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# The crustal structure of Norway from inversion of teleseismic receiver functions

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## Abstract

Teleseismic body waves from seismic broadband and short period stations were used to investigate the crustal structure of Norway through inversion of the receiver functions. The Moho depths of the Baltic Shield are quite well known from previous studies including seismic experiments and spectral ratio technique. However, the results on the details of the crustal structure are inconsistent. This study provided more detailed crustal structure information at 16 locations than previously known and generally confirmed Moho depth results obtained in earlier studies. Significant differences are seen at a few sites. The Moho for the various sites was found at depths between 28 and 44 km. In summary, the crustal thickness increases from the West Coast of Norway, away from the continental margin, towards the centre of the Baltic Shield and from Southwest to the Northeast. This corresponds to the increasing age of the crust. The P velocities in the crust at most sites show a gradual increase from about 6.0 to 7.1 km/s, without clear layering.

## Introduction

The Baltic Shield, since its Archaean origin, has seen an interesting geologic history and hence has been subjected to extensive geo-scientific studies. Today's uplift of Fennoscandia presents the latest event of the geodynamic processes. Over the past four decades, a large number of scientists have been devoted to mapping the crustal structure of Fennoscandia, which has provided a great amount of valuable information and resulted in a better understanding of the tectonic evolution. The results, however, are somewhat inconsistent (Bungum et al., 1980; Kinck et al., 1993). Bungum et al. (1980) applied the spectral ratio method (Phinney, 1964) to teleseismic long-period data from Scandinavian seismic stations with the goal of finding more consistent results. The study provided Moho depths for Fennoscandia based on the same method for a large area in combination with other data, but did not allow for resolving any details of the crustal structure. The technique has evolved into the receiver function analysis (Langston, 1979; Ammon, 1991), which based on broadband and short-period data, resolves fine details of the crust. In this paper, we present the application of the receiver function method to data from seismic stations on the western Baltic Shield. The main goal is to apply this method to a large area in order to improve or confirm the knowledge on the crustal structure of mainland Norway, the islands of Bjørnøya and Spitsbergen. This study is the first country wide receiver function study for Norway. Due to the relatively large number of stations analysed in this study, we did not investigate the dependence on back azimuth for individual stations. The results have to be seen in context with existing geophysical data. A possible application of the results is to improve the location of local and regional earthquakes.

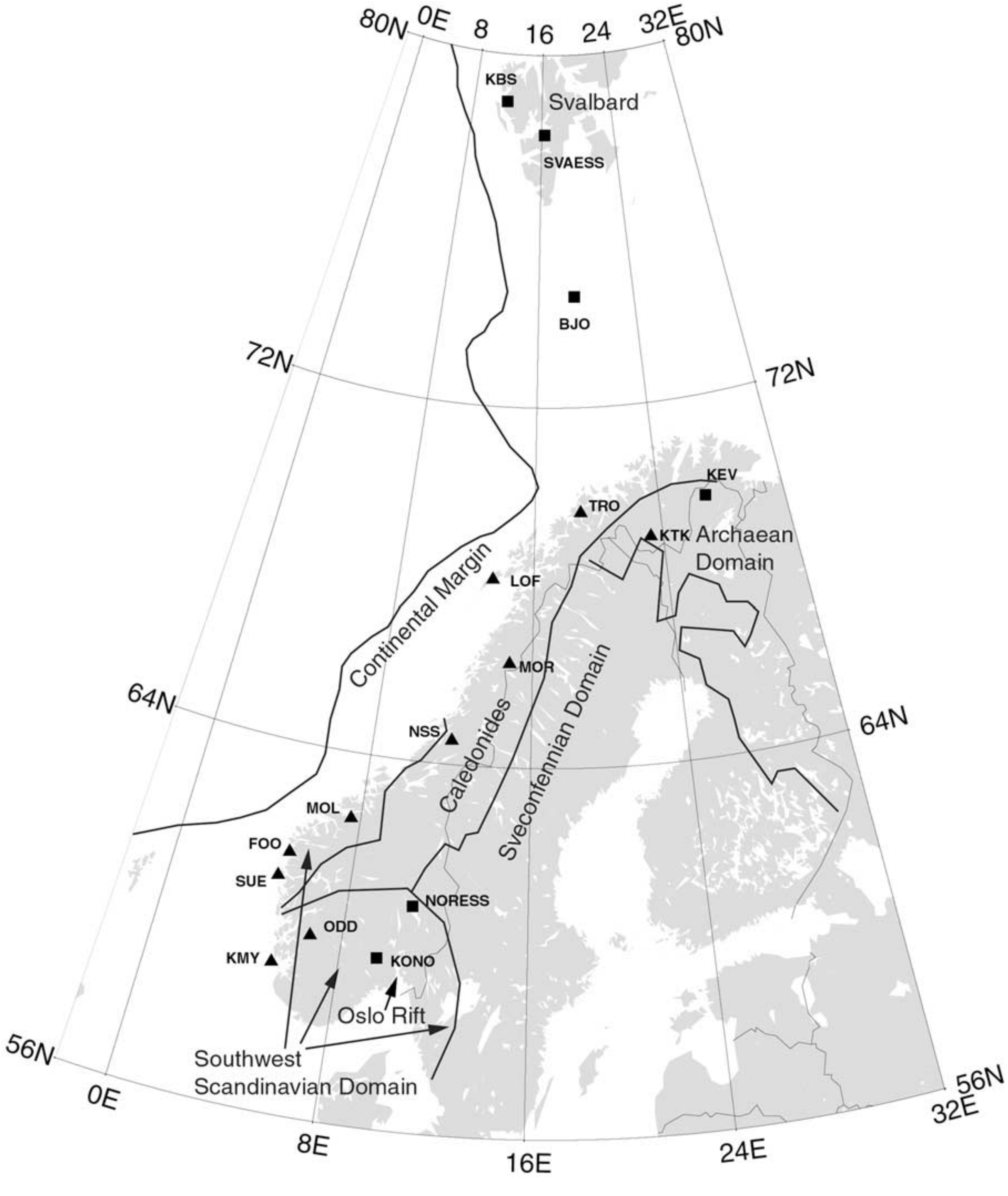

*Figure 1.* Simplified tectonic map showing the main geologic units of Norway after Gaal and Gorbatschev (1987) and Gorbatschev and Bogdanova (1993). Locations of short period and broadband seismic stations are plotted as triangles and squares respectively.

## Tectonic settings

The Norwegian mainland lies on the western edge of the Baltic Shield. Formation of continental crust of the Baltic Shield occurred in three main Precambrian orogenic events starting 3.5 Ga ago (Gaal and Gorbatschev, 1987; Gorbatschev and Bogdanova, 1993). Hence the Baltic Shield is divided into three main geochronologic zones, namely the Archaean Domain in the Northeast (3.1-2.9 Ga and 2.9-2.6 Ga), the Svecofennian Domain in the central part (2.0-1.75 Ga) and the Southwest Scandinavian Domain in the Southwest (Figure 1). The Southwest Scandinavian Domain had been affected by the Sveconorwegian-Greenvillean event (1.25-.9 Ga) and later by the Caledonian orogeny (600-400 Ma). The Caledonian orogeny has been the last major tectonic event affecting Norway and creating the mountain ranges in the western part of the country (Frisch and Loeschke, 1993).

West off the Norwegian coast, the Baltic Shield is confined by the continental margin, which has been extensively investigated (Planke et al., 1991; Faleide et al., 1991; Eldholm et al., 1994; Kodaira et al., 1995). Cutting the Southwest Scandinavian Domain, the Oslo Rift system is believed to be an extension of the Trans-European rift system (Tryti and Sellevoll, 1977; Kinck et al., 1993). North of mainland Norway, the western Barents Sea-Svalbard continental margin extends from 70 to 80° N (Faleide et al., 1991). The island of Bjørnøya is situated in the Barents Sea on the Svalbard Platform, and lies on the edge of the continental crust. Further north, the margin passes west of the Svalbard archipelago.

## Studies of Norwegian crustal structure

The continental crust of Fennoscandia as well as the continental margin offshore Norway have been extensively studied over the past four decades. A comprehensive overview of geophysical studies with respect to crustal structure carried out in Fennoscandia was given by Kinck et al. (1993). The main result of their review study was the Moho depth distribution map for Fennoscandia (Figure 8 in Kinck et al., 1993), which we use for comparison. Their map is based on seismic refraction and reflection experiments as well as long-period spectral ratio studies. Earlier maps had been presented by Sellevoll (1973) and by Luosto (1997). The Moho depths range from about 30 km along the coast, increasing towards the centre of the Baltic Shield and reaching 45 km in Northern Norway. However, due to the lack of data in certain areas, it is expected that the map will see future modifications.

*Table 1.* Overview of seismic refraction and reflection experiments in Norway

| # | Name/Location | References |
|---|---|---|
| 1 | NSDP84-3 | Fichler and Hospers, 1990 |
| 2 | NSDP84-2 | Fichler and Hospers, 1990 |
| 3 | NSDP84-4 | Fichler and Hospers, 1990 |
| 4 | NSDP84-1 | Fichler and Hospers, 1990 |
| 5 | Southern Norway, CANOBE | Cassell et al., 1983 |
| 6 | Fedje-Grimstad | Sellevoll, 1968;<br>Kanestrøm and Nedland, 1975 |
| 7 | Oslo Rift, Falkum-Orud | Tryti and Sellevoll, 1977 |
| 8 | Oslo Rift, Falkum-Semestad | Tryti and Sellevoll, 1977 |
| 9 | Flora-Åsnes | Sellevoll and Warrick, 1971;<br>Kanestrøm and Nedland, 1975 |
| 10 | Trondheim-Oslo | Kanestrøm Haugland, 1971a and 1971b |
| 11 | Årsund-Otta | Mykkeltveit, 1980 |
| 12 | Crustal transect off Norway, a | Planke et al., 1991 |
| 13 | Crustal transect off Norway, b | Planke et al., 1991 |
| 14 | Blue Norma | Avedik et al., 1984 |
| 15 | Blue Road | Hirschleber, 1975 |
| 16 | Crustal Transect across the Lofoten margin | Kodaira et al., 1995;<br>Eldholm et al., 1994 |
| 17 | Lofoten | Drivenes et al., 1984 |
| 18 | Lofoten-Vesterålen | Sellevoll, 1968;<br>Kanestrøm, 1971;<br>Sellevoll, 1983 |
| 19 | Tromsø-Muonio | Sellevoll et al., 1964 |
| 20 | FENNOLORA | Galson and Mueller, 1986;<br>Lund, 1987 |
| 21 | POLAR Profile | Luosto et al., 1989 |
| 22 | Deep seismic transects, A | Faleide et al., 1991 |
| 23 | Deep seismic transects, B | Faleide et al., 1991 |
| 24 | Deep seismic transects, C | Faleide et al., 1991 |
| 25 | Deep seismic transects, D | Faleide et al., 1991 |

Our interest is the crustal structure of mainland Norway and the islands of Bjørnøya and Spitsbergen. In these regions the coverage with seismic profiles is relatively high (Table 1 and Figure 2). However, the early studies used few, low-sensitivity instruments and simplified horizontal layer models in the analysis.

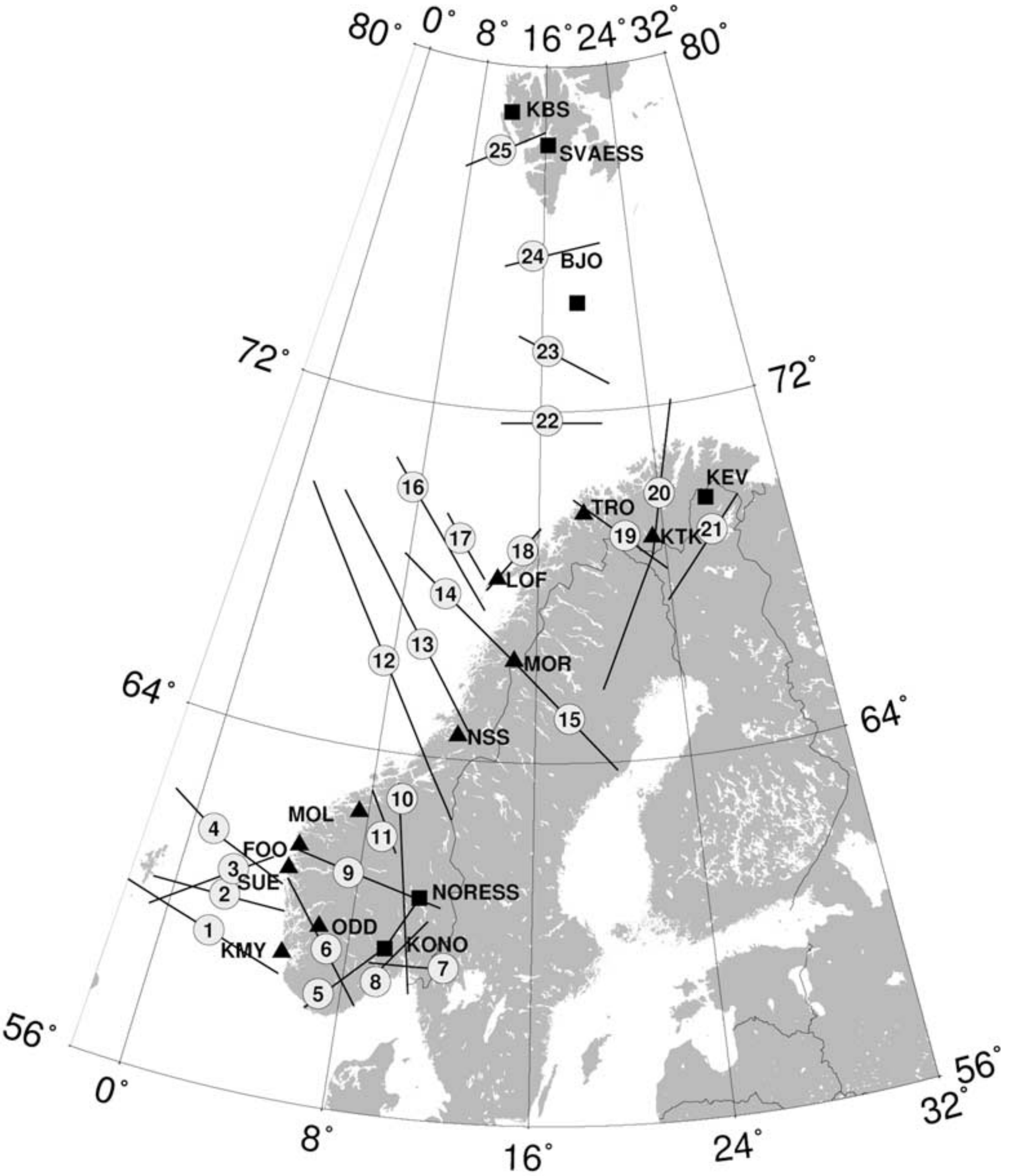

*Figure 2.* Location of cross sections obtained from seismic experiments, the numbers refer to Table 1.

Thus they were not able to resolve the fine details of the crustal structure (Kinck et al., 1993). The resolution has improved in more recent studies, however, most of the continental Norwegian crust has been studied with early methods.

The spectral ratio method was carried out for several seismic long-period stations in Fennoscandia (Table 2). The main result from these studies was the determination of Moho depths beneath the receivers, which generally agree with data obtained from seismic refraction experiments.

The results of the various studies are somewhat inconsistent, mainly with respect to the details within the continental crust. Most studies divide the crust into 2–4 layer models and distinguish between lower and upper crust, with velocities of 6.0–6.4 km/s for the

*Table 2.* Moho depths obtained through spectral ratio method for selected seismic stations

| Station | Moho depth in km | Reference |
|---|---|---|
| NORSAR array | 32–36 | Berteussen, 1977 |
| KONO | 34 | Bungum et al., 1980 |
| KEV | 46 | Bungum et al., 1980 |
| Kapp Laila (Spitsbergen) | 35 | Guterch et al., 1978 |
| Longyearbyen (Spitsbergen) | 35 | Guterch et al., 1978 |

upper part and 6.6–7.2 km/s for the lower part. The inconsistency between different studies seems to indicate that the crust is highly inhomogeneous and still not sufficiently resolved by existing studies.

## Computation and inversion of receiver functions

The method of calculating and inverting receiver functions applied in this paper follows Ammon et al. (1990) and Ammon (1991). For a detailed presentation of the theory, the reader is also referred to Langston (1979), Owens (1984) and Cassidy (1992). The receiver function method is a routine tool that has been widely used by Owens et al. (1987), Langston (1989), Ammon et al. (1993), Mangino et al. (1993), Sandvol et al. (1998), Mangino et al. (1999) and Ligorria (2000), to mention but few. Before broadband stations became available, long-period data were used, giving low resolution, while today mainly broadband or short period data (e.g. Julia et al., 1998) are used, providing higher resolution.

The radial and transverse receiver functions are time series obtained through source equalisation, which is the deconvolution of the vertical from radial and transverse components respectively. The deconvolution was done in the frequency domain using the water-level method (Clayton and Wiggins, 1976), which is given by

$$E_R(\omega) = \frac{D_R(\omega) D_Z^*(\omega)}{\Phi(\omega)} G(\omega) \tag{1}$$

where

$$\Phi(\omega) = max\left\{D_Z(\omega) D_Z^*(\omega),\ c \times max\left\{D_Z(\omega) D_Z^*(\omega)\right\}\right\} \tag{2}$$

and

$$G(\omega) = exp(-\frac{\omega^2}{4a^2}) \tag{3}$$

$E_R(\omega)$ is the radial receiver function in the frequency domain (FD), $D_Z(\omega)$ and $D_R(\omega)$ are the vertical and radial components of ground motion in the FD respectively, * indicates the complex conjugate, $c$ is the water level parameter and $a$ is the width factor for the Gaussian filter $G(\omega)$.

The receiver function is sensitive to changes in the S-velocity structure beneath the seismic station. In the linearised inversion, velocities are determined for P-waves using a constant Poisson ratio of 0.25 (Ammon, 1991), which facilitates the comparison with other studies. The structure is assumed to consist of horizontal layers of constant wave velocity. The results of the inversion are non-unique and heavily depend on the starting model (Ammon et al., 1990). The resolution of the method is expected to be 0.2–0.4 km/s for S-velocity and 2–5 km for depth (Cassidy, 1992). Due to the uncertainties, we present the Moho depth results by giving a depth range.

## Data-set and processing

Teleseismic events (Figure 3) within a distance range to the receiver of 30 to 90 degrees were selected. The data-set consisted of 182 teleseismic events recorded on 15 seismic stations in Norway and one station in Northern Finland (Table 3). Most of the stations are part of the Norwegian National Seismic Network, which is operated by the University of Bergen. Two of the stations are part of the seismic arrays operated by NORSAR. Three stations are operated by IRIS or IRIS/GEOFON. Five of the stations are broadband, while the remaining 11 stations are short period (1-second) instruments. For the stations BJO, NAO01 and SPB4 the available data-set was relatively small and it should be noted that the use of a larger data-set would provide more reliable results.

The pre-processing including data quality check, reading of P onset, tapering, rotation and cutting of data was done with the SAC software (Goldstein, 1999). The deconvolution and inversion of receiver functions was done using the programs by Ammon and others (Ammon, 1991). Receiver functions were calculated for every event-receiver pair using a water level constant of c = 0.001 and a Gaussian filter constant of a = 3.0, which corresponds approximately to a gain of 0.1 at 1 Hz. For every station, the events were grouped with respect to great circle arc distance and

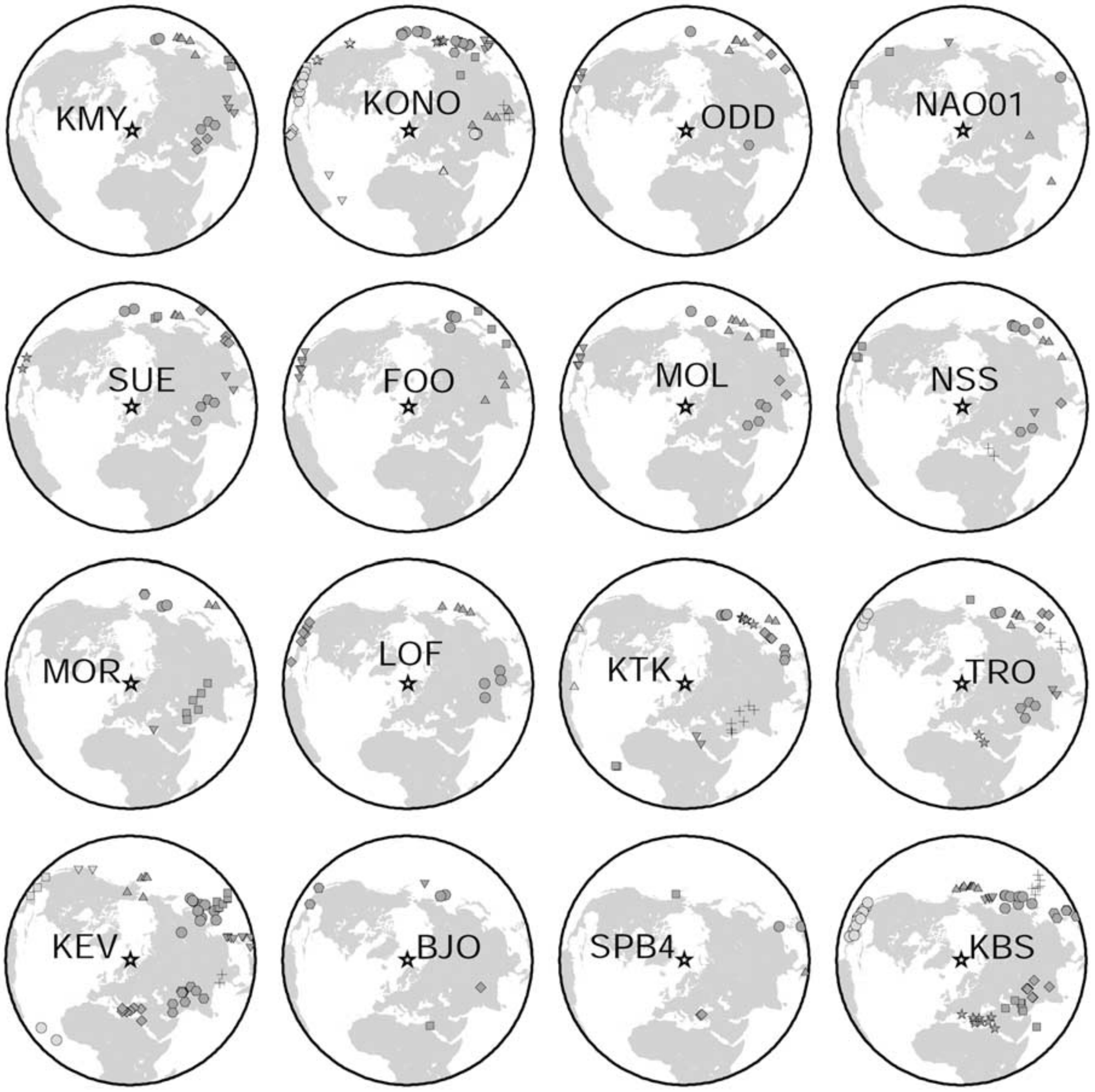

*Figure 3.* Epicentre maps of events used for computation of receiver functions from the individual stations (star). Clusters of events are indicated by different symbols and shade. The receiver functions from the event clusters were stacked before inversion for the layer velocities.

back azimuth (Figure 3). It is seen, that the number of events used per station is quite different for the various stations (Figure 3). The event clusters were stacked, if they included more than one event, to improve the signal to noise ratio. The stacked receiver functions were then inverted to obtain the plane layered crustal velocity structure.

Due to the non-uniqueness of the problem (Ammon et al., 1990) and the uncertainties in the a priori knowledge, several starting models were used for each receiver, ranging from simple three or four layer models to an approximation of a gradient model with 2 km thick layers. In general, the rather complicated receiver functions could not be modelled based on simple 3 or 4 layer models. For brevity, therefore,

*Table 3.* Overview of seismic stations used (Abbreviations are: BB = Broadband, FI = Finland, GEOFON = GeoForschungsNetz, IRIS = Incorporated Research Institutions for Seismology, Lat = Latitude, Lon = Longitude, SP = Short Period, UiB = University of Bergen)

| Sation Code | Type of sensor | Location | Operated by | Lat (° N) | Lon (° E) |
|---|---|---|---|---|---|
| BJO | SP | Bjørnøya | UiB | 74.502 | 18.999 |
| FOO | SP | Florø | UiB | 61.597 | 5.042 |
| KBS | BB | Kingsbay | IRIS/GEOFON | 78.917 | 11.917 |
| KEV | BB | Kevo (FI) | IRIS | 69.755 | 27.007 |
| KMY | SP | Karmøy | UiB | 59.202 | 5.241 |
| KONO | BB | Kongsberg | IRIS | 59.649 | 9.598 |
| KTK | SP | Kautokeino | UiB | 69.011 | 23.235 |
| LOF | SP | Lofoten | UiB | 68.132 | 13.541 |
| MOL | SP | Molde | UiB | 62.570 | 7.547 |
| MOR | SP | Mo i Rana | UiB | 66.285 | 14.735 |
| NAO01 | BB | NORESS array, Hamar | NORSAR | 60.844 | 10.886 |
| NSS | SP | Namsos | UiB | 64.531 | 11.967 |
| ODD | SP | Odda | UiB | 59.911 | 6.627 |
| SPB4 | BB | SVAESS array, Spitsbergen | NORSAR | 78.179 | 16.349 |
| SUE | SP | Sulen | UiB | 61.057 | 4.761 |
| TRO | SP | Tromsø | UiB | 69.634 | 18.908 |

we only show the results from the inversion using starting models with 2 km layer thickness. In addition synthetic forward modelling was used to refine initial models. In the inversion, a smoothness factor of $\sigma = 0.1$ was used, which gave approximately the same RMS fit between synthetic and observed receiver functions as for the pre-signal noise. Allowing for larger roughness will lead to a better fit, but at the same time it may lead to unrealistic solutions (Ammon, 1991).

## Results

The results from the inversion of the receiver functions are presented for four groups of seismic stations corresponding to the main geologic units. The resulting average P-velocity profiles determined from all stacks are shown for all individual stations in Figure 4. The plots show standard deviation to express the non-uniqueness and uncertainties involved. Table 4 summarises the depth range of the Moho in comparison to the Moho depth distribution compiled by Kinck et al. (1993). The receiver functions for selected individual stations, are shown as stacks of

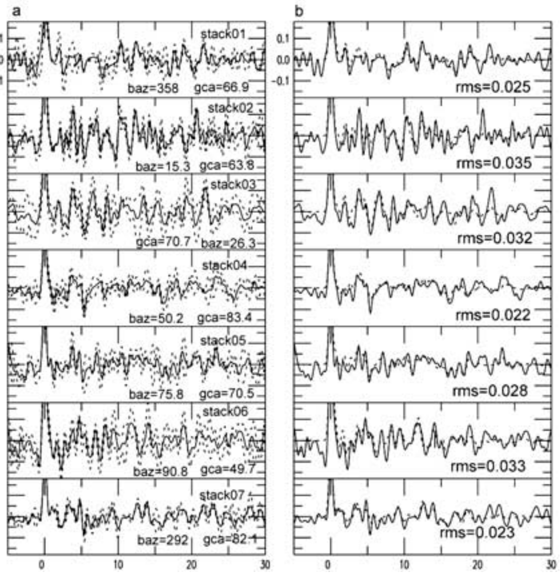

*Figure 5.* Receiver functions for station SUE, a: stacked radial components with standard deviation, and b: observed (solid line) and inverted synthetic radial components (dashed line), the abbreviations used are baz = back azimuth, gca = great circle arc distance, rms = root mean square value.

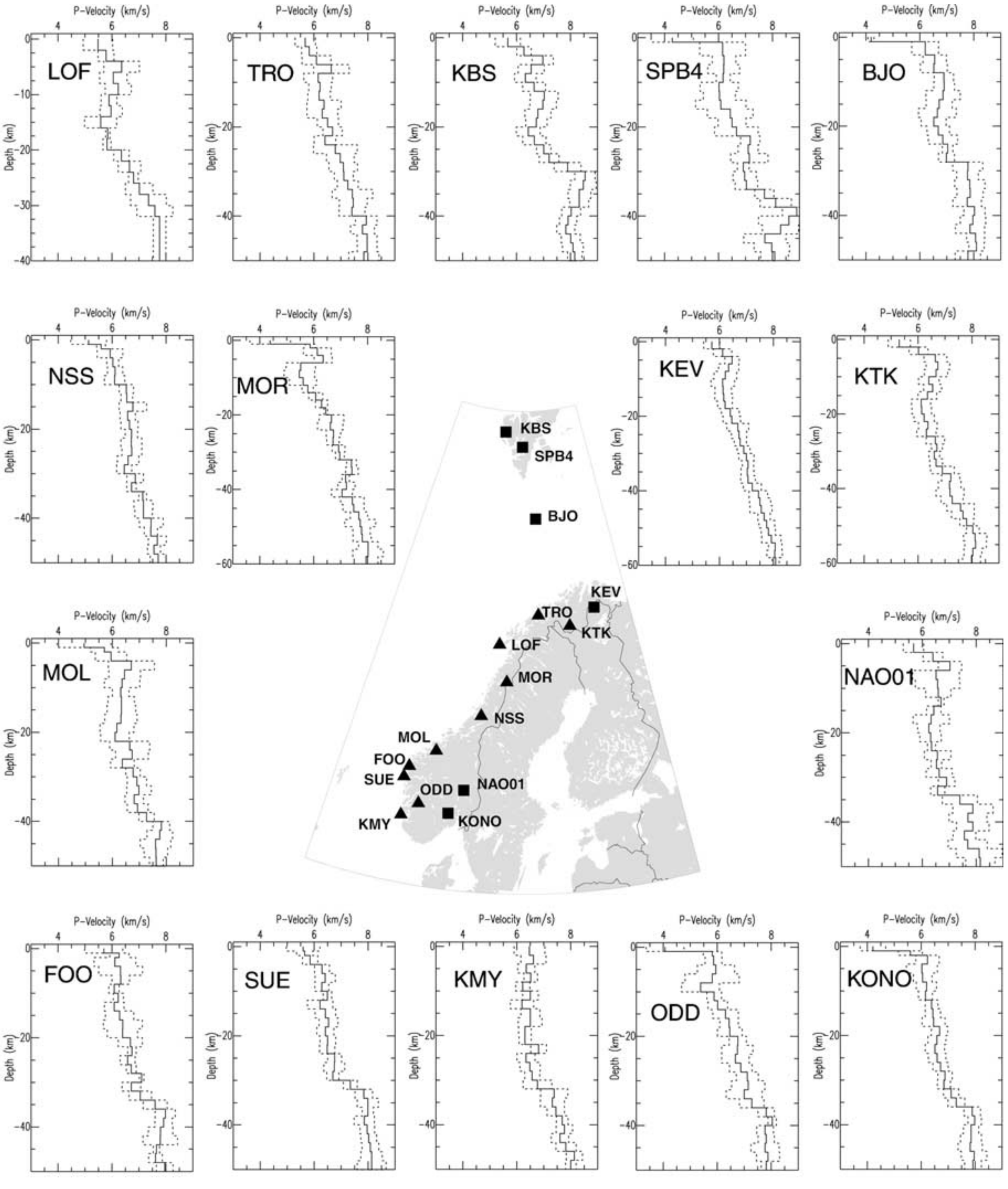

*Figure 4.* Average velocity profiles (solid lines) and standard deviation (dashed lines) obtained from the inversion of receiver function stacks for the individual sites.

*Table 4.* Moho depths for seismic stations based on receiver function analysis in comparison with Moho depth results of Kinck et al. (1993)

| Station Code | Moho depth range in km, this paper | Moho depth in km, Kinck et al., 1993 |
|---|---|---|
| BJO | 28 | – |
| FOO | 32–36 | 30 |
| KBS | 28–32 | – |
| KEV | 44–50 | 44 |
| KMY | 30–32 | 34 |
| KONO | 34–36 | 34 |
| KTK | 44–50 | 44 |
| LOF | 28–32 | 24 |
| MOL | 38–40 | 30 |
| MOR | 42–46 | 36 |
| NAO01 | 34–36 | 34 |
| NSS | 38–42 | 36 |
| ODD | 34–38 | 36 |
| SPB4 | 34–38 | – |
| SUE | 30–34 | 30 |
| TRO | 36–40 | 36 |

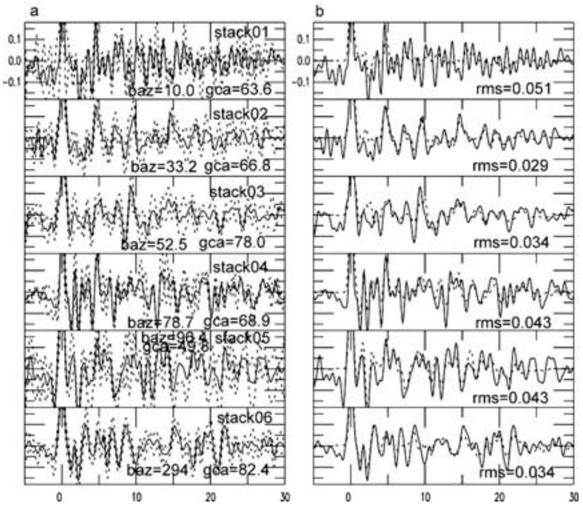

*Figure 6.* Receiver functions for station MOL (see Figure 5 for explanation)

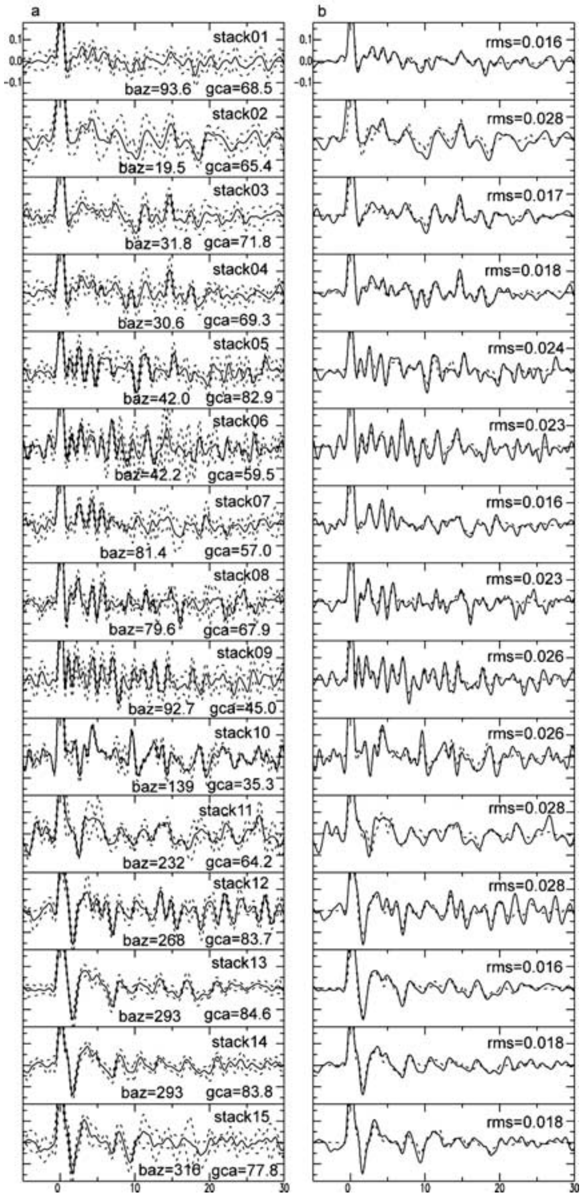

*Figure 7.* Receiver functions for station KONO (see Figure 5 for explanation)

radial component plus/minus one standard deviation and observed compared to the inverted synthetic radial component (Figures 5 to 10).

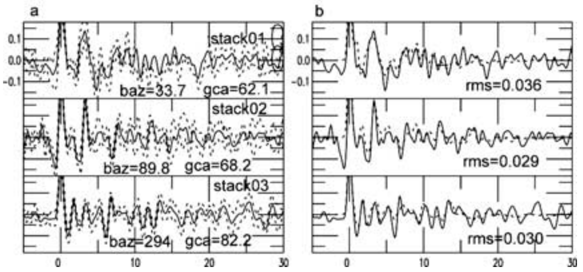

*Figure 8.* Receiver functions for station LOF (see Figure 5 for explanation)

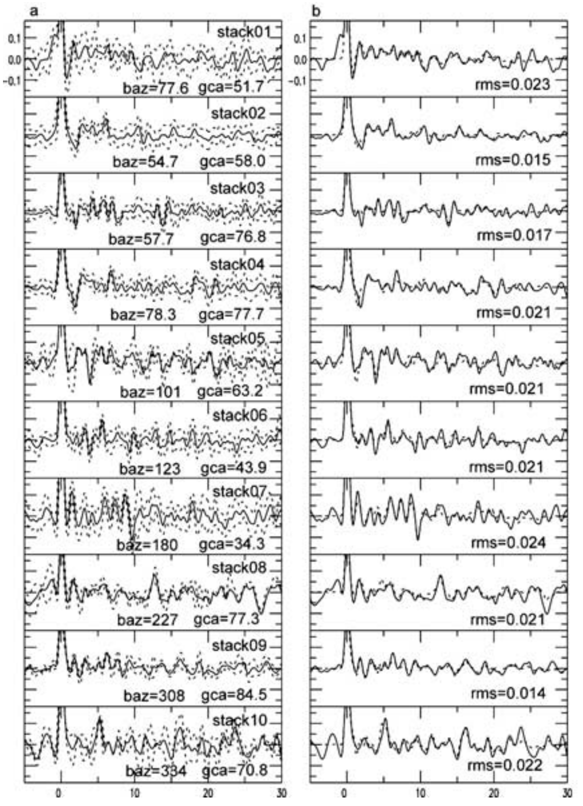

*Figure 9.* Receiver functions for station KEV (see Figure 5 for explanation)

### *Southwest Scandinavian Domain, Stations: KMY, SUE, FOO, MOL, ODD, KONO and NAO01*

The KMY (Karmøy) station presents the southern most in the line of coastal seismic stations. The Moho was found at a depth of 30–32 km. Minor changes in the velocities within the crust can be approximated by a constant velocity of 6.4 km/s. These values agree with Fichler and Hospers (1990).

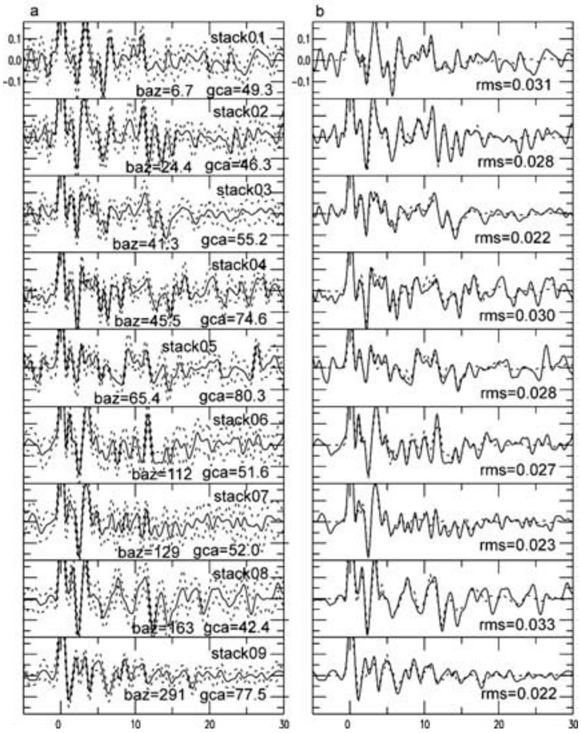

*Figure 10.* Receiver functions for station KBS (see Figure 5 for explanation)

Further north the stations FOO (Florø) and SUE (Sulen) are located within a short distance from each other. The receiver functions for SUE are plotted in Figure 5. The Moho was found at 32–36 km and at 30–34 km for FOO and SUE respectively. The crust shows a gradual velocity increase from 6.1 to 7.0 km/s and 6.3 to 6.9 km/s for FOO and SUE respectively. Earlier studies (Sellevoll and Warrick, 1971; Kanestrøm, 1971; Fichler and Hospers, 1990) reported similar results with a Moho depth of 30 km. As one would expect, the velocity models obtained for both sites are very similar.

The MOL (Molde) site presents the northern most station of the Southwest Scandinavian Domain. The Moho for this site lies at a depth of 38 to 40 km, which is slightly greater than the 35 km reported by Kanestrøm (1971). The Moho map of Kinck et al. (1993) for this site shows a depth of 30 km. Due to the large discrepancy with our results, receiver function forward modelling was done to test our results. It was confirmed that the strong Ps converted phase, which arrives about 4.5 seconds after the P phase (Figure 6) is likely to be generated by an interface at a depth of 38–40 km.

For the ODD (Odda) station, north-east of KMY, the Moho was found in the range 34–38 km. This result agrees with Sellevoll and Warrick (1971). The velocity in the crust increases from 5.8 to 7.1 km/s.

The largest data-set with 15 clusters (Figure 7) was available for the KONO (Kongsberg) station, which is located in southeastern Norway close to the western edge of the Oslo rift system (Figure 1). Due to the location on the edge of a rift system, strong differences in the derived velocity models with respect to azimuth may have been expected. This was not the case, however, the stacks 11 to 15 (back-azimuth 232-316) are significantly different from the rest, showing a large trough following the P wave. This can possibly be explained by lateral inhomogenities in the near-surface layers. The Moho is found at a depth of 34–36 km, in agreement with Tryti and Sellevoll (1977), Bungum et al. (1980) and Cassell et al. (1983). The velocity in the crust gradually increases from 6.0 to 7.1 km/s.

NAO01 is part of the NORSAR array, which is north-east of KONO. The Moho was found at 34–36 km and thus in agreement with Moho depths reported by Sellevoll and Warrick (1971), Kanestrøm (1971), Berteussen (1977) and Cassell et al. (1983). The crust shows an increase in velocity from 6.3 to 6.8 km/s without distinct velocity contrasts, which corresponds to the results of Cassell et al. (1983).

In summary, the crustal models obtained for the Southwest Scandinavian Domain are similar to earlier studies. However, a clear division into upper and lower crust or simple layer models as reported by most studies was not seen from the inversion results. Our results suggest a more inhomogeneous velocity structure, which can be approximated by a gradual increase of velocity from approximately 6.0 to 7.0 km/s. The Moho depths found for most of the sites agree with existing knowledge, except for the MOL site.

### *Caledonides, Stations: NSS, MOR, LOF and TRO*

The NSS (Namsos) seismic station is the southernmost of four sites situated on the Caledonides of Northern Norway. The thickness of the crust was found to be 38–42 km, indicating a slightly deeper Moho than shown in the Moho map compiled by Kinck et al. (1993). From the velocity profile (Figure 4) it is seen that the crust can be divided into three layers with boundaries at 10 and 30 km. The velocity in the first layer increases from near-surface low velocities to 6.1 km/s. The second layer is characterised by an almost constant velocity of 6.5 km/s and the last layer shows an increase up to 7.1 km/s. Planke et al. (1991) had presented a Moho depth of 30 km for the endpoint of profile 13 (Figure 2), which is near the coastline and from where the Moho depth seems to increase rapidly towards the NSS site. They presented a similar pattern of velocity layers.

Further north, MOR (Mo i Rana) is situated, near the border with Sweden. The Moho for this site has a depth of 42–46 km, which is slightly deeper than for NSS. Earlier studies by Hirschleber (1975) and Mykkeltveit (1980) presented the Moho at 39 and 40 km, respectively. In the upper crust, our results suggest a low velocity zone (LVZ) between 6 and 14 km. Mykkeltveit (1980) had reported a LVZ between 14 and 18 km. Since the receiver functions are only sensitive to the product of velocity and depth, we cannot conclude the exact dimensions of the LVZ, however there is strong evidence for its presence. Below 14 km, the velocity increases gradually up to 7.2 km/s at a depth of 42 km.

The LOF (Lofoten) seismic station is located on the southern end of the Lofoten islands. The receiver functions (Figure 8) clearly show a Ps converted phase at 3.3 seconds, which corresponds to a shallow Moho at 28 to 32 km (Figure 4). This result agrees with Kodaira et al. (1995), who reported a Moho depth of 28 km at the southern end of profile 16 (Figure 2). The shallow Moho depth is expected since the station is close to the continental margin. The crust can be divided into two parts, a rather constant velocity of 6.0 km/s down to 20 km and a gradual increase to 7.0 km/s from 20 to 28 km depth.

TRO (Tromsø) is situated near the coastline and is the northern most station on the Caledonides presented here. The Moho was resolved in the depth range of 36 to 40 km. This result agrees with the Moho depth of 36 km presented by Kinck et al. (1993). An estimation of 33 km Moho depth given by Sellevoll et al. (1964) seems rather shallow. The crust is characterised by a gradual velocity increase from 6.0 to 7.2 km/s.

In summary, we obtained the velocity structure for four sites on the Caledonides of Northern Norway. The dependence of the structure on the distance from the continental margin is clearly seen. The increase of depth with distance from the margin is seen in the sequence LOF, TRO, NSS and MOR.

### *Archaean Domain, Stations: KTK and KEV*

The stations KTK (Kautokeino) and KEV (Kevo) are situated in the Archaean Domain, the oldest part of the Baltic Shield (Figure 1). The receiver functions for KEV are shown in Figure 9. The velocity plots for both stations show similar characteristics. The Moho is in the depth range of 44 to 50 km, which presents the deepest Moho resolved beneath seismic stations in this study. This result agrees with earlier studies (Bungum et al., 1980; Lund, 1987; Luosto et al., 1989) where the Moho was presented at a depth of 45–46 km. The transition from crust to mantle appears to be broader than for the other sites. A LVZ is seen at both sites at depths of 10 to 22 km and 8 to 18 km for KTK and KEV, respectively. The LVZ is also seen in the results published by Luosto et al. (1989) for profile 21 (Figure 2). Beneath this zone the velocity increases gradually from 6.3 to 7.2 km/s. The change in velocity is smoother for the KEV station, possibly explained by the larger data-set available for the KEV station.

### *Svalbard and Bjørnøya, Stations: KBS, SBP4 and BJO*

The stations KBS (Kingsbay) and SPB4 (SVAESS array) are located on the arctic island of Spitsbergen. The KBS site lies closer to the margin than SPB4. The receiver functions for the KBS site are shown in Figure 10. From the inversion of the receiver functions the Moho depth was revealed at 28–32 km for KBS and at 34–38 km for SPB4. Faleide et al. (1991) showed a cross section across the margin into Svalbard with similar Moho depths. The increase of Moho depth between KBS and SPB4 shows a thickening of the crust away from the continental margin. The velocities revealed for the KBS site show a decrease with depth between 6 to 10 km and 16 to 22 km. Between 22 and 28 km the velocity increases gradually. The velocity at SPB4 is almost constant down to 16 km where it starts to increase gradually down to 34 km. For both sites the velocity values in the crust range from approximately 6.0 to 7.0 km/s.

The BJO station is located on the Bjørnøya island between Svalbard and mainland Norway. The island lies on the continental shelf close to the margin. The Moho appears at a depth of about 28 km. The crust is characterised by an almost gradual increase in velocity from 6.1 to 6.9 km/s. This result agrees with the cross sections along profiles 23 and 24 (Figure 2) presented by Faleide et al. (1991).

## Discussion

The results from this study generally agree with earlier studies. Therefore, it seems that our results can generally be trusted, including cases where the results differ from earlier studies. Due to inconsistencies between the earlier studies, for some sites both agreement and disagreement between our results and earlier studies may coexist. The benefits from this study are that the use of a single method for a large area facilitates the comparison between various sites. However, due to the non-uniqueness of the problem, a model close to the real structure can only be obtained through a combination of various geophysical techniques.

The resolution in layer thickness for the various sites on average was about 2–5 km and thus was sufficient to resolve major changes in the velocity structure. The attempt to reduce the number of layers in the resulting velocity profiles failed, because it was found that simpler models were not sufficient to explain the rather complicated receiver functions. The main feature seen at all sites is the gradual increase in velocity from about 6.0 km/s in the near-surface layers to 7.0 km/s above the Moho. The velocity profiles from some sites indicate the existence of low velocity zones. The combination of the velocity gradient with the varying Moho depth distribution presents a good approximation to the crustal structure on a regional scale and would help to improve earthquake locations.

The Moho depths revealed in this study at most sites correlate with existing knowledge and thus provide a welcome confirmation using a different method. However, we have found differences, which in most cases were small. Significantly larger Moho dephts were obtained for MOL and MOR compared to earlier studies (e.g. Kinck et al., 1993). The map of Kinck et al. (1993) was produced through contouring of Moho depths from earlier studies and suffers from the sparse spacing of profiles on mainland Norway, the lower resolution of earlier studies and the differences between the various experiments. This study provided data from additional locations as well as revised values for some sites, and thus helps to improve the general understanding of the crust in Fennoscandia.

## Conclusion

The main conclusions and results differing from previous studies are as follows:

- results confirm general knowledge of Moho depth in Fennoscandia
- a gradient model for the velocities in the crust is probably a better approximation to the true model than simple layered models
- MOL: Moho at depth of about 38–40 km
- NSS: Moho at depth of about 40 km
- MOR: Moho at 42–46 km, LVZ between 6–14 km
- KTK and KEV: LVZ in the upper crust, deepest Moho compared to the other sites in this study with 44–50 km
- BJO: Moho at 28 km, velocity gradient in the crust
- results from short period stations are reliable

After almost four decades of seismic investigation in Norway, knowledge of the continental crustal structure is still limited. This study has shown that the analysis of teleseismic body waves can provide additional valuable information on the nature of the continental crust in the Baltic Shield. Future application of the receiver function method for the remaining part of the Baltic Shield would provide additional data, which together with existing knowledge would give an improved image of the continental crust in this region. Data from temporarily deployed local networks may provide additional valuable data. The main goal of this study has been to improve the crustal velocity model beneath seismic stations in Norway. The approach was to present average velocity models for the individual sites. Smaller scale studies could examine the receiver functions as a function of back azimuth, which could reveal the dipping nature of the Moho and other near-site inhomogenities.

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## References

Ammon, C.J., Randall, G.E. and Zandt, G., 1990, On the nonuniqueness of receiver function inversions, *J. Geophys. Res.* **95**, 15303–15318.

Ammon, C.J., 1991, The isolation of receiver effects from teleseismic P wave fronts, *Bull. seism. Soc. Am.* **81**, 2504–2510.

Ammon, C.J. and Zandt, G., 1993, Receiver structure beneath the southern Mojave block, California, *Bull. seism. Soc. Am.* **83**, 737–755.

Avedik, F., Brendsen, D., Fucke, H., Goldflam, S., Hirschleber, H., Meissner, R., Sellevoll, M.A. and Weinrebe, W., 1984, Seismic investigations along the Scandinavian 'Blue Norma' profile, *Annales Geophysicae* **2**(5), 571–578.

Berteussen, K.-A., 1977, Moho depth determinations based on spectral-ratio analysis of NORSAR long-period waves, *Physics of the Earth and Planetary Interiors* **15**, 13–27.

Bungum, H., Pirhonen, S.E. and Husebye, E.S., 1980, Crustal thickness in Fennoscandia, *Geopys. J. R. astr. Soc.* **63**, 759–774.

Cassell, B.R., Mykkeltveit, S., Kanestrøm, R. and Husebye, E.S., 1983, A North Sea southern Norway seismic crustal profile, *Geohpys. J. R. astr. Soc.* **72**, 733–753.

Cassidy, J.F., 1992, Numerical experiments in broadband receiver function analysis, *Bull. seism. Soc. Am.* **82**, 1453–1474.

Clayton, R.W. and Wiggins, R.A., 1976, Source shape estimation and deconvolution of teleseismic body waves, *Geophys. J. R. astr. Soc.* **47**, 151–177.

Drivenes, G., Sellevoll, M.A., Renard, V., Avedik, F. and Pajchel, J., 1984, The continental margin/crustal structure off the Lofoten islands, Northern Norway, In: Spencer, A.M. et al. (eds), *Petroleum Geology of the North European Margin*, Norwegian Petroleum Society, Graham & Trotman, London, pp. 211–216.

Eldholm, O. and Grue, K., 1994, North Atlantic volcanic margins: dimensions and production rates, *J. Geophys. Res.* **99**, 2955–2968.

Faleide, J.I., Gudlaugsson, S.T., Eldholm, O., Myhre, A.M. and Jackson, H.R. 1991, Deep seismic transects across the western Barents Sea continental margin, *Tectonophysics* **189**, 73–90.

Frisch, W. and Loeschke, J., 1993, Plattentektonik, Wissenschaftliche Buchgesellschaft, Darmstadt.

Fichler, C. and Hospers, J., 1990, Deep crustal structure of the Northern North Sea Viking Graben: results from deep reflection seismic and gravity data, *Tectonophysics* **178**, 241–254.

Gaal, G. and Gorbatschev, R., 1987, An outline of the Precambrian evolution of the Baltic Shield, *Precambrian Res.* **35**, 15–52.

Galson, D.A. and St. Mueller, 1986, An introduction to the European Geotraverse project: first results and present plans, *Tectonophysics* **126**, 1–30.

Goldstein, P., 1999, SAC user's manual, Lawrence Livermore Laboratory, University of California.

Gorbatschev, R. and Bogdanova, S., 1993, Frontiers in the Baltic Shield, *Precambrian Res.* **64**, 3–21.

Guterch, A., Pajchel, J., Perchuc, E., Kowalski, J., Duda, S., Komber, J., Bojdys, G. and Sellevoll, M.A., 1978, *Seismic Reconnaissance Measurement on the Crustal Structure in the Spitsbergen Region*, Seismological Observatory, University of Bergen, 61 pp.

Hirschleber, H.B., Lund, C., Meissner, R., Vogel, A. and Weinrebe, W., 1975, Seismic investigations along the Blue Road traverse, *J. Geophys.* **41**, 135–148.

Julia, J., Vila, J. and Macia, R., 1998, The receiver structure beneath the Ebro basin, Iberian Peninsula, *Bull. Seism. Soc. Am.* **88**, 1538–1547.

Kanestrøm, R., 1971, *Seismic Investigations of the Crust and Upper Mantle in Norway*, Swedish Natural Science Research Council, Stockholm, 27 pp.

Kanestrøm, R. and Haugland, K., 1971a, *Crustal Structure in Southeastern Norway from Seismic Refraction Measurements*, Scientific Report No. 5, Seismological Observatory, University of Bergen, Norway, 72 pp.

Kanestrøm, R. and Haugland, K., 1971b, The Trans-Scandinavian seismic profile of 1969: profile section 3–4, In: Vogel, A. (ed.), *Deep Seismic Sounding in Northern Europe*, Swedish Natural Science research Council, Stockholm, pp. 76–91.

Kanestrøm, R. and Nedland, S., 1975, *Crustal Structure of Southern Norway, A Reinterpretation of the 1965 Seismic Experiment*, Institue Report, Seismological Observatory, University of Bergen, 47 pp.

Kinck, F.R., Husebye, E.S. and Larsson, 1993, The Moho depth distribution in Fennoscandia and the regional tectonic evolution from Archean to Permian times, *Precambrian Res.* **64**, 23–51.

Kodaira, S., Mjelde, R., Sellevoll, M.A., Hirschleber, H.B., Iwasaki, T., Kanazawa, T. and Shimamura, H., 1995, Crustal Transect across the Lofoten Volcanic Passive Continental Margin, N. Norway, Obtained by Use of Ocean Bottom Seismographs, and Implications for its Evolution, *J. Phys. Earth* **43**, 927–745.

Langston, C.A., 1979, Structure under Mount Rainier, Washington, inferred from teleseismic body waves, *J. Geophys. Res.* **84**, 4749–4762.

Langston, C.A., 1989, Scattering of teleseismic body waves under Pasadena, California, *J. Geophys. Res.* **94**, 1935–1951.

Ligorria, J.P., 2000, *The Mantle-Crust Transition beneath North America*, Ph.D. Thesis, Department of Earth and Atmospheric Sciences St. Louis University.

Luosto, U., Flueh, E.R., Lund, C.-E. and Working Group, 1989, The Crustal structure along the Polar Profile from seismic refraction investigations, *Tectonophysics* **162**, 51–85.

Luosto, U., 1997, Structure of the Earth's Crust in Fennoscandia as Revealed from Refraction and Wide-Angle Reflection Studies, Published by Finnish Geophysical Society, Helsinki, *Geophysica* **33**, 1–14.

Lund, C.-E., 1987, Crustal structure along the Northern 'FENNOLORA' profile, *Precambrian Res.* **35**, 195–206.

Mangino, S.G., Zandt, G. and Ammon, C.J., 1993, The receiver function structure beneath Mina, Nevada, *Bull. Seism. Soc. Am.* **83**, 542–560.

Mangino, S.G., Priestly, K. and Ebel, J., 1999, The receiver structure beneath the China digital seismograph network stations, *Bull. Seism. Soc. Am.* **89**(4), 1053–1076.

Mykkeltveit, S., 1980, A seismic profile in southern Norway, *Pure and Appl. Geophys.* **188**, 1310–1325.

Owens, T.J., 1984, *Determination of Crustal and Upper Mantle Structure from Analysis of Broadband Teleseismic P-Waveforms*, PhD thesis, Department of Geology and Geophysics, The University of Utah.

Owens, T.J., Taylor, S.R. and Zandt, G., 1987, Crustal structure at regional seismic test network stations determined from inversion of broadband teleseismic P waveforms, *Bull. seism. Soc. Am.* **77**, 631–662.

Phinney, R.A., 1964, Structure of the Earth's crust from spectral behaviour of long-period body waves, *J. Geophys. Res.* **69**, 2997–3107.

Planke, S., Skogseid, J. and Eldholm, O., 1991, Crustal structure off Norway, 62° to 70°, *Tectonophysics* **189**, 91–108.

Sandvol, E., Seber, D., Calvert, A. and Barazangi, M., 1998, Grid search modelling of receiver functions: Implications for crustal structure in the Middle East and North Africa, *J. Geophys. Res.* **103**, 26899–26917.

Sellevoll, M.A. and Pentilä, E., 1964, *Seismic Refraction Investigation in Northern Scandinavia,* Årbok. Mat. Naturv. Series No. **9**, University of Bergen.

Sellevoll, M.A., 1968, *Report on Crustal Studies in Norway*, Seismological Observatory, Bergen, Norway.

Sellevoll, M.A. and Warrick, R.W., 1971, A refraction study of the crustal structure in southern Norway, *Bull. Seism. Soc. Am.* **61**, 457-471.

Sellevoll, M.A., 1973, Mohorovicic discontinuity beneath Fennoscandia and adjacent parts of the Norwegian Sea and North Sea, *Tectonophysics* **20**, 359–366.

Sellevoll, M.A., 1983, A study of the Earth's crust in the island area of Lofoten-Vesterålen, northern Norway, *Nor. Geol. Unders.* **380**, 235–243.

Tryti, J. and Sellevoll, M.A., 1977, Seismic Crustal Study of the Oslo Rift, *Pure and applied geophysics* **115**(4), 1061–1085.