

**POLENET/LAPNET teleseismic P-wave  
traveltime tomography model**

H. Silvennoinen et al.

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# POLENET/LAPNET teleseismic P-wave traveltime tomography model of the upper mantle beneath northern Fennoscandia

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Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



## Abstract

The POLENET/LAPNET broadband seismic array was deployed in northern Fennoscandia (Finland, Sweden, Norway, and Russia) during the third International Polar Year 2007–2009. The array consisted of roughly 60 seismic stations. In our study we estimate the 3-D architecture of the upper mantle beneath the northern Fennoscandian shield using high-resolution teleseismic P-wave tomography. For this purpose 111 clearly recorded teleseismic events were selected and the data from the stations hand-picked and analysed. Our study reveals a highly heterogeneous lithospheric mantle beneath the northern Fennoscandian shield though without any large high P-wave velocity area that may indicate presence of thick depleted lithospheric “keel”. The most significant feature seen in the velocity model is a large elongated negative velocity anomaly (up to  $-3.5\%$ ) in depth range 100–150 km in the central part of our study area that can be followed down to a depth of 200 km in some local areas. This low-velocity area separates three high-velocity regions corresponding to the cratons and it extends to greater depth below the Karelian craton.

## 1 Introduction

Recently, dense two-dimensional (2-D) arrays of broadband seismic instruments have proved to be a most effective mean to study the 3-D structure of the lithosphere and the lithosphere–asthenosphere boundary (LAB) (Trampert and Van der Hilst, 2005). One of such arrays was the POLENET/LAPNET broadband seismic array (<http://www.oulu.fi/sgo-oty/lapnet>) deployed in northern Fennoscandia (Finland, Sweden, Norway, and Russia) during the third International Polar Year 2007–2009. The project was a part of POLENET (POLar Earth observing NETwork, <http://www.polenet.org>) consortium. The array consisted of 37 temporary and 21 permanent seismic stations (Fig. 1). All the stations, except of 2 temporary stations, were the broadband ones.

SED

7, 2527–2562, 2015

## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion







cofennian orogeny formed a large unit of new Palaeoproterozoic crust, similarly to the Lapland-Kola orogeny, its northern part mainly comprises of reworked Archean crust. The orogeny began from north, where the Karelian Craton and Norbotten craton collided at ca. 1.92 Ga (Lahtinen et al., 2008, 2015).

After these two orogenies there has been no major tectonic events in our study area, but there are still some smaller volume magmatism with ages from Palaeoproterozoic to Devonian (Downes et al., 2005). These rocks are often from relatively deep source and, though quite diverse in rock types, are some of the most important carriers of mantle xenoliths (Woodard, 2010).

### 3 Dataset

The main data set used in this study is the data of POLENET/LAPNET array that recorded during May 2007–September 2009. For this study we selected 96 teleseismic events that were located at the epicentral distances between 30 and 90° from our station array and were clearly recorded by most of the stations. Most of the selected events have magnitudes larger than 6.0, but events with magnitudes larger than 5.5 were also considered in order to improve the azimuthal coverage. In this way we obtained a good coverage over all back-azimuths (see Fig. 2); the largest azimuthal gap in events recorded by POLENET/LAPNET array is smaller than 15 degrees. The first arrivals of P-waves were picked manually using Seismic Handler software (<http://www.seismic-handler.org/>). We used the WWSSN short period simulation filter for easier comparison between waveforms recorded by different types of sensors. The traveltimes residuals were calculated using IASP91 reference model (Kennett and Engdahl, 1991). While picking the arrivals, the uncertainty of each arrival time was also estimated. The picked arrival times were then divided into 3 quality classes based on individual uncertainty estimates. Table 1 shows the error estimates for each class and the number of arrival times attributed to different classes. Example wave forms with arrival time picks of different quality are shown in Fig. 3.

## POLENET/LAPNET teleseismic P-wave traveltimes tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



“Raw” residuals calculated as differences between observed and theoretical arrival times using IASP91 velocity model and ISC catalogue hypocentre parameters exhibit large amplitudes while showing only small azimuthal variations between stations (see Fig. 4a). This is the combined effect of deep mantle velocity variations located outside our study volume and of near-source structure or teleseismic hypocentre parameter uncertainties. To separate those effects from our observations, the average traveltime residual over all stations was calculated for each event and reduced from all corresponding residuals. The resulting residuals contain the effects of the velocity variations within our study volume with reference to IASP91 model. Fig. 4b) shows the azimuthal distributions of residuals for eight selected stations representing different regions within our study area.

In addition to the new dataset from the POLENET/LAPNET array we included to our study previous data from the northern part of the SVEKALAPKO array (Hjelt et al., 2006) located south of POLENET/LAPNET array but still within our study area. This additional data included 15 events recorded at 31 stations yielding 360 P-wave residual times (Plomerová et al., 2006). The quality classes of the SVEKALAPKO data are shown in Table 1. All seismic stations included in this study are shown in Fig. 1b) and the event distribution is shown in Fig. 2.

#### 4 The inversion method

In seismic tomography, the basic system of equations relating velocity perturbations inside the study volume to traveltime residuals is

$$\mathbf{d} = \mathbf{G}\mathbf{m}, \quad (1)$$

where  $\mathbf{d}$  is a data vector composed of traveltime residuals,  $\mathbf{m}$  is a model parameter vector composed of velocity perturbations in each cell of the velocity model and  $\mathbf{G}$  is a matrix of partial derivatives defining the coupling between data and model parameters (e.g., Kissling, 1988). The method searches the velocity perturbations inside the



Best results for resolution estimates, however, are obtained by synthetic data testing in combination with hit count, derivative weighted sums, and resolution matrix (Kissling, 1988).

#### 4.1 Model parameterization and regularization

The POLENET/LAPNET study area is approximately 550 km in east-west direction and 600 km in north-south direction. For this study area we selected an 80 km × 80 km × 60 km inversion grid. The horizontal size of the cells is slightly larger than the average distance between stations to guarantee at least one station for each surface cell within the array space. Thus, the minimum dimension of a resolvable anomalous body beneath POLENET/LAPNET array is approximately 100 km. The uppermost 60 km are mostly made of crust and are not included in the high-resolution teleseismic inversion (e.g., Arlitt et al., 1999). In the case of real data, a priori crustal corrections are applied to travel time residuals (see Sect. 4.2). For the inversion of synthetic data (see Sect. 4.3) no crustal correction was applied as the crustal layers were assumed to have velocities according to the IASP91 reference model. The first layer inverted for below 60 km crustal layer is 40 km thick and all deeper layers are of thickness 60 km. For better performance of the ray tracer, the study volume was surrounded on all sides and on bottom with large cells of respective IASP91 velocities that were kept fixed during inversion. The P-wave velocities of the initial model correspond to the IASP91 reference model. Because of the uneven ray geometry the kernel matrix ( $\mathbf{G}$  in Eq. 1) can be singular and needs to be regularized. The regularization can be done by selecting appropriate damping value for the dataset ( $\varepsilon^2$  in Eq. 2). We analysed the effect of the damping values by performing multiple inversions with different damping values both with synthetic data and real data to find the best damping value for our data set. The final damping value 70 was selected by investigating the trade-off curve between model and data variance (see Fig. 5 for the trade of curve of the real data and Fig. 6 for examples of the crustal corrected residuals).

## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



## 4.2 Crustal correction model

Due to high angles of incident and subsequent near-lack of cross firing, teleseismic rays are largely incapable to resolve upper lithosphere structure such as Moho topography and 3-D crustal velocity variations. Lateral variation of crustal structure, however, may significantly affect travel times of teleseismic rays (Arlitt et al., 1999). Hence, when illuminating with high-resolution teleseismic tomography the structures at upper mantle depths, it is necessary to apriori correct the data for the effect of crustal structures as documented for southern Finland by Sandoval et al. (2003). A new map of the crustal thickness was established by Silvennoinen et al. (2014) based on previous and new controlled source seismic and receiver function results in our study area. This map in combination with an averaged 1D-velocity-depth function is used in this study to construct a 3-D crustal model for the purpose of correcting travel times for crustal effects.

The 3-D crustal model used in this study was established by using Bloxer software (<https://wiki oulu.fi/display/~mpi/Block+model+maintenance>). Bloxer is software designed to build 3-D block models with different parameters (density, seismic velocity, magnetization etc.). The Moho depth map by Silvennoinen et al. (2014) was imported to Bloxer. All blocks below the Moho to a maximal model depth of 60km were given a P-wave velocity value  $8.05 \text{ km s}^{-1}$  that is the P-wave velocity at uppermost mantle of IASP91 reference model.

As our study area has not enough P-wave velocity information available to estimate a true 3-D distribution of seismic velocities in the crust between CSS profiles, we used an average P-wave velocity for crustal depths defined from major refraction seismic profiles, namely FENNOLORA (Guggisberg, 1986; Guggisberg et al., 1991; Luosto and Korhonen, 1986) in Sweden, POLAR (Luosto et al., 1989; Janik et al., 2009) in Finland, and PECHENGA-KOSTOMUKSHA (Azbel et al., 1989; Azbel and Ionkis, 1991; Azbel et al., 1991) in Russia. A map of crustal correction times for vertical incident rays arriving at each station is shown in Fig. 6a. The crustal effect established by Sandoval

SED

7, 2527–2562, 2015

### POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion







to checkerboard tests, we see that the resolution in western part of the study area is not as good as in the eastern part. In the areas with fairly good resolution leakage may generally extends into the next layer up or down from the anomalous body, while in areas with poor resolution the leakage can extend to 2 layers (i.e., up to 120 km).

5 From the test (Fig. 10) we can also see that we obtain slight positive anomalies around the negative anomalous body and vice versa marking an “overswinging” effect Kissling et al. (2001).

## 5 Results and discussion

The main results of inversion with the real data are shown in Figs. 11 and 12. Our study revealed highly heterogeneous lithospheric mantle beneath the northern Fennoscandian Shield, without any large high P-wave velocity area that may indicate presence of thick depleted lithospheric “keel” revealed beneath southern part of the shield in Sweden (Shomali et al., 2006) and beneath the SVEKALAPKO study area (Sandoval et al., 2004) and in some other shield areas (e.g., Pasyanos and Nyblade, 2007; Priestley et al., 2008; Villemaire et al., 2012). As can be seen, P-wave velocities in our model are generally close to the reference IASP91 model. This is in agreement with the recent result of joint analysis of P- and S-wave receiver functions by (Vinnik et al., 2015). They showed that absolute average values of seismic velocities in the lithospheric mantle beneath southern Finland are generally higher than those beneath the northern Finland. Because of limitations of the data, they did not analyse the lateral heterogeneities in the lithospheric mantle. However, in our model we can recognize several high velocity anomalies in the upper part of the lithospheric mantle (down to depths of about 120–200 km) that spatially correlate with the Karelian, Kola and Norrbotten cratons (Fig. 11a, b). We interpret these high-velocity anomalies as non-reworked fragments of cratonic lithosphere preserved since the Archaean.

The non-reworked part of the Karelian craton can be recognized as a fast velocity anomaly (+3 %) from the Moho down to a depth of about 160 km. Below this depth we

### POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



observed velocities lower than those in the reference model. At the margin of the Kola craton we also see a positive (+3.5%) anomaly roughly down to 180 km. Below it the velocity values are close to those from the reference model (Fig. 11d). The Norrbotten Craton is also seen as a high velocity anomaly (+1.5%) starting from the Moho down to a depth of about 200 km (Fig. 11). These upper mantle heterogeneities can be explained by three major factors, namely temperature variations, compositional variations or seismic anisotropy.

Seismic anisotropy beneath the POLENET/LAPNET area was studied independently by Plomerová et al. (2011) and Vinnik et al. (2014). Plomerová et al. (2011) analysed the relative P-wave travel-time deviations of selected teleseismic events and lateral variations of shear-wave splitting. Their results demonstrate the lithosphere of the study area consists of distinctive domains of anisotropic structures. One of their findings was that the patterns related to the Proterozoic-Archean transition zone in central Finland continued to POLENET/LAPNET study area. The Archean mantle block interpreted by Plomerová et al. (2011) is collocated with the south-eastern upper mantle high velocity anomaly of our model. Vinnik et al. (2014) analysed the southern half of the POLENET/LAPNET study area using joint inversion of P-wave receiver functions and SKS recordings. They obtained an averaged depth distribution of the magnitude and azimuthal direction of the anisotropy in the area. They found an anisotropic layer with approximate S-wave anisotropy of 2.5% from below Moho to approximately 110 km with fast direction of 40–60° as well as another, slightly less anisotropic layer with same azimuthal direction starting from approximately 220 km. Above this lower layer they modelled another anisotropic layer with contrasting azimuthal direction of 110°. Radial anisotropy found by Vinnik et al. (2014) does not influence significantly travel times of teleseismic P-waves, propagating in sub-vertical direction.

As shown by Lee (2003), James et al. (2004) and Schutt and Leshner (2006), compositional variations between depleted in Fe and more fertile mantle peridotites can explain up to 1–2% velocity anomalies of P-wave velocities. In order to explain larger anomalies, one would need the combined effect of major element chemistry and tem-

perature, as shown by Hieronymus and Goes (2010). That is why the lowered seismic velocities in the lithospheric mantle in our study area are most probably due to the combined effect of more fertile composition and generally higher temperature.

The most significant feature seen in the velocity model is a large negative velocity anomaly (up to  $-3.5\%$ ) in the central part of our study area that can be followed down to a depth of 160–200 km (Fig. 11). In the upper part of the model this low-velocity area separates three high-velocity region corresponding to the cratons and it extends to the greater depth below the Karelian craton (Fig. 11b, c, d). As is shown in several previous studies (e.g., Pedersen et al., 2013; Plomerová and Babuška, 2010; Lebedev et al., 2009), the lithospheric thickness in POLENET/LAPNET study area in northern Fennoscandia must be at least roughly 150 km, which makes it unlikely the upper boundary of the low velocity anomaly seen in our model at depth shallower than 100 km were the lithosphere-asthenosphere boundary. In Fig. 13 we show a north-south directed cross-section through a model based on surface waves by Pedersen et al. (2013) together with the location of the low velocity zone found in our study.

The low velocity zone is spatially overlapping in the East with the Kola alkaline province (see Fig. 14), in which the crust has been intruded by alkaline magmas (Downes et al., 2005) during several metasomatic events. Kempton et al. (1995) proposed that at least one of them was ancient, while the latest occurred in Devonian. The same events would result also in extensive reworking and refertilization of the originally depleted Archean mantle keel, while the latest Devonian magmatism would explain also higher mantle temperatures.

On the other hand, the area of low seismic velocities in the central part of the array separates Norrbotten, Kola and Karelian Cratons from each other and is spatially correlating with the northern part of the 1.9–1.8 Ga N–S trending Baltic-Bothnia Megashear (BBMS) stretching below the Baltic Sea and Gulf of Bothnia and continuing to the Caledonides in Norway (Berthelsen and Marker, 1986, see Fig. 13;). The northern part of the BBMS was later named Pajala shear zone by Kärki et al. (1993). This shear zone is about 40 km wide represented by a complex set of N–S striking shear and thrust

## SED

7, 2527–2562, 2015

### POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



zones. Lahtinen et al. (2015) proposed that the Pajala shear zone originated as a divergent plate boundary due to collision of two Archean continental units (Norrbotnen and Karelian), and it was multiply reactivated after continental collision with both lateral and vertical movements before 1.83 Ga. The vertical seismic velocity distribution in the upper mantle obtained in our model (e.g. low velocity over high velocity) is similar to that revealed by Bruneton et al. (2004) and Sandoval et al. (2004) (see Fig. 14), we suggest that the Pajala shear zone may continue to the south beneath the Gulf of Bothnia, as was originally proposed by Berthelsen and Marker (1986).

However, the metasomatic processes and refertilization of the upper mantle in the Palaeoproterozoic alone would not produce such low seismic velocities as we observed in our model and combined effect of temperature and composition would be necessary (Hieronymus and Goes, 2010). Thus we may speculate that since the Palaeoproterozoic the whole BBMS was reactivated by a later tectonothermal event (or multiple events), during which the cratonic lithosphere was partly destroyed. The time of those events is not clear, but it could occur the same time with the Paleozoic post-collisional alkaline magmatism in the Kola alkaline province caused by a plume activity (Marty et al., 1998; Downes et al., 2005; Kogarko et al., 2010).

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## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



- Downes, H., Balaganskaya, E., Beard, A., Liferovich, R., and Demaiffe, D.: Petrogenetic processes in the ultramafic, alkaline and carbonatitic magmatism in the Kola Alkaline Province: A review, *Lithos*, 85, 48–75, 2005. 2531, 2540, 2541
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**POLENET/LAPNET  
teleseismic P-wave  
traveltime  
tomography model**

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



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## SED

7, 2527–2562, 2015

### POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

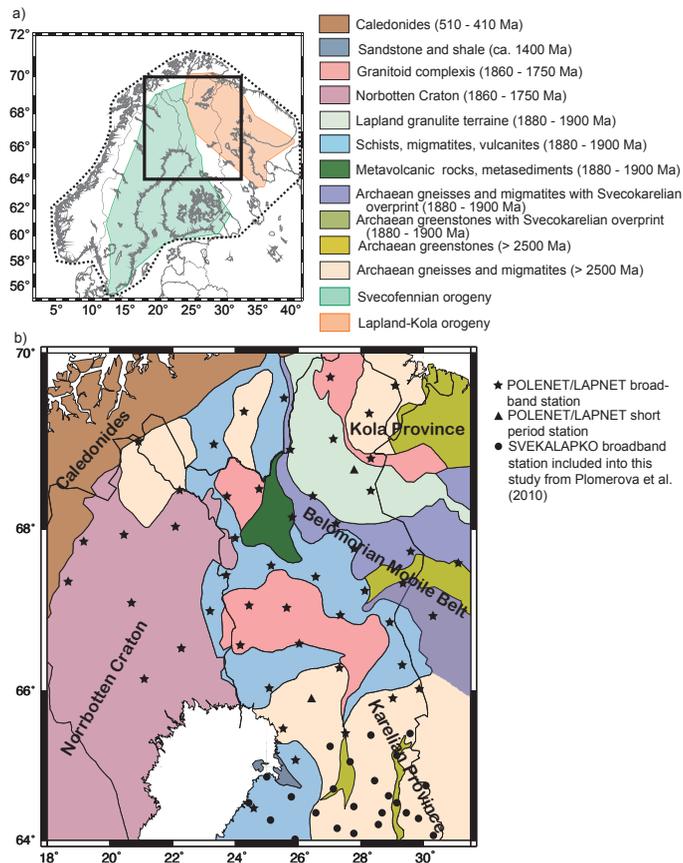
Interactive Discussion





## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.

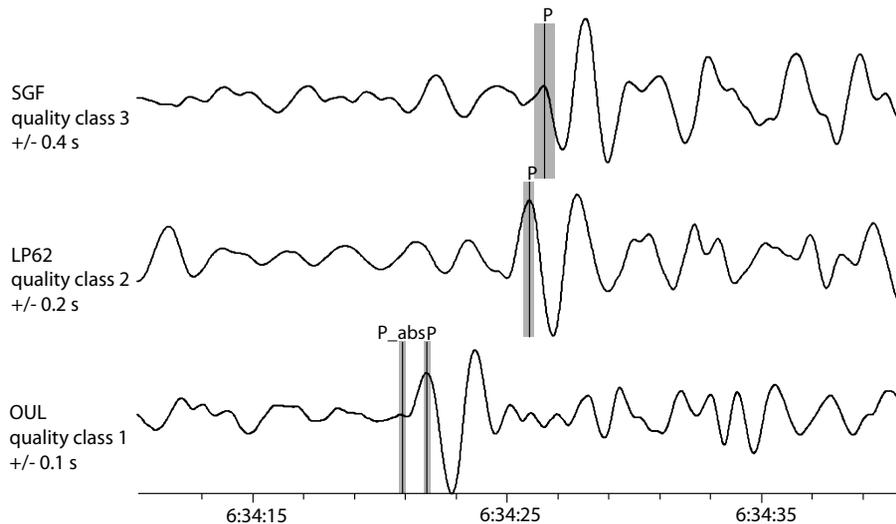


**Figure 1.** The map of the study area. **(a)** Simplified map of orogenies (based on Lahtinen et al. (2015)) in northern Fennoscandia. The Fennoscandian crustal block is outlined by black dotted line and our study area a rectangle. **(b)** The simplified geological map is based on 1 : 2 000 000 geological map of Fennoscandia (Koistinen et al., 2001). The seismic stations are shown with black stars, triangles and dots and main geological provinces are named.



## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.



**Figure 3.** Example traveltime picks of different quality from same event observed at three different stations. The recorded waveforms have been filtered with the WWSSN short period simulation filter. The travel time observation from the station OUL was attributed a quality class 1, the one from station LP62 a quality class 2 and from station SGF a quality class 3. For OUL the absolute travel ( $P_{\text{abs}}$ ) time was picked as well as the relative travel time ( $P$ ). For each pick the timing uncertainty estimate of the quality class is shown with grey rectangle.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

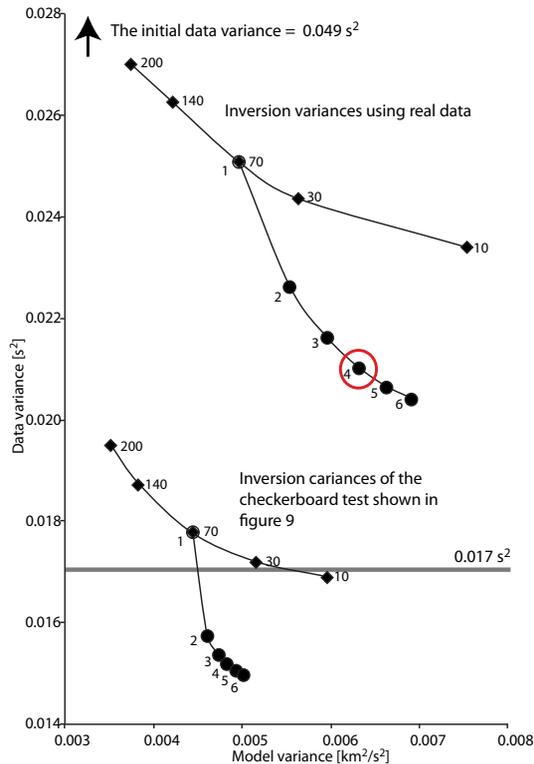
Interactive Discussion





## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.



**Figure 5.** A comparison between data and model variance with different damping values. The selected damping value was first found using one inversion round only and the results are shown with diamonds with the damping value used shown on the right side of the symbols. The number of iterations was also optimized (dots with the number of iterations marked to their left side) for the selected damping value 70. The final number of iterations, 4, is marked with red circle. The grey line denotes the overall data uncertainty,  $0.017 \text{ s}^2$ , calculated as the average of all observations.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

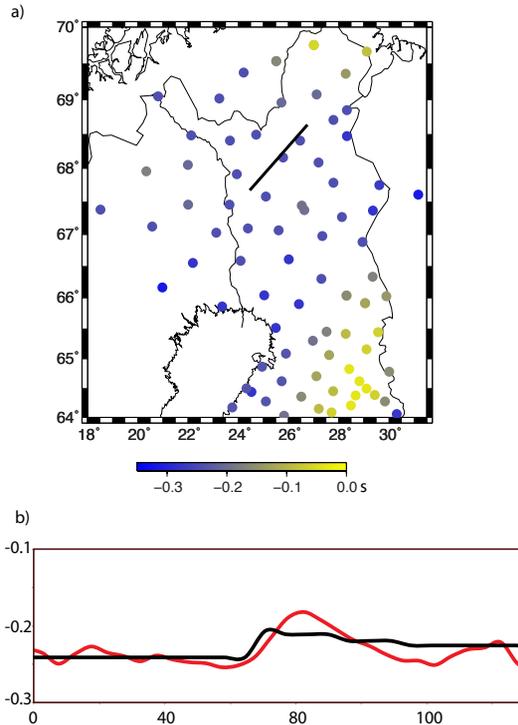
Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion





**Figure 6.** Crustal correction map. Panel **(a)** shows crustal correction values for all of the station. The values are based on a travel time of a vertical seismic ray travelling through the crustal correction model. Panel **(b)** shows the comparison of the travel times through semi-2-D crustal correction model used in this study (black line) and southern part of POLAR wide-angle reflection and refraction profile (red line, Janik et al., 2009). The location of the comparison is shown as a black line in panel **(a)**.

**POLENET/LAPNET  
teleseismic P-wave  
traveltime  
tomography model**

H. Silvennoinen et al.

Title Page

Abstract Introduction

Conclusions References

Tables Figures

◀ ▶

◀ ▶

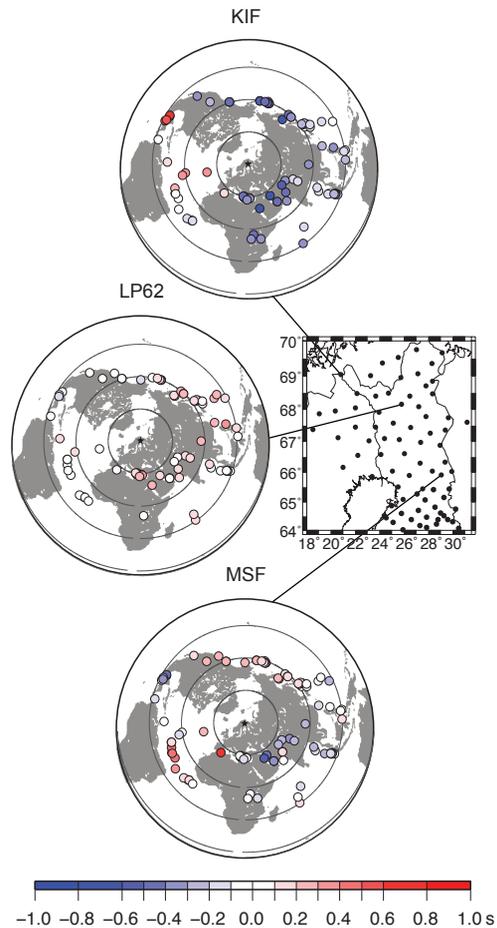
Back Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion





**Figure 7.** Distribution of crustal corrected travel time residuals at few example stations from different parts of POLENET/LAPNET array.

**POLENET/LAPNET  
teleseismic P-wave  
traveltime  
tomography model**

H. Silvennoinen et al.

Title Page

Abstract Introduction

Conclusions References

Tables Figures

◀ ▶

◀ ▶

Back Close

Full Screen / Esc

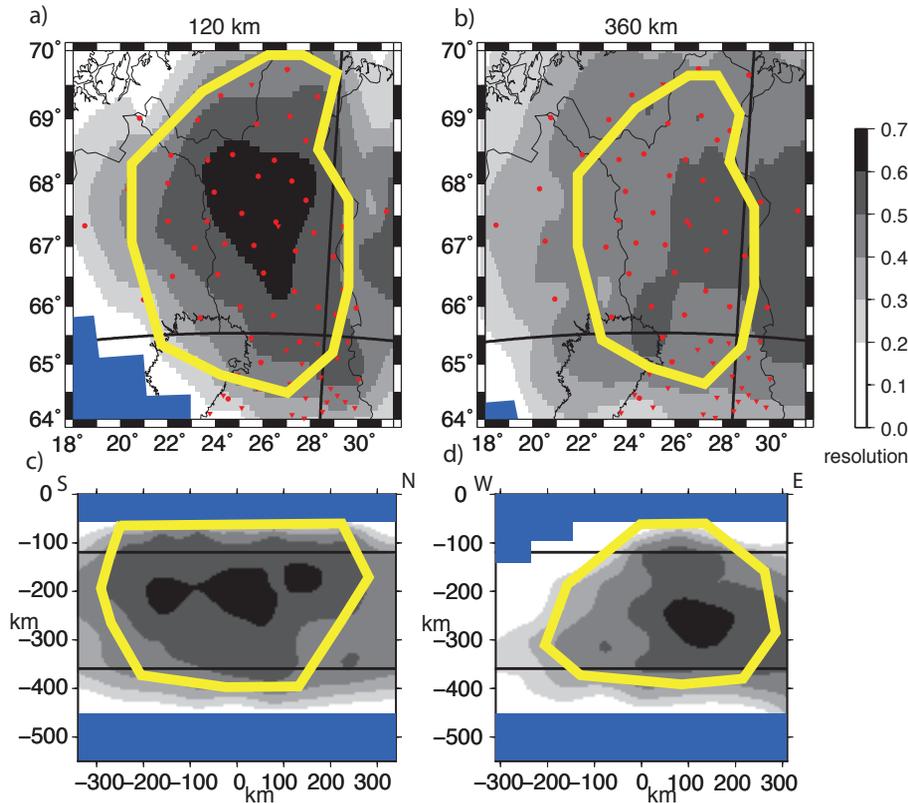
Printer-friendly Version

Interactive Discussion



## POLENET/LAPNET teleseismic P-wave traveltime tomography model

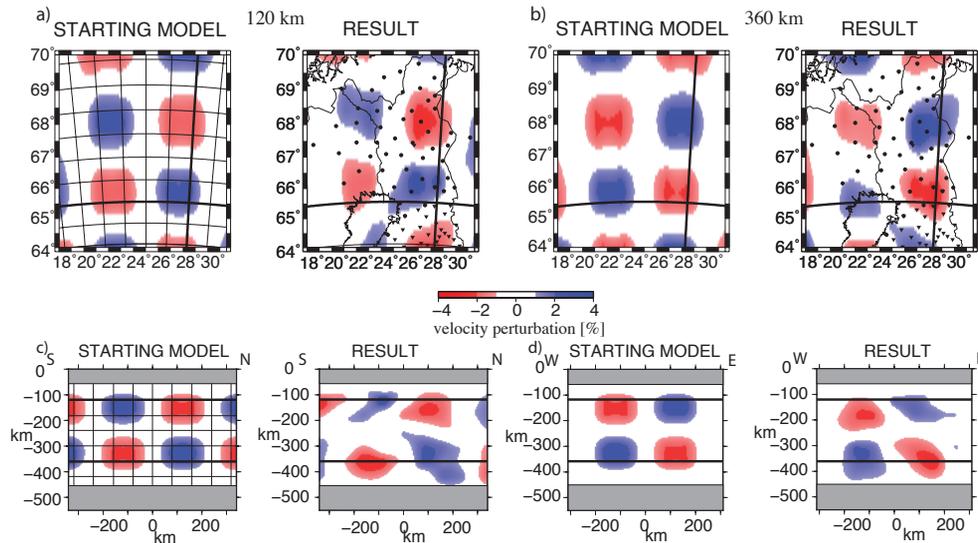
H. Silvennoinen et al.



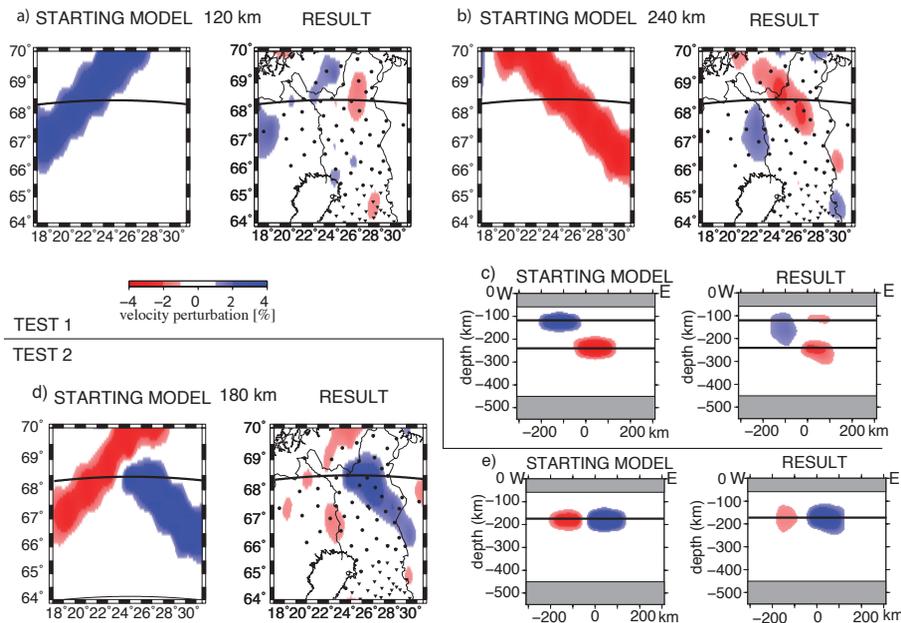
**Figure 8.** Diagonal elements of the resolution matrix (RDE). **(a)** and **(b)** show RDE for horizontal layers at depths of 120 and 360 km, respectively, **(c)** RDE distribution along a north-south directed vertical section, and **(d)** RDE along a west-east directed vertical section. The locations of the vertical sections are marked in horizontal sections with black lines and vice versa. The blue blocks show areas that were not inverted, as there was no data available. Red dots and triangles show the locations of POLENET/LAPNET and SVEKALAPKO stations, respectively.

## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.



**Figure 9.** Checkerboard test results. Two horizontal sections are shown at the depths of 120 and 360 km (**a** and **b**) as well as two vertical sections, one in SN direction (**c**) and one in EW direction (**d**). For each section both the model used to compute the synthetic data set and the result after inversion are shown. The model plots of (**a**) and (**c**) show the inversion grid used. In horizontal plots the locations of vertical sections are shown with thick black lines and vice versa. The areas that were not inverted are marked with grey rectangles.

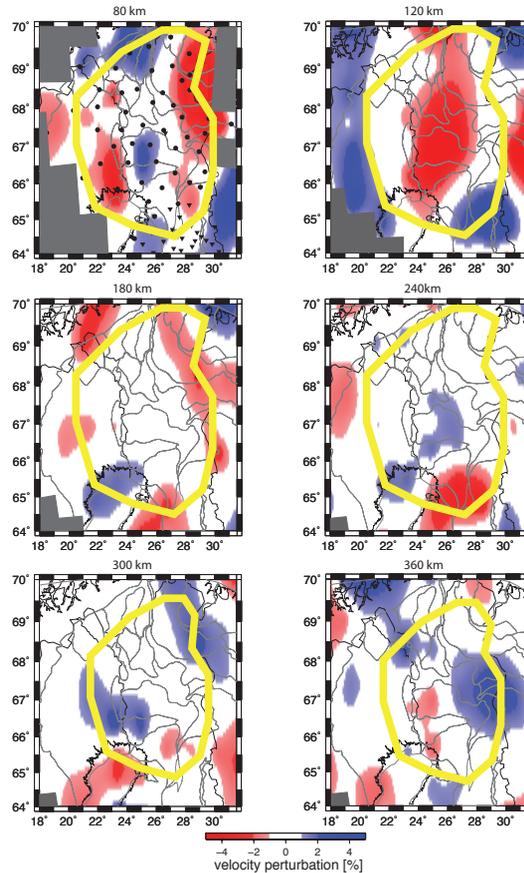


**Figure 10.** Synthetic test result examples. Two synthetic tests are shown. Test 1 (two horizontal sections in **a**, **b** and a vertical section in **c**) shows a test with crossing anomalous bodies at the depths of 120 and 250 km. The low velocity anomaly is located beneath Belomorian Mobile Belt, a hypothetical location for an anomaly and the high velocity body above it crosses it perpendicularly. Test 2 (a horizontal and vertical panel in **d** and **e**, respectively) shows another test with anomalies horizontally in same locations but now at the same depth. The locations of vertical sections are shown in horizontal sections with thick black lines and vice versa. The thin lines show the horizontal inversion grid. In horizontal results plots the seismic stations are marked with dots. In vertical sections the cells not inverted are marked with grey rectangles.

[Title Page](#)
[Abstract](#)
[Introduction](#)
[Conclusions](#)
[References](#)
[Tables](#)
[Figures](#)
[⏪](#)
[⏩](#)
[◀](#)
[▶](#)
[Back](#)
[Close](#)
[Full Screen / Esc](#)
[Printer-friendly Version](#)
[Interactive Discussion](#)

**POLENET/LAPNET  
teleseismic P-wave  
traveltime  
tomography model**

H. Silvennoinen et al.



**Figure 11.** Horizontal sections through our final inversion result. The depth of each section is marked on top of the figure. The grey corners show areas that were crossed by no rays and consecutively were excluded from inversion. The fairly well resolved area of the model is shown for each depth with yellow line.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

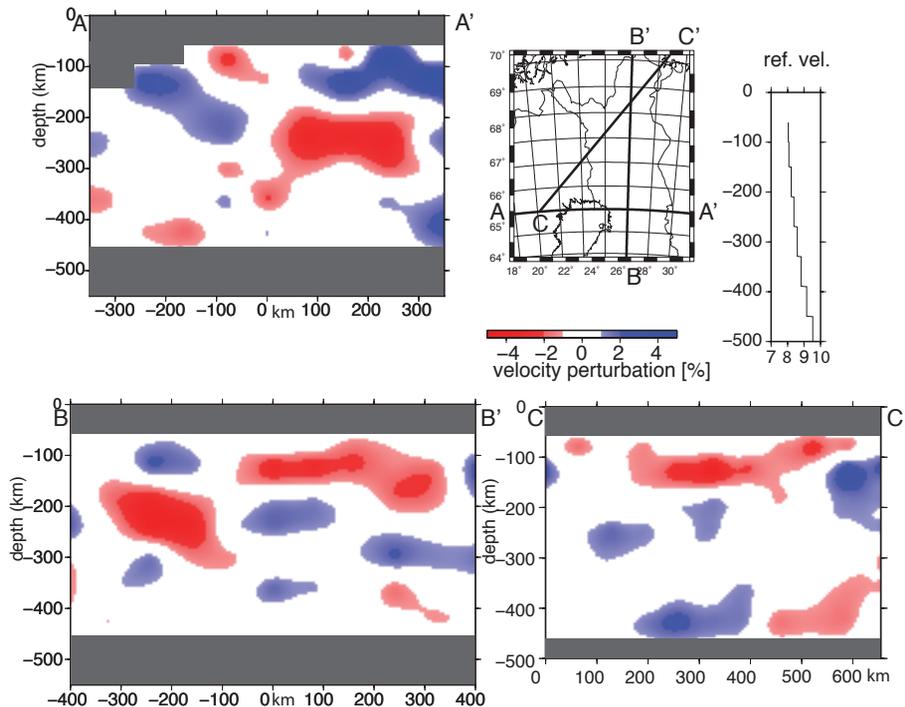
Printer-friendly Version

Interactive Discussion



**POLENET/LAPNET  
teleseismic P-wave  
traveltime  
tomography model**

H. Silvennoinen et al.



**Figure 12.** Vertical sections through our final inversion result. The locations of the sections are shown as black lines on the map also showing the horizontal grid. In addition to vertical section we show here the reference velocity curve used (modified from IASP91 reference model, Kennett and Engdahl, 1991). The areas where no inversion was done are shown in grey colour.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

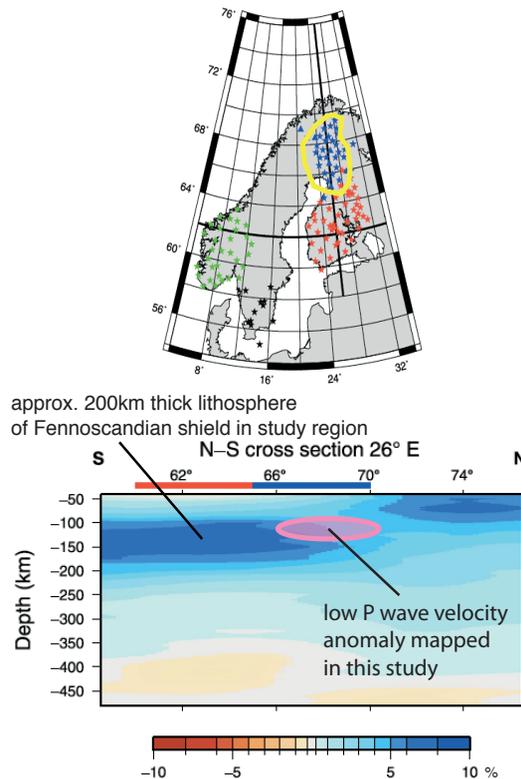


## SED

7, 2527–2562, 2015

**POLENET/LAPNET  
teleseismic P-wave  
traveltime  
tomography model**

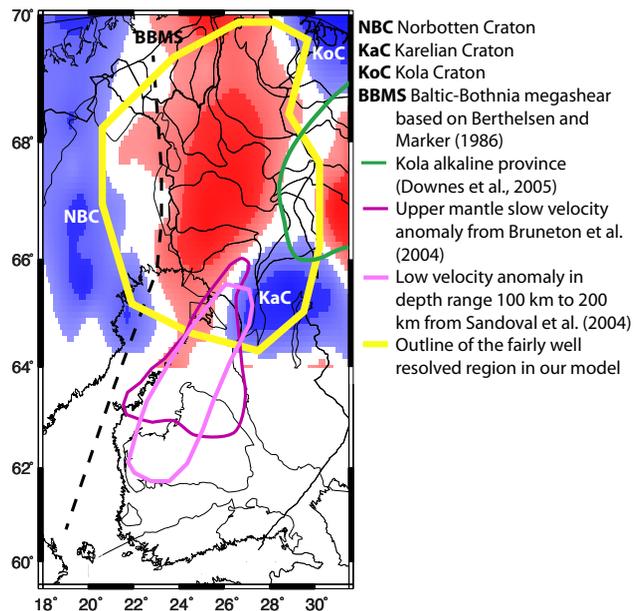
H. Silvennoinen et al.



**Figure 13.** Comparison to previous lithosphere thickness results. The figure has been modified from Pedersen et al. (2013) as it shows the S-wave velocity structure obtained in that study in comparison to global S-wave velocity model by Debayle and Ricard (2012). The area outlined with yellow line shows the well-resolved part of our model at the depth of the low velocity anomaly found in the upper mantle.

## POLENET/LAPNET teleseismic P-wave traveltime tomography model

H. Silvennoinen et al.



**Figure 14.** Our results at the depth of 120 km together with cratonic units in our study area and locations of the Kola alkaline province, an upper mantle low velocity anomaly found in SVEKALAPKO data by Bruneton et al. (2004) and Sandoval et al. (2004) and Baltic-Bothnia megashear.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

