds, respectively.

The character of the terms suggests that it may be possible to make an iterative development. We can recast the coupled equations in the form

\[
\begin{pmatrix}
H_{\phi\phi} & H_{\phi\beta} \\
H_{\beta\phi} & H_{\beta\beta}
\end{pmatrix}
\begin{pmatrix}
\phi \\
\beta
\end{pmatrix}
=
\begin{pmatrix}
H_{\phi\phi} & 0 \\
0 & H_{\beta\beta}
\end{pmatrix}
\begin{pmatrix}
I \\
H_{\beta\beta}^{-1} H_{\phi\phi}
\end{pmatrix}
\begin{pmatrix}
\phi \\
\beta
\end{pmatrix}
\]

The required inverse can thus be written

\[
\begin{pmatrix}
H_{\phi\phi} & H_{\phi\beta} \\
H_{\beta\phi} & H_{\beta\beta}
\end{pmatrix}^{-1}
=
\begin{pmatrix}
I & H_{\phi\beta}^{-1} H_{\phi\phi} \\
H_{\beta\beta}^{-1} H_{\phi\phi} & I
\end{pmatrix}
\begin{pmatrix}
H_{\phi\phi} & 0 \\
0 & H_{\beta\beta}^{-1}
\end{pmatrix}
\]

\[
=
\left[
\begin{pmatrix}
1 & 0 \\
0 & 1
\end{pmatrix} + \begin{pmatrix}
0 & -H_{\phi\beta}^{-1} H_{\phi\phi} \\
-H_{\beta\beta}^{-1} H_{\phi\phi} & 0
\end{pmatrix}
\right] +
\begin{pmatrix}
H_{\phi\phi}^{-1} H_{\phi\beta}^{-1} H_{\phi\phi} & 0 \\
0 & H_{\beta\beta}^{-1} H_{\phi\phi} H_{\phi\beta}^{-1} H_{\phi\phi}
\end{pmatrix}
\begin{pmatrix}
I & H_{\phi\beta}^{-1} H_{\phi\phi} \\
H_{\phi\phi}^{-1} H_{\phi\beta}^{-1} H_{\phi\phi} & I
\end{pmatrix}^{-1}
\begin{pmatrix}
H_{\phi\phi} & 0 \\
0 & H_{\beta\beta}^{-1}
\end{pmatrix}
\]

If we truncate the expansion of the inverse

\[
\begin{pmatrix}
H_{\phi\phi} & H_{\phi\beta} \\
H_{\beta\phi} & H_{\beta\beta}
\end{pmatrix}^{-1}
=
\left[
\begin{pmatrix}
1 & 0 \\
0 & 1
\end{pmatrix} + \begin{pmatrix}
0 & -H_{\phi\beta}^{-1} H_{\phi\phi} \\
-H_{\beta\beta}^{-1} H_{\phi\phi} & 0
\end{pmatrix}
\right] +
\begin{pmatrix}
H_{\phi\phi}^{-1} H_{\phi\beta}^{-1} H_{\phi\phi} & 0 \\
0 & H_{\beta\beta}^{-1} H_{\phi\phi} H_{\phi\beta}^{-1} H_{\phi\phi}
\end{pmatrix} + ...
\begin{pmatrix}
H_{\phi\phi} & 0 \\
0 & H_{\beta\beta}^{-1}
\end{pmatrix}
\]
$$\begin{pmatrix} H_{\phi\phi}^{-1} & 0 \\ 0 & H_{\beta\beta}^{-1} \end{pmatrix} + \begin{pmatrix} 0 & -H_{\phi\phi}^{-1}H_{\phi\beta}H_{\beta\phi}^{-1} \\ -H_{\phi\phi}^{-1}H_{\phi\beta}H_{\beta\phi}^{-1} & 0 \end{pmatrix} + \begin{pmatrix} H_{\phi\phi}^{-1}H_{\phi\beta}H_{\beta\phi}^{-1}H_{\phi\phi}^{-1} & 0 \\ 0 & H_{\beta\beta}^{-1}H_{\beta\phi}H_{\phi\phi}^{-1}H_{\beta\phi}^{-1} \end{pmatrix} + \ldots$$

So we can construct a series of approximations for $\phi$, $\beta$ in successive powers of $H_{\beta\phi}$.

$$H_{\beta\beta}$$

$$\phi_0 = H_{\phi\phi}^{-1} \theta_0, \quad \beta_0 = H_{\beta\beta}^{-1} \theta_0$$

$$\phi_1 = H_{\phi\phi}^{-1}(\theta_0 - H_{\phi\phi}H_{\beta\phi}^{-1} \theta_0) = H_{\phi\phi}^{-1}(\theta_0 - H_{\phi\phi} \beta_0)$$

$$\beta_1 = H_{\beta\beta}^{-1}(\theta_0 - H_{\beta\beta}H_{\phi\phi}^{-1} \theta_0) = H_{\beta\beta}^{-1}(\theta_0 - H_{\beta\phi} \phi_0)$$

$$\phi_2 = H_{\phi\phi}^{-1}(\theta_0 - H_{\phi\phi} \beta_0 + H_{\phi\phi} H_{\beta\phi}^{-1} H_{\phi\phi} H_{\phi\phi}^{-1} \theta_0)$$

$$\beta_2 = H_{\beta\beta}^{-1}(\theta_0 - H_{\beta\phi} \phi_0 + H_{\beta\phi} H_{\phi\phi}^{-1} H_{\beta\phi} H_{\phi\phi}^{-1} \theta_0)$$

Now

$$\phi_0 - \phi_1 = H_{\phi\phi}^{-1} \theta_0 - H_{\phi\phi}^{-1} \theta_0 + H_{\phi\phi}^{-1} H_{\phi\phi} H_{\beta\phi}^{-1} H_{\phi\phi} \theta_0 = H_{\phi\phi}^{-1} H_{\phi\phi} H_{\beta\phi}^{-1} \theta_0$$

$$\beta_0 - \beta_1 = H_{\beta\phi}^{-1} \theta_0 - H_{\beta\phi}^{-1} \theta_0 + H_{\beta\phi}^{-1} H_{\phi\phi} H_{\beta\phi}^{-1} H_{\phi\phi} \theta_0 = H_{\beta\phi}^{-1} H_{\phi\phi} H_{\beta\phi}^{-1} \theta_0$$

so

$$\phi_2 = H_{\phi\phi}^{-1}(\theta_0 - H_{\phi\phi} \beta_0 + H_{\phi\phi} (\beta_0 - \beta_1)) = H_{\phi\phi}^{-1}(\theta_0 - H_{\phi\phi} \beta_0)$$

$$\beta_2 = H_{\beta\phi}^{-1}(\theta_0 - H_{\beta\phi} \phi_0 + H_{\beta\phi} (\phi_0 - \phi_1)) = H_{\beta\phi}^{-1}(\theta_0 - H_{\beta\phi} \phi_0)$$

and we have an iterative development

$$\phi_0 = H_{\phi\phi}^{-1} \theta_0$$
Note 1. \[ \beta_0 = H_{\beta \beta}^{-1} \theta_\beta \]

To compute estimates of \( \beta \) and \( \phi \) data, we have employed the first but derivative matrices such as

\[ \phi_1 = H_{\phi \phi}^{-1} (\theta_\phi - H_{\phi \beta} \beta_0) \]

\[ \beta_1 = H_{\beta \beta}^{-1} (\theta_\beta - H_{\phi \beta} \phi_0) \]

\[ \phi_2 = H_{\phi \phi}^{-1} (\theta_\phi - H_{\phi \beta} \beta_1) \]

\[ \beta_2 = H_{\beta \beta}^{-1} (\theta_\beta - H_{\phi \beta} \phi_1) \]

and in general

\[ \phi_r = H_{\phi \phi}^{-1} (\theta_\phi - H_{\phi \beta} \beta_{r-1}) \]

\[ \beta_r = H_{\beta \beta}^{-1} (\theta_\beta - H_{\phi \beta} \phi_{r-1}) \]

We can regard \( H_{\phi \phi}^{-1} \) and \( H_{\beta \beta}^{-1} \) as representing the operation of solving the block diagonal equations and the coupling from the off-diagonal blocks is introduced by successive modification of the right-hand sides of the equations. In effect each successive iteration aims to work with a vector on the right-hand side which represents just the contribution from a single wave type by correcting for the influence of the other wave type.

In the practical implementation of *) and **) we do not actually compute the inverse but regard *) and **) as representing the solutions of the equation

\[ H_{\phi \phi} \phi_r = \theta_\phi - H_{\phi \beta} \beta_{r-1} \]

\[ H_{\beta \beta} \beta_r = \theta_\beta - H_{\phi \beta} \phi_{r-1} \]

where the right-hand sides of the equations are updated as the iteration proceeds.
Note 2:

To compute variance of $P$ and $S$ data, we have employed the Fréchet derivative matrix such as

$$
\begin{pmatrix}
G_{P\Phi} & G_{P\beta} \\
0 & G_{S\beta}
\end{pmatrix}
\begin{pmatrix}
\phi^p \\
\beta
\end{pmatrix} =
\begin{pmatrix}
\Delta t_p \\
\Delta t_s
\end{pmatrix}
$$

Then the data variance is

For $P$: $\text{Var}(P) = \sum_{i=1}^{n_P} \Delta t_{p(i)}^T C_{pp}^{-1} \Delta t_{p(i)}$ and

For $S$: $\text{Var}(S) = \sum_{i=1}^{n_S} \Delta t_{s(i)}^T C_{ss}^{-1} \Delta t_{s(i)}$

where $n_P$ and $n_S$ are the number of $P$ and $S$ data, respectively.

The derivation above represents a formal proof of the Jacobi iteration for a block partitioned matrix (see Conte & De Boor 1980 for further discussion). Consider

$$
\begin{pmatrix}
H_{\phi\phi} & H_{\phi\beta} \\
H_{\beta\phi} & H_{\beta\beta}
\end{pmatrix}
\begin{pmatrix}
\phi \\
\beta
\end{pmatrix} =
\begin{pmatrix}
\theta_{\phi} \\
\theta_{\beta}
\end{pmatrix}
$$

Initially setting the off diagonal blocks to zero we can build a partitioned form of the Jacobi iteration in the following way:

First iteration,

$$
\begin{pmatrix}
H_{\phi\phi} & 0 \\
0 & H_{\beta\beta}
\end{pmatrix}
\begin{pmatrix}
\phi_{-o} \\
\beta_{-o}
\end{pmatrix} =
\begin{pmatrix}
\theta_{\phi} \\
\theta_{\beta}
\end{pmatrix}
$$

so that

$$
\phi_{-o} = H^{-1}_{\phi\phi} \theta_{\phi} \quad \text{and} \quad \beta_{-o} = H^{-1}_{\beta\beta} \theta_{\beta}
$$
Now find the residual from the original equation

\[
\begin{pmatrix}
  r_\phi \\
  r_\beta
\end{pmatrix} = \begin{pmatrix}
  \theta_\phi \\
  \theta_\beta
\end{pmatrix} - \begin{pmatrix}
  H_{\phi\phi} & H_{\phi\beta} \\
  H_{\beta\phi} & H_{\beta\beta}
\end{pmatrix} \begin{pmatrix}
  \phi_o \\
  \beta_o
\end{pmatrix}
\]

which comes just from the off-diagonal blocks

\[
r_\phi = -H_{\phi\beta} \beta_o \\
and \\
r_\beta = -H_{\beta\phi} \phi_o
\]

We look for a perturbation from the initial estimates to account for the off diagonal blocks

\[
\bar{\phi} = \phi_o + \delta \phi \\
and \\
\bar{\beta} = \beta_o + \delta \beta
\]

for which \( \delta \phi \) and \( \delta \beta \) are to be determined from

\[
\begin{pmatrix}
  H_{\phi\phi} & H_{\phi\beta} \\
  H_{\beta\phi} & H_{\beta\beta}
\end{pmatrix} \begin{pmatrix}
  \delta \phi \\
  \delta \beta
\end{pmatrix} = \begin{pmatrix}
  r_\phi \\
  r_\beta
\end{pmatrix}
\]

Second iteration, (set off diagonal blocks to zero again)

\[
\begin{pmatrix}
  H_{\phi\phi} & 0 \\
  0 & H_{\beta\beta}
\end{pmatrix} \begin{pmatrix}
  \delta \phi \\
  \delta \beta
\end{pmatrix} = \begin{pmatrix}
  r_\phi \\
  r_\beta
\end{pmatrix}
\]

then

\[
\delta \phi = H_{\phi\phi}^{-1} r_\phi \\
\delta \beta = H_{\beta\beta}^{-1} r_\beta
\]

Adding \( \begin{pmatrix}
  \phi_o \\
  \beta_o
\end{pmatrix} \) to both sides gives,


Continuing the process of successive corrections we obtain recursive expressions,

\[ \phi_1 = H_{\phi\phi}^{-1}(\theta_0 - H_{\phi\phi}\beta_0) \]

\[ \beta_1 = H_{\beta\beta}^{-1}(\theta_0 - H_{\beta\phi}\phi_0) \]

which are exactly the same as *) and **)
Appendix 8B

Results of the LSQR inversion of $P$ data (the $P_{LSQR}$ model)
Figure 8B.1(a-e). $P$ velocity anomaly maps from the $LSQR$ inversion (the $P\_LSQR$ model). The letters (a-e) depict different layers in the upper mantle, transition zone and lower mantle. They plotted with contours scales -1.0% to +1.0% relative to the $ak135$ model for structure in the full mantle.
Slices at depths of 1300, 2100 & 2700 km

Figure 8B.1(c-e). (continued)
Figure 8B.2. (a) Plot of maximum and minimum amplitude of the $P$ wavespeed perturbations relative to the $ak135$ model for the $P_{LSQR}$ model. (b) Plot of the r.m.s. velocity perturbations for the same model. The values are plotted at the centre of each of 18 layers in the model parameterization.
Appendix 8C

Results of the joint inversion by excluding the $SKS$ information
Figure 8C.1[1-5]. Bulk-sound velocity anomaly maps from the direct extraction and joint inversion. (a) The Q\_DIRECT model. (b) The Q\_JOINT model. The numbers (1-5) depict different layers in the upper mantle, transition zone and lower mantle.
Slice at 500 km depth

Figure 8C.1[2]. (continued)
Figure 8C.1[3]. (continued)
Slice at 2100 km depth

Figure 8C.1[4]. (continued)
Slice at 2700 km depth

Figure 8C.1[5]. (continued)
Figure 8C.2. (a-b) Plots of maximum and minimum amplitude of the bulk-sound speed perturbations for the $Q_{DIRECT}$ and $Q_{JOINT}$ models. (c-d) Plots of the r.m.s. velocity perturbations for each of the two models. The values are plotted at the centre of each of 18 layers in the model parameterization.
Figure 8C.3[1-5]. Shear velocity anomaly maps from the $LSQR$ and joint inversions. (a) The $S_{LSQR}$ model. (b) The $S_{JOINT}$ model. The numbers (1-5) depict different layers in the upper mantle, transition zone and lower mantle.
Figure 8C.3(2]. (continued)
Slice at 1300 km depth

Figure 8C.3[3]. (continued)
Slice at 2100 km depth

Figure 8C.3[4]. (continued)
Slice at 2700 km depth

Figure 8C.3[5]. (continued)
Figure 8C.4.  (a-b) Plots of maximum and minimum amplitude of the shear wavespeed deviations from the ak135 model for the S_LSQR and S_JOINT models.  (c-d) Plots of the r.m.s. velocity deviations for each of the two models. The values are plotted at the centre of each of 18 layers in the model parameterization.
9 Thesis summary and future work

9.1 SUMMARY

9.1.1 Global synthesis of regional studies

We have established a high-resolution regional tomographic imaging technique by embedding the region of interest within a lower-resolution global inversion with the objective of avoiding the mapping of structure outside the region into apparent structure. This technique has successfully applied to probing the subduction zone structure worldwide. We have demonstrated that the results provide relatively detailed images of slab structures in the upper mantle, the transition zone and the top of the lower mantle.

From these studies we are able to get some impression of the variety of styles of processes associated with subduction. We see for example: (a) slab penetration just into the upper mantle, e.g. beneath the Makran subduction zone associated with relatively recent subduction; (b) a sub-horizontal subducted slab in the upper mantle beneath north America; (c) a double subduction zone with slabs subducting in opposite directions beneath the Molucca Collision Zone; (d) slab deflection in the transition zone, e.g. beneath Banda, Izu Bonin and central Aleutian; (e) a slab detachment beneath Sumatra; (f) slab penetration into the lower mantle with a kink (change of slope), e.g. beneath Java and northern Tonga, and (g) without a kink, e.g. beneath Kermadec and Mariana (see chapters 3 and 4).

The set of regional inversions provides a high-resolution image of the most of the regions of the Earth associated with subduction zones and by combining these images we...
can provide a uniform representation of the structure of much of the upper mantle. In Fig. 9.1 we therefore display the set of results for the inversions for the ten subduction zone regions which have been displayed separately in chapters 3 and 4 in a unified global display with a 1° x 1° resolution. We can extend the regional inversion approach into other regions where there is a relatively high density of either seismic events or seismic stations, by once again using a high-resolution 1° x 1° mesh embedded in a low resolution global model. We have used this approach for Australia (where we have supplemented permanent station data with data from SKIPPY arrays), and for the area around China extending from the subduction beneath Burma to the Philippines. We have also looked at the surrounding of two major hot spots, Iceland in the north Atlantic and Afar in east Africa. We will not discuss the four additional inversions in any detail in this thesis, however, their inclusion means that the images in Fig. 9.1 cover almost all zones for which high resolution is justified. The resulting images can be compared with the recent P1200 model of Zhao (1996) in which 1° x 1° model for the whole upper mantle down to 1200 km is presented from a multicell inversion procedure. Note that Zhao (1996) makes no allowance for lower-mantle structure, whereas each regional inversion included an inversion for the structure in the rest of the Earth.

In the mantle images in Fig 9.1 we have used the regional results directly with no attempt to smooth structures across boundaries between regions, but have allowed the subduction zone results to overlap those for the four additional inversions. We see that in general there is a relatively good match of structures resulting from each of the fourteen regional inversions. The discrepancies which do occur lie at the edges of the study regions and are most likely due to limited resolution.

The images for the structure in the upper mantle (110-160 km) shown in Fig. 9.1 (a) depict narrow high-velocity anomalies beneath most of the well-known subduction zones, as would be expected from our discussion in chapters 3 and 4. In particular we see the high velocities in Tonga, Indonesia, northwest Pacific, the Aleutian Islands and south America. In addition we see clearly the fast P wavespeeds beneath parts of stable continental shields. Low velocity anomalies characterizing the back-arc regions around
the Pacific are also well imaged. The extension of the regional studies to cover east Africa and Iceland gives a clear indication of the presence of low velocities beneath the hot spots. The African anomaly appears to extend to the transition zone but this may be due, in part, to vertical smearing owing to the event and station distribution. Further, the infill of the China/Burma zone has improved the definition of structures in the Philippines and the slab structures associated with the subduction of the Philippine plate.

In the transition zone (Fig. 9.1b), the lateral displacement associated with subduction leads to fast wavespeeds lying beneath the back-arc regions at the surface, especially in the western Pacific. A narrow region of fast wavespeed propagation is observed beneath central America, the Caribbean region and south America. A large region of slow wavespeeds is imaged beneath India and east Africa in contrast to the more subdued velocity structure beneath the Australian shield (cf. Fig. 9.1a).

In the upper part of the lower mantle (Fig. 9.1c) we see the traces of pronounced fast slabs in the lower mantle beneath southern Eurasia, Tonga and parts of the Americas which strongly support the inferences we have drawn from our global imaging at the lower resolution of $2^\circ \times 2^\circ$. The prominent nearly linear feature in the Americas can be traced across three separate regional inversions through south America, central America and into the north Atlantic zone. Likewise there is a reasonable link between the high velocity feature beneath Indonesia and the higher velocities in the inversion for the Indian region. It is clear that the regional high-resolution inversions continue to be useful to considerable depth.

We have not included any results from the global inversions in Fig. 9.1 so that the regional inversions can be clearly displayed. However, there is a very close correspondence between the patterns of anomalies revealed by the regional inversions and the global $P$ wave results. Higher resolution has been achieved in the regional results, but it is likely that a global inversion with flexible parameterization could reveal further intriguing detail of the Earth's internal structure.
Figure 9.1. Results of 14 different regional tomographic inversions plotted using a Mollweide projection. (a-c) Solutions representing structures in the upper mantle, the transition zone and the lower mantle with contour scales -2.0% to +2.0%, -1.5% to +1.5% and 1.0% to +1.0% relative to ak135, respectively. Blue (red) colors represent fast (slow) wave propagation. Note that there is no smoothing applied across the boundaries between regions.
Layer: 10
490-570 km

Figure 9.1b (continued)
Layer: 17
1130-1250 km

Figure 9.1c (continued)
9.1.2 Global models

We also have been able to exploit the high quality reprocessed data set for arrival times to produce relatively high-resolution global models, in which we have been able to extend detailed imaging for structures in the entire lower mantle. We have applied this approach to invert not only global $P$ travel times but global $S$ travel-time data as well. Although the models for $P$ and $S$ derived using totally different data sets, these two models show a reasonably close comparison through much of the mantle which reinforces the reliability of the images.

The resulting $P$ images provide a snapshot of the seismic velocity distribution in the mantle, which can be interpreted in terms of convection in the mantle by association of the velocity anomalies with the thermal effects. The patterns of fast (cold) subduction and slow (hot) upwelling provide clear evidence for deep mantle circulation as has been discussed in detail by Van der Hilst, Widiyantoro & Engdahl (1997) and Grand, Van der Hilst & Widiyantoro (1997). Although less detail can be extracted from the $S$ data because of the more limited path coverage from the reported data, the $S$ images are generally consistent with the $P$ images. We have been able to improve the quality of the inversion for $S$ in the lowermost mantle by selective inclusion of SKSac core-phase data in the global $S$ data inversion. In particular, the velocity deviations from the reference model increase significantly as the core-mantle boundary is approached and reach levels close to those in the upper mantle (Widiyantoro, Kennett & Van der Hilst 1997), which is consistent with the results of global inversions from long-period waveform studies e.g. $SH13\_WM13$ (Su et al. 1994).

The results from the individual inversions of the global $P$ and $S$ data for the depth interval of 1200-1400 km are shown in Fig. 9.2. The two different inversions were designed to make the optimum possible use of the available global data sets for $P$ and $S$ waves separately and consequently do not have readily comparable $P$ and $S$ ray coverage in the mantle. Despite the differences in resolution imposed by the available ray paths, the images of velocity heterogeneity provide an important observation through the clarity
of narrow compressional and shear velocity anomalies in the mid mantle. These features are resolved by both \( P \) and \( S \) data and geographically correspond to some zones of former subduction as well as parts of the present-day subduction zones in the upper mantle (see chapters 6 and 7) which lead us to draw a conclusion that there must be significant slab penetration into the lower mantle. This conclusion requires there to be significant mass transfer between the upper mantle and the lower mantle associated with the slab penetration. Although deflection of slabs is also observed, this phenomenon seems to occur just locally, and is significant on a time scale that is shorter than that for mantle wide overturn (Christensen 1996; Van der Hilst et al. 1997).
Depth: 1200 - 1400 km

Figure 9.2. Results of global $P$ and $S$ individual inversions using the $LSQR$ method. (a) $P$ wavespeed relative to $ak135$ from the $P$ model presented in chapter 6. (b) $S$ wavespeed relative to $ak135$ from the $S_{SKS105}$ model presented in chapter 7.
One of the major new results of the research undertaken in this thesis work is that we have been able to produce global bulk-sound and shear velocity heterogeneity models through joint inversion of global $P$ and $S$ travel-time data. The joint inversion procedure is relatively efficient in term of the computational load since we solve sequentially for bulk-sound and shear wavespeeds and introduce cross-correction terms between the two partial inversions. With this new class of information we are able to look at deep structures in the mantle of our planet in a different way, and begin to test some of the standard assumptions with regard to the interpretation of tomographic images. In particular, an advantage arises from the differing sensitivity of the bulk and shear moduli to temperature. Thermal effects seem to be more strongly marked for shear wavespeed than bulk-sound speed, but when there is a consistent pattern between bulk-sound and shear speeds we can expect that the origin is thermal. However, where pronounced anomalies in bulk sound do not coincide with those for shear waves this is likely to indicate chemical heterogeneity.

We display, in Fig. 9.3, the images of bulk-sound and shear wavespeeds for the mid mantle (1200-1400 km; the same depth range as in Fig. 9.2). The bulk-sound and shear wavespeed models were produced by the joint inversion using a common $P$ and $S$ ray coverage in order to achieve approximately the same resolution in the resulting two models (see chapter 8). At this depth the heterogeneity in bulk sound is somewhat smaller than for shear, but most of the major features in the shear wave image can be discerned in the bulk-sound image which suggests that the pronounced anomalies in shear wavespeed are indicative of thermal heterogeneity. However there are some zones of discrepancy that might be associated with chemical heterogeneity. The shear model displayed in Fig. 9.3 (b) is fortunately very similar to that in Fig. 9.2 (b), even though these two shear models were derived using different inversion techniques and different ray coverage. Some of the differences are likely to arise from the fact that we have extracted shear wavespeed information from $P$ wave times in the joint inversion (this is introduced by cross-correction terms in the partitioned inversion scheme), as well as from differences in ray sampling.
Figure 9.3. Results of the joint inversion of $P$ and $S$ data using the LINBCG method presented in chapter 8. (a) Bulk-sound speed (the $Q_{SKS\_JOINT}$ model). (b) Shear wavespeed (the $S_{SKS\_JOINT}$ model; cf. Fig. 9.2b).
9.2 FUTURE WORK

We have successfully imaged global slab structure using the improved quality of travel-time data provided by the new data base of high-resolution earthquake locations and arrival times of Engdahl et al. (1997). The present style of parameterization, however, may not be the most efficient in terms of the representation of the different styles of structure which appear in the tomographic images. In order to reduce the dimension of the matrices involved in the inversion and the CPU time it is desirable to use a flexible grid rather than just shrinking the cell-size.

A suitable parameterization for future work, would appear to be the natural neighbours technique introduced by Sambridge et al. (1995). This method has a very useful property i.e. the ability to represent large variations in the scale-lengths of the interpolated function, and so is potentially suitable as a basis for global tomographic inversion. The flexibility of using various cell sizes and configurations may enable us to parameterize the region of interest, e.g. subduction zones, in more detail without necessarily increasing the overall dimension of the matrices required. The work of Spakman & Bijwaard (personal communication, 1997) suggests that the use of a smaller cell size than the $1^\circ \times 1^\circ$ used in the current regional inversions can lead to even clearer imaging of narrow slabs in the upper mantle.

The choice of the ak135 reference velocity model (Kennett et al. 1995) for this study is sensible, in particular for the global imaging because this model represents a close match to the data set used (personal communication with B.L.N. Kennett, 1997) and so improves the linearity of the inversion. However, any such global model will not necessarily be the most suitable reference model for particular areas of the globe. For instance, the regional inversion for central America presented in chapter 4 indicates that ak135 is perhaps too fast for the upper mantle structure beneath the region. Therefore, in any future more detailed study it will be desirable to develop 1-D reference models for such regional inversions as have been presented in chapters 3 and 4.

From the results of the comparisons between our global P and S models we found
good general agreement but observed some differences as well, notably in the lowermost mantle. In chapter 8, we have made an attempt to explain these differences to some extent in terms of the variation in bulk-sound speed as a preliminary interpretation. Discrepancies between the different classes of models generally occur in regions where one model and/or the other do/does not have good resolution, but may also arise from the nature of $P$ and $S$ themselves which need further study and further improvement in the models.

In this work, we only have used data from body-waves (and mostly from permanent stations) to try to constrain three-dimensional structure and so there are still large gaps in the data coverage. This encourages new projects such as SKIPPY (Van der Hilst et al. 1994); and the 'Japan Indonesia Broad-band Seismic Network' (JISNET) project, the 'Ocean Hemisphere Network' (OHP) project and the SPANET project mentioned in chapter 7 which are conducted by Japanese groups to introduce new stations into under-sampled areas. A partial solution would be to adopt a 3-D reference model such as the $SH13\_WM13$ model by Su et al. (1994) instead of a 1-D model (such as $akl35$). In this way, the regularization procedures in the inversion will bias toward the starting 3-D model but in regions where we have good ray coverage, e.g. beneath subduction zones, the new global data set will potentially improve the 3-D starting model. Note that such a 3-D model, e.g. by Su et al. (1994), provides constraints on long-wavelength anomalies, since it was produced based on spherical harmonics. An alternative approach would be to seek to include times from multiply reflected phases as has been undertaken by Grand (1994) for $SH$ waves.

In this thesis, we have produced many regional and global models but have not attempted a detailed interpretation of all the different models. We have made an attempt to interpret the regional inversion results for Indonesia in some detail (Widiyantoro & Van der Hilst 1996; 1997), but a comparable level of interpretation lies beyond the time frame of this Ph.D. and will be conducted elsewhere.

Interpretation of aspects of the global models have already been published by Van der Hilst, Widiyantoro & Engdahl (1997); and Grand, Van der Hilst & Widiyantoro
(1997) but we are aware that further analysis is desirable. Furthermore, we only have been able to make an early stage of interpretation of our new results i.e. the global bulk-sound and shear velocity models. Detailed interpretations of this new view of the Earth's interior will obviously need further investigation which most likely will involve the integration of information from mineral physics.
Appendix 9

CD-ROM

In order to provide readers with access to our models, we present a complete set of layer anomaly maps of most of the models presented in this thesis in the attached CD. All of the figures are plotted using the same color scheme, in which red colors indicate slower than average and blue colors indicate faster than average seismic velocity.

The complete set of the ten regional inversion results for subduction zones worldwide can be found in the sub-directories under `ps/reg`.

The results of the various global inversions are located in `ps/glb`. For the global inversion results, we only display layer anomaly maps that have not been otherwise presented in the chapters of this thesis due to the limited space on the disc. The global P model presented in chapter 6 is contained in `ps/glb/p`. We present the `S_SKS105` model from chapter 7 in `ps/glb/s`. The joint inversion results from chapter 8 i.e. `Q_SKS_JOINT` and `S_SKS_JOINT` are in `ps/glb/joint`.

A GIF version of the above postscript files for the regional and global models is available in the sub-directories under `gif/reg` and `gif/glb`, respectively.
BIBLIOGRAPHY


