Analyses of Seismic Wave Conversion in the Crust and Upper Mantle beneath the Baltic Shield

SVERKER OLSSON
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Abstract

Teleseismic data recorded by broad-band seismic stations in the Swedish National Seismic Network (SNSN) have been used in a suite of studies of seismic wave conversion in order to assess the structure of the crust and upper mantle beneath the Baltic Shield. Signals of seismic waves converted between P and S at seismic discontinuities within the Earth carry information on the velocity contrast at the converting interface, on the depth of conversion and on P and S velocities above this depth.

The conversion from P to S at the crust-mantle boundary (the Moho) provides a robust tool to constrain crustal thicknesses. Results of such analysis for the Baltic Shield show considerable variation of Moho depths and significantly improve the Moho depth map. Analysis of waves converted from S to P in the upper mantle reveals a layered lithosphere with alternating high and low velocity bodies. It also detects clear signals of a sharp velocity contrast at the lithosphere-asthenosphere boundary at depths around 200 km.

Delay times of P410s, the conversion from P to S at the upper mantle discontinuity at 410 km depth, were used in a tomographic inversion to simultaneously determine P and S velocities in the upper mantle. The polarisation of P410s was also used to study anisotropy of the upper mantle. Results of these analyses are found to be in close agreement with independently derived results from arrival time tomography and shear-wave splitting analysis of SKS.

The results presented in this thesis demonstrate the ability of converted wave analysis as a tool to detect and image geological boundaries that involve sharp contrasts in seismic properties. The results also show that this analysis can provide means of studying aspects of Earth’s structure that are conventionally studied using other types of seismic data.

Keywords: converted waves, Moho, upper mantle, Baltic Shield, tomography, anisotropy, receiver function

Sverker Olsson, Department of Earth Sciences, Villav. 16, Uppsala University, SE-75236 Uppsala, Sweden

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This thesis is a compilation of the following four papers:


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1. Introduction

From the early days of instrumental seismology, important conclusions regarding the Earth’s interior have been drawn from analysis of recorded seismograms. From the observed absence of direct body wave arrivals at epicentral distances beyond ~103°, Oldham already in 1906 suggested the presence of the Earth's core. In 1909 Mohorovicic deduced a rapid increase of seismic velocities at an estimated depth of ~50 km from travel time observations (Mohorovicic, 1992). This velocity discontinuity has later been named after its discoverer as theMohorovicic discontinuity or simply the Moho. Since its discovery it has been observed globally with a variation in depth from ~5 km beneath oceans to 30-60 km beneath continents and is recognized as a major lithological boundary separating the Earth's crust and mantle. In the following 100 years, the increased number of seismic stations in the world together with improved instruments has led to a more and more detailed image of the Earth's structure. This evolution has accelerated in the last few decades through the use of digital data and more powerful computers.

Even long before seismic waves had ever been instrumentally recorded, principles of seismic wave propagation were well studied in theory. In the early 19th centuries, major advances had been made by mathematicians and theoretical physicists such as Young, Navier, Cauchy, Green, Stokes and others. The French mathematician Poisson suggested around 1830 that two modes of wave propagation are excited by a source in an elastic medium. These modes differ in the particle motion of the wave. The compressional wave (P) has its particle motion in the direction of wave propagation, while the particle motion in the shear wave (S) is perpendicular to it. From a simple model Poisson also deduced that the P wave travelled with a speed ~√3 times the speed of the S wave. It was not until around 70 years later that these wave phases could be reliably identified in recorded seismograms.

Conversion between these two modes of wave propagation will occur at seismic velocity discontinuities at various depths in the Earth's interior. The amount of seismic energy to be converted is dependent on the magnitude and sharpness of seismic velocity contrasts. Hence the analysis of converted waves allows us to detect and map major structural boundaries within the Earth. Furthermore, wave speeds of the converted wave and the direct, non-converted wave are different. This means that the study of their relative arrival times also has the potential to resolve velocities of overlying layers.
The present thesis is a compilation of four papers where converted waves have been analysed to study the crust and upper mantle beneath the Baltic Shield. The data used in the studies are teleseismic data recorded by broadband stations in the Swedish National Seismic Network (SNSN). In the first paper, P to S conversion at the crust-mantle boundary is studied to map variation in Moho depth in the area. The analysis of waves converted from S to P in Paper II reveals significant layering of the lithospheric mantle beneath the Baltic Shield. A tomographic approach is used in Paper III to infer upper mantle velocity variation from delay times of P410s, the conversion from P to S at the upper mantle discontinuity at 410 km depth. Delay times and polarizations of P410s are also used in Paper IV to map velocities and anisotropy in the upper mantle.
2. Elements of converted wave analysis

This thesis is based on analysis of the conversion between P and S modes of wave propagation that will occur at reflecting and refracting interfaces in the subsurface. For a full treatment of the theoretical aspects of seismic wave conversion and its mathematical formulation, the reader is referred to seismology textbooks such as Dahlen and Tromp (1998) or Aki and Richards (2002). In this introduction I will restrict the presentation to the elements of converted wave analysis essential for the understanding of the analyses in this thesis.

Receiver functions

Converted waves are easily detected and discriminated from other wave phases through deconvolution. The converted wave can be viewed as the convolution of the associated direct wave with some operator describing the phase conversion (Ammon, 1991). This conversion operator is commonly referred to as a receiver function. From the components \((P, SV\) and \(SH)\) of the seismogram, we can form the convolution

\[
SV(t) = P(t) * R_{SV}(t)
\]

(1)

to define the radial P receiver function \(R_{SV}(t)\) describing the conversion from P wave motion to SV polarised S wave motion. Similarly

\[
SH(t) = P(t) * R_{SH}(t)
\]

(2)

defines the tangential P receiver function \(R_{SH}(t)\) and

\[
P(t) = SV(t) * R_{SVP}(t)
\]

(3)

defines an S receiver function \(R_{SVP}(t)\) describing the conversion from SV polarised S motion to P wave motion. To study a certain mode of wave conversion, we thus compute the receiver function through deconvolution of the appropriate components.
In the ideal case, the deconvolved receiver function (RF) consists of a number of peaks each representing seismic wave conversion. Amplitudes of RF peaks describe the amplitudes of converted waves relative to that of the associated direct wave. The amount of energy converted is dependent on the contrast in seismic impedance at the converting discontinuity and on the incidence angle of the direct wave. The timescale of a receiver function gives the relative arrival times of the converted phase and the direct wave. P to S converted phases arrive at the recording station later than the direct P, so time shifts in the P receiver function are delay times. S receiver functions have negative time shifts indicating advance times of S to P converted phases.

In the studies in this thesis, two different techniques have been used for deconvolution. For deconvolution in the frequency domain, the “water level” method suggested by Langston (1979) has been used in several applications of the receiver function technique. This method uses a “water level”, $c$, to stabilize the spectral division. For the radial P receiver function the formulation is:

$$ R(\omega) = \left| \frac{SV(\omega)P(\omega)}{\max\{P(\omega)P^*(\omega), c \cdot \max_{\omega} |P(\omega)^*P(\omega)|\}} \right| G(\omega), $$

where $^*$ has been used to denote the complex conjugate and a low pass Gaussian filter, $G(\omega)=\exp(-\omega^2/4a^2)$, defined by its filter width $a$, is used to suppress high frequency noise. Stabilization of the spectral division is hence obtained by setting a threshold for the spectral amplitude of the denominator. In this thesis, this frequency domain deconvolution method has been used for the analyses in Papers I, II and IV. It has the advantage over other deconvolution methods in that its computation is very fast. It has been observed to produce less reliable results with noisy data and suffers from acausal troughs that may decrease amplitudes of significant peaks in the trace. It also requires the value of the water level $c$ to be defined from analysis of data quality.

An alternative method for deconvolution is the time domain iterative deconvolution method proposed by Ligorría and Ammon (1999). In the application of this technique to the radial P receiver function, the largest peak of the deconvolved function is found from cross correlation of P and SV components. The convolution of the deconvolved peak with the P component is then subtracted from the SV component and the next cross correlation is performed. This is repeated and the deconvolved P receiver function is constructed from the time lags and amplitudes of the peaks found. The iterative time domain deconvolution method was used for the analysis in Paper III. Its major drawback is that its computation is far more time consuming than the
frequency domain method. It does however have several advantages. First, it
does not require the \textit{a priori} setting of a water level value. It has also been
considered to produce more reliable results for noisy data (Ligorría and
Ammon, 1999) and it is capable of reproducing a flat long period level,
which helps the ability to assess the significance of low amplitude peaks
from visual inspection of the deconvolved data.

Isolation of the converted phase

In equations 1-3, P and S components are assumed to be completely sepa-
rated. However, since P and S waves travel at different wave speeds their
incidence angles will not be the same. Also, the interaction of P and S waves
with the Earth’s surface is different. Hence observed P and S wave particle
motions will not be orthogonal and rotation to obtain complete separation
between the two is not possible. In some cases approximation of the P and S
polarisations with vertical and horizontal components will be sufficient. This
holds especially if we are studying the fairly high amplitude wave conver-
sions associated with the impedance contrast at the crust-mantle boundary.
In other cases it is advisable to rotate components in the vertical plane in
order to isolate the converted phase as much as possible prior to deconvolu-
tion. For the case of a P to S conversion at upper mantle discontinuities, the
converted S may have an amplitude of only 1-2 percent of that of the direct P
wave. We therefore want to clean the SV component from any P motion by
rotating into a coordinate system that minimizes the energy on the SV com-
ponent from the direct P arrival. Effectively this means that we rotate com-
ponents into a coordinate system given by the apparent angle of incidence of
the P wave. Similarly, to study S to P conversion components are rotated to
the incidence angle of the S wave.

There are several methods for determining the angle of component rota-
tion. A rotation to a coordinate system that minimizes the amplitude power
on the radial component (Yuan et al., 2006) is often used in receiver function
studies. In the study of S to P converted waves in this thesis (Paper II), com-
ponents were rotated to a coordinate system that minimizes cross correlation
between components within a time window including the direct S arrival.
This method was found to give more consistent rotation angles than rotations
based on minimum amplitudes on the P component. Whatever method is
used, it will certainly not provide a very exact solution for the optimum rota-
tion angle. The minimum of the quantity we seek to minimise in the deter-
mination will not be very sharp. This in turn is not a major problem, since
having the rotation angle within 5-10° of the optimum will give a very small
difference in amplitudes of deeper converted phases. With some reasonable
idea of P and S near surface velocities at the recording site, we can use the
horizontal slowness of the direct wave (calculated from the hypocentre loca-
tion and a global velocity model) to calculate its apparent angle of incidence
and use this for rotation of components. For the analysis in Paper III, components of the direct P arrival were used to estimate station surface velocities and these were used to calculate rotation angles for the analysis of P410s delay times. It even turns out that an approximation with a linear relation between horizontal slowness and rotation will give sufficient accuracy for component rotation angles, with errors of only a few degrees. This approximation was used in the analysis of P410s polarizations presented in Paper IV.

Delay times of converted phases

While amplitudes of major peaks in a receiver function are interpreted in terms of velocity contrasts, the time shifts are related to depth of conversion, and to P and S velocities. Interpretation of delay time variation as representing variation in depth to the converting interface by applying a linear time-to-depth conversion is well justified in the analysis of fairly shallow conversions. Figure 1 shows P receiver functions calculated for station ask (58.9°N, 14.8°E). In the section, traces have been sorted according to back azimuth. The sharp peak seen at delay times ~6-7 seconds is interpreted as the phase conversion at the Moho. The observed variation in delay time with back azimuth in this example should certainly be interpreted as geographical variation in depth to the converting interface.

Figure 1. P receiver functions calculated for the SNSN station ask (58.9°N, 14.8°E). Traces have been arranged by back azimuth. Black is used for positive RF amplitudes and light grey for negative. \( t = 0 \) corresponds to the direct P arrival. The coherent signal at delay times 6-7 s is interpreted as the Moho P to S conversion.
For deeper phase conversion, violations to this one-to-one relation between receiver function time shifts and conversion depth will be more severe. For the event distance range used in e.g. P receiver functions (30°-95°), the differences in incidence angle of the incoming P wave (or of the horizontal slowness) will cause significant differences in the delay time of the associated RF peak. In the study of P to S conversion at the 660 km discontinuity this time differences may be up to 9 seconds. Velocity contrasts suggested for upper mantle discontinuities will produce small conversion amplitudes that might require stacking for detection. For coherent stacking traces have to be corrected to account for the effect of moveout. Moveout correction is applied to RF data in order to align phases that have been converted at one common depth. It is thus analogous to moveout correction in reflection seismic processing where phases reflected at a common depth are aligned prior to stacking.

For the studies in this thesis, two different approaches have been used for moveout correction of the data. In the analysis of waves converted from S to P (Paper II), peaks in the computed S receiver functions are interpreted as subsurface converting discontinuity structures. In this analysis moveout correction has therefore been achieved by depth-migrating the data. For each time step in the RF, the depth of a corresponding S to P conversion is calculated using the IASP91 velocity model and the RF time series are resampled as depth series. In P receiver function studies, the equation

$$T_{Pd} (d) = \int_0^d \frac{1}{V_S^2(z)} - p^2 - \frac{1}{V_p^2(z)} - p^2 \right]dz$$

is often used for time to depth conversion (Dueker and Sheehan, 1995). It relates the delay time $T_{Pd}$ to the depth of conversion $d$, given P and S velocities ($V_p$ and $V_S$) in overlying layers. This formulation assumes a constant horizontal slowness $p$ and may be referred to as a plane wave approximation. In P receiver function analysis this approximation will lead to negligible errors of delay times calculated for upper mantle wave conversion. For S to P converted waves however, the geometry of wave paths will lead to a bigger difference in the slownesses of the converted wave and the associated direct wave. In this case, depth migration using a plane wave approximation will introduce more severe errors even for phase conversion in the uppermost mantle. In the analysis of S to P converted wave included in this thesis (Paper II) this approximation was therefore not used. Instead the travel times of the direct S and the S to P converted wave were calculated from ray tracing using the TauP toolkit (Crotwell et al., 1999).

In two of the studies in this thesis (Papers III and IV) the phase associated with P to S conversion at the 410 km discontinuity has been studied. The
interpretation of the delay time of this phase has been in terms of velocities of overlying layers rather than depth. Representation of the moveout corrected RF data on a depth scale given by a reference velocity model would therefore have been misleading. Instead the data has been presented on a time scale relative to the reference model. For each event the delay time of P410s predicted by IASP91 (Kennet and Engdahl, 1991) was calculated. Aligning the data to this time will correct sufficiently for moveout and allow for stacking in a narrow time window surrounding the predicted delay time. The data represented as time series on this reduced time scale was used in the analysis in Papers III and IV.

Trade-off between P and S velocities

It follows from equation 5 that receiver function lag times are not fully capable of resolving the trade-off between conversion depth and velocities. To what degree this will complicate the interpretation depends on what depth range is being studied and what resolution we aim to achieve. Depths of shallow converters can be determined from lag times with an accuracy of a couple of km even with a very vague idea of P and S velocities. The uncertainty in conversion depths due to uncertainties in velocities will grow with depth. In Paper III, delay times of P410s were found to be 1-2 seconds less than predicted by the IASP91 reference model. Using this velocity model for depth migration will lead to an estimated depth of the 410 km discontinuity of 390-400 km. Most of this apparent shallowing could probably be attributed to upper mantle velocities rather than true depth of conversion.

Even with a fixed depth of conversion, receiver function lag times cannot resolve the velocities of overlying layers due to the trade-off between P and S velocities that follows from equation 5. This is illustrated in Figure 2. For a range of perturbations to the upper mantle velocities of IASP91, the delay times of P to S conversions at 410 km are calculated and compared to the delay time predicted by IASP91. Trade-off curves for fixed delay times are shown in the figure calculated for a horizontal slowness of 6 sec/°.

The relation between lag time, conversion depth and velocities will change slightly for data with different horizontal slowness. It would therefore appear possible to constrain both the depth and the P and S velocities in the overlying layers for say the 410 km discontinuity, by using a wide range of slownesses. This has been attempted by some authors by stacking RF data in a space spanned by \( V_p, V_s \) and conversion depth (Gurrola et al., 1994). However, this possibility probably only exists in theory and in the presence of a truly one-dimensional velocity structure. It is clear from Figure 2 that large changes in delay times can be given by fairly small perturbations to \( V_p \) and \( V_s \). Using a wide range of horizontal slowness, the data will sample a large volume beneath the station that will most certainly contain such small
velocity variations. Also the limited slowness range (5-8 sec/°) of data used in RF analysis will only make a small change in slope of velocity-depth trade off curves based on equation 5.

In an attempt to resolve P and S velocities from P410s delay times despite the severe trade-off inherent in this kind of data, a tomographic approach was used. Assuming that the depth of conversion is fixed and that any variation seen in P410s delay times is due to variation in overlying velocities, receiver function delay times can easily be set up as a tomographic equation to solve both for P and S velocities. In Paper III in this thesis, the ability of tomographic inversion of P410s delay times was studied using a linearised 2-D inversion. While the resolution is limited due to the trade-offs identified, the method potentially provides access to absolute velocities, as opposed to the relative velocity perturbations normally produced from teleseismic arrival time tomography.
3. Methods of converted wave analysis

In the analyses of receiver function data presented in this thesis, various methods have been employed. In the following section, some fundamental aspects of these methods will be discussed in further detail.

Determination of Moho depth and \( \frac{V_p}{V_S} \) from stacking of receiver functions

To estimate the bulk crustal properties below seismic stations, the method suggested by Zhu and Kanamori (2000) was used in Paper I. This method uses the delay times of three phases that are often clearly visible in the receiver functions to constrain a solution for Moho depth and average \( \frac{V_p}{V_S} \).

The phases used are the Moho P to S conversion (\( P_{ms} \)) and the converted crustal multiples (\( P_{pPms} \) and \( P_{pSms} \)) as illustrated in Figure 3a. The delay time of the phase converted from P to S at Moho depth \( m \) is given by

\[
\tau_{P_{ms}} = \frac{m}{\sqrt{\frac{1}{V_S^2} - p^2}} \left( \frac{1}{\sqrt{V_p^2 - p^2}} \right)
\]

(6)

where \( V_p \) and \( V_S \) are the (time integrated) average P and S velocities above \( m \) and \( p \) is horizontal slowness. Similarly

\[
\tau_{P_{pPms}} = m \left( \sqrt{\frac{1}{V_S^2} - p^2} + \sqrt{\frac{1}{V_p^2} - p^2} \right)
\]

(7)

and

\[
\tau_{P_{pSms}} = 2m \sqrt{\frac{1}{V_S^2} - p^2}
\]

(8)

give the delay times of the multiples \( P_{pPms} \) and \( P_{pSms} \). For a given pair of
Figure 3. a) Phases used to estimate crustal properties. b) $V_p/V_S - m$ trade-off curves for the phases.

$m$ and $V_p/V_S$ values, a weighted sum of the amplitudes of the receiver functions for different events at the delay times corresponding to the phases given by the above equations is computed. A grid search is performed in the space spanned by $m$ and $V_p/V_S$ and the best estimate of the average value of these parameters for the crust beneath a station is taken to be the pair for which the amplitudes stack coherently and the weighted sum reaches a maximum (Figure 3b).

Tomographic inversion of P410s delay times

In Paper III, a linear inversion of P410s delay times is used to simultaneously solve for both P and S velocities. A constant depth of conversion at 410 km is set as the base of the model and the model is split into cells in the normal tomographic manner using a 2-dimensional representation.
With the notation of Figure 4, the delay time of P410s can be written as

$$T_{Pds} = \sum B_i \frac{1}{\beta_i} - \sum A_i \frac{1}{\alpha_i} - p_s X_S + p_p X_P$$

where $A_i$ and $B_i$ are the distances travelled in the $i$:th cell by the direct P and the converted S respectively. $\alpha_i$ and $\beta_i$ denote the P and S velocities in the $i$:th cell and $p_p$ and $p_s$ are the horizontal slownesses of the P and converted S. Horizontal offsets of the wave paths at the base of the model are $X_P$ and $X_S$. The first two terms represent the travel times of the P and S legs through the model. The latter two terms correct for the difference in arrival time of the P and converted S wave paths at the base of the model. To linearise the equation, we use the background model to calculate $A_i$ and $B_i$ and assume the sum of the last two terms of the equation to be constant. With these assumptions, we can form the delay time anomaly relative to that predicted by the reference model as

$$T^\text{obs}_{Pds} - T^{0}_{Pds} = \sum B_i \left( \frac{1}{\beta_i} - \frac{1}{\beta_i^0} \right) - \sum A_i \left( \frac{1}{\alpha_i} - \frac{1}{\alpha_i^0} \right)$$
With N data we have N such equations and form the matrix equation:

\[ Gm = d \]  \hspace{1cm} (11)

where

\[
G = \begin{bmatrix}
B_{11} & \cdots & B_{1m} & -A_{11} & \cdots & -A_{1m} \\
\vdots & \ddots & \vdots & \vdots & \ddots & \vdots \\
B_{n1} & \cdots & B_{nm} & -A_{n1} & \cdots & -A_{nm}
\end{bmatrix},
\]

\[
m = \begin{bmatrix}
\frac{1}{\beta_i} - \frac{1}{\beta_m^0} \\
\vdots \\
\frac{1}{\beta_m} - \frac{1}{\beta_m^0} \\
\frac{1}{\alpha_i} - \frac{1}{\alpha_i^0} \\
\vdots \\
\frac{1}{\alpha_m} - \frac{1}{\alpha_m^0}
\end{bmatrix}
\]

and \( d = \begin{bmatrix} \delta T_1 \\ \vdots \\ \delta T_n \end{bmatrix} \).

We thus arrive at a system of linear equations that can be inverted in the least square sense using standard matrix manipulation routines. To stabilise the inversion, smoothing and damping of the inversion is achieved by solving:

\[
\begin{bmatrix}
G & w_1 \mathbf{D}_h \\
w_2 \mathbf{D}_v & m = \\
w_3 \mathbf{I}_s & 0 \\
w_4 \mathbf{I}_p & 0
\end{bmatrix}
\begin{bmatrix}
d \\
0 \\
0 \\
0
\end{bmatrix}
\]

\hspace{1cm} (12)

where \( \mathbf{D}_h \) and \( \mathbf{D}_v \) are second derivative matrices for the horizontal and vertical directions providing smoothing, \( \mathbf{I}_s \) and \( \mathbf{I}_p \) are identity matrices for S and P model length damping and \( w_k \) are damping weights.
Analysis of shear-wave splitting of upper mantle converted waves

In Paper IV, the polarisation of P410s is analysed to study upper mantle anisotropy in a manner similar to the analysis of shear-wave splitting of SKS. To enhance the weak signal of P410s, P receiver functions are stacked in clusters of events that are closely spaced in back azimuth and slowness. This is to ensure that wave paths of included events sample the same rock volume. The polarisation of the stacked data is represented in an azimuth-time domain. At 1° increment in azimuth, the projections of the $SV$ and $T$ components of the moveout corrected receiver functions are summed according to

$$S(t,\psi) = \sum_i [T_i(t)\sin(\psi - \phi_i) - SV_i(t)\cos(\psi - \phi_i)],$$

where $\phi_i$ is the back azimuth of the $i$-th event, $t$ delay time and $\psi$ is the variable azimuth.

To retrieve the splitting parameters, two methods are suggested. Both methods apply back-rotations and back-shifts of components in order to reconstruct the signal prior to splitting. In the first method, the corrected signal is assumed to contain minimum tangential noise in full analogy to the method often used in the analysis of shear-wave splitting of SKS (Savage and Silver, 1993).

The residual tangential noise is formed by the sum

$$\tau(dt,\psi) = \sum_t [\sin(\psi - \phi)S(t,\psi) + \cos(\psi - \phi)S(t - dt,\psi + 90°)]^2,$$

where the summation is computed in an interval $[-2 \leq t \leq +2]$ surrounding the position of the maximum peak found on the radial component, assumed to be P410s. For the event back azimuth $\phi$, the median of the stacked data was used. A grid search is performed for time shifts in the interval $[-2 \leq dt \leq +2]$ and for azimuths within 90° counter-clockwise from the radial direction of the stacked RF. The position of the minimum in the sum is interpreted as the split time between the fast and slow axes and the orientation of the fast or slow axis depending on the polarity of the time split.
The second method instead searches for a maximum peak on the radial component of the corrected signal. For each time shift $dt$ and polarisation angle $\psi$, the corrected components are summed and the radial peak is assigned to the maximum projection on the radial component:

$$\rho(dt, \psi) = \max \left\{ \cos(\psi - \varphi)S(t, \psi) + \sin(\psi - \varphi)S(t - dt, \psi + 90^\circ) \right\} \quad (15)$$

A grid search is then done for the splitting parameters $dt$ and $\psi$ that give the maximum radial peak.
4. Data

SNSN

The data analysed in the four studies included in this thesis are all teleseismic data recorded by broadband seismological stations (Figure 5) in the Swedish National Seismic Network (SNSN). The number of SNSN stations has grown rapidly in the last few years. In 1998 modern digital instruments were installed at six sites where an older analogue seismograph network had been operated by Uppsala University since the 1960's. Since then, several generations of seismic stations have been added to the network. In 1999-2000, 12 stations were installed along the coast of the Gulf of Bothnia. 20 more stations came into operation in south-east Sweden in 2000-2001. Further additions to the network were made in the far north of Sweden in 2003-2004 and in the south-west in 2006. At present, the number of SNSN stations in operation is 59.

Stations are equipped with Guralp CMG 3T broadband seismometers. These are flat in response to velocity for frequencies between 30 seconds and 50 Hz. Within the network, routine analysis such as event detection, location and source parameter estimation for local events is performed by automatic processes similar to the SIL-system in Iceland (Böðvarsson and Lund, 2003). Due to financial and technical limitations, most of the stations have so far been connected via dial-up telephone modem and only a few stations have provided continuous data via broadband data connection. However, the number of on-line stations is likely to increase in the coming years. Teleseismic data (the data used in the studies in this thesis) are automatically saved from stations based on event hypocenter and origin time information contained in e-mail messages from the National Earthquake Information Center (NEIC) at United States Geological Survey (USGS).

Data distribution

Event locations for data used in this thesis are shown in Figure 6. Figure 6a shows the locations of 78 events providing high quality P wave data (SNR > 8.0) that were used in the tomographic inversion of P410s delay times in Paper III. For P receiver function studies, events in a broad distance range (30º-90º) can be used. This gives a fair azimuthal coverage of the data, al-
Figure 5. SNSN stations in operation 2007.

though events to the south are rare. Earthquakes in the North Pacific form the dominant event cluster in this dataset and these data are often of very high quality. Back azimuths of these events at SNSN stations are north to north-east. Other source regions of events used for P receiver function studies are South East Asia, Middle East Asia and Central America.

The distribution of 48 events used in the study of S to P conversions (Paper II) is more sparse. Firstly, to see waves converted from S to P at depths down to the Lithosphere-Asthenosphere boundary, epicentral distances are required to be at least 65°. Secondly, the S arrival may be obscured by the arrival of other phases or other phases may complicate the interpretation of S
to P conversions. The events providing S data for Paper II are shown in Figure 6b. In this analysis, the events in northernmost Pacific could not be used since short event distances did not permit study of deep enough conversions. The dataset for this study is dominated by events in two clusters in South-East Asia and along the Pacific coast of East Asia, arriving at SNSN stations at back azimuths to the east and north-east.

Figure 6. Distribution of events providing the data for this thesis. a) P data. b) S data.
5. Crust and upper mantle of the Baltic Shield

The Baltic Shield, also often referred to as the Fennoscandian Shield, is an old and stable environment where only the outmost rims of its ca. 1000 km radius have been exposed to any major tectonic activities in the last billion years. Starting from its Archaean core to the northeast, the Baltic Shield is viewed as having accreted through a series of Precambrian orogenies (Gaal and Gorbatschev, 1987). The area covered by the SNSN stations providing the data for the studies in this thesis lies mainly in the Svecofennian Domain (SD). The Svecofennian orogen has been interpreted as a complex sequence of collisional and extensional phases that in a relatively short time interval (1.95-1.80 Ga) built up large parts of the interior of the Baltic Shield. The later accreted Southwest Scandinavian Domain (1.7-1.5 Ga) is separated from the SD by the Transscandinavian Igeous Belt and the Protogine Zone. The Baltic Shield is delimited to the south by the Tornquist Zone and to the west by the Caledonides.

Several active seismic experiments have been conducted to study the tectonic evolution of the Shield. The FENNOLORA (Lund, 1980) seismic refraction profile formed the northern segment of the European Geotraverse (Blundell, 1999). Owing to the length of the profile (~1800 km), data from this experiment have been used to study not only crustal structure (Guggisberg et al., 1991; Lund et al., 2001), but also the upper mantle down to depths of almost 400 km (Guggisberg and Berthelsen, 1987; Perchuć and Thybo, 1996; Abramowitz et al., 2002). The reflection seismic BABEL experiment (BABEL working group 1993a, 1993b), shot off-shore in the Baltic Sea and the Gulf of Bothnia, was focused to image the crustal imprints of the many tectonic episodes that built the Baltic Shield. Seismic experiments on a (relatively) smaller scale have been aimed at specific tectonic boundaries. The BLUE ROAD (Lund, 1979) and CCT (Juhojuntti et al., 2001) experiments studied the structure of the Scandinavian Caledonides. EUGENO-S (EUGENO-S Working Group, 1988) investigated the transition from the Precambrian Shield to the younger tectonic provinces to the south. Several seismic profiles (Grad and Lousto, 1987; Korsman et al., 1999) have also studied the Archaean-Proterozoic boundary in the Finnish part of the Baltic Shield.

In recent years passive seismological experiments using temporary arrays that have been carried out on the Baltic Shield and neighbouring areas include TOR (1996-1997) designed to image the Trans-European Suture Zone.
from northern Germany to southern Sweden, and SVEKALAPKO (1998-1999) across the Proterozoic-Archaean suture in Finland. The teleseismic data recorded at these arrays have been used for travel time tomography (Shomali et al., 2002, Sandoval et al., 2004), surface waves (Bruneton et al., 2002) and receiver function studies (Gossler et al., 1999; Wilde-Piórko et al., 2002; Alinaghi et al., 2003). The data from the SNSN stations provides a completely new teleseismic data set bridging the gap between the TOR and SVEKALAPKO areas.

The distribution of SNSN stations gives the potential to observe signals of crust and upper mantle structures related to various stages of the tectonic evolution of the Baltic Shield. In the north of Sweden, the Archaean-Proterozoic boundary is sampled by SNSN data and to the south, the transition to younger tectonic provinces. Within the Svecofennian Domain, several tectonic subprovinces have been identified and the interpretation of the teleseismic SNSN data can certainly be made in relation to those.
6. Summary of papers

Paper I

**Moho depth variation in the Baltic Shield from analysis of converted waves**

Mapping the variation of crustal thickness, or of the seismological observable associated with it, the Moho depth is a useful tool for the interpretation of the tectonic evolution of regions at length scales appropriate for e.g. The Baltic Shield. Previous estimates of Moho depths in the Baltic Shield area have mostly been based on interpretation of controlled source seismic experiments, such as FENNOLORA (Lund, 1979) and BABEL. For large areas of the Baltic Shield Moho depth data have not been available. In this study, new Moho depth data from broadband stations in the SNSN are presented. These Moho depth estimates are based on analysis of teleseismic data using the receiver function (RF) technique.

The method of Zhu and Kanamori (2000) was used for the analysis. This method uses the Moho P to S conversion and reverberations within the crust to constrain average Moho depth and $V_p/V_s$ of the crust beneath the recording station. It is described in more detail in chapter 3 in this thesis. Data from 52 stations were analysed. Reliable solutions for Moho depth and $V_p/V_s$ could be obtained for 41 stations. Solutions obtained at five stations were not fully constrained due to weak amplitudes of crustal reverberations. However, the Pms was clear and Moho depths were constrained assuming a fixed value for $V_p/V_s$. The noisy data recorded at six stations did not permit reliable Moho depth determination.

Results of the analysis are discussed in comparison to previous Moho depth estimates from e.g. refraction studies. A contour map of Moho depths from a compilation of previously available data is shown in Figure 7. In Figure 8 these data have been complemented with the new Moho depth estimates from the analysis of RF data. The new Moho depth estimates are found to be in excellent agreement with previously available Moho depth data where these have existed, confirming the reliability of both these new and previous results. In areas where Moho depth data have not been available, the inclusion of these new data has significantly constrained Moho depth variation in the Baltic Shield.
Figure 7. Moho depth map contoured from a compilation of previously available data (Korsman et al., 1999).
Figure 8. Moho depth map contoured from the dataset used for Figure 7 complemented with Moho depth estimates from this analysis. Triangles show locations of SNSN stations.
In this paper four specific regions are discussed where previous data have been sparse and hence the addition of new data provides the most significant contribution. For the eastern Bergslagen area (marked C in Figure 8), the addition of new data was found to confirm the image perceived by contouring the previously available sparse data. New Moho depth data obtained for several sites in the far north of Sweden (A in Figure 8) agree with a general view of the large scale structure of the area and with previous data where available. It was observed however that using only previously available data, the sparse sampling in this area might produce misleading artefacts in the contoured Moho depth map. Including also the new data significantly improved the map. Two zones of enhanced crustal thickness (B and D) were confirmed by the new Moho depth data. Addition of new Moho depth data also helps constrain the geometry of these zones. For the zone of enhanced crustal thickness in the southern Gulf of Bothnia area (B), the apparent double Moho signal in the RF data agrees with the interpretation of the crustal thickening as being due to mafic underplating (Nironen, 1997) established as a 10-15 km layer of intermediate seismic velocity in the lowermost crust. The Moho signal seen in RF data sampled in southern Sweden suggests a different nature of the crustal thickening seen in this region. An intrusive body extending up to depths of 20-25 km has been proposed (Korja and Heikinnen, 2005) based on reflection data and would agree with the conversion signals seen in RF data.

Paper II

Analysis of waves converted from S to P in the upper mantle beneath the Baltic Shield

Seismic refraction experiments on the Baltic Shield have revealed considerable layering of the lithospheric mantle (Guggisberg and Berhelsen, 1987; Perchü and Thybo, 1996). Lithospheric structure seen in refraction data have also been confirmed by analysis of surface waves (Bruneton et al., 2004). In the present study, the lithospheric upper mantle has been studied using analysis of waves converted from S to P. The S receiver function (SRF) technique is very well suited for study of the lithosphere since the depth range 50-200 km will not be obscured by signals from crustal reverberations as is the case for the traditional P receiver function.

To image upper mantle structures causing S to P conversions, the amplitudes of the SRF are migrated into their spatial position in depth. At 1 km increments in depth, the advance times of corresponding phase conversions and their positions are calculated assuming IASP91 velocities. The depth migrated SRF are then projected onto a 2D-profile along the network and amplitudes are summed in 1km x 1km grid cells. Structural variation is imaged by plotting amplitudes using a polar colour scale ranging from blue for
negative conversion amplitude to red for positive amplitudes. Considering particle motions of the direct S wave and the converted P wave implies that negative conversion amplitudes correspond to phase conversions at positive velocity contrasts (velocity increases with depth), whereas positive amplitudes suggest negative velocity contrasts.

The results (Figure 9) of this study suggest the existence of pronounced layering of the uppermost mantle of the Baltic Shield at depths down to at least 200 km. This is based on the coherent structures seen by S to P conversions, in some cases continuous for up to 1000 km. For the southern and central parts of the profile, a sharp contrast is seen at depths around 160 km with a polarity indicating that velocity increase with depth (marked C in Figure 9). This coincides with the velocity contrast interpreted by several authors as the bottom of a low velocity zone (Thybo, 2006). Signals at depths around 100 km (A and B) might be related to a high velocity lid on top of this zone. However, the SRF data clearly suggests that the proposed LVZ has internal structure (D1 and D2). A clear feature which can be identified as the lithosphere-asthenosphere boundary is imaged at depths around 200 km (L). Deepening of this and other observed features of the profile occur around 65°N which in this part of the Baltic Shield corresponds to the boundary between Archaean and Proterozoic terranes.

Figure 9. The upper mantle beneath the Baltic Shield as seen by S to P converted waves. Spatially smoothed amplitudes of depth migrated S receiver functions are shown using a polar colour scale. Blue represents negative SRF amplitudes, corresponding to phase conversion at a contrast with a velocity increase with depth. Red represents positive amplitudes from a velocity decrease. Structures marked in the figure are discussed in the text.
Paper III

**Tomographic inversion of P410s delay times for simultaneous determination of P and S velocities of the upper mantle beneath the Baltic Shield**

Depths to the upper mantle discontinuities at 410 km and 660 km have been observed to show little variation in tectonically stable areas such as the Baltic Shield. It has also been demonstrated that on a global scale there exists a clear correlation between delay times of upper mantle converted phases observed in receiver functions and upper mantle velocities found from tomography (Chevrot et al., 1999). This suggests that observed anomalies in delay time (relative to a reference model) can be attributed to anomalous upper mantle velocities rather than anomalous depth of conversion. Although this has been observed and noted on by several receiver function studies, attempts to assess upper mantle velocities using receiver function delay times have not been made. In Paper III of this thesis, tomographic inversion of P410s delay times has been used to resolve both P and S velocities of the upper mantle beneath the Baltic Shield.

The formulation of the tomographic equation is presented in further detail in chapter 3 in this thesis. A 2-dimensional simplification was used for the model parameterisation. This was considered justified by the narrow coverage of the 33 stations providing the data used in this study. A fully linearised approach was used and no iterations were done to update wave paths. The P410s delay times used as the input data for the inversion could be read from individual receiver functions after employing a strict data quality criterion (SNR > 8.0). A suite of synthetic tests was performed to study the resolution and potential artefacts of the method and to determine appropriate damping and smoothing parameters. Synthetic tests demonstrate that the method is capable of resolving P and S velocity variation, despite the observed trade-off between these. Lateral resolution is good for both \( V_p \) and \( V_s \). Vertical resolution is poorer for \( V_s \), which is certainly due to the steeper incidence angles of the converted S waves.

Results from application of this method to real data appear to be consistent with independently produced mantle velocity structures deduced from traditional arrival time tomography. For the P velocity model (Figure 10a), a north-dipping body of (relatively) low velocity is found for the central part of the profile at 58°-64°N. A similar feature is observed in P-wave first arrival tomography (Eken et al., 2007a). Within the central part of the profile, where resolution is best due to the geometry of the system, even smaller details of the topography of the deep low velocity structure appear to correspond. Due to poorer vertical resolution, the S velocity model (Figure 10b) is more difficult to interpret. Also, no S velocity model was available for com-
Figure 10. Velocity models obtained from tomographic inversion of P410s delay times using data from SNSN stations. a) P velocity model. b) S velocity model. c) VP/VS model.

In the VP/VS model (Figure 10c), the low velocity zone seen clearly in the VP model is evident, but apparently broader in depth, which might be considered consistent with vertical smearing in the VS model. To the north, a very clear apparently near-vertical feature is seen. Since ray coverage in this part of the profile is relatively poor, this feature may not be reliably resolved. Nevertheless, it is interesting to note that it corresponds to the known surface major boundary between the Proterozoic and Archaean part of the area.
Paper IV

**Velocities and anisotropy as seen by upper mantle converted waves**

This manuscript demonstrates two aspects of upper mantle seismic structure as seen by upper mantle converted waves. The phase used is P410s, the P to S conversion at the upper mantle discontinuity at ~410 km.

In the first part of this study, upper mantle velocity and velocity variation is seen in the P receiver function as the delay time and delay time variation of the peak associated with P410s. Delay times of the peaks are read after stacking. Data are stacked in fairly wide spatial windows according to their 210 km depth point of the converted shear wave. Before stacking, receiver functions are moveout corrected by aligning them to their P410s delay time predicted by the IASP91 reference model. The residual delay times of the spatially stacked data projected on the map show variations that can be interpreted as reflecting large scale variations in the average velocity structure of the upper mantle (Figure 11).

Three zones of (relatively) late arrivals can be seen in the image. To the north, the sharp contrast in P410s delay times around 66°N corresponds to the border between Archaean and Proterozoic parts of the Baltic Shield. This is interpreted as a transition to relatively low P and S velocities. The same interpretation is made for the southern tip of the array, where later arrivals appear to correlate with the transition to younger tectonic terranes at 56°-57°N. These interpretations are in agreement with velocity models independently derived from P and S arrival time tomography (Eken et al, 2007a, 2007b). A narrow band of late arriving P410s is also imaged across the profile at 62°-64°N. This is interpreted as a zone of relatively high $V_p/V_S$.

A second part of this study investigates the possibility to study upper mantle velocity anisotropy from the polarisation of P410s. Receiver function stacks are formed from event clusters closely spaced in back azimuth and slowness. The polarisation of the stacked data is interpreted as splitting of the converted shear wave due to velocity anisotropy in the upper mantle. Two methods are used to determine splitting parameters. The first method is based on a criterion of minimum tangential noise on the initially radially polarised, non-split pulse. In the other method, the criterion used is instead a maximum radial peak. Both methods are described in more detail in chapter 3 in this thesis.

Estimated splitting parameters were judged as reliable when both methods gave consistent solutions. From 184 data stacks recorded at 52 stations, reliable solutions were obtained for 116 stacks at 49 of the stations. 58 were considered null solutions and 68 showed a split of the converted shear wave (Figure 12).
Figure 11. Map of the variation of P410s delay times.
Figure 12. Split parameters determined from analysis of P410s. Arrows show fast split directions. Lengths of arrows are proportional to the estimated time split between fast and slow directions. Crosses show the possible split axes in null solutions. Arrows and crosses are projected to the 210 km depth points of the P410s.
Split orientations determined for stations in the Archaean part of the Shield north of 66°N appear homogeneous in the image, with fast directions striking NE-SW. This agrees with results from shear wave splitting analysis of SKS-data recorded by SVEKALAPKO-stations in the Archaean part of Finland (Vecsey et al., 2007). Results obtained around 65°N yield null solutions. However, due to the back azimuths of the events studied, it is impossible to assess whether this change in the imaged pattern reflects a true feature of the Archaean-Proterozoic transition. Further south, NE-SW striking fast split directions are again found for stations down to 62°N. Central parts of the array (57°-62°N) show a complex pattern of determined split orientations. Results for the southernmost stations (south of 57°N) agree with anisotropic models suggested from analysis of data from the TOR experiment (Plomerova et al., 2002).
7. Concluding remarks

As with any geophysical method, the analysis of converted waves has its strengths and its weaknesses. Observed wave conversion may be directly interpreted in terms of subsurface converting structures that have geological significance. The use of deconvolution when processing data also simplifies the analysis. We do not rely on consistent picking of phases in the seismograms. We also do not have to consider instrument responses, provided of course that they are equal on all components. Weaknesses of converted wave analysis are primarily in the control of velocities. Receiver functions are in a way sensitive to velocity contrasts rather than absolute velocities. There are also problematic trade-offs between conversion depths and P and S velocities in observed delay times.

In some of the studies presented in this thesis, I have made direct use of the strength of converted wave analysis as a tool to detect and map discontinuities within the Earth. In others, I have made methodological development to overcome its inherent trade-offs and to provide means of studying aspects of Earth’s structure that are conventionally studied using other types of seismic data.

The analysis of P receiver functions in Paper I was used to map Moho depth variation. For a few stations, the method failed, partly because receiver functions have little control on absolute velocities. For a majority of stations, however, clearly identifiable Moho conversions could be observed and Moho depths could be estimated. The main contribution of this study is the addition of new Moho depth data that helps constrain Moho depth variation in areas where previous estimates have not been available.

In Paper II waves converted from S to P were used to map discontinuities in the lithosphere. Since detecting discontinuities is what converted phases do best, this analysis serves as a valuable complement and an independent confirmation of lithosphere models that have been suggested from analysis of refraction seismic data.

In Paper III P410s delay times were used for velocity inversion. Synthetic tests indicate that the method has the potential to resolve both P and S velocities. Application to real data also shows agreement with results from arrival time tomography. The focus of this study was to investigate the capability of the method. The results obtained suffer from limited resolution. This could in part be explained by the 2-D parameterisation used. To some extent the resolution should be attributed to the trade-offs between P and S veloci-
ties. One important observation is, however, that the formulation of this inversion uses absolute velocities rather than relative velocities that are often used in teleseismic arrival time tomography. Information about absolute velocities may be very important in the mineralogical interpretation of the observed velocities. As a tool to gain further geological insight by imposing constraints on absolute velocities, the tomographic formulation using delay times of converted waves could therefore be used jointly with P and S arrival time data.

The analysis in Paper IV is also more method-oriented. Shear-wave splitting parameters were determined from P410s and interpreted as being due to velocity anisotropy. This interpretation constrains the depth of the anisotropy causing the observed splitting to within the uppermost 410 km. In the more traditional method of shear-wave splitting analysis using SKS-data, the depth of anisotropy cannot be directly constrained by the observed splitting. For some parts of the area covered by these data, previous splitting results are available from analysis of SKS-data recorded by passive seismic experiments. The good agreement found between the new and the previous results suggests that the shear wave splitting observed in the analysis of SKS can indeed be attributed to anisotropy in the uppermost mantle. The results of this analysis should also be compared to splitting parameters obtained from analysis of SKS in data recorded by the SNSN stations. These data have not yet been used for shear wave splitting analysis, but are likely to be in the near future.

The ambition of the studies presented in this thesis has been to explore the capabilities of converted wave analysis, rather than to obtain an optimal assessment of the Earth's structure. Any attempt to do so properly should involve a combined interpretation of results from studies using different methods and independent datasets. The use of converted waves in such analysis can serve as a valuable complement to other geophysical data.
Trots att seismologin är en relativt ung vetenskap, har den under de senaste 100 åren starkt bidragit till vår förståelse av jordens inre. Redan i den instrumentella seismologins ungdom kunde Oldham år 1906 sluta sig till att jorden har en flytande kärna som skapar en skuggzon för seismiska vågor på avstånd över $102^\circ$. Den kroatiske seismologen Mohorovicic visade 1909 på en skarp kontrast i seismiska hastigheter mellan Jordens yttre skorpa och dess mantel. Denna seismiska diskontinuitet bär numera namn efter sin upptäckare, på seismologisk jargon förkortat till ”Moho”. Utvecklingen av seismometrar samt den kraftiga ökningen i antalet seismometrar runtom jorden har sedan dess givit en allt klarare bild av jordens uppbyggnad. Denna utveckling har under senare årtionden accelererat genom användningen av digitalt dataformat och allt kraftfullare datorer.

De senaste tio åren har antalet stationer i Uppsala universitets seismiska nätverk (Swedish National Seismic Network, SNSN) ökat kraftigt. 1998 installerades moderna digitala seismometrar på sex platser där Uppsala universitet sedan 1960-talet drivit ett nätverk av analoga stationer. Under 2007 kommer antalet stationer i universitetets nät att nå 60. De data som registrieras på dessa nya stationer gör det möjligt att med hög upplösning studera strukturer i jordens inre. Denna avhandling är en sammanläggning av fyra studier där seismiska data registrerade av stationer i SNSN har använts för studier av jordens skorpa och mantel.

Analys av konverterade vågor

I seismologin identifieras ett antal vågfaser utifrån egenskaper hos vågutbredningen. Man skiljer tex mellan ytvågor, som utbredes sig längs Jordens yta, och volymsvågor, som forplantar sig genom Jordens inre. En än finare indelning av volymsvågor kan göras i longitudinala vågor (P-vågor) som har sin partikelrörelse i vågens utbredningsriktning, samt transversella vågor (S-vågor) med en partikelrörelse vinkelrätt mot utbredningsriktningen. En gemensam nämnare för de fyra studier som presenteras i denna avhandling är att de alla bygger på analys av konverterade faser. Med en konverterad vågfas avses en omvandling mellan olika typer av vågutbredning. Vid seismiska diskontinuiteter kommer en del av energin att omvandlas, konvertera från till exempel P- till S-våg. Analys av konverterade vågor ger oss därför ett verktyg för att detektera och kartlägga geologiska strukturer som ger
skarpa kontraster i bergsmassans seismiska egenskaper. Den konverterade vågen och den direkta, icke-konverterade vågen utbreder sig dessutom med olika hastigheter. Observationer av skillnaden i ankomsttid mellan dessa kommer därmed även att avspeglja seismiska hastigheter.


Sammanfattning av artiklar

I denna avhandling studeras strukturer på olika djup under den Baltiska, eller Fennoskandiska urbergsskölden genom analys av konverterade vågor i seismiska data registrerade av stationer i SNSN. De fyra delstudierna presenteras i separata artiklar, sorterade efter det djup som analysen fokuserar på. I artikel I presenteras en kartläggning av Mohodjupets variation från analys av konvertering från P till S vid den skarpa seismiska diskontinuitet som skiljer jordens skorpa från dess mantel. I artikel II analyseras istället Sp-konverterade vågor för att påvisa strukturer och skiktning i litosfären (den allra översta delen av övre manteln). I de båda avslutande delstudierna används fasen P410s, en omvandling från P till S vid en diskontinuitet i övre manteln på ett djup av 410 km. Tidsfördröjningen på P410s används i artikel III som indata för en tomografisk inversion för att bestämma hastighetsmodeller för den övre manteln. I artikel IV analyseras polariseringen av P410s för att studera seismisk anisotropi i övre manteln.

Artikel I

Kartläggning av variationer i jordskorpans tjocklek är en betydelsefull komponent för tolkningen av den tektoniska utvecklingen av den Baltiska skölden. Tidegare uppskattningar av Mohodjup i området har i stor utsträckning grundats på resultat från refraktnsseismiska samt reflektionsseismiska experiment, såsom FENNOLORA och BABEL. I denna studie presenteras en kartläggning av Mohodjupets variation baserad på analys av teleseismiska data registrerade av stationer i SNSN.

Resultaten diskuteras i jämförelse med tidigare uppskattade Mohodjup i området (Figur 7 och Figur 8). God överensstämmelse uppvisas i de delar där tidigare uppskattningar funnits tillgängliga. De mest betydelsefulla bidragen från denna analys fås i områden som tidigare inte undersöks i seismiska profiler. I artikeln pekas speciellt fyra områden ut. För övre Norrland samt östra Bergslagen har nya data bidragit till att Mohodjup kunnat kartläggas där tidigare djupdata saknats. Två zoner av kraftigt förtjockad jordskora kan i studien observeras i södra Bottenviken samt i östra Götaland. Analyser av PRF \( (P \text{ receiver functions}) \) tyder på att olika tektoniska mekanismer varit aktiva i skapandet av den tjocka skorpan i dessa två områden.

Artikel II

Refraktionsseismiska experiment har visat på betydande strukturer i litosfären under den Baltiska skölden. I artikel II studeras litosfären genom analys av Sp-konverterade vågor med SRF-metoden \( (S \text{ receiver function}) \). SRF-metoden är mycket väl lämpad för studier av strukturer i litosfären då det aktuella djupintervallet (50-200 km) inte störs av signaler från multipla reflektioner i jordskorn, vilket är fallet för PRF-metoden \( (P \text{ receiver function}) \).

För att åskådliggöra konverterande strukturer i djupet, migreras SRF med hastigheterna i den globala referensmodellen IASP91 (Kennet och Engdahl, 1991), projiceras på en 2-dimensional profil längs det seismiska nätet och utjämnas. Resultatet blir en bild (Figur 9) där blått används för negativa amplituder, vilket för SRF innebär en Sp-omvandling vid en diskontinuitet där.seismiska hastigheter ökat med djupet. På samma sätt kommer rött att svara mot lägre hastighet med djupet.

Resultaten visar på ett flertal koherenta strukturer i de övre 200 km, i vissa fall sammanhängande i upp till 1000 km. I södra delen av profilen syns en skarp kontrast på ett djup omkring 160 km. Denna sammanfeller med en struktur som av vissa forskare tolkats som basen för en utbredd låghastighetszon i continental mantel. En signal som med största säkerhet kan identifieras som harrörande från gränsen mellan litosfären och den underliggande astenosfären syns tydligt på djup omkring 200 km. Mot norra delen av profilen tycks denna och andra strukturer fördjupas omkring 65°N, vilket sammanfeller med gränsen mellan Proterozoiska och Arkeiska domäner av den Baltiska skölden.
Artikel III

De seismiska diskontinuiteterna i övre manteln på djup omkring 410 km och 660 km visar liten variation i djup i tektoniskt stabila områden som den Baltiska skölden. Det har även observerats att tidsfördröjningen på vågfaser som konverterat vid dessa diskontinuiteter korrelerar starkt med seismiska hastigheter bestämda med tomografi. Det finns därför goda skäl att förmoda att observerade variationer i tidsfördröjningen på P410s (Ps-omvandling vid 410 km) till största del avspeglar variationer i den övre mantelns seismiska hastigheter snarare än i djupet till diskontinuiteten.


Resultat från inversion av data från SNSN (Figur 10) visar god överensstämelse med resultat från traditionell tomografi baserad på ankomsttider. P-hastighetsmodellen visar en zon av (relativt) låga hastigheter under centrala delar av profilen vid 58°-64°N. En liknande struktur har observerats i tomografi med P-ankomster (Eken et al., 2007a). S-hastigetsmodellen är med anledning av sin sämre vertikala upplösning svårare att tolka. Vp/Vs-modellen liknar i stora delar P-hastighetsmodellen. En nästan vertikal kontrast i profilens norra delar sammanfaller med gränsen mellan Arkeiska och Proterozoiska delar av Baltiska skölden.

Artikel IV

Detta manuskript beskriver två aspekter av övre mantelns struktur som kan studeras genom analys av konverterade vågor. Den vågfas som använts är P410s, omvandling från P till S i en seismisk diskontinuitet i övre mantlen på ett djup omkring 410 km.

I manuskriptets första del kartläggs variation i tidsfördröjningen för P410s. Under antagandet att den observerade variationen avspeglar variation i seismiska hastigheter ovanför diskontinuiteten snarare än variation i dess djup, kan kartan över tidsfördröjningen tolkas som en karta över övre mantelns hastigheter. Kontraster i hastigheter kan i sin tur tolkas som geologiska strukturer. Resultatet av analysen blir således en karta över storskaliga strukturer i övre manteln under den Baltiska skölden (Figur 11).

Resultat från analysen projicerade på kartan (Figur 12) visar på ett mönster av anisotropi som i stor utsträckning sammanfaller med kända storskaliga strukturer i berggrunden. I vissa delar av det område som täcks av SNSN har tidigare tolkningar av anisotropi baserade på analys av shear-wave splitting av SKS-data samt ankomsttider för P-vågor varit tillgängliga. I dessa områden visar de nya resultaten på mycket god överensstämmelse med tidigare. Detta bekräftar att den shear-wave splitting som i tidigare studier observerats i SKS-data kan hänföras till anisotropa strukturer i den övre manteln.

För en mer pålitlig tolkning av anisotropi i de centrala delarna av profilen, där analysen visar på ett mycket komplettert mönster samt för att ytterligare säkerställa pålitligheten i analysmetoden bör resultaten dessutom jämföras med resultat från analyses av SKS-data från stationerna i SNSN.

Slutkommentar

Analys av konverterade vågor har liksom alla geofysiska metoder sina starka och svaga sidor. Å ena sidan ger den oss ett utmärkt verktyg att detektera och kartlägga geologiska strukturer med skarpa konstraster i seismiska hastigheter. Den ger också möjligheten att direkt djupbestämma anisotropier och hastighetsvariationer. Å andra sidan har vi i studiet av konverterade vågor dålig kontroll på absoluta hastigheter. Amplituden i våra data är främst känsliga för kontraster i seismisk hastighet och den tidsförskjutning vi kan uppmätta i data avspeglar en kombination av P- och S-hastigheter. Jag har i delar av mitt arbete direkt använt mig av fördelarna med denna typ av data. I andra, mer metodinriktade delar har jag sökt överkomma dess svårigheter.

Ambitionen i detta arbete har varit att utreda vad analys av konverterade vågor kan bidra till i studiet av jordens inre, snarare än att presentera en fullständig geologisk tolkning. En fullständig tolkning måste grunda sig på ett flertal skilda analysmetoder och oberoende data. I en sådan analys bör konverterade vågor ingå som ett självklart komplement till andra typer av geofysiska data.
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Scientific work is seldom the fruit of one person’s effort alone. This thesis is no exception. Several people have contributed to the work I have presented and should be duly acknowledged.

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I dedicate this work to my wonderful girls, Disa and Ebba.
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