Isotropic and Anisotropic P and S Velocities of the Baltic Shield Mantle

Results from Analyses of Teleseismic Body Waves

TUNA EKEN
Abstract


The upper mantle structure of Swedish part of Baltic Shield with its isotropic and anisotropic seismic velocity characteristics is investigated using teleseismic body waves (i.e. P waves and shear waves) recorded by the Swedish National Seismological Network (SNSN).

Nonlinear high-resolution P and SV and SH wave isotropic tomographic inversions reveal velocity perturbations of ± 3% down to at least 470 km below the network. Separate SV and SV models indicate several consistent major features, many of which are also consistent with P-wave results. A direct cell by cell comparison of SH and SV models reveals velocity differences of up to 4%. Numerical tests show that differences in the two S-wave models can only be partially caused by noise and limited resolution, and some features are attributed to the effect of large scale anisotropy.

Shear-wave splitting and P-travel time residual analyses also detect anisotropic mantle structure. Distinct back-azimuth dependence of SKS splitting excludes single-layer anisotropy models with horizontal symmetry axes for the whole region. Joint inversion using both the P and S data reveals 3D self-consistent anisotropic models with well-defined mantle lithospheric domains. These domains of differently oriented anisotropy most probably retain fossil fabric since the domains' origin, supporting the idea of the existence of an early form of plate tectonics during formation of continental cratons already in the Archean.

The possible disturbing effects of anisotropy on seismic tomography studies are investigated, and found to be potentially significant. P-wave arrival times were adjusted based on the estimates of mantle anisotropy, and re-inverted. The general pattern of the velocity-perturbation images was similar but changed significantly in some places, including the disappearance of a slab-like structure identified in the inversion with the original data. Thus the analysis demonstrates that anisotropy of quite plausible magnitude can have a significant effect on the tomographic images, and should not be ignored. If, as we believe, our estimates of anisotropy are reasonably correct, then the model based on the adjusted data should give a more robust and correct image of the mantle structure.

Keywords: teleseismic tomography, mantle lithosphere, seismic anisotropy, teleseismic earthquakes, shear wave splitting

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ISSN 1651-6214
urn:nbn:se:uu:diva-102501 (http://urn.kb.se/resolve?urn=urn:nbn:se:uu:diva-102501)
Dedicated to:
My parents and my love Tülay
List of Papers

This thesis is based on the following papers, which are referred to in the text by their Roman numerals.


III. Eken, T., J., Plomerová, R. Roberts, L. Vecsey, V. Babuška, H. Shomali, R. Bodvarsson (2008), Seismic anisotropy of the mantle lithosphere beneath the Swedish National Seismological Network (SNSN), *Submitted to Tectonophys*.


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<tr>
<td>ACH</td>
<td>Aki-Christoffersson-Husebye</td>
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<td>EEC</td>
<td>East European Craton</td>
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<tr>
<td>HTI</td>
<td>Horizontal Transversely Isotropic</td>
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<tr>
<td>LAB</td>
<td>Lithosphere- Asthenosphere Boundary</td>
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<tr>
<td>LPO</td>
<td>Lattice Preferred Orientation</td>
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<td>LQT</td>
<td>Longitudinal-Radial-Transverse</td>
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<td>MOP</td>
<td>Multi-objective Optimization Procedure</td>
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<td>PDE</td>
<td>Partial Differential Equation</td>
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<td>SNSN</td>
<td>The Swedish National Seismic Network</td>
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<td>SVD</td>
<td>Singular Value Decomposition</td>
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<td>TESZ</td>
<td>Trans-European Suture Zone</td>
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<td>TI</td>
<td>Transversely Isotropic</td>
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<td>TIB</td>
<td>Trans-Scandinavian Igneous Belt</td>
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<td>TSVD</td>
<td>Truncated SVD</td>
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<td>USGS</td>
<td>United States Geological Survey</td>
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<tr>
<td>VTI</td>
<td>Vertical Transversely Isotropic</td>
</tr>
<tr>
<td>WWSSN</td>
<td>World Wide Standardized Seismographic Network</td>
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<tr>
<td>ZRT</td>
<td>Vertical Radial Transverse</td>
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1. Introduction

One of the aims of Earth Scientists is to be able to generate reliable models regarding the physical properties (e.g. velocity, density etc.) of the Earth’s interior. This can be difficult, especially if the given modeling problem is associated with the deeper part of the Earth. There are numerous methods for obtaining physical models which depend either on advanced forward or inverse modeling techniques and which build upon geophysical or geological data, and sometimes some assumptions. Ultimately, none of these methods can be considered to be “ideal” or “true” unless there is direct confirmation of the observed information from a direct sample of the region of interest.

We have direct access via drilling to only the uppermost few kilometers of the Earth. For the deeper Earth - in our case depths of about 70 km to 500 km - drilling for samples is today technically impossible and we must thus generally rely on indirect methods including various types of geophysical or geological modeling. One such method is to utilize seismic waves. Seismic waves are by far the best tool for scrutinizing the Earth’s current deep structure (Figure 1.1). Over many decades robust knowledge about layering within the Earth and deviations from such layering, i.e. velocity heterogeneities, has been developed using the travel time or amplitude information extracted for various seismic phases in a seismogram. Such data provides, of course, not only information regarding current structure but also important clues to understanding past geodynamic processes linked to the plate tectonic paradigm of the Earth.

Seismic tomography was introduced by Aki and Lee (1976) who first applied methods from medical tomography to geoscientific problems. Such tomography is essentially an inverse problem and has become a well-established technique in constraining the three dimensional (3-D) properties of the Earth affecting seismic wave propagation, namely the elastic, anelastic, anisotropic and density properties (Thurber and Ritsema, 2007). Seismic waves passing through the Earth propagate energy which can be measured and quantified in various ways such as source-receiver travel times and amplitudes on seismic recordings. Such data can be used to infer information regarding the physical properties (e.g. density, attenuation, velocity etc.) of the layers in which seismic waves propagate. Resulting tomographic images are used in the analysis of the subsurface – lithology, temperature, fracturing, fluid content, etc (Thurber and Ritsema, 2007). There are numerous examples of the application of seismic tomography to different geographical
and depth regions by using different types of measurements. Which of the various tomography techniques can be applied in a specific case depends on the geometry of the measurement situation, including the source and receiver positions and the depth range of interest (local earthquake tomography, teleseismic tomography, global tomography etc.). Different types of data can be utilized in the inversion procedure including absolute or relative arrival times at recording stations, waveforms of body waves, or surface wave data. More comprehensive information regarding the principles of different seismic tomography methodologies and applications can be found in Aki and Lee (1976), Aki et al. (1977), Evans and Achauer (1993), and Thurber and Ritsema (2007).

Figure 1.1 Left: A cartoon illustration of the Earth’s interior (adapted after Beatty, 1990). Right: The ray paths of four different teleseismic phases (P, S, PcP, and ScS) (Figure is courtesy of Edward Garnero).

A proper estimate of seismic velocity distributions is one of the key tools for a better understanding of lithospheric and mantle dynamics and rheology. The effects of temperature, composition, the presence of partial melt or water and seismic anisotropy are parameters which affect the speed of seismic waves within the mantle (Goes et al. 2000). In general, regional or teleseismic tomography studies performed in different parts of the world have resulted in models with velocity perturbations of ~±2-4% down to depths of...
450-500 km for P and S waves (Evans and Achauer, 1993; Bijwaard et al., 1998; Thurber and Ritsema, 2007; Sandoval et al., 2004; James et al., 2001; Shomali et al., 2006). Although in some cases mapped deep seismic velocity structures are in good accordance with surface geological features, in many cases the true causes of observed velocity variations remain ambiguous.

According to Dziewonski et al. (1977), temperature differences are significant in relation to the laterally varying seismic velocities resolved in global tomography models (between the depths of ~25-30 km and ~2800 km), since the models are well correlated with active mantle convection. Examples are subducting bodies underlying convergent plate margins (with prominent high velocity anomalies) and large scale low velocity regions where it is believed that hot material is rising through the mantle (plumes, Bijwaard and Spakman, 2000, Li et al., 2008, Garnero, 2004).

Observations show that on a regional scale the effect of temperature is more dominant for the S waves than P waves. According to Goes et al. (2000), thermal variations below the solidus cause the largest effects for seismic velocities in the upper mantle. For instance, velocity decreases of 0.5-2 % for P and 0.7-4.5 % for S are predicted for a 100 C° increase in temperature. However, at relatively low temperatures typical of the lithosphere, anelastic effects are not important and the thermal effect on seismic velocity is probably only about –0.75 % for Vp and –0.8% for Vs for each 100°C increase in temperature (Karato, 1993; Cammarano et al., 2003). Therefore, if the observed seismic velocity differences in the mantle below the Baltic Shield are the result of thermal effects, lateral temperature differences of a few hundred degrees Celsius would be required (Hieronymus et al., 2007).

The effect of chemical composition on seismic velocities within the upper mantle is more debated. Jordan (1979) and Sobolev et al. (1996) claim that significant changes in the uppermost mantle composition, which is constrained by information from the composition of xenoliths, result in relatively small variations in seismic velocities. Using high-resolution stacks of precursors to the seismic phase SS, Schmerr and Garnero (2007) investigated seismic discontinuities associated with mineralogical phase changes approximately 410 and 660 kilometers beneath South America and the surrounding oceans. Their detailed maps of phase boundary topography revealed deep 410- and 660-km discontinuities in the down-dip direction of subduction, inconsistent with purely isochemical olivine phase transformation in response to lowered temperatures. These authors explored mechanisms invoking chemical heterogeneity within the mantle transition zone to explain this feature. They observed, in some regions, multiple reflections from the discontinuities, consistent with partial melt near 410-km depth and/or additional phase changes near 660-km depth which may imply that the origin of upper mantle heterogeneity has both chemical and thermal contributions and is associated with deeply rooted tectonic processes. Griffin et al. (1999) reported, in addition to the compositional variations, temperature
differences from 400 to 500 °C as a source of lateral variation in seismic velocities of sub-continental lithosphere. If significant heterogeneity within the upper mantle was only thermal in origin, then the P and S velocity images should generally be expected to be very well spatially correlated. Thus for regions where a mismatch between P and S velocity anomalies is detected, then compositional differences should be considered in explaining the observed anomalies (Grand et al., 1997; Su and Dzienowski, 1997 and Becker and Boschi, 2002).

Few other methods which can directly provide information about present-day deep structure exist. One of these is the family of electromagnetic methods which use the natural induction of deep electrical currents to provide information about the electrical conductivity at depth. While such studies can be difficult, and electrical conductivity can depend on several different factors, electromagnetic studies can provide important constraints regarding the mantle properties and its geometry, complementing the seismological data (Jones, 1999; Ledo and Jones, 2005).

Over the last four decades, the number of seismic recording stations has increased dramatically, and computational techniques for the analysis of waveform data have become far more advanced. Seismologists now know that the idea of an Earth whose interior is composed of isotropic layers is not viable, even if these layers may be laterally heterogeneous. Observations of different types of waves (P, S, SKS, surface waves etc.) with different frequency content yield strong evidence that the Earth’s interior is significantly anisotropic (see Savage, 1999; Fouch and Rondenay, 2004; Plomerová et al., 2008). In other words, the velocity of seismic waves differs depending upon the direction of propagation. Love (1927) and Andersson (1961) define an isotropic elastic medium which can be described using only two parameters (λ and μ, the Lamé parameters). However a full description of an arbitrarily anisotropic medium requires 21 independent elastic constants. Anisotropy is a crucial concept in interpreting the deep structure of the Earth because it reflects the texture of rocks (Babuška and Cara, 1991). This is steered by the mechanical construction of (crystal orientations within) the material, which in turn can reflect the material’s deformation history. Thus both seismic velocity heterogeneity and anisotropy can illuminate past and ongoing geodynamic events and the tectonic evolution of the Earth’s interior.

The focus of this thesis is to investigate the characteristics of seismic body wave velocities (P, S, and SKS waves) within the upper mantle of the Earth (~ between 70 and 500 km) in isotropic and anisotropic conditions. Our region of interest is the Swedish part of the Baltic Shield, the exposed part of the Fennoscandian Shield forming the northwestern part of the East European Craton which hosts the oldest rocks known in Europe. As waveform data, we utilize high-quality teleseismic observations recorded by the Swedish National Seismic Network (SNSN). The outcomes are documented in four papers. In Paper I and II, we modeled isotropic 3-D velocity hetero-
geneities of compressional and shear waves within the upper mantle using non-linear teleseismic travel time tomography derived from the original ACH (ACH after Aki, Christoffersson, and Husebye) algorithm of Aki et al. (1977). The resulting P and S models indicate ±3-4 % velocity perturbations down to at least 470 km. Lateral velocity heterogeneities reveal several major features, many of which are also consistent when models based on P and S waves are compared. Conventional tomography studies use only SH wave arrival time, despite the fact that there may be different arrival times on the radial (SV) and transverse (SH) components. However the high quality of data from the SNSN allows robust identification of the arrival times of S waves with different polarizations (SH and SV). Paper II presents a comparison of isotropic velocity models derived independently from SV and SH data and discusses the discrepancies between these models. This study indicates that seismic anisotropy is significant within the mantle, to the degree that the tomographic images may be slightly distorted. In Paper III more conventional approaches including the directional dependency of P wave travel time residuals and shear wave birefringence (or SKS splitting) are used to estimate the type, level and orientation of mantle anisotropy. A joint inversion of SKS and P residual data allows estimation of a self-consistent 3-D anisotropic structure for the region. Finally, in Paper IV we investigate in more detail the possible effect of anisotropy on P velocity estimates produced under the assumption of isotropic structure.

The following chapters summarize some basic concepts and theory behind the applications presented in the attached papers, draw a depictive picture of study area with its geological settings, and briefly discuss the history of seismological studies of the deep lithosphere beneath the Baltic Shield.
2. The region of interest: The Baltic Shield

2.1 Major Tectonics of the Region

The lithosphere underlying the SNSN forms a part of the East European Craton (EEC), which has remained geologically stable since the mid-Proterozoic. The EEC is composed of the Fennoscandian, Sarmatian and Volgo-Uralian segments (Gorbatschev and Bogdanova, 1993, see inset in Figure 2.1) and it is bounded by Phanerozoic mobile belts. The Fennoscandian is exposed in the northern and central parts of the ECC (Fennoscandian Shield) and it comprises an Archean nucleus in the NE to which Proterozoic terranes have successively been accreted along the southern and western flanks. We concentrate on the Swedish part of the Fennoscandian Shield, which according to Gaál and Gorbatschev (1987) can be divided from the north to south into the Karelian Province, Svecofennian Domains, Trans-Scandinavian Igneous Belt, Southwest Scandinavian Domain and Caledonides (Figure 2.1).

The Karelian Province, consisting of Archean granitoid-gneiss complexes and supracrustal rocks (e.g., greenstones), is the only part that has ages in excess of 3 Ga. Available data indicate that the crust started to form more than 3.5 Ga ago (Gorbatschev and Bogdanova, 1993). From the Late Archean onwards, the Karelian Province formed a relatively stable nucleus against which the Paleoproterozoic Svecofennian orogen was moulded.

Post-Archean development started with rifting of the craton interior and the margin at 2.45 Ga (e.g. Park et al., 1984; Gaál and Gorbatschev, 1987). The rifting created several ‘microcontinents’ and it was followed at 2.1-1.93 Ga by formation of several arc systems on the margins of the craton in the SW, including the Bergslagen arc (Andersson et al., 2006; see Figure 2.1). Some of these microcontinents contain unexposed Archean basement. The main Early Svecofennian rock-forming episode (1.91-1.86 Ga) resulted in reworking of the early arc systems, partly including rift-related volcanism, and in the formation of juvenile crust in areas between the microcontinents. Lahtinen et al. (2005) describe the Palaeoproterozoic part of the Fennoscandian Shield as consisting of island arcs, microcontinental fragments and the formation of sedimentary basins between these structures (see Figure 2.2).
Final accretion of the Svecofennian composite collage to the craton occurred at about 1.86 Ga (Andersson et al., 2006 and references therein). Subsequent long-term north(east)ward subduction resulted in reworking of the newly formed crust during the Late Svecofennian (~1.85-1.75 Ga). The reworking caused the Svecofennian Complex to be intruded and covered along the SW margin by plutons and extrusive rocks of the NNW-SSE trending Trans-Scandinavian Igneous Belt (TIB in Figure 2.1). The TIB is more than 1500 km long and runs from south-east Sweden to north-central Norway. Of three age groups of volcanic and plutonic rocks, those of 1.81-1.77 Ga are by far the most voluminous and constitute the southernmost part of the belt (Korja et al., 2006).

Andersson et al. (2006) subdivided the major Svecofennian processes into: (1) The ‘proto-Svecofennian’, the earliest oceanic arc crust formation (2.1-1.91 Ga). (2) An early Svecofennian period of reworking and main juvenile
crust formation (~1.91-1.86 Ga) (3) The late Svecofennian period of underplating and reworking of the juvenile and cratonic crust (~1.85-1.75 Ga).

Based on lithological associations, Gaál and Gorbatschev (1987) divided the Svecofennian Domain into Northern, Central and Southern subprovinces (Figure 2.1). After the formation and cratonization of the Svecofennian Domain, growth of the Fennoscandian crust continued towards the present west, creating the Southwest Scandinavian Domain. This domain is separated from the rest of the Fennoscandian Shield by the TIB and the Protogine Zone (Figure 2.1), a major belt of shearing and faulting (Gaál and Gorbatschev, 1987). The shield thus becomes younger towards the west where Gothian evolution took place between 1.75 and 1.55 Ga and rocks in the westernmost part were reworked during the Sveconorwegian-Grenvillian orogeny at 1.15-0.9 Ga (Gorbatschev and Bogdanova, 1993). The Sveconorwegian orogenic belt, which forms the southernmost part of the region of interest, has been interpreted as a polyphase imbrication of terranes at the margin of Fennoscandia, as a result of a continent-continent collision, followed by relaxation between 0.96 and 0.90 Ga and by syn- and post-collision magmatism that increases towards the west (Bingen et al., 2008). Neoproterozoic development is related to lithosphere extension and formation of sedimentary basins along the margins of Fennoscandia. The Caledonide orogen is exposed in Norway and western Sweden, outside the region covered by the SNSN network.

Apart from the un-reworked relics of the Middle Archean crust in the Karelian Province, which so far do not permit geodynamic reconstruction, the character and spatial arrangement of most crustal units in Fennoscandia allow identification with well-known plate-tectonic patterns. The paper of Korja et al. (2006) based on an integrated study of geological and geophysical data provides a tectonic model of the Svecofennian orogen, which forms the major part of our region of interest. The authors characterized the Svecofennian orogen as a collage of microcontinents and island arcs formed during five partly overlapping orogenies and concluded that the orogenies evolved from accretionary to collisional stages in the Paleoproterozoic, which can be compared to ongoing orogenies elsewhere today. Some of these tectonic events are reflected in the large-scale fabrics of the deep lithosphere (Plomerová et al., 2001; 2006; Vecsey, 2007).
2.2 An overview of previous investigations of crust and upper mantle structure beneath the Fennoscandian Shield

To investigate the crustal structure of the Baltic Shield, several controlled source seismic investigations have been carried out (e.g., Guggisberg et al., 1991; Kinck et al., 1993; Grad and Luosto, 1987; Balling, 2000; Lund et al., 2001; Abramovitz et al., 2002; Juhlin et al., 2002; and Korja et al., 2005). FENNOLORA (a seismic refraction experiment, 1979) and BABEL (a wide-angle seismic experiment, 1993) are especially important in earlier studies of the large-scale structures. More recently, using the Receiver Functions (RF) technique, Olsson et al. (2008) estimated Moho depths well consistent with the results from previous seismic experiments where these existed and providing data in hitherto unsampled areas (Figure 2.3a). Results of these studies indicate that the Moho deepens towards the center of Fennoscandia,
although there is no apparent corresponding topographical signature. The length of the FENNOLORA profile of ~1800 km allowed investigations of both the crust and the upper mantle down to ~400 km (Guggisberg and Berthelsen, 1987; Perhuc and Thybo, 1996; Abramovitz et al., 2002). Studies from up to three decades ago provide rough images of velocity perturbations in the upper mantle (Hovland et al., 1981), lithosphere thickness estimates (Sacks et al., 1979; Calcagnile, 1982; Babuška et al., 1988), velocities in the crust and the uppermost mantle (Guggisberg and Berthelsen, 1987) and interpretations of regional tectonic evolution (e.g., Kinck et al., 1993; Thybo et al., 1993). Constraints from a small-scale passive experiment in south-central Sweden (Värmalnd’91, Plomerová et al., 1996; 2001) and two large passive seismic arrays, TOR and SVEKALAPKO (operated during 1996-1999; Gregersen et al., 2002; Hjelt et al., 1996) improved substantially our knowledge of the structure of the Fennoscandian Shield. Data from these experiments has been analyzed using a suite of methods including teleseismic and local tomography, shear wave splitting analysis, P-wave anisotropy, surface wave studies and receiver functions (e.g., Sandoval et al., 2004; Hyvönen et al., 2007; Plomerová et al., 2002; Cotte et al., 2002; Alinagni et al., 2003). Stacking of S to P converted phases in the upper mantle indicates clear signals of several horizons in the 50-200 km depth range which has been interpreted as a layered lithosphere with alternating high and low velocities (Olsson et al., 2007).

Recent high-resolution body-wave tomography studies covered also regions adjacent to the Baltic Shield and revealed its sharp termination to the south-west (the TOR experiment; Arlitt et al., 1999; Shomali et al., 2002). The P-velocity tomography based on the SVEKALAPKO data (Sandoval et al., 2004) located the cratonic root of the oldest part of the Shield (Karelia), but did not detect the Proterozoic-Archean boundary in the mantle beneath southern Finland (between ~58.5° and 64° N). This boundary has been modelled through changes in apparent seismic anisotropy reflecting variations in mantle lithosphere fabrics (Plomerová et al., 2006, Vecsey et al., 2007—see Figure 2.3b). Olsson et al. (2007) detected the seismic anisotropy by splitting evaluations from P410s converted phases. They claim that resulting delay times and fast polarization directions could be due to the seismic anisotropy located within the upper 410 km below Sweden.

In the Värmland area (south-central Sweden), different orientations of dipping anisotropic structures have been found within subcrustal lithosphere blocks on either side of the Precambrian Protogine zone, although isotropic tomography did not show any differences in isotropic P velocity perturbations between these two domains (Plomerová et al., 2001). Beneath southern Finland velocity-depth distributions from Rayleigh waves exhibit a regional grouping (Bruneton et al., 2004) similar to findings from body waves beneath southern Finland (Plomerová et al., 2006). Azimuthal and polarization anisotropy analysis of surface waves distinguished high velocities oriented to
the NE in the upper mantle (Pedersen et al., 2006) which is compatible with findings from body waves in the Archean part of the Shield (Vecsey et al., 2007). A vertical cross section of a preferred model after Vecsey et al. (2007) is shown in Figure 2.3b.

Figure 2.3a The map of Moho depth variation in the Baltic Shield obtained from analysis of converted waves (Olsson et al., 2008).
Figure 2.3b Model for mantle lithosphere obtained from body wave analyses along the SW-NE profile across the Proterozoic and Archean parts of the central Fennoscandia. Hexagonal aggregates with the high-velocity a axis dipping in the NE azimuth (in the Archean), or with the high-velocity (a,c) foliation plane dipping to the SE (in the Proterozoic) represent anisotropic structure of the mantle lithosphere and shown in velocity distributions presented in polar projections of the lower hemisphere (Vecsey et al., 2007). The model is also supported by alternating Proterozoic-Archean-Proterozoic ages of xenolith samples (Peltonen and Brügmann, 2006).
3. Data

3.1 The Swedish National Seismic Network (SNSN)

The three component waveform data utilized in this work comes from the Swedish National Seismic Network (SNSN). Almost continuous seismological recordings from Sweden exist for over 100 years, but the number of observation points was limited. Since 1998, Uppsala University has installed a large number of new stations to construct the new SNSN. The network is currently one of the most modern broadband seismic networks in the world. It now operates 60 3-component seismic stations (Figure 3.1a), which are mostly equipped with Guralp CMG-3TD sensors with flat velocity response in the period range from 0.02 to 30 s. (Some stations are instead configured to 120s). As the network is still under construction, data from all of 60 current stations was not available for my studies, and mostly I have used data from the 45 stations then available. The array extends over an area of about 450 km wide and 1450 km long, with station spacing differing in different provinces. The highest concentration of stations is along the Baltic coast. A comprehensive description of the network can be found in Bödvarsson (1999), Olsson (2007), and Paper I and Paper II of this thesis.

3.2 Event distribution and preparation of the data before processing

In both Paper I and Paper II we have used the same 52 high quality teleseismic earthquakes with magnitudes larger than 5.5 and epicentral distances between 30° and 90°. Locations were based on the event catalog reported by the USGS corrected as reported by Engdahl et al. (1998). Because of the rather long and narrow shape of the SNSN array, and because the major axis of the SNSN is believed to be approximately perpendicular to the major tectonic features in the area, during the event selection primarily teleseismic earthquakes whose great circle paths were roughly in line with the long axis of the SNSN were considered. Limiting the data set in this manner can alleviate possible complications related to significant three-dimensionality. Before starting P and S-phase data picking to determine the relative-arrival times, all seismograms were filtered to simulate a long period WWSSN station (the World Wide Standardized Seismographic Network) with a domi-
nant period of 1 and 10 sec (Oliver and Murphy, 1971) for P and S waves respectively. We rejected data with a very low signal-to-noise ratio. The relative phase picking procedure includes overlaying the waveform of a clear recording from one reference station on other detectable peaks or troughs of data from other stations (see Shomali et al., 2002; Paper I). This is necessary in order to avoid cycle skipping (Evans and Achauer, 1993) and is considered to provide more accurate time picks than e.g. picking first (P or S-wave) breaks. An example of the data picking can be seen in Figure 3.1b. The geological conditions in Sweden, with sedimentary cover often thin or completely absent, provide generally very high signal-to-noise ratio at all stations. However the signal-to-noise ratio of the data varies depending upon the source position. Available data means that generally the minimum signal-to-noise ratio for the events which approach the array from the north-east is four times larger than for those approaching from the south-west. During the data preparation and phase picking process, various different software packages including Seismic Handler (SHM, see Stammler, 1993) and SAC2000 were used (Goldstein et al., 2001).

In Paper III, the anisotropy within the upper mantle was evaluated by using two different data sets. The first data set consists of about 4200 arrival times of P waves of 136 teleseismic earthquakes at epicentral distances from 30° to 90° with magnitudes between about 5.5 and 7.6. The P wave data set is an extended version of that used in Paper I and II, with more events from more geographic locations. The same picking procedure as in Paper I was applied manually to the first amplitude peaks correlated for all P-wave trains of an event. For subsequent calculations each pick was assigned a weight describing its quality. P relative time residuals estimated for the Paper III study have been also used in Paper IV.

The second data set in Paper III consists of high-quality broad-band shear waveforms. Altogether we have inspected recordings of 285 events, which occurred between 2002 and 2008. However, only 25 of them possess sufficiently large signal/noise (S/N) ratio to be considered as suitable for the SKS splitting analysis. The minimum suitable magnitude was about 6.5. We have used only SKS phases from epicentral distances between 65° and 140°, which are well separated from other phases (e.g. S, P to S, or ScS). The frequency content of shear waveforms for various events differs. Sometimes even spectra for the same event but recorded at different stations are significantly different. To examine the frequency content of each signal, we first calculate a Morlet wavelet spectrum (Daubechies, 1992) that images the frequency as a function of time and shows frequency ranges of useful signal. Wavelet analysis of the SKS waveforms at the SNSN indicated useful signals in three overlapping period ranges: 1-10 s, 2-20 s and 4-20 s. Secondly, we applied 3rd order band-pass Butterworth filtering for the three frequency ranges. A detailed application of such filtering for SKS splitting measurements can be found in Vecsey et al. (2008). Appropriate filtering of individ-
ual traces improves the S/N ratio and the evaluation of the splitting. Figure 3.1c presents an application of such processing with an example waveform at station OCT before and after filtering.

a)

b)

![Map and diagram with foci and stations]
Figure 3.1. An example of an SH waveform recorded at station ARN and at a reference station BAC. Inter-station offset between BAC and ARN is ~180 km. b) Distribution of 136 teleseismic earthquakes recorded by the SNSN (left) during 2002-2008 which were used either in P-wave anisotropy analysis (red circles) and/or in the shear-wave splitting measurement (25 earthquakes, stars). Plate boundaries are after Bird, (2003). c) One example of an SKS phase recorded at the station OST from an earthquake of 20030526-2313 with magnitude 6.9 Mw and epicentral distance 92.3º (Top). The bottom figure shows an example of wavelet time-frequency analysis used for determining the width of band-pass filtering.
4. The physics of seismic waves

As my work investigates the effect of Earth structure on seismic waves, it is necessary to understand the theoretical background regarding seismic wave propagation. Here, I summarize some of the most important concepts, largely following Scales, (1997) and Stein and Wysession, (2003).

4.1 Seismic wave equation

When an earthquake happens, the seismic energy radiated from the source is transported by seismic waves which cause an elastic displacement within the medium. In a whole-space the solution of the equation of motion results in two types of seismic (elastic) waves, compressional and shear waves, which propagate with different velocities depending in different ways in the elastic properties of the material.

The equation of motion for a homogeneous isotropic elastic medium can be written in terms of the displacements as a function of time and position,

\[(\lambda + 2\mu)\nabla(\nabla \cdot u(x,t)) - \mu \nabla \times (\nabla \times u(x,t)) = \rho \frac{\partial^2 u(x,t)}{\partial t^2}\]  (4.1)

where \(\lambda\) and \(\mu\) are Lame’s constants and represent the elastic material, \(u(x,t)\) is the displacement as a function of spatial position \((x)\) and time \((t)\), \(\rho\) is the density of the medium, and \(\nabla\) is the Nabla operator given by,

\[\nabla = e_x \cdot \frac{\partial}{\partial x} + e_y \cdot \frac{\partial}{\partial y} + e_z \cdot \frac{\partial}{\partial z}\]  (4.2)

where \(e_x\), \(e_y\), and \(e_z\) are the unit vectors.

The displacement term, \(u\) of equation 4.1 can be expressed in terms of two other functions, \(\phi\) and \(\gamma\) known as potentials,

\[u(x,t) = \nabla \phi(x,t) + \nabla \times \gamma(x,t)\]  (4.3)
Equation 4.3 is a decomposition of the displacement into the gradient of the scalar, \( \phi(x,t) \) and the curl of the vector, \( \gamma(x,t) \) potentials. These correspond to compressional (for P waves) and transverse (for S waves) movement, respectively. Using the identities

\[
\nabla \times \phi(x,t) = 0 \quad \text{and} \quad \nabla \cdot \gamma(x,t) = 0
\]

we can say that compressional waves do not result in shear motion while shear/transversal waves do not cause any volume change. According to Lamé’s theorem the seismic wave propagation can be written in terms of the scalar and vector potentials as follows,

\[
\alpha^2 \nabla^2 \phi(x,t) = \frac{\partial^2 \phi(x,t)}{\partial t^2}
\]

with the velocity \( \alpha = \sqrt{\lambda + 2\mu}/\rho \) for P waves,

\[
\beta^2 \nabla^2 \gamma(x,t) = \frac{\partial^2 \gamma(x,t)}{\partial t^2}
\]

with the velocity \( \beta = \sqrt{\mu/\rho} \) for S waves.

### 4.2 Basic concepts of ray theory and seismic travel times

Generally speaking, the rules of geometrical optics work successfully for seismic rays. Ray theory constitutes a basis for many of the applications used in investigations of crustal and mantle velocity structures, and for the determination of the spatial distribution (locations) of earthquakes.

Geometric ray theory deals with the behavior of seismic waves by considering ray paths associated with them. Although it does not fully correspond to all important aspects of the wave propagation, it is a powerful and often more than adequate approximation.

For plane waves in homogeneous isotropic perfectly elastic medium, the rays, along which the elastic energy of the wave flows, are perpendicular to the wave fronts (Figure 4.1). A non-linear partial differential equation (PDE), the eikonal equation describes the propagation of a wave front through a medium (Scales, 1997):

\[
(\nabla T)^2 = \left( \frac{1}{c(r)} \right)^2
\]
where $T$ represents the travel time of the wave front and $c(r)$ is the local wave velocity at position $r$. The eikonal equation for the waves works fine for the cases in which the elastic properties of the given medium change slowly.

![Figure 4.1](image)

Figure 4.1 An illustrative figure to show the distortion of the expanding wave fronts as a result of the velocity structure increasing with depth. Note that "ray path", mentioned in many places within the current chapter, the normal to the wave fronts, is bent (Figure is courtesy of Edward Garnero).

Seismic ray theory is mostly used in computing travel times i.e. the time for a wave to travel from one point to another one. The travel time is the length of the given ray path divided by the velocity, allowing for possible variation of velocity along the ray path. In a homogeneous medium, velocity is constant and ray paths are straight. Where velocity changes the ray will generally alter direction. A simple case is a laterally homogeneous multi-layered Earth i.e. a structure with several homogeneous layers whose seismic velocity characteristics change only at the horizontal boundaries. Except for rays traveling exactly vertically or horizontally, ray direction will change stepwise at each boundary. However complex a ray path is, total travel time can always be calculated by summing up the travel times for each portion of the ray path,

$$t_{\text{total}} = \int_{S_{(c)}} \frac{1}{c(r)} \, ds$$  \hspace{1cm} (4.8)

Equation 4.8 is a generally valid equation giving the total time spent along the ray path $S$ for any medium (not just layered media). If the ray path (and velocity structure) is known, the travel time can easily be calculated by numerical integration. However, not only travel time but also the ray path itself is a function of the velocity structure, so to calculate travel time we must first find the ray path.
According to Fermat’s principle, the ray path between two points located on the opposite sides of a plane interface is that for which the travel time is a minimum (Stein and Wysession, 2003). The direction of the travel paths of rays in each layer is controlled by Snell’s law given by,

\[
\frac{\sin(i_i)}{c_i} = \text{constant}
\]  

(4.9)

where \( c_i \) represents the velocity of the \( i \)th layer and \( i_i \) is the incidence angle or the angle of the ray to the normal at the \( i \)th interface. Physically, the constant corresponds to the apparent horizontal velocity of propagation of the wave i.e. the velocity estimated from the arrival time of the signal at two points at the same depth divided by the distance between these points. The reciprocal of this is called the horizontal slowness, or often just slowness. According to equation 4.8, the ratio of the sine of the incidence angle of each wave to the corresponding velocity is constant (Stein and Wysession, 2003).

In normal ray theory, we implicitly assume that the wave has infinitely high frequency. In this case, all boundaries are essentially “plane” to the incoming ray, and Snell’s law is valid for all boundaries, given that the local boundary orientation and appropriate angle of incidence are used. Ray tracing is also easily applied even where velocity changes gradually with position. This can be done by approximating the medium by layers (or cells) of constant velocity and thus with straight ray path, and applying Snell’s law at each boundary. In principle, any desired accuracy of calculation can be achieved by making the layers (or cells) sufficiently small.

4.3. Seismic waves in a spherical earth

In the preceding section we touched on the basics of the ray theory for a given laterally homogeneous multilayered Earth structure. Such a model can be valid if the ray paths between the source and receiver are short enough, i.e. if we can ignore the curvature of the Earth. For the analyses of seismic travel times to investigate deeper parts of the Earth such as the mantle we deal with greater distances between the source and receiver. Hence we must use seismic ray theory adapted to the case of a spherical earth.

4.3.1. Ray paths

Starting from the rules of seismic ray theory for an Earth structure composed of flat layers as discussed above, we can easily adapt to spherical geometry and spherical symmetry (velocity only as a function of depth or radius). An illustration of ray geometry is given in Figure 4.2 which depicts the portion of
a seismic ray path between two points located at radial distances \( r_1 \) and \( r_2 \) from the center of the Earth. Assuming \( c_1 \) and \( c_2 \) are the velocities above and below the point located at \( r_1 \), and \( i_1, i'_1, \) and \( i_2 \) are the angles of incidence as shown on the figure, then Snell’s law can be written (Stein and Wysession, 2003):

\[
\frac{r_1 \sin i_1}{c_1} = \frac{r_1 \sin i'_1}{c_2} = \frac{r_2 \sin i_2}{c_2} = p
\]  

(4.10)

A generalized form of equation 4.10 is

\[
\frac{r \sin i}{c} = p
\]  

(4.11)

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{figure4_2.png}
\caption{Parameters used in Snell’s law for a spherical earth (Figure is courtesy of Stein and Wysession, 2003).}
\end{figure}

where \( p \) represents the ray parameter. In a layered spherical Earth, \( p \) is constant for a given ray. The only changing factors in this case are the velocity of the medium and radial distance from the center to the given point. Compared to Snell’s law for a flat layer case (equation 4.9), the only difference is the addition of the parameter \( r \) as a multiplier in equation 4.11. In the spherical case of Snell’s law, the factor \( r \) corrects for the change along the path of the orientation of the normal to the interface, which is the radius. For cases where \( r \) varies slowly, then its variation can be ignored. Thus using a flat Earth approximation is suitable for near-surface refraction and reflection studies (Stein and Wysession, 2003).
4.3.2. Travel times

The calculation of the total travel time for a ray traveling from one point to another one in a spherical earth (see Figure 4.3) is again simply integration along the ray path as previously defined in equation 4.8. In spherical geometry, however, the ray path can be described in terms of the polar coordinates \((r, \theta)\).

\[ T(p) = \int ds / c = 2 \int_{r_p}^{r_s} \frac{\zeta^2 dr}{r(\zeta^2 - p^2)^{1/2}} \]  

where \(p\) is ray parameter, \(\zeta\) is \((r/v)\), and \(c\) indicates the velocity. The integrals in equation 4.12 are from the surface to the deepest point at P. Hence doubling is necessary to account for the upward paths. A derivation of the relationship given in equation 4.12 can be found in Stein and Wysession (2003).
5. Teleseismic travel time tomography

5.1 General definition of the system as an inversion problem

As emphasized in previous sections, the present thesis work aims at elucidating internal velocity structure of the upper mantle part of the Earth by applying some mathematical techniques (i.e. inversion) on seismic observations. Seismic tomography is a well-known method for producing approximate estimates of the 3-D distribution of the physical properties which affect seismic-wave propagation: elastic, anelastic, anisotropic parameters and anisotropy. Models from tomography experiments are important since they provide constraints on the temperature, subsurface-lithology, fracturing, fluid content etc. (Thurber and Ritsema, 2007). The first applications of seismic tomography were initiated in the mid-1970s by Keiiti Aki and his co-workers using seismic body wave observations on a regional-teleseismic or local scale. Starting in 1975 Dziewonski and his colleagues developed a tomographic approach to resolve the seismic velocity characteristics of body-waves on a global scale. During preparation of this section regarding the principles of teleseismic tomography algorithms and inversion theory I have benefited mainly from Nolet (1987), Menke (1989), Thurber and Ritsema (2007).

The current work presents an application of teleseismic tomography to the relative residuals of the arrival times of body-waves. Due to the source-receiver geometry, teleseismic tomography can be considered as a restricted array procedure because the receiver array does not cover all distances from the source (Evans and Achauer, 1993). As with other types of travel time tomography (i.e. local or global tomography), the basic principle of teleseismic tomography is to estimate the spatial-extension and magnitude of velocity anomalies by back-projection of (relative) arrival-time residuals (Weiland et al. 1995). The system is described by the following equation (c.f. equation 4.8),

$$ T = \int_{S} \frac{dS}{V(r)} $$

(5.1)

where T and S represent travel-time and ray path respectively. These depend on the velocity of the medium, $V(r)$. As it is obvious from equation 5.1 the
system is non-linear since an unknown variable $V(r)$ explicitly affects the ray-path, $S$. However the level of non-linearity in the tomographic inversion depends upon the geometry and velocity structure. In teleseismic tomography a linearized approach such as the ACH method (Aki et al. 1977) is commonly adopted, though often applied stepwise to allow updating of ray paths.

According to Evans and Achauer (1993), ACH is considered to be a robust method since ray turning points are excluded from the modelled volume. Thanks to the availability of spherically symmetric reference Earth models (e.g. IASPEI91, PREM etc.), $V(r)$ can be approximated with high accuracy leading to deviations rarely exceed about 2 seconds in total travel times of 1000 sec or more (Nolet, 1987). Therefore symmetric Earth models (see Figure 5.1) can be improved by small deviations from this symmetry within the upper mantle.

![Figure 5.1](image)

*Figure 5.1* 1-D IASP91 model (Kennett and Engdahl, 1991; solid lines) is given with a comparison of the classic Jeffreys-Bullen earth model (Jeffreys and Bullen, 1940). Figure is courtesy of Stein and Wysession (2003).

Using an initial model we can estimate the total travel times for a given ray in a similar way to equation 5.1,

$$ T^0 = \int_{s^0}^{S} \frac{dS}{V_0} $$

(5.2)
where \( V_0 \) is the starting model and thus \( S_0 \) is the ray path based on this starting model. The input of the tomography procedure delay times then can be defined,

\[
\Delta t = T - T^0 = \int_s \frac{dS}{V(r)} - \int_{S_0} \frac{dS}{V_0(r)} = \int_{S_0} dS \left( \frac{1}{V(r)} - \frac{1}{V_0(r)} \right)
\]

(5.3)
equation 5.3 can be approximated as follows,

\[
\Delta t = -\int_{S_0} \left( \frac{\Delta V(r)}{V_0(r)^2} \right) dS
\]

(5.4)
where \( \Delta V(r) \) is simply \( V - V_0 \). By employing the deviations from a reference Earth model, the problem can formulated as a linear scheme as given in equation 5.4. In other words a linear relationship exists now between the seismic data (i.e. travel time anomalies of body waves, \( \Delta t \)) and expected perturbations from the wave velocity in the reference model. Then equation 5.4 can be first discretized and then written in a matrix form,

\[
\Delta t_i = \sum_j G_{ij} u_j
\]

(5.5)
where \( G_{ij} \) represents the distance the \( i^{th} \) ray travels in the \( j^{th} \) block, and \( u_j \) is the slowness perturbation in the block (see Figure 5.2). In the inverse problem of travel time tomography, estimates of the slowness perturbation via observed travel time perturbations can be described in a matrix form,

\[
d = Gm
\]

(5.6)
where \( d \) is the data vector of length \( N \) and \( m \) is the vector of length \( M \) for model parameters. \( G \) is known as the Frechet matrix and represents the matrix of the partial derivatives of the data with respect to the model parameters and is given by,

\[
G = \frac{\partial d_i}{\partial m_j}
\]

(5.7)
From equation 5.7 it is obvious that \( G \) describes the distance the \( i^{th} \) ray travels in the \( j^{th} \) block which is the partial derivative of the ray’s travel time with respect to the slowness in the block (Stein and Wyssession, 2003). The \( G \) matrix defines the relationship between data and model perturbations.
Once the system is linearized, various different inversion methodologies can be employed (Menke, 1989). In the present work one of the methods used to solve the linearized system is damped least squares. Generally speaking in seismic tomography we deal with more data than the unknowns ($N >> M$). However, because of data errors and possibly inadequacies in the form of the model, equation 5.6 becomes inconsistent and we can not obtain an exact solution. Therefore we minimize the misfit between the empirical and model data to find the optimum estimate (Thurber and Ritsema, 2007),

$$\begin{align*}
(Gm - d)^T (Gm - d)
\end{align*}$$

(5.8)

where equation 5.8 is actually a least squares problem which is defined in the Euclidean norm. If the system is solved based on damped least squares then we minimize,

$$\begin{align*}
(Gm - d)^T (Gm - d) + \varepsilon^2 m^T m
\end{align*}$$

(5.9)

Here the term $\varepsilon^2 m^T m$ related to the damping is responsible for penalizing models $m$ with a large norm (Thurber and Ritsema, 2007). Minimization of equation 5.9 with respect to model parameters is given by equation 5.10,

$$\begin{align*}
m^{est} = [G^T G + \varepsilon^2 I]^{-1} G^T d
\end{align*}$$

(5.10)

If the weights $W^h$ and $W^d$, which are related to the model and data respectively, are added into the system then solution for $m$ becomes

$$\begin{align*}
m^{est} = [G^T W^d G + \varepsilon^2 W^m]^{-1} G^T W^d d
\end{align*}$$

(5.11)
where $m^{\text{est}}$ is the model vector that includes velocity perturbations. In this equation, $W^d$ stands for data error, determined during the data picking. In practice, $W^d$ also executes the removal of the arrival means and the source effects (Weiland et al. 1995). $W^m$ is a smoothing operator. For discrete model parameters one can use the difference between physically adjacent model parameters as an approximation of solution roughness, or alternatively the second derivative of travel-time with respect to model parameters. $W^m$ then stands for solution roughness, $\epsilon^2$ is the damping factor which ensures the stability of the inversion. $W^b$ and $\epsilon^2$ are defined a priori to the inversion.

A more generalized form of solution can be written as

$$m^{\text{est}} = G^{-g}d$$

where $G^{-g}$ is defined as the ‘generalized inverse’. Using singular value decomposition (SVD) of $G=U\Lambda V^T$, $G^{-g}$ can be written as

$$G^{-g} = VFA^{-t}U^T$$

Here $A^{-t}$ is described as the pseudoinverse of $A$ (with diagonal elements $1/\lambda_i$) and $F$ is the filter factors with elements defined by (Aster et al., 2005; Thurber and Ritsema, 2007):

$$f_i^2 = \frac{\lambda_i^2}{(\lambda_i^2 + \epsilon^2)}$$

For an assessment of the quality of the solution, two measures are important. These are the model resolution matrix indicating the ‘model blurriness’ and the model covariance matrix representing the uncertainty and covariation among the model parameters (see Aster et al., 2005; Thurber and Ritsema, 2007). One of those measures, the model resolution matrix $R_m$ can be found by inserting $d=Gm$ into equation 5.13,

$$m^{\text{est}} = G^{-g}Gm = R_mm$$

while the model covariance matrix $C_m$ is defined as follows (Thurber and Ritsema, 2007)

$$C_m = G^{-g}C_d(G^{-g})^T$$
where \( C_d \) is the data covariance matrix and has a diagonal form in equation 5.16. Considering a damped least square solution for solving an inverse problem, then \( R_m \) and \( C_m \) can be defined by,

\[
R_m = VFV^T \quad \quad \quad (5.17)
\]

\[
C_m = VF\Lambda^{-p}U^T C_d U (\Lambda^{-i})^T F^T V^T \quad \quad \quad (5.18)
\]

It is obvious from equations 5.14, 5.17, and 5.18 that there is an inevitable trade-off in which the estimated model resolution gets closer to the identity matrix when small damping parameters are employed. In other words model resolutions can be improved by choosing small damping parameters. However, one disadvantage of small damping values is an increase in the model uncertainty i.e., increasing size of (the diagonal elements of) \( C_m \). Thus during the selection of an optimum damping value, a trade-off analysis can be applied (Eberhart-Phillips, 1986). In Paper I and Paper II, we use an L curve analysis (Aster et al., 2005) whereby the variance of the model parameters versus data misfit plot is used in choosing a suitable damping value (Figure 5.3).

Using truncation during the singular value decomposition (SVD) of the generalized inverse is another approach to the issue of damping. When using truncated SVD (known as TSVD), an approximate damped generalized inverse matrix can be obtained by using the \( p \) largest singular values of \( G \),

\[
G^{-g} = V_p F_p \Lambda^{-i}_p U_p^T \quad \quad \quad (5.19)
\]

A comprehensive discussion on the least-squares, pseudoinverse and TSVD solutions can be found in Aster et al. (2005).
5.2. Data

The data for the body wave travel time tomography experiments are arrival times of teleseismic events whose epicentral distances to each station of our 2-D array are usually greater than 20°. In the work presented, we selected teleseismic earthquakes with magnitudes larger than 5.5 and epicentral distances between 30° and 90° based on the event catalog reported by the USGS. It is helpful if the source location estimates are accurate because more accurate hypocenter estimates mean less error in the travel time residuals data which is used as input to the inversion procedure. Since the P waveforms may vary while propagating across the large array, it is crucial to be able to reliably select a first break, or a precise peak or trough point for each P-waveform. This can usually be achieved by taking one reference seismogram with a good signal-to-noise ratio and overlaying this on the others in order to pick a corresponding peak or trough relative to the reference station (Figure 5.4). This procedure is useful in avoiding cycle skipping (Evans and Achauer, 1993).
and reducing the problems related to the wave propagation. The normally very good correlation of first arrival wavetrains over the array, even over great distances, suggests that the risk of bias from picking peaks (as opposed to first breaks) is small when compared to the poorer precision in picking first breaks in data in the presence of noise. Basically, our final relative arrival-time residuals are obtained by first subtracting observed arrival time from the calculated theoretical arrival time at each station, and then subtracting the associated mean for each event from all arrival-time residuals,

\[ \Delta t_{rel} = \Delta t - \Delta t_{ref} \]  

Residuals defined by equations 5.3 and 5.4 may carry some undesired effects originating in deep mantle paths or around the source, as well as mislocation effects which might be up to several seconds. Such effects usually occur on all rays belonging to the same event. Removing the averaged arrival-time residuals for each event is useful in suppressing undesirable propagation path effects originated from the lower mantle (Dueker et al., 1993; Evans and Achauer, 1993). This, indeed, reflects the main assumption which underlies the ACH method (Aki et al., 1977), that any travel-time residuals originating from velocity deviations from spherical symmetry outside the target region are considered to be almost constant across the seismic array (Masson and Trampert, 1997). In other words, only the upper legs of incoming rays are inverted. Therefore there is a potential risk of leakage of deeper mantle anomalies into the target volume (Masson and Trampert, 1997).

However the results of recent studies (e.g. Shomali et al., 2002; Sandoval et al., 2004) and also the current study imply that these effects are minor in terms of revealing the major transitions and blocks of the upper mantle. The relative arrival time-residuals do not yield the absolute velocity for the target volume since the whole ray path is not inverted. In other words, the output model represents essentially the velocity perturbations from an unknown layered Earth model. Although during the forward modeling the theoretical arrival times are calculated according to a known background model (starting model), the final velocity perturbation obtained from the inversion should not normally be considered to be relative to the known background model, due to the nature of the relative arrival-time residuals. In principle, difficulties may arise if the starting model is inadequate, but there are no indications of significant such problems in our analyses.
Figure 5.4 a) An example of the data record section for the event of 20031112-0826 which occurred at 33.189°N and 137.044°E with a magnitude of 6.4 Mw and an epicentral distance of 74° (as great circle arc length) to the centre of the network. The P arrival-times are computed based on the IASP91 travel-time model. b) An example of the relative picking process on two selected waveforms. Top: the waveform obtained at the reference station NIK (blue solid line). Bottom: Overlaid waveform of the reference station (dotted blue line) on the readable peak of another station, SJU (red solid line).
5.3 Model Parameterization

One of the crucial steps for seismic tomography surveys is to represent the Earth’s seismic velocity structure. This can be done in different ways and each of them aims at providing a good approximation to the real 3-D structure of the target volume which will be inverted. Choosing an appropriate model approximation is, however, not always an easy task due to the Earth’s heterogeneity on a large range of spatial scales, which includes complicated structures such as faults, discontinuities, layering, intrusions, zones of elevated temperature and partial melt etc. Seismic anisotropy is another complexity which is discussed more below. Imaging the spatial scale of heterogeneity is limited mainly by the density of wave-sampling, with a lower bound proportional to the minimum wavelength of recorded seismic wave energy (Thurber and Ritsema, 2007).

There is no a unique way of representing heterogeneities. Following Aki et al. (1977) the use of constant velocity uniform blocks has become a common approach for regional and global scale tomography. Over the years there have been several efforts to improve upon the original ACH algorithm. Evans and Achauer (1993), for instance, introduced varying block thickness.
with depth in their model parameterization. \textit{VanDecar} et al. (1995) used nodes with spline interpolation. Using major seismic velocity discontinuities along with seismic velocity heterogeneity allows the inclusion of converted phases in modeling the position of discontinuities (\textit{Zhao} et al., 1992). In a similar fashion, \textit{Zhao} (2004) combined a larger set of seismic discontinuity data (especially the Moho, the 410 and 660 km discontinuities) for a global tomography experiment.

One of the critical issues affecting model parameterization and thus the resolution of inverted models is the uneven data coverage. There are several reasons for this including uneven distribution of events used as sources, non-uniform station distribution and ray bending (\textit{Thurber and Ritsema}, 2007). In the case of uneven data coverage then there is always risk for some cells/nodes not to be sampled while some others are very well sampled. The mismatch between the data distribution and the cells/nodes can be compensated for by employing irregular cell and adaptive mesh approaches. The purpose is in general to provide an adaptive way of distributing the inversion cells or grid to be able to match the resolving power of the data and to better condition the inversion problem (\textit{Thurber and Ritsema}, 2007). For the curious reader we strongly recommend the recent review of \textit{Sambridge and Rawlinson} (2005) regarding the usage of irregular cell and adaptive mesh methods for seismic tomography. We should also recall that the various forms of smoothing applied in inversion (eigenvector truncation, regularization, direct constraints on model smoothness etc) reduce the true number of degrees of freedom in the model and thus have a similar effect to using a sparser grid.

It is obvious that model parameterization is a long and detailed story in seismic tomography studies and there are numerous approaches that we do not have enough time and space here to mention. During the discretization of the target volume to represent a 3-D Earth structure, one can naturally ask what might be the optimal form of model parameterization. One approach which may have some advantages is to employ a two-stage parameterization in which a traditional parameterization with a regular grid is used in the initial phase of modeling. Resulting model parameters after the first inversion can be employed as a second starting model within an inversion scheme of adaptive irregular model parameterization. Such an approach also allows a comparison between resulting models after first and second inversions, revealing the effects of the different gridding. When using regular grid approaches, a reasonable way to decide optimum model parameterization in the vertical and horizontal dimensions is to conduct resolution tests in which the ability to recover a hypothetical pattern of heterogeneity is tested. Paper I, and Paper II have benefited from the latter option in our model parameterization story. Based on the station spacing and a number of numerical tests, a lateral grid size of 75x75 km was found to be suitable for most of the model. The model and inversion are fully three-dimensional, but the extended form of the array, which is believed to lie with its major axis largely perpendicular
to most major tectonic features in the area, means that most of our results are presented as simple cross-sections. In the northern part of the profile where the station spacing is sparser the lateral grid size was adjusted to be 100x100 km. The gridding imposes a spatial smoothing effect since 75 km is larger than the smallest resolvable structure defined by velocity and the dominant frequency of the data. However, the primary target for the current study is large scale structures in the upper mantle and this block size is appropriate in relation to the structures of interest. As depicted in Figure 5.5, some (velocity) nodes in the model were fixed since they are not crossed by (many) seismic rays, while others were allowed to vary in the inversion since they were crossed by many upcoming rays. Our final model, which we consider to be that which best resolves structures, was chosen to have 19 layers of varying thickness (30 to 70 km) from the surface down to a depth of 470 km (Figure 5.5). Many tests were conducted, including using larger horizontal grid spacing such as 100x100 km, in order to investigate the possible effects of grid spacing. The resulting models were in good agreement and show close similarities in terms of resolving major blocks and transitions. The model parameterization was mostly 75x75km in the horizontal plane (partly 100x100) with a vertically changing block size (30 to 70 km), leading to 1287 floating nodes to be inverted out of 3553 associated model parameters in total. We kept this setting to be same for both P- and S-wave inversions. The size of the SNSN array means that the Earth’s curvature must also be taken into account. This was done by correcting the background velocities based on an Earth Flattening Transformation (Shearer, 1999). The velocity interpolation between nodes is linear (Steck and Prothero, 1991). A 3-D ray-tracing calculation was used to estimate the positions of rays inside the 3-D target volume. We used Steck and Weiland’s simplex-based ray tracer (Steck and Prothero, 1991; Weiland et al., 1995).

5.4 3-D ray tracing

The prediction of travel times for a given 1-D reference Earth model can be achieved using a ray-tracer. We use that which was originally developed by Steck and Prothero (1991). The ray tracing algorithm is based on Fermat’s principle which states that rays follow minimum-time paths. Using an initial 1-D reference Earth model ray paths for minimum travel times are computed starting from a planar wave front at any given depth to a certain point at the surface, usually where a seismic station is located. For obtaining the minimum travel times, the initial straight ray is systematically distorted until the minimum travel time ray path is found. Within an iterative process, 3-D ray paths are computed by successively adding higher harmonics to the initial ray path in the both horizontal and vertical ray directions (see Steck and Prothero, 1991; Nelder and Mead, 1985; Jordan, 2003 for more details).
Figure 5.5 A simple illustration of ray tracer algorithm in this study. The figure is adapted from Steck and Prothero (1991).

5.5 Resolution assessment

Up to now we have dealt with some basic principles of the inversion theory behind the seismic tomography and some major procedures of tomographic modeling. One remaining question after the inversion of the data regards the reliability of the models. In other words, the question of how well derived models represent the true Earth. This issue is not simple and definite answers are rarely possible. Especially in large-scale tomography investigations, where there is no “ground truth” (e.g. borehole data) to confirm and calibrate results, an analysis of the spatial resolution of the data is crucial in assessing the reliability of the images. Therefore resolution analysis was an important part of our studies. Tomographic models often represent distorted images of the real Earth. The resolution of resulting images is finite due to, for instance, the choices in model parameterization (e.g. block size) and regularization (smoothing). It also spatially depends on the heterogeneous data coverage (Thurber and Ritsema, 2007).
5.5.1 Synthetic tests

There are several methods for the assessment of the resolution capabilities of tomographic data. One way is to perform synthetic tests. The main idea of such tests is to understand how well a hypothetical structure of the Earth can be recovered when a given inversion scheme applied. We investigated a number of synthetic models, including checkerboards (i.e., Bijward and Spakman, 2000; Shomali et al., 2002; Lippitsch et al., 2003; Sandoval et al., 2004). In Paper I, for instance, we present the inversion of synthetic data associated with a hypothetical structure but based on the same ray geometry as the real data. Gaussianly distributed noise with standard deviation equal to 0.1 sec, which was observed for P waves, was added into the synthetic data. The synthetic data were then inverted by employing the same inversion parameters (i.e. block size, damping, etc.) as used for the real data. The hypothetical structures are, in general, recovered properly (Figure 5.6). The spatial positions of assumed structures are resolved well. The amplitudes of the positive and negative anomalies are less well resolved, partly because of the damping and smoothing included in the inversion. We note, however, that, in this example, the synthetic test was not successful at recovering a relatively small and deep positive anomaly located beneath the northern part of the model. This is due to poor geometry (ray coverage) in this part of the model. Another popular type of hypothetical test is the ‘model restoration’ test in which synthetic observations calculated based on a model obtained from the inversion of real data are themselves inverted using the same method and parameters as those used for the real data (see Husen and Kissling, 2001, Paper II of the current thesis study). Such an approach can be very useful especially when extended to include comparative assessments of models lacking some of the key features (Thurber and Ritsema, 2007). However, using the inversion results as an input model to calculate synthetic travel time data may have some drawbacks. For instance areas of low resolution are identified as areas of good resolution due to the good recovery of the input model which either shows low-amplitude anomalies or even artifacts in areas of low resolution. Thus to assess the reliability of the model we decided to use a number of different methods including synthetic tests, constrained quadratic programming inversion, cell hit-count analysis etc.

5.5.2 Data Coverage

Data coverage has a great impact on the gross characteristics of the estimated tomographic models (Thurber and Ritsema, 2007). Thus cell hit-count and total ray length analyses, when considered together, are useful tools in evaluating resolution. Based on our analyses, the number of rays in each cell decreases significantly at greater depth as the rays spread laterally.
with increasing depth because their ray paths are not completely parallel to the major axis of the array (see Figure 5.7).

![Figure 5.6. (a) A hypothetical synthetic model. (b) Resolved model at the 4th iteration. (+) and (-) signs show the areas with higher and lower velocities relative to the IASP91 background model.](image)

Rapid near surface variations in hit-count can arise depending on our choice of profile relative to the locations of stations on the particular row of cells. These effects, together with the considerable smoothing involved in the inversion procedures means that metrics such as hit-count can not be directly rigorously interpreted to quantify resolution. However, the image does give an indication of where resolution is likely to be best i.e. in the central part of the array where station density is greatest.
5.5.3 Data Fit: Variance Reduction

One criterion which is very commonly used as a measure of goodness of data fit in seismic tomography studies is variance reduction, i.e. the ratio of aggregated data-model misfit before and after inversion. It is helpful in deciding how satisfactory the final solution is since during the whole inversion procedure resulting models can be influenced by several factors including the regularization, vertical and horizontal parameterization, etc. After the inversion of P waves extracted from in-line events (see Paper I), the variance reduction after the fourth iteration was around 65% while the inversion of shear waves gave ~50% (Paper II). Especially interesting is the large increase in variance reduction observed in our last paper (Paper IV) where the observed P relative residuals are corrected for the effects of anisotropy before inversion. Data variance reduction significantly increased from around 50% to 90%. This is a vast improvement of ~40% and implies that images produced after correcting the observed data for anisotropy much better represent the real Earth.

5.5.4 An auxiliary approach for inversion: Quadratic programming

For the assessment of the resolution and reliability of resolved models, another method is to use an auxiliary technique for the inversion of the observed data under different constraints. Such testing helps us to understand if major features (i.e., boundaries and blocks) apparently resolved by the inversion are genuinely required by the data or if they are in reality unconstrained or have been artificially introduced by the inversion process (see Shomali, 2001). Paper I presents an example using a quadratic programming method in order to assess the resolution and reliability of the model obtained. Quadratic programming is a well-established method in solving inverse prob-
lems, and its application to teleseismic tomography is discussed in detail by Shomali et al. (2002). One advantage of the method is that it facilitates the inclusion of inequality constraints into the inversion, which allows flexibly defined constraints to be included in the inversion to e.g. assess to what extent the data demands the inclusion of a particular feature in the model. The method was applied partly to see whether the application of a completely different inverse method reveals the same structures, or if some of these have been artificially caused by the inversion. As discussed above, in order to calculate the model parameters given in equation 5.11, damping, spatial smoothing, and eigenvalue rejection were applied to stabilize the inversion. In the application of quadratic programming, we used the same symmetric positive-definite objective function as used for equation 5.11 but without any spatial smoothing operator, $W^m$ (equation 5.11) and due to nature of the method no eigenvalue rejection was needed. The only requirement is now the lower and upper bounds of unknown model parameters, which were specified a priori to be ±3 % in velocity. This figure was chosen based on the characteristics of previous inversions of the same data, on results from previous studies (such as TOR), and on considerations of the likely petrophysics of the mantle materials (Kukkonen et al. 2003). Figure 5.8 presents two models obtained after inversions based on SVD and quadratic programming. Comparison of these two models tells us that the major blocks and transitions are required by the data and are not artifacts caused by the inverse algorithm. A possibility remains that while the features are not caused by the inversion algorithm as such, they are artifacts due to the fundamental character of the data, including possible noise. However, our various resolution tests and the use of independent inversion algorithms strongly suggest that the major features seen in the inverted images do represent significant lateral velocity variations at depth.
Figure 5.8 Tomographic models along the SNSN a) resolved after the SVD b) quadratic programming.
6. The concept of anisotropy

6.1. Seismic waves in anisotropic media

So far we have dealt with the velocity of seismic waves assuming that they propagate through an Earth structure made of purely isotropic and linearly elastic material in which the elastic properties of the material are the same in all directions and the stress-strain relationship is linearly proportional as defined by Hooke’s law (equation 6.1) (Stein and Wysession, 2003; Maupin and Park, 2007; Babuška and Cara, 1991):

\[ \sigma_{ij} = c_{ijkl} e_{kl} \quad (6.1) \]

where \( \sigma_{ij} \) and \( e_{kl} \) are the components of the stress (load) and strain (deformation) tensors respectively. \( c_{ijkl} \) is the elastic tensor containing a total of 81 terms. This represents the elastic properties of the material. Equation 6.1 actually corresponds to the partial derivative of the density of internal energy with respect to the strain tensor (Babuška and Cara, 1991). In an isotropic material the 81-term tensor of elastic moduli reduces to a form containing only two independent constants, \( \lambda \) and \( \mu \) (Lamé’s parameters) since the elastic properties of the material are the same in all directions. Considering an Earth structure with isotropic properties is a useful simplified approximation in many seismological studies. However in reality significant deviations from isotropy can and do exist. In other words, parts of the Earth are significantly anisotropic and the elastic constants of the material, and thus the reaction to stresses, are different in different directions. This means that the velocity of seismic waves propagating through an anisotropic material depends upon their direction of propagation (Stein and Wysession, 2003). A simple form of anisotropy can be found in a medium with a stack of parallel isotropic layers since the strength of such material differs in directions parallel to or perpendicular to the layers (Babuška and Cara, 1991). This type of (transversely isotropic) anisotropy has been considered as a suitable model for the mantle by e.g. Toksoz and Anderson (1963).

When the material is anisotropic, the form and complexity of the elastic stiffness tensor relating the stress to the strain via Hooke’s law become dependent upon the degree of the symmetry of the material. Symmetry means
that of the 81 elements, only at most 21 of these are independent and,

\[ c_{ijkl} = c_{jikl} = c_{ijlk} = c_{klij} \]  

(6.2)

Having \( c_{ijkl} = c_{jikl} \) yields 6 independent pairs \((i,j)\) while similarly, \( c_{ijkl} = c_{klij} \) results in 6 independent pairs \((k,l)\). Such symmetries can be used to introduce a 6 x 6 matrix \( C_{mn} \) describing the elasticity of the medium. \( C_{mn} \) is derived from \( c_{ijkl} \) by assigning the pairs of indices \((i,j)\) and \((k,l)\) to the indices \(m\) and \(n\) varying from 1 to 6 by taking values of \((1,1), (2,2), (3,3), (2,3), (1,3), \) and \((1,2)\), respectively (Voigt, 1928):

\[
C_{mn} = \begin{pmatrix}
C_{11} & C_{12} & C_{13} & C_{14} & C_{15} & C_{16} \\
C_{21} & C_{22} & C_{23} & C_{24} & C_{25} & C_{26} \\
C_{31} & C_{32} & C_{33} & C_{34} & C_{35} & C_{36} \\
C_{41} & C_{42} & C_{43} & C_{44} & C_{45} & C_{46} \\
C_{51} & C_{52} & C_{53} & C_{54} & C_{55} & C_{56} \\
C_{61} & C_{62} & C_{63} & C_{64} & C_{65} & C_{66}
\end{pmatrix}
\]  

(6.3)

\( C_{mn} \) is diagonally symmetric, i.e. \( C_{12} = C_{21}, C_{13} = C_{31}, C_{14} = C_{41}, C_{15} = C_{51}, C_{16} = C_{61}, C_{23} = C_{32}, C_{24} = C_{42}, C_{25} = C_{52}, C_{26} = C_{62}, C_{34} = C_{43}, C_{35} = C_{53}, C_{36} = C_{63}, C_{45} = C_{54}, C_{46} = C_{64}, \) and \( C_{56} = C_{65} \). Thus the number of independent elastic constants becomes 21.

In general any sample of the Earth will exhibit such anisotropy. However differences between some of the moduli will be small and only in exceptional cases can 21 different moduli be measured (Maupin and Park, 2007). In many cases simpler forms of anisotropy are well adequate to describe the properties of a given sample. For orthorhombic symmetry \( c \) contains 9 independent parameters, and 6 for tetragonal symmetry (see Babuška and Cara, 1991). Quite often, a simple form of anisotropy with an axis of symmetry and 5 independent elastic constants is considered to be a realistic approximation for parts of the Earth. Olivine crystals, the main contributor to upper mantle anisotropy, are considered to orientate via mechanisms leading to aggregate anisotropy with this type of symmetry (Maupin and Park, 2007). The nomenclature around anisotropy can be confusing. In cases where rotation around one particular axis does not change the elastic tensor, the material is sometimes said to be transversely isotropic (TI). The terms axisymmetry and cylindrical symmetry are also used, as is hexagonal symmetry. Where the symmetry axis is vertical, it may be termed radially anisotropic (Figure 6.1). This is one of the most common types of anisotropy observed in the Earth and corresponds to when a medium is composed of different isotropic layers which are perpendicular to the symmetry axis. In such a case the material behaves isotropically in directions transverse to the symmetry axis. According to the direction of symmetry axis, TI can be defined as hori-
zontal (HTI), vertical (VTI) or tilted. Garnero et al., (2004) investigates how the waveform characteristics of two shear waves (SH and SV) are affected by TI with a tilted axis (Figure 6.1).

Using Love notation, a TI medium can be described with five independent elastic coefficients A, C, F, L and N representing its aggregate properties. When the direction of symmetry is along $x_3$ (see Figure 6.2), then the properties in that direction becomes different from those in the $x_1$-$x_2$ plane. Thus equation 6.3 can be written in the following matrix form (Stein and Wysession, 2003):

\[
C_{mn} = \begin{pmatrix}
A & (A - 2N) & F & 0 & 0 & 0 \\
(A - 2N) & A & F & 0 & 0 & 0 \\
F & F & C & 0 & 0 & 0 \\
0 & 0 & 0 & L & 0 & 0 \\
0 & 0 & 0 & 0 & L & 0 \\
0 & 0 & 0 & 0 & 0 & N
\end{pmatrix}
\] (6.4)

where the independent components of $C_{mn}$ ($A$, $C$, $F$, $L$, and $N$) matrix are given as follows,

- $C_{11} = C_{22}$ (or $c_{1111} = c_{2222}$ of the stiffness tensor) is used for $A$,
- $C_{33}$ (or $c_{3333}$ of the stiffness tensor) is used for $C$,
- $C_{31} = C_{32}$ (or $c_{3311} = c_{3322}$ of the stiffness tensor) is used for $F$,
- $C_{44} = C_{55}$ (or $c_{2323} = c_{1313}$ of the stiffness tensor) is used for $L$,
- $C_{66}$ (or $c_{1212}$ of the stiffness tensor) is used for $N$.

![Figure 6.1](image)

*Figure 6.1. How orthogonal SV and SH shear waves are affected by propagation through a transverse isotropic material with different orientations of the symmetry axis (Garnero et al., 2004; Figure is courtesy of Edward Garnero).*
When the propagation of direction of waves is in the $x_j$ direction, P velocity and two orthogonal shear wave velocities are given by,

$$P_j = \frac{A}{\rho}, \quad \nu_j = \frac{N}{\rho}, \quad S_j = \frac{L}{\rho} \quad (6.5)$$

The velocity of shear waves traveling along $x_j$ is controlled by their particle motions. As a consequence shear waves split with waves polarized in one plane traveling faster than those polarized in the other. This is why we observe shear wave splitting on seismograms (Figure 6.3). Irrespective of the direction of the wave propagation in the $x_j$-$x_2$ plane, the same phenomena occurs since the physical properties in this plane are independent of direction (Stein and Wysession, 2003).

Figure 6.2. Illustration of a transversely isotropic (TI) medium with layering and two cases in which the direction of wave propagation is in the $x_1$ (top) and $x_3$ (bottom) directions. In both case the axis of symmetry is $x_3$ (Figure is courtesy of Stein and Wysession, 2003). The original figure is from Babuška and Cara (1991).
Figure 6.3. An illustration of SKS splitting after Crampin (1981). Blue (fast) and red (slow) components which are bifurcated upon encountering an anisotropic region (Figure is courtesy of Edward Garnero).

For a TI medium with an axis of symmetry along $x_3$, $S_1$ (SH) wave velocity is considered to be greater than $S_2$ (SV) since SH propagates preferentially along the fast layers while SV samples both fast and slow layers provided that the direction of propagation is in the $x_1$ direction.

Considering propagation parallel to $x_3$ in the same TI medium, then shear waves velocities for $S_1$ and $S_2$ do not show any difference. In such a case P velocity is given by,

$$P_2 = \sqrt{\frac{C}{\rho}}$$  \hspace{1cm} (6.6)

Similarly to the comparison between SH and SV waves explained above, for layered media $P_1$ and $P_2$ are P-waves velocities which vary depending on the direction of propagation (see Figure 6.2) and $P_1$ is generally greater than $P_2$ since the waves propagates preferentially in faster layers.

The degree of transverse anisotropy can be quantified using,

$$\xi = \frac{N}{L} = \left(\frac{S_1}{S_2}\right)^2, \phi = \frac{C}{A} = \left(\frac{P_2}{P_1}\right)^2, \text{ and } \eta = \frac{N}{(A-2L)}$$  \hspace{1cm} (6.7)

In an isotropic material, $\xi = \phi = \eta = 1$ while typically $\xi > 1$ and $\phi < 1$ for layered structures.

In fact, unlike in the isotropic case, in an anisotropic medium in general the direction of particle motion will not be perpendicular (for S) or parallel to (for P) the direction of propagation of the signal, and more correctly the waves should be termed quasi-S and quasi-P. However, if the medium is only weakly anisotropic, then the directions of polarization of the seismic waves are expected not to deviate significantly from the perpendicular or...
parallel to the direction of propagation \((\text{Babuška and Cara}, 1991)\). In such cases the angular dependence of the body wave velocities can be determined using first-order perturbation theory \((\text{Babuška and Cara}, 1991)\). Backus (1965) derived a first order description of the seismic velocity by introducing the perturbation stiffness tensor, \(\delta c_{ijkl}\),

\[
\delta c_{ijkl} = c_{ijkl} - c_{ijkl}^0 \quad (6.8)
\]

where \(c_{ijkl}^0\) is the stiffness tensor for a medium which is isotropic but otherwise similar to the anisotropic medium. This leads to an equation for the angular variation of the square of P velocities in the anisotropic and the corresponding isotropic medium, \(P(\theta)\) and \(P_0\), respectively:

\[
P(\theta)^2 - P_0^2 = A_1 + A_2 \cos 2\theta + A_3 \sin 2\theta + A_4 \cos 4\theta + A_5 \sin 4\theta \quad (6.9)
\]

where \(\theta\) is the azimuth of the wave propagation with respect to the axis \(x_1\) (see Figure 6.4) and the constants \(A_1, A_2, A_3, A_4, \text{ and } A_5\) are derived from the elastic coefficients and are defined by \((\text{Babuška and Cara}, 1991)\):

\[
A_1 = (3\delta c_{1111} + 2\delta c_{1122} + 4\delta c_{1212} + 3\delta c_{2222}) / 8\rho \quad (6.10)
\]
\[
A_2 = (\delta c_{1111} - \delta c_{2222}) / 2\rho \quad (6.11)
\]
\[
A_3 = (\delta c_{1112} + \delta c_{1212}) / \rho \quad (6.12)
\]
\[
A_4 = (\delta c_{1111} - 2\delta c_{1122} - 4\delta c_{1212} + \delta c_{2222}) / 8\rho \quad (6.13)
\]
\[
A_5 = (\delta c_{1112} - \delta c_{2122}) / 2\rho \quad (6.14)
\]

\textbf{Figure 6.4.} In a weakly anisotropic medium, the above Cartesian coordinate system is used to determine the azimuthal variation of P wave velocities (adapted from \textit{Babuška and Cara}, 1991).
6.2 Anisotropy of upper mantle

One of the main aims of this thesis is to investigate seismic anisotropy in the upper mantle below the Baltic Shield. Therefore we will look briefly at anisotropy in upper mantle - its origin, behavior (type of deformation, effects of water, partial melt etc.). We also review some published examples of anisotropy (both P and S waves) which are interpreted to be mainly due to the upper mantle. Much of this chapter is based on the book of Babuška and Cara (1991), which provides a detailed documentation of seismic anisotropy observations from different part of the world. Other very relevant publications are those of Savage (1999), Fouch and Rondenay (2006), Maupin and Park (2007), and Plomerová et al. (2008).

6.2.1. General considerations

While seismic anisotropy exists at many places within the Earth it is often considered to be particularly important within the upper mantle and many studies of this have been published (see e.g. see Babuška and Cara, 1991; Savage, 1999; Fouch and Rondenay, 2006; Maupin and Park, 2007; Plomerová et al., 2008). Seismic anisotropy here is generally associated with lattice preferred orientation (LPO) of the anisotropic minerals forming the mantle. The symmetry and strength of the large-scale upper mantle anisotropy reflects a portion of the mantle minerals whose crystallographic axes have been oriented by some deformation process (Mainprice and Nicolas, 1989; Babuška and Cara, 1991; Mainprice et al., 2005), acting in the past (fossil anisotropy) or at present. Olivine represents the major constituent of the upper mantle and as a highly anisotropic mineral plays an important role in defining its mechanical properties and seismic anisotropy (e.g. Karato, 1987; Nicolas and Christensen, 1987; Karato and Wu, 1993; Mainprice and Silver, 1993; Ben-Ismail and Mainprice, 1998; Holtzman et al., 2003). Kumazawa and Anderson (1969) have reported that single-crystal olivine undergoes variations in the wave speed of 24.6% for P and 22.3% for S waves depending on direction.

Crystals are characterized by three axes, which are not necessarily mutually perpendicular, denoted a-, b-, and c- respectively (Babuška and Cara, 1991). Many studies in the literature accept that the a-axis of the olivine crystals represents the direction in which P waves have highest velocities. If the a-axis is close to the horizontal plane this corresponds to the polarization direction of the fast SKS-waves. Hence the a-axis is usually called ‘the fast axis’. For the b- and c- axes seismic velocities do not show large differences and compared to the a-axis it is more difficult to detect their directions from the seismological data (Maupin and Park, 2007).
According to Jung and Karato (2001) and Holtzman et al. (2003) relating the a-axis to the direction of mantle flow may not be simple since both numerical models and observations from the rock samples suggest that the relationship between deformation and mineral orientation is controlled by several factors such as temperature, pressure, the presence of melt and water, strength and type of the deformation etc. Figure 6.5, for example, is intended to illustrate that uniaxial compression can lead to an orientation of the slow b-axis with the compression direction while the fast a-axes to will orient randomly in the plane normal to compression, resulting in slow velocities in the compression direction and faster velocities in the plane normal to it. In shear and for large deformation, then the fast axis tends to align with the shear direction and the slower axes in the plane perpendicular to it (Kaminski and Ribe, 2001; Maupin and Park, 2007).

Figure 6.5. Long term preferred orientation of dry olivine in simple shear and in uniaxial compressive deformation. The figure is taken from Maupin and Park (2007).

The effect of water on the relation between strain and orientation of the ‘fast axis’ has been studied by Jung and Karato (2001). According to these authors, wet conditions may lead to a different type of LPO in which the direction of the a-axis can orient at 90° from the strain direction. In other
words the new type of orientation becomes perpendicular to the direction expected for dry olivine under the same deformation process.

Holtzman et al. (2003) implied that a similar effect can be observed because of strain partitioning within partially molten regions. Such complications imply that even for homogeneous strain conditions, different bulk anisotropy may be induced depending on the conditions at in different volumes. Thus, even under simple strain complicated patterns of anisotropy may be produced.

Delay times (see Figure 6.3) measured between fast and slow polarized shear waves from the mantle are typically around ~1 s (e.g., Plenefisch et al., 2001; Babuška et al., 2008; Paper III). Observation of anisotropy-induced shear wave splitting can not directly tell us which part of the ray path is responsible for the effect, and all waves must traverse the crust to reach our sensors (Babuška et al., 2008). However, Barruol and Mainprice’s (1993) investigation of splitting and delay times caused by heterogeneous crust with small scale anisotropic bodies in a tectonic unit suggests effects only of the order of 0.2-0.3 s. Although small-scale observations of high frequency seismic waves (see Vavrycuk, 1993; Ruzek et al., 2003) indicate the significance of crustal anisotropy, especially within the upper few kilometers, contributions from crustal anisotropy could be only a small fraction of the observed split times for long-wavelength teleseismic SKS phases seen in broadband recordings. In this case the question is where the major anisotropy in the upper mantle resides (Babuška et al., 2008). Is it due to the mantle lithosphere or is it of sub-lithospheric origin? As emphasized previously the fabric of the sub-lithospheric part of the upper mantle is traditionally believed to be related to the LPO of olivine mineral. In such cases, the fast symmetry axis detected, commonly from shear wave splitting measurements, will orient with the present day asthenospheric flow which is considered to sub-horizontal in global scale. However, in continental regions where there is thicker lithosphere and often an undulated relief of the lithosphere-asthenosphere boundary (LAB), things may be more complicated, and especially so e.g. around the margins of convergent plates (Plomerová et al., 2008 and references therein). Babuška et al. (2008) claim that the delay times measured from shear wave splitting at the western Bohemian Massif are due both to the E-W strike of dipping mantle lithosphere fabrics and to E-W oriented present day flow in the asthenosphere. Their findings confirm results of previous workers (e.g., Levin et al., 1999; Fouch et al., 2000; Gung et al., 2003; Walker et al., 2004; Fouch and Rondenay, 2006) which imply that the anisotropy within the upper mantle of stable continental regions has both lithospheric and sub-lithospheric origin. Generally speaking, mantle anisotropy, deformation and flow can be affected by both asthenospheric and tectonic phenomena (Plomerová et al., 2008 and references therein).
Although there is not consensus regarding the maximum depth of upper mantle anisotropy, Karato and Wu (1993) claim that the dislocation creep regime for olivine implies that deformationally induced LPO may exist to depths of 200-400km. Other studies such as Fouch and Fischer (1996), Montagner and Kennett (1996), and Tommasi et al. (2004) claim to detect anisotropy at mantle transition zone depths (i.e. between 410-670 km).

6.2.2. Major techniques exploiting P- and S-waves observations for lithospheric and upper mantle anisotropy

6.2.2.1 Shear wave splitting measurements

One of the most popular tools for investigating seismic anisotropy is shear wave splitting analyses based SKS or SKKS waves (e.g., Kind et al., 1985, Vinnik et al., 1989, 1992, Özlalaybey and Savage, 1995; Silver, 1996, Savage, 1999; Babuška et al., 1993, 2008; Plomerová et al., 2002, 2006, 2008a). Due to the increasing number of data collected during passive seismic experiments and the development of fast computers and different analysis techniques, this technique has become common for direct estimation of seismic anisotropy in many regions with different tectonic settings such as continental collisions (e.g. Lev et al., 2006), strike slip faults (e.g. Sandvol et al., 2003; Levin et al., 2006), stable cratonic regions (e.g. Fouch and Rondenay, 2006; Plomerová et al., 2008), subduction zones (e.g. Greve et al., 2008) and rift zones (e.g. Kendall et al., 2005). Shear waves show distinctive behavior in an anisotropic medium (Maupin and Park, 2007) since they undergo birefringence so that quasi-shear waves with polarizations aligned with fast and slow orientations propagate at different speeds (Levin et al., 2008).

Ando et al. (1980) introduced an analysis technique based on measuring travel time delays between two split and orthogonally polarized shear waves. This phenomenon, analogous to light birefringence, is a common and well-established technique for studying the Earth’s structure. As described above, when shear waves propagate through anisotropic media they split into two quasi shear waves – fast and slow. In general an initial short time interval of linear particle motion will be observed corresponding to the linearly polarized fast phase. Thereafter interference between the fast and slow phases results in elliptical particle motion. If the original (unsplilt) S-wave is linearly polarized and the waves are affected only by a zone of consistent anisotropy, then the recorded S-wave can be rotated such that two very similar phases, apart from scaling and a simple time delay, are seen on the orthogonal components. These orthogonal phases can not be expected to be exactly identical, because of noise, interfering phases, the effect of near-receiver structure and the effect of the free surface, but are often very similar. The orientation of the anisotropy controls the rotation angle, and the time delay between the
phases shows the integrated effect of the velocity anisotropy. These parameters carry information about the mineral orientation of the mantle material.

Various techniques exist for evaluating splitting parameters, including eigenvalue methods, transverse energy minimization methods and correlation approaches (Silver and Chan, 1991; Savage and Silver, 1993; Vinnik et al., 1989; Levin et al., 1999). Some methods focus only on identifying the initial period of linear particle motion, others are based on the idea of recovering the particle motion of the initial shear waves, which in the case of the core-mantle refracted shear waves (e.g. SKS) is a linear radial (SV) polarization in the ray-path plane. A “correction for anisotropy” is achieved by rotating coordinate systems (LQT or ZRT, see below) and applying an appropriate time shift to the orthogonal polarization components. Among the techniques mentioned above, the cross correlation approach aims at detecting the maximum similarity between the linearly polarized fast and slow components (Iidaka and Niu, 1998; Levin et al., 1999). Another approach is a grid search inversion for finding the incoming polarization direction and removing the split delay time by maximizing the aspect ratio of the covariance matrix eigenvectors (Silver and Chan, 1988). Several further developments of the methods mentioned above are in use. For instance, the multi-event approach developed by Wolfe and Silver (1998) combines data from events approaching from various directions. This method aims at computing the best solution under the assumption a single anisotropic layer with horizontal fast axis. Chevrot (2000) introduced splitting intensity estimations as an additional measure of the anisotropy by analyzing the back-azimuthal variations of shear wave data to examine dipping or multiple layers of anisotropy. Reviews of methodologies are presented in e.g. Savage (1999) and Fouch Rondenay (2006). In the present thesis work the transverse energy minimization method was used and applied to SKS phases, According to Vecsey et al. (2008) this yields more stable solutions than other techniques.

As previously mentioned, conventional studies tend to assume azimuthal anisotropy only and approximate the anisotropic medium by symmetries with a horizontal ‘fast’ a-axis. The splitting parameters (the fast S polarization defined by azimuth φ and delay time δt of the slow split shear wave) are evaluated in the ZRT (Z vertical, R-radial, T-transversal) coordinate system only from the horizontal components (NS and EW) of the shear waves. The simplest forms of analysis implicitly assume a vertical angle of incidence. However in reality even for steeply incident SKS the angle of incidence is not zero and can be up to several degrees (e.g. up to 10-12º in this study). This means the possibility of significant signal on the vertical component. Taking into consideration also a general orientation of anisotropic structures in 3-D, we therefore evaluate the splitting in the ray-coordinate LQT coordinate system (L longitudinal, Q – radial, normal to the L-T plane, T transversal). In this case the fast polarization direction ψ is searched for in the Q-T plane, perpendicular to the ray-path plane L-Q by taking into consideration.
the incidence angle in the 3-D approach (Vecsey et al., 2008). By rotating the ray-path coordinate system by an angle $\psi$ in the Q-T plane and by shifting the Q and T components by the split delay time $\delta t$, we search for the orientation of the new coordinate system in which the energy content of the T component is minimized. The pair of the splitting parameters ($\psi$, $\delta t$) is obtained as a minimum of a misfit surface. The polarization $\psi$ of the fast component is usually defined by two (Euler) angles - azimuth $\phi$, measured from the north, and inclination $\theta$ measured from vertical upward (see Šílený and Plomerová, 1996; Vecsey et al., 2008). To evaluate the reliability of the splitting parameters we apply a bootstrapping technique and calculate standard deviations $\sigma_{\psi}$, $\sigma_{\delta t}$ for $\psi$ and $\delta t$, respectively (Sandvol and Hearn, 1994). An example of the shear-wave splitting evaluation procedure is shown in Figure 6.6.

Vecsey et al. (2008) describe the theoretical equations for a split wave considering a simple model with a homogeneous layer with a weak anisotropy in which incoming shear waves split into ‘fast (F)’ and ‘slow (S)’ waves which are polarized perpendicularly to each other in a plane (Q-T) normal to the direction of wave propagation. If we assume that a linearly polarized shear wave arrives at the bottom of an anisotropic layer, then in the ray parameter coordinate system the radial (Q) and transversal (T) components can here be described by,

$$Q_1(t) = A_0 f(t) \quad \text{and} \quad T_1(t) = B_0 f(t)$$

where $f(t)$ indicates the signal in the time domain with amplitudes $A_0$ and $B_0$. $A_0$ and $B_0$ are the parameters controlling the angle of linear particle motion. When shear waves encounter an anisotropic layer, then splitting occurs resulting in two quasi-shear waves, fast and slow. Then at the bottom of the layer the quasi-shear waves can be written as follows (Vecsey et al., 2008):

$$F_1(t) = Af(t) \quad \text{and} \quad S_1(t) = Bf(t)$$

where $A = A_0 \cos \psi_0 + B_0 \sin \psi_0$ and $B = -A_0 \sin \psi_0 + B_0 \cos \psi_0$. The relation between the old amplitudes ($A_0$ and $B_0$) and new ones ($A$ and $B$) is determined by the angle between Q and F axes ($\psi_0$).

Through the anisotropic layer the slow wave ($S_1$) is delayed by time $\delta_0$, thus slow and fast waves can be described by,

$$F_2(t) = F_1(t) \quad \text{and} \quad S_2(t) = S(t - \delta_0)$$
A rotation of $F_2$ and $S_2$ into the Q-T plane gives the equations for the split shear waves (Vecsey et al., 2008):

\[
Q_2(t) = A\cos\psi_0 f(t) - B\sin\psi_0 f(t - \delta_0) \tag{6.18}
\]

\[
T_2(t) = A\sin\psi_0 f(t) - B\cos\psi_0 f(t - \delta_0) \tag{6.19}
\]

The main idea behind this kind of shear wave splitting analysis is the reconstruction of the linearity of polarized shear waves at the bottom of the anisotropic layer. Thus we can approximate quasi-shear waves starting from equation 6.21 to 6.17 with a backward procedure. In this case recovered fast and slow components $F'_1$ and $S'_1$ can be written as,

\[
F'_1(t) = A\cos\psi_m f(t) + B\sin\psi_m f(t - \delta_0) \tag{6.20}
\]

\[
S'_1(t) = -A\sin\psi_m f(t + \delta) + B\cos\psi_m f(t + \delta - \delta_0) \tag{6.21}
\]

where $\psi_m = \psi - \psi_0$ and general values ($\delta$ and $\psi$) are searched for (Vecsey et al., 2008).

For e.g. SKS shear wave splitting analysis the near-vertical ray path means that lateral resolution is good. For instance, the Fresnel zone for an SKS phase with 12 s dominant period is around 90 km in radius at 150 km depth assuming a Fresnel zone approximation using the half-wavelength criteria for a vertically incident ray (Fouch and Rondenay, 2006). On the other hand, shear wave splitting is sensitive to the smaller-scale structures which can also be detected for tightly spaced arrays. In the case of very small scale anisotropic variations, however frequency dependent shear wave splitting effects must be evaluated (Fouch and Rondenay, 2006 and references therein). For any analysis of more complex anisotropic structures, e.g. dipping or multiple anisotropic layers, adequate back-azimuthal and incidence angle coverage is necessary (Fouch and Rondenay, 2006).

Although shear wave splitting methods are simple and convenient, the quality of published results varies. Especially for SKS data the range of available incidence angles and back-azimuths are limited due to the distribution of suitable sources. Limited ray-path coverage means that the splittings themselves are generally insufficient to allow retrieval of the 3-D orientation of the structures, even if the Earth were laterally homogeneous and had a single orientation of anisotropy. To some extent, this can be improved by including more data with a better geographical distribution of foci. However, as the method requires signals with good signal-to-noise ratio which can be obtained only from large events (M~6.5) the number of events available remains limited. Moreover shear wave analyses do not normally provide any constraints regarding the depth of the anisotropic structure because splitting may occur at any part of the Earth along the path of propagation.
Figure 6.6 Example of shear wave splitting evaluation. a) radial and transversal components of an SKS phase recorded at the station OST from an earthquake of 20030526-2313 with magnitude 6.9 Mw and epicentral distance 92.3°; b) Particle motions (PM), misfit function and splitting parameters from the transverse energy minimization method (Vecsey et al., 2008); c) Error estimates of the evaluated splitting parameters by the bootstrap method (Sandvol and Hearn, 1994); d) Schematic illustration of the LQT coordinate system and angles used in the SKS splitting measurements (Šílený and Plomerová, 1996).
6.2.2.2 P-wave anisotropy measurements

As well as affecting shear-wave propagation, seismic anisotropy also plays an important role in the propagation of P waves. First observations regarding seismic anisotropy within the Earth were obtained from the azimuthal variation of seismic velocities of Pn waves (Hess, 1964). Such azimuthal variations were confirmed later on by e.g. findings from Shearer and Orcutt (1986) and Gaherty et al. (2004) who reported ~5.5% and 3-4% variations below the Pacific and Atlantic Oceans, respectively. Compared to the convergent plate margins, detected anisotropy from Pn waves beneath the continental regions shows more complicated patterns which make interpretation difficult (Bamford, 1977, Hearn, 1999, Smith and Ekström, 1999). Observations from Pn waves in oceanic environments indicate that the direction of maximum velocity of this phase coincides with the direction of paleo-spreading, implying a straightforward relation between two directions. Beneath intra-plate structure, the fast axis of the Pn waves is reported to align with suture zones (Nemeth et al. 2005), orogenic regions (Hearn, 1999) and subduction arcs (Smith and Ekström, 1999, see Figure 6.7).

Figure 6.7. Figure from Smith and Ekström (1999) showing the fast direction of maximum velocity of Pn waves observed around the subducting arc in Japan.

Hearn (1999) observed that regions of low velocity heterogeneity beneath the Mediterranean Sea region are associated with high anisotropy. This may imply that water responsible for the low velocity also encourages LPO, and thus as a consequence, high anisotropy.

The variations of the Pn velocity are considered to be a direct measure of the uppermost mantle anisotropy. For cases where a sufficient regional coverage exists then Pn waves can provide good constraints for some parts of the Earth which have not yet been investigated otherwise in terms of anisotropy. However, the variations observed from Pn waves are limited in depth since they propagate close to the Moho. Moreover, measurements may suffer from poor lateral resolution due to the geometry of the system with long, near horizontal, ray paths. In general, Pn can only be analyzed for anisotropy.
under the assumption of horizontally orientated anisotropy, thus ignoring possible geometrical complexities (Maupin and Park, 2007).

In order to analyze P-wave anisotropy at greater depth Oreshin et al. (2002) use earthquakes from larger epicentral distances whose P waves sample the deeper parts of the upper mantle beneath Siberia. Their analyses indicate that P-wave anisotropy is as strong as ~3% and it decreases to ~2% towards the shallower depths. These results correspond well to those observed from global surface wave data for the continental regions (Maupin and Park, 2007).

The analysis of relative travel time residuals as function of back-azimuth is also an important tool which can provide useful additional constraints regarding seismic anisotropy. This type of analysis usually depends on relative delay time measurements of body waves (usually P waves) from events at regional or teleseismic distances in order to elucidate their azimuthal variations at single stations (see Babuška et al., 1984; Babuška and Cara, 1991). Though P waves carry primarily information about velocity heterogeneities, effects due to a directional dependence of velocities (anisotropy) can be extracted as well. The use of P waves for relative residual analysis allows a much better geographical distribution of foci than SKS. Moreover, measuring first arrival times of the P waves is often feasible for lower-magnitude events than those suitable for SKS splitting analysis. Relative delay times are calculated by the subtraction of the actual arrival times of a given phase from corresponding predicted ones for a reference background model (e.g. IASP91). Calculated directional means are normalized to a directional mean for each station, then resulting residuals are plotted in a lower hemisphere projection (P-spheres) as a function of back azimuth and angle of upward propagation (‘incidence’) within the upper mantle (Babuška et al., 1984).

The pattern of negative (relatively early arrivals) and positive (relatively late arrivals) is often bipolar, indicating clear fast and slow orientations. Distributions of relative residuals can be analyzed using visual inspection to define the principal directions or analytically by inversion of relative-delay time data to determine the anisotropic parameters such as the orientation of the principal axes of anisotropy (see Babuška and Plomerová, 2006, Babuška et al., 2008; Plomerová et al., 2001, 2002, 2006). The main advantages of the relative residuals analysis can be summarized as follows: (1) relative delay times are simple and rapid to compute in order to have an immediate first-order understanding of the anisotropic pattern within the upper mantle; (2) they can be a good candidate in characterizing structures with dipping axes of anisotropy; (3) the anisotropic parameters modeled via the delay time analyses exhibit a good lateral resolution controlled by the Fresnel Zone, which has a diameter of ~50-100 km for P waves at lower lithosphere depths (Fouch and Rondenay, 2006).

Although such analyses can yield significant information regarding anisotropic structures within the upper mantle, the trade-off between heterogenei-
ty and anisotropy effects in the P-residuals can create difficulties and must be taken into account. Variations of isotropic velocity on local or regional scales may lead to azimuthally varying delay times which could be similar to those caused by anisotropy (Fouch and Rondenay, 2006). Babuška et al. (2008) correct relative P-residuals to allow for variations of thickness of the crust and of distant heterogeneities. Effects from the deep mantle and focal areas are reduced by normalization, i.e. by calculating P residuals relative to a reference level which for each event is the mean residual computed either from residuals of a set of high-quality well distributed reference stations or from residuals of all stations of the array.

One can think that the most guaranteed way of obtaining exact information about the anisotropic structure within the Earth must be to invert using fully anisotropic Earth models. However, this is difficult. Firstly, the larger number of degrees of freedom in the model (compared to an isotropic model) will increase the underdeterminedness of the inversion problem. Inversion of P waves in crosshole surveys has shown that the problem is extremely ill-conditioned even with the restriction to TI structures (Pratt and Chapman, 1992; Maupin and Park, 2007). Secondly, such inversion approaches are strongly dependent on adequate data coverage (Wysession, 1996; Chevrot, 2000; Fouch and Rondenay, 2006).

In conclusion we note that the various methods for examining seismic anisotropy (e.g. shear wave splitting, delay times analyses etc.) and seismic velocity heterogeneities (e.g. seismic tomography) are often applied independently from each other (Fouch and Rondenay, 2006). However joint analyses of all available methods and datasets will possibly reduce the trade-off between anisotropy and heterogeneity (Babuška et al., 2008). Moreover, such joint analyses combining methods with different depth sensitivities and structure/anisotropy trade-offs (travel time residuals, SKS splitting, receiver functions, tomography etc.) should allow enhanced detection of vertical and lateral variations in anisotropy including non-horizontal symmetry axes (Plomerová et al., 2008).
This chapter is a summary of papers included in this thesis. I describe the main motivation behind each study, the methodology, main results and conclusions. Generally speaking, for all of the four papers, the main focus is to attain a robust estimation of some of the physical characteristics of the upper mantle beneath Sweden, which sits on the oldest part of the European continent, the Fennoscandian Shield. Robust models of seismic velocity heterogeneity of the Earth’s interior are key factors for better constraining other physical/chemical properties of the Earth such as density, thermal properties, viscosity, and composition. In other words, if our ambition is to understand the physics, chemistry and evolution of the deep Earth, we must first understand its present day seismic structure, including lateral velocity variations and possible anisotropy.

7.1. Paper I: Upper mantle structure of the Baltic Shield below the Swedish National Seismic Network (SNSN) Resolved by Teleseismic Tomography

Paper I deals with an application of an ACH-based teleseismic tomography method to travel time residuals obtained from teleseismic earthquakes recorded by the SNSN network. The tomography algorithm estimates the spatial extent and magnitude of velocity anomalies by back-projection of relative arrival-time residuals (Weiland et al., 1995).

7.1.1. Motivation

Robust images of the seismic velocities in the upper mantle are required to help constrain geodynamic models which are necessary to attain better insight into the evolution of Baltic Shield. Until recently, information on deep structure below the Baltic Shield, the oldest and tectonically stable part of the European continent, was based on a limited number of reflection/refraction profiles, primarily FENNOLORA, and a small number of earthquake seismological studies based on rather limited data sets. High resolution body wave tomography methods were first applied in the area in the two passive seismic experiments, TOR (1996-1997) and SVEKALAPKO.
(1998-1999). The TOR experiment investigated the deep lithosphere-
esthenosphere structure beneath the Trans European Suture Zone (TESZ) from northern Germany to the southern part of Sweden and revealed the presence of a thinner lithosphere under northern Germany (of the order of 100 km) compared to southern Sweden (over 200 km) (see Arlitt, 1999; Cotte et al., 2002; Plomerová et al., 2002; Shomali et al., 2002; Shomali et al., 2006). To the north-east, another passive experiment, SVEKALAPKO imaged P and S velocity variations across the Archean-Proterozoic tectonic suture in Finland. SVEKALAPKO isotropic seismic velocity models (e.g., Sandoval et al., 2004; Bruneton et al., 2004) indicate no clear deep boundary between the Archean and Proterozoic provinces (e.g., Sandoval et al., 2004; Bruneton et al., 2004), but investigations of lateral variations of upper mantle seismic anisotropy (Plomerová et al., 2006) indicate a deep transition between the Archean and Proterozoic domains.

There is a geographic gap of several hundred kilometers between the TOR and SVEKALAPKO arrays (Figure 7.1). Part of this gap is now covered by stations of the Swedish National Seismic Network (SNSN), but offshore between Sweden and Finland there is still relatively little data. In Paper I P-wave arrival time data are inverted to elucidate structures primarily in the depth range of 70 to 680 km (i.e. in the upper mantle) to yield an initial high resolution P-wave velocity model to complement results from the TOR and SVEKALAPKO arrays.

![Figure 7.1](image_url) Station distribution of the SNSN (red triangles) shown with TOR (blue circles) and SVEKALAPKO (black diamonds) arrays.
7.1.2. Method

7.1.2.1 Model Parameterization
The inversion was performed using a model parameterization consisting of a 75x75km block size horizontally and a vertically changing block size (30 to 70 km). 1287 out of total 3553 associated model parameters were chosen to be a floating nodes that were allowed to vary in the inversion since they were crossed by many upcoming rays. The starting model velocities are based on the IASP91 model (Kennett and Engdahl, 1991). Owing to the length of the SNSN array the Earth’s curvature has been taken into account. This was done by correcting the background velocities based on an Earth Flattening Transformation (Shearer 1999). Figure 7.2 represents the model parameterization in horizontal and vertical directions for both P (Paper I) and S (Paper II). The velocity interpolation between nodes is linear (Steck and Prothero, 1991). A 3-D ray-tracing calculation was used to estimate the positions of rays inside the 3-D target volume. We used Steck and Weiland’s simplex-based ray tracer to calculate minimum travel time ray paths in the 3-D volume (Steck and Prothero, 1991; Weiland et al., 1995).

7.1.2.2 Inversion
The data to be inverted is the relative residuals between the observed travel times extracted from the seismogram and calculated travel times which are computed through the simplex-based ray tracer based on the background model. Travel time perturbations within the model are calculated by updating ray paths during each iteration. The final inverted model shows primarily lateral velocity deviations with respect to a background model which remains unknown. Various different inversion methodologies were used, including an iterative weighted damped least squares approach. The approach includes a smoothing operator based on the second difference matrix (see e.g. Menke, 1989).

Due to the nature of the observed data sets and the geometry of the problem the system is very close to linear and the main deviations from the starting model are achieved during the first iteration, with only minor refinements in further iterations. Previous studies (i.e., Shomali et al., 2002; Lipsticks et al., 2003; Sandoval et al., 2004) using the teleseismic algorithm have shown that the optimal solution for the significant model parameters can be obtained normally within four iterations, consistent with experiences in this study. During the inversion a truncated SVD approach was used by discarding the smaller eigenvalues that probably do not contain useful information about Earth structure (i.e. those whose magnitude was below a chosen threshold level). The truncation limit of 150 for the smallest eigenvalue was chosen for the final inversion. The suitability of this value was later confirmed by the good correlation of models obtained using SVD inversion and a quite different method based on quadratic programming. The damping
parameter which controls the amount of minimization of the data misfit with respect to the model roughness and stabilizes the inversion is another important inversion parameter. Assessment of the optimal damping parameter was made based on inversions using a set of various damping values between 5 and 1000 sec$^2$ %$^{-2}$ and constructing trade-off curves between data variance and model length (normalized squared summation of the model perturbations) computed for different damping values. A damping level of 100 sec$^2$ %$^{-2}$ was chosen, as this appeared to correspond to a change in gradient in the curve. We also show that our selected damping is a reasonable choice for the recovery of a hypothetical structure containing low and high velocity anomalies (–3.0% to +3.0%) consistent with the observed anomalies.

Figure 7.2 Model parameterization a) Horizontal grid distribution used for both P and S inversions. The black solid line is the profile AA’. b) Layers and corresponding starting velocity models for P and S phases (based on the IASP91 model). Only velocities of the floating nodes are estimated during the inversion.

7.1.3. Results
The final P velocity perturbation model is presented in Figure 7.3. The figure shows the result of inversion of relative arrival-time residuals corrected
based on available information about the crustal structure and with an optimum damping value of 100 sec$^2$ %. The vertical cross-section shown in Figure 7.3 represents estimated velocity perturbations along a profile aligned with the major axis of the SNSN array produced using the IASP91 travel-time model as a starting model. Blue and red are regions with relatively higher and lower velocities. In most areas the crustal correction has only a limited effect on the derived model.

**Figure 7.3** Final model obtained with a damping of 100 sec$^2$ % after crustal correction.

Major differences between images resolved with and without crustal corrections are observed below the southern part of the network where there is a significant change in Moho-depth. The structures resolved at the central part of the model (between latitudes about 60°-64° N.) given in Figure 7.3 can be mainly divided into a shallow part (less than about 300 km depth) and a deeper part. In the shallower part the main characteristic is an apparently slab-like positive velocity anomaly dipping gently towards the north, where it underlies a structure of anomalously low relative velocity. Due to the limited resolution of the data, this apparently continuous “slab” could in fact be two or more spatially separate anomalies, and we should be cautious about interpreting the image as e.g. a continuous remnant lithospheric slab. The deeper part shows a pronounced (relatively) low velocity feature starting from the depth of 250-300 km and extending to the base of the model (between latitudes about 60°-64° N.). Shallower (up to about 100km below the surface) velocity perturbations decrease in magnitude from North to South. There is
an apparently clear and sharp near-vertical transition at latitude about 57.5° N and depths of from 70 to 250 km.

7.1.4 Conclusions
Non-linear teleseismic tomography was used to invert to P-wave residual arrival times along the SNSN network. During the inversion, regularization parameters (such as damping, a threshold for dropping the eigenvalues, data and model weighting matrices) were applied. Two different methods, Singular Value Decomposition (SVD) and Quadratic Programming (QP) were used in order to investigate the possible sensitivity of the derived models to the inversion algorithm. In terms of the major features in the models, the results from the different inversions are very consistent. A comparison of the models obtained before and after crustal correction indicates the crustal correction plays an important role in increasing the resolution and removing biased velocity variation estimates caused especially by significant changes in the Moho depth. Our various resolution analyses, including synthetic tests, ray density and quadratic programming, indicate that different blocks and transitions resolved in our tomographic images are required by the data and are not artifacts. The most important feature resolved in the central part of the model is an apparently slab-like positive velocity anomaly dipping gently towards the north, where it underlies a structure of anomalously low relative velocity and coincides with a possible boundary between Proterozoic and Archean lithospheric domains. This boundary separating the two different lithospheric domains is also resolved in a horizontal map of the vertically averaged P velocity perturbations. Although the difference in relative velocities across the presumed tectonic suture is considerable, it is not yet clear whether or not this feature is truly resolved. A pronounced relatively low velocity structure from depths of the order of 250-300 km to the bottom of the model and latitudes between about 60°-64° N characterizes the central part of the tomographic image. One characteristic appears to be common in to both the SNSN and TOR models. This is a near-vertical transition between latitudes of about 56.5° and 58.5° N, and depths 70 to 250 km. The sharpness of the well-resolved features in our tomographic model with lateral velocity changes of several percent over some tens of kilometers would appear to require an explanation in terms of composition rather than temperature or possible effects of anisotropy.
7.2 Paper II: S and P velocity heterogeneities within the upper mantle below the Baltic Shield

Paper II examines shear wave (SH and SV) velocity heterogeneities within the upper mantle of the Baltic Shield using the same type of non-linear tomographic inversion that we used in Paper I. For this purpose we have used the same type of model parameterization and teleseismic data as in Paper I. 1532 good quality S-wave relative arrival times obtained from both radial and tangential waveforms were initially analyzed independently and later in a combined manner.

7.2.1 Motivation

Geological conditions in the Swedish part of the Shield region allow very high quality recording of waveform data. Thus, in contrast to conventional shear wave regional and teleseismic tomography surveys, in this study we are able to invert the travel time residuals from both radial and tangential movements along the ray direction. The resulting models can be compared block by block. The ray geometry is essentially identical for the radial and tangential data inversions. After careful consideration of resolution issues, we consider that the significant observed differences between the images produced from the radial and tangential data is probably due to the influence of seismic anisotropy. The presence of seismic anisotropy in the region has been detected on various scales previously by Wylegalla et al., (1999), Plomerová et al., (2002, 2006), Vecsey et al., (2007). Our information on shear wave velocity perturbations is based on a similar data set (geometry) to that used in Paper I. This assists in evaluating consistencies and inconsistencies between the derived P and S models as a basis for further interpretations regarding e.g. what effects (thermal or chemical) may cause the lateral heterogeneities.

In the next part of this summary we will not touch upon any technical details about the method itself since these have already been mentioned in the summary of Paper I. Instead we discuss the results and possible conclusions of this work.

7.2.2 Results

In general, the S and P models present some striking similarities e.g. the near vertical transition at the southern end of the profile. In the centre of the profile (between the latitudes of ~59° and 62° N.), the shallower part of the P and S models (approximately down to 250 km) show similar anomalies except for a small scale relatively low velocity feature in the P model which does not appear in the SH model and is only weakly seen in the SV model. Another prominent structure which seen in the isotropic P inversions; the north-dipping slab-like positive anomaly in the P cross section, is not seen in
these S models. As can be seen from Figure 7.4a, to the north, below the Archean region, the SH model is well correlated with that for P except for a prominent positive anomalous body which can be traced up to 200 km from the bottom of the model (between the latitudes of ~66°-68° N). A similar structure is seen in the SV model in the same depth range, albeit with low amplitude (see Figure 7.4b). Analysis of the arrival time residuals shows that for over 90% of the data, SV and SH residuals are less than 1.5 sec apart. When rays with large SH-SV arrival time differences (> 1.5 sec) are excluded prior to inversion, the SH and SV images have automatically become more similar (Figures 7.4c,d). More surprisingly, these “isotropic” S-wave images now also become very similar to the P wave image (Paper I).

A cell by cell subtraction between two shear wave models (Figure 7.5) indicates that the level of difference between the SV and SH models is large (±3 or 4 %) which is not inconsistent with what we might expect due to shear wave anisotropy in the upper mantle (Silver, 1996; Savage, 1999; Smith and Ekström, 1999). The SH and P models show slightly greater similarity than SV and P (Figure 7.5). The main differences between the SH and SV models are mostly at shallower depth (down to 150-200km). Another significant observation is a significant SH-SV anomaly coincident with the top of the slab-like relative high velocity structure observed in the P-wave model. A correlation analysis which is performed layer by layer confirms that the S-wave models are most similar at greater depth (starting from 200 km).

In order to understand one possible effect which may result in different shear wave models for SH and SV data, synthetic tests with different levels of random noise were performed. We used noise levels corresponding to those in the real data (standard deviations of 0.55s and 0.7s, respectively for SH and SV). The recovered S models, in general, resolve the general characteristics of the synthetic model. There are some slight differences between two recovered models but these differences were only about ± 1% i.e. significantly smaller than the differences observed for the real data.
Figure 7.4 A cross-section of SH and SV velocity perturbation models along the profile AA' obtained from the Singular Value Decomposition (SVD) inversion. a) Final model from tangential (SH) phase inversion. b) Final model from radial (SV) phase inversion. c) Final isotropic model from tangential (SH) phase inversion after excluding rays with large SH-SV arrival time differences (> 1.5 sec.). d) Final isotropic model from radial (SV) phase inversion after excluding rays with large SH-SV arrival time differences (> 1.5 sec.). In both Figs. a and b, yellow contours indicate the borders of (-) or (+) Vp anomalies observed by Eken et al., (2007). (TIB: Transcandinavian Igneous Belt, Prt: Proterozoic, Arc: Archean).

7.2.3 Conclusions

The shear wave velocity models show a surprising degree of complexity, especially at small scales. The structures appear to be present in both the shear wave models and their difference images. Some small scale features at shallower depths, up to about 200 km in the south and slightly less in the north, show good correlation with geological models of the evolution of the area, e.g. Lahtinen et al., (2005), thus indicating a likely interpretation in terms of mostly Proterozoic tectonics. One of the most distinctive features detected in various ways in our models (i.e. P, S tomography and shear wave tomography differences) is the boundary between Archean-Proterozoic domains, although there does not seem to be a simple geochemical explanation for such boundaries observed underneath cratons. The observed seismic velocity heterogeneities of ±3-4 % for P and S waves are difficult to explain in
terms of changes in chemical composition. A plausible explanation is lack-
ing as to how some the proposed tectonic structures translate into the ob-
erved velocity structures. Comparing the differences between the SH and
SV images produced using the whole data set (Figures 7.4a,b) and excluding
rays with large SH-SV residuals (Figures 7.4c,d) reveals that the apparently
resolved differences between the respective SH and SV sections are consis-
tent, albeit with significantly lower amplitude, in the limited data set. We
suggest that these results demonstrate that the differences in SH-SV arrival
times are, at least primarily, due to anisotropy. By excluding the data most
affected by anisotropy, we stabilize the inversion and produce results more
consistent with the independent P wave data. However, even the remaining
data may be affected by anisotropy, though to a lesser degree. Therefore, that
the SH-SV differences remain but with lower amplitude supports our hypo-
thesis. Moreover both layer by layer correlation analysis and cell by cell
comparison of radial and tangential models yield larger scale differences.
Such differences appear as poor SV-SH correlation at shallower depths. We
attribute this to the effects of anisotropy which may be a result of ancient
strains related to the continental accretion. For a better understanding of
whether or not seismic anisotropy exists in the region, investigation of aniso-
tropic parameters with various conventional geophysical techniques (i.e.
SKS splitting, directional P-waves analysis, Receiver Function analyses,
surface wave analysis etc.) or performing tomography based on fully aniso-
tropic inversions is required.
7.3 Paper III: Seismic anisotropy of the mantle lithosphere beneath the Swedish National Seismic Network (SNSN)

In Paper III we perform a body wave analysis including shear-wave splitting and P-travel time residuals in order to detect anisotropic structure of the upper mantle beneath the Swedish part of Fennoscandia. Moreover body-wave anisotropic parameters are jointly analyzed to produce 3-D self-consistent anisotropic models of well-defined mantle lithosphere domains with differently oriented fabrics approximated by hexagonal aggregates with plunging symmetry axes.
7.3.1 Motivation

In addition to various geochemical and geological constraints, detecting and modeling 3-D anisotropy and mapping fabrics of lithosphere domains contributes to understanding past geodynamic events and the tectonic evolution of the Earth’s interior. Seismic anisotropy is of major significance in understanding the deep structure of the Earth because it reflects the texture of rocks (Babuška and Cara, 1991).

Previous results observed in various parts of the region are in accord with a domain-like structure of the mantle lithosphere, retaining large-scale fossil fabric related to the origin of the Precambrian continental fragments (Plomerová et al., 2001; 2002). The main purpose of this study is to test the capacity of such a 3-D approach in modelling seismic anisotropy beneath the Swedish part of the Shield and to complement knowledge regarding the mosaic of upper mantle structure beneath Fennoscandia, including the mantle fabric around the northern continuation of the Proterozoic-Archean contact zone from south-central Finland (Vesey et al., 2007). Moreover, after isotropic inversions of separate SH and SV data considerable discrepancies between shear-wave velocity perturbations have indicated anisotropic structure in the upper 200 km of the upper mantle (Paper II). In this work, however, our main aim is to conduct a pioneering detailed analysis of seismic anisotropy parameters beneath the SNSN network with the use of conventional techniques on teleseismic observables (e.g. shear waves splitting, P residual analyses). Moreover in a final stage, Paper III examines the capabilities of a 3-D self-consistent anisotropy modeling approach for the Swedish part of the Shield using joint inversion and interpretation of the splitting with independent anisotropic parameters extracted from P-wave arrival times.

7.3.2 Methods

7.3.2.1 Shear wave splitting

As described in section 6.2.2.1 Ando et al., (1980) first introduced the shear wave splitting method. The method simply depends on measuring travel time delays between two split linearly polarized shear waves which are generated when traveling through anisotropic media. When core-mantle refracted shear waves (SKS) propagate through anisotropic media they split into two quasi-shear waves – fast and slow. Generally, an initial short time interval of linear particle motion will be observed corresponding to the linearly polarized fast phase. Interference between the fast and slow phases then generally causes elliptical particle motion. If the original (unsplit) S-wave is linearly polarized and the waves are affected only by a zone of consistent anisotropy, then the recorded S-wave can be rotated such that two very similar phases, apart from scaling and a simple time delay, are seen on the orthogonal components. Due to the presence of noise, interfering phases, the effect of near-receiver struc-
ture and the effect of the free surface, these orthogonal phases may not be expected to be exactly identical. However they are often very similar. The resulting rotation angle is controlled by the orientation of the anisotropy and the time delay between the phases is associated with the integrated effect of the velocity anisotropy. These parameters are important in terms of giving information about the mineral orientation of the mantle material. Due to the increasing amount of data collected during passive seismic experiments and the development of fast computers and different analysis techniques, SKS splitting analysis has become a well-developed and famous technique to detect the anisotropic characteristics of the mantle in many regions and in different tectonic settings (e.g., Kind et al., 1985, Vinnik et al., 1989, 1992, Özalaybey and Savage, 1995; Silver, 1996, Savage, 1999; Fouch and Rondenay, 2006; Park and Levin, 2002; Plomerová et al., 2002, 2006, 2008a).

7.3.2.2 Evaluation of P-residual spheres
Shear wave splitting is one of the most common tools in measuring seismic anisotropy. However SKS phases arrive from a limited range of incidence angles and back-azimuths, thus the method has a limitation making it difficult to model the 3-D orientation of the structures. This is significant considering that the structure of the upper mantle is often more complicated than a model with a single anisotropic layer with a horizontal ‘fast’ symmetry axis. To overcome such problems requires adding more data with a much better geographical distribution of available foci. The propagation of P waves is also affected by seismic anisotropy. Compared to SKS they show a better azimuthal coverage. Moreover, measuring first arrival times of the P waves is often feasible for lower-magnitude events than those suitable for SKS splitting analysis. Thus Paper III deals with analyses of P waves by a method described first by Babuška et al., (1984). The pattern of relative P travel time residuals plotted in a lower hemisphere projection (P-spheres) as a function of back azimuth and angle of upward propagation (‘incidence’) within the upper mantle is analyzed. In other words, the P-spheres constructed for each station actually show the directional terms of relative residuals. In general, positive residuals in P-spheres (relatively delayed arrivals) represent low-velocity directions of propagation, while negative values (relatively earlier arrivals) indicate high-velocity directions. The anisotropic pattern which could be extracted from the P-spheres may indicate the possible existence of consistent fabrics of individual lithosphere domains over large continental provinces (Babuška and Plomerová, 2006, Babuška et al., 2008; Plomerová et al., 2001, 2002, 2006).
7.3.2.4 Modeling anisotropic structure- joint inversion

An assumption of homogeneous anisotropy within individual domains in the upper mantle overlying the sub-lithospheric mantle is common for most of the anisotropic modeling. A type of anisotropy of peridotite aggregates with elastic tensors derived from data from mantle xenoliths is considered to be a suitable approximation for the large scale upper mantle (Christensen, 1984, Nicolas and Christensen, 1987, Ben-Ismail and Mainprice, 1998, Babuška and Cara, 1991; Babuška et al., 1993). To obtain information about anisotropic structure within the continental mantle lithosphere in this paper a joint analyses of P and S data was performed. Such analyses enabled us to reduce some of ambiguities and seeming inconsistencies (Babuška et al., 1993; 2008) in modelling anisotropy within the mantle. The main idea in modeling is to minimize the differences between anisotropic parameters which were measured through SKS splitting and P residual analyses and their corresponding theoretical values. One routine procedure for minimization is a direct joint inversion of both the shear-wave splitting parameters and the directional terms of relative P residuals. However, due to the fact that we have different data-sets which are different in character, a simple weighted summation of misfits may not be the optimal method. Therefore a multiobjective optimization procedure (MOP) is used for solving the inverse problem. In this case a set of Pareto optimal solutions are searched for in the global minimization (see Kozlovskaya et al., 2007 for more details). Such a minimization approach allows us to use combined solutions of individual pairs of misfit functions or a minimum of a single one.

7.3.3 Results

Our results from SKS splitting analyses indicate that the fast S polarizations of waves from the west and east differ at most of stations. Lateral variations in anisotropy along the network are obvious since different polarizations were measured at different stations. However, closely located stations generally show a similar pattern. In order to map lateral variations the region was divided in to 4 groups according to the character of the observed polarizations at different stations. Figure 7.6 shows the geographic variation of evaluated splitting using an example of a single shear wave approaching the SNSN from the ~E (i.e. the epicentral distance does not change much along the array). Evaluated polarizations for this sample event indicate the direction of anisotropy points to the SW in Region 1 and to about W in Region 2. Nulls dominate in Region 3, while polarizations to the NW characterize Region 4 in the south. Observed split delay times are up to 2s. In this work, split delay times less than 0.2 s are considered as nulls and they are not taken into account. The mean delay time of 223 split shear wave measurements is 0.91±0.19s.
Figure 7.6 Lateral variations of evaluated splitting parameters across the array. Shown are fast polarization azimuths $\phi$ and split times $\delta t$ of the SKS phase. Four regions with similar anisotropic signal expressed in shear-wave splitting characteristics are colored.
As an auxiliary tool, P residual spheres for all stations with sufficient data were constructed. This enables us to infer velocity variations beneath each of them as a function of the direction of wave propagation. The derived directional terms range between $\pm 0.5$ sec ($-0.5 < R_{i,k} < 0.5$) and often indicate a systematic back-azimuthal dependency. Stations used in this study mostly show ‘bipolar patterns’, with negative and positive residuals consistently in separate halves of the lower hemisphere projection. Negative residuals are characteristics of the faster waves relative to the isotropic velocity mean (directional mean) beneath each station. A visual inspection of P spheres based on their patterns allows us to divide our study area into five geographical groups of stations with a similar P sphere pattern. For some stations there was no observable systematic back-azimuthal variation of the residuals (black triangles, Figure 7.7). Such behavior is observed at some stations around 59-60°N along the coast in the middle part of the array. For the northernmost group of stations (grey stations in Figure 7.7) we could not detect a ‘bipolar character’, although the residual distribution was not random. Consistent orientation of directions of relatively higher and lower velocities beneath each station enabled us to group the stations based on visual inspection of the patterns of P spheres. The characteristics of groups changes when crossing from one domain to another.

Modelling of the anisotropic fabric of several mantle domains beneath the SNSN was performed in two steps. First the difference between the observed residuals from P spheres and their corresponding synthetics were minimized. During the minimization stations exhibiting no clear pattern were not included in the inversion. Inversions of the P residuals alone resulted in well-defined mantle lithosphere domains with differently oriented fabrics approximated by hexagonal aggregates with plunging symmetry axes. Models may have either fast a- or slow b-axes. In the southern and central parts of the region, the high-velocity foliation $(a,c)$ in the b-axis models or lineation $a$ in the a-axis models dip to the NE. They are orientated with approximately orthogonal azimuths in the south-central and south-north parts – they dip to the SW and NW, respectively. In general, single P inversions yield models with dipping $(a,c)$ foliations which seem to be more suitable for the lithosphere beneath the SNSN. The second step in modeling consists of the inversion of SKS splitting parameters and joint inversion of all type of body wave anisotropic parameters with all possible combinations of misfit functions $f_1$, $f_2$ and $f_3$ and isotropic velocities of the models in the range of 8.0 – 8.5 km/s (see Table 2 of Paper III). Solutions in the table are from the stable ones satisfying the independent P and S wave information in the data. Criteria of stability were small differences between results from a summation of misfit functions and optimum Pareto optimum solutions, as well as the lowest values of misfit functions in the inversion outputs. Final Pareto optimal solutions usually confirm the b-axis models as the most suitable approximation of the anisotropic structure of the individual
mantle listhosphere domains (see Figure 7.8) with only one exception (small Group 4-North Svecofennian, see Table 1 of Paper III).

Figure 7.7 Stations of the SNSN grouped according to their P–sphere type and examples of typical P spheres for each group. The spheres show azimuth–incidence angle dependent terms of relative residuals related to anisotropy. Negative terms (blue) mark relatively high-velocity directions of propagation, whereas positive terms (red) correspond to low-velocity directions relative to an isotropic mean velocity beneath each station. Dashed curves indicate boundaries between the mantle lithosphere domains with similar fabric as seen by the P waves.
7.3.4 Conclusions

The first significant conclusion after the interpretations of the results from such detailed analyses of body waves of teleseismic observables is the existence of seismic anisotropy beneath the Swedish part of Baltic Shield as previously reported at various parts of the Fennoscandian Shield (Wylegalla et al., 1999; Plomerová et al., 2002, 2006; Vecsey et al., 2007; Olsson et al., 2007). Individual and joint analyses of SKS and P waves have shown that the mantle lithosphere is formed by several domain-like structures with their own fabric. These fabrics were modeled successfully using the approximation of 3-D self-consistent anisotropic structures with plunging symmetry axes. The domains in the Proterozoic part of Fennoscandia seem to be sharply bounded by sutures cutting the whole lithosphere and their boundaries are also in good accordance with the boundaries of main crustal terrains. In addition to the detection of seismic anisotropy in Paper III, we observed a good correlation between the isotropic tomography models (P, SH, and SV models) and domain boundaries resolved after SKS splitting and P residual analyses, i.e. the boundaries often correspond to both a change in bulk velocity and a change in anisotropic fabric. The revealed domain-like structure of the mantle lithosphere beneath Sweden, which retains fossil anisotropy which probably originated prior to assembly of its fragments, supports the idea of the existence of an early form of plate tectonics during formation of continental cratons already in the Archean and the Proterozoic.
Figure 7.8 3-D self-consistent anisotropic models of individual mantle lithosphere domains derived by joint inversion of body-wave anisotropic parameters. Boundaries of the domains (dashed green) correlate with those of major crustal terranes (orange curves). The mantle lithosphere domains of the Proterozoic part of Fennoscandia beneath the SNSN seem to be sharply bounded and separated by a narrow and steep contact (suture) zone cutting the whole.
7.4 Paper IV: Effects of seismic anisotropy on P–velocity tomography of the Baltic Shield

Paper IV is an attempt to investigate how significantly seismic anisotropy beneath the Baltic Shield may influence tomographic images produced from P-wave data if an isotropic medium is assumed in the inversion. The concept we investigate is based on “correcting” each individual arrival time under the assumption that the information deduced about anisotropy by using the methods described in Paper III is correct. This corrected data can then be inverted and the resulting models compared.

7.4.1 Motivation

Conventional studies aimed at elucidating the 3-D seismic velocity structure of the Earth usually assume that the Earth is isotropic. However, we know very well from seismological data that the Earth is often significantly anisotropic, especially in the lithosphere. Convincing evidence for anisotropy is supplied by the well-established SKS splitting technique (Vinnik et al., 1984; Ando et al., 1980; Šílený and Plomerová, 1996; Savage et al., 1999; Fouch and Rondenay, 2006), directional variations of P-wave travel time delays (Babuška et al., 1984; Babuška and Plomerová, 2006; Plomerová et al., 1996; Hearn, 1996; Eberhart-Phillips and Henderson, 2003; Hirahara and Ishikawa, 1984; Ishise and Oda, 2008) and surface waves (e.g., Anderson, 1961; Aki and Kaminuma, 1963; Anderson and Dziewonski, 1982; Montagner, 1998; Bruneton et al., 2004). Obviously, seismic anisotropy gives rise to a more complex pattern of wave travel times than for a corresponding isotropic medium. Despite the fact that teleseismic tomography is usually a near-linear inverse problem, it is not immediately obvious how arrival time perturbations caused by anisotropy will map into a tomographic image produced under the assumption of isotropy. Clearly, the perturbing effect of anisotropy will depend on both the Earth structure and available ray paths. (e.g., Sobolev, 1999; Lloyd and van der Lee, 2008) Therefore the sensitivity of the inversion may be different in different areas, and here we investigate the possible effects of anisotropy on data from the Swedish National Seismic Network.

7.4.2 Method

Teleseismic P-wave travel time residuals are assumed to carry perturbations caused by seismic anisotropy along the propagation path. The anisotropic component in the observed (arrival time) data can be estimated by using the technique of Babuška et al., (1984). This is designed to separate the arrival time data into a part related to travel time perturbations corresponding to an isotropic medium and a part related to anisotropy. With a limited amount of
data and unknown Earth structure, there is always the risk of “cross talk” between components related to anisotropy and those related to lateral inhomogeneity of a spatial scale comparable to or smaller than the zone of anisotropy. However, previous studies and also Paper III of this thesis indicate that the method can be robust. Babuška and Plomerová, (1992) decompose relative residuals corrected for crustal effects into an isotropic part and a directionally dependent part presumed to be related to anisotropy. Data used consists of arrival times of teleseismic earthquakes from sources at distances of ~30-90° with angles of incidence ~22-38°. The analysis provides an estimate of a directional mean representing the reference level for calculating the directional terms (azimuth-incidence-angle dependent constituents of the relative residuals). More detail about the method can be found in Babuška and Plomerová, (1992). If the identification of the anisotropic component in the data has been successful, then the effect of anisotropy can in principle be removed by subtracting directional terms from individual relative residuals. The “corrected” data can then be inverted using the same parameterization as for the initial (“isotropic”) inversion.

7.4.3 Results

The directional terms generally exhibit systematic dependence on back azimuth. The observed patterns are often consistent at neighboring stations, but are different in different geographical areas. Most of the directional terms are in the range -0.5 and +0.5 s, but with some extreme values of the directional terms of up to ~+1 or -1s. Figure 7.9 shows an example of the distribution of directional terms calculated at the station VXJ. From the figure, we can say that the positive residuals from late arriving waves are dominant for events with back azimuths of ~0° to 130°-140°. The majority of the negative residuals appear in the range of back azimuths of ~180°-360°. For many stations in the SNSN, such a sharp separation of directional terms with back azimuth is observed, supporting the contention that the difference is largely or completely due to anisotropy. While results at different stations vary, most stations do show a consistent pattern of directionality in the residuals with a significant “bipolar” component, i.e. a component compatible with large scale anisotropy, suggesting that it may be possible to improve the tomographic inversions by compensating for the identified effect of anisotropy. We apply this concept to the data inverted in Paper I, and apparently significant changes in the tomographic images are observed.
7.4.4 Discussions and conclusions

The effect of seismic anisotropy on isotropic tomography images has been investigated previously by several studies using various seismic waves. For instance, Sobolev (1999) has shown that synthetic models of anisotropic bodies with a dipping olivine “a” axis associated with fossil subduction can produce bias in the velocity images while the effect of a pure shear deformation type anisotropy due to extension or compression may be ignored. In recent work, Lloyd and van der Lee (2008), however, claim that the influence of anisotropy on shear wave velocity images is weak.

Generally speaking, isotropic seismic tomography studies conducted in and around the Baltic Shield indicate velocity perturbations of $\sim \pm 2\text{-}3\%$ for P and S waves (Shomali et al., 2002, 2006, Sandoval et al., 2004; Bruneton et al., 2004; Paper I-II-III). According to normal resolution analyses, these studies show significant and resolved lateral variations in velocity to greater depths than appears to be consistent with our current understanding of the physical properties of mantle materials and our preconceptions about plausible variations in temperature, composition etc in the area. Several recent studies in the region (Wylegalla et al., 1999; Plomerová et al., 2001; 2002, 2006; Vecsey et al., 2007; Paper III) demonstrate the existence of seismic anisotropy in the upper mantle beneath the Shield.

Figure 7.9. A view of the directional terms of relative P residuals observed for VXJ station vs. backazimuth. Directional terms are considered to characterize well the P-wave anisotropy beneath the network. The larger directional terms are visually enhanced by scaling the symbols according to magnitude.
When inversions based on data uncorrected and corrected for the effect of anisotropy are compared, the latter appears to be slightly more simple (less laterally inhomogeneous) at greater depths. This suggests that the significant lateral velocity variations deduced from the isotropic analysis for depths of 250km or more could be artifacts produced by the presence of anisotropy (plausibly within the upper 250km). This is clearly an issue which should be further investigated.

Irrespective of the correctness of our estimates of the anisotropic parameters, the estimated magnitudes of mantle anisotropy (Paper III) are clearly reasonable when compared to existing knowledge from mineral physics. If we assume that our analysis procedure itself does not introduce bias or artifacts, then the differences in the models demonstrate that anisotropy can significantly distort the P-wave tomographic image. These differences can be significant. Furthermore if our estimates of the anisotropy of the identified domains are reasonably reliable (not just in general magnitude) then the new inversions should provide a more reliable image of the deep earth than the models. Clearly, this could have important implications for many teleseismic tomography studies, even in other parts of the world.
Isotropa och anisotropa P- och S-vågshastigheter hos manteln under den Baltiska skölden: Resultat från analys av teleseismiska volymsvågor

Under flera årtionden har tolkningen av den tids- och amplitudinformation som för olika vågfaser kan identifieras i ett seismogram gett oss betydande kunskap om Jordens uppbyggnad. Seismiska data hjälper oss inte bara att förstå Jordens nuvarande uppbyggnad utan bidrar även till vår tolkning av de plattektoniska processer som föråldras den.

Vår kunskap om den övre mantelns 3-dimensionella struktur under den Baltiska skölden, tektoniskt den mest stabila delen av den Europeiska kontinenten, har ökat enormt genom två tidigare passiva seismiska experiment, TOR och SVEKALAPKO. De data som registrerades av dessa nätverk kan nu kompletteras med högkvalitativa trekomponentregistreringar från det nya permanenta Svenska Nationella Seismiska Nätet (SNSN). Syftet med denna avhandling är att belysa den isotropa och anisotropa hastighetsstrukturen hos den övre manteln, främst under den svenska delen av Baltiska skölden. Detta har gjorts genom analys av teleseismiska volymsvågor (dvs. P- och S-vågor) med metoder såsom seismisk tomografi och shear wave splitting.


Vi inverterade ankomsttider från 52 jordbävningar registrerade av stationer i det Svenska Nationella Seismiska Nätet (SNSN). För att minimera komplikationer på grund av tredimensionella strukturer valdes teleseismsar...
parallella (±30º) med nätets längdaxel. Resultaten från seismisk tomografi visar variationer i P-våghastigheten på ± 3 % ned till ett djup av minst 470 km. Nätverkets storlek ger oss möjlighet att studera strukturer på än större djup och lateral hastighetsvariationer ned till 680 km djup kan upplösas. Under nätets centrala delar (60º - 64º N) där geometrin ger bäst täckning, visar data på ett stort område med relativt låga hastigheter på djup större än 300 km. På djup under 250-300 km innehåller modellen ett antal strukturer, bl.a. spår av subduktion i form av en struktur med svag lutning mot norr.

På samma sätt analyserades S-vågsankomster. Dock kunde vi här, till skillnad mot liknande tidigare studier använda högkvalitativa registreringar både av den radiella (SV) och tangentiella (SH) komponenten av S-vågen. Inversion av de relativa ankomsttiderna utfördes separat för de två komponenterna med samma modellparametrar. De resulterande modellerna har flera gemensamma storskaliga strukturer, av vilka många också återfinns i resultatet av P-vågsanalysen. Lateral hastighetsvariationer på flera procent (±3-4 %) kan observeras på djup av minst 470 km. Korrelation mellan SH och SV-modellerna visar på ett mönster av små men tydliga skillnader ned till 150-200 km, under vilket djup modellerna väsentligen är desamma. Direkt jämförelse av hastigheter i enskilda celler av modellerna visar ett liknande mönster med skillnader mellan modellerna på upp till 4 %. Numeriska tester visar att skillnaden mellan de två S-vågsmodellerna bara delvis kan tillskrivas brus och begränsningar i upplösning, och att den i vissa delar beror på storskaliga anisotropiska strukturer. Ett område där modellerna skiljer sig väsentligt sammanfaller med den förmodade gränsen mellan Arkeiska och Proterozoiska delar av Baltiska skölden, och antyder därmed olika karaktär av anisotropi.

för tanken att plattektoniska processer var verksamma i uppbyggandet av kontinenternas sköldområden redan under Arkeisk tidsålder. På samma sätt som andra geologiska och geokemiska metoder kan därför modellering och kartläggning av den 3-dimensionella anisotropa strukturen hos Litosfären bidra till vår tolkning av plattektoniska skeenden i jordens urtid.

Möjliga störningar från anisotropi på seismisk tomografi undersöktes och befanns vara potentiellt signifikant. Ankomsstider för P-vågen justerades utifrån uppskattningar av mantelns anisotropi och inverterades på nytt. Även om mönstret i hastighetsmodellerna i stora delar inte ändras, ändras det betydligt i vissa delar. Till exempel försvinner de spår av en struktur relaterad till subduktion som kunde identifieras genom inversion av originaldata. Analysen visar därmed att anisotropi av rimlig storlek kan ha betydande effekt på de tomografiska bilderna och inte kan förbises. Om vår uppskattning av storleken på anisotropin är någorlunda riktig bör modellen baserad på korrigerade data ge en mer robust och korrekt bild av mantelns struktur.
Baltık Kalkanı manto yapısının izotropik ve anizotropik P-ve S-hızları: Telesismik cisim dalgaları analizlerinden elde edilen sonuçlar


Doktora çalışmanın ilk parçası olarak, Baltık Kalkanı’nın üst mantosuna ait telesismik P dalga hızlarını modellemek için P dalgalarının seyahat zamanını gözlemek için kullanlan sismik tomografi yöntemini kullandık. Teleseism, uzak mesafeli depremlerden elde edilen sismik dalgalı olup ve bizim kayıtlarımızı tarafından kaydedilmeden önce uzun bir yol katdederek yerin derinliklerine kadar nüfuz etmişlerdir. Sismik tomografi modeli, medikal tomografiden jeo-bilimsel problemlere 1976 yılında Aki ve Lee tarafından uyarlanmıştır. Sismik tomografi genel hatlarıyla bir ters çözüm problemidir ve de elastik, anizotropik, izotropik ve yoğunluk özellikleri gibi sismik dalga yayılmasını etkileyen yerin üç boyutlu özellikleri hakkında bilgi veren iyi oluşturmuş bir tekniktir. Sismik tomografının değişik coğrafik ve derin bölümlere ilişkin değişik tip ölçümlerle uygulanmış olan birçok örneği vardır. Bu çeşitli tomografi teknikleri ölçüm istasyonunun yerine yani kaynak ve alıcı derinlik alanna bağlı olarak belli bir durumda uygulanabilir (örneğin yerel deprem tomografisi, telesismik tomografi, global tomografi vb.).
Başlangıç olarak, üst manto yapısını tasvir etmek için İsviç Ulusal Sismolojik Ağı (SNSN)'ın kaydettiği 52 telesismik depremden kaydedilen bağıl ulaşma zamanları farklıını ters çözüm işleminine tabi tuttuk. Bu 52 telesismik depremi seçerken bu depremlerin sismik yaklaşım yapısı yanal halleri izin veriyor ve 680 kilometre kadar derinliklere kadar heterojenite gösterdi. Sismik ne tworkun boyutu daha derinlerde de artıyor ve yerlerde yaklaşıklık bir özelliğin olmak üzere radial (SV) ve tangential (SH) parçaları içindedir ve iki model oluşturuldu. 250-300 kilometreden daha küçük derinliklerde, modelin görünür kuzeye doğru dalan bir özellik sahip olduğu görülüyor.


Son kırk senede, sismik kayıtlar kayıtların sayısıındaki artış ve dalga formu verilerini analiz etmedeki hesaplama yöntemlerinde önemli ilerlemler oldu. Sismologlar şimdi yerin iç yapısının “yanal olarak heterojen” izotropik katmanlardan meydana geldiği gerçekini geçerli olmayını bilebiliyolar. Çeşitli frekans içeriğin sahip çeşitli dalga tipi gözlemleri (P, S, SKS, Yüzey dalgaları vb.) yerin iç yapısının önemli ölçüde anizotropik bir şekilde sahip olduğu sonucuna bizi götürüyor. Başka bir deyişle, sismik dalga hızı davranış yönüne bağlı olarak değişiklik gösteriyor. Sismik anizotropinin olmadığında var olup olmadığını anlayabilmek ve Baltık kalkanı altındaki olası yapıyı modellledebilmek için biz çalışma sırasında


Böylece, analiz sonuçları şu ortaya koydu ki makul büyüklikteki bir anizotropi izotropik varsayılara bağlı olarak elde edilen tomografik şekiller üzerinde oldukça büyük bir etki gösterebilir ve bu nedenle de ihmal edilmemelidir. Eğer, bizim bu çalışmada gözlemlediğimiz gibi, anizotropik tahminlerimiz makul ölçüde ise, o zaman düzeltilmiş veriler bize manto yapısı ile ilgili daha güvenilir ve doğru bir model vermiş olmalıyız.
Uppsala is a great university city and I met with many people-friends from different parts of the world during all the years of my stay in this cozy city. I have shared numerous unforgettable memories with them and now I do not want to miss any single one while writing my appreciation down in this chapter. Thus, now I think I am in the hardest part of this thesis.

To start with, I will not break common rule and first will thank Roland Roberts, who is my main supervisor from the beginning of my PhD. He is one who always believes that it is never late even if there are minutes to the end of any deadline. The first moment when I was Uppsala, I remember that I asked how I will find him if necessary. Then he replied me like this: “I will find you if it is necessary”. Himm, am I working with a kind of “mafia group”? I asked myself 😊 YES … He did. Whenever I needed him, he was always showing up. He was very satisfactory during our meetings to discuss on the progress of my studies. I always enjoyed while talking about scientific matters with him. I am thankful him since he has always shared his brilliant ideas, his never-ending support, beliefs and patience about the future of my study here. I just like to thank you Roland for your great mental tutorship and enormous help in making my bad writings 😊 always a masterwork (at least for me).

Then Roland introduced me Hossein Shomali. The first Persian I have ever met. By time we realized Turks and Persians have many things in common (food, language, music etc.). He helped me lot in getting used to this new city of my life. He had lot of assistances for me to find my way especially as a new person who started to be interested in “seismic tomography”. I also like to thank him for being an intimate friend of mine to listen my problems whenever I needed.

In some part of this thesis, I worked with my Czech colleagues. Without their valuable guidance some of the works would not be present right now in this thesis. First person who I like to give my deepest gratitude is Jaroslova Plomerová (Jarka 😊). You were always supportive, and insisting on teaching me about the theory of seismic anisotropy. I appreciate you for correcting my manuscripts (as your term “terrible”). Also I would like to thank you for providing an important amount of financial support during my visits to your beautiful city of Prague. Vladislav Babuška was always patient in sharing his boundless knowledge on seismic anisotropy. I would like to thank Ludek Vecsey since he was always bountiful in teaching me the technical details of
SKS splitting evaluations. Hanka was a good friend and also office-lunch mate in the institute.

Coming back to Uppsala again, I am here especially like to thank Reynir Bödvarsson about his financial support especially towards the end of my studies. Considering that Sweden is a very expensive country and without financial support “I am nothing here”, so his support is very laudable for me. Christoph Hieronymus played an important role in improving the discussion part of the Paper II with his valuable ideas. I would like to thank him for the Geodynamic course too. I am grateful to Chris Juhlin for his very valuable teaching efforts during a few courses. I would like to thank him for many coffee breaks with us. I am also grateful to Dan Dyrelius and Laust Pedersen for their teaching efforts.

Mattias Lindman, my first office-mate, enormously helped at the early times of my PhD. He helped a lot for solving my accommodation problem. He answered my stupid Matlab related questions. I think I cannot count his numerous favors here because of the limited space here but thanks a lot for everything. I will always wish the best for you and your wife San and now new born cute daughter. My second office-mate Sverker Olsson; I am grateful to you for your friendship, and also nice discussions about the upper mantle of Baltic Shield. Sverker are especially thanked also for taking care of the Swedish summary of this thesis. Sawasdee: my first Thai friend, one of the best football player I ever seen and my intimate at Flogsta. Thousands of thanks for your friendship. My special thanks are for Hossein, Palmi Erlendsson, Arnaud Pharasy, Björn Lund, Ari Tryggvason, Kristin Jonsdottir, Baldur and Jon, Eva Karlsson, Lijam Zemichael since all of them are always nice to me as a corridor mate in the department. I can never forget Palmi and Arnaud’s never-ending patience whenever I have a PC-related problem. They were always kind and always helpful. Arnaud is also one of my loyal ‘fika friends’. Palmi and Kristin, I also appreciate you guys for your kind dinner invitations a few times. I also like to thank you Lijam and Hossein for our early times meetings, for especially dinners at the downtowns. Those were really useful for me to get used to Uppsala. I cannot forget many funny moments that we had in those ‘hanging outs’.

During almost five years, I have shared many moments with many people; some of them are my fellows, in the department. I am sure I will forget now counting their names but for those who I forget somehow please forgive me. Anyway my deepest gratitude goes to Mattias, Lijam, Sawasdee, Thomas, Hossein, Sverker, Hasse, Arnaud, Björn Lund, Cedric, Zurab, Juan Diego, Maria, Abi, Ester, Frances, Imma, Peter Schmidt, Palmi, Artem, Zuzana, Ari, Howri, Faramarz, Peter Hodac, Jahiris, Aså Frisk, Elena, Nina, Alireza, Azita, Armita, Christoph Hieronymus, David Gee, Valentin Troll, Ota Kulhanek, Olafur, Thora, Conny, Leif Persson, Björn Bergman, Anna, Hemin Koyi, Nazli, Kristin, Juliane, Majid, Mehrdad, Lasse, Niklas Juhojuntti, Luis, Niklas Linde, Jochen, Hesam, Marie, Anna, Jarek, Alicia, Patric, Mimmi, Hen-
ning, Emil, Nauman, Saeid, Maxim, Björn Sundquist, Can Yang, Fengjiao Paiboon, Johiris, Michael. I would like to thank all of you guys and also to those who I forgot to mention here. Taher, Anna Callerholm, Siv, Ingegerd, Tomas Leif Nyberg, and Susanne Paul were always kind and patient to me in solving various technical and administrative issues.

Since the early times of my stay here I also have met with many Turkish friends who made me feel almost like ‘at home’ by many occasions with a full of warm smiles, nice jokes, delicious foods and drinks 😊 and cheerful chats and useful advices. Some of those friends living here permanently or temporarily: Fatma, Selcuk, Derya and Deniz (footballmania 😋) – thank you for always being ready for every kind of help making my lazy life continuous without any problem. You were like my family here and I cannot forget any single moment that we had especially in your cosy flat (parties, barbecues, relaxing amateurish musical trials). I will be always wishing a long living friendship for me and you. I had numerous funny times during my all meetings with the member of Uppsala-TR gangs: Arzu, Nazım, Mehmet Yıldız (not the one famous soccer player), Çağrıhan, Selçuk (Selçık 😊), Ali Temiz, Mert, Candan, Murat, Turgay Duman, Fatma teyze, Hamdi abi, Serpil-Somer-Tuna-Nil Bekiroğlu, Ö zgür, Esat-İlkınr Pehlivan, Vehbi abi, Türkcan ayla, Gülşen abla, his all family, Songül, Ece, Filiz, Habibe, Mustafa Aslan, Yasin, Celal abi, Demet, Gürkan, Cuneyt abi 😊, Armağan, Önder, İlke, Esra Bayoğlu Flener, Mark Can, Deniz, Pierre Flener, semi Turkish Fredrik (Fiko 😊), Arash, and Julia. Thank you for your friendship and always being around. You never made me feel alone. I am hoping for having many cheerful moments with many of you in future too. Here I like to appreciate Kuvvet Atakan for his hospitality and accepting me to a mini workshop in Bergen University as one of his colleagues. He was very kind by affording accommodation expenses during my stay at Bergen. I am grateful to Ege and Ulaş since they were helpful during my visit Oslo for the IGC conference. Mehmet Yıldız, Nazım and Esat are particularly thanked here since they were assisting me in technical issue in the last minutes of my writing this summary with lots of funny jokes. I really enjoyed lot.

Two out of my friends was always supportive and advised me never giving up even if the things go wrong in my life: Fatih and Birsen. Thank you for everything that you did and especially during my trips to Istanbul. Fatih, Necmiye (his wife) and little Timur were always friendly and we had lot of fun together in Berlin. I especially like to thank Fatih for many hours on the phone which I will always remember moments of our cheerful chats (lots of jokes). Thank you Fatih, my good friend for everything. Gonca Örgülü was another person who is always interested in my studies and supportive. I had always a positive motivation and encouragement people from my former institute, KOERI. So I would like to express my deepest gratitude to Niyazi Türkelli, Mustafa Aktar, Hayrullah Karabulut, Cemil Gürbüz, Bülent Tank, and all staffs of KOERI.
Whenever I was in Istanbul, I got enormous support from my friends Serkan-Ahu Metin, Tolga-Yuli Yüksekl and of course genious Erhan☺, Hatice, Özlem, Zafer and Murat. Mehmet-Eylem Kanik, Baran, Berken, and Turgay Bagdu are also appreciated since they never feel me alone either by phone or by visiting me. Bahar anne, Türkan and Nuray are especially thanked for their hospitalities and for accepting me as one of your sons. Our cheerful chats with Selahattin (kankam), Recep, and Berkay about our teenager ages were always nostalgic and nice.

I owe many things to my parents. Sevgili güzel anacığım, Sema Eken and babacığım Kemal Eken. Benim için yaptığınız her şey ve katlandığınız her zorluk için sonsuz kere teşekkürler ediyorum sizlere. My sister, Suna gave one of the biggest supports in many senses especially during the all years of mine in Uppsala. My uncle Mehmet Sinmaz was another person who always encouraged me to study. He always wished the best for me and my sister. Thank you for taking care of us when we really needed. Tülay, my LOVE and also my best friend always shared many things with me. Thank you for always listening to my problems and giving the best support, advises and your love. You never let me walked alone. Thank you for everything. I like to thank god also for keeping me alive during every bad situation that I faced with at work or at social life.

Finally I just apologize if I forget to mention some people by accident. For those if I really did, I would like to thank all of you too.

Tuna Eken
Uppsala, 3 May, 2009
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Acta Universitatis Upsaliensis

Digital Comprehensive Summaries of Uppsala Dissertations from the Faculty of Science and Technology 653

Editor: The Dean of the Faculty of Science and Technology

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