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Seismic Investigations of the Subcrustal Lithosphere Beneath Fennoscandia

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Abstract. A detailed *P*-velocity-depth model SCA1 of the structure of the subcrustal lithosphere beneath Scandinavia was derived using data from an approximately 600-km-long seismic profile running across the northern part of Scandinavia, the BLUE ROAD traverse. After examining the effect of a laterally varying Moho on later arrivals it was possible to reconstruct a *P*-velocity model containing five layers *P*1 to *P*5 of high velocity in the depth range from 50 to 100 km which appear to be embedded in material of normal to low seismic *P*-velocity. The amplitude ratios of these arrivals were compared with those from synthetic seismograms derived from the model.

These results were then used for a partial reinterpretation of Russian nuclear explosion observations recorded at NORSAR. The King and Calcagnile (1976) model KCA of monotonously increasing velocities in the subcrustal lithosphere produces significant discrepancies between observed and theoretical amplitudes which were found to be resolved by the insertion of the BLUE ROAD model. In addition, evidence for a further low velocity zone between 170.0 and 190.0 km depth was found.

Taking all findings into account the subcrustal lithosphere and upper mantle beneath Fennoscandia seem to consist of alternating layers of low and high seismic velocoty, their lateral extensions possibly not exceeding 100–160 km.

Key words: Lithosphere-Asthenosphere transition beneath Fennoscandia – Deep seismic sounding – Synthetic seismograms – Amplitude ratios.

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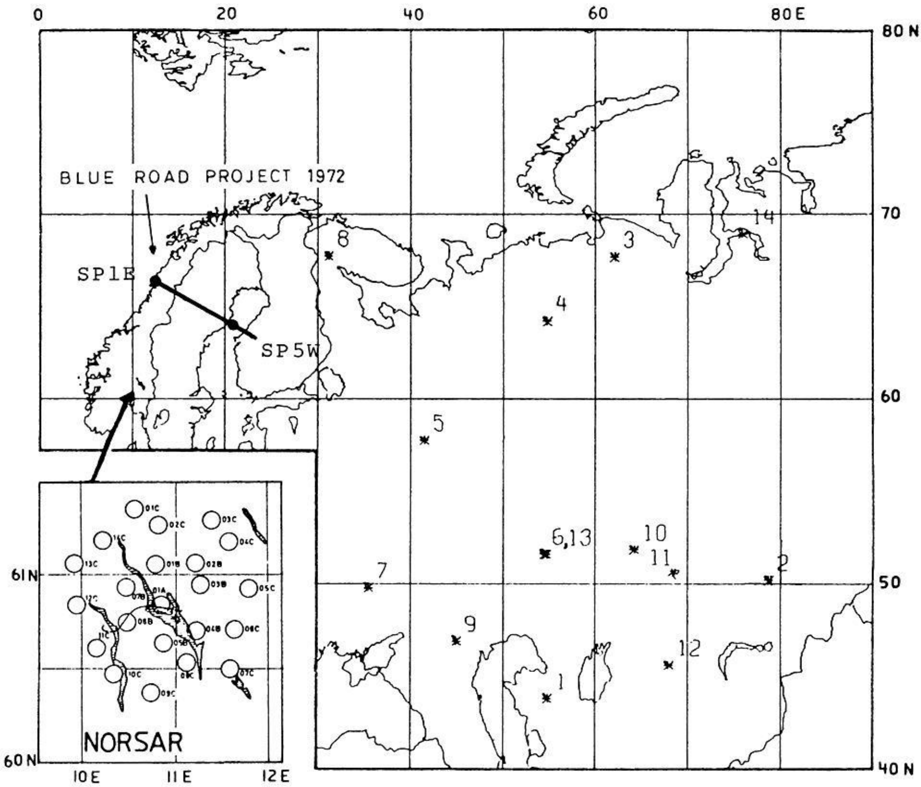


Fig. 1. Locations of seismic events in the USSR recorded at NORSAR (after King and Calcagnile 1976). *Top left*: BLUE ROAD traverse. *Inset*: NORSAR array location and subarray configuration

Introduction

Existing models of the postglacial uplift of the Baltic Shield incorporate an elastic layer superimposed over either a viscous half space (Cathles 1975) or a thick viscous asthenosphere (Lliboutry 1971; Bott 1971). In order to obtain a more thorough understanding of the dynamics responsible for the uplift a more accurate description of the structure and physical properties of the lithosphere and asthenosphere is required.

The seismic record sections used in the first part of this study were from the 'BLUE ROAD PROJECT 1972', comprising an explosion seismic profile which runs from the Norwegian coast near Mo i Rana over the Caledonides and across the Gulf of Bothnia between Umeå and Vasa before continuing for another 100 km into Finland (Hirschleber et al. 1975; Vogel 1976) (Fig. 1).

Five shots were located so as to make possible a detailed survey of the crustal structure along the profile (Hirschleber et al. 1975). The two longest record sections, SP 1 E (600 km) and SP 5 W (420 km) allow reversed coverage of the subcrustal lithosphere to a depth of about 100 km and both contain clear first arrivals from the Moho. The data which has been filtered by a band pass of 1–20 Hz supplies ample evidence for a number of later arrivals.

The second part of this study constitutes a detailed examination of data recorded at NORSAR in the years 1970 to 1974. These recordings were provided by 14 nuclear explosions in the USSR within a distance range of $11° \lesssim \Delta \lesssim 40°$.

Taking observational gaps into account, King and Calcagnile (1976) constructed ar record section up to a distance of nearly 4,000 km. This was done by constructing profiles over the entire NORSAR area for each of the 14 events in various parts of the USSR (Fig. 1). These profiles cover a distance range between 1,250 and 4,295 km which enabled King and Calcagnile (1976) to calculate a P-velocity model to a depth of approximately 1,000 km. The resulting model, however, contains no low velocity zones. This fact is in contrast to existing models of the postglacial uplift of Fennoscandia, comprising viscous flow in the upper mantle which should indicate a region of low P-velocity.

For a more detailed description of the preparation of data used in both studies the reader is referred to Hirschleber et al. (1975) and King and Calcagnile (1976).

Interpretation of the Blue Road Data

For the interpretation the crustal structure derived by Hirschleber et al. (1975) was used. The crustal models consist of two homogeneous layers with a velocity contrast of 6.18 to 6.7 km/s at a depth of 18.6 km for SP 1 E and a velocity contrast of 6.08 to 6.5 km/s at a depth of 12.8 km for SP 5 W. Sellevoll (1973) has shown such a crustal model consisting of a granitic and a basaltic layer separated by the Conrad discontinuity to be a good first approximation for Scandinavia.

As can be seen in Figs. 2–4, first arrivals from 250 km onwards seem to undulate in a time interval of approximately 0.4 s. Using these arrivals Hirschleber et al. have applied a wave-front method (Meissner 1965) from which laterally varying Moho depths have been derived.

In order to show the effect of such an undulating Moho (P_n-velocity 8.1 km/s) on later arrivals, travel-times have been calculated using a ray tracing computer program allowing for the propagation of body waves in laterally inhomogeneous media with curved interfaces (Červený et al. 1974). During the course of this study several additions have been made to the program in order to enhance its efficiency (Cassell 1978). Not only are the undulating first arrivals reproduced but so are 'shadow zones' in the distance interval of 350–400 km (Fig. 2).

Due to the relatively constant Moho depths from a distance of 420 km onwards the undulating effect on arrivals later than $P1$ for SP 1 E is negligible, as is the effect on all later arrivals for SP 5 W. The depths of the subcrustal layers corresponding to $P1$ and $P2$ in Fig. 2 have been derived by means of a computer program developed by G. Müller which calculates travel-times for laterally homogeneous models (Figs. 3 and 4). In order to account for the non-linear first arrivals for SP 1 E and SP 5 W averaging P_n travel-time branches have been drawn through them. Although the amplitudes in the seismogram

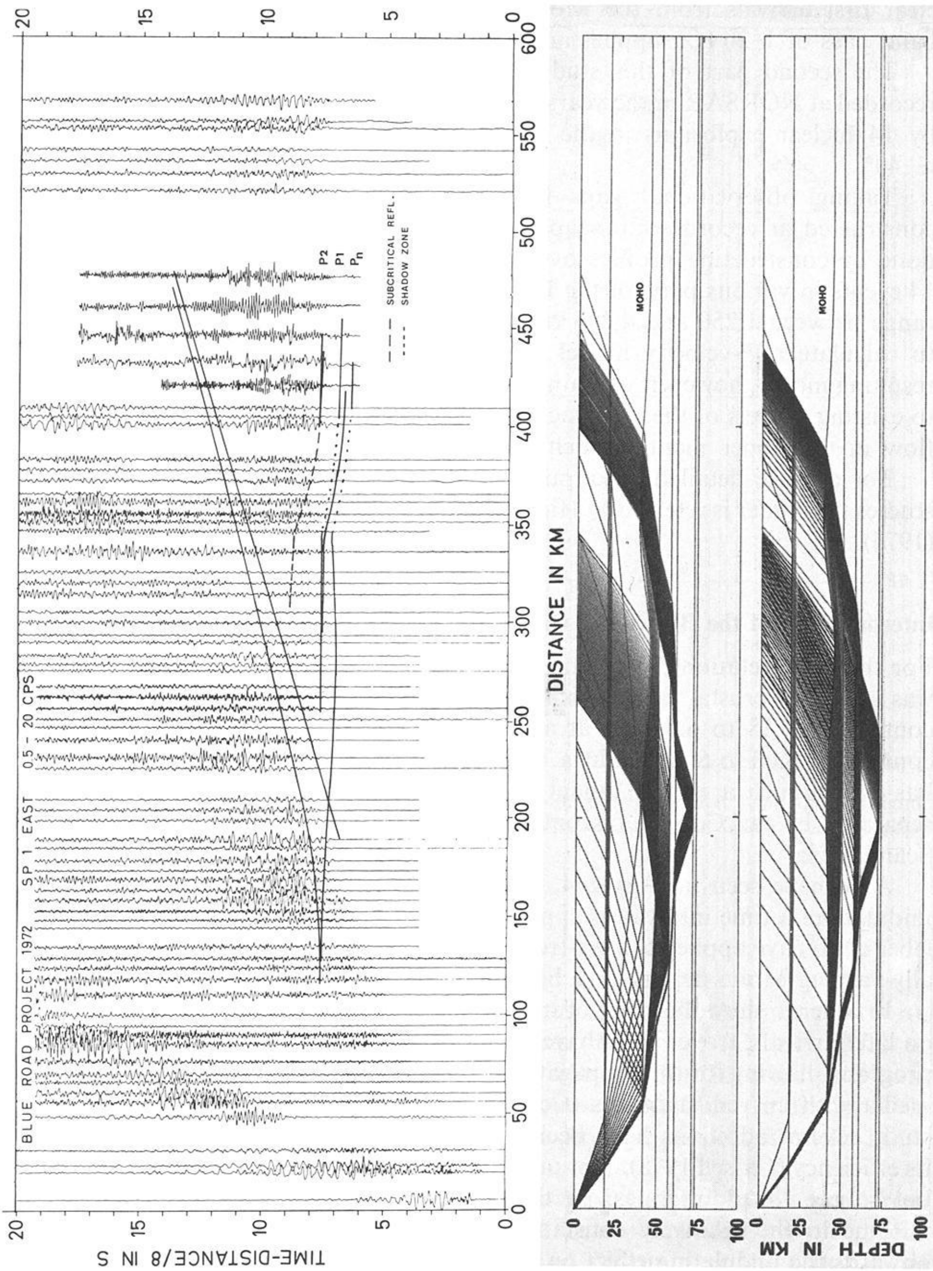


Fig. 2. Ray diagrams and travel-time curves showing the effect of an undulating Moho on later arrivals for SP 1 E. *Centre*: Reflections. *Bottom*: Refractions. The Moho relief was taken from Hirschleber et al. (1975) and the depth of the layers corresponding to *P*1 and *P*2 have been determined by calculating for a laterally homogeneous Earth. Depths are modified by allowing for a flatearth-transformation. The travel-time branches do not exceed 450 km due to the numberical method used

sections are not normalised, the relatively large-amplitude Moho reflections which should be near the inner cusp (Fig. 3) (Červený and Ravindra 1971) suggest the introduction of a 6-km-thick transition zone for the Moho at DP 1 E.

The large number of later arrivals for SP 1 E, especially in the recordings of the Gulf of Bothnia stations, suggested the correlation of a maximum of five nearly parallel travel-time branches with average velocity differences of 0.04 km/s (Fig. 3). These have been numbered $P1$ to $P5$. The possibility of these being multiple reflections has been ruled out by the unsuccessful attempt to reconstruct them by ray tracing and they could not be interpreted as such using the crustal structure as given by Hirschleber et al. (1975).

Due to the low velocity contrast between the subcrustal arrivals it was necessary to introduce low velocity zones in order to move the critical points to the small distances required. Owing to the sufficient velocity contrast between P_n and $P1$, $P1$ would be the *only* travel-time branch which could possibly be explained by a velocity increase beneath the Moho at a depth of 66.0 km without an inversion zone. The small critical distances and relatively strong amplitudes of the other arrivals make it necessary to strengthen the velocity contrasts at the interfaces by including low velocity layers ($\sim$7.7 km/s) between them. In the case of $P1$ the travel-time data could not resolve the presence of a first order discontinuity or a low velocity zone. For this study, layer $P1$ was taken to incorporate the same characteristics for use in further calculations, i.e., a thin layer embedded in low velocity material (Fig. 5). The average thicknesses of the layers is 1.8 km, in any case at least as large as a wavelength at the depths in question. The transition zones underneath the layers were introduced to reduce the effect of multiples in the computation of synthetic seismograms. For SP 5 W (Fig. 4) it was only possible to detect two later arrivals, $P1$ and $P2$ (Fig. 5). As explained earlier on for SP 1 E, arrival $P1$ for SP 5 W could also be modeled by either using a velocity increase at a depth of 66.0 km without an inversion zone or a thin layer embedded in low velocity material. In this case the latter model has again been used for reasons of consistency. Ray tracing tests (Fig. 2) show that the lateral extensions of the reflecting layers need not exceed 160 km.

The depths of layers $P1$ and $P2$ for SP 5 W coincide roughly with those of SP 1 E. In order to lessen the degree of non-uniqueness due to the inversion of the data, amplitude ratios have been calculated. The amplitude ratios of observed data and computed synthetic seismograms (Reflectivity method by Fuchs and Müller 1971) have been compared in order to verify the agreement between observation and theory (Fig. 6). For the computation of the synthetic seismograms the dominant frequency of the source signal was approximately 6.8 Hz, consisting of a roughly sinusoidal waveform with two extrema. This frequency is the average frequency of wave groups in the observational data. Since the uncertainty of the characteristics of the true source signal must also be taken into account a source signal with four extrema was tried without achieving an improvement. The observational amplitudes were determined by measuring the maximal amplitudes of wave groups directly behind arrivals. The main problem however, was the determination of critical distances since the non-normalised amplitudes made it impossible to distinguish between subcri-

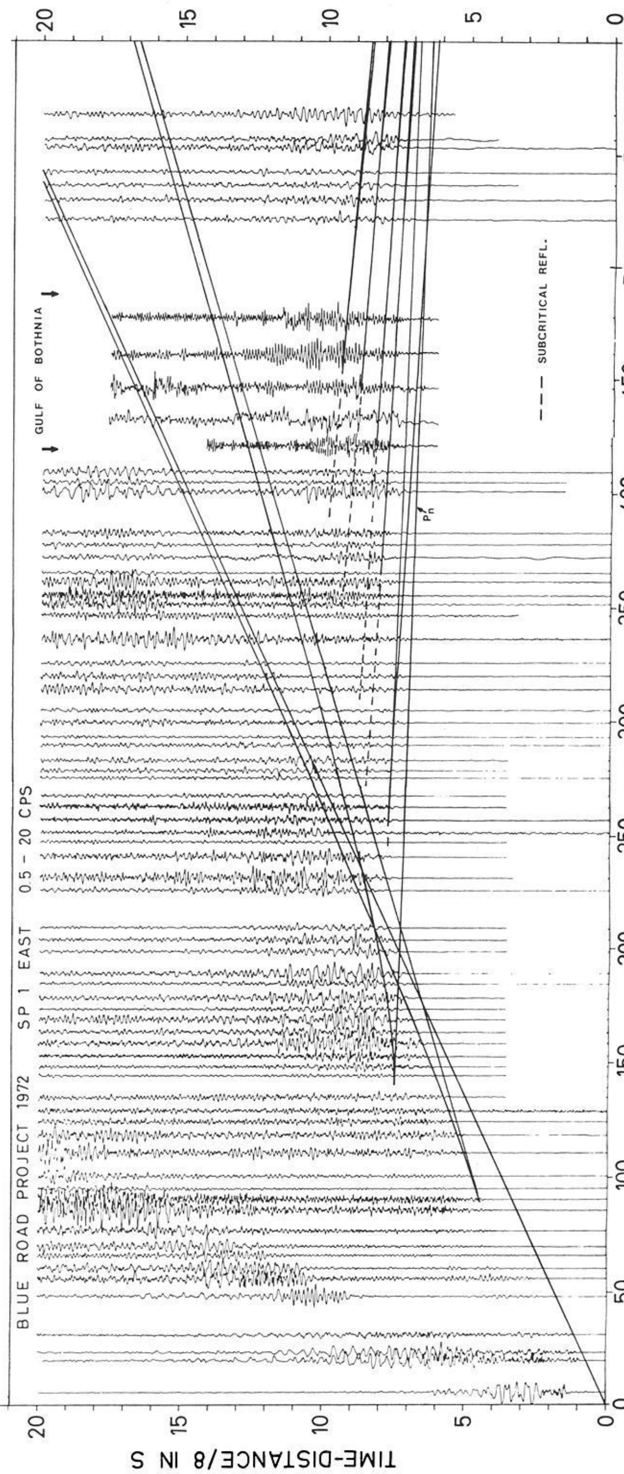


Fig. 3. Travel-time curves and record sections for SP 1 E. Arrivals later than P_n are labeled *P*1 to *P*5. Crustal arrivals calculated after a model by Hirschleber et al. 1975. Moho and upper mantle arrivals correspond to the model presented in Fig. 5. P_n branch approximates undulating first arrivals

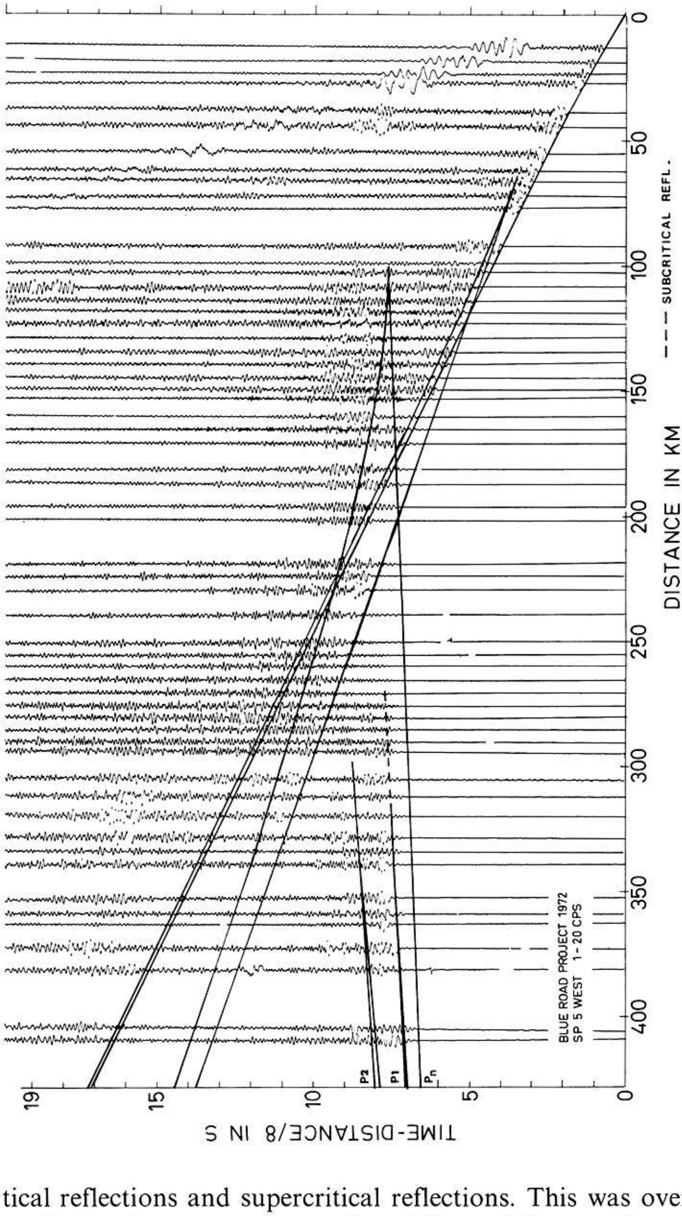


Fig. 4. Travel-time curves and record section for SP 5 W. Crustal arrivals calculated after a model by Hirschleber et al. 1975. Corresponding model presented in Fig. 5. P_n branch approximates undulating first arrivals

tical reflections and supercritical reflections. This was overcome by computing synthetic seismograms for small variations of the models until the best amplitude ratio correlations were found. To reduce the scatter of observed amplitudes, ratios of arrivals on the same seismogram were used. Figure 7 shows the final synthetic seismogram section and amplitude ratio fit for SP 5 W. Even with ratios the amount of scatter is considerable but the solid line follows the trend of the observations.

We have made no attempt to study the effect of long-wavelength heterogeneities in the Earth's crust on amplitude data. The short-wavelength scatter is certainly due to near-surface heterogeneities and may also reflect heterogeneities

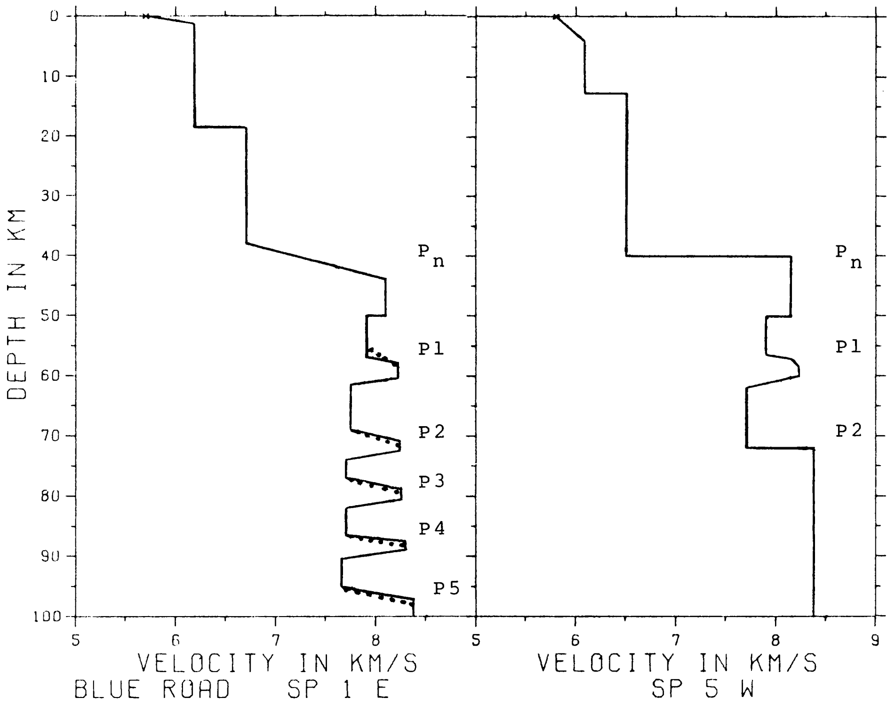


Fig. 5. Proposed velocity-depth models for SP 1 E and SP 5 W. Notations at the high velocity layers correspond to those of the travel-time branches in Figs. 3 and 4. *Dotted lines* in the left diagram specify deviations of the model which produce the dotted lines in Fig. 8. The Moho was found to consist of a 6-km-thick transition zone for SP 1 E

within the subcrustal lithosphere. Velocity-depth models are to be seen in Fig. 5. In Fig. 8 three possibilities for theoretical amplitude ratios for SP 1 E are shown. The deviations were achieved by varying the velocity gradients at the tops of the layers. The solid line in Fig. 8 is taken to be the best fit although discrepancies still remain which could not be resolved by calculating for a laterally homogeneous Earth. The corresponding model variation for the dotted line in Fig. 8 is to be seen as a dotted line in model SP 1 E in Fig. 5. The dashed line in Fig. 8 is produced by an assumed decrease of the velocity gradient on top of layer *P*1, not shown in Fig. 5. Since it was only possible to construct models by using methods allowing for laterally homogeneous structures a better fit for the amplitude ratios was not to be obtained.

At this point it should be emphasized that the detailed subcrustal structures produced by the correlations of the travel-time curves, especially by *P*3 to *P*5 for SP 1 E are in any case not the only possibilities but have been taken to be the most likely as evidence pointing to a complex subcrustal structure. Small variations of the correlation would not distort the main features of the model. A vertically homogeneous model of the upper mantle as suggested by Hirschleber et al. (1975) is not compatible with the observations. By using BLUE ROAD data, Lund (1979) has derived similarly structured *P*-velocity models

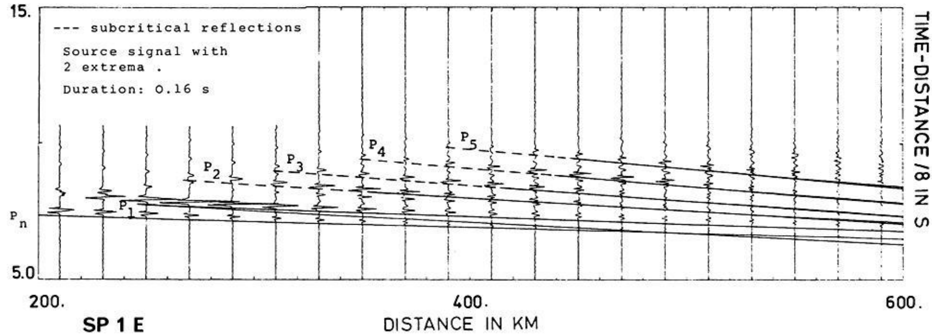
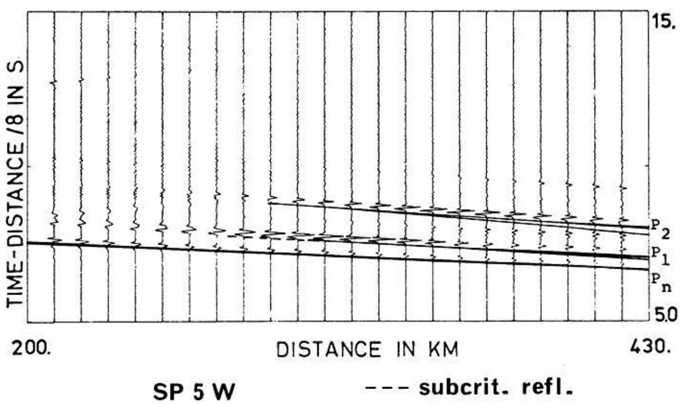


Fig. 6. Synthetic record section for SP 1 E. Travel-time braches calculated after model in Fig. 5. The dominant frequency of the source signal is approximately 6.8 Hz



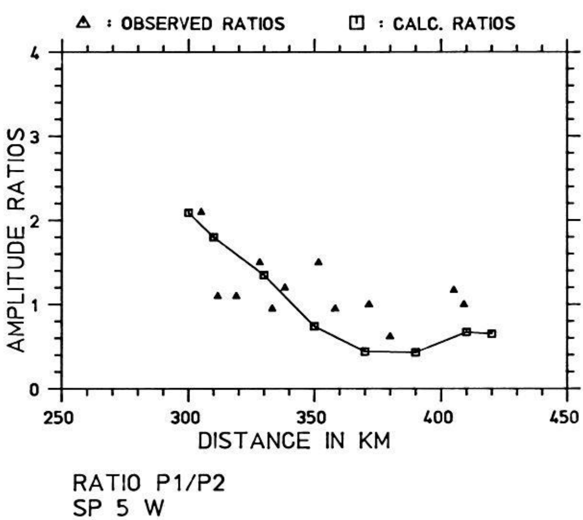


Fig. 7. *Top*: Synthetic record section for SP 5 W. Travel-time branches calculated after model in Fig. 5. The dominant frequency of the source signal is approximately 6.8 Hz and has two extrema. *Bottom*: Comparison of observed and theoretical amplitude ratios for SP 5 W

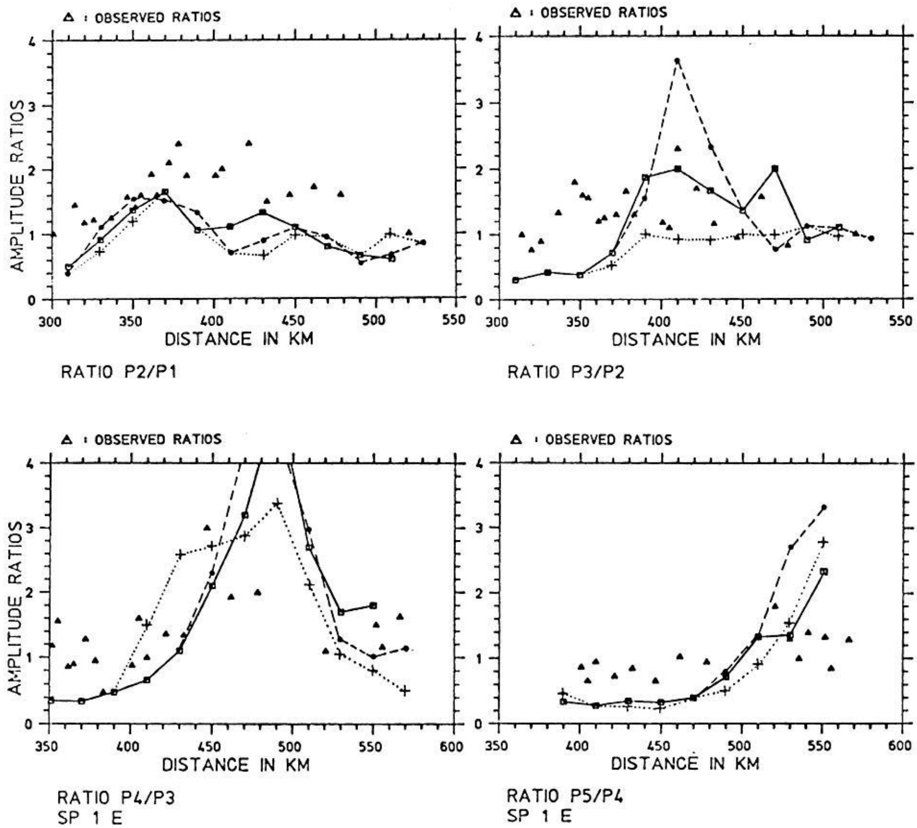


Fig. 8. Comparison of observed and theoretical amplitude ratios for SP 1 E. *Dotted lines* correspond to the *dotted lines* in Fig. 5. The *dashed curve* is produced by decreasing the velocity gradient at the top of layer P1 in Fig.5

of the subcrustal lithosphere. These were supplemented by an analysis of *S*-waves.

Reinterpretation of the NORSAR Data

King and Calcagnile (1976) constructed record sections using NORSAR data from 14 selected nuclear explosions in western Russia. The record sections with lengths of 110 km, the diameter of the array, could be put together in such a fashion as to obtain a total record section in a distance range 1,000 km $\lesssim \Delta \lesssim$ 4,500 km (Fig. 9). The amplitude variability across the array could partially be due to the effect of varying Moho depths (Berteussen 1975) and three-dimensional velocity variations of the lithosphere (Aki et al. 1977).

By subtracting approximately 4 s from the travel-times for event nos. 3 and

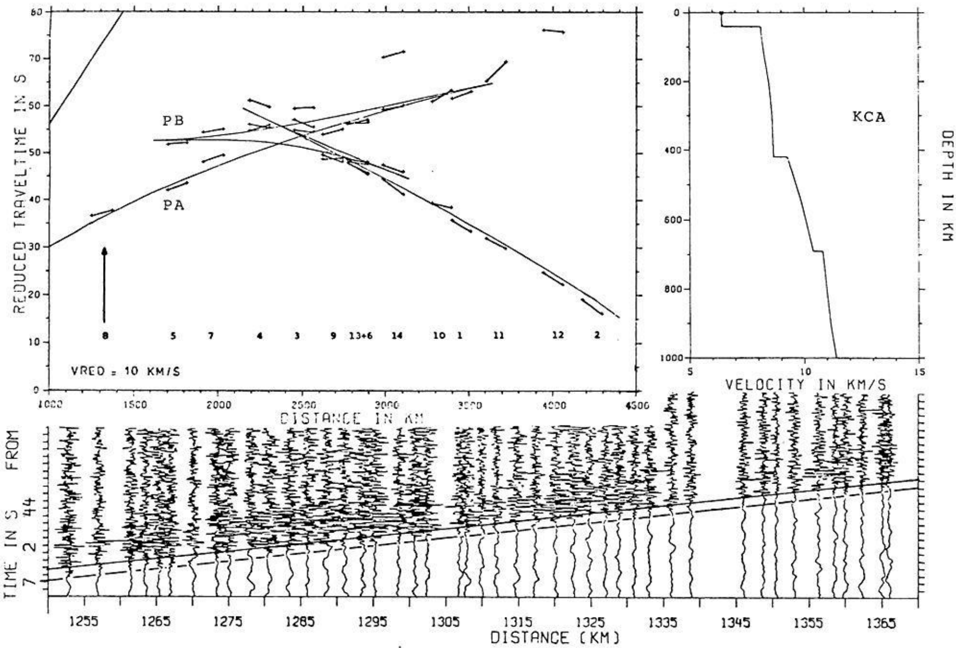


Fig. 9. *Bottom*: Record section of selected unfiltered NORSAR data for event 8. The strong secondary arrivals (*solid line*) were not considered by King and Calcagnile (1976). *Top right*: KCA model. *Top left*: Travel-time curves corresponding to the KCA model. *Short lines* represent arrivals for each numbered event. Event numbers as in Fig. 1. For a description of arrivals PA and PB see text

4 it was possible for King and Calcagnile to extrapolate continuous curves through the strongest arrivals in the data, these incorporating two (T, Δ) triplications (KCA model). The discrepancies in the correlation become evident by plotting the record sections at a reduced velocity of 10 km/s. These discrepancies are possibly due to varying precision in the determination of event locations and origin times. The azimuthal differences are probably also a contributing factor. The gaps which are incurred by the nature of the data make it difficult if not impossible to locate cusps with a reasonable degree of precision.

The KCA model consists of a relatively small positive velocity gradient below the Moho to 420 km depth followed by a first order discontinuity over a further positive velocity gradient leading to another first order discontinuity at 690 km depth.

Since there was no data available for the distance range necessary to draw conclusions on the nature of the Moho, an average depth of 40 km was chosen. Crustal effects were not considered in the calculations since their influence upon the deep structures of interest in this study is negligible. The data from event 8 (Fig. 9) suggests a strong secondary onset, arriving about 1–2 s after the first onset. This secondary arrival (solid line in Fig. 9, bottom), which has not been taken into account by King and Calcagnile gives rise to the

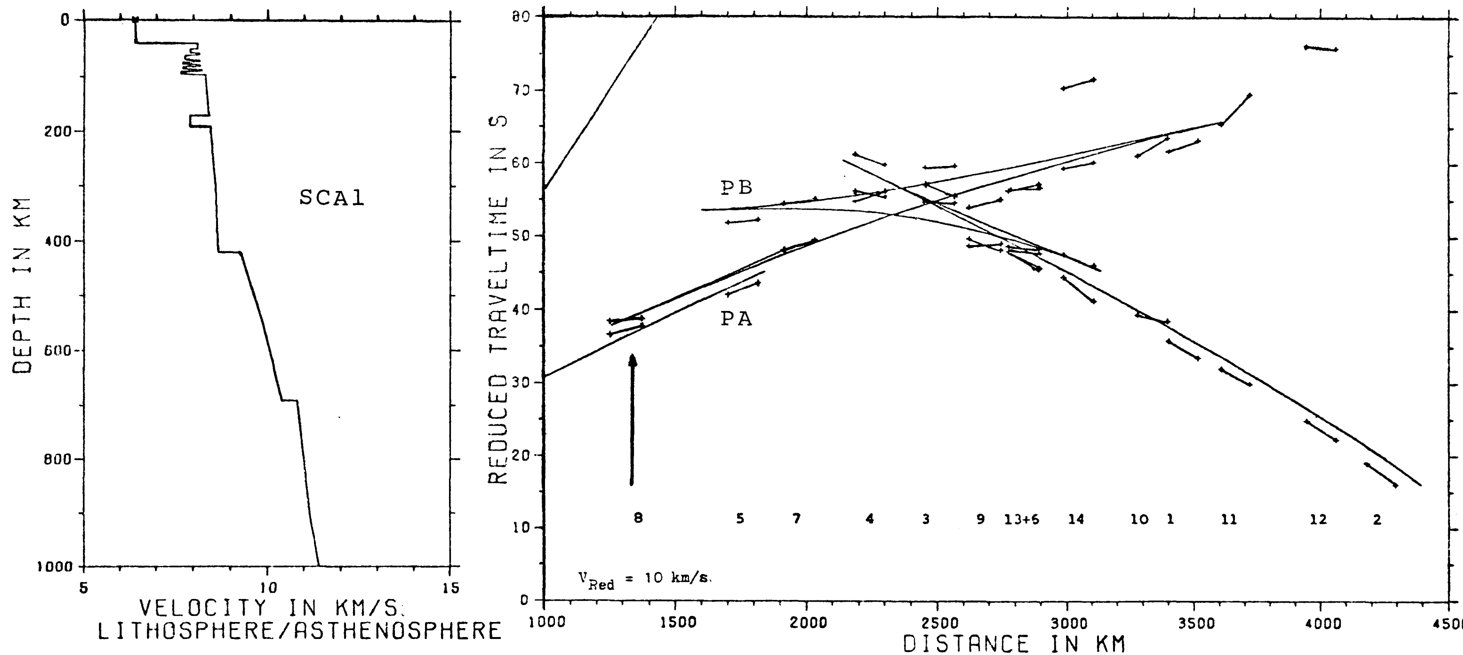


Fig. 10. *Left*: Modified version of the KCA model of the subcrustal lithosphere and upper mantle containing the BLUE ROAD results and a low velocity zone at 170 km depth. (SCA1). *Right*: Corresponding travel-time curve. *Short lines* as in Fig. 9. The secondary arrival for event 8 is to be seen in the bottom part of Fig. 9

Table 1. *P*-velocity model of the subcrustal lithosphere beneath Fennoscandia

z (km)	$v(z)$ (km/s)	z (km)	$v(z)$ (km/s)
0.0	6.40	86.5	7.70
40.0	6.40	87.5	8.30
40.0	8.10	89.0	8.30
50.0	8.10	90.5	7.65
50.0	7.90	95.0	7.65
57.0	7.90	97.0	8.32
58.0	8.22	100.0	8.32
60.5	8.22	170.0	8.40
61.5	7.75	170.0	7.90
69.0	7.75	190.0	7.90
71.0	8.24	190.0	8.45
72.5	8.24	200.0	8.45
74.0	7.70	200.0	8.59
77.0	7.70	420.0	8.66
78.8	8.26	420.0	9.27
80.5	8.26	550.0	9.88
82.0	7.70	690.0	10.38

construction of a low velocity zone between 170.0 and 190.0 km depth (Fig. 10). This is also in accordance with results published by Nolet (1977) who deduced a velocity reversal between 150 and 230 km by means of a Rayleigh wave dispersion analysis.

Although the travel-time curve produced by the relatively sharp structure of the low velocity zone satisfies the data in Fig. 9 its exact dimensions are not considered to be final.

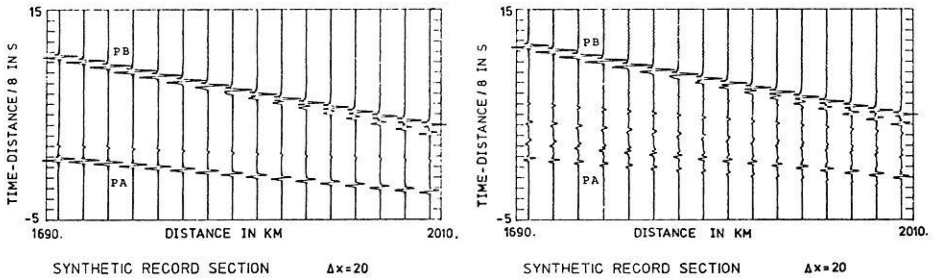
The inferred low velocity zone might also explain the discontinuation of the observed first arrival branches for events 5 (PA) and 7 in the case that the origin times of these events have been determined with sufficient precision (Fig. 10).

After having calculated synthetic seismograms for the KCA model it was found that the amplitude ratios of first (PA) and secondary (PB) arrivals were not at all in agreement with those of the observational data for event 5.

In an attempt to reduce the amplitudes of the first arrivals, the BLUE ROAD model and the low velocity layer at 170 km depth were introduced to the KCA model which resulted in our final model SCA1 and which we believe to be representative for Fennoscandia according to the presently available data (Fig. 10, Table 1).

This resulted in an essential improvement of the amplitude ratios (Fig. 11). The incurred deviations from the KCA model travel-times are small as compared to the precision in the construction of the travel-time curves themselves.

The source signal for the synthetic calculations had a dominant frequency of 2.0 Hz.



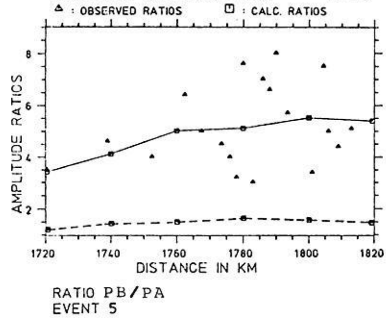


Fig. 11. *Top*: Synthetic record sections for the KCA model (*left*) and the modified model SCA1 (*right*). Dominant frequency for the source signal is approximately 2.0 Hz. *Bottom*: Amplitude ratios for both models. The *dashed line* corresponds to the KCA model while the *solid line* corresponds to the theoretical calculations for model SCA1, which produces an optimal agreement

Discussion

Other long range explosion seismic profiles have revealed the presence of relevant structural features in the lower lithosphere beneath Fennoscandia.

Massé and Alexander (1974) compared the upper mantle structure of the Canadian Shield with results of an interpretation of Russian nuclear explosion data recorded at NORSAR; in part the same data as later used by King and Calcagnile (1976). The Massé and Alexander model (MA) exhibits similar features to the KCA model, although the former includes a low velocity zone beneath the Moho.

Kulhánek and Brown (1974), using earthquake and explosion data recorded by the Uppsala array, proposed velocity depth models, including one with a low velocity zone between 50 and 150 km depth.

Sacks et al (1977), by examining SP conversions of incident teleseismic *S*-waves, found a velocity reversal at the lithosphere-asthenosphere boundary at a depth of 250 km. The data used in the present study provided no evidence to confirm their results.

The velocity-depth models shown in Fig. 5 give rise to a subcrustal lithosphere-asthenosphere consisting of alternating regions of low and high *P*-velocity.

This study did not take into account *S*-wave data and it is quite likely that the *S*-velocity-depth model would differ in detail. However, the complexities of the *P*-velocity structure require that the simple homogeneous model of a subcrustal lithosphere and asthenosphere as found by surface wave analysis must be modified for geodynamical calculations.

Using ray tracing methods (Cassell 1978) it has been shown that the minimum lateral extensions of the reflectors *P*1 and *P*2 are approximately 100–160 km which leads to the conclusion that they must not necessarily exceed that amount in order to produce the corresponding travel-time curves. A similar model of the lower lithosphere has been presented by Fuchs and Schulz (1976) who have investigated the penetration of tunnel waves through thin high-velocity layers.

Oxburgh and Parmentier (1978) propose a process allowing for the generation of continental lithosphere by diapiric accretion of low-density, depleted mantle bodies. This could lead to a plausible explanation of the alternating subcrustal structure suggested in this study.

Fuchs (1977) argues that shear stresses may create preferred orientations of olivine crystals which would cause maximum velocity orientation.

Conclusions

Beneath the Moho ($\sim$8.15 km/s) which was found to consist of a 6-km-thick transition zone under the Caledonides and a first order discontinuity to the east, the derived velocity-depth model pertaining to the BLUE ROAD data shows the subcrustal structure beneath Scandinavia to consist of a series of alternating high and low velocity layers down to a depth of 100 km.

The presence of critical reflections from the high velocity layers, which are to be observed at near distances, requires the insertion of low velocity zones of 7.9 to 7.65 km/s. Throughout the high velocity layers the *P*-wave velocity increases with depth within the range of 8.22 to 8.38 km/s.

The reinterpretation of the NORSAR long-range data infers the existence of an additional low velocity zone ($\sim$7.9 km/s) between 170.0 and 190.0 km depth.

Furthermore, the new model SCA1 of the upper mantle in Fennoscandia assures an optimal agreement between the observed and theoretical travel-times and amplitude ratios of first and secondary arrivals. Unfortunately the data were insufficient to provide information on the velocity-depth configuration in the depth range of 100 km to 170 km. This raises the question as to whether this depth range contains a continuation of the pattern of the above mentioned alternating layers.

With the presently available data it has not been possible to detect a sharp lithosphere-asthenosphere boundary. A homogeneous model does not agree with the observations. The main conclusion, however, is that the subcrustal lithosphere forms a region of clear vertical and possibly also horizontal variations characterised by high and low velocity material.

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