Crustal structure of central Norway and Sweden from integrated modelling of teleseismic receiver functions and the gravity anomaly

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SUMMARY

Receiver functions have been calculated from teleseismic events recorded by an array of seismometers deployed on an E-W transect between the coasts of central Norway and Sweden. Forward and inverse modelling and migration of the receiver functions yields models for the subsurface velocity structure along the profile which have the crust thickening from c. 32 km at the Norwegian coast to c. 43 km beneath the central Scandinavian mountain range and then remaining constant beneath Sweden. There is some evidence for a low-velocity layer in the upper 10 km of the crust beneath parts of Norway and western Sweden and good evidence for a high-velocity lower crust underlying much of Sweden which thins to the west beneath Norway. Inverting the seismic velocities to density results in a very good correspondence between calculated and observed gravity anomalies. The results of this study do not support the presence of a significant crustal root providing buoyant support for the mountain range. Low topography and thick crust beneath Sweden are maintained by the high-velocity, high-density lower crustal layer. The upper crustal low-velocity layer is consistent with models based on existing refraction profiling and known geology and physical properties of the crust. There is no direct correlation between properties of the crust and topography suggesting that recent epeirogenic uplift has not resulted from modification of the crust.

Key words: Gravity anomalies and Earth structure; Intra-plate processes; Crustal structure; Europe.

1 INTRODUCTION

The Scandinavian mountain range running the length of Norway and extending east into Sweden (Fig. 1) lies at least 400 km from the nearest plate boundary. It is principally formed from the eroded core of the Scandinavian Caledonides and can be divided into the southern Scandes in SW Norway, with peaks reaching c. 2 km in the Jotunheim area, and the northern Scandes in northern Norway and Sweden with peaks reaching c. 1.4 km. These areas of very high topography are separated by the central Scandes in the Trondheim region which has more subdued topography reaching c. 1 km. The last major plate deforming event in the region was the opening of the northeast Atlantic at c. 55 Ma which involved extensive rifting and associated volcanism, and prior to this the formation of the North Sea rift system. Both these events involved lithospheric extension rather than continental collision which is more likely to have formed the mountain range. Consequently, the mountains might be regarded as an example of epeirogenic uplift. However, they have also been interpreted as the long-lived eroded remnants of the Caledonian mountain chain which formed in the late Silurian to early Devonian (Nielsen et al. 2009).

One of the major problems interpreting the origin of the mountain range is the absence of onshore Mesozoic sedimentary rocks in Norway which could be used to demonstrate that the Caledonian mountain range had been peneplaned, as has been argued by Lidmar-Bergström et al. (2000) and others. The application and interpretation of fission track dating to determine the timing of uplift is strongly debated, principally because of differences in modelling of track length data to determining cooling rates (Rohrmann et al. 1995; Redfield et al. 2005; Nielsen et al. 2009; Osmundsen et al. 2010). However, these data reveal two areas of up to 3 km of exhumation beginning in the Mesozoic (c. 205 Ma from Apatite fission track data (summarized in Redfield et al. 2005)) that are similar to the shape of the southern and northern Scandes. Offshore, the sedimentary record contains evidence, in the form of unconformities and prograding sediment packages, for periods of uplift followed by erosion and deposition during the Miocene and Pliocene. The ages of these events are difficult to determine and appear to be asynchronous along the continental margin, suggesting that the uplift is not synchronous along the length of the mountain range. What is clear is that the latest event has produced the most voluminous sediments and this is generally associated with the onset of Plio-Pleistocene
The unconformity at the base of the Plio-Pleistocene sediments cuts down into Miocene and older prograding wedges of sediment (Faleide et al. 2002) which presumably were built outwards from an elevated hinterland. Consequently, most researchers favour a Cenozoic age for the uplift to form the present mountain range which created the topographic and climatic conditions necessary for the Scandinavian ice sheet to nucleate. The alternative is that the present mountain range is the remnants of the Caledonian orogen which has been gradually eroded over time (Nielsen et al. 2009).

Various explanations and models have been proposed for this uplift, of which the most cited and supported by data are: Asthenospheric diapirism (Rohrmann & van der Beek 1996) which is supported in part by low upper mantle velocities identified in an analysis of surface wave tomography (Weidle & Maupin 2008); marginal flexure and associated faulting (Redfield et al. 2005; Osmundsen et al. 2010) which is based on field studies, flexural modelling and fission track data; rift flank uplift on the margins of the North Atlantic rift (Torske 1972; Doré 1992) based on tectonostratigraphic studies of the continental margin; and extended erosion of the Caledonian mountains (Nielsen et al. 2009) also based on fission track data.

The purpose of this contribution is to present an interpretation of seismic and gravity data along a profile crossing the central Scandinavia in central Norway and Sweden. This is one of a series of integrated active and passive seismic experiments (e.g. CEN-MOVE: Svenningsen et al. 2007; MAGNUS REX: Stratford et al. 2009; MAGNUS: Weidle et al. 2010) designed to determine the present day crust and mantle structure beneath Scandinavia and its relationship to present day topography. Once the crust and upper mantle structure is fully characterized by this series of experiments it may be possible to deduce the cause of the uplift.

2 THE SCANLIPS EXPERIMENT

In order to determine if there is a relationship between crustal structure, topography and uplift the SCANLIPS (SCANdinavian Lithosphere P and S) and SCANLIPS2 (in progress) experiments were designed to complement the work of Svenningsen et al. (2007) in southern Norway (Fig. 1). The two profiles in southern Norway cross the region of highest topography and uplift, the SCANLIPS profile crosses central Norway and Sweden where the topography and uplift is lowest and the SCANLIPS2 profile crosses intermediate topography and uplift in northern Norway and Sweden. In combination, the results from these profiles should enable a comparison to be made between varying levels of topography and uplift and the thickness and physical properties of the crust.

The SCANLIPS experiment builds on seismic refraction (Schmidt 2000) and reflection profiling (Hurich & Roberts 1997; Juhojuntti et al. 2001) in central Norway and Sweden which provide a priori constraints on the velocity structure of the crust and detailed information on the crustal structure beneath the Caledonides, respectively. The profile also crosses the FENNOLORA refraction profile in Sweden (Guggisberg et al. 1991).

Thirty-one Guralp 6TD 30 s period broadband 3 component seismometers (sampling at 100 Hz) were deployed between May and October 2006 along a 465 km long profile, striking c. 098 (Fig. 1). In order to try to avoid major complications from anisotropy in the geological structure the instrument profile was deployed as close as possible orthogonal to geological strike, which is oriented NNE—SSW. Of these, 28 instruments were on the west to east profile described in this paper. These stations were numbered N6001 at the western (Norwegian) end of the profile to N6029 at the eastern (Swedish) end of the profile. Station N6015 was not deployed and there is no station with this number. One instrument (N6012), deployed 12 km west of Storlien, failed completely, so the profile is formed from an array of 27 instruments separated by an average spacing of c. 20 km. Details of the instrument locations and relative distances along the profile are given in Table 1. In Norway and western Sweden most instruments were buried 1 m beneath the surface. The instruments at sites 20 to 25 and 27 to 29 were deployed on bedrock at the surface, protected by concrete pipes. Estimates of signal to noise at the different station sites were made from power spectral density plots of 2-d-long event-free time intervals of data recorded at the different sites. These plots were compared to mean and low- and high-seismic noise levels and between stations (England et al. 2008). Short period (0.1 to 1 Hz) cultural noise

Figure 1. (a) Topographic map of Scandinavia showing the Scandes mountains, the location of the SCANLIPS profile (red squares), the profiles described by Svenningsen et al. (2007) (black dots and blue triangles) and the SCANLIPS2 profile (white squares); thin blue lines are the trace of existing refraction profiles and white line joins the first and last shot points of the CABLES profile. (b) Locations of the SCANLIPS instruments plotted onto a map of the Bouguer gravity anomaly.
processing were calculated using spherical geometry. Back azimuths used to rotate the different components during velocity structure within the aperture sampled at each recording station, and Philippine trenches. This limits estimates of anisotropy of the 3 DATA PROCESSING

Data recovery from the instruments was around 95 per cent. Raw data were transcribed into miniseed after applying instrument response corrections and are archived with IRIS (Incorporated Research Institutions for Seismology) under the temporary array code YF.

A total of 25 good teleseismic events with body wave magnitudes greater than or equal to M b 6.0 and within an angular distance suitable for calculating receiver functions (30° to 90°) were recorded during the 6 month deployment. The locations of these events are shown in Fig. 2 and examples are shown in Fig. 3. As a result of the short duration of recording these events provide a limited range of back azimuths and a longer recording duration (ideally c. 18 months), not possible at the time of the experiment, would have yielded a larger number of useful events. Most events occurred to the north and east of the profile along the Japan, Marianas and Philippine trenches. This limits estimates of anisotropy of the velocity structure within the aperture sampled at each recording station. Back azimuths used to rotate the different components during processing were calculated using spherical geometry.

After applying a 0.3 to 3 Hz band pass filter to attempt to reduce noise levels ZRT (vertical, radial and transverse component) receiver functions were calculated for all 25 good events using the method of Ammon (1991). This method uses the source equalization procedure of Langston (1979), which involves rotating the three component waveforms to the back azimuth of the source path, isolating the near source effects on the vertical component and then deconvolving this from the horizontal components to leave the radial and transverse receiver functions. Langston (1979) used a ‘water level’ or spectral whitening to avoid division by zeros in the deconvolution. This has the disadvantage of attenuating frequencies with low amplitude, and as a consequence the water level was determined by testing to keep its value as low as possible. Good or useful receiver functions were determined visually as those radial functions with signals in the 30 or so seconds following the direct P arrival with amplitudes above that of the noise preceding the direct P arrival. The optimum water level used in this analysis was determined by finding the ‘averaging function’ (deconvolution of the vertical component with the water level applied by the vertical component without) which was closest to a Gaussian. The method also uses a low pass Gaussian filter to remove high-frequency noise which can be introduced by the water level deconvolution. This was again determined by visual examination of the radial receiver function and the averaging function. The optimum filter was found to have a width of 2 which approximates to a low pass frequency of 1 Hz. Examples of ZRT receiver functions from 3 stations showing a range of signal to noise ratios are given in Fig. 3.

In addition, following an initial migration, angles of incidence were accurately determined for each event/station pair and the seismograms were rotated into their LQT components (where L is the principle direction of particle motion of the P-wave; Q is the particle motion of a S-wave in the radial direction and T is the transverse component as before) and receiver functions were calculated for those teleseismic events with the highest signal to noise ratios (estimated by comparing amplitudes prior to and following the P arrival).

In theory, calculation of the LQT or P receiver function should isolate most of the Sv wave energy on the Q component (Langston 1977). The most noticeable effect of this is the removal of the direct P arrival observed at the start of ZRT receiver functions. Removal of this large amplitude component from the signal leaves P to S conversions from the upper crust visible in the first few seconds.

Table 1. Location of SCANLIPS seismic stations.

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<th>Longitude</th>
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Figure 2. Stereographic projection of the Earth centred on the SCANLIPS profile showing the locations of teleseismic events of magnitude greater than or equal to M b 6.0 occurring between 30 and 95 degrees which were recorded by the seismic stations on the profile.
of data after the direct $P$ arrival time. These mode conversions can then be used to better resolve the structure of the upper crust.

A further beneficial effect of the LQT transformation is the isolation of $S_{II}$ energy on the transverse component ($T$) of the receiver function. The magnitude of residual energy on this component is dependent on the generation of $S_{II}$ energy at interfaces in the subsurface. In an ideal case of crust in which the velocity structure varies only vertically (i.e. interfaces are horizontal) no $S_{II}$ energy should be produced from an incoming $P$-wave, so the presence of energy on the transverse component of the receiver function indicates non-horizontal and/or non-planar interfaces in the subsurface or anisotropy. Amplitude of energy on the transverse component is typically 1/10 of that on the radial ($R$) or $Q$ component suggesting that some anisotropy is present but that it has been minimized by the choice of profile orientation. The ideal case would rarely be approached in the crust due to the complexity of structures imaged on the reflection and refraction profiles (Hurich & Roberts 1997; Juhojuntti et al. 2001). However, detailed examination of the data

**Figure 3.** Teleseismic $P$-wave arrivals from events shown in Fig. 2 with $M_b$ greater than or equal to 6.0 recorded at stations N6001 (top), N6019 (middle) and N6020 (bottom). Arrival times are aligned at 0 s and the events are plotted in order of backazimuth, instrument gain corrections have been applied to give relative amplitudes. The plots of the right hand side of the figure show receiver functions calculated from the events on the left, using the method of Ammon (1991).
showed considerable variation in traces with different offsets and backazimuths, suggesting complexity in the crustal structure which the experiment had been designed to minimize.

4 MODELLING AND INTERPRETATION

4.1 Forward modelling

Receiver functions suffer from the typical non-uniqueness of geophysical inverse problems. Similar receiver functions can arise from complementary variations in seismic velocity/Poisson’s ratio and crustal layer thickness (Ammon et al. 1990). In order to overcome this, a priori data constraining the velocity structure were taken from the CABLES P-wave refraction profile (Schmidt 2000). The SCANLIPS profile closely follows this profile, except for the western 100 km (Fig. 1), so it should provide a reliable starting velocity model. However, the limited number of shots along the CABLES profile means that the ray coverage of the crust is not high, particularly at each end, causing large uncertainties in the modelled velocity depth structure. A further uncertainty in this model is the presence of a modelled low-velocity layer between 5 and 10 km depth along the central part of the profile. This causes uncertainties in ray paths traced through the model and weakens constraints on depth to interfaces and layer velocities below it.

One of the principle objectives of the SCANLIPS experiment was to obtain a more accurate thickness for the crust along the profile than was available from the existing CABLES data (Schmidt 2000). Due to poor signal to noise ratios in the data (Fig. 3) and complex crustal structure (discussed below), it was not easy to unambiguously identify PpPhs, PpShs and PsShs multiples in the receiver functions calculated from the teleseismic arrivals (Fig. 3) even by comparison with simple synthetic functions derived from an initial forward model. Consequently, it was decided to forward model the crustal velocity structure using the CABLES profile as a starting model rather than move to H-κ stacking (Zhu & Kanamori 2000). H-κ stacking would provide an estimate of average $V_p/V_s$ and crustal thickness but because these values are estimated from the differential moveout of $P_s$ and the crustal multiples with varying $V_p/V_s$, it depends on clearly resolved multiple events for accuracy.

For the forward modelling a number of initial starting models for $V_p/V_s$ versus depth, based on the CABLES profile (assuming a constant P-wave to S-wave velocity ratio of 1.73:1), were created at the following receiver positions along the SCANLIPS profile where the topography and known geology suggested a change in velocity structure: N6009 at the edge of the Scandes mountain range; N6011 the maximum topographic elevation along the profile; N6020 at the eastern edge of the Caledonian deformation front. While the velocity functions were based on the CABLES profile, they did not include the low-velocity zone at the near surface. Where there was no constraint from the CABLES profile because of its raypath geometry, that is the eastern and western most 60 km of the receiver function profile, the velocity structure was continued from the last good constraint. Synthetic three-component waveforms were calculated from the initial models using the respknt code written by Randall (1994). These waveforms were then processed in an identical way to the real data and synthetic ZRT receiver functions were calculated for each velocity model. These models were then visually compared to the real functions (Fig. 4). Particular emphasis

Figure 4. Examples of forward/synthetic modelling (in red) of receiver functions (in blue) at stations N6004 (a) and N6016 (b) in each case the Ps converted arrival from the base of the crust is labelled. The events following this are multiples from the crust or Moho. The lower plot is the $P$-wave velocity-depth profile formed from constant velocity layers derived from forward modelling. White dots are nodes in the velocity model.

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was paid to achieving a good fit between the observed and synthetic receiver functions in the first 20 s after the direct $P$ arrival during which time the synthetic functions suggested the highest amplitude multiples should occur. An effort was made to visually align peaks and troughs in the forward models with the observed data and less effort was applied to matching amplitudes, although in practice a reasonable match in amplitudes was also achieved at short delay times (10 to 15 s after the direct $P$ arrival). Examples are provided in Fig. 4. The most appropriate models for the crust beneath each receiver station were selected on the basis of the fit between the $P$s and multiple arrivals. If necessary the starting models were updated and the process repeated. In practice excellent fits were obtained without significant modification of the CABLES derived starting velocities (Fig. 4). Wherever possible the best fit model from adjacent stations was extrapolated to the next station along the profile. This resulted in a simple smoothed velocity model for the crust (Fig 4). One station, N6016, 233 km along the profile (Fig. 4) proved particularly difficult to fit with adjacent velocity functions. The reason for this is unclear. It is thought that it may result from complex subsurface geology. There are comparably higher energy levels on the transverse component of the receiver functions for events recorded at this station compared with many others.

4.2 Inverse modelling

While it is well known that inversion of receiver functions is highly non-unique (Ammon et al. 1990), the availability of a priori information from the CABLES profile and the set of well fitting forward models suggested that constrained inverse modelling using the well understood Ammon code might be possible. Using the forward models as a starting point for the inversion would help constrain the model space but some simple tests demonstrate the initial model dependence of this code. One of its limitations is that layer thicknesses do not vary during the inversion. To overcome this problem the starting models, which were based on the forward modelling results, were created with 2 km thick layers. The velocity in each layer was then allowed to vary by up to 20 per cent. The models were extended to 50 km depth to ensure that velocities could vary within the region of the Moho and thus also constrain its depth. Fig. 5 illustrates an example of how the initial layer thickness constraint can affect the final velocity model. In this case, the thickness of the upper crustal layer was fixed for one inversion and in the other it was allowed to vary. Synthetic receiver functions from both models fit the observations satisfactorily but a far better fit is achieved (for smoothed inversions) with the inversion using finer steps in velocity structure.

Before inversion the real data were decimated with an anti-alias filter to 10 Hz sample rate and the synthetic functions calculated during the inversion were filtered with the same parameters. Due to the relatively high noise levels in the data the inversion was only performed using receiver functions with the highest signal to noise ratio and relatively heavy smoothing constraints were applied to the velocity models. Parameter testing showed that convergence usually occurred within 4 to 6 iterations. Inversions were performed at each receiver station and the resulting best fit models were

![Figure 5](https://academic.oup.com/gji/article-abstract/191/1/1/585045)
averaged to give a 1-D velocity versus depth function beneath each station.

4.3 Migration

In order that a comparison with the receiver function profiles produced by Svenningsen et al. (2007) could be made, those LQT receiver functions with the highest signal to noise ratio and comparable P and multiple arrivals with a range of backazimuths from 9 to 40 degrees and offsets between 65 and 82 degrees were depth migrated using the common conversion point technique of Jones & Phinney (1998) and Kosarev et al. (1999). Each Q component of the receiver functions is back projected from the surface along an estimated incoming ray path calculated from a chosen velocity function. A similar velocity and Poisson’s ratio versus depth function as that used by Svenningsen et al. (2007) was used (i.e. 6.30 km s\(^{-1}\) and 0.26 between 0 and 20 km and 6.85 and 0.27 between 20 km and the Moho as determined from the forward modelling), although more detailed velocity-depth functions became available as the data interpretation progressed. However, it is unlikely that a more detailed velocity function would significantly change subsurface locations of mode conversion points in the crust, as demonstrated by the sensitivity test of Svenningsen et al. (2007). These authors showed that a 3 per cent variation in Vp and Poisson’s ratio resulted in an uncertainty in the Moho depth of ±1 km. Following migration the receiver functions were trace normalized to remove the effects of variable mean values between traces and to suppress particularly high amplitudes which would otherwise dominate the stack. The traces were then stacked and plotted (Fig. 6) using a regular grid-ded binning and Gaussian averaging technique based on the Fresnel zone for a wavelet with the dominant 1 Hz frequency of the data, again similar to that used by Svenningsen et al. (2007) in order to maintain comparability between the two studies. Elevation corrections were also applied at this stage so depths to mode conversion points are measured from sea level.

4.4 Interpretation

The depth migration of the receiver functions (Fig. 6) reveals a relatively simple crustal structure beneath Sweden. The Ps arrival (Moho P to S conversion) occurs at 40 to 45 km depth between 150 and 464 km profile distance. West of 150 km the Ps arrival is poorly defined, as can be seen in Fig. 3 from a comparison of the receiver functions from site N6001 (at 0 km) with those from N6019 and N6020 (at 292 and 315 km). However, the forward modelling allows it to be identified (see modelling of N6004 in Fig. 4), rising from c. 42 km at c. 100 km profile distance to c. 34 km at the coast. The short wavelength variation over the length of the profile is due to a lack of consistency in the receiver functions between adjacent stations and the limited number of good noise free events included in the stack, which leaves unfilled bins in the grid over which the data are combined. This is a particular problem to the west of 150 km and would only be resolved with more events with high signal to noise ratio and possibly using seismometers with lower internal noise levels, which was not possible on this temporary deployment. Within the limits of resolution of the data and the accuracy of the velocity model which, following Svenningsen et al. (2007) are estimated at ±1 km (see above), there is no evidence in the migrated data for a clear crustal root beneath the Scandinavian mountain range which could support the observed topography (Fig. 4).

The forward modelling results (Fig. 4) yield a velocity-depth model for the crust composed of constant velocity layers and enable identification of the Ps conversion in the migrated data. Beneath Norway, (0 km to c. 100 km) surface velocities are relatively low (5.6 km s\(^{-1}\)) compared with the rest of the profile (5.8 km s\(^{-1}\)) and the lower crustal velocities are also lower. Between 100 km and c. 250 km the near surface velocities are higher (5.8 km s\(^{-1}\)) and the 6.0 km s\(^{-1}\) contour lies at a greater depth. Lower crustal velocities also increase to the east over this interval. The modelling suggests a small crustal root of maximum amplitude of c. 5 km in this area.
East of 250 km the mid crustal velocities increase significantly and a high velocity \((V_p > 7.2 \text{ km s}^{-1})\) lower crustal layer approximately 8 km thick is required to achieve an acceptable fit to the observations. In this area a velocity model which broadly fitted the observations was determined and used for all receiver stations (Fig. 4). Although it should be noted that at some locations there are misfits between observed and synthetic receiver functions, particularly in fitting the multiple arrivals at travel times in excess of 20 s after the direct \(P\) arrival.

In general terms, the inversion results yielded a similar but smoother velocity model than the forward modelling with little variation between adjacent profiles. The inversion did not recover the high-velocity lower crustal layer but the gradient of the resulting velocity-depth models is shallow in this region and the Moho is also poorly defined (Fig. 5). This suggests the need for high velocities in this depth range. One of the surprising results of this analysis was that a low-velocity zone was required in order to match the observations between c. 100 and 200 km along the profile. This low-velocity zone was also observed on the CABLES profile over the same distance. Fig. 5 shows two inversions for station N6029, at the east end of the profile, which illustrate that a single thick high-velocity layer can be replaced by a series of thin layers with an average lower velocity. The fit between the observed and synthetic receiver functions is not as good for the thick high-velocity layer model. Given that an acceptable fit between observed and calculated synthetic receiver functions can be achieved by forward modelling using thick layers suggests that the low-velocity zone may be an artefact of the inverse modelling. However, a comparison with geological data (see discussion) indicates that this feature is justifiable.

The \(P\)-wave velocity models derived from the receiver functions and the CABLES wide angle data are broadly similar. In both cases the velocities beneath Norway, west of the Scandinavian mountain range, are lower. Both models show a clear three layer crustal structure in this area. The velocities in the receiver function model are consistently 0.1 to 0.3 \text{ km s}^{-1} lower than the CABLES model beneath Norway. A similar observation can be made for the crust beneath Sweden but the velocity structures indicated by the forward model from the receiver functions and the CABLES data are different. The receiver function model shows a relatively low-velocity upper and middle crust extending to c. 20 km depth underlain by a high-velocity lower crust and a very high-velocity lowermost crust \((V_p > 7.2 \text{ km s}^{-1})\) east of the Scandinavian mountains. The CABLES model, in contrast, does not contain the very high-velocity lower crust and instead the velocities throughout the crust are higher, with the exception of the near-surface low-velocity layer. The middle and upper crust is much shallower in the CABLES data beneath Sweden than beneath Norway, whereas in the receiver function derived model the middle and upper crust extend to deeper levels. The crust is about 43 km thick in both models. There is a small difference in the mean \(P\)-wave velocities of the crust beneath Sweden derived from the receiver function and wide-angle models which are 6.39 and 6.43 \text{ km s}^{-1}, respectively, but this is not considered significant given the likely uncertainties in the modelling methods.

5 Gravity Modelling

In order to test if the velocity model determined from the receiver functions was realistic the gravity anomaly along the profile was modelled using the velocities determined as a criterion on density. Any significant difference between the modelled and observed anomaly would suggest either an inappropriate velocity structure or a poor correlation between velocity and density in the region of the model. To do this, the layered velocity depth functions derived from forward modelling of the receiver functions was converted to density and the gravity effect of the model was then calculated. The velocity structure from the forward modelling is used because the Moho, which usually coincides with a major change in density, is best defined in this data set. The velocity to density conversion was made using the linear velocity-density equations developed by Christensen & Mooney (1995) and corrected by Zoback & Mooney (2003) for crystalline crust.

The gravity effect of the model shows a reasonable fit to the observed Bouguer anomaly on a regional scale (Fig. 7) without
any adjustment. No attempt has been made to iteratively forward model the gravity anomaly. The crustal configuration results in a minimum of −80 mGal as observed in the Bouguer anomaly (Olesen et al. 2010). To adjust for an E-W level misfit, which is also expressed in a steep gradient in the geoid (e.g. Ebbing & Olesen 2005), an east-west trend correction of 0.15 mGal km$^{-1}$ was necessary which mimics the eastwards deepening of the base of the lithosphere (e.g. Plomerova & Babuska 2006), which is not resolved by our model. After applying the trend correction, the model shows mostly local differences to the observed Bouguer anomaly. The modelled gravity minimum extends further to the west than the observed gravity, which is most likely an effect of near-surface geology (see discussion). The steep gradients in the observed Bouguer anomaly cannot be reproduced by the regional model derived from forward modelling of the receiver functions.

In general the velocity model results in densities which are higher in the eastern and central part of the profile than in the west, reflecting the depth dependency of the velocity-density conversion. However, the most significant contribution to the gravity anomaly is a high-density lower crustal layer beneath Sweden.

The top of the high-density lower crustal layer is, however, very similar to the Airy-isostatic Moho. The density model is similar to those presented in Ebbing (2007) and Pascal et al. (2007), where it was demonstrated that the high-density lower crust is necessary to achieve isostatic balance of the lithosphere below the Scandian mountains and to model the low in the Bouguer gravity anomaly which correlates with them. The observations based on the model described here confirm those previously obtained, but redefine the high-velocity layer, which has a maximum thickness of ~9 km. A more detailed discussion of the isostatic state of the whole of the Scandes is given in Ebbing et al. (2011).

6 DISCUSSION

The results presented here illustrate the crustal structure along the SCANLIPS profile on a broad scale and address the question of support for the observed topography. A study of the details of the crustal structure is ongoing but one of the surprising results of the inversion of the receiver function data was the presence of the low-velocity zone in the upper 10 km of the crust, which is in agreement with the best fit model to the seismic refraction data (Schmidt 2000). While there are uncertainties in modelling refraction data with low-velocity layers and inversion of receiver functions due to the trade off between velocity and layer thickness it is noted that early geophysical work on the Scandinavian Caledonides by Wolff (1984) recognized that high-density rocks rest on low-density rocks in the region to the east of Trondheim. Rocks of ophiolitic affinities in the upper allochthon with densities in the range 2730 to 2830 kg m$^{-3}$ were interpreted as overlying allochthonous basement rocks with densities in the range 2650 to 2750 kg m$^{-3}$. Simple 2-D gravity modelling, by Wolff (1984), using these values suggested the upper dense layer was between 8 and 12 km thick, depending upon the combination of densities used. Converting the densities to velocities would produce a model with a velocity inversion similar to that observed in the upper part of the inverse models derived from the receiver functions. However, given the uncertainties in the refraction and inverse modelling we have not incorporated the low-velocity layer into the model for the gravity data which is based on the forward modelling of the seismic data. The anomaly associated with this density inversion is visible as a region of slightly elevated gravity anomaly (arising from the relatively dense surface rocks) between 60 and 120 km within the overall gravity low associated with the topography. This anomaly is not reproduced by the large-scale gravity model presented in Fig. 7.

Within the limits of the data, the results of the forward modelling and the migration of the receiver functions show no evidence for a significant crustal root which could provide isostatic support beneath the mountains (Figs 4 and 6). Beneath the topographic high the crust is about 42 km thick and it remains at this thickness to the east beneath Sweden. Toward the Norwegian coast the forward modelling of the receiver functions and the gravity data suggest that the crust thins to about 34 km. This is also consistent with the velocity structure imaged by the CABLES seismic refraction profile (Schmidt 2000). Ottemöller & Midzi (2003) calculated receiver functions from stations across Norway. The nearest stations to the SCANLIPS profile (Molde (MOL) and Namsos (NSS)) are 160 and 140 km away but also suggest thinner (38 to 42 km) crust towards the coast. The positive arrivals from stations N6003, N6004, N6005 and N6007 seen in the migrated profile at depths of 40 to 42 km (Fig. 6) are the result of constructive stacking of a probablemultiple and destructive stacking of a shallower and weak P$_{0}$ (e.g. station N6004 in Fig. 4). On this basis the crust is not considered to be thick (i.e. 40 to 42 km) towards the coast. This change in crustal thickness could be explained by stretching, associated with rifting along the Norwegian margin. However, while there is some evidence for post Caledonian extension onshore (Osmundsen et al. 2010) there are no sedimentary basins which would record both timing and possibly the amount of extension. It is possible that the thinning of the crust along coastal Norway was the result of extensional collapse of the Caledonian orogen. Although, the elevation at which the Devonian basins, which are currently exposed near sea level, formed at the time of collapse of the orogen (Serrane & Seguret 1987), is not known. Until better timing constraints on the extension are available it is not possible to constrain when thinning of the crust beneath onshore Norway occurred.

As noted above, the crust has constant thickness beneath Sweden (Figs 3, 4 and 6) at about 42 km but the topography decreases gently from c. 800 m near the border with Norway to 0 m on the Gulf of Bothnia (Fig. 3). Modelling of the gravity anomaly (Fig. 6) suggests that this relatively low and gently decreasing topography over near constant thickness crust is compensated by the presence of a lower crustal layer of high-density rock. It is impossible to determine the age of this layer, but its high-seismic velocity suggests, by analogy with similar deep crustal high-velocity bodies observed on magmatic continental margins (White & McKenzie 1989), that it may be a layer of magmatic underplate that accumulated over time. The absence of post-Caledonian mafic magmatism in the form of feeder dykes or pipes or any volcanism across Sweden suggests that if this is a layer of magmatic underplate, it was added, in whole or in parts, prior to the orogeny. Larsen & Tullborg (1998) attribute post-Caledonian heating, indicated by Pb isotope and fission track studies, to burial of the present basement surface beneath sediments derived from the erosion of the Caledonide mountains, rather than to magmatic activity. Alternatively, it is noted that the crustal structure of the central part of the Fennoscandian shield is dominated by a high-velocity lower crust, which is thought to have formed during the Palaeoproterozoic Svecofennian orogeny (Kukkonen et al. 2008). These authors suggest that the present Moho dates to late Svecofennian time and was created by delamination of high-density eclogitized lower crust after tectonic collisional thickening. Densities deduced from seismic and gravity data suggest a very small
or no density contrast between the present lower crust and uppermost mantle. New geological evidence from kimberlite-hosted garnet xenocrysts supports the presence of a garnet population with compositions consistent with crustal eclogites. Hence, Kukkonen et al. (2008) interpret the high-velocity lower crust to be related to the presence of eclogite. Irrespective of its origin, the layer continues but thins beneath the highest topography which suggests a pre-Caledonian age and that it was modified by post-Caledonian extension. The crustal model presented here is consistent with the isostatic model of Ebbing (2007) which suggested the continuation of a high-density lower crustal body beneath central Norway, compensating topography. Consequently, our preferred explanation for the geometry of the Moho along the SCANLIPS profile is that it is the result of crustal extension which occurred during and/or after the end Caledonian orogenic shortening and crustal thickening. Seismic studies of the crustal thickness beneath the high topography of southern Norway (Svenningsen et al. 2007; Stratford et al. 2009) provide evidence for a small crustal root. The magnitude of this root is not sufficient to provide Airy isostatic support to the mountain range (Ebbing 2007; Ebbing et al. 2011) and it is offset relative to topography. It is possible that the root was originally larger but that it too was modified by post-Caledonian extensional collapse or rifting along the western margin of Norway and the Oslo graben.

The average seismic velocity and the velocity and density structure of the crust along the SCANLIPS profile (Figs 4 and 6) do not correlate with the pattern of crustal uplift and exhumation. There is no evidence for magmatic underplating solely beneath the mountain range, which would produce uplift (Cox 1980). The opposite is observed. The high-velocity lower crustal layer which could be magmatic underplately thins beneath the mountains and is most probably pre-Caledonian in age. A lateral change in crustal velocities and densities from higher to lower values immediately to the west of the mountain range is consistent with the observed gravity data (Fig. 6) but neither correlates with the published pattern of exhumation (Rohrman et al. 1995; Rohrman & van der Beek 1996). The low-velocity layer in the upper crust can be explained in terms of the Caledonian crustal structure and this also does not relate to the pattern of exhumation. These observations lead to the interpretation that the present day topography may not be the result of direct modification of the crust.

7 CONCLUSIONS

Migration, forward modelling and inverse modelling of receiver functions have been used to produce a model for the crustal velocity structure beneath central Norway and Sweden. This model shows thin crust beneath Norway transitioning to thick crust beneath Sweden below the Scandes mountains. Upper- and middle-crustal velocities are found to be lower beneath Norway and higher beneath Sweden, which is also underlain by a high-velocity lower crustal layer which thins beneath the mountains. A low-velocity zone in the upper crust is observed in the inversion results which correlates with a similar feature modelled from existing seismic refraction data. This velocity inversion can be explained in terms of relict structure from the Caledonian orogeny, but given the uncertainties in inverse modelling of receiver functions and the difficulties posed by low-velocity layers in refraction profiles this feature is treated with caution.

Conversion of the forward model for the velocity structure into density yields a calculated gravity anomaly which compares closely with the observed anomaly. No forward modelling of the gravity anomaly by modification of the velocity derived density structure was considered necessary.

Within the constraints of the data, there is no evidence for a significant root beneath central Norway and Sweden which would be supporting topography. Instead the variation in crustal thickness is interpreted as the result of crustal extension, the exact age of which is not known. There is also no correlation between the velocity and density structure of the crust and published estimates of exhumation across the mountain range.

Combining the observations that the present day topography is not related to crustal structure and the lack of a crustal root supporting the topography leads to the conclusion that the support for the present topography is either the result of a strong crust and/or lies deeper within the mantle lithosphere or the asthenosphere. This issue is the subject of larger, ongoing regional scale studies.

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