

Crust-mantle boundary in the central Fennoscandian shield: Constraints from wide-angle P and S wave velocity models and new results of reflection profiling in Finland

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[1] Our paper presents analysis of the crust-mantle boundary in the central Fennoscandian shield based on new P and S wave two-dimensional velocity models of the SVEKA'81, SVEKA'91 and FENNIA wide-angle reflection and refraction profiles and on results of a new seismic reflection experiment in Finland (Finnish Reflection Experiment (FIRE)). In this area, the crust is extremely thick (50–60 km), and the Moho boundary is difficult to detect using methods based on interpretation of P waves only (near vertical profiling and wide-angle experiments). However, the S wave reflections from the Moho boundary (SmS) are frequently more pronounced in wide-angle data than the P wave reflections (PmP). In order to infer the crust-mantle transition, we developed new P and S wave velocity models using reprocessing of the old data and compared them to record sections of collocated reflection profiles and to published values of V_p and V_p/V_s for the main types of lower crustal and upper mantle rocks. On the basis of the lateral variations of V_p , V_p/V_s , and reflectivity in the lower crust and upper mantle, three main types of the crust-mantle boundary were distinguished. The first type corresponds to eclogitized lower crust that overlies peridotitic upper mantle. In this case, the Moho coincides with the lithological crust-mantle boundary. The second type corresponds to lower crust composed of mafic garnet granulites overlying the peridotitic upper mantle, for which the Moho and the lithologic crust-mantle boundary coincide as well. The third type corresponds to the mafic garnet granulites underlain by a layer of eclogitic upper mantle. In the latter case, the lithological crust-mantle boundary is deeper than the Moho.

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1. Introduction

[2] The origin and evolution of the crust-mantle boundary in Precambrian areas has been debated for many years but remains the subject of debate. Generally, connection between the crust-mantle boundary and the seismically defined Mohorovičić discontinuity (Moho) [Mohorovičić, 1910] is not straightforward, although an abrupt increase of seismic velocity at the Moho ($>7.6 \text{ km s}^{-1}$) suggests such connection. It has been proposed that the velocity increase at the Moho is explained by the change in bulk chemical composition or transition in mineral phases [e.g., Ito and Kennedy, 1971; Green and Ringwood, 1972; Griffin and O'Reilly, 1987]. Since the mid-1970s, application of the seismic reflection profiling technique to the study of the deep crust and upper mantle has provided a large database with which the structure and physical properties of the

crust-mantle transition can be analyzed. On the basis of these data, mechanical decoupling of the crust and mantle across the Moho was proposed as the main explanation of its reflectivity [Mooney and Meissner, 1992; Meissner and Mooney, 1998]. Observations of the Moho variability from the Lithoprobe near-vertical incidence seismic reflection surveys suggested that the Moho reflectivity may be correlated to lithospheric evolution and that this correlation can be extended into the Archaean [Hammer and Clowes, 1997].

[3] On the basis of the general property of the Moho boundary (e.g., increase of seismic velocities), the depth to the crust-mantle boundary (the Moho boundary) is usually defined by seismic techniques, e.g., wide-angle reflection and refraction profiling, near-vertical reflection profiling and teleseismic receiver function studies. However, different seismic techniques employed to study the Moho boundary may give conflicting results [Eaton, 2006] and interpretation of these results in terms of rock composition is difficult. The main reason for this is the long geological history of Precambrian regions that resulted in the complicated lithosphere structure.

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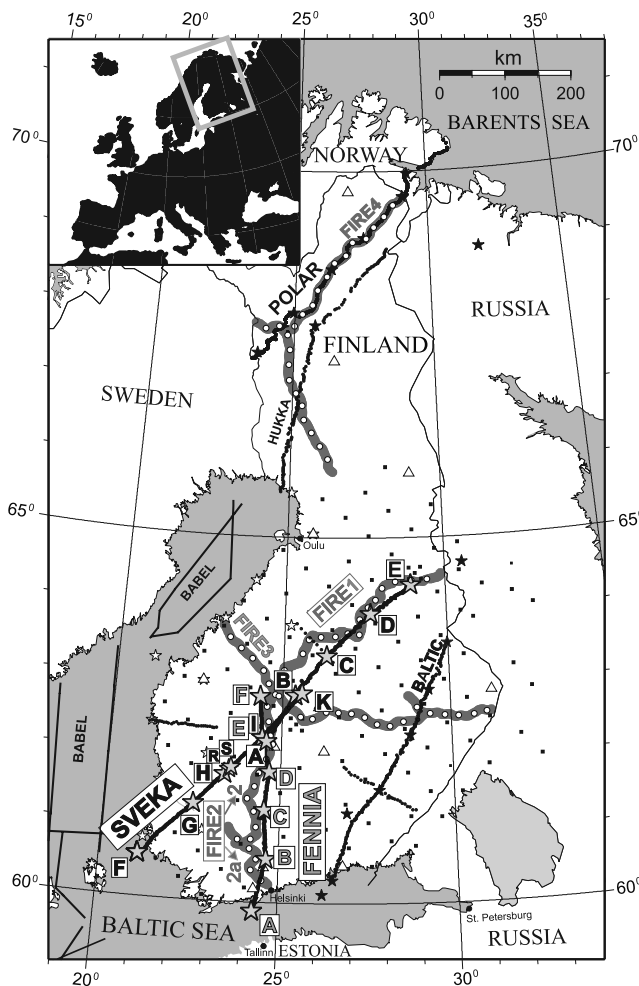


Figure 1. Location of main onshore and offshore seismic experiments in the eastern part of the Fennoscandia. Investigated, SVEKA transect (SVEKA'81 (SPA, SPB, SPC, SPD, SPE) profile and SVEKA'91 (SPF, SPG, SPH, SPR, SPI, SPK) profile) and FENNIA profile, and FIRE1 (20,000 CDP points), FIRE2 (line 2a 5600 CDP points and line 2 9000 CDP points) profiles display more details. Stars represent the shot points of the profiles; black dots represent recording stations of DSS experiment; gray dots represent receiver stations; and large white dots represent selected CDP points of the deep seismic reflection experiment FIRE which are described on zoomed part of Figure 2. Positions of record stations of the SVEKALAPKO temporary array are denoted by rectangles (temporary broadband and short-period stations) and by triangles (permanent stations). Inset map shows the location of target area in Europe.

[4] One such area is the central part of the Precambrian Fennoscandian shield (Figures 1 and 2), where the crust mantle boundary has been studied by all the seismic methods mentioned above. In the 1980s–1990s the central part of the shield was studied by BALTIC, SVEKA'81, SVEKA'91 and FENNIA wide-angle reflection and refraction profiles [FENNIA Working Group, 1998; Grad and Luosto, 1987, 1994; Heikkinen and Luosto, 2000; Luosto,

1991, 1997; Luosto *et al.*, 1984, 1990, 1994; Yliniemi, 1991]. In 1989, a marine seismic survey recorded more than 2000 km of deep reflection profiles around the Gulf of Bothnia and the Baltic Sea [BABEL Working Group, 1990, 1993a, 1993b]. The Baltic And Bothnian Echoes from the Lithosphere (BABEL) survey revealed a clear reflection Moho (“bright Moho”) that appears as a transition from highly reflective lower crust to transparent upper mantle. These surveys (Figure 1) revealed strong variations of the structure and thickness of the crust in the suture zone between the Svecofennian and Archaean domains. A map of the Moho boundary was compiled from the results of these two-dimensional (2-D) surveys by Luosto [1991, 1997]. In this map, the thickness of the crust varies between 42 and 65 km, the Svecofennian crust is generally thicker than the crust of the Karelian domain, and the thickest crust was found in the contact between Archaean and Proterozoic domains.

[5] The SVEKALAPKO (1998–1999) multidisciplinary passive seismic array [Bock and SVEKALAPKO Seismic Tomography Working Group, 2001], covering the area from 59° to 68°N and 18° to 34°E (Figure 1), was deployed to find the signatures of the Proterozoic–Archaean suture in the upper mantle [Bock and SVEKALAPKO Seismic Tomography Working Group, 2001]. A detailed description of the SVEKALAPKO seismic field experiment and data processing procedures was presented by Sandoval [2002] and Sandoval *et al.* [2003, 2004], and as a part of these studies, a new map of the Moho boundary from the results of the previous deep seismic sounding (DSS) profiles. The new map of the crust-mantle boundary beneath the SVEKALAPKO array was obtained from the *P* wave receiver function (RF) [Bock and SVEKALAPKO Seismic Tomography Working Group, 2001; Alinaghi *et al.*, 2003]. The depth to the Moho obtained from analysis of *P*-to-*S* conversions generally differs slightly from that presented by Luosto [1991, 1997] and Sandoval *et al.* [2003], and in some places it is about 10–15 km deeper.

[6] New information about the crust-mantle boundary was also obtained by Yliniemi *et al.* [2004], who presented results of forward ray trace modeling of reflected and refracted *P* waves of the strongest local events registered by the SVEKALAPKO array. Recordings of these events demonstrated strong single (*PmP*) and double (*PmPPmP*) reflections from the Moho boundary. This study showed that the Moho depression beneath central Finland has a different shape than that presented by Luosto [1991, 1997], which is in agreement with the result by Alinaghi *et al.* [2003].

[7] In 2001–2003, the Finnish Reflection Experiment (FIRE) project collected data along four reflection seismic transects with a total length of 2104 km crossing the main geological units of the central part of the Fennoscandian shield (Figure 1) [Kukkonen *et al.*, 2006]. The FIRE1 and FIRE3 profiles revealed complicated reflection patterns in the crust along the boundary of Archaean and Proterozoic domains, with no coherence between the orientation of upper and lower crustal structures. In contrast to the BABEL survey, this new experiment revealed no clear reflection Moho boundary (e.g., reflecting lower crust overlying transparent upper mantle) at the contact of Archaean and Proterozoic domains.

[8] This new and sometimes contradicting information about the Moho boundary in the central Fennoscandian

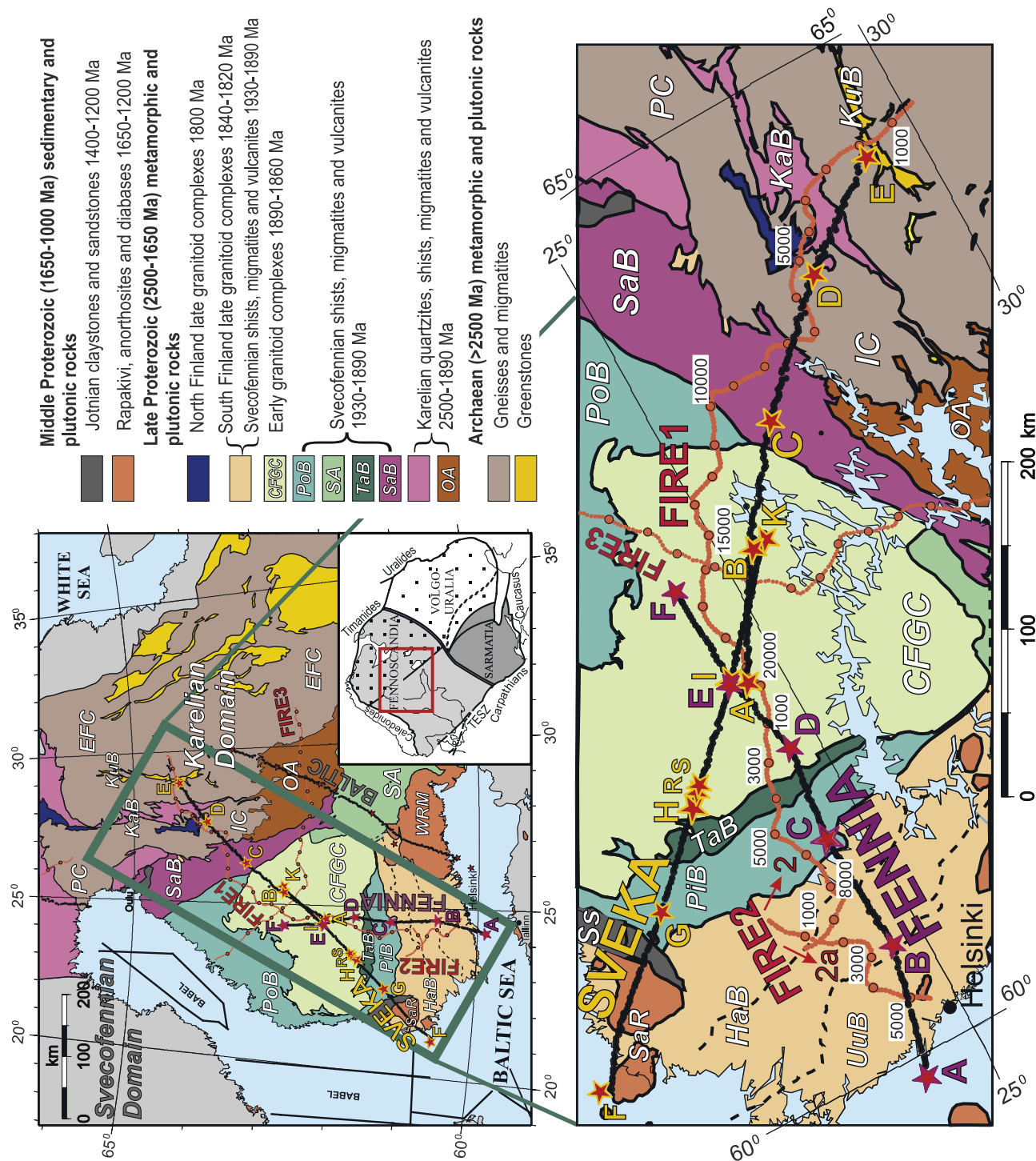


Figure 2

shield was the main motivation for our study, in which we tried to explain the apparent inconsistencies between the Moho obtained from wide-angle data, receiver function studies and reflection profiling. Thus we undertook a new analysis of the crust-mantle boundary in the central Fennoscandian shield using both P and S wave wavefields of previous SVEKA and FENNIA wide-angle reflection and refraction profiles. As shown by *Musacchio et al.* [1997], *Thybo et al.* [2003], and *Wang et al.* [2003], analysis of V_p/V_s ratio can provide important constraints on compositional variations at the Moho boundary. In our analysis, we compared the new P and S wave velocity models to record sections from the FIRE reflection experiment in Finland and to petrophysical data about seismic velocities of the main types of lower crustal and upper mantle rocks.

2. Geological Setting

[9] The Precambrian Fennoscandian shield (Figure 2) was formed as a result of several episodes of crustal accretion which started about 3.5 Ga [*Gorbatshev and Bogdanova*, 1993]. The main tectonic feature in the central part of the shield is a tectonic suture separating Paleoproterozoic Svecofennian (2.1–2.3 Ga) and Archaean Karelian (2.6–3.1 Ga) domains (Figure 2). The Karelian domain was formed during the Saamian (3.1–2.9 Ga) and Lopian (2.8–2.6 Ga) orogenies and is composed of migmatitic trondhemitic-tonalite-granodiorite (TTG) gneisses with minor greenstone rocks intruded by later granitoids. During 2.5–1.9 Ga metasediments and metavolcanic rocks were deposited on an eroded Archaean craton. The Svecofennian domain consists mainly of granitoid batholiths surrounded by schist belts and was formed from mantle-derived juvenile material between 1.9 and 1.75 Ga as a result of accretion of island arcs and microcontinents to the craton [*Lahtinen et al.*, 2005]. The Svecofennian crust was intruded by anorogenic rapakivi batholiths, mafic dike swarms and gabbro-anorthosites between 1.65 and 1.54 Ga in an extensional tectonic setting [*Korsman et al.*, 1999].

3. Data Acquisition

3.1. DDS Profiles

[10] Seismic measurements along the SVEKA, SVEKA'91, and FENNIA DSS profiles were carried out in 1981, 1991, and 1993, respectively. Hereafter we use the names “SVEKA profile” for SVEKA'81 profile and “SVEKA transect” for SVEKA'81 and SVEKA'91 profiles. Details of the experiments and interpretations were described by *Luosto et al.* [1984, 1994], *Grad and Luosto* [1987], and *FENNIA Working Group* [1998]. Ten large shots along the SVEKA transect and six shots along the FENNIA profile, with the charge of 200–1400 kg of TNT,

at intervals from 51 to 113 km were registered by mobile three-component seismic stations with 2–3 km spacing of recording sites (Figures 1 and 2). Additionally, two small (80 kg of TNT) shots, R and S, were fired to the northeast from regular shot H (800 kg of TNT) along the SVEKA'91. All shots were fired at the nighttime to minimize the level of seismic noise. The shots were detonated at shallow water at the depths of 5–11 m. Acquisition was undertaken in 5–7 deployments along each of the profiles.

3.2. FIRE Experiment

[11] The Finnish Reflection Experiment (FIRE) was managed by a consortium comprising of the Geological Survey of Finland, the Helsinki University (Institute of Seismology), and the University of Oulu (Department of Physics, Division of Geophysics and Sodankylä Geophysical Observatory) [*Kukkonen et al.*, 2006]. The experiment collected a total of 2135 km of vibroseis seismic reflection data along four profiles, FIRE1–4 crossing major geological units of the Fennoscandian shield (Figures 1 and 2). Acquisition of data started in September 2001 and ended in June 2003. The measurement was done by the company Spetsgeofizika (Moscow) using five 10 ton vibrators, 360 groups of 12 geophones with a split-spread geometry of 18.4 km length and giving approximately a 90-fold coverage. The sweep signal was 30 s long with linearly increasing frequency from 12 to 80 Hz. In addition, two high-resolution profiles (with signal frequencies up to 120 Hz) were measured across two ore formations. Standard data processing using the PROMAX system was applied [*Kukkonen et al.*, 2006]. All sections were migrated using one-dimensional average velocity-depth function of SVEKA'81 [*Grad and Luosto*, 1987]. The FIRE1 profile generally follows the NE part of the SVEKA transect, while the FIRE2 profile is a continuation of the FIRE1 to the south along the FENNIA profile (Figures 1 and 2). The reflection profiles were measured along the existing roads, thus in some places they are 20–50 km apart from the DSS profiles.

4. P and S Wavefield

[12] The seismic data of high quality were obtained along all the investigated profiles [*Luosto et al.*, 1984; *Grad and Luosto*, 1987; *Luosto et al.*, 1994; *FENNIA Working Group*, 1998]. Three-component recordings were available for the FENNIA profile, while for the SVEKA transect only vertical component was available. Record sections were prepared using a band-pass filter of 2–15 Hz for P and 1–6 Hz for S waves. We found the vertical component as good as the horizontal ones for detection of S waves (Figures 3a, 3b, 3c, 3d, and 3e). The traveltimes of the first arrivals of refracted P and S waves for all shots

Figure 2. Location of refraction and wide-angle reflection deep seismic sounding profiles on the simplified tectonic map of the study area (adapted from *Korsman and Koistinen* [1998]; description of the legend after *Korsman and Koistinen* [1998]). Inset map shows location of the target area with respect to the East European Craton [after *Bogdanova et al.*, 2001]. Paleoproterozoic (Svecofennian Domain): CFGC, Central Finland Granitoid Complex; SA, Saimaa area; PoB, Pohjanmaa area; PiB, Pirkkala Belt; TaB, Tampere Belt; HaB, Hämeenlinna Belt; UuB, Usimaa Belt; SaB, Savo Belt; SaR, Satakunta Rapakivi area; Ss, Satakunta sandstone; WRM, Wyborg Rapakivi Massif; Archaean (Karelian Domain): EFC, Eastern Finland Complex; PC, Pudasjärvi Complex; IC, Iisalmi Complex; OA, Outokumpu area; KaB, Kainuu Belt; KuB, Kuhmo Belt. Other explanations are the same as for Figure 1.

are compared in Figure 4. Small differences in traveltimes of first arrivals between vertical and horizontal components visible on Figure 4a suggest weak anisotropy in the crust and upper mantle beneath the FENNIA profile, which agrees with the results by Plomerová *et al.* [2006]. Surface waves, very well visible on seismic record sections were interpreted by Grad and Luosto [1994].

[13] Both refracted and reflected P and S waves were used in 2-D ray tracing modeling. Resulting P and S wave velocity models and the V_p/V_s distribution are shown in Figure 5. Examples of the interpreted phases, synthetic seismograms, and the fit between the observed and calculated traveltimes are shown in Figures 6–10.

4.1. P_g and S_g Phases

[14] The P wave refracted from the crust that appears as the first arrival (P_g) has rather high apparent velocities of $6.0\text{--}6.9\text{ km s}^{-1}$ up to the offset of 235 km along all the profiles. Lower velocities of $5.75\text{--}5.9\text{ km s}^{-1}$ are observed only from few shots points at offset up to 50 km (SP_A, SP_D of the FENNIA profile and SP_D, SP_G of the SVEKA transect). Apparent velocities higher than 7 km s^{-1} are seen in most of the record sections at offsets from 185 km up to 290 km (SP_G and SP_I of the SVEKA'91 profile and all sections of the FENNIA and the SVEKA'81 profiles, except for shots located in central parts of the profiles). Usually, these velocities are $7.0\text{--}7.25\text{ km s}^{-1}$, but velocities of $7.5\text{--}7.75\text{ km s}^{-1}$ are observed at SP_I additionally to these values. At SP_B and SP_D only velocities of $7.5\text{--}7.75\text{ km s}^{-1}$ are observed. Variations of the apparent velocity of the first arrivals with an offset can be seen in many record sections (e.g., SP_A, SP_B of the FENNIA profile). Observed differences in arrival times of P_g do not exceed 1 s (Figure 4).

[15] It is usually difficult to separate the “first” arrival of S wave (S_g) from the background noise and P wave coda. However, it was possible to pick the S_g with almost the same accuracy as the P_g on most of the record sections. An apparent velocity of the S_g is $3.4\text{--}3.9\text{ km s}^{-1}$ at offsets up to 250 km for all profiles. Higher apparent velocities of $4.0\text{--}4.1\text{ km s}^{-1}$ are observed at offsets of 190–270 km at SP_A, SP_E, SP_F of the FENNIA profile and at SP_F, SP_G, SP_I, SP_D of the SVEKA transect.

4.2. Crustal Phases in Later Arrivals, PmP and SmS Phases

[16] Strong and clear phases reflected from the crust (PcP and ScS) were detected even at far offsets on record sections of SP_B and SP_C of the FENNIA profile and SP_A of the SVEKA transect. It was possible to find plenty of other arrivals of PcP and ScS , but they were either not clear or could be correlated only over short distances (SP_H, SP_R, and SP_S of the SVEKA transect) (Figure 3c).

[17] In many record sections the P wave reflection from the Moho boundary (PmP) cannot be detected as a sharp arrival. If seen, the PmP starts from offsets of 120–160 km. Particularly clear reflections from the Moho are recorded at SP_B, SP_E of the FENNIA profile and SP_S, SP_A, SP_D of the SVEKA transect.

[18] In contrast to P waves, the recordings of S waves demonstrate clearer or similar quality of reflections from the Moho (SmS). The phase is clearly visible in all of the longest seismic sections, namely, SP_A, SP_F of the FENNIA profile and SP_F, SP_K, SP_A, SP_E of the SVEKA transect.

[19] Crustal waves continue as strong phases into over critical distances up to the offset of 250–350 km in many cases (e.g., SP_E, SP_K of the SVEKA transect).

4.3. Pn , Sn , and Reflected Mantle Phases

[20] The P waves refracted beneath the Moho boundary (Pn) with apparent velocities of $8.0\text{--}8.5\text{ km s}^{-1}$ are weak, but they are clearly visible at larger offsets (>250 km) in some of the longest sections (see, e.g., Figures 3a, 3b, 3c, 3d, 3e and 4).

[21] Amplitudes of S waves refracted beneath the Moho boundary (Sn) are comparable to those of Pn , but a reliable determination of Sn is difficult, if possible at all, due to higher level of background noise. Anyway, the Sn phases with apparent velocities of $4.5\text{--}4.8\text{ km s}^{-1}$ were picked on all the longest branches of record sections from the offset of 245 km (SP_A, SP_F of the FENNIA profile and SP_F, SP_K, SP_A, SP_E of the SVEKA transect) (see, e.g., Figures 3a, 3b, 3c, 3d, 3e and 4).

[22] Strong reflections from the upper mantle (P_1P and S_1S) are seen in record sections of SP_A, SP_E, SP_F of the FENNIA profile and SP_F, SP_G, SP_I, SP_K, SP_B, SP_E of the SVEKA transect at offsets of 170–230 km and 360–470 km (see, e.g., Figures 3s, 3b, 3c, 3d, 3e). Mantle refracted waves arriving at reduced times of 7–9 s were observed at offsets of 250–270 to 350 km.

5. Seismic Modeling by Forward Ray Tracing

[23] The 2-D forward ray-tracing modeling was undertaken to obtain the P and S wave velocity models of the crust along the SVEKA transect and the FENNIA profile. A total of 12 and 5 seismic record sections and traveltimes of all refracted and reflected P and S waves, respectively, were used. Starting models were based on our data analysis and results of the previous studies. The method used was iterative, manual traveltime fitting, using the ray-tracing package SEIS83 [Červený and Pšenčík, 1983], enhanced by the graphical user interface programs MODEL and XRAYs [Komminaho, 1998] and ZPLOT [Zelt, 1994; Šroda, 1999]. The forward modeling proceeded so that the synthetic traveltimes are calculated and compared to the observed

Figure 3a. Example of trace-normalized, vertical component seismic record sections for P and S waves along the FENNIA profile filtered by band-pass filters (2–15 and 1–6 Hz). P_g , S_g , seismic refractions from the upper and middle crystalline crust; $P_{7.25}$, refractions from high-velocity lower crust; P_{ov} , overcritical reflections; PcP , ScS , reflections from the middle crustal discontinuities; PmP , SmS , reflections from the Moho; Pn , Sn , refractions from the sub-Moho upper mantle; P_1P , P_2P , P_3P , S_1S , S_2S , S_3S , P and S wave phases from the upper mantle; P_MP and P_n , reflected and refracted waves from the Moho boundary, respectively. The reduction velocities are 8.0 and 4.52 km s^{-1} for P and S waves, respectively.

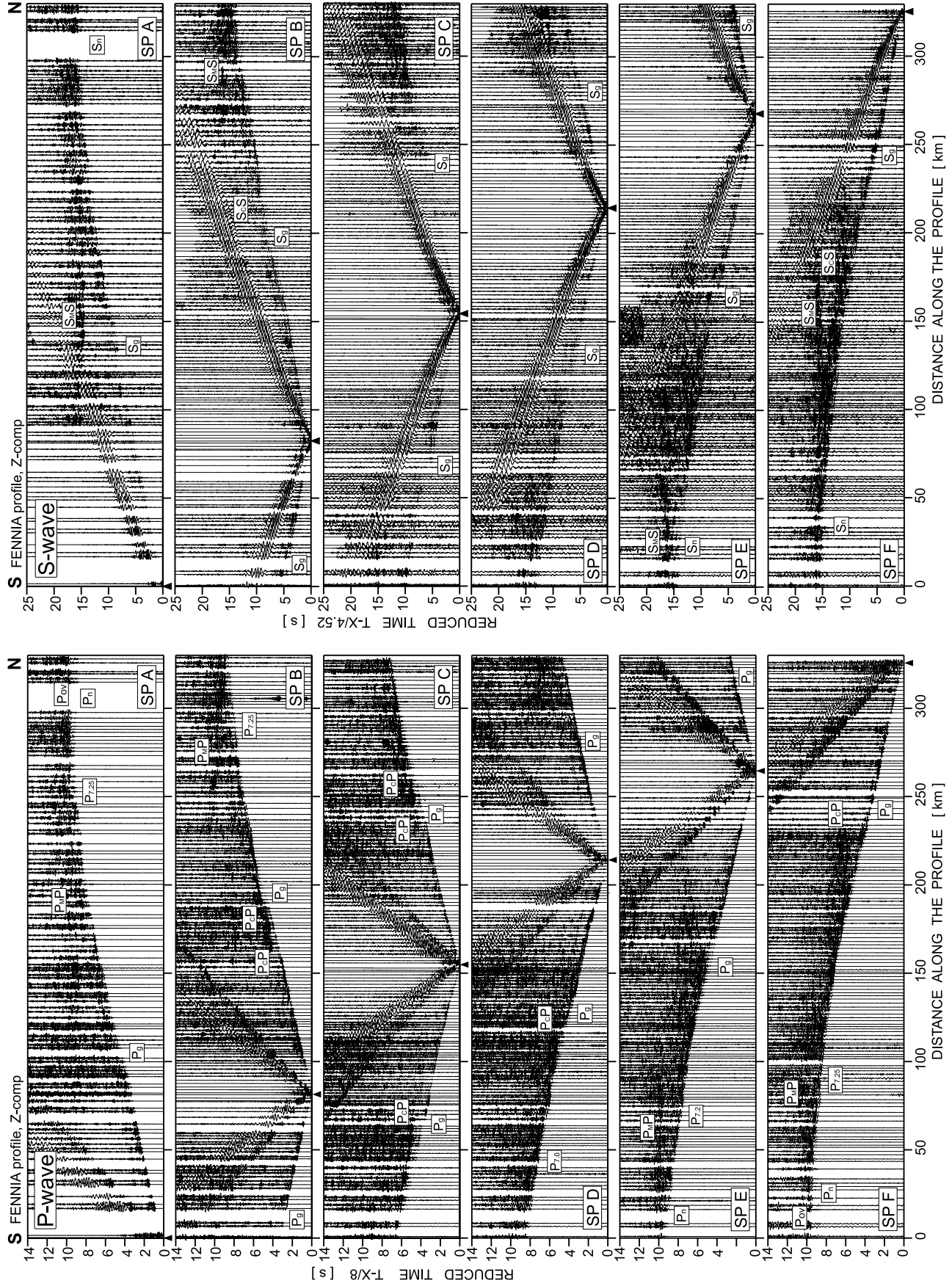
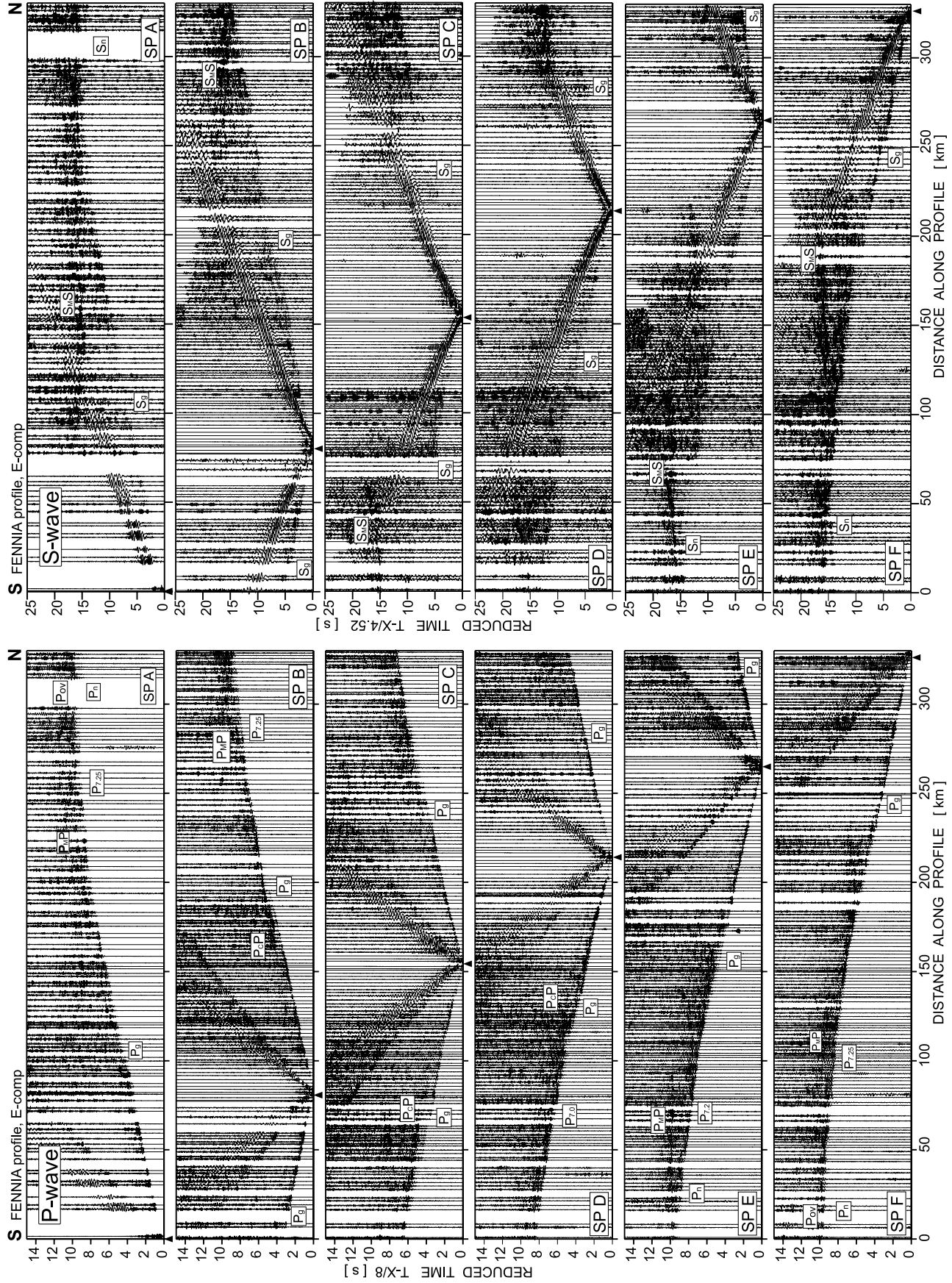


Figure 3a



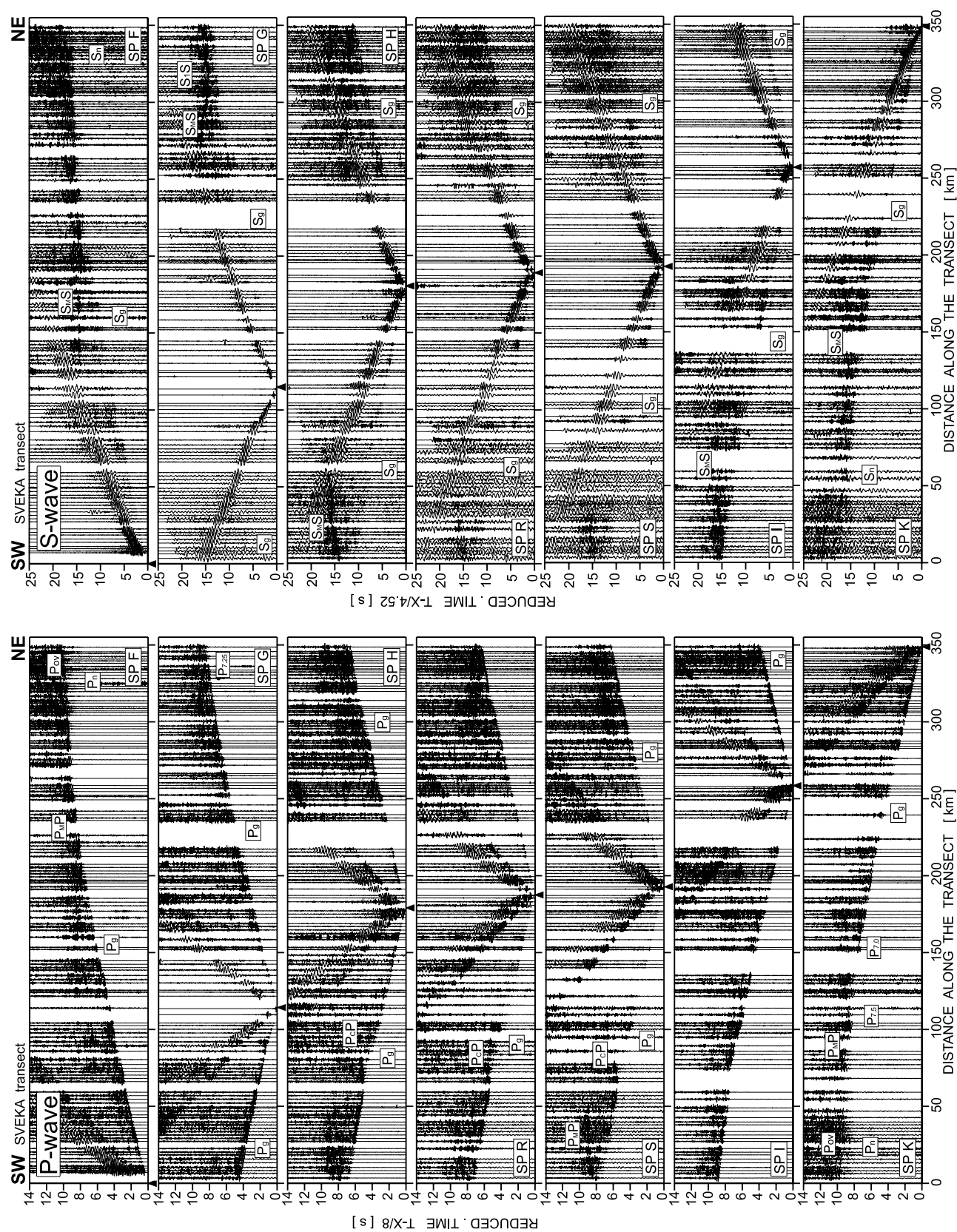


Figure 3c. Example of trace-normalized, vertical component seismic record sections for P and S waves along SVEKA'91 profile. Parameters of the band-pass filters and abbreviations are the same as on Figure 3a.

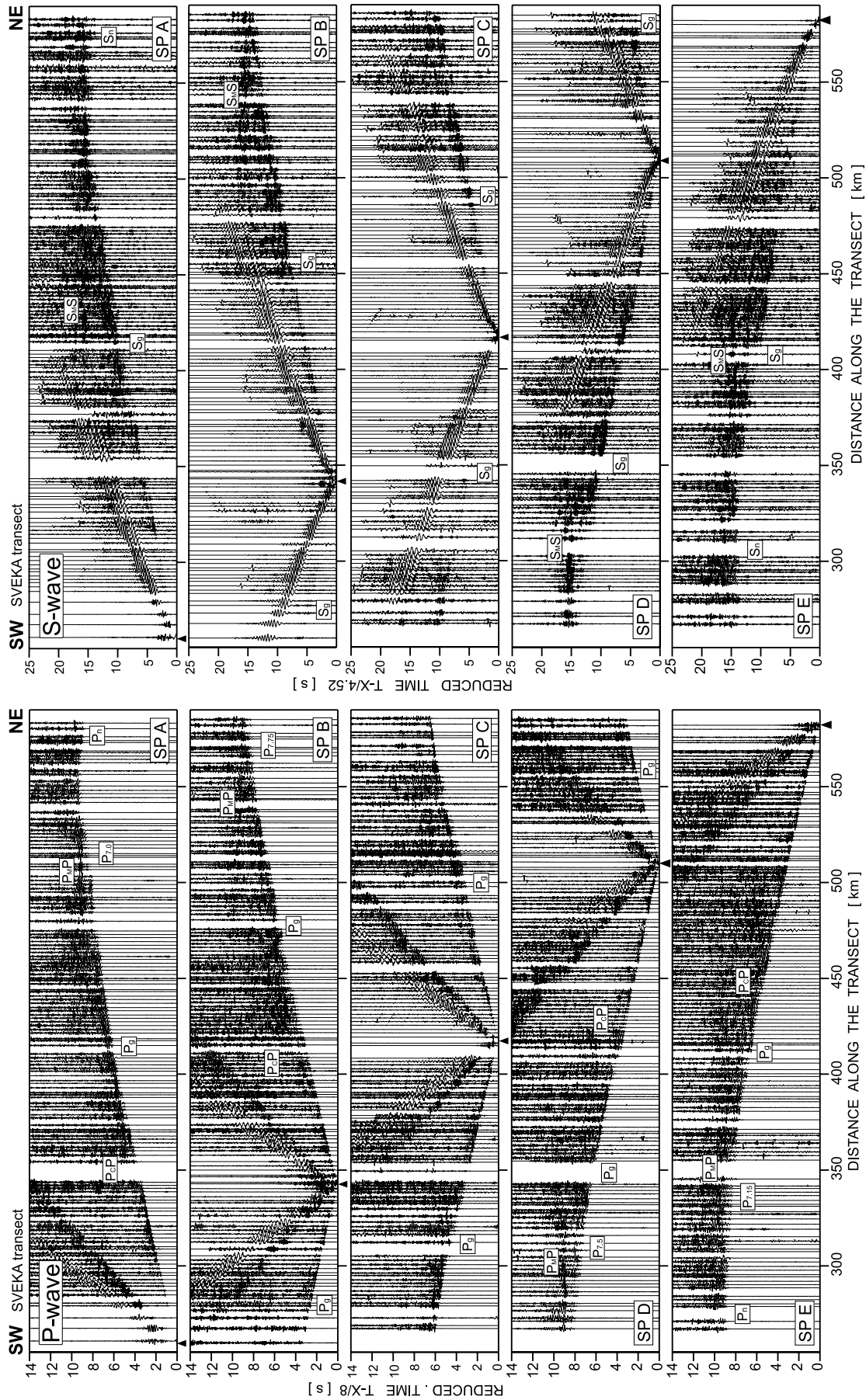


Figure 3d. Example of trace-normalized, vertical component seismic record sections for P and S waves along SVEKA'81 profile. Parameters of the band-pass filters and abbreviations are as on Figure 3a.

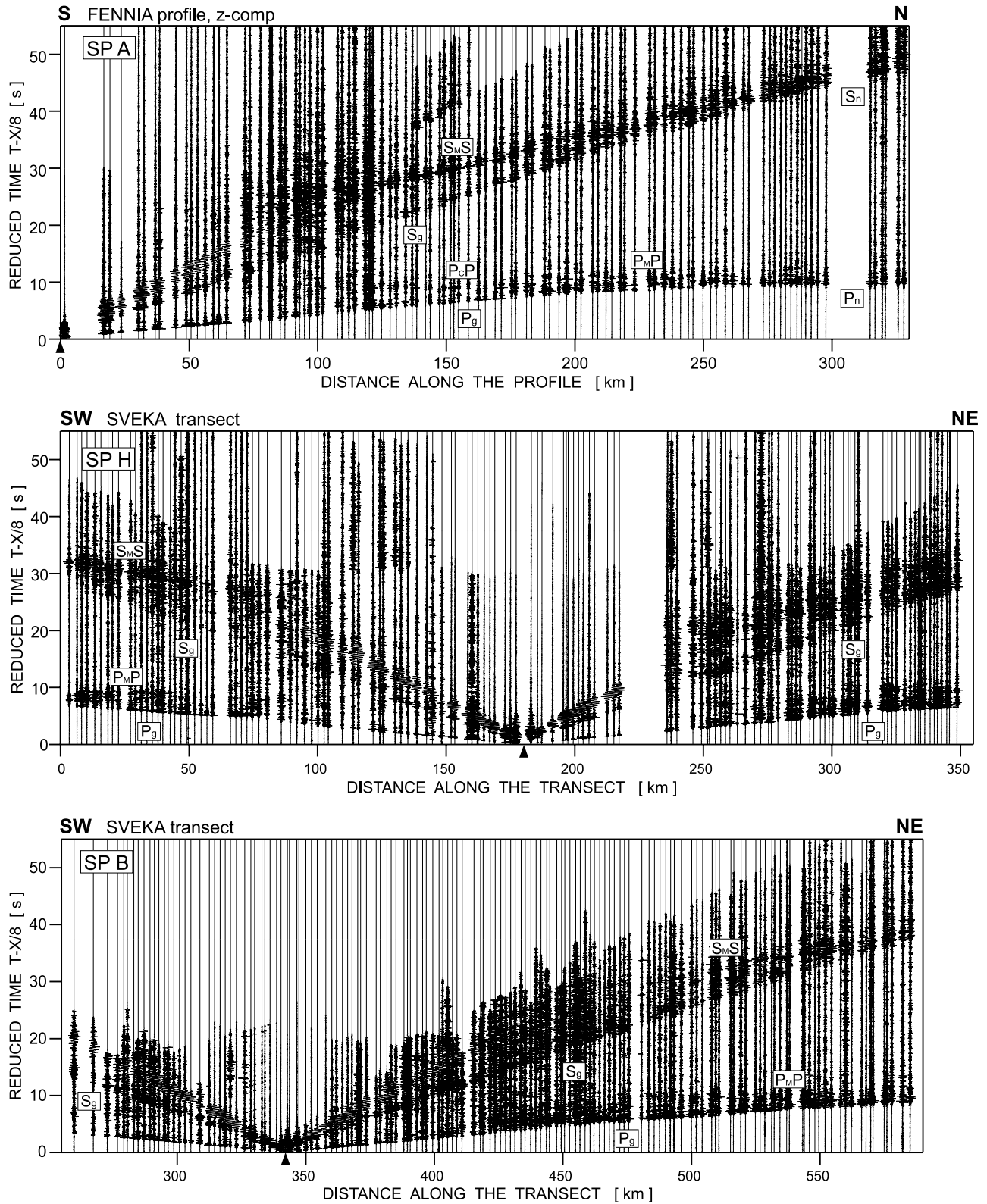


Figure 3e. Example of trace-normalized, vertical component, common seismic record sections for P and S waves along SVEKA transect (SP_H, SP_B) and FENNIA profile (SP_A). A band-pass filters (2–15 Hz). Abbreviations are as on Figure 3a.

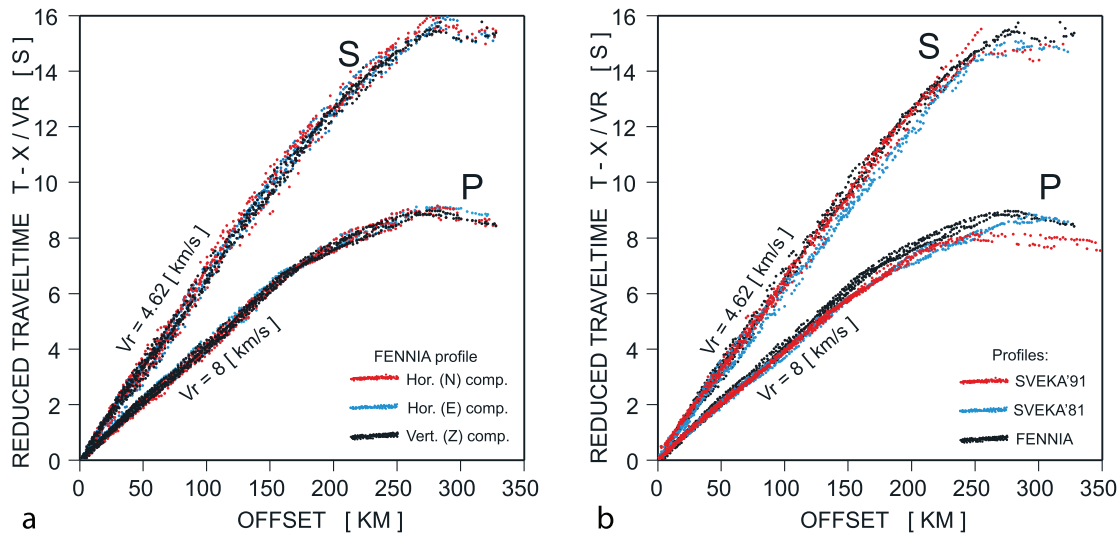


Figure 4. (b) P and S waves first arrival traveltime picks along the SVEKA'91 (red dots), SVEKA'81 (blue dots) and FENNIA (black dots) profiles. Highest differences, >1 s are visible for P_n waves. It suggests differences of the crustal structure and thickness along these profiles. Reduction velocities are 8.0 and 4.62 km s^{-1} for P and S waves, respectively. (a) Comparison of P and S waves first arrival traveltime picks of vertical (black) and two horizontal (north and east components are marked by red and blue colors, respectively) components along FENNIA profile. High differences are not observed. Reduction velocities are 8.0 and 4.62 km s^{-1} for P and S waves, respectively.

traveltimes and the model is successively altered at each iteration step. This procedure was continued until an agreement of the order of 0.1–0.2 s was achieved between the observed and model-derived traveltimes. During the modeling, the depth to the velocity discontinuities and velocity values were compared at the crossing of the SVEKA transect and the FENNIA profile. The derived 2-D velocity models are shown in Figure 5.

[24] In addition, synthetic seismograms of P waves were calculated, in order to control velocity gradients within the layers and velocity contrast at the seismic boundaries. The final synthetic seismograms show good qualitative agreement with the relative amplitudes of the observed wavefield. Examples of ray-tracing modeling are shown in Figures 6–10.

[25] In contrast to earlier interpretations of the SVEKA'81 [Grad and Luosto, 1987], SVEKA'91 [Luosto et al., 1994] and FENNIA [FENNIA Working Group, 1998], which neglected lateral variations of V_p/V_s ratio, we also investigated lateral variations of V_p/V_s inside the layers.

5.1. Velocity Model for the SVEKA Transect

[26] Figure 5 (left) presents the final models generated by our 2-D forward ray-tracing modeling along the SVEKA transect.

5.1.1. P Wave Velocity Model

[27] The most complicated structure was found in the upper crust, down to the depth of 10–15 km. The velocity at the surface is usually ≥ 6.0 km s^{-1} . A thin layer with velocity of about 5.8 km s^{-1} was revealed beneath SP_D and SP_G. An area with the same velocity down to the depth of 2 km was detected between SP_F and SP_G.

[28] A low-velocity layer (LVL) with velocities of 6.05–6.1 km s^{-1} , ~ 5 km thick, was found at depths of 5–14 km at a distance of 20 km from the end of the transect. Its thickness is higher (up to 7 km) in the southwest (40–240 km). Together with the part of a deeper layer, that is slightly thicker, but has similar geometry, the LVL forms characteristic “trough” structure continuing to the depth of 18 km. The inclined boundaries limiting this structure are documented by very strong PcP phases (see SP_H, SP_R, SP_S). To the left of it, the velocity of 6.3 km s^{-1} was detected at the depth of 5 km, which is the highest value for the uppermost crust. Depth to the deeper layer in the upper crust with velocities of 6.25–6.35 km s^{-1} varies from 12 to 18 km along the whole transect.

[29] The middle crust, which we define as a layer with velocities of 6.45–6.65 km s^{-1} , is relatively thin, ~ 7 –12 km,

Figure 5. Two-dimensional models of seismic P and S wave velocity in the crust and upper mantle derived by forward ray-tracing modeling using the SEIS83 package [Červený and Pšenčík, 1983] along (left) the SVEKA transect and (right) the FENNIA profile. (a) P wave velocity models. Thick, black lines represent major velocity discontinuities (interfaces). (b) S wave velocity models. Those parts of the first-order discontinuities that have been constrained by reflected or/and refracted arrivals of S waves are marked by thick lines. (c) Models of V_p/V_s ratio distribution. Those parts of the first-order discontinuities that have been constrained by reflected or refracted arrivals of P or/and S waves are marked by thick lines. Thin lines represent velocity isolines with values in km s^{-1} shown in white boxes. Position of large-scale crustal blocks is indicated. Arrows show positions of shot points. The crossing points of the SVEKA transect with the FENNIA profile are marked by blue arrows.

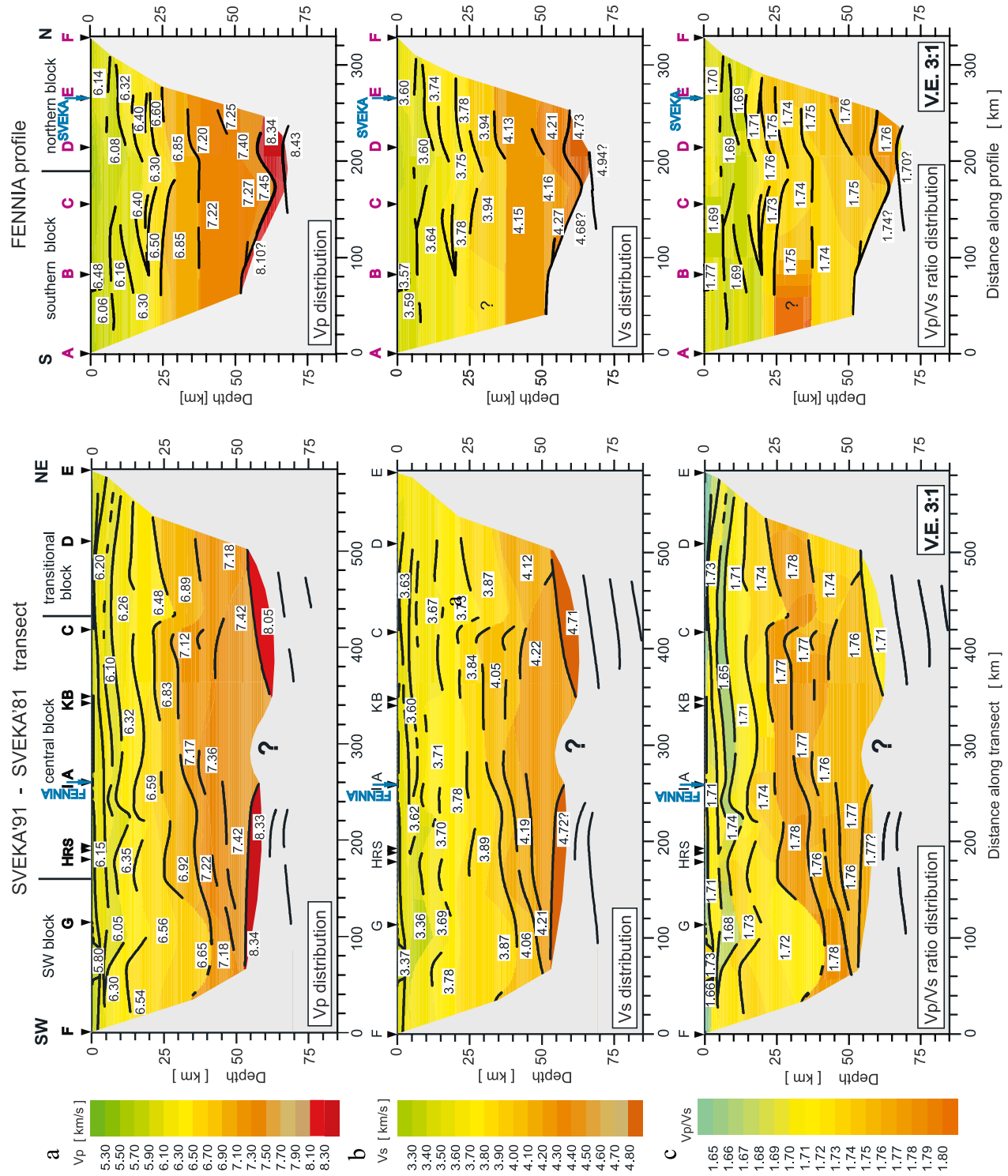


Figure 5

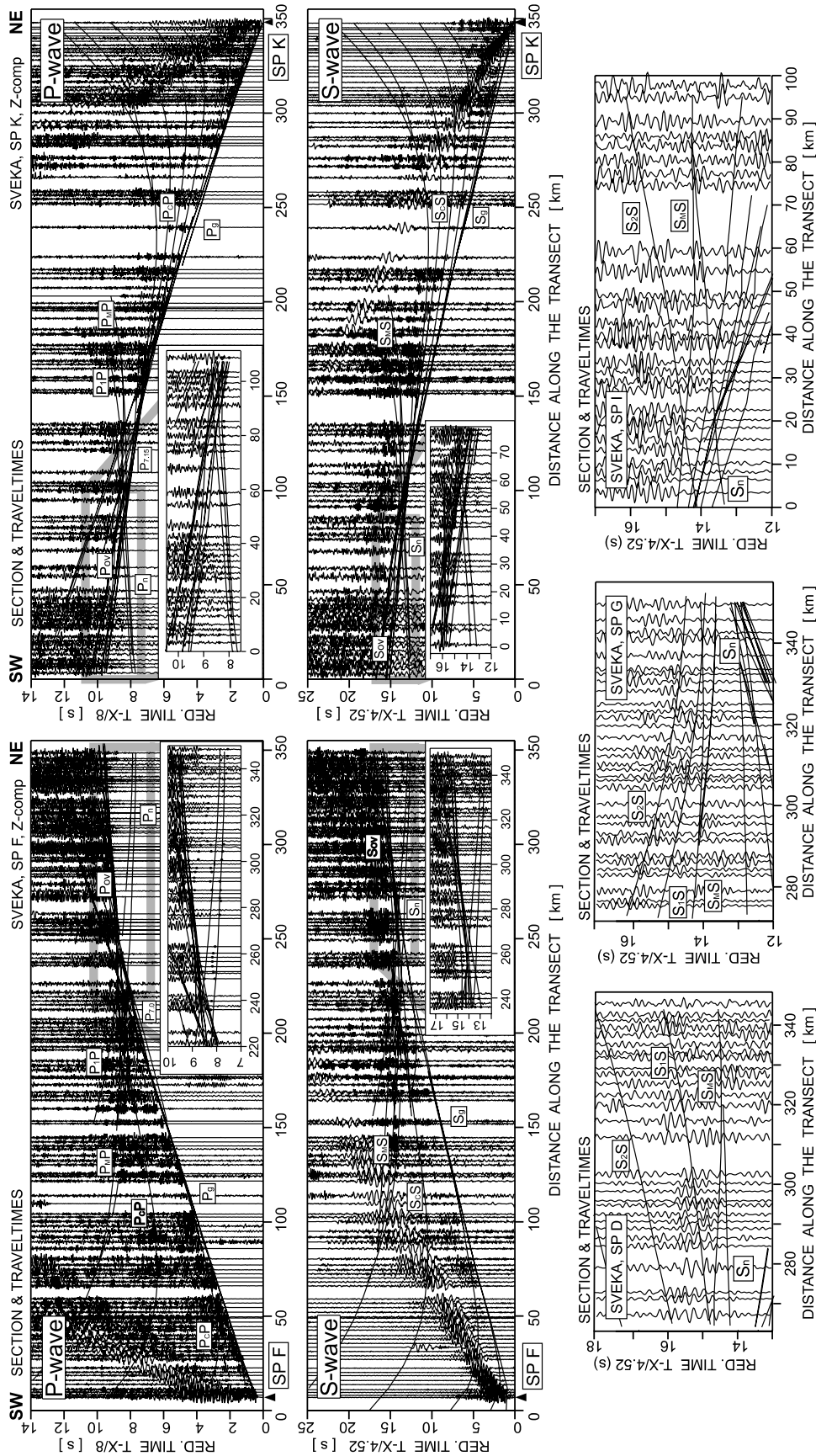


Figure 6. Examples of amplitude-normalized vertical component record sections with theoretical traveltimes derived using the SEIS83 ray-tracing technique for SP_F, SP_K, SP_D, SP_I, and SP_B of the SVEKA transect. (top) The data have been filtered using the band-pass filter of 2–15 Hz and displayed using the reduction velocity of 8.0 km s^{-1} for *P* waves, (middle and bottom) For *S* waves we used the band-pass filter of 1–6 Hz and reduction velocity 4.52 km s^{-1} . All examples were calculated for the model presented on Figure 5. Other abbreviations are as on Figures 3a, 3b, 3c, 3d, 3e.

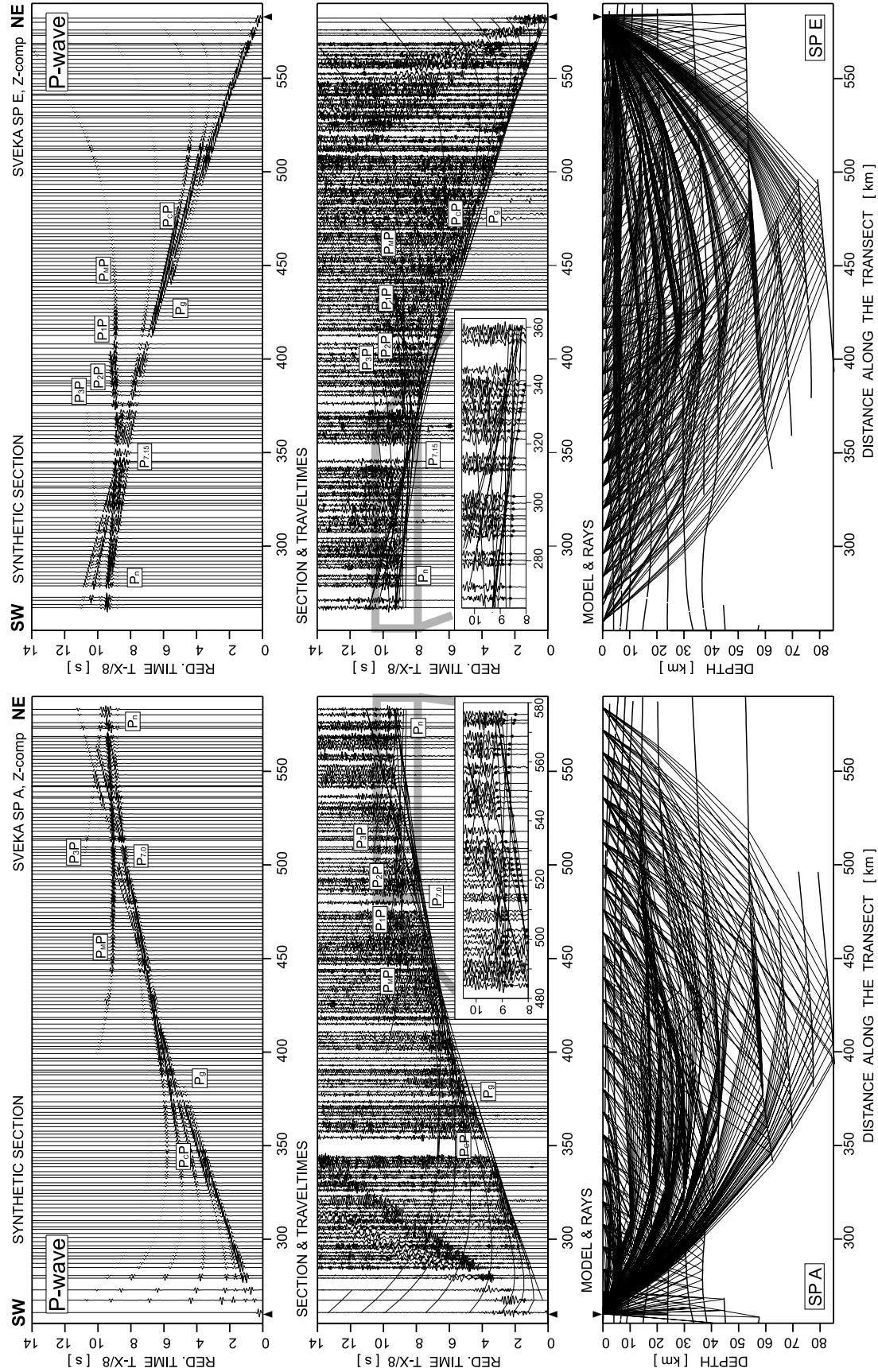


Figure 7. Examples of results derived using the SEIS83 ray-tracing technique for SP_A and SP_E of SVEKA transect: synthetic seismograms and amplitude-normalized seismic record section with theoretical traveltimes and ray diagram of P waves. Reduction velocity is 8.0 km s^{-1} . Other abbreviations are as on Figures 3a, 3b, 3c, 3d, 3e and 6.

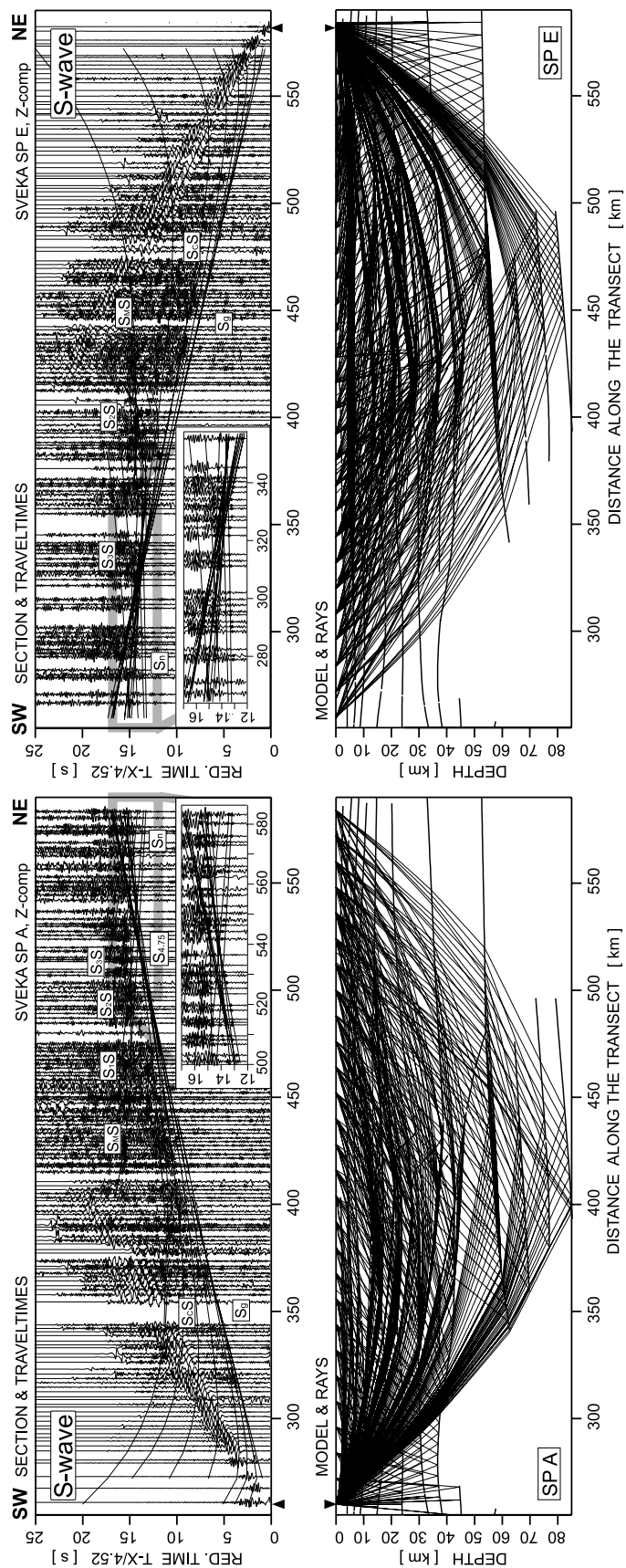


Figure 8. Amplitude-normalized seismic record sections and theoretical traveltimes and ray diagram of S waves (SVEKA, SP_A SP_E). Reduction velocity is 4.52 km s^{-1} . Other abbreviations are as on Figures 3a, 3b, 3c, 3d, 3e and 6.

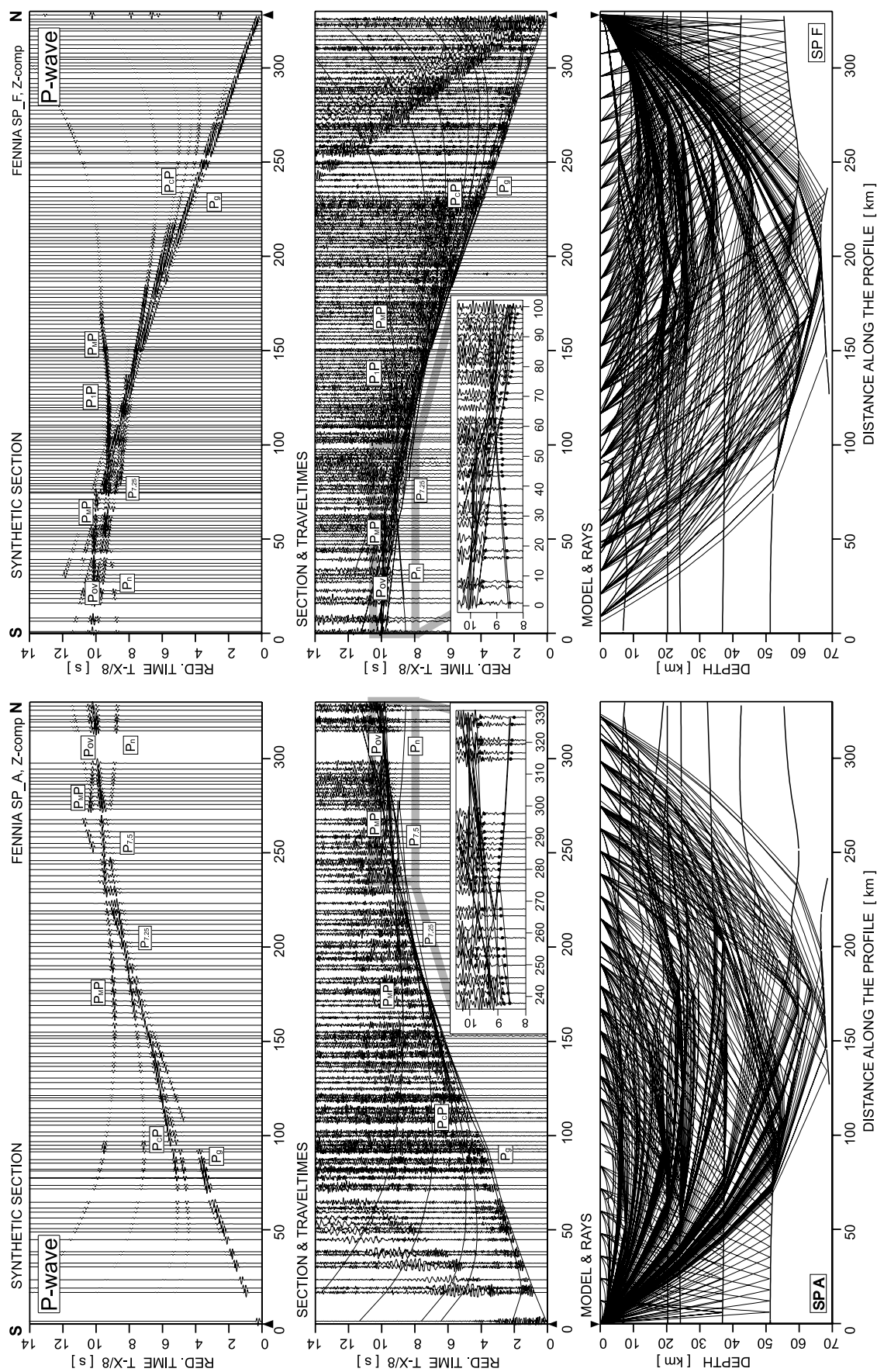


Figure 9. Examples of results derived by ray-tracing modeling for SP_A and SP_F of the FENNIA profile: synthetic seismograms and amplitude-normalized seismic record section with theoretical traveltimes and ray diagrams of P waves. Reduction velocity is 8.0 km s^{-1} . Other abbreviations are as on Figures 3a, 3b, 3c, 3d, 3e and 6.

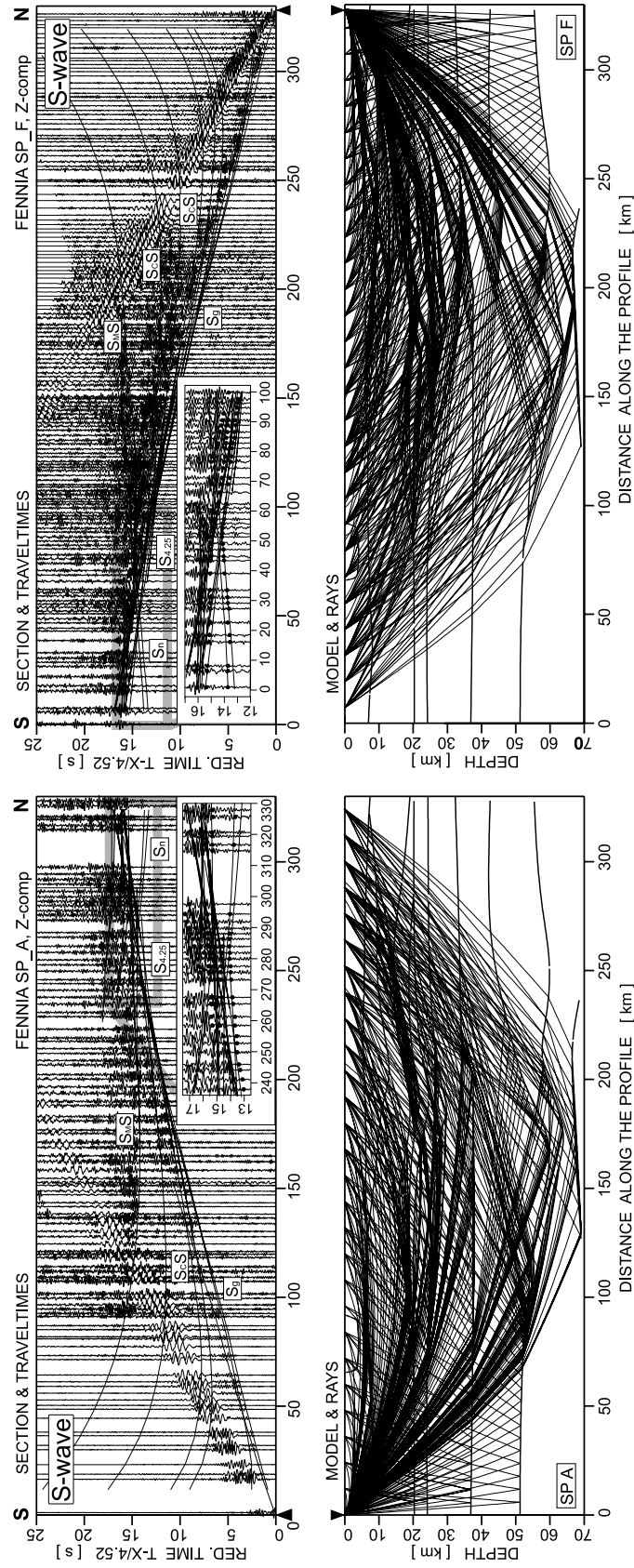


Figure 10. Amplitude-normalized seismic record sections, theoretical traveltimes and ray diagrams of *S* waves (FENNIA, SP_A and SP_F). Reduction velocity is 4.52 km s^{-1} . Other abbreviations are as in Figure 5. All examples were calculated for the model of the crust presented on Figure 5 km s^{-1} . Other abbreviations are as on Figures 3a, 3b, 3c, 3d, 3e and 6.

except for the southwestern part (up to the distance of 120 km), where it is extremely thick, ~ 30 km.

[30] The thickness of the third, deeper layer with velocities of $6.8\text{--}7.4\text{ km s}^{-1}$, defined as the lower crust, varies strongly, from ~ 12 km in the SW part to $25\text{--}30$ km in the central and NE parts of the transect. Several more or less documented reflectors divide the lower crust into three layers. Uppermost, ~ 10 km thick layer with the velocities of $6.8\text{--}6.9\text{ km s}^{-1}$ well constrained by refracted waves, was detected at the depth of ~ 25 km in the central and NE parts of the profile (Figure 5). A deeper layer of thickness of $5\text{--}15$ km and relatively high velocities of $7.1\text{--}7.2\text{ km s}^{-1}$ was observed along the whole transect. We denote this layer as a high-velocity lower crust of the first type (HVLC1). The velocity in this layer is well constrained by refracted waves recorded in the NW part of the transect (SP_A, SP_B, SP_D, SP_F) and overcritical strong phases recorded in its southeastern part (SP_I). The higher crustal velocities exceeding 7.4 km s^{-1} were detected up to distance of 460 km in the underlying layer that we define as a high-velocity lower crust of the second type (the HVLC2 in Figure 5). The thickness of this layer increases to ~ 15 km in the center of the transect. Velocities in this layer were constrained only by strong overcritical phases (SP_F and SP_K), however.

[31] The depth to the Moho boundary increases from $52\text{--}55$ km in southeastern and northeastern ends of the model to $57\text{--}62$ km in its central part. Because of the lack of the data, the structure of the lowermost crust and upper mantle in the central part of the transect is not resolved. Very strong *PmP* phase is visible at the distance from 360 to 440 km in all record sections from this part of the transect. The reflections are not so clear starting from 440 km and further to the NE. At the distance from 80 to 230 km the interpretation of *PmP* arrivals from different shotpoints is ambiguous.

[32] The sub-Moho velocities are constrained by *Pn* and vary from $\sim 8.3\text{ km s}^{-1}$ in the SW (SP_F, SP_K; Figure 6), to more than 8.1 km s^{-1} in the NE of the transect (SP_A, SP_E; Figure 7).

[33] A few *P* wave reflections from the upper mantle (*P₁P*, *P₂P*, *P₃P*) were detected at the distances of $170\text{--}230$ km and $340\text{--}470$ km (Figures 3a, 3b, 3c, 3d, 3e and 6–10). The velocities beneath these reflectors are constrained only by amplitudes of the reflection waves.

[34] The *P* wave velocity model generally agrees with the previous models of the SVEKA'81 [Grad and Luosto, 1987; Kozlovskaya and Yliniemi, 1999] and SVEKA'91 [Luosto et al., 1994]. The largest difference was found in the southwestern part, where the middle crustal velocities are 0.1 km s^{-1} lower compared to the previous model by Luosto et al. [1994]. This reduced velocity was necessary to achieve the delay of about 0.5 s for overcritical phases recorded from SP_A at the distance of ~ 250 km, which agrees better with the data.

5.1.2. *S* Wave Velocity Model and V_p/V_s Ratio

[35] It was unexpected that most of the detected *P* phases had correspondent *S* phases. Traveltimes of correlated reflected and refracted *S* waves were used to estimate the *S* wave velocity model (Figure 5), from which the distribution of V_p/V_s ratio was calculated for the principal layers of the crust and the upper mantle (Figure 5). The geometry of the first-order discontinuities in the *S* wave velocity model was inherited from the *P* wave velocity model (Figure 5).

[36] *S* wave velocity has the values of $3.6\text{--}3.7\text{ km s}^{-1}$ beneath the surface along the transect, except for $V_s \sim 3.35\text{ km s}^{-1}$ in a basin-shape area with $V_p \sim 5.8\text{ km s}^{-1}$. These values indicate V_p/V_s ratio of $1.71\text{--}1.7$ for most parts of the upper crust. The lowest values of V_p/V_s ratio in the model ($1.65\text{--}1.66$) were found in the LVL and also in the uppermost crust in the SW part of the transect. In the middle crust, V_s varies from 3.73 to 3.87 km s^{-1} , but values of V_p/V_s ratio vary from 1.71 to 1.74 .

[37] The *S* wave velocities in the lower crust are of $3.85\text{--}3.90\text{ km s}^{-1}$, $\sim 4.05\text{ km s}^{-1}$ or 4.1 km s^{-1} , and $\sim 4.2\text{ km s}^{-1}$, respectively, for the layers in the lower crust defined earlier from the *P* wave velocity model. These values indicate V_p/V_s ratio of $1.76\text{--}1.78$ for all these layers with one exception. Namely, the V_p/V_s ratio of 1.74 was obtained in the high-velocity lower crust in the northeastern part of the transect. Basing on this relatively low value of the V_p/V_s , we define this layer as a high-velocity lower crust of the third type (HVLC3).

[38] In the upper mantle, modeled *S* wave velocities are similar for the whole transect ($4.71\text{--}4.72\text{ km s}^{-1}$), but this value is not well constrained in the SW part because of less clear arrivals of *Sn*. In contrast to V_s , values of V_p/V_s ratio are different, 1.77 for SW and 1.71 for NE parts of the transect (Figure 5). *S* wave reflections from the upper mantle (*S₁S*) are observed at the same or even larger offsets (in particular, in the NE of the transect) as the correspondent *P* wave reflections (e.g., Figures 6–8).

[39] We observed very strong *SmS* phase from this part of the transect in the same distance range (from 360 to 440 km) as the *PmP*. From 440 km to the NE the *SmS* reflections from some of the shotpoints are not clear. Similar to *PmP*, the interpretation of *SmS* arrivals from different shotpoints is ambiguous at the distance from 80 to 230 km.

[40] A few *S* wave reflections from the upper mantle (*S₁S*, *S₂S*, *S₃S*) were detected particularly for distances of $170\text{--}230$ km and $340\text{--}470$ km (e.g., Figures 3c, 3d, 3e, 6, and 8). *S₁S* reflections are stronger than *SmS* for some shotpoints (SP_B, SP_G, SP_I, see Figure 6).

[41] The values of *P* and *S* wave velocities and V_p/V_s ratio generally agree with the previous models of the SVEKA'81 profile [Yliniemi et al., 1996; Kozlovskaya and Yliniemi, 1999]. The Moho boundary beneath the central part of the transect is a little deeper than in the previous interpretations by Grad and Luosto [1987] and Luosto et al. [1994], but this agrees well with the result of receiver function study by Alinaghi et al. [2003].

5.2. Velocity Model for the FENNIA Profile

[42] Figure 5 (right) presents the final velocity models of the FENNIA profile generated by our 2-D forward ray-tracing modeling.

5.2.1. *P* Wave Velocity Model

[43] Similarly to the *P* wave velocity model of the SVEKA, the most complicated structure was found in the upper crust, which extends to the depths of 15 km in northern and 20 km in southern parts of the profile. The velocity increases from 5.9 km s^{-1} and $6.05\text{--}6.15\text{ km s}^{-1}$ at the surface to 6.3 km s^{-1} at the depth of 20 km. A low-velocity layer (LVL) with velocity of $6.05\text{--}6.1\text{ km s}^{-1}$, ~ 5 km thick, was found at the depth of $5\text{--}14$ km at the distance of $210\text{--}300$ km. In the southern part of the profile,

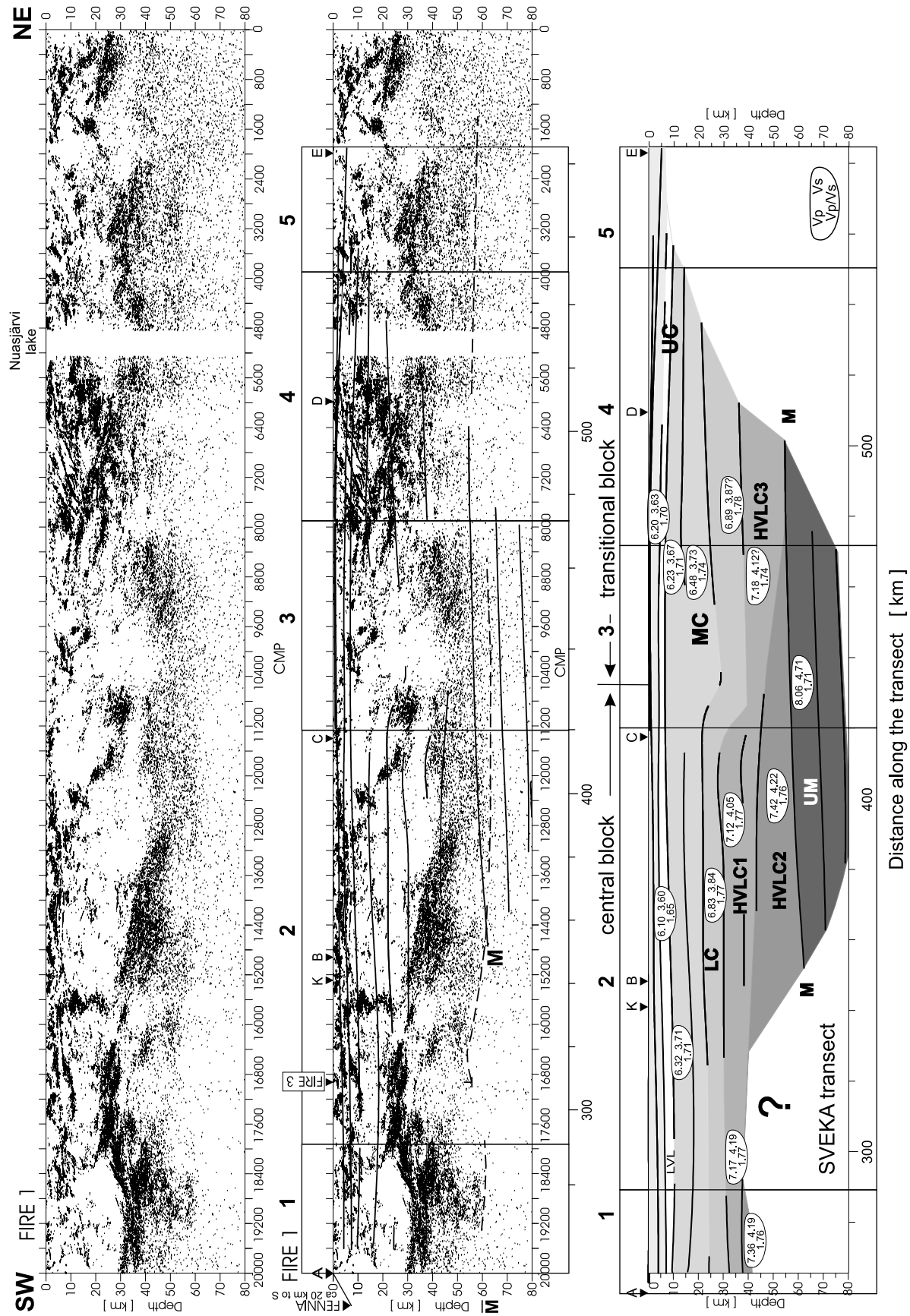


Figure 11a

near SP_B, a high-velocity body (HVB) with $V_p \sim 6.4 \text{ km s}^{-1}$ was detected.

[44] The middle crust, with velocity of $6.4\text{--}6.6 \text{ km s}^{-1}$, is relatively thin, $\sim 10 \text{ km}$. A low-velocity layer (LVL) with velocity of 6.3 km s^{-1} was detected at a depth of 20 km . This LVL corresponds to the “shadow zone” and also to a strong reflector visible in record section of SP_B at the offset of $\sim 200 \text{ km}$.

[45] The velocities in the lower crust are constrained by refracted waves recorded from SP_A, SP_B, SP_D (Figures 3c and 9). However, the velocity in the deepest part of the lower crust was controlled only by strong overcritical phases observed on SP_E. According to the values of P wave velocity, the lower crust can be divided into two layers. In the upper part, at depths of $25\text{--}37 \text{ km}$, velocity of $\sim 6.85 \text{ km s}^{-1}$ was revealed. The velocity in the lowermost layer (HVLC) is relatively high ($7.2\text{--}7.45 \text{ km s}^{-1}$). In contrast to the HVLC in the model of the SVEKA transect, the HVLC beneath the FENNIA profile is rather transparent: only few reflectors were detected.

[46] The depth to the Moho varies significantly along the profile, from 51 km in the south to $\sim 63 \text{ km}$ at the distance of 180 km and to 61 km at the northern end of the model. Generally, all crustal discontinuities are shallower in the northern part of the profile, starting from distances of $\sim 200 \text{ km}$, than the same discontinuities in its southern part.

[47] Sub-Moho P wave velocities vary from $\sim 8.1 \text{ km s}^{-1}$ in the southern part ($70\text{--}140 \text{ km}$) (Figure 9, SP_A) to $>8.3 \text{ km s}^{-1}$ in the northern part (Figure 9, SP_F).

[48] The values of P wave velocities and the depth to the Moho boundary generally agree with the previous model [FENNIA Working Group, 1998]. However, the difference from the previous models mainly appears in the vicinity of crossing with the SVEKA transect. The shape of the Moho boundary slightly differs from the previous model.

5.2.2. S Wave Velocity Model and V_p/V_s Ratio

[49] Similar technique, as was described in section 5.1.2, was used to estimate the S wave velocity model and the V_p/V_s ratio distribution (Figure 5). The S wave velocity is $3.57\text{--}3.74 \text{ km s}^{-1}$ in the upper crust. These values result in V_p/V_s ratio of $1.69\text{--}1.7$ in most parts of the upper crust. The highest value of V_p/V_s ratio (1.77) was detected in the HVB. The V_s varies from 3.75 to 3.78 km s^{-1} in the middle crust, while the values of V_p/V_s ratio vary from 1.71 to 1.76 .

[50] The S wave velocities in the lower crust are 3.94 km s^{-1} and 4.15 or $\sim 4.2 \text{ km s}^{-1}$, respectively, for layers determined from the P wave velocity model. These values indicate the V_p/V_s ratio of $1.74\text{--}1.76$ in all these layers. Velocity in the upper mantle varies from $\sim 4.68 \text{ km s}^{-1}$ in the southern part ($100\text{--}140 \text{ km}$) (Figure 5 and 9, SP_A) to 4.73 km s^{-1} in the northern part (Figures 5 and 9, SP_F). The SmS reflections are of same or better quality than PmP for all shotpoints where they are recorded.

5.3. Uncertainties of the Models Obtained by Ray Tracing

[51] For kinematic modeling, a consistent approach of carefully matching traveltimes of refracted and reflected phases was used. The time differences at reciprocal points do not exceed 100 ms , which is the uncertainty of picked arrival times of phases refracted from the upper crust (Pg and Sg). The uncertainty of picking of deeper phases is larger; and we estimate it to be of the order of 150 ms for phases refracted from the middle and lower crust as well as for PmP and SmS . The uncertainty of picking of intracrustal reflections, Pn and Sn we estimate to be of about 200 ms . As we pointed out earlier, uncertainty of Sn in the SVEKA'91 is larger than that of Pn .

[52] The quality of the data and accuracy of 2-D forward ray trace modeling of the SVEKA transect and FENNIA profile are comparable or even better than those achieved in the modeling of the POLONAISE'97, CELEBRATION 2000 and EUROBRIDGE'97 wide-angle reflection and refraction profiles, for which analyses of accuracy were presented by Janik *et al.* [2002], Grad *et al.* [2003], Thybo *et al.* [2003], and Šroda *et al.* [2000]. Uncertainties of velocities and of depths to the boundaries are of the order of $\pm 0.1 \text{ km s}^{-1}$ and $\pm 1 \text{ km}$, respectively, for the parts of models with a relatively simple structure and clear phase correlations. For the areas with complicated structure, the accuracy of velocities and depths should be about $\pm 0.2 \text{ s}$ and $\pm 2 \text{ km}$, respectively, but V_p/V_s ratio was defined with the accuracy of $\pm 0.01\text{--}0.02$.

[53] The 2-D modeling does not take out-of-plane refractions and reflections into account. The latter can affect the seismic wavefield in the structurally complex areas, which we revealed in the centers of both investigated lines. In addition, many of details in the crust and upper mantle were not resolved because of the large space between shots and not sufficient lengths of profiles.

6. Discussion

6.1. Correlation of Velocity Models With Major Geological Units in the Crust

[54] Pronounced lateral variations of V_p , V_s , and V_p/V_s in the crust along the SVEKA transect and of the depth to the Moho boundary can be associated with three large blocks of the crust, marked as SW, central and transitional in Figure 5.

[55] The SW block is composed of three layers. The upper crust (UC) is laterally inhomogeneous, as can be judged from the values of V_p , V_s , and V_p/V_s . The low-velocity region ($V_p \sim 5.8 \text{ km s}^{-1}$) in the uppermost crust corresponds to Jotnian sedimentary rocks (Figure 2), but the layer in SW with $V_p \sim 6.3 \text{ km s}^{-1}$ is probably associated with a rapakivi granite intrusion exposed at the surface

Figure 11a. Seismic reflection profiles: FIRE1 and FIRE2 with the velocity model along NE part of the SVEKA transect, superimposed. There is general agreement between the models, regarding the shape and thickness of the lower crust and the Moho depth. The reflection profiles are located semiparallel, from -20 to $+50 \text{ km}$ apart the DSS profiles; hence one should not expect complete correspondence. (top) Raw seismic section, (middle) velocity models overlaid on CDP sections, and (bottom) P and S wave velocity model with V_p/V_s distribution. Dashed lines represent our interpretation of Moho boundary for FIRE1 section. HVLC(1,2,3), different layers of high-velocity lower crust (see results of ray trace modeling); LC, lower crust; MC, middle crust; UM, upper mantle; M, Moho boundary.

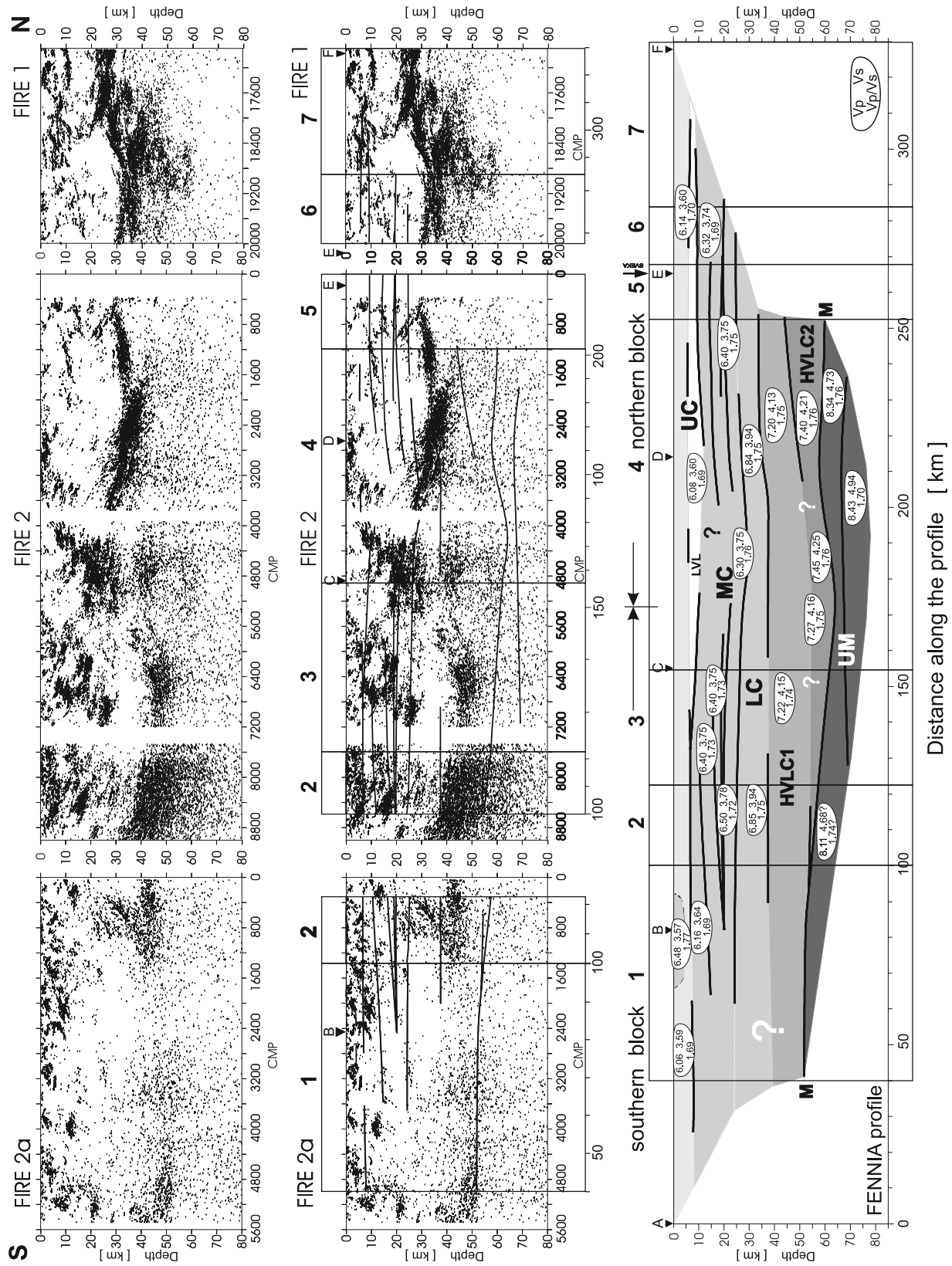


Figure 11b. Same as Figure 11a except for the FENNIA profile.

(Figure 2). The middle crust (MC) is homogeneous down to the depth of 40 km and has low P wave velocities and V_p/V_s ratio. The boundary between the MC and the lower crust (LC) is sharp, and the LC is composed of only one layer.

[56] The central block is spatially coincident with the Central Finland Granitoid complex (CFGC). It is limited on the NW by the Savo belt and on the SW by Tampere Schist belt. Granitoids of the CFGC are spread in the uppermost layer of the crust down to the depth of 5–10 km, but they are underlain by metamorphic rocks of amphibolite facies, as can be concluded from the values of V_p and V_p/V_s ratio. The most spectacular feature of this block is the thickened lowermost crust composed of three layers denoted as LC, HVLC1 and HVLC2 in Figure 5.

[57] The transitional block corresponds to transitional zone between Archaean and Proterozoic domains and is limited on the southwest by the Savo belt. In this block we can distinguish four layers. The values of V_p , V_s , and V_p/V_s in the upper crust (UC) are generally higher than the correspondent values for Archaean granitoids and gneisses exposed at the surface [Kern *et al.*, 1993], indicating that felsic Achaean rocks are present only as a thin surface layer (0–5 km). Multiple reflectors in the UC cannot be correlated to any surface structures; they are rather explained by a generally high reflectivity of the upper crust, which is also seen in the section of the FIRE1 reflection profile (Figures 11a and 11b).

[58] The middle crust (MC) in this block is composed of amphibolite facies rocks (most probably, pelitic and mafic gneisses), while the lower crust (LC) is composed of mafic garnet-free granulites, as can be judged from the comparison of V_p , V_s , and V_p/V_s to the values for these rocks reported by Weiss *et al.* [1999] and by Kozlovskaya *et al.* [2004]. The HVLC3 has P wave velocity typical for mafic garnet granulites, but lower V_p/V_s ratio than the overlying layer. The depth to the Moho boundary increases from 52 km in the northeast to 60 km in the southwest.

[59] In the seismic model of the FENNIA profile (Figure 5) we can see two blocks with different distribution of V_p , V_s , and V_p/V_s . The lateral variations of these values in the crust and in the depth to the Moho boundary can be associated with two large blocks of the crust, marked as southern and northern in Figure 5.

[60] The southern block can be subdivided into three layers. The uppermost layer with low P wave velocities is 20 km thick. The high-velocity anomaly at the surface coincides with Hyvinkää gabbroic and ultramafic intrusion, as can be judged from values of V_p , V_s , and V_p/V_s . The middle crust down to a depth of 35 km has V_p , V_s , and V_p/V_s corresponding to amphibolite facies rocks [Weiss *et al.*, 1999]. In the lower crust, only one thick layer can be recognized. The V_p in this layer is similar to the value in the HVLC1 beneath the SVEKA transect. However, the value of V_s in this layer is higher, resulting in lower a V_p/V_s ratio similar to that in the HVLC3 beneath the SVEKA transect. Generally, the velocity structure in the southern block (e.g., between SP_A and SP_D) varies significantly from the velocity structure of the SW block beneath the SVEKA transect (e.g., between SP_F and SP_H). That is surprising, because these profiles cut the same geological structures exposed at the surface, namely, the Pirkkala, Häme, and Uusimaa belts (Figure 2). These belts extend

in an east–west direction, and, according to Lahtinen *et al.* [2005], belong to the Southern Svecofennian terrain of the Fennoscandian shield. However, the significant difference in crustal structure suggests that the SW block beneath the SVEKA transect and the southern block beneath the FENNIA profile belong to different crustal units and there should be an unexposed boundary separating these units and stretching in NS direction (Figure 2).

[61] The northern crustal block corresponds to the CFGC. It is marked by the Moho depression and limited from the south by the Tampere Schist Belt. The velocity distribution within the crust is in a good agreement with the velocity structure in the central block revealed by the SVEKA transect. In particular, both HVLC1 and HVLC2 can be recognized at the crossing with the SVEKA transect.

6.2. Comparison of Velocity Models to Record Sections of FIRE Profiles

[62] Figures 11a and 11b present comparison of the velocity models obtained for the SVEKA transect and the FENNIA profile to record sections for the FIRE1 and FIRE2 reflection profiles. The northeastern part of SVEKA transect was compared with the FIRE1 profile and the FENNIA profile was compared with FIRE2a and FIRE2 profiles. Upper diagrams present original line drawing reflection sections and the bottom ones present the main features of the velocity models along profiles obtained by 2-D forward modeling. In the middle diagrams both sections are overlain. However, the reflection and refraction profiles do not coincide exactly (Figure 2). In some places they are separated by more than 50 km (see Figures 1 and 2). Thus we decided to divide the velocity models into fragments (7 for the SVEKA and 5 for the FENNIA), defined by projections of the extreme points of FIRE 1 and FIRE 2 lines to the profiles. These selected fragments were extended or compressed, respectively, and overlain on the FIRE sections (middle diagrams).

[63] For the SVEKA transect reflectivity in the upper and middle crust is generally strong, and its pattern is very complicated. Although direct comparison to the velocity sections is not possible, there is some agreement between the top of the HVLC1, HVLC3 and the high reflectivity in the lower crust in the central part of the FIRE1 profile. The Moho boundary (dashed line, Figure 11a) is seen as a boundary between reflective crust and transparent mantle beneath the central block and is coincident with the Moho obtained from ray tracing of P and S waves. However, the Moho is not clear beneath the contact with the transitional block. The differences in the Moho depths are no more than 5 km in places where profiles are up to 50 km away from each other. The second explanation of the difference may be too high velocity used to process the reflection data. The reflectors below the Moho boundary are not seen in the reflection data.

[64] For the FENNIA profile reflectivity in the upper and middle crust is generally strong, and shows a very complicated structure. It cannot be compared directly to the velocity sections due to differences in position of the lines. However, the southern block appears as generally transparent in the section of the FIRE2a profile as well as in the FENNIA profile. There is some agreement between the top of the HVLC3 and the high reflectivity in the lower crust in the central part of the FIRE2 profile. A strong reflector in

the HVLC beneath the SP_D is not seen in the reflection section. The Moho boundary cannot be identified in the FIRE2 profile, except for its southern end.

6.3. Different Types of the High-Velocity Lower Crust

[65] As can be seen, the HVLC was detected in both profiles considered in our study. In previous studies, this layer was considered to be homogeneous, resulting from magmatic underplating and intraplating [Korja *et al.*, 1993; Korsman *et al.*, 1999]. However, our results demonstrate that this layer is laterally and vertically inhomogeneous, because seismic velocities and, in particular, V_p/V_s ratio within it differ significantly. This suggests a multigenetic origin of the HVLC and the Moho boundary in the area.

[66] Recently, Eaton [2006] analyzed results of seismic reflection and refraction surveys recorded as part of the Canadian Lithoprobe program and proposed four different explanations of the origin of the continental Moho boundary. The relict Moho hypothesis proposes that the oceanic Moho is preserved during continental assembly. In this case the Moho appears as fossil subduction with continuous upper plate and lower plate reflections. The magmatic underplating hypothesis proposes formation of a new Moho by episodic emplacement of sill-like intrusive bodies. Presence of high-velocity lower crust (HVLC) with $V_p > 7.0 \text{ km s}^{-1}$ and downdip truncation of lower crustal reflections is typical for this type of Moho. The metamorphic (or metasomatic) front hypothesis proposes that the Moho is overprinted by a phase transformation to eclogite or by serpentinization of the upper mantle. In this case, the Moho topography is preserved without reciprocal surface topography, relict layering of the upper mantle is observed at wide angle and the reflection Moho is weak. The regional decollement hypothesis proposes that the Moho behaves as a structural detachment. This is indicated by dipping reflections in the upper and the middle crust that are listric into subhorizontal reflectivity in the lower crust, and by flat, strong reflection Moho. Eaton [2006] also pointed out that this is a very general classification and other explanations of origin of the continental Moho may exist.

[67] As can be concluded from Figure 11a and 11b, hypotheses of relict Moho and regional decollement can be excluded from consideration, because no corresponding reflection images are observed in the record sections of the FIRE1 and FIRE2 profiles. Therefore, here we consider two other explanations, namely, the metamorphic (or metasomatic) front and magmatic underplating hypotheses, because both of them can explain the high seismic velocities in the lower crust.

[68] The explanation of the HVLC beneath the wide-angle reflection and refraction Fennoscandian Long-Range Project (FENNOLORA) profile in Sweden by eclogite facies transition was proposed by Henkel *et al.* [1990], who interpreted the high-velocity lower crust with velocity and density of $7.1\text{--}7.5 \text{ km s}^{-1}$ and $3.15\text{--}3.27 \times 10^3 \text{ kg m}^{-3}$, respectively, as composed mainly of rocks that have reached transition to eclogite facies. Although depth to this transition depends on the rock composition, the average depth was estimated to be at about 36.5 km under present pressure and temperature conditions of the shield [Henkel *et al.*, 1990]. This value is significantly less than the depth to the Moho boundary beneath the SVEKA and FENNIA profiles,

thus the Moho boundary cannot be associated with the eclogite phase transition. However, this value agrees well with the depth to the HVLC estimated in our study. In addition, the sections of FIRE1 and FIRE2 profiles demonstrate that this depth range (e.g., 30–40 km) is also associated with a change in reflectivity pattern between the middle and lower crust, although the change is not uniform. This suggests that transition to eclogite can explain the origin of the HVLC in some places. To locate this transition, analysis of V_p/V_s ratio can be used. As it was demonstrated by Sobolev and Babeyko [1994], the rocks that underwent high-pressure metamorphic reactions generally have higher V_p and lower V_p/V_s ratio than corresponding magmatic rocks of the same chemical composition (Figure 12).

[69] As can be seen from Figure 11a, the upper boundary of the HVLC3 beneath the northeastern part of the SVEKA transect coincides with the lower limit of a highly reflective area in the upper and middle crust. Low reflectivity and low V_p/V_s ratio (1.74) of this layer suggest that this layer may represent ancient Archaean crust of intermediate composition transformed to quartz-bearing eclogite (Figure 12a). A similar explanation can be proposed for the HVLC3 beneath the FENNIA profile that has values of V_p/V_s equal to 1.74, although in this case the Late Proterozoic crust may have been transformed to eclogites.

[70] The estimated depth to the eclogite facies transition is coincident with the upper boundary of the HVLC1 also. This layer is seen as transparent beneath SP_C, while it appears as a strongly reflecting layer beneath SP_A of the SVEKA transect, indicating that it underwent nonuniform tectonic deformation. However, both the HVLC1 and HVLC2 generally have higher V_p/V_s ratios than rocks that underwent high-pressure metamorphic reactions (Figure 12a). These layers appear to be the lower part of dome-shaped structure, composed of granulite-facies (e.g., high-temperature) rocks with high seismic velocities and V_p/V_s ratios. Therefore the eclogitization (e.g., low-temperature metamorphism) cannot be accepted as an explanation for the HVLC1 and HVLC2 that appear to be composed of high-pressure, high-temperature mafic garnet granulites. O'Brien and Rötzler [2003] pointed out that the high temperatures necessary for formation of high-pressure granulites can be reached in the lower crust as a result of short-lived tectonic events that led to crustal thickening or subduction of the crust into the mantle. Such high-pressure granulites may also represent overprinted eclogites.

[71] Samples of high-pressure granulites from different locations in Finland have been studied by Hölttä and Paavola [2000] and Hölttä *et al.* [2000], who demonstrated that these rocks have different origins. The mafic xenoliths from the Lahtojoki kimberlite pipe located at the surface contact of Archaean and Proterozoic domains are mainly mafic granulites with varying garnet content; however, they contain large amounts of amphibole (up to 50%). Their composition suggests that their protholiths were crystallized from basaltic melts, suggesting magmatic underplating as the main reason for thickening of the crust. Another type of mafic garnet granulites from central Finland are mafic granulites, resulting from partial melting of amphibolites under high pressures (granulitic residues) composed primarily of garnet, pyroxenes and plagioclase. We estimated

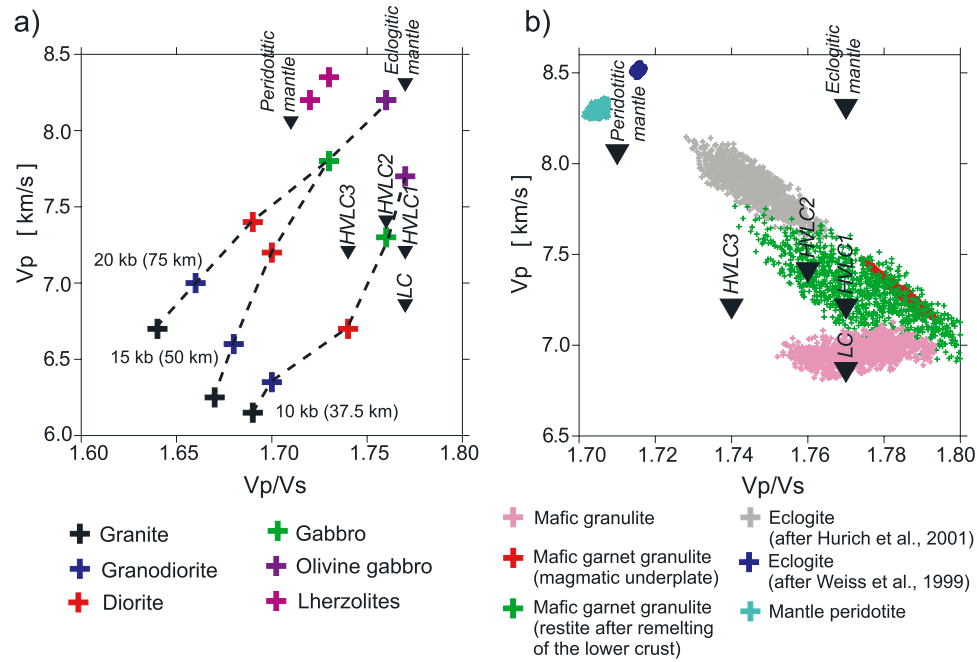


Figure 12. Comparison of V_p and V_p/V_s ratio in different layers within the lower crust and upper mantle beneath the SVEKA transect and FENNIA profile (black triangles) to the correspondent properties of the main types of lower crustal and upper mantle rocks. (a) Comparison to the V_p - V_p/V_s diagram for anhydrous magmatic rocks of various bulk chemical composition that underwent metamorphic transition from low-pressure/high-temperature assemblages to high-pressure/low-temperature assemblages [after Sobolev and Babeyko, 1994]. (b) Comparison to the V_p - V_p/V_s diagram estimated by Monte Carlo simulation from modal mineralogy of metamorphic rocks presented in Table 1.

elastic properties of these rocks from modal mineralogy using Monte Carlo simulation technique described by Kozlovskaya *et al.* [2004]. The modal mineralogy is listed in the Table 1. Results of Monte Carlo modeling are shown on V_p density and V_p/V_s density plots. As can be seen, the elastic properties of these two groups of granulites generally overlap. Both rock types have values of V_p and V_p/V_s ratio similar to the correspondent properties of the HVLC1 and HVLC2. From this it can be concluded that the nonuniform and thick HVLC beneath the CFGC was most probably formed by a multistage evolution that included both partial melting of the lower crust and magmatic underplating. This agrees with recent results of Bogdanova *et al.* [2006], who demonstrated that magmatic underplating should not be considered as a uniform explanation of the origin of the HVLC. The values of V_p , V_s , and V_p/V_s in the lower crust correspond to mafic garnet-free granulites (LC) and mafic

garnet granulites (HVLC1 and HVLC2). However, the garnet content in the HVLC2 is higher than in the HVLC1.

6.4. Different Types of Crust-Mantle Transition

[72] Analysis of V_p , V_s , and V_p/V_s ratio beneath the SVEKA transect makes it possible to analyze the lithological crust-mantle boundary that we define as transition from lower crustal lithologies (mafic granulites and eclogites) to peridotitic upper mantle. Generally, the upper mantle peridotites have lower values of V_p/V_s than corresponding values for mafic garnet granulites and eclogites (in particular, quartz-free eclogites), as the latter have higher values of isotropic P wave velocity [Sobolev and Babeyko, 1994; Kern *et al.*, 2002] (Figure 12). In addition, eclogites with high content of amphibole have higher V_p/V_s ratio compared to that of mafic garnet granulites. This is illustrated by Table 1 and Figure 12, where the result of Monte Carlo simulation of

Table 1. Modal Mineralogy for Selected Rock Types Used for Monte Carlo Simulation^a

	Qtz	Pl (An60)	Cpx	Opx	Ol	Amph	Grt	Source
Mafic granulite	0–4	40–50	4–20	8–12	0	20–30	0–9	after Kozlovskaya <i>et al.</i> [2004]
Mafic garnet granulite (restite)	0	20–40	30–50	0	0	5–15	5–15	after Hölttä and Paavola [2000]
Mafic garnet granulite (magmatic underplate)	0	12–60	6–40	1–20	0	4–70	1–30	after Hölttä and Paavola [2000] and Hölttä <i>et al.</i> [2000]
Eclogite	0	0	5–41	6–16	3–15	13–57	20–40	after Hurich <i>et al.</i> [2001]
Eclogite	0	0	40–60	6–7	3–7	1–2	40–60	after Weiss <i>et al.</i> [1999]
Mantle peridotite	0	0	5–10	10–20	65–90	0	0	after Peltonen <i>et al.</i> [2002]

^aModal mineralogy is in percent. An in Pl is shown in parentheses. Mineral abbreviations are after Kretz [1983].

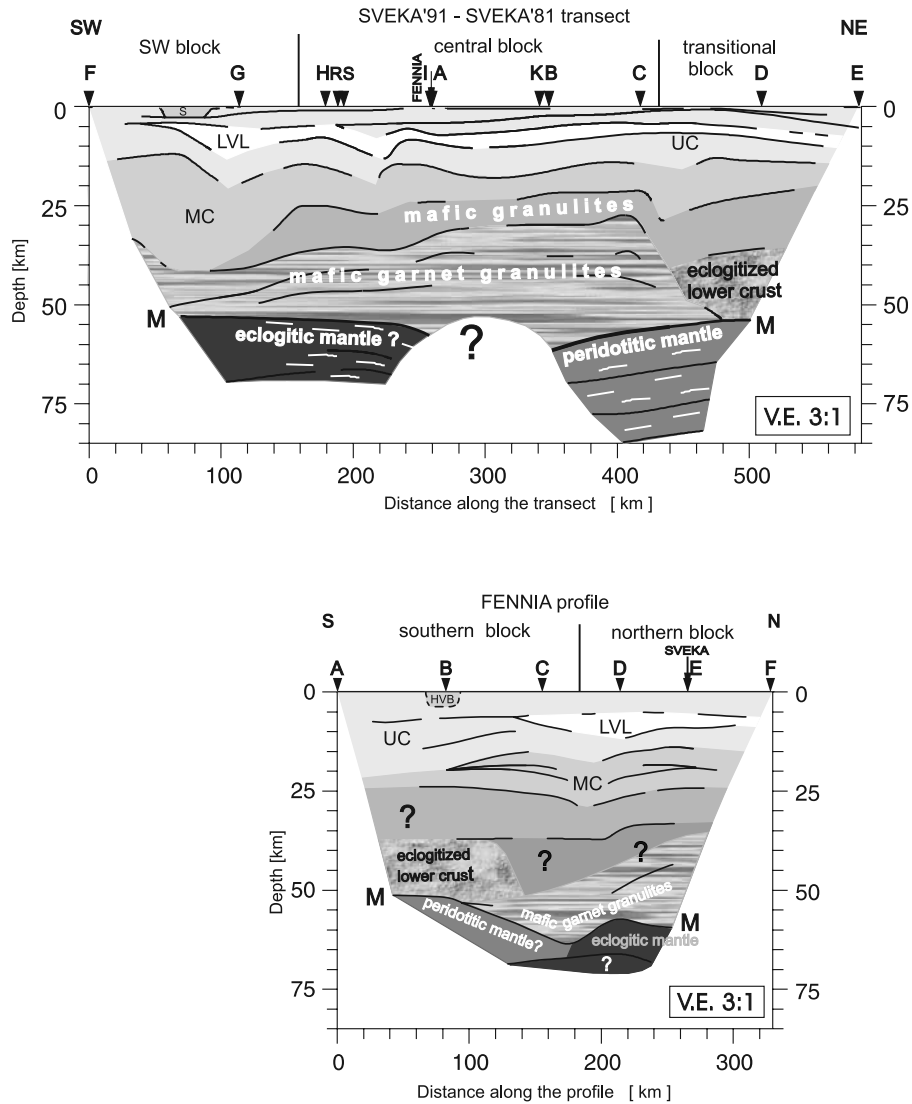


Figure 13. Simplified sketch of the crustal structure along the SVEKA transect and FENNIA profile developed by forward ray tracing. Thick lines are velocity discontinuities. Dotted lines show boundaries between four blocks with different structure. UC, upper crust; MC, middle crust; LVL, low-velocity layer; M, Moho boundary.

two different types of eclogites (e.g., amphibole-bearing and amphibole-free) are shown. Therefore the lithological crust-mantle boundary can be defined by the drop of V_p/V_s ratio from 1.76–1.78 (mafic garnet granulites and quartz-free eclogites) to 1.71–1.73 (mantle peridotites). As can be seen from Figure 13, two types of crust-mantle transition beneath the transect can be distinguished. The Moho boundary beneath the transitional block corresponds to the lithological contact of mafic garnet granulites (HVLC3) with the upper mantle peridotites, having V_p , V_s , and V_p/V_s equal to 8.05 km s⁻¹, 4.71 km s⁻¹, and 1.71, respectively. On the contrary, the Moho boundary beneath the central and SW blocks corresponds to the contact of mafic garnet granulites (HVLC2) with eclogites, having V_p , V_s , and V_p/V_s equal to 8.33 km s⁻¹, 4.72 km s⁻¹, and 1.77, respectively. These values are typical for amphibole-bearing eclogites (Table 1 and Figure 12) and they also are confirmed by the FENNIA profile (Figure 5). Therefore no drop of V_p/V_s ratio is

observed at the Moho boundary, suggesting that the lithological crust-mantle boundary is deeper. It may correspond to the reflector in the upper mantle at the depth of 60–70 km, as can be judged from the low V_p/V_s ratio beneath this reflector (about 1.70), although this value is not well constrained by our data (Figure 5).

7. Conclusions

[73] The main conclusions from our study are as follows:

[74] 1. New combined seismic models of the SVEKA transect and FENNIA profile demonstrates pronounced lateral variations of values of V_p , V_s , and V_p/V_s ratio in the crust and upper mantle and of the depth to the Moho boundary that can be associated with large blocks of the crust. These blocks correlate well with major geological units in the crust except of the SW block beneath the SVEKA transect and the southern block beneath the FENNIA profile.

[75] 2. A LVL was detected at depths of 5–12 km in central and northern parts of FENNIA profile and along all SVEKA transect, with exception for most of the southwestern part. The origin of the LVL at this depth is not completely clear and requires further studies.

[76] 3. Significant differences in the crustal structure of the SW block beneath the SVEKA transect and the southern block beneath the FENNIA profile suggests that these blocks belong to different crustal units and that there should be an unexposed N-S trending boundary separating these units.

[77] 4. The Moho boundary is overlain by a high-velocity lower crust (HVLC) with $V_p > 7.15 \text{ km s}^{-1}$ and $V_s > 4.0 \text{ km s}^{-1}$. The thickness of the HVLC and the value of V_p/V_s ratio in this layer varies from 10 to 30 km and from 1.74 to 1.78, respectively.

[78] 5. There is a general agreement between SVEKA and FENNIA velocity models and FIRE1 and FIRE2 reflection sections, regarding the shape and thickness of the lower crust and the Moho depth.

[79] 6. Beneath the CFGC, the thickness of the HVLC is extremely high, and two layers can be distinguished, HVLC1 and HVLC2. The P and S wave velocities in the lower HVLC2 ($V_p > 7.3 \text{ km s}^{-1}$, $V_s > 4.15 \text{ km s}^{-1}$) are higher than in the upper HVLC1 ($V_p > 7.15 \text{ km s}^{-1}$, $V_s > 4.0 \text{ km s}^{-1}$). The values of V_p/V_s are similar (1.76–1.77) within the HVLC1 and the HVLC2. This indicates that both HVLC sublayers are probably composed of mafic garnet granulites.

[80] 7. The HVLC beneath the Archaean-Proterozoic transition and beneath the southern part of the FENNIA profile is composed of one layer (HVLC3) with a V_p/V_s ratio of 1.74. This suggests that in both cases the HVLC3 is composed of rocks that have reached the transition to eclogite facies.

[81] 8. The deepest Moho (>60 km) was found beneath the central part of the CFGC. The thick crust and nonuniform and thick HVLC beneath the CFGC was most probably formed by a multistage evolution that included both partial melting of the lower crust and magmatic underplating.

[82] 9. The sub-Moho velocities of $V_p > 8.3 \text{ km s}^{-1}$, $V_s = 4.72\text{--}4.73 \text{ km s}^{-1}$ and V_p/V_s ratio of 1.76–1.77 were detected below southwestern part of the SVEKA transect and the northern part of the FENNIA profile. The values of $V_p = 8.1 \text{ km s}^{-1}$ and $V_p = 8.05 \text{ km s}^{-1}$ were detected below the southern part of the FENNIA profile and the northeastern part of the SVEKA transect, respectively. The values of V_s and V_p/V_s were not well constrained there. The differences in sub-Moho velocities indicate different composition of the upper mantle below profiles.

[83] 10. Three main types of the Moho boundary were distinguished: The first type corresponds to eclogitized lower crust (HVLC3) that overlies peridotitic upper mantle. In this case the Moho boundary coincides with the lithological crust-mantle boundary. The second type corresponds to HVLC2 composed of mafic garnet granulites overlying peridotitic upper mantle, for which the seismic Moho and the petrophysical crust-mantle boundary coincide as well. The third type corresponds to the contact of mafic garnet granulites (HVLC2) and eclogites. In this case the Moho and the petrophysical crust-mantle boundary do not coincide and the latter is located deeper than the Moho boundary.

[84] 11. Upper mantle reflectors were detected at the depth of 60–80 km. They could represent relicts of ancient subduction and collision processes, as has been proposed by Balling [2000] and Yliniemi et al. [2004].

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