

Upper-mantle structure of the Baltic Shield below the Swedish National Seismological Network (SNSN) resolved by teleseismic tomography

Tuna Eken,¹ Z. Hossein Shomali,^{1,2} Roland Roberts¹ and Reynir Bödvarsson¹

¹Department of Earth Sciences, Uppsala University, SE-752 36 Uppsala, Sweden

²Geological Survey of Sweden, SE-751 28 Uppsala, Sweden. E-mail: Tuna.Eken@geo.uu.se

Accepted 2007 January 8. Received 2006 December 5; in original form 2006 August 30

SUMMARY

Upper-mantle structure under the Baltic Shield is studied using non-linear high resolution teleseismic *P*-phase tomography. Observed relative arrival-time residuals from 52 teleseismic earthquakes recorded by the Swedish National Seismological Network (SNSN) are inverted to delineate the structure of the upper mantle. The network consists of 47 (currently working) three-component broad-band stations located in an area about 450 km wide and 1450 km long. In order to reduce complications due to possible significant three-dimensionality of Earth structure, events chosen for this study lay close to in-line with the long-axis of the array ($\pm 30^\circ$). Results indicate *P*-wave velocity perturbations of ± 3 per cent down to at least 470 km below the network. The size of the array allows inversion for structures even at greater depths, and lateral variations of velocity at depths of up to 680 km appear to be resolved. Below the central part of the array (60° – 64° N), where ray coverage is best, the data reveals a large region of relatively low velocity at depths of over about 300 km. At depths less than about 250–300 km, the models include a number of features, including an apparent slab-like structure dipping gently towards the north.

Key words: ACH method, Baltic Shield, lithosphere, teleseismic tomography.

1 INTRODUCTION AND TECTONIC SETTING

For a better understanding of the evolution of Baltic Shield, robust imaging of the seismic velocities in the upper mantle is required to help constrain geodynamical models. The Baltic Shield is the exposed part of the Fennoscandian Shield which constitutes the northwestern part of the East European Craton and is composed of Precambrian crystalline rocks (Fig. 1). The East European Craton includes a range of lithospheric provinces from Archaean to late Proterozoic (Artemieva 2003). The evolution of the Fennoscandian Shield can be described in a few stages (Gorbatshev & Bogdanova 1993): First the formation of the Karelian in the Archaean province over 3.5 Ga. Then in the northeast part of the shield area the formation of the late Archaean–Paleoproterozoic Lapland Kola mobile belt. The high topography of such Archaean–early Proterozoic cratons is interpreted as implying depleted, low density underlying lithospheric mantle (Abbott *et al.* 1997). Semi-simultaneously, there was accretion to the southwest of the Proterozoic Svecofennian domain. To the west, the Svecofennian Orogenic zone is delineated by the ~ 1.8 – 1.65 Ga Transcandinavian Igneous Belt. The final stage of crustal shield formation occurred by accretion to western Scandinavia between ~ 1.75 and 1.5 Ga ago. A major part of Sweden lies in the Svecofennian part of the shield. Nironen *et al.* (2002) explain

the evolution of the Svecofennian domain by five subsequent orogenies alternating with extensional activity between 1.92 and 1.79 Ga.

Crustal structures of the Baltic Shield have been extensively investigated in several controlled source seismic studies including Guggisberg *et al.* (1991); Kinck *et al.* (1993); Balling (2000); Lund & Smirnov (2001); Abramovitz *et al.* (2002); Juhlin *et al.* (2002) and Korja & Heikkinen (2005). Two large-scale projects in the region, namely FENNOLORA (a seismic refraction experiment, 1979) and BABEL (a wide-angle seismic experiment, 1989) are especially important in earlier studies of the large-scale structures. The structure of upper mantle in the Baltic Shield is an interesting target for deep seismic studies (i.e. Calcagnile 1982; Gee & Zeyen 1996; Perchuc & Thybo 1996; Thybo & Perchuc 1997; Plomerova *et al.* 2002a; Shomali *et al.* 2002; Bruneton *et al.* 2004; Sandoval *et al.* 2004; Shomali *et al.* 2006) partly due to the presence of at least two potentially different lithospheres in the shield (the Archaean Karelia and Proterozoic Svecofennian), which has been tectonically stable since 1.2 Ga. Another noteworthy geodynamic phenomenon in the area is the glaciation and deglaciation related to the Late Pleistocene glacial cycle that is responsible for large scale redistributions of ice and water on the Earth's surface (Kaufmann *et al.* 2000). Current crustal deformation in Fennoscandia is dominated by postglacial isostatic adjustment (Milne *et al.* 2001). Ongoing postglacial

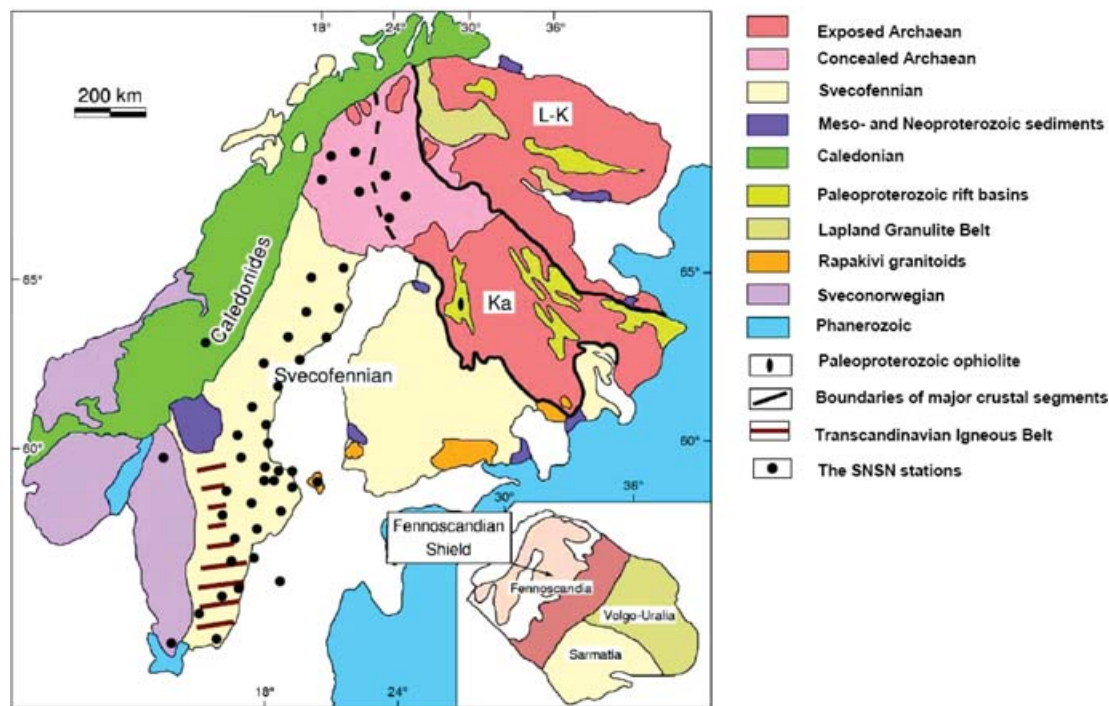


Figure 1. Tectonic sketch map of the Fennoscandian Shield (modified from Gee & Zeyen 1996). The SNSN station distribution in the SNSN network is also shown (dots).

rebound results in a maximum uplift rate of about 9 mm per year over the Bothnian Sea (Ekman 1996), and may be partly responsible for recent and ongoing seismicity in Fennoscandia (Wu *et al.* 1999). The recent analysis of recent postglacial rebound rates of Milne *et al.* (2001) estimates of the elastic thickness of the lithosphere beneath the shield to be in the range of 90–170 km, depending on the average viscosity (5×10^{20} to 1×10^{21} pascal seconds) of the upper mantle.

Recently, the lithospheric structures of the Baltic Shield and its adjacent regions have been studied during the EUROPROBE program (Gee & Zeyen 1996). To the south of the shield, the TOR experiment (1996–1997) investigated the deep lithosphere–asthenosphere structure beneath the Trans European Suture Zone (TESZ) from northern Germany to the southern part of Sweden. According to the model obtained, thinner lithosphere exists under northern Germany (of the order of 100 km) compared to southern Sweden (over 200 km) (Arlitt *et al.* 1999; Cotte *et al.* 2002; Plomerova *et al.* 2002b; Shomali *et al.* 2002, 2006). Another passive experiment, SVEKALAPKO (1998–1999), was designed to image *P* and *S*-wave velocity variations across the Archaean–Proterozoic tectonic suture in Finland. The results of a regional teleseismic tomography study (Sandoval *et al.* 2004) based on the SVEKALAPKO data set revealed a significant high-velocity zone in the upper mantle below the central part of the array down to ~300 km. No sharp velocity transition between the Archaean and Proterozoic areas is indicated in that study. Surface wave studies based on the same data set (Bruneton *et al.* 2004) also observe shear velocity perturbations up to ± 3 per cent. Such variations can be attributed predominantly to compositional effects since the thermal variations are considered to be small beneath Finland.

There is a geographic gap of several hundred kilometres between the TOR and SVEKALAPKO arrays (Fig. 1). Part of this gap is now covered by stations of the Swedish National Seismological Network (SNSN), but offshore between Sweden and Finland there is still relatively little data. In the future, this remaining gap will be filled largely using land-stations on some of the many islands in the

area. Because of the data gap, our current work focuses only on the Swedish side of the Baltic Shield. We invert *P*-wave arrival time data to elucidate structures primarily in the depth range 70–680 km, that is, in the upper mantle. The station spacing, which varies between 30 and 100 km, means that crustal structures cannot be resolved using teleseismic tomography. Previous studies (see e.g. Shomali *et al.* 2002) suggest that crustal structure may have a derogatory effect on the final upper-mantle models. To reduce the risk of possible bias and loss of resolution, observed relative arrival-time residuals were corrected for the effects of crustal structure using a crustal model based on data from various previous studies (see e.g. Korsman *et al.* 1999).

2 DATA SELECTION

Almost continuous seismological recordings from Sweden exist for over 100 yr, but the number of observation points was limited. The current Swedish National Seismological Network (SNSN) has been operated by Uppsala University since 1998 (Bödvarsson 1999). It is the one of the most modern broadband seismic networks in the world, currently operated 47 three-component seismic stations (soon to be about 60), which are equipped with Guralp CMG-3TD sensors with flat velocity response in the period range from 0.02 to 30 s. Most SNSN stations lie on the Svecofennian part of the shield that is presumed to have a Proterozoic lithosphere, exceptions being seven stations in the north which lie in the Archaean area (Fig. 1). 52 high quality teleseismic earthquakes were selected for the current study (Fig. 2). We selected teleseismic earthquakes with magnitudes larger than 5.5 and epicentral distances between 30° and 90° based on the event catalog reported by USGS. The hypocentres of the events were corrected as reported by Engdahl *et al.* (1998). Since the geometrical configuration of the SNSN array is rather long and narrow, covering an area about 450 km wide and 1450 km long, the great circle paths

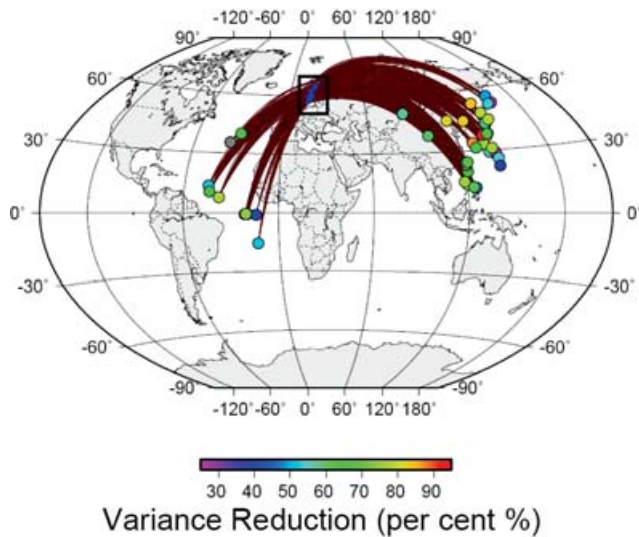


Figure 2. Earthquake distribution of the events used. Events are color coded to show the corresponding variance reduction. The SNSN network is shown in blue within the black rectangle.

of all of the earthquakes chosen were approximately in line with the long axis of the SNSN array ($\pm 30^\circ$) in order to alleviate possible complications related to significant three-dimensionality (see Fig. 2). Thus our data set can be mainly divided in two groups. The events in the first group are mainly in the vicinity of Japan where activity is dominated by the ongoing subduction. These signals approach the SNSN from the north with relatively high signal-to-noise ratio (Fig. 3a). The second group of earthquakes in our data set is from the Atlantic Ocean. These signals approach the array from the southwest and are characterized by shallow (around 10 km) sources and poorer signal-to-noise ratio. The geological conditions in Sweden, with sedimentary cover often thin or completely absent, means that very high signal-to-noise ratios are normal. For instance the minimum signal-to-noise ratio for the events used in this study which approach the array from northeast is around 30 dB while it is about 7 dB for those approaching from the southwest. During data preparation, all seismograms were restituted to mimic a short period the World Wide Standardized Seismographic Network station (WWSSN) with a dominant frequency of 1 Hz (Oliver & Murphy 1971). Data with a very low signal-to-noise ratio were rejected from our data set. Only high quality data recorded by many stations along the array were kept for the further processing. 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 (Fig. 3b). This procedure is useful in avoiding cycle skipping (Evans & 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. While phase picking, a weight is assigned to each pick. The assigned weighting is actually a measure of how well the observation of each phase is matched with others for a given earthquake. Theoretical arrival-time data is calculated for each event and station based on the IASP91 traveltimes model (Kennett & Engdahl 1991). Basically, our final rel-

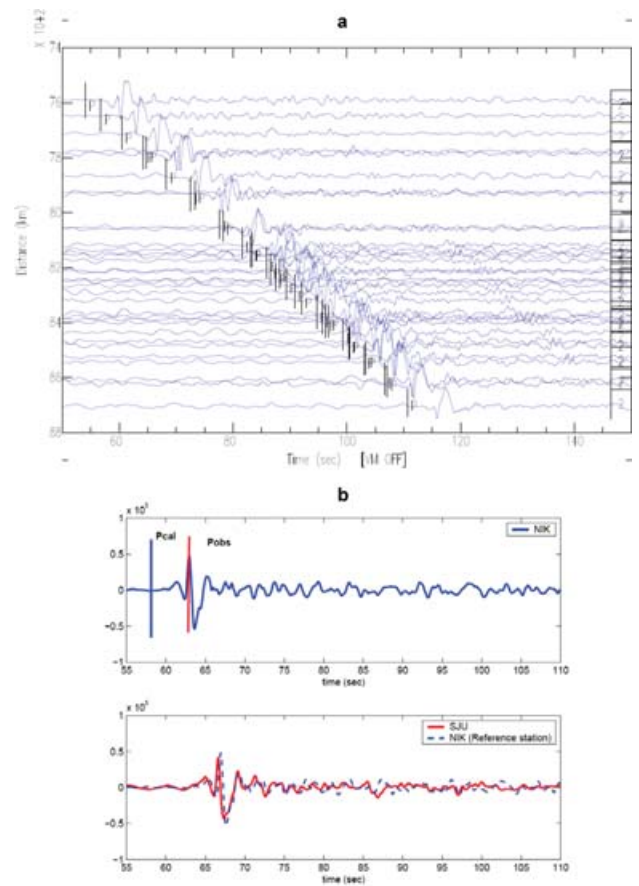


Figure 3. (a) An example of the data record section for the event of 20031112-0826 occurred at 33.189°N and 137.044°E with a magnitude of 6.4 Mw and an epicentral distance of 74° (as great circle arc length) to the centre of the network. The P arrival-times are computed based on the IASP91 traveltimes 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).

ative 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. 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 & 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 & 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 & 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 averaged earth model. Although during the forward modelling 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. Difficulties may arise if the starting model is inadequate. For example, due to the large size of the array, an incorrect initial (1-D) model could mean that the traveltime residuals were incorrectly calculated, but in practice for our configuration such problems are of minor significance. In the current study, we used the IASP91 traveltime model (Kennett & Engdahl 1991) as the starting model in the inversion. SAC2000 was used for signal processing (Goldstein *et al.* 2001) and phase picking was carried out using the Seismic Handler software (Stammler 1993).

3 THEORY

In teleseismic tomography the spatial-extension and magnitude of velocity anomalies are estimated by back-projection of relative arrival-time residuals (Weiland *et al.* 1995). The system is described by the following equation,

$$T = \int_S \frac{dS}{V(r)}, \quad (1)$$

where T and S represent traveltime and ray path, respectively. Both of them depend on the velocity of the media, $V(r)$. 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. While ray tracing is done based on the background model, the final output model of inversion reflects velocity deviations with respect to a background model which remains unknown. Once the system is linearized, various different inversion methodologies can be employed (Menke 1989). Here, as a first attempt we used a weighted damped least squares method to solve the linearized system:

$$X^{\text{est}} = (A^T W^b A + \varepsilon^2 W^X)^{-1} A W^b b, \quad (2)$$

where X^{est} is the model vector that includes velocity perturbations, A is a matrix of partial derivatives of traveltime with respect to unknown velocity parameters, and b is a vector which includes our observed data, namely relative arrival-time residuals. In this equation, W^b stands for data error, determined during the data picking and W^X is a smoothing operator. In practice, W^b also executes the removal of the arrival means and the source effects (Weiland *et al.* 1995). 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 traveltime with respect to model parameters. W^X stands for solution roughness, ε^2 is the damping factor which ensures the stability of the inversion. W^b and ε^2 are defined *a priori* to the inversion. We used singular value decomposition (SVD) to calculate the inversion of the matrix given in eq. (2). Small eigenvalues which may have poor information about the structure and may cause some instability in the inversion were discarded from the calculation. In choosing smoothing parameters, such as solution roughness, damping and eigenvalue truncation level, we must consider that there is a trade-off between model resolution and variance. The smoothing parameters are chosen based on considerable numerical testing, in

order to produce models which are considered to have as high spatial resolution as possible from the given data set without leading to undesirably large variance of the velocity estimates.

4 CRUSTAL CORRECTION

The main purpose of this study is to investigate large-scale upper-mantle structures. Traveltime perturbations due to crustal structures may be significant relative to the effects of mantle structures and may, therefore, disturb the results. Moreover, station spacing and the lateral horizontal grid spacing (75×75 km) used in this study are too large to be useful in resolving crustal structures using traveltime tomography. Due to glaciation, significant sedimentary cover is absent below most of the array. However, the variations in the structure of the crystalline crust, including Moho depth, still generate significant arrival time perturbations. Thus for optimal inversion of the traveltime residuals it is desirable to correct teleseismic relative arrival-times residuals for the traveltime perturbations caused by the crust prior to inversion.

Most of the seismic reflection/refraction experiments of the Baltic Shield and surrounding area are compiled by Korsman *et al.* (1999). The compiled dataset consists of BABEL-2, BABEL-6, BALTIC-BLACK SEA, BALTIC SEA, BARENTS SEA, BLUE ROAD, FENNIA, NW POLAND, RUBIN and SYLE-PORVOO profiles and data from the Bothnian Sea and Lake Ladoga—Lake Onega regions. We used these datasets to construct a 3-D crustal model for the studied area. The input parameters were Moho depths, P -velocity beneath the Moho and depth to the upper-lower crust boundary at different discrete points. We used mean P -wave velocities of the order of 6.6 and 7.3 km s⁻¹ for the upper and lower crust, respectively (Guggisberg *et al.* 1991). A linear interpolation scheme was then implemented to construct a 3-D crustal model. Traveltime variations due to the crustal model were then calculated for each event and seismic station based on a simplex-based ray tracer algorithm (Steck & Prothero 1991). These were used to correct each observed relative arrival time residual. A cross-section along the SNSN profile of the crustal model is depicted in Fig. 4. The depth of the Moho varies greatly along the profile (changes approximately between 35 and 50 km). Particularly significant changes in the Moho depth are observed in the southern part of profile where variations between around 35 and 50 km exist over a distance of only 200 km. A teleseismic receiver function study is ongoing (Olsson *et al.* 2004) and even preliminary results from this analysis have been considered in constructing the crustal model.

We can expect the crustal correction to improve our results for the deeper structure, under the assumption that the crustal model we use is reliable. Data quantity is limited and in most cases the structure beneath a given station requires interpolation between robustly estimated data from the previous studies. There could, therefore, be a risk of introducing some bias into our results if some errors or inadequacies exist in the crustal model. Test show that inverting without using crustal corrections but instead allowing the inversion to estimate even the near surface structure does not fundamentally alter the major structures observed at depth, but does slightly degrade resolution, as we would expect if our crustal model is reliable. We conclude that it is advisable to apply the crustal corrections, and that future results might be further improved when more information about crustal structure becomes available.

5 MODEL PARAMETRIZATION

Inter station spacing of the SNSN varies from 30 to 100 km. It becomes sparser towards the northern part of the network. In general

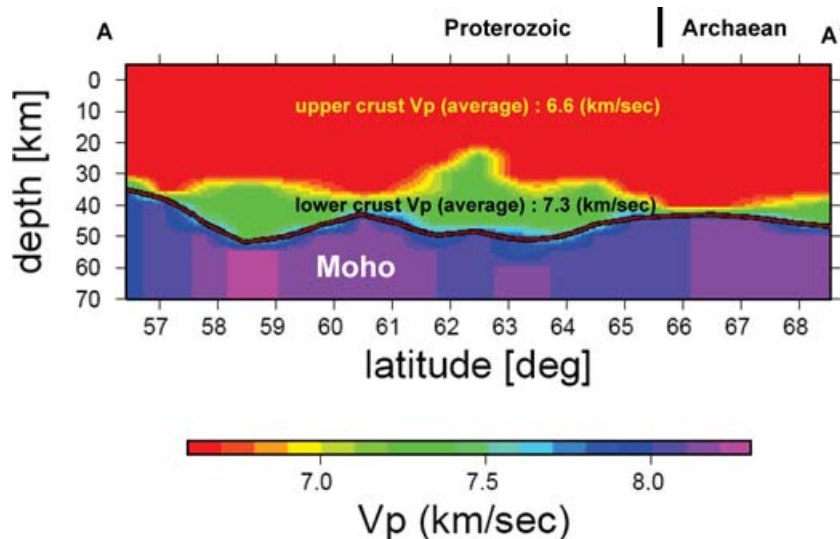


Figure 4. A cross-section along AA' (Fig. 5) of our 3-D P -velocity crustal model.

three major characteristics of the data set affect the resolution of the inversion results: (1) dominant station spacing, (2) frequency content of the data and (3) ray geometry. Additionally, we must choose a suitable balance between resolution and variance in the results. These considerations affect suitable design for our inversion procedure. For example, the ratio between horizontal and vertical grid spacing has an effect on the ray coverage of the cell blocks (Arlitt *et al.* 1999). Based on the station spacing and a number of numerical tests, a lateral grid size of order of 75×75 km was found to be suitable for most of the model. The model and inversion are fully 3-D, but the extended form of the array, which is believed to lie with its major axis largely perpendicular to the 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 adapted to be 100×100 km. This imposes a spatial smoothing effect since 75 km is larger than least resolvable structure defined by roughly average 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 Fig. 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 (30–70 km) from the surface down to a depth of 470 km (Fig. 5). Many tests were conducted, including using larger horizontal grid spacing such as 100×100 km, in order to investigate possible effects of grid spacing. The resulting models were in a good agreement and show close similarities in terms of resolving major blocks and transitions. The results of the inversion with grid spacing 100×100 and 75×75 km are given in Table 1. The model parametrization (75×75 km) includes a vertically changing block size (30–70 km) with 1287 nodes and 3553 associated model parameters in total. The size of the SNSN array means that the Earth's curvature (Fig. 5) 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 & 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 & Prothero 1991; Weiland *et al.* 1995).

6 INVERSION

Due to the nature of observed data sets and 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. Earlier studies (i.e. Shomali *et al.* 2002; Lippitsch *et al.* 2003; Sandoval *et al.* 2004) using the teleseismic algorithm indicate that the optimal solution of 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 insignificant eigenvalues that probably do not contain useful information about Earth structure (i.e. those whose magnitude was below a chosen threshold level). For selection of the optimum truncation level a set of inversions with different truncation levels was conducted. Because the threshold is based on the magnitude of the eigenvalues, a different number of eigenvalues may be found to be acceptable for different iterations in the same inversion. In cases where the number of accepted eigenvalues changed significantly from iteration to iteration, it was considered that there was a risk that the inversion was unstable due to the inclusion of eigenvalues which were too insignificant, and this threshold level was then considered to small. We therefore required a truncation level which did not lead to such instability. We finally chose a truncation limit of 150. The stability of this value was later confirmed by good correlation of the models obtained by SVD and quadratic programming, a completely different inversion approach (see below). Another important inversion parameter is damping. It controls the amount of minimization of the data misfit with respect to the model roughness and stabilizes the inversion. Assessment of the optimal damping parameter was made based on inversions using a set of various damping values between 5 and $1000 \text{ s}^2 \text{ per cent}^{-2}$ and constructing trade-off curves between data variance and model length (normalized squared summation of the model perturbations) computed for different damping values (Fig. 6). A damping level of $100 \text{ s}^2 \text{ per cent}^{-2}$ was chosen, as this appeared to correspond to a change in gradient in the curve. We also show below that our selected damping is also a reasonable

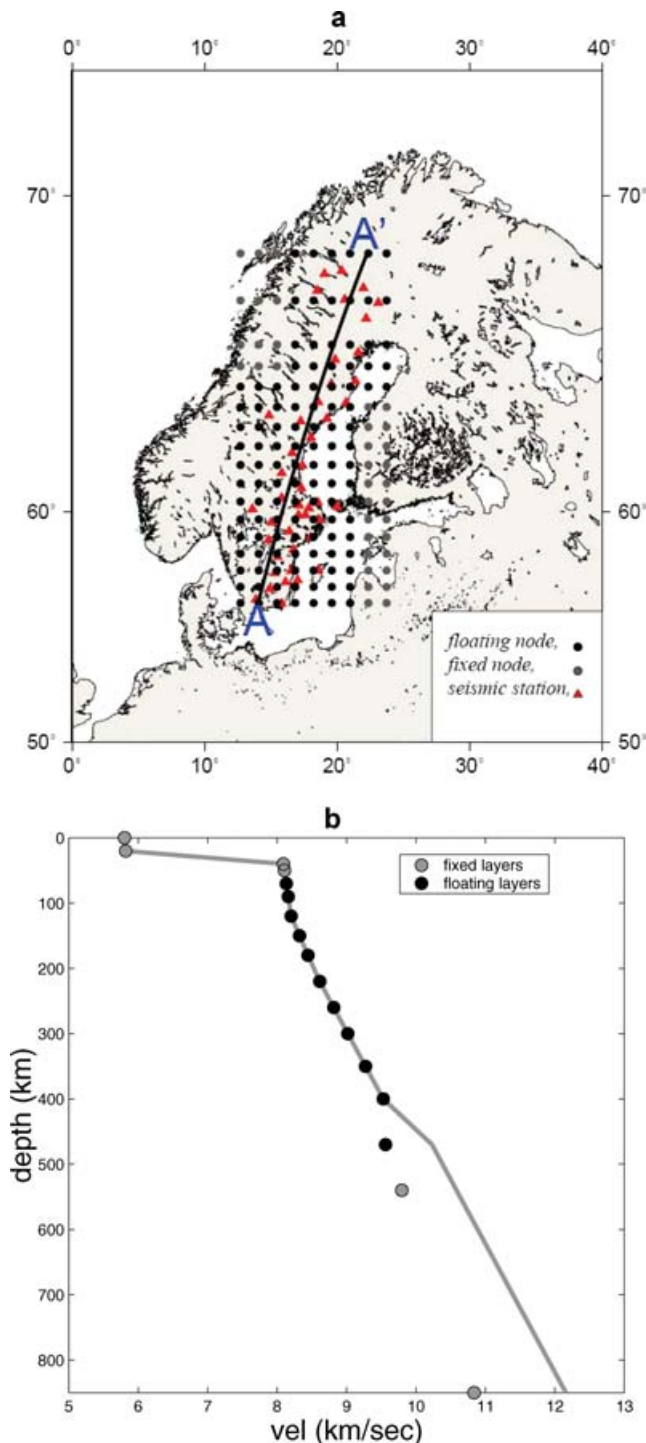


Figure 5. Model parametrization (a) The horizontal grid spacing and node configuration used. The black solid line is the profile AA'. (b) Layers used in the inversion which allows lateral velocity variations to 470 km. The velocities of the starting model nodes are taken from the IASP91 model, adjusted using an Earth Flattening Transformation.

choice in the recovery of a hypothetical structure containing low and high velocity anomalies (-3.0 per cent to $+3.0$ per cent) consistent with the observed anomalies. In Fig. 7 we present three images. The first two figures show the inversion results of uncorrected and corrected relative arrival-time residuals based on the crustal structure with our optimum damping value of $100 \text{ s}^2 \text{ per cent}^{-2}$. The third model applies a crustal correction and a larger damping factor.

The results shown represent estimated velocity perturbations along a profile aligned with the major axis of the SNSN array produced using the IASP91 traveltime 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. Major differences are observed on the southern part of the network where there is a significant change in Moho-depth (see Fig. 4). The inversion without crustal correction (Fig. 7a) includes small scale negative (low velocity) anomalies extending down to the depth 250–300 km between the latitudes of $\sim 56.5^\circ$ and 58° N. Applying crustal corrections, a similar low velocity feature is observed, but is now larger in size and amplitude and extends deeper. Thus it seems that the crustal correction does enhance our results, at least in this area, and we base our interpretations on the corrected data. The results of the model obtained from inversions with and without crustal corrections can be compared through Table 1.

The structures resolved at the central part of the model (between latitudes about 60° – 64° N) given in Fig. 7(b) 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 due to two or more spatially separate anomalies, and we should be cautious about interpreting the image as, for example, 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 100 km 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. Poor ray coverage in this region means that resolution is limited and it is not yet possible to completely exclude the possibility that the feature is an artifact. Resolution below the northern part of the array is limited by the relatively larger station spacing (in order of 100 km), and our different resolution tests (discussed later) indicate that this area suffers from significant vertical smearing. However, the dominant feature in this area is large, and it is likely that a significant lateral change in velocity occurs in this region, somewhere in the depth range 70–250 km. For comparison, Fig. 7(c) presents the model obtained using a damping factor of $300 \text{ s}^2 \text{ per cent}^{-2}$. In this overdamped image the observed structures are smoothed, but the dominant features of the images correlate, implying that our chosen damping is not introducing artifacts into the images.

The results of the inversion with the damping values of 100 and $300 \text{ s}^2 \text{ per cent}^{-2}$ are summarized in Table 1. One of the major questions while conducting the inversions is how deep the model should be. In fact determining a reasonable depth for the bottom of the target volume seems to be an enigma in many teleseismic tomography applications (Masson & Trampert 1997; Sandoval *et al.* 2004; Shomali *et al.* 2006). If the tomographic model is too shallow, there is a risk of importing velocity anomalies from outside (at the bottom of the model) into the target volume. On the other hand, making the model deeper will degrade results if, for example, the extra structure included is in reality homogeneous, not least because of the risk of vertical smearing (Weiland *et al.* 1995; Waldhauser *et al.* 2002). Thus, extending the model to greater depths may produce better results, but could also lead to poorer resolution in the model. The large size of our array (1450 km long) suggests that even very deep structures may be resolvable. The effect of model depth was tested by conducting a suite of inversions extending to depths from 300 to

Table 1. Inversion results.

Case	Horizontal grid spacing (km × km)	Inversion depth (km)	Crustal correction	Damping (s ² per cent ⁻²)	Perturbation model length ^{a,b} (km s ⁻¹) ²	Remaining data variance ^b (s) ²	Variance reduction ^b (per cent)
Case I	100 × 100	470	yes	100	0.0010	0.0478	63
Case II	75 × 75	470	yes	100	0.0058	0.1136	65
Case III	75 × 75	470	no	100	0.0068	0.1090	64
Case IV	75 × 75	470	yes	300	0.0040	0.1367	59
Case V	75 × 75	680	yes	100	0.0039	0.1459	58

Note: Centre of the model: 60.8060°N17.5493°E. Error associated with data (s): 0.1. No. of events: 52. No. of seismic stations: 45. No. of observations (P): 1565. No. of model parameters: 1287.

^aPerturbation model length is normalized to the number of corresponding singular value used in the inversion.

^bData are based on iteration no. 4.

680 km. The model inverted down to a depth of 470 km resulted in the highest reduction in the variance normalized by the degrees of freedom (Fig. 8, Table 1). A different way of assessing a suitable thickness of the inversion model is to examine the ray pattern. Such an analysis suggests that useful resolution may be attained even for the greater depths. Therefore, we present inversion result down to 680 km deep in Fig. 9. The large scale blocks-transitions resolved in the inversion, including deeper layers, are correlated well with model presented in Fig. 7(b).

7 RESOLUTION STUDY

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 the assessing the reliability of the images, and resolution analysis was an important part of our study. There are several methods of the assessment of the resolution capabilities of tomographic data. One way is to perform synthetic tests. We investigated a number of synthetic models, including checkerboards (i.e. Shomali *et al.* 2002; Lippitsch *et al.* 2003; Sandoval *et al.* 2004). Here we present the inversion of synthetic data associated with a hypothetical structure but based on the same ray geometry as the real data (Fig. 10a). Gaussianly distributed noise with standard deviation equal to 0.1 s 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 in Fig. 10(b) are, in general, recovered properly. 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.

Hit count and total ray length analyses, when considered together, are also useful tools in evaluating resolution. In Fig. 11, such data is presented for a profile parallel to the major axis of the array. As the rays spread laterally with increasing depth because their ray paths are not completely parallel to the major axis of the array, the number of rays in each cell decreases significantly at greater depth. The rapid near surface variations in hit count depend 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 images 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, that is, in the central part of the array where station density is greatest.

We also inverted the data using a quadratic programming method in order to assess the resolution and reliability of model obtained. Quadratic programming is a well-established method in solving inverse problems, 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, for example, 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 eq. (2), 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 eq. (2) but without any spatial smoothing operator, W^X 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 inversion to ± 3 per cent s⁻¹. 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). The model obtained based on quadratic programming is shown in Fig. 12. By comparing Figs 7(b) and 12 it can be concluded 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.

8 AN AVERAGED VELOCITY MODEL

The final inversion result shown in Fig. 7(b) represents velocity deviations nominally from the IASP91 starting model along the SNSN array. Our inversions suggest that a significant boundaries between major tectonic units exist, for example, between the Archaean and Proterozoic lithospheres in the north of the study area. As an alternative method for assessment of the possible reliability of these

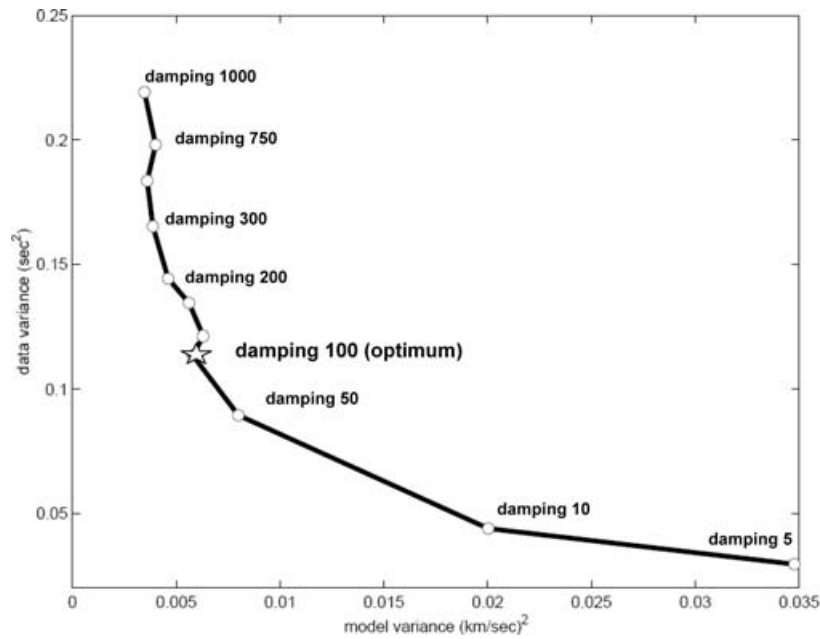


Figure 6. Trade-off curve for various damping factors at the 4th iteration. Our preferred data and model variance relation is given by the inversion with a damping of $100 \text{ s}^2 \text{ per cent}^{-2}$ (star).

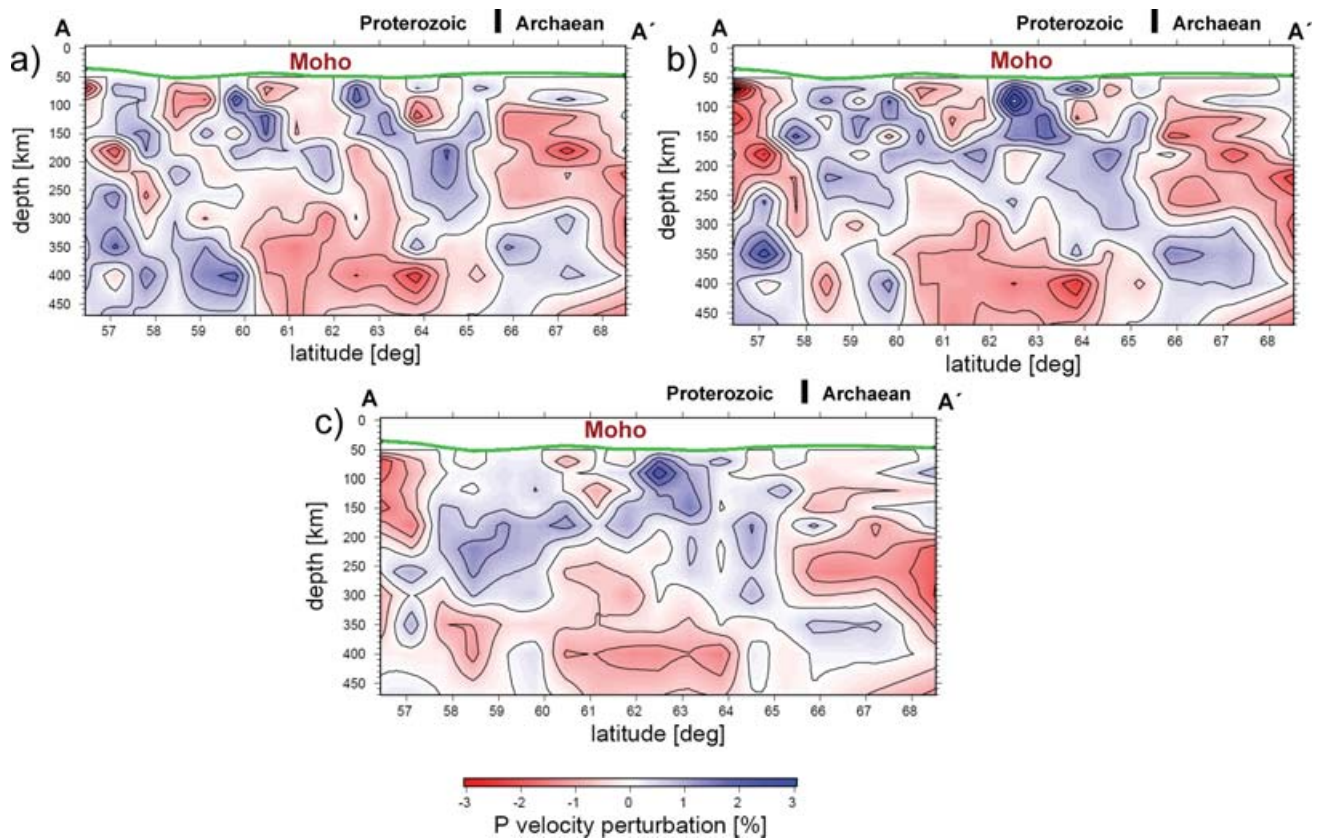


Figure 7. A cross-section of the P velocity perturbation along the profile AA' obtained from the SVD inversion. (a) Model obtained with a damping of $100 \text{ s}^2 \text{ per cent}^{-2}$ before crustal correction. (b) Model obtained with a damping of $100 \text{ s}^2 \text{ per cent}^{-2}$ after crustal correction. (c) Model obtained with a damping of $300 \text{ s}^2 \text{ per cent}^{-2}$ after crustal correction. All results are given with contour lines at an interval of 0.5 per cent.

features, we also investigate the vertically averaged velocity model. The velocity for each vertical column of cells in the final model was averaged, weighted to allow for the thickness of each cell. Essentially, this provides an image showing the hypothetical (model)

traveltime residual at each surface point for a vertically arriving ray. We choose to present the data in terms of velocity (essentially the inverse of traveltime residual) to allow direct comparison with the other velocity images (Fig. 13). Also shown on the figure are the

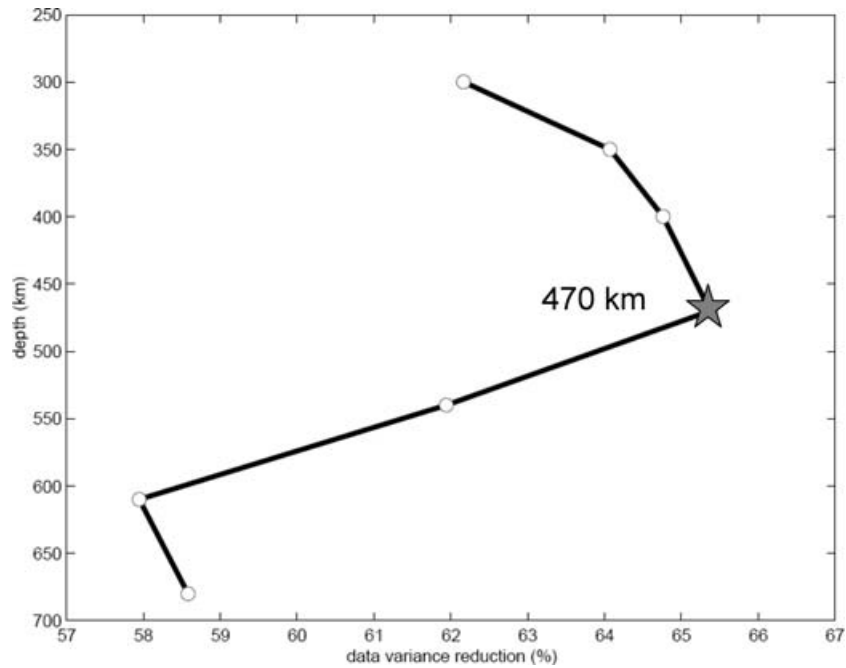


Figure 8. Data variance reduction (normalized using the effective number of degrees of freedom in the model) as a function of inversion depth at the 4th iteration. Inversion depths are 300, 350, 400, 470, 540, 610 and 680 km, respectively. The highest data variance reduction is achieved at 470 km (star).

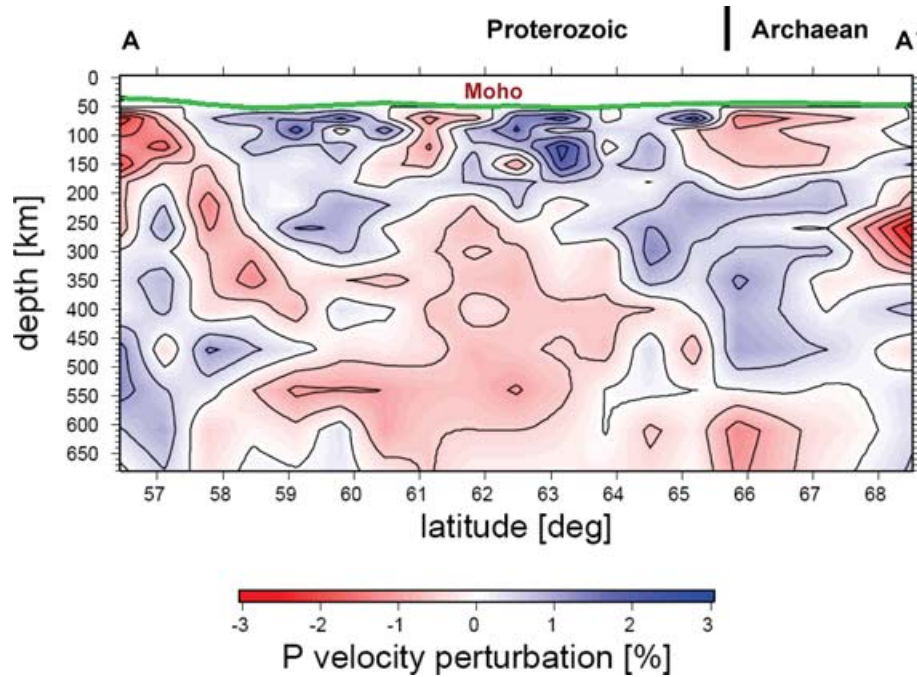


Figure 9. SVD inversion (four iterations) with variable velocity allowed to a depth of 680 km, crustal corrections applied, horizontal grid spacing of 75×75 km and damping factor $100 \text{ s}^2 \text{ per cent}^{-2}$.

averaged raw traveltimes residuals at each station. This presentation allows direct comparison of the raw data with a summary of the model, and confirms that there does appear to be a very significant change in residual time across the presumed Archaean–Proterozoic boundary. The magnitude of the effect is such that it appears untenable to associate it only with relatively shallow structure, implying that a significant velocity contrast at depths at least of the order of

200 km is required to explain the data, consistent with our inverted images (Fig. 7b).

9 DISCUSSION

Analysis of FENNOLORA long-range refraction data indicates significant lateral velocity changes in the upper mantle, which

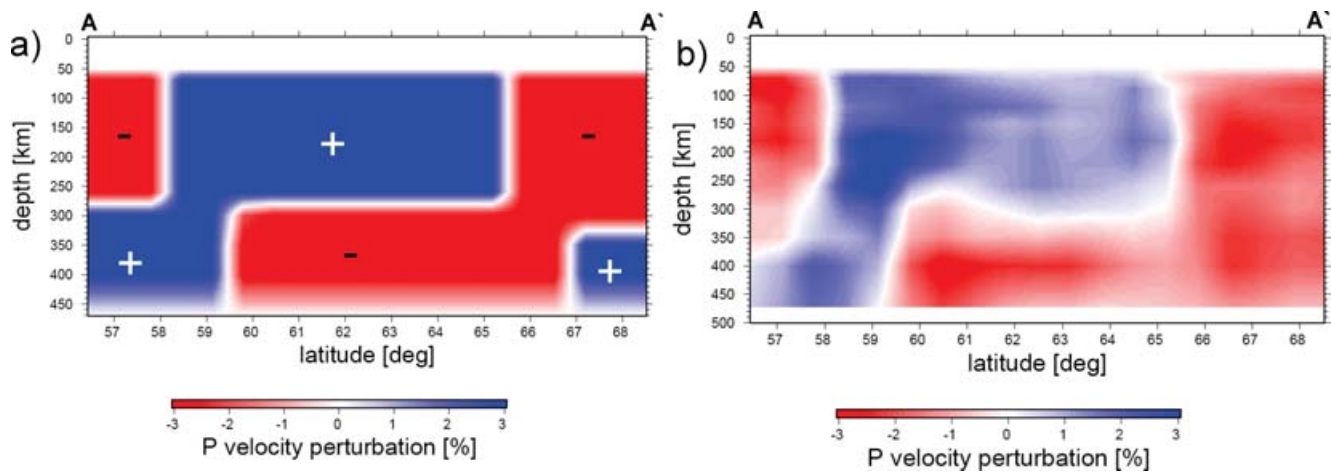


Figure 10. (a) A hypothetical synthetic model. (b) Resolved model at the fourth iteration. (+) and (–) signs show the areas with higher and lower velocities relative to the IASP91 background model.

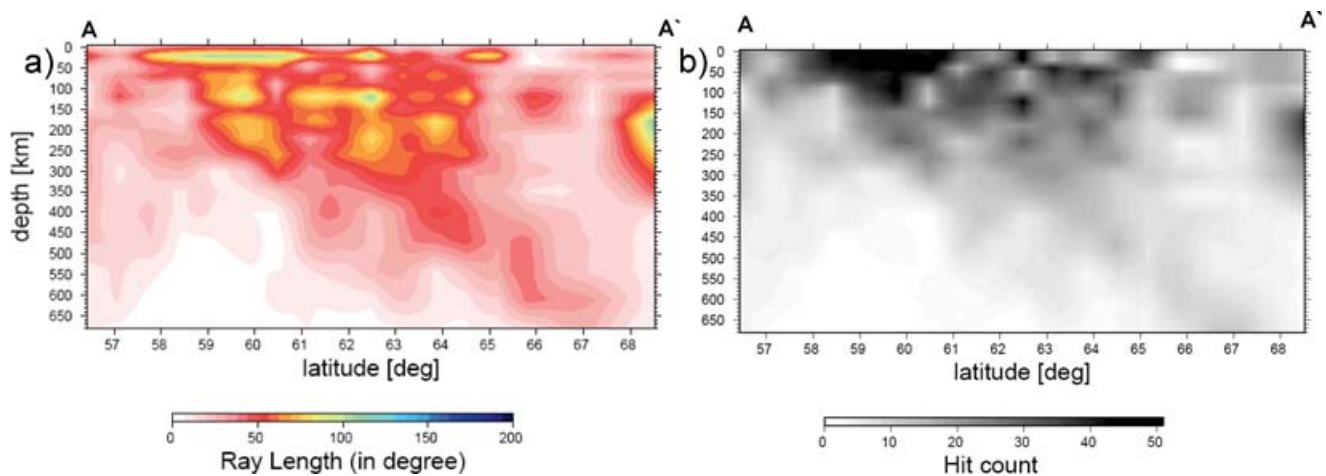


Figure 11. (a) A vertical cross-section of the total ray length in each cell along the profile AA' (b) A vertical cross-section of the total number of hits (number of rays) in each block along the profile.

includes several alternating high-low velocity layers (Guggisberg & Berthelsen & Berthelsen 1987). A lithosphere–asthenosphere boundary (LAB) was proposed at a depth of about 180–200 km by Guggisberg *et al.* (1991), who also reported the existence of a thin low velocity zone (LVZ) with a thickness of 50 km beneath this boundary. A similar LVZ in a stratified mantle, about 40 km thick and 100 km depth, is reported by Thybo & Perchuc (1997) based on long range refraction/reflection data. Their results are consistent with inversions of *P*- and *S*-wave traveltime data obtained from FENNOLOGRA (Abramovitz *et al.* 2002). On the other hand, Calcagnile (1991) presented a LVZ extending between 140 and 350 km beneath Sweden. Husebye & Ringdal (1978) found that a relative high velocity structure was dominant down to a depth of 250–300 km from *P*-wave residuals. Using the *P*- and *S*-wave seismic refraction data, a thickness of about 250 km was found for the Baltic Shield (Sacks *et al.* 1979). Correlation of the lithospheric thicknesses with heat flux suggests that the shield continues down to 200 km (Pollack & Chapman 1997). According to the SVEKALAPKO results (Sandoval *et al.* 2004) a well pronounced relative high velocity anomaly seems to be the most prominent feature down to 250 km in the centre of the array beneath the Finnish part of the Baltic Shield. They relate this positive anomaly to the

existence of a depleted tectosphere which was first proposed by Jordan (1978). Our final model obtained in this study is well correlated with both results (Husebye & Ringdal 1978; Sandoval *et al.* 2004) since it shows dominant relatively high velocities in the centre of the SNSN (at latitudes ~60–64° N) for depths less than ~300 km. Moreover, this higher velocity area modelled in the current study tends to significantly dip towards the northeast down to the bottom of the model.

In the Baltic Shield, the absence of a distinct LAB was proposed by Dost (1990) based on the fundamental and higher mode surface waves data from NARS array. In the southern part of the Baltic Shield, according to the results of the TOR project, a sharp contrast in the lithospheric thickness and structure under the Tornquist Zone was found (Arlitt *et al.* 1999; Gregersen *et al.* 1999; Cotte *et al.* 2002; Kissling 2002; Plomerova *et al.* 2002b; Shomali *et al.* 2002, 2006).

According to the teleseismic tomography studies along the TOR profile, the upper-mantle structure was mainly characterized by a relative high velocity zone down to 300–350 km beneath Sweden. *P*-phase teleseismic tomography results along the TOR array reveal that a significant transition zone between thin lithosphere in Germany and thick lithosphere in Baltic Shield (Southern part of

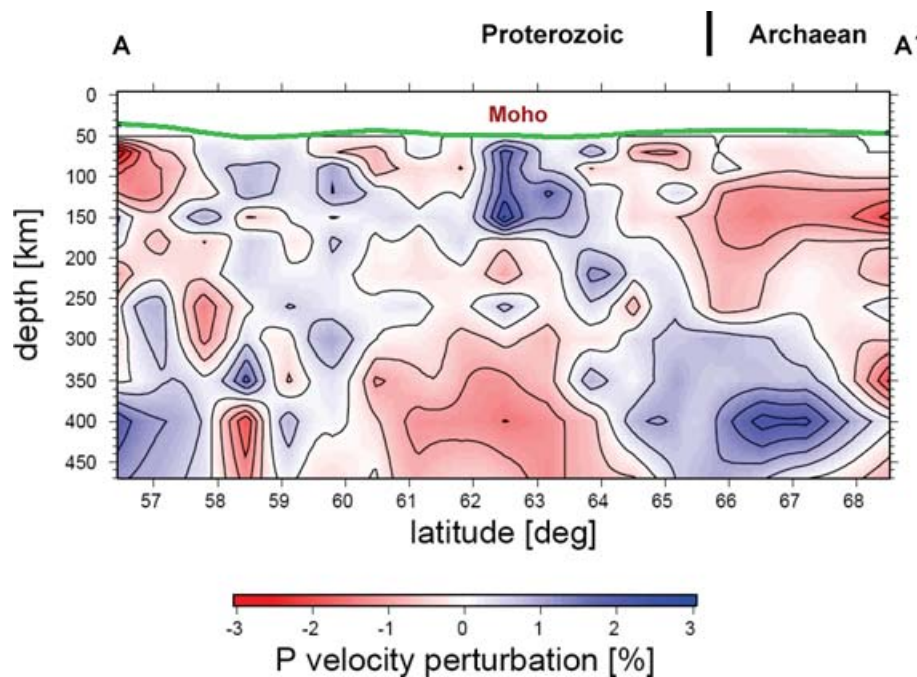


Figure 12. Inversion result from quadratic programming (four iterations) along the profile AA'. Crustal corrected relative-arrival time data. Horizontal grid spacing of 75×75 km. Damping $100 \text{ s}^2 \text{ per cent}^{-2}$.

Sweden) is not gradual but occurs in two very sharp and steep steps (Arlitt *et al.* 1999; Shomali *et al.* 2002; Kissling 2002; Shomali *et al.* 2006). The lithosphere beneath the northern Germany is thin but reaches to intermediate thicknesses of about 120 km in the Tornquist Zone area while a continental lithosphere extends to depth more than 200 km farther north in the Baltic Shield (Shomali *et al.* 2002). There is a common geographical part between tomographic models along the TOR and the SNSN data (between the latitudes of $\sim 56.5^\circ$ and 59.5° N). For this part of the model, the TOR analysis shows a predominant positive velocity anomaly, while the SNSN inversion dominated by negative relative velocities. This can be explained since the velocity perturbations are relative to different background models. However, the horizontal gradient of the velocity perturbations at this overlapping part of two models show a similar sharp transition between about 56.5° and 58.5° N for depths 70–200 km. In other words, where the models overlap they are consistent, with velocity perturbations increasing northwards.

Traveltime inversion results from the Kapvaal and Zimbabwe cratons (James *et al.* 2001) indicated an increase in *P*-wave velocity perturbations between -1 per cent to and 1 per cent down to 300 km. For the Slave craton in northwest Canada Bank *et al.* (2000) report a major fast slowness perturbation around $+1.0$ per cent down to 500 km beneath the craton. These results match well with the dominant positive velocity anomaly which was found to be located at the centre of our study area, except that the anomalies obtained in the current study (see Fig. 7b) have significantly larger velocity perturbations which may be explained based on the use of different inversion schemes which may have differences in resolution, parametrization and damping (Sandoval *et al.* 2004).

As discussed above, an important tectonic suture exists beneath the SNSN between the Proterozoic Svecofennian in the centre and the Archaean domain in the northeast. This composite craton uncovered by sediments and tectonically stable since 1.2 Ga, is an ideal place to test the differences between Archaean and Proterozoic

lithospheres (Bruneton *et al.* 2004). While isotropic seismic velocity studies of SVEKALAPKO data revealed no clear deep boundary between the Archaean and Proterozoic provinces (e.g. Bruneton *et al.* 2004; Sandoval *et al.* 2004), recently Plomerova *et al.* (2006) detected the possible transition between the Archaean and Proterozoic boundary by investigating the lateral variations of seismic anisotropy in the upper mantle. In our data, the suture between the Proterozoic and Archaean domains (see Fig. 1) is clearly seen in the form of a north dipping, relatively high velocity structure (e.g. Figs 7b and 9), and is also very clear on Fig. 13.

Generally speaking, the heterogeneity of the mantle in terms of seismic velocity variations can be explained by variations in temperature, composition or anisotropy (Goes *et al.* 2000). No major thermal event has been reported for the Baltic Shield during the last 1.5 Ga. As Bruneton *et al.* (2004) reported this time period is sufficient for thermal conduction to remove significant temperature differences between lithospheric blocks. That the upper mantle has a relatively homogeneous temperature is also supported by Kukkonen (1993) and Kukkonen & Jöeleht (1996) using heat flow and geochemical data and by Alinaghi *et al.* (2003) using receiver function for SVEKALAPKO data. Bruneton *et al.* (2004) concluded that as the thermal variations beneath Finland are negligible the observed lateral variation resolved in their shear wave tomographic model must be explained by compositional effects. The sharpness of the well resolved features in our tomographic model, with lateral velocity changes of several percent over some tens of kilometres (e.g. at latitude 57.5° , depth 70–200 km of Figs 7b, 9), would appear to require an explanation in terms of composition rather than temperature and the possible effect due to the anisotropy is discussed below. In fact, Kukkonen *et al.* (2003) observed that the seismic velocities which have been modelled by using the mineral compositions of the xenolith and mineral elastic parameters show excellent agreement with the velocity models (up to ± 2 per cent) obtained from the SVEKALAPKO tomography studies. Moreover, they found that rheological models based on the xenolith-derived

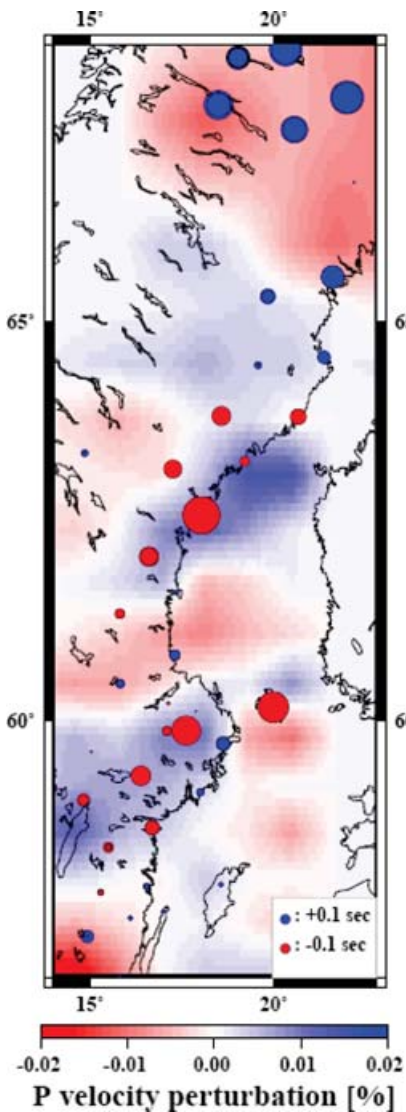


Figure 13. Map of the layer averaged velocity perturbations with the mean value of the observed relative arrival time residuals observed at each station. Observed mean residuals correlated well with the average velocity perturbations, that is, red circles (early arrivals) fit very well with blue contour regions, and vice versa.

geotherm and geophysical data do not indicate any partial melting within Fennoscandian mantle down to a depth of ~ 250 km.

The Baltic Shield is rapidly rising due to post glacial rebound (Kaufmann *et al.* 2000). The region where maximum uplift is observed is characterized by a relatively dominant positive velocity anomaly at least down to a depth of 200 km (between latitudes of 62° and 65° N). Moreover, in the map of the layer averaged velocity perturbations, the maximum uplift occurs above the part of mantle with relative higher velocities (Fig. 13). Uplift rate is normally, and plausibly, related to the pattern of loading (ice thickness), implying that any apparent relationship between ongoing uplift and the deep mantle is coincidental. However, as the rheology of the mantle and possible lateral variations in this clearly do affect rebound, there is perhaps a possibility that more complex mantle-rebound relationships exist than is normally presumed.

In the current study we have assumed an isotropic velocity structure. At mantle pressures observed directionally dependent veloci-

ties are usually believed to be related to the preferred orientations of mineral crystals. According to Babuska *et al.* (1998) the source of anisotropy in the upper mantle, different olivine crustal preferred orientations, may be caused by the different processes which result in the formation of the continental lithosphere in our region, for example, in the Archaean compared to the Proterozoic. Based on the analysis of *P*-wave residual arrival times, various authors have suggested significant *P*-wave anisotropy even at depth. For instance, Martynov *et al.* (2004) suggest *P*-wave velocity anisotropy of 2–3 per cent at depths of several hundred kilometres beneath Northern Tien Shan. It seems likely that genuine *P*-wave anisotropy within the mantle will be often be of rather limited magnitude. Any anisotropy which does exist may bias isotropic velocity estimates where waves sample predominantly in one direction (Goes *et al.* 2000). In southcentral part of Sweden, Värmland, Plomerova *et al.* (2001) found different orientations of dipping anisotropic structures within the subcrustal lithosphere blocks across the Precambrian Proterogene zone. Plomerova *et al.* (2002a) demonstrated anisotropy in the sublithospheric Baltic Shield mantle using SVEKALAPKO data. More recently, Vecsey *et al.* (2006) reported sharp spatial changes in the fast shear-wave polarization and in the time delay in the region. They derived a non-unique anisotropic model for the upper mantle beneath the region from SKS splitting analyses. However, Wylegalla *et al.* (1999) from SKS splitting, Babuska & Plomerova (2004) from the joint analysis of shear-wave splitting and directional dependence of *P*-wave residuals and Shomali *et al.* (2006) from first arrival-time residual analysis suggested that no significant anisotropy exists at the northernmost TOR station in Sweden. According to Wylegalla *et al.* (1999), the small SKS phase delay times between 0.2 and 0.5 s observed at the southern part of Sweden can be considered to be controlled by anisotropy in a thickened crust. Moreover, given the experimental configuration (i.e. station and event distribution) it seems likely that the velocity perturbations such as those produced by our inversion are not greatly perturbed by possible anisotropy structures within the model volume. All the rays used travel quite close to the vertical and in roughly the same plane, that is, in roughly the same direction relative to any anisotropy (see e.g. Shomali *et al.* 2006). However, if there is a strong anisotropy, interpretations of the model velocity as isotropic velocities may be misleading. The size of the observed anomalies suggests that the bulk changes in velocity are required by the data, even there may be some contribution related to the anisotropy.

For a further insight into the possible evaluation of the observed velocity images, it is highly desirable to have corresponding images derived from *S*-phase data. Another project related to *S*-phase inversion has been, recently, initiated and we are planning to publish its results in a separate paper. A combined inversion of TOR and SNSN data will reveal a consistent image of upper-mantle structure from northern Germany through the Baltic Shield, and is a clear ambition for further studies.

10 CONCLUSION

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 crustal model compiled using data from other studies of the area, primarily controlled source seismic data, was used to remove the crustal effect from the observed relative arrival time residuals. 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 by especially 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 Archaean lithospheric domains. This boundary separating the two different lithospheric domains is also resolved in a map view of the vertically layer 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.

In the southern part of profile, a near-vertical transition is resolved between latitudes of about 56.5° and 58.5° N and depths 70–250 km in both SNSN and TOR models. The sharpness of the well resolved features in our tomographic model with lateral velocity changes of several percent over some tens of kilometres would appear to require an explanation in terms of composition rather than temperature or possible effects due to anisotropy.

ACKNOWLEDGMENTS

During this study, the waveform data recorded at the SNSN were used. We wish to thank Mr. Sverker Olsson, Mr. Conny Holmquist, Mr. Hans Palm, Mr. Palmi Erlendsson and all other workers of the SNSN. Dr. Toivo Korja from University of Oulu, Finland is, especially, thanked for providing the Moho Depth data of the study area. We appreciate Dr. Jaroslava Plomerova and Dr. Annakaisa Korja for their valuable comments and suggestions.

REFERENCES

- Abbott, D.H., Drury, R. & Mooney, W.D., 1997. Continents as lithological icebergs: the importance of buoyant lithospheric roots. *Earth Planet. Sci. Lett.*, **149**, 15–27.
- Abramovitz, T., Thybo, H. & Perchuc, E., 2002. Tomographic inversion of seismic *P*- and *S*-wave velocities from the Baltic Shield based on FENNOLORA data. *Tectonophysics*, **358**, 151–174.
- Aki, K., Christofferson, A. & Husebye, E., 1977. Determination of the three-dimensional seismic structure of the lithosphere. *J. Geophys. Res.*, **82**, 277–296.
- Alinaghi, A., Bock, G., Kind, R., Hanka, W., Wylegalla, K., TOR. & SVEKALAPKO Groups, 2003. Receiver function analysis of the crust and upper mantle from the North German Basin to the Archaean Baltic Shield. *Geophys. J. Int.*, **155**, 641–652.
- Arlitt, R., Kissling, E., Ansorge, J. & TOR Working Group, 1999. 3-D crustal structure beneath the TOR array and effects on teleseismic wavefronts. *Tectonophysics*, **314**, 309–319.
- Artemieva, I.M., 2003. Lithospheric structure, composition, and thermal regime of the East European Craton: implications for the subsidence of the Russian platform. *Earth Planet. Sci. Lett.*, **213**, 431–446.
- Babuska, V., Montagner, J.-P., Plomerova, J. & Girardin, N., 1998. Age-dependent large-scale fabric of the mantle lithosphere as derived from surface-wave velocity anisotropy. *Pure. appl. Geophys.*, **151**, 257–280.
- Babuska, V. & Plomerova, J., 2004. Sorgenfrei-Tornquist Zone as the mantle edge of Baltica lithosphere: new evidence from three-dimensional seismic anisotropy. *Terra Nova*, **16**, 243–249.
- Balling, N., 2000. Deep seismic reflection evidence for ancient subduction and collision zones within the continental lithosphere of northwestern Europe. *Tectonophysics*, **329**, 269–300.
- Bank, C.G., Bostock, M.G., Ellis, R.M. & Cassidy, J.F., 2000. A reconnaissance teleseismic study of the upper mantle and transition zone beneath the Archaean Slave craton in NW Canada. *Tectonophysics*, **319**, 151–166.
- Bödvarsson, R., 1999. The new Swedish Seismic Network. *ORFEUS Newslett.*, **1**(3), 22.
- Bruneton, M., Pedersen, H.A., Farra, V., Arndt, N.T., Vacher, P. & the SVEKALAPKO Seismic Tomography Working Group, 2004. Complex lithospheric structure under the central Baltic Shield from surface wave tomography. *J. Geophys. Res.*, **109**, B10303.
- Calcagnile, G., 1982. The lithosphere asthenosphere systems in Fennoscandia. *Tectonophysics*, **90**, 19–35.
- Calcagnile, G., 1991. Deep structure of Fennoscandia from fundamental and higher mode dispersion of Rayleigh waves. *Tectonophysics*, **195**, 139–149.
- Cotte, N., Pedersen, H.A. & TOR Working Group, 2002. Sharp contrast in lithospheric structure across the Sorgenfrei-Tornquist Zone as inferred by Rayleigh wave analysis of TOR1 project data. *Tectonophysics*, **360**, 75–88.
- Dost, B., 1990. Upper mantle structure under western Europe from fundamental and higher mode surface waves using the NARS array. *Geophys. J. Int.*, **100**, 131–151.
- Dueker, K., Humphreys, E. & Biasi, G., 1993. Teleseismic imaging of the western United States upper mantle structure using the simultaneous iterative reconstruction technique, in *Seismic Tomography: Theory and Practice*, Chapter 11, pp. 265–298, eds H.M., Hirahara, K., Chapman and Hall, London.
- Ekman, M., 1996. A consistent map of the postglacial uplift of Fennoscandia. *Terra Nova*, **8**, 158–165.
- Engdahl, E.R., Van der Hilst, R.D. & Buland, R.P., 1998. Global teleseismic earthquake relocation with improved traveltimes and procedures for depth determination. *Bull. seism. Soc. Am.*, **88**, 722–743.
- Evans, J. & Achauer, U., 1993. Teleseismic velocity tomography using the ACH method: theory and application to continental-scale studies, in *Seismic Tomography: Theory and Practice*, pp. 319–360, eds Iyer, H.M. & Hirahara, K., Chapman & Hall, London.
- Gee, D.G. & Zeyen, H., 1996. *EUROPROBE 1996-Lithosphere Dynamics: Origin and Evolution of Continents*, p. 138, EUROPROBE Secretariat, Uppsala University, Uppsala.
- Goes, S., Govers, R. & Vacher, P., 2000. Shallow mantle temperatures under Europe from *P* and *S* wave tomography. *J. geophys. Res.*, **105**, 11 153–11 169.
- Goldstein, P., Dodge, D. & Firpo, M., 2001. SAC2000: Signal processing and analysis tools for seismologists and engineers, UCRL-JC-135963, Invited contribution to the IASPEI, *International Handbook of Earthquake and Engineering Seismology*.
- Gorbatshev, R. & Bogdanova, S., 1993. Frontiers in the Baltic shield. *Pre-cambrian Res.*, **64**, 3–21.
- Gregersen, S. & Tor Working Group, 1999. Important findings expected from Europe's largest seismic array. *EOS Trans.*, **80**, 1–2.
- Guggisberg, B. & Berthelsen, A., 1987. A two-dimensional velocity model for the lithosphere beneath the Baltic Shield and its possible tectonic significance. *Terra Cognita*, **7**, 631–638.
- Guggisberg, B., Kaminski, W. & Prodehl, C., 1991. Crustal structure of the Fennoscandian Shield: a traveltime interpretation of the long-range FENNOLORA seismic refraction profile, in *The European Geotraverse*, Part 7, eds R. Freeman, M., Huch, St. Mueller. *Tectonophysics*, **195**, 105–137.
- Husebye, E.S. & Ringdal, F., 1978. Seismic mapping of the Fennoscandian lithosphere and asthenosphere with special reference to the Oslo graben

- region, in *Tectonics and Geophysics of Continental Rifts*, pp. 297–311, eds Ramberg L.B., Neuman, E.-R., Reidel, Dordrecht.
- James, D.E., Fouch, M.J., VanDecar, J.C. & van der Lee, S., 2001. Tectospheric structure beneath southern Africa. *Geophys. Res. Lett.*, **28**, 2485–2488.
- Jordan, T.H., 1978. Composition and development of continental tectosphere, *Nature*, **274**, 544–548.
- Juhlin, C., Elming, S.-Å., Mellqvist, C., Öhlander, B.P., Weihed, P. & Wikström, A., 2002. Crustal reflectivity of the Archaean–Proterozoic boundary and comparison with BABEL lines 2 and 3, northern Sweden, *Geophys. J. Int.*, **150**, 180–197.
- Kaufmann, G., Wu, P. & Li, G., 2000. Glacial isostatic adjustment in Fennoscandia for a laterally heterogeneous earth, *Geophys. J. Int.*, **143**, 262–273.
- Kennett, B.L.N. & Engdahl, E.R., 1991. Traveltimes for global earthquake location and phase identification, *Geophys. J. Int.*, **105**, 429–465.
- Kinck, J.J., Husebye, E.S. & Larsson, F.R., 1993. The Moho depth distribution in Fennoscandia and the regional tectonic evolution from Archaean to Permian times, *Precambrian Res.*, **64**, 23–51.
- Kissling, E., 2002. Tomographic images of Europe—an overview. Keynote lecture, European Seismological Commission (ESC) 28th General Assembly.
- Korja, A. & Heikkinen, P., 2005. The accretionary Svecofennian orogeny—insight from the BABEL profiles. *Precambrian Res.*, **136**, 241–268.
- Korsman, K., Korja, T., Pajunen, M., Virransalo, P. & the GGT/SVEKA Working Group, 1999. The GGT/SVEKA transect—structure and evolution of the continental crust in the Palaeoproterozoic Svecofennian Orogen in Finland, *Int. Geol. Rev.*, **41**, 287–333.
- Kukkonen, I.T., 1993. Heat-flow map of northern and central parts of the Fennoscandian Shield based on geochemical surveys of heat producing elements, *Tectonophysics*, **225**, 3–13.
- Kukkonen, I.T. & Jöeleht, A., 1996. Geothermal modelling of the lithosphere in the central Baltic Shield and its southern slope, *Tectonophysics*, **255**, 24–45.
- Kukkonen, I.T., Kinnunen, K.A. & Peltonen, P., 2003. Mantle xenoliths and thick lithosphere in the Fennoscandian Shield, *Phys. Chem. Earth*, **28**, 349–360.
- Lippitsch, R., Kissling, E. & Ansorge, J., 2003. Upper mantle structure beneath the Alpine orogen from high-resolution teleseismic tomography, *J. geophys. Res.*, **108**, 2376. doi:10.1029/2002JB002016.
- Lund, C.-E. & Smirnov, A., 2001. A seismic model of the Precambrian crust along the coast of southeastern Sweden: The Coast Profile wide-angle airgun experiment and the southern part of Fennolara revisited, *Tectonophysics*, **339**, 93–111.
- Martynov, V.G., Vernon, F.L., Kilb, D.L. & Roecker S.W., 2004. Directional Variations in TravelTime Residuals of Teleseismic P Waves in the Crust and Mantle beneath Northern Tien Shan, *Bull. seism. Soc. Am.*, **94**, 650–664.
- Masson, F. & Trampert, J., 1997. On ACH, or how reliable is regional teleseismic delay time tomography?, *Phys. Earth planet. Inter.*, **102**, 21–32.
- Menke, W., 1989. *Geophysical Data Analysis: Discrete Inverse Theory*, Academic Press, Inc., New York.
- Milne, G.A., Davis, J.L., Mitrovica, J.X., Scherneck, H.-G., Johansson, J.M., Vermeer, M. & Koivula, H., 2001. Space-geodetic constraints on glacial isostatic adjustment in Fennoscandia, *Science*, **291**, 2381–2385.
- Nironen, M., Lahtinen, R. & Korja, A., 2002. Paleoproterozoic tectonic evolution of the Fennoscandian shield—comparison to modern analogues, in *Lithosphere 2002—Second Symposium on the Structure, Composition and Evolution of the Lithosphere in Finland*, pp. 95–97, eds Lahtinen, R., Korja, A., Arhe, K., Eklund, O., Hjelt, S.-E., Pesonen, L.J., Programme and extended abstracts, Espoo, Finland, Institute of Seismology, University of Helsinki, report S-42.
- Oliver, J. & Murphy, L., 1971. WWNSS: Seismology's global network of observing stations, *Science*, **174**, 254–261.
- Olsson, S., Roberts, R.G. & Shomali, Z.H., 2004. Imaging upper mantle discontinuities using P to S converted phases, *European Seismological Commission XXIX General Assembly, poster presentation*, Potsdam, Germany.
- Perchuc, E. & Thybo, H., 1996. A new model of upper mantle P-wave velocity below the Baltic Shield; indication of partial melt in the 95 to 160 km depth range, *Tectonophysics*, **253**, 227–245.
- Plomerova, J., Arvidsson, R., Babuska, V., Granet, M., Kulhanek, O., Poupinet, G. & Sileny, J., 2001. An array study of lithospheric structure across the Protogine zone, Varmland, south-central Sweden—signs of a paleocontinental collision, *Tectonophysics*, **332**, 1–21.
- Plomerova, J., Hyvonen, T., Sandoval, S., Babuska, V. & the SSTW Group, 2002a. Structure of the Fennoscandian mantle from seismic velocity anisotropy, in *Geophys. Res. Abs.*, **Vol. 4**, 02173 27th General Assembly of the European Geophysical Society, Nice, France, 21–26 April, 2002.
- Plomerova, J., Babuska, V., Vecsey, L., Kouba, D. & TOR Working Group, 2002b. Seismic anisotropy of the lithosphere around the Trans-European Suture Zone (TESZ) based on teleseismic body-wave data of the Tor experiment, *Tectonophysics*, **360**, 89–114.
- Plomerova, J., Babuska, V., Vecsey, L., Kozlovskaya, E., Raita, T. & SSTWG., 2006. Proterozoic–Archaean boundary in the upper mantle of eastern Fennoscandia as seen by seismic anisotropy, *J. Geodyn.*, **41**, 400–410.
- Pollack, H.N. & Chapman, D.S., 1997. On the regional variation of heat flow, geotherms and lithospheric thickness, *Tectonophysics*, **38**, 279–296.
- Sacks, I.S., Snoke, J.A. & Husebye, E.S., 1979. Lithosphere thickness beneath the Baltic Shield, *Tectonophysics*, **56**, 101–110.
- Sandoval, S., Kissling, E., Ansorge, J. & the SVEKALAPKO Seismic Tomography Working Group, 2004. High-resolution body wave tomography beneath the SVEKALAPKO array: II. Anomalous upper mantle structure beneath central Baltic Shield, *Geophys. J. Int.*, **157**, 200–214.
- Shearer, P.M., 1999. *Introduction to Seismology*, Cambridge University Press, Cambridge.
- Shomali, Z.H., Roberts, R.G. & the TOR Working Group, 2002. Non-linear body wave teleseismic tomography along the TOR array. *Geophys. J. Int.*, **148**, 562–575.
- Shomali, Z.H., Roberts, R.G., Pedersen, L.B. & the TOR Working Group, 2006. Lithospheric structure of the Tornquist Zone resolved by nonlinear P and S teleseismic tomography along the TOR array, *Tectonophysics*, **416**(1–4): 133–149.
- Stammler, K., 1993. Seismic Handler—Programmable multichannel data handler for interactive and automatic processing of seismological analyses, *Comput. Geosci.*, **2**, 135–140.
- Steck, L.K., Prothero, J.R. & W.A., 1991. A 3-D raytracer for teleseismic body-wave arrival times, *Bull. seism. Soc. Am.*, **81**, 1332–1339.
- Thybo, H. & Perchuc, E., 1997. The seismic 8 degrees discontinuity and partial melting in continental mantle, *Science* **275** (5306), 1626–1629.
- Vecsey, L., Plomerova, J., Kozlovskaya, E. & Babuska, V., 2006. Shear-wave splitting as a diagnostic of variable upper mantle structure beneath south-eastern Fennoscandia, *Tectonophysics*, submitted.
- Waldhauser, F., Lippitsch, R., Kissling, E. & Ansorge, J., 2002. High-resolution teleseismic tomography of upper mantle structure using an a priori 3D crustal model, *Geophys. J. Int.*, **150**, 141–403.
- Weiland, C.M., Steck, L.K., Dawson, P.B. & Korneev, V., 1995. Nonlinear teleseismic tomography at Long Valley caldera, using three-dimensional travel time ray tracing, *J. geophys. Res.*, **100**, 20 379–20 390.
- Wu, P., Johnston, P. & Lambeck, K., 1999. Postglacial rebound and fault instability in Fennoscandia, *Geophys. J. Int.* **139**, 657–670.
- Wylegalla, K., Bock, G., Gossler, J., Hanka, W. & TORWorking Group, 1999. Anisotropy across the Sorgenfrei–Tornquist zone from shear wave splitting, *Tectonophysics*, **314**, 335–350.