The deep structure of the Scandes and its relation to tectonic history and present-day topography

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A B S T R A C T

We review the results of the TopoScandiaDeep project, a component of the TOPOEUROPE project, in which we have studied the crustal and upper mantle structures of southern Norway in relation to its high topography. The Scandinavian Mountain Chain (the Scandes) is an intracontinental mountain chain at the western edge of the Baltic shield, and has its southern part located in southern Norway. The timing as well as the processes causing the formation of the Scandes are disputed. We bring new geophysical constraints to this issue by providing crustal and mantle seismic models for the area and by integrated modeling of the lithosphere and its potential deformation.

New maps of Moho depth and crustal seismic velocities have been compiled using data from refraction lines, P-receiver functions and noise cross-correlation. These results show a thickening of the crust from southwest to northeast and a small crustal root not directly located below the topographic high. P-, S- and surface wave tomography infer seismic mantle velocities lower than in normal shield structure, with a possible sharp boundary close to the Oslo Graben. These low velocities are imaged in the lithosphere and in the underlying mantle down to the 410 km discontinuity.

Integrated modeling of seismic models and gravity data shows that the low velocities below southern Norway are compatible with a change in lithosphere thickness from c. 100 km under southern Norway to nearly 200 km under southern Sweden, with possible additional differences in composition. The study also indicates that the topography can be isostatically sustained by the density distribution in the crust and lithospheric mantle. We argue that the lithospheric lateral variation has been present for at least 300 My and has had a significant influence on the localization of the topography, independently of the mechanism for uplift.

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1. Introduction

This paper summarizes the main results of the TopoScandiaDeep project, in which we have analyzed the crust and upper mantle structures of the southern Scandinavian mountain range (southern Scandes) in order to bring new constraints in the study of the link between the deep crustal and upper mantle properties and the present-day topography.

TopoScandiaDeep is part of TOPOEUROPE, the vision of which is to study the relation between Earth’s topography and surface, mantle and plate tectonics processes (Cloetingh et al., 2007). It is well established that the highest topography on Earth is located close to plate boundaries and results from the interaction between tectonic plates. The Alpine−Caucasus system in Europe is the result of such an orogenic process and it is the subject of many studies in TOPOEUROPE (ref to articles in same issue). Uplift and high topography areas are
however also found away from plate boundaries, as for example in the Transantarctic mountains (van Wijk et al., 2008), the Southern Rocky mountains (Eaton, 2008), southern Africa (Partridge and Maud, 1987), Brazil (Gallagher et al., 1994), Urals (Puchkov and Danukalova, 2009), and the Scandes (this study, Fig. 1). High topography far away from plate boundaries can be a remnant of ancient mountain belts or may have been created by epeirogenic uplift not related to a classical plate boundary setting. Most continental regions, and in particular cratonic areas, have a long and complex geological history. They have been formed by the accretion of plates and show the imprint of successive collisional, orogenic and rifting processes, leading to a complex structure and a non-uniform rheology. The consequences of forces acting on systems with a complex rheological structure may not be simple and modeling may be misleading if the modeled rheology is too simplified (Burov et al., 2007).

Pronounced topography is observed far from plate boundaries around the northeast Atlantic. High topography along the Scandinavian Peninsula, including most of Norway, with surface elevation up to around 2000–2500 m is among the most impressive topographic features of this region. Eastern Greenland also features high topography in the Palaeogene North Atlantic Igneous Province, which today has an average elevation of 2000 m with the highest peak (Gunnbjørn Fjeld) at around 3700 m altitude, higher than the maximum elevation of the ice cap. However, there is no general agreement about the processes of formation or the age of this topography.

The areas of high topography in Scandinavia are located mainly along or close to the trace of the Caledonides, a major mountain belt that formed at around 430 to 410 Ma (Mid Silurian to Early Devonian) due to the collision of Laurentia with Baltica/Avalonia (e.g. Roberts, 2003; Torsvik and Cocks, 2005) and whose remnants have been separated by the opening of the Atlantic Ocean and are presently located in eastern North America, Greenland, Scandinavia, Scotland and Ireland. A debate persists (see Anell et al., 2009 and references therein) whether high topography is old and basically a remnant of the mountain building processes of the Caledonian orogeny as recently argued by Nielsen et al. (2009), or whether the topography is geologically much younger, as is widely believed (e.g., Anell et al., 2010; Bonow et al., 2007; Dorè, 1992; Japsen and Chalmers, 2000; Japsen et al., 2012; Lidmar-Bergström et al., 2000; Stuevold and Eldholm, 1996). Faleide et al. (2002) provide evidence for strong recent rotation of sedimentary sequences offshore southern Norway and Anell et al. (2010) observe a recent increase in sedimentation rate from the sedimentary record offshore Norway which is indicative for uplift or increased erosion during the latest 5 My. In eastern Greenland, the highest topography is found within the area covered by the Palaeogene igneous province in which a pile of basaltic lavas up to 7 km thick was extruded close to sea level at around 60 Ma (Brooks, 2011). The presence of the lavas provides a maximum age, and perhaps a cause, for the observed topography. However, there are no comparable lavas on the Scandinavian peninsula. Several mechanisms

![Topography of Northwestern Europe](image)

**Fig. 1.** Topography and bathymetry of Northwestern Europe. Note the high elevations in southern Norway. The black rectangle indicates the area studied and corresponds to the frames of the tomographic maps in Figs. 12 and 14.
of uplift have been proposed in order to explain the presence of the topography, most of them being related to the rifting event and opening of the North Atlantic at around 55 Ma. Among the most important suggestions are asthenospheric diapirism, lithospheric delamination, magmatic underplating, intraplate compression and rift flank uplift (see model suggestions and reviews in Chery et al., 1992; Clift et al., 1998; Cloetigh et al., 1990; Gabrielsen et al., 2005; Nielsen et al., 2002, 2009; Rohrmann and van der Beek, 1996). In the model of Nielsen et al. (2009) climatic forcing is responsible for a time varying and long erosion history and uplift is explained as due mainly to isostatic response to erosional unloading.

The different hypotheses imply different types of isostatic support for the topographic load: e.g. crustal isostasy, lithospheric isostasy, dynamic support, and flexural support. In order to distinguish between the different models and potential mechanisms, detailed knowledge of the crustal and upper mantle structure is needed.

Here, we present an overview of the results of the TopoScandiaDeep project so far. A major part is the detailed mapping of the seismic structure of the crust and upper mantle of southern Norway. This has been done using both controlled-source seismic data onshore and partly offshore and passive seismic data. The passive data have been analyzed with various methods: P- and S-receiver functions, P- and S-body wave tomography, surface wave dispersion analysis from noise interferometry and from earthquake data. The results of these investigations from the surface down to 410 km depth will be presented. The implications of the seismic structure in terms of possible models of lateral variations in composition and temperature have been analyzed quantitatively by geophysical–petrological modeling that also constrains the model by its fit to the observed gravity, and the results of this modeling will be presented. Analog modeling used to test the implication of lateral lithospheric variations on uplift due to intracontinental stress will also be presented, before we summarize by discussing the possible implications of our results for the presence of the topography in southern Norway.

2. Tectonic setting

The Scandes are located at the western edge of the Fennoscandian Shield (also referred to as the Baltic Shield). The main structure of the study region can be seen in Fig. 2, where the main blocks that amalgamated to form the shield from Archean to Middle Proterozoic are displayed together with the Caledonian nappe structures. In southernmost Scandinavia, we find crust with a younger Phanerozoic cover.

Archean rocks with ages of up to more than 3 Ga are located in the northeastern part of the shield and there is a general age decrease from northeast to southwest (Gorbatchev and Bogdanova, 1993). The main area that has been studied here, the southern Scandes, is located in Southern Norway, in the Sveconorwegian province. This part was accreted to the shield during the 1.64–1.52 Ga Gothian and 1.52–1.48 Ga Telemarkian accretionary events and during the Sveconorwegian orogeny, dated to 1.14–0.97 Ga (Bingen et al., 2008). A major unit located both within and to the east of our study region is the Transscandinavian Igneous belt (TIB), a large granitoid belt that formed about 1.85 to 1.65 Ga (Lahtinen et al., 2008). This belt is a prominent feature in magnetic maps and has a significant influence on the gravity field due to its generally reduced density in the shallow crust (Henkel and Eriksson, 1987; Olesen et al., 2002; Pascal et al., 2007).

There is evidence for magmatic activity in the area at about 580 Ma (Dahlgren, 1994). Ultramafic dikes with carbonatite affinities were emplaced to the west of the present-day location of the Oslo Graben but the lateral extension of the area subject to magmatic activity may have been larger than indicated by the present-day surface traces. The formation of the Hedmark Basin at that time north of the Oslo Graben suggests an extensional regime in the Late Proterozoic (Dahlgren, 1994).

This phase of extension was followed by the crustal collision that resulted in the Caledonian orogeny that affected the western border of the Baltic Shield. This major orogeny, with a main phase between 430 and 410 Ma (Roberts, 2003; Gorsvik and Cocks, 2005), resulted from the closure of the Iapetus Ocean and collision of Laurentia (incl. Greenland) with Baltica. Extensive nappes cover the western edge of Baltica (Fig. 2). These nappes usually form the upper few kilometers of the crust, and the crust and mantle beneath are considered as autochthonous units. The Western Gneiss Region, along the northwest coast of southern Norway (Fig. 2), has been strongly reworked during the orogeny and is composed of crustal and mantle materials that were deeply buried during the orogeny and subsequently exhumed (Austreim et al., 1997).

Slightly before the closure of the Iapetus Ocean, at about 440 Ma, Baltica was affected by the closure of another sea, the Tornquist Sea, on its southwestern margin (Cocks and Gorsvik, 2006). The remnant of the closure of this ocean is probably located in the southeastern part of the North Sea (Balling, 2000; MONA LISA Working Group, 1997) and in the south of Denmark where dipping structures identify the suture zone (Abramovitz et al., 1997; Abramovitz and Thybo, 2000). To the east, southern Sweden this suture intersects the very well-marked boundary between the East-European platform and the younger Phanerozoic terranes of Central and Western Europe, known as the Tornquist–Teisseyre Zone (TTZ). A clear step in lithospheric thickness and thermal state, as well as a sharp transition in mantle velocity and density have been imaged along this transition zone, which makes it one of the major features of the European lithosphere (Berthelsen, 1988; Grad et al., 2002; Ziehlhuis and Nolet, 1994).

From about 300 to 240 Ma, southern Norway and Denmark were subject to rifting and magmatism, which formed the north–south oriented Oslo Graben in the southeastern part of South Norway (Fig. 2) and a series of other rift grabens in the North Sea region. These rifts
are part of a larger system of rifts that affected northwest Europe at that time (Neumann et al., 1992). The extensive magmatism associated with the rifting is an important source of information about the composition and thermal state of the mantle in the area at that time.

The last major tectonic event that affected the area is the opening of the North Atlantic Ocean; more precisely the Norwegian–Greenland Sea. This occurred in the early Eocene (55 Ma) after a long period of extension and rifting that lasted for the entire Mesozoic (Osmundsen and Ebbing, 2008 and ref. therein). The rifting took place parallel to the strike of the Caledonian suture, to form the rifted continental margin that we observe today.

3. Seismological models

3.1. Data and methodology

Location of most of the active and passive seismic data collected for this study is summarized in Fig. 3. A large part of the seismological data was acquired during the MAGNUS experiment: the temporary deployment, from 2006 to 2008, of a network of broadband seismological

stations in southern Norway (Weidle et al., 2010), supplemented by recordings at permanent stations. Data were also acquired for a shorter period of time in Sweden (DANSEIS experiment). Tomographic and receiver function studies also took advantage of the existence of recordings from previous temporary deployments whenever possible (see Medhus et al., 2012 for a complete list). In addition, three controlled-source seismic profiles (MAGNUS-REX) were acquired over southern Norway in October 2007 in order to add information about the crustal structure and sub-Moho seismic velocities in this region (Stratford et al., 2009). One of these lines was supplemented to the west by three OBS-profiles on the continental platform offshore in order to link the continental part to the well-studied offshore basins. The study also includes the acquisition of broadband data over two lines which traverse the mountain range further north (SCANLIPS 1 and 2, England and Ebbing, 2012). The seismological data are complemented by potential field data (Ebbing et al., 2012, and references therein).

The seismological data are analyzed by several teams using different techniques in order to map different characteristics of the subsurface, and to compare the independently obtained results. The controlled-source lines were used to constrain the P-wave and S-wave velocities in the crust and refine Moho depth in the study area (Stratford and Thybo, 2011a, 2011b; Stratford et al., 2009). Dispersion characteristics of short-period surface waves were extracted from ambient seismic noise (Köhler et al., 2011). Combined with earthquake surface wave data, these dispersion relations were used to constrain S-wave velocity (Vs) and anisotropy in the crust and upper mantle, as well as Moho depth (Köhler et al., 2012b). Earthquake surface wave data were also used to generate a model of the average upper mantle S-wave velocity below southern Norway (Maupin, 2011). Moho depth and mantle discontinuities were analyzed using P-wave (Frassetto and Thybo, 2010, submitted for publication) and S-wave receiver functions (Wawerzinek, 2012). In addition, receiver functions were analyzed along the SCANLIPS lines further north (England and Ebbing, 2012). A tomographic inversion of the arrivals of the P-waves from teleseismic earthquakes was used to constrain both the relative variations in mantle P-wave velocities (Vp) and their absolute values (Medhus et al., 2009, 2012). A 3-D model of the mantle S-wave velocity contrasts was determined by teleseismic S-wave tomography (Wawerzinek et al., 2013). Analysis of SKS-splitting was also performed to analyze seismic anisotropy (Roy and Ritter, 2013).

The different methods were applied by the different research groups to generate individual models independently of each other. As we will show, these different models show a remarkable coherency. This gives us confidence that the different data sets give robust information. Still, it is important to keep in mind that the different techniques used here have different sensitivities and resolutions and consequently different strengths and weaknesses. In order to benefit fully from the complementarity of the different data sets, it is desirable in a second stage to combine all the data and seek a model that can explain all the data together. Considering the coherency that we have obtained so far, we are confident that such an analysis will be fruitful and will help in constraining variations in the physical properties of the crust and upper mantle, in particular in a better evaluation of the magnitude of the velocity anomalies. The magnitude is a key parameter required for a quantitative geodynamical interpretation of the models put forward to explain the topography but is often difficult to ascertain in individual studies. In this article, we concentrate on summarizing and comparing the results of the individual studies with emphasis on the models obtained. We refer to the individual articles for details concerning the analysis procedures, the resolution tests and for more details about the models.

3.2. Crustal velocities

P- and S-wave crustal velocities as well as Poisson’s ratio were determined from the controlled-source data along three profiles
The location of the interfaces between layers found by analysis of the P-wave data was used without modification in the S-wave models. The S-wave velocity models (Fig. 4) are here compared with corresponding profiles in the 3-D S-wave velocity model obtained from teleseismic and ambient noise surface wave data (Köhler et al., 2012b) (Fig. 5). The exceptional quality of the S-wave arrivals in the MAGNUS-Rex data made it possible to construct S-wave profiles along the three lines down to Moho depth, providing an excellent opportunity to directly compare the models derived from the controlled-source experiment to those derived from the surface waves. The strong S-waves are mainly observed for explosions in the area with Caledonian structures, and we ascribe their generation to the conversion of P to S waves due to stronger impedance contrasts in the extensive series of thin nappes in the area. The vertical sections through the 3D surface-wave model are straight lines following closely, but not exactly, the refraction lines (see Fig. 3). Line 2 in particular follows the maximum topography and its southern part is further east than the refraction profile to remain in a well resolved part of the surface-wave model. An additional EW section, called line 4 and situated in the center of the study area, is shown in Fig. 5. In addition, we show a horizontal cross-section at 5 km depth of the 3D model in Fig. 6.

The two methods give very consistent Vs models, showing a crustal structure without strong lateral variations. A large portion of the crust consists of a layer with velocities increasing from 3.65 km/s at the top to about 3.85 km/s at the bottom. This layer is bound at the top by a layer with velocities of around 3.4 to 3.6 km/s and thickness of a few kilometers and at the bottom by a thin lower crustal layer with velocities of about 4.0 km/s. This last layer has not been observed directly by seismic refraction analysis, suggesting a thickness not larger than 2 km. The Poisson’s ratio increases from 0.25 in the upper half of the crust to 0.27 in the thin bottom layer (Stratford and Thybo, 2011a, 2011b). This result is consistent with similar observations at presently active rift zones in Siberia and east Africa (Thybo and Nielsen, 2009; Thybo et al., 2000). The relatively high velocities in the upper part of the crust are consistent with the absence of sediments in the area. The lower crust is particularly thin. It thickens below the Oslo Graben where we do not have an indication of a significant Moho shallowing but rather of a thickening of the lower crust. The velocities in the Oslo Graben upper crust are lower than in the surrounding areas, as can be seen in the horizontal section (Fig. 6). Upper crust S-wave velocities are also lower in western Norway. We have verified that the dispersion of the surface waves is not biased by the large relief present in this area (Köhler et al., 2012a) and that these reduced velocities are therefore not an artifact of the relief. Low velocities are also found in the controlled-source profiles, where the low-velocity region has a slightly higher Poisson’s ratio of 0.255 as compared to the values of 0.25 further east (Fig. 4). A similar transition from west to east is also seen in line 4 where we observe a sharp transition, located at horizontal distance 300 km on Fig. 5, from a thick low-velocity upper layer in the west to a very thin layer in the east, overlying a middle crust of low velocity compared to the west. The low S-wave velocities and high Poisson’s ratio in the upper crust in western Norway could result from reworking of the basement at the time of the Caledonian orogeny (Stratford and Thybo, 2011b).

The seismic structure of the onshore and offshore transition into the thinned crust of the Møre basin along controlled-source line 1 (Fig. 7) shows that the transition from normal crustal thickness of about 40 km to a thinned crystalline crust (light blue layer labeled “Crust” in Fig. 7) of no more than 8 km occurs over an approximately 150 km wide transition zone. The zone of strongest thinning is located west of the coastline. Beneath and landward of the coastline, evidence is found for a two-layered crystalline crust, with Vp of 5.5–6.0 km/s and 6.6 km/s, respectively. These velocities are indicative of granitic–granodioritic rocks. The average crystalline Vp beneath the Møre Basin is 6.4 km/s, suggesting that these felsic rocks extend westward to the Møre Marginal High. The thinned crust in the Møre Basin is overlaid by a sedimentary package of up to 10 km thickness, similar to the observations in the Voring Basin further north (e.g. Mjelde et al., 2009). The sedimentary succession is dominantly of Cretaceous age, but also contains significant amounts of Devonian–Jurassic and Cenozoic rocks.
3.3. Moho depth

Moho depth is constrained by the controlled-source profiles and surface wave velocities discussed in the previous sections. Additional information is obtained from P- and S-wave receiver functions (Frassetto and Thybo, submitted for publication; Wawerzinek, 2012). Fig. 8 is a comparison of the Moho depths found independently by controlled-source seismics, P-receiver function and surface wave dispersion analysis. The map based on the controlled-source experiments results from interpolation between the three MAGNUS-REX lines and a series of older refraction profiles summarized by Kinck et al. (1993). The location of the complete set of lines (Stratford et al., 2009) shows that the map is based on a good coverage of data except for two spots: the high topography area at 7–9°E and 60–61°N; an area of the Western Gneiss Region that will be filled when the analysis of the data from the profiles shown in Fig. 7 will be finalized. The P-receiver function analysis was performed with data from all MAGNUS stations and other temporary and permanent stations in the area (Frassetto and Thybo, submitted for publication) and covers the whole area. The results provide an areal coverage that extends the Moho depth models obtained by P-wave receiver functions along the two CENMOVE lines (Svønningen et al., 2007).

In southern Norway, in the southwestern part of the study region, there is good agreement between the Moho depths found by the three methods. The seismic refraction map is smoother than the two others due to its construction by interpolation between profiles and this does not mean that it infers a smoother Moho discontinuity. In general, the receiver functions indicate a 2–3 km deeper Moho than the refraction seismic model and the surface wave model shows intermediate values. The uncertainty for Moho depth estimates along the refraction lines is ±1 km or ±2 km, depending on the location (Stratford et al., 2009). The values derived from receiver functions have an uncertainty estimated at ±2.5 km, mainly due to uncertainty in mean crustal velocities. The uncertainty is more difficult to evaluate for surface wave studies due to larger trade-off with velocity. Moho depth is at around 30 km depth along the coast and deepens to about 40 km towards the northeast. The crustal "root" at 42–43 km depth in the receiver function model at 8–10°E and 60–61°N cannot be confirmed by refraction studies due to a lack of data in this region. Its continuation further north, at about 62°N, is however located in an area well covered by seismic refraction lines and there seems to be a discrepancy of about 5 km in this region that cannot be simply ascribed to the uncertainties of both methods. All maps show thinning of the crust in the Oslo Graben area with some variation possibly related to a misinterpretation of the phases related to the top of the lower crust in the receiver functions. The limited crustal thinning associated with the Oslo Graben (<3 km crustal thinning), is similar to the findings in other rift grabens (Kirschner et al., 2011; Thybo and Nielsen, 2009; Thybo et al., 2000).

The correlation between the three maps is poorer in Sweden and in the northeastern corner of our study area where Moho depth is 45 km in the models derived from refraction and surface wave data, whereas the receiver functions infer an up to 60 km thick crust. The analysis of the receiver functions shows however that the converted...
S-waves are weak in the area of strong discrepancy, indicating a lack of strong impedance contrast at the base of the crust. The presence of a very high velocity layer in the lower crust (e.g. BABEL Working Group, 1993) or an extended gradient from the lower crust to the mantle (Olsson et al., 2008) could reduce the Moho to a second-order discontinuity that does not generate significant phase conversions. We notice that the sharp increase in Moho depth eastwards in the receiver function model corresponds to the boundary between the Sveconorwegian domain to the west and the Transscandinavian Igneous Belt (TIB) in the Baltic Shield proper in the east. In the northeastern most corner of the area covered with the receiver functions, eastward of the TIB, Moho depths are again around 40 km, in accordance with the results obtained with P-receiver functions by Olsson et al. (2008). Our results are also in general agreement with those from Olsson et al. (2008) in the small geographical overlap area to the east of the Oslo Graben. In conclusion, in the areas with high resolution, the Moho depth maps obtained by the three methods match on average and show a deepening towards the north-east. In the TIB area, the Moho is a weak converter and therefore less well constrained. This region is poorly covered by controlled source seismic data. Receiver functions indicate that Moho depth might vary locally more than shown in the interpolated map based on seismic refraction data. The Moho depths obtained by S-receiver functions (Wawerzinek, 2012) are not shown since they are not as well resolved as in the three methods presented here, but they indicate an average depth of 37 km, in good agreement with the other methods, and a slight deepening towards northeast.

The thickening of the crust towards the NE does not correlate well with the maximum topography, and the small root is shifted by about 60 km eastwards of the maximum topography (see Fig. 4 line 1 for ex.). This implies that the topography in Norway is not simply compensated by Airy type isostasy. However, the Bouguer anomaly low over southern Norway correlates strongly with the high topography, which indicates a high degree of isostatic compensation (e.g. Balling, 1980).

Fig. 7. P-wave velocity model across the Møre continental platform along profile 1. See inset for the location of the profile. This profile is the continuation of Magnus-Rex profile 1 shown in Figs. 4 and 5.

Fig. 8. Moho depths maps obtained from a) refraction-seismic profiles (Stratford et al., 2009) combined with the compilation of Kinck et al. (1993), b) P-wave receiver functions (Frassetto and Thybo, submitted for publication) and c) surface wave dispersion analysis (Köhler et al., 2012b). The color scale has been cut at 55 km in order to give appropriate contrasts although a few points in panel b) are at depths between 55 and 60 km. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
Lateral variation within the crust is a key to resolve this apparent contradiction. A significant contributor to the gravity in this region is a high-density lower crust located to the east of the Scandes, which tapers out beneath the high topography, and which ensures that isostatic calculations for the Fennoscandian shield show a high degree of isostatic compensation (Ebbing, 2007; Ebbing et al., 2012; Frassetto and Thybo, submitted for publication; Stratford and Thybo, 2011a). Ebbing et al. (2012) show that an Airy–Pratt isostatic model with a laterally varying crust and lithospheric thickness may largely explain the gravity field.

Further north, the presence of the high-velocity lower crust was confirmed along the central Norway SCANLIPS profile (profile SCANLIPS 1 on Fig. 3) (England and Ebbing, 2012). By forward and inverse modeling and migration of the receiver functions a model was established which shows the crust thickening from about 32 km at the Norwegian coast to about 43 km beneath the central Scandinavian mountain range and then remaining constant beneath Sweden (Fig. 9). The model shows furthermore good evidence for a high-velocity lower crust underlying much of Sweden which thins to the west beneath Norway. Preliminary results for the northern SCANLIPS 2 profile show a similar change in Moho depth from the coast towards Sweden. A high-velocity lower crust is not detected as a discrete layer as for the southern profile, but is also required here on the basis of isostatic and gravity considerations.

These results stress the importance of taking several factors into account for assessing the isostatic balance of an area. The 60 km eastward shift of the maximum of the topography compared to the maximum Moho depth indicates that flexural rigidity of the lithosphere may play a role in sustaining the topography, or, alternatively, that other mechanisms sustain the asymmetric topography.

3.4. Upper mantle seismic velocities

The seismic velocities in the upper mantle were mostly derived from teleseismic P-wave tomography (Medhus et al., 2012), teleseismic S-wave tomography (Wawerzinek et al., 2013), and surface wave analysis (Köhler et al., 2012b; Maupin, 2011). Sub-Moho seismic velocities are also constrained by the controlled-source experiments that infer velocities of 8.05 to 8.15 km/s for P-waves and relatively high velocities of 4.65 to 4.70 km/s for S-waves in the center of southern Norway. A larger range of sub-Moho seismic velocities is observed from the surface wave dispersion tomography (Fig. 5) but the trade-off with Moho depth in dispersion analysis makes it difficult to interpret these values with confidence, independently of other information. This trade-off is also apparent in the models derived from the analysis of the average dispersion of the Rayleigh waves in southern Norway (Maupin, 2011): for average Moho depth varying from 35 to 41 km, the inferred average sub-Moho S-wave velocity varies from 4.45 to 4.6 km/s. Although the model with Moho depth fixed at 41 km has the best agreement with the average lower crustal and mantle sub-Moho velocities found by the controlled-source experiments, we show in Fig. 10 the model obtained by fixing Moho depth from 35 to 39 km, in better agreement with the Moho depth maps presented above.

The most notable feature of the model in Fig. 10, and which is not strongly affected by a trade-off with Moho depth, is a low velocity zone in the depth range 100 to 200 km. The depth-averaged S-wave velocity in this depth range is estimated with an uncertainty of about 2% (Maupin, 2011). Shear-wave velocities of 4.4 km/s are inferred, lower than the velocities in model ak135 (Kennett et al., 1995) by 2%.
and notably lower (almost 4%) than the range expected for Proterozoic fold belts of similar age (4.5 to 4.75 km/s in the depth range between 100 km and 150 km (Lebedev et al., 2009)) or in older cratonic areas where S-wave velocities range between 4.60 km/s and at least 4.80 km/s are expected (Lebedev et al., 2009; Pedersen et al., 2009). This average S-wave model (Fig. 10) has been built with emphasis on using surface-wave recordings of up to 200 s periods to obtain the best possible depth resolution at the expense of the lateral resolution. The lateral variations are analyzed in a tomographic study using surface waves up to 67 s period (Köhler et al., 2012b). This tomographic model also shows low velocities at 110 km depth (Fig. 11), albeit the anomaly is weaker than in the averaged model. It shows S-wave velocities varying from about 4.50 km/s under southern Norway (at the upper limit of uncertainty of the average model) to 4.60 km/s across the Swedish border at 110 km depth.

The teleseismic S-wave tomography model also contains lower velocities below southern Norway than further east beneath the Baltic Shield (Figs. 12 and 13). The velocity contrast reaches about 3 to 4% and dips downwards from west to east (Fig. 13). At shallow depths (35–120 km), the reduced S-wave velocities are located in the western half of southern Norway and to the north. The location of this low S-wave velocity correlates very well with an area of low mantle density (Artemieva et al., 2006) and to where Ebbing et al. (2012) find the largest imbalance in the isotropic equilibrium of their model and propose the presence of a low-density mantle material to explain the gravity field in addition to the lithospheric thickness variation. The general velocity increase from west to east is in overall agreement with findings from surface wave data in the central area where the surface-wave tomography has good resolution (Fig. 11).

The low S-wave velocity anomaly increases in amplitude with depth (Figs. 12 and 13) and reaches its maximum amplitude at about 400 km depth. Resolution tests (Wawerzinek, 2012; Wawerzinek et al., 2013) support this finding which may indicate an increased temperature at the top of the mantle transition zone and which then would predict a deeper 410 km discontinuity compared to average Earth models (Bina and Helffrich, 1994).

Depth slices of the velocity model from the P-wave teleseismic body wave tomography (Fig. 14; Medhus et al., 2012) also show a transition from lower velocities in the west below southern Norway to higher velocities to the east. Recordings from additional stations in Denmark compared to the S-wave tomography provide resolution further south and show that the low upper mantle velocities extend further south, beneath the deep basins in Denmark. The teleseismic tomography inversion method used here (Medhus et al., 2012) includes both relative and absolute travel-times, and therefore provides both the usual Vp model relative to an unknown regional average and an absolute model with respect to the ak135 model (Kennett et al., 1995). The relative and absolute velocity models are not fundamentally different and lead to similar interpretations. It is shown that the relative model has best resolution in the upper part of the model (100–200 km depth range) and that the absolute model, shown in Fig. 14, has best resolution at a depth of 300–400 km (Medhus et al., 2012). The difference in resolution between the two models results from the absolute traveltime method being able to include less steep rays and therefore additional crossing points at depth. The major contrast across the Tornquist–Sorgenfrei transition zone is well imaged in the 100 to 300 km depth range, in addition to a low-velocity anomaly west of the Oslo Graben, where the velocity is lower than in the ak135 reference model by 1–2%. Amplitude variation of up to about ±2% persists down to at least 400 km depth, where the model shows an increase in velocity from west to east in the southern part of southern Norway and northern Denmark.

There is a clear indication that the mantle below southern Norway has lower seismic velocities than expected beneath a Proterozoic fold belt. This is a very consistent result that we obtain with all the techniques used to analyze the seismological data and is valid for P-waves as well as for S-waves. The exact amplitude and location of the low velocity are however not completely consistent between studies. The surface wave average model infers low velocities in the depth range 100 km down to at least its maximum resolution depth of 200–250 km. In combination with the results from the controlled source experiments, this model also suggests that the low velocities do not extend up to Moho depth. The P- and S-body-wave tomographic models suffer from lack of resolution above 100 km depth and infer low velocities from 100 km depth down to at least 400 km depth. The locations of the anomalies centered below southern Norway are in agreement between 200 and 350 km, depth but the shallowest part of the anomaly is shifted eastwards in the P-wave model compared to the S-wave model and the deeper parts of the P-models show highest anomaly amplitudes shifted to the southwest. The overall amplitudes of about 3% of the relative variations are very similar in the two body-wave tomographic models, an unexpected feature for thermally induced anomalies (e.g. Goes et al., 2004), suggesting the possibility of additional compositional variations. The amplitude of the velocity anomalies does not decrease with depth and even increases in the S-wave body-wave tomographic model (Fig. 12). The discrepancy between these models may be related to different resolutions and a joint data analysis is necessary to study if surface wave dispersions and P- and S-delay times are compatible with a unique model. In considering the fit of amplitudes, it is noted that the P-wave velocity (Fig. 14) is particularly slow in the 100–200 km depth range below southern Norway, and that this property correlates with the low-velocity zone in the average surface wave model (Fig. 10). The lower contrast of 1% in the S-wave body-wave tomography model (Fig. 12) in this depth range is on the other hand in better agreement with the smaller contrast found in the surface wave tomographic model (Fig. 11).

Although not documented as clearly and unequivocally before, a low-velocity mantle anomaly in southern Norway is not unique to our models. Low P-wave velocity in the sub-crustal mantle below southern Norway was proposed by Bannister et al. (1991) although

![Fig. 11. Horizontal slice at 110 km depth of the SV-wave velocity model obtained from the analysis of surface wave dispersions (Köhler et al., 2012b).](image)
their method did not permit a precise depth constraint on the low velocities. A strong upper mantle low-velocity anomaly is also present in the area in the regional surface wave tomographic model of Weidle and Maupin (2008). This low velocity anomaly can also be traced in several recent global models (Ritsema et al., 2011 in particular at 400 km depth); Amaru, 2007; van der Meer et al., 2010) and in large-scale...
The P-wave velocity below southern Norway is also at about 4.4 km/s, but is a transition region from lower velocities to the north to higher velocities to the south. A low-velocity develops in this model below southern Norway only from 200 km down to 330 km depth. The results of the present project have constrained a sharp pronounced transition between the low velocities in southern Norway and normal shield velocities further east. This first order observation is crucial for interpretation of the tectonothermal history of the lithosphere and possible implications for the evolution of high topography. We should also note that similar low velocity zones have previously been identified in the whole Baltic Shield (Abramovitz et al., 2002; Perchuck and Thybo, 1996) and even globally (Rychert and Shearer, 2009; Thybo, 2006).

3.5. Upper mantle seismic discontinuities

The P- and S-wave receiver functions (Frassetto and Thybo, 2010; Wawerzinek, 2012) address discontinuities in the entire upper mantle. Besides Moho, the S-wave receiver functions image mostly negative discontinuities down to a depth of 250 km (Fig. 15). From 58°N to 60°N, we observe a major negative discontinuity (decreasing velocity with depth) at 70 to 90 km depth with a decreasing velocity contrast dipping towards north and another one at 100 to 150 km depth. These discontinuities have a tendency to deepen towards the east. The P-wave receiver functions show a similar trend of two sets of negative discontinuities, deepening to the north east from 50 to 100 km depth and from 150 to 200 km depth (Fig. 16), perhaps with a weak positive discontinuity in between. It is not straightforward to associate any of these discontinuities with the lithosphere-asthenosphere boundary (LAB). In addition, detailed analysis of the compatibility of the two data sets requires a common data inversion that has not been carried out yet. Still, the receiver functions clearly indicate the prevalence of negative discontinuities, in agreement with the profiles obtained from the surface wave dispersions (Fig. 10) that show a low-velocity zone in the depth range of 100 to 200 km, and a deepening of the pattern towards the northeast. This is also in agreement with previous inferences of the thickening of the lithosphere in this direction (Artemieva et al., 2006; Calcañile, 1982) and negative discontinuities found further east beneath Sweden (Olsson et al., 2006).

The P-receiver functions also image the 410-km discontinuity (Fig. 16), which shows a deepening of about 20 km in the center of the study area that correlates quite well in location and amplitude with the maximum of low-velocity in the teleseismic S-wave model (Fig. 12). The P-wave velocity model also shows a low-velocity in this area, but further southwest (Fig. 14). The 410-km discontinuity has been found to be at a normal depth of about 410 km in northern Germany (Alinaghi et al., 2003; Grunewald et al., 2001) as well as in the northern North Sea, just to the west of our study area where a depth of 414 ± 5 km has been inferred (Helffrich et al., 2003).

A 20 km downward shift of the 410-discontinuity corresponds to a temperature increase of about 200 °C (Helffrich, 2000). Since migration velocities have been taken from the tomographic models, this shift is not related to errors caused by too high migration velocities. A 200 °C temperature contrast at that depth is predicted to produce velocity variations of ± 1.0% for S-waves and of ± 0.5% for P-waves (Goes et al., 2004). The velocity variation will be slightly larger at 200 km for the same temperature contrast: ± 1.1% for S-waves and ± 0.7% for P-waves. In the asthenosphere, the S-wave velocity variations (of ± 1.5% to 2%) that are observed are mostly consistent with rather large temperature variations of the order of 200 to 300 °C. The observation that Vp varies more than Vs, however, is not compatible with purely thermal variations and would require additional elements such as variations in composition.

4. Temperature, density and composition of the mantle

An interpretation of seismological models in terms of thermo-chemical anomalies of the mantle requires knowledge of the density structure given different sensitivities of seismic velocities and density to variations in temperature, mineral composition, fluid phase and melt. In order to study how the upper mantle velocity variations relate to the thermal structure and composition of the lithosphere, thermo-isostatic modeling and geophysical–petrological forward modeling were performed. The main ingredients in these modeling methods are observed topography and gravity and/or geoid anomalies that are used together with the seismic velocities to constrain heat-transfer and thermodynamically–consistent models of the composition and temperature of the lithosphere. In a 1-D inversion study by Kolstrup et al. (2012) subsurface temperature, temperature-
dependent density and seismic velocities were calculated from a set of input parameters (topography, Moho depth, geoid height and heat flow) assuming isostatic and thermal equilibrium. Their results suggest a thickening of the thermal lithosphere below southern Norway from west to east. The western part is found to have higher temperatures, lower densities and lower synthetic S-wave velocities than the eastern part, compatible with results from the P-wave travel time tomographic study (Medhus et al., 2012). Comparison of the synthetic S-velocity profiles beneath southwestern Norway with velocity profiles inverted from Rayleigh wave dispersion data suggests that the higher temperatures associated with a thinner lithosphere can explain parts of the seismic low-velocity anomaly.

Fig. 14. P-wave tomographic model of Medhus et al. (2012). This figure shows the absolute variations with respect to model ak135 in 4 different depth ranges.
This study was subsequently extended to 3D using a direct modeling approach and a self-consistent 3D subsurface model was developed to investigate the observed velocity variations between Norway and Sweden in conjunction with gravity and geoid observations (Gradmann et al., submitted for publication). This modeling approach makes it possible to analyze the thermal and compositional properties of the lithospheric mantle that are compatible with an isostatic state of equilibrium, observed topography, gravity, and seismic velocities. The model was built using combined geophysical–petrological forward modeling of the lithosphere and sublithospheric upper mantle (LitMod3D: Afonso et al., 2008; Fullea et al., 2009) and covers southern Norway and southern Sweden (Fig. 17). The Moho depth is based on the results of the MAGNUS-REX survey (Stratford et al., 2009) as well as the compilations by Kinck et al. (1993). The geometry of the Lithosphere–Asthenosphere Boundary (LAB), defining the position of the 1300 °C isotherm, is adjusted during the modeling since it is not well constrained from previous studies (Artemieva, 2006; Calcagnile, 1982; Plomerová et al., 2008). Thermophysical properties of the mantle are derived from several possible characteristic chemical compositions (Afonso et al., 2008). The chemical compositions are adjusted during the modeling to fit the observed data. In the model presented here (Fig. 17), the isostatic state and gravity field can be explained by the crustal architecture and long-wavelength trends in uppermost mantle structure and composition.

A major temperature difference between southern Norway and southern Sweden is required in order to reproduce the variations in seismic velocity observed in the area, in particular in the P-wave tomographic model of Medhus et al. (2012) and in the velocity-depth profiles of Maupin (2011) compared to profiles further east (Cotte et al., 2002; Pedersen et al., 2009). Considering the lithosphere as a thermally conductive layer in steady-state equilibrium, temperature distribution in the lithosphere is directly controlled by its thickness. The observed velocity variation requires a major change in lithosphere thickness from approximately around 100 km under southern Norway to nearly 200 km under southern Sweden (Fig. 17b).

In addition, in order to satisfy isostatic equilibrium and to match geoid and gravity data, a change in mantle composition from Phanerozoic-type (in the sense of Griffin et al., 1998) under southern Norway to Proterozoic-type (more depleted, lighter) under southern Sweden is required. The seismic velocity is here primarily controlled by the temperature field and only to a small degree by the composition, i.e. by changes in density and elastic moduli (Fig. 16). We do not consider the effects of compositional variations in minor elements or water content (Goes et al., 2000). Minor changes in the thermophysical properties of the SubContinental Lithospheric Mantle (SCLM) are considered, as they additionally control the temperature field (Hieronymus et al., 2007). The gravity data and topography signature also suggest that the lateral transition from thick Proterozoic to thin Phanerozoic SCLM is not subvertical, and can be represented by older SCLM overlying younger SCLM in an ~100 km wide transition zone (Fig. 17).

The present model has been constrained to satisfy the S-wave velocities obtained by the analysis of surface wave dispersion (Fig. 10). The two body-wave tomographic models (Figs. 12 and 14) also show low velocities in southern Norway although with some differences compared to the surface-wave-based model. The P-wave tomography of Medhus et al. (2012) (Fig. 14) shows a velocity variation under Norway and Sweden in the 100–200 km depth range which is in good agreement with the integrated 3D model, both in the location of the maximum gradient and in amplitude. The S-wave tomography depicts a sharp, eastward dipping velocity contrast, extending from the base of the crust to 400 km depth (see Fig. 13). The location of the largest velocity gradient is located 200 km westward compared to the integrated 3D. Both tomographic models show low-velocity anomalies which continue to at least 400 km depth, a feature that cannot be reproduced with a uniform sub-lithospheric mantle as assumed in the present modeling. This however does not rule out
variations in density of the underlying asthenosphere contributing to the gravity signal. Such a sub-lithospheric gravity signal is generally of low amplitude compared to the lithospheric field and its uncertainties, and only affects the ultra-long wavelength of the gravity field which is filtered out from the present data. The comparison of the seismologically and numerically derived velocity models is only considered qualitatively here. A quantitative approach is not feasible with the unresolved differences in magnitude of P- and S-wave anomalies (see Section 3) as well as the simplification of vertically homogeneous SCLM composition employed in the numerical models.

The temperatures derived here correlate with the temperatures derived for the upper mantle across Europe by Goes et al. (2000). Their study is also based on the transformation of P- and S-wave seismic tomographic velocities into mantle temperatures. While most of the Baltic Shield shows low temperatures, with for example 1000 °C at 100 km depth and a lithospheric thickness of 200 km for southern Sweden, southern Norway stands out with high temperatures of 1200 to 1300 °C at 100 km depth, in good agreement with our results.

A step in the lithosphere implies large contrasts in viscosity and tectonothermal history of the adjacent domains. The region of the modeled boundary zone in the sub-lithospheric mantle largely coincides with the Permian Oslo Graben, with the main deformation zone during the Neoproterozoic Sveconorwegian orogeny, and with the edge of the Paleoproterozoic Svecofennian domain. While the different tectonic ages suggest independent evolution, their spatial correlation together with today’s lithospheric step suggests multiple reactivation events. This indicates that the presence of the lithospheric step observed today resulted, possibly indirectly, from structural heterogeneities inherited from earlier Phanerozoic and Proterozoic events.

5. Experimental modeling of Scandes mountain building

The presence of an anomalous mantle below southern Norway, as indicated by models derived from seismological data and obtained by integrated geophysical modeling, could be important for its deformation style. In order to test this, analog experiments were carried out to investigate the effect of different lithospheric structures on the style of lithospheric deformation which could be compared with present-day southern Norway (Agostini et al., submitted for publication). The hypothesized (abrupt) lateral variations in composition, temperature and thickness of the crust and lithosphere across southern Fennoscandia imply related variations in rheological strength. Abrupt contrasts in rheological strength and/or the presence of anomalously weak or strong zones can have a strong influence on the localization

![Fig. 16. Vertical profiles of Common-Conversion-Point stacks of P-receiver functions from Frassetto and Thybo (2010). Station location and topography are shown at the top. Insets show the locations of the profiles. The plotting conventions are the same as in Fig. 15 but the color scale saturates at ±0.05. The depth of 410 km is indicated with a black line. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)](image)
Fig. 17. Results of integrated 3D geophysical modeling. a) Overview of model domain. b) Observed and calculated elevation and Bouguer anomaly along a cross section at 61°N. c) Cross section at 61°N showing model geometry and densities. d) Cross section at 61°N showing seismic S-wave velocities. e) Comparison of modeled (red, blue) and observed (gray) velocity-depth plots. f) Cross section at 61°N showing temperature. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
and mode of deformation of lithosphere that is subjected to horizontal tectonic loading (i.e. Willingshofer et al., 2005).

4D lithosphere-scale analog experiments were carried out to investigate the role of a weak lower lithosphere for the localization of mountain building (Agostini et al., submitted for publication). The experimental models incorporate three distinct domains from west to east (Fig. 18): (1) an offshore continental shelf between the oceanic-continental lithosphere boundary (COB) and the shoreline; (2) a continental area about 300 km wide overlying variable thickness lithosphere; and (3) a cratonic area, about 200 km wide, which could be considered as analogs for the continental shelf, the area in which the southern Scandes mountains have formed and the Fennoscandian craton respectively. As the width of the continental shelf (from the COB to the coastal area) is not uniform in the study area, an average value of about 350–400 km is adopted. In each domain the experimental lithosphere consists of two or three different layers floating on a viscous fluid representing the asthenosphere. The top layer is composed of quartz sand as an analog of the brittle upper crust, whereas the deeper layers consist of mixtures of various silicon putties as analogs of the ductile lower crust and mantle lithosphere. The models are characterized by a complex geometry with strong lateral transitions in thickness and brittle–ductile strength variations (Fig. 18) designed as analogs for the northeastern Atlantic margin with and without a weak lower lithosphere beneath the Scandes. The analog models are scaled following the principles of geometric and dynamic-rheological similarity (Agostini et al., submitted for publication; Weijermars and Schmeling, 1986).

Experiments were performed in a transparent tank with one moving wall (acting as an analog for the oceanic lithosphere), which induces compression on the analog lithosphere (Fig. 18). The applied compression simulates the intracontinental horizontal stress arising from mid-Atlantic sea-floor spreading, which is considered as one possible driving mechanism responsible for the inversion structures characterizing the Norwegian Sea, as well as for the building of the Scandes (e.g. Anell et al., 2009; Fejerskov and Lindholm, 2000; Pascal and Gabrielsen, 2001; Vagnes et al., 1996).

In summary, the 4D analog experiments show that deformation in the relatively strong domains tends to be dominated by long-wavelength lithosphere folding, whereas deformation in the weaker domains is characterized by localized uplift and subsidence (e.g. Fig. 19) (Agostini et al., submitted for publication). The main conclusion from the analog modeling experiments is that only models with a weak upper mantle rheology successfully predict the development of a mountain range at the correct location of the present-day Scandinavian Mountain Chain (Fig. 19, right), and with comparable morphological characteristics such as steep slopes dipping towards the coastline and gentle slopes dipping towards the Fennoscandian inland areas (e.g. Lidmar-Bergström et al., 2000). Model results indicate that the largest part of the deformation at the boundary between shelf and Scandes is distributed in a narrow belt (in the order of 4–6 cm, scaled to natural values of ~80–120 km). However, the observed uplifted areas of the Scandes are 2–3 times wider. A possible explanation of this discrepancy is that unlike the actual lithosphere the model did not incorporate or inherit a long and complex deformation history. The structures developed during the different tectonic phases, which led to the present configuration, allowed wider distribution of deformation. In nature, deformation is not randomly distributed but avoids strong regions and concentrates along pre-existing weakness, reactivating and focusing on them. The high number of preexisting structures that could be reactivated may have resulted in the observed wide uplifted areas.

6. Implications for tectonic history and discussion

There are two main questions that arise in connection with the results presented here: what is the origin of the lithospheric and sublithospheric lower velocities below southern Norway? What is their relation to the present-day topography?

6.1. Mantle diapir intrusion in the lithosphere?

The presence of low seismic velocities below southern Norway has previously been used as an argument in favor of a mantle diapir model for the uplift of the Scandes (Rohrmann and van der Beek, 1996; Rohrmann et al., 2002). Following this model, uplift would be created by intrusion into the lithosphere of some anomalously warm asthenospheric material flowing from the Icelandic hotspot 30 to 20 My ago. Pascal and Olesen (2009) showed that the proposed diapir would have to be too shallow and too warm to be solely responsible for the gravity low observed in southern Norway. Gravity modeling (Ebbing et al., 2012; see below) taking into account the laterally varying crustal structure, shows however that the short-wavelength part of the gravity low is well explained by the crustal structure and that only the long-wavelength component of the gravity low needs to originate from the upper mantle. The gravity field is shown in Section 4 to be consistent with a composition and temperature anomaly below southern Norway. The gravity field alone cannot therefore be used as an argument against the diapir model.

The underlying assumption behind the diapir model is that the transit of the Icelandic hotspot from below Greenland to the west of the Atlantic Ocean, about 30 My ago, resulted in a light warm asthenospheric material flowing eastwards, reaching the base of the Scandinavian continental lithosphere and that a Rayleigh–Taylor instability allowed this buoyant material to rise into the lithosphere. The understanding of the consequences of impingement of hot mantle material below continental lithosphere has greatly improved in recent years due to better modeling of the complex rheology of the continental lithosphere (Burov and Guillaud-Frottier, 2005; Burov et al., 2007). It has been shown that the hot material from a plume head can create Rayleigh–Taylor instabilities if the lithosphere is warm (equivalent to a thermal age of 100 My) and the viscosity contrast between lithosphere and asthenosphere is not too large. In the case of lower asthenospheric viscosity, the plume head erodes the lithosphere but is not able to destabilize it, and in the case of a colder lithosphere (equivalent to 400 My age or more), the asthenospheric material ponds at the bottom but does not erode the lithosphere significantly. In addition, the consequence of such Rayleigh–Taylor instabilities in terms of topography is more complex than the dome shape observed in southern Norway. Jurine et al. (2005) have emphasized the role of the buoyancy ratio between the lithosphere and the plume material in the behavior of the plume and also have shown that lithospheric penetration is unlikely for Archean and Proterozoic lithospheres unless an exceptionally strong mantle plume is involved. It seems therefore unlikely that some warm asthenospheric material from Iceland could have destabilized an initially cold lithosphere below southern Norway and caused uplift, and that the low velocities observed today in the lithosphere are a result of this process.

6.2. Mantle diapir below the lithosphere?

However, even if intrusion of the lithosphere by a mantle diapir does not seem likely, it does not exclude the possibility that the low velocities observed below the lithosphere and down to 400 km depth could be related to a general large area of low velocities below the northern Atlantic and that this area is related to the Iceland plume. In the tomographic model of Amaru (2007), the region of low seismic velocity located below Iceland has a branch towards the Baltic craton in the depth range 100 to 400 km that corresponds well with the anomaly that we image in the body-wave tomography. The P-tomography models indicate that also a deep connection to the North Sea region to the southwest may exist (Fig. 14). A similar asthenospheric feature has been proposed to explain the nearby combination of moderate
Fig. 18. (upper) Cartoon showing the initial set-up for an analog experiment investigating lateral variations in crustal and lithosphere thickness, composition and rheology across southern Fennoscandia. UC = brittle upper crustal layer; LC = ductile lower crustal layer; UM = upper mantle layer. The black arrow indicates the direction of compression. Bottom panel shows rheological strength profiles at locations a, b and c. (lower) Same analog experiment, however, with a weaker upper mantle layer (green mix 2) for the central Scandes domain. (after Agostini et al., submitted for publication).
Fig. 19. Experimental results of (upper panels) model with a homogeneous upper mantle composition across southern Fennoscandia, and (lower panels) model with an anomalous weak upper mantle for southern Norway. (a) Top views of the model after 1% and 4% of bulk shortening. The black arrows indicate the direction of compression. (b) Shaded relief pictures of the surface topography after 1% and 4% of bulk shortening. (c) Topographic evolution profiles taken from the center of the model parallel to the compression direction after different amounts of bulk shortening. (d) Cross section of the model (trace is the dotted black line in panel a).
topography and crustal structure in the northern UK (Davis et al., 2012), although the connection to the Iceland plume has less support in this area than in southern Norway according to regional tomographies (e.g. Pilidou et al., 2005).

It is however unlikely that a potential sublithospheric diapir has created the present-day topography in the Scandes. According to Burov and Guillou-Frottier (2005), sublithospheric material below an old lithosphere (from 400 Ma) produces surface topography of larger wavelengths (1500 km in their model) and smaller amplitude (400 m) than observed in southern Norway.

In addition, the lithospheric structure is likely to be close to iso-static equilibrium, reducing the need for a deep mantle component in sustaining the topography. Ebbing (2007) and Ebbing et al. (2012) showed that, although Moho topography alone cannot isostatically explain the topography of the Scandes, a more complex compensated Airy–Pratt model that takes into account the presence of a laterally varying high-density lower crust, varying lithospheric thickness and low-density crustal material in the Transscandinavian Igneous Belt (TIB) can together explain the gravity field very well. The additional constraints on Moho depth and crustal structure that have been achieved in the present study, and that are largely already accounted for in Ebbing et al. (2012), do not contradict these conclusions. The more detailed knowledge of crustal structure that we have achieved in this and other recent studies enables us to state that the area is likely to be close to isostatic equilibrium, a result suggested earlier based on the anti-correlation between regional topography and Bouguer gravity anomaly (e.g. Balling, 1980).

6.3. Relation between metasomatism and the lateral variation in lithospheric structure

The gradual thickening of the lithosphere from Norway to Sweden in the model of Ebbing et al. (2012), which explains the long wavelengths of the gravity data, is taken from the early studies of Rayleigh wave dispersion along a few paths in Fennoscandia (Calcagnile, 1982). This lateral variation has been confirmed by the present study and modeled with an improved horizontal and vertical resolution (Figs. 12, 13 and 14). Surface wave and gravity data have been shown to be consistent with a relatively sharp increase of 100 km in lithospheric thickness close to the eastern border of the Oslo Graben associated with a contrast in elemental depletion (Section 4, Gradmann et al., submitted for publication). A change from fertile (zero-depletion) mantle below southern Norway to Proterozoic mantle below southern Sweden (0.6–0.9% depletion) is also required by buoyancy analysis (Artremieva, 2007).

The geometry and location of the contrast proposed here fit rather well with the results of the P-wave tomography but overpredict the amplitude of the contrasts in the S-wave tomographic model by about a factor of 2, which we can consider as satisfactory considering the limitations of tomographic models and lithospheric modeling.

A contrast of 4% in S-wave velocity between southern Sweden and southern Norway may be explained by a temperature difference of the order of 400 °C in the 100 to 200 km depth range. The S-wave tomography implies a contrast of 200 °C over a lateral transition zone of 100 to 200 km. It is commonly argued that large temperature contrasts over small distances cannot survive in the lithosphere over long periods due to the equilibration effect of thermal conduction and that compositional variations or the presence of melt is necessary to explain abrupt seismic velocity variation in old lithospheres. The direct effect of compositional variations on seismic velocities has, however, been shown to be able to produce variations of about 1% only (e.g. Goes et al., 2004) and temperature is in many cases far from the solidus, making it unlikely that partial melt will be present and significantly affect seismic velocities. Regardless of other possible effects, temperature variation is therefore expected to play a significant role in producing the seismic velocity contrasts that we observe. Compositional variations may however have an indirect influence on seismic velocities by affecting either viscosity contrasts (Hieronymus et al., 2007) or concentration in radioactive elements (Hieronymus and Goes, 2010), that have a direct influence on the resulting temperature.

Several mechanisms can produce increased mantle temperatures. Whereas ponding of hot mantle material (such as produced by mantle plume, convection instability or delamination) would increase mantle heat flow supplied to the lithosphere, infiltration of magmas into the lithosphere may produce its reheating from inside by the supply of material enriched in radioactive elements. (Hieronymus and Goes, 2010). The timescales of the two processes are fundamentally different, the reheating by increased heat flow being transient while reheating by radioactive sources being basically steady-state due to the long half-life of the radioactive elements involved. Hieronymus and Goes (2010) showed that different degrees of lithospheric depletion are associated with mild variations in the concentration of radioactive elements that suffice to maintain steady-state temperature differences of 100 °C to 300 °C at mid-depth in the lithosphere. Their model can explain an S-wave velocity contrast of almost 2% over a distance of 200 km by juxtaposing a depleted Archean lithosphere with a less depleted Proterozoic lithosphere. They infer absolute S-wave velocities that fit well with those observed in Archean cratons and Proterozoic fold belts in the depth range 100 to 200 km (see Fig. 10).

In our case, the tectonic history and the results presented in Section 4 show that a juxtaposition of a Proterozoic-type and a Phanerozoic-type lithosphere would probably be more appropriate. Considering that this implies an overall lower degree of depletion, we anticipate an overall reduction in the absolute velocities but a similar amplitude of lateral variation. The velocities observed here may therefore partially be the result of different degrees of depletion and the sharpness of the contrast does not require that the Norwegian lithosphere has been locally replaced or heated recently. Hieronymus and Goes (2010) show that the surface heat-flow resulting from their model is hardly affected by the deep temperature variations due to the insulating effect of the surface layers. An interpretation of the upper mantle velocities in terms of temperature differences does not therefore contradict the rather uniform and moderate surface heat-flow in southern Norway (Slagstad et al., 2009).

The formation of the Oslo Graben within the Baltic Shield has been proposed to be associated with a localization of the deformation during Permian extension by the presence of a step in lithospheric thickness of at least 45 km just east of the Graben (Pascal and Cloetingh, 2002). Albeit smaller, this step is remarkably similar to the one we see in tomographic models today, suggesting that the step is older than 300 Ma. Enrichment in radioactive elements in southern Norway could be native or related to one or several subsequent geological events. One candidate is local metasomatism in the lower lithosphere related to the magmatic activity in the area at about 580 Ma. Metasomatism has been proposed to explain the geochemical and petrological data of the Oslo Graben magmatic rocks (Neumann et al., 1992) that have been shown to derive from two mantle sources with different degrees of depletion. Although the metasomatism inferred by Neumann et al. (1992) was recorded in the area of the Oslo Graben, it might have affected a larger area but be visible at the surface only in areas where volcanism brought some material close to the surface. On the other hand, geochemical analysis of mantle-derived rocks indicates that the lithosphere is depleted in the whole Baltic Shield, including southern Norway (Andersen and Sundvoll, 1995). Very high degrees of depletion have also been inferred in the Western Gneiss Region (Beyer et al., 2006), where peridotites also bear evidence for the several tectono-magmatic events that have formed the lithosphere and its complex history (Beyer et al., 2006; Lappen et al., 2009). There is therefore no unequivocal geological evidence for an enrichment of the lithosphere below southern Norway that could contribute to increase its temperature.
6.4. Metacratonization of the lithosphere

Another candidate for the rejuvenation of the lithosphere is the formation of the Caledonides and the associated subduction of the western margin of the Baltic craton, presently southern Norway. It has been proposed that the margins of cratons involved in intercontinental collisions may partly subduct but still preserve major cratonic characteristics, and that southern Norway is a good candidate for such a metacratonic area (Liégeois et al., 2012). The consequence of “metacratonization” is reduced seismic velocities at the bottom of the lithosphere, as shown in the Saharan Craton where reduced velocities occur below 100 km depth (Adbelsalam et al., 2011). These authors propose several possible mechanisms at the origin of this velocity modification, in particular lithospheric delamination and convective removal.

6.5. Comparison with other intracontinental mountain chains

Several other regions worldwide show high topography away from a recent plate boundary. We have already mentioned Greenland, located on the margin conjugate to Scandinavia. The Urals Mountains are also a recent (0–5 My) topographic feature (Puchkov and Danukalova, 2009) situated at the location of an ancient Hercynian orogenic structure, and away from any present-day plate boundary. Their well-preserved crustal root (Carbonell et al., 1996) distinguishes them from the Scandes and could be a remnant of the Hercynian structure (Leech, 2001), but the present-day topography is inferred to result from intraplate deformation (Puchkov and Danukalova, 2009). The Transantarctic Mountains have a seismic structure more similar to that of southern Norway. They are located at the outer edge of the Eastern Antarctic craton, next to the younger tectonic area of the Ross Sea. They display a well-developed topography of 3500 m associated with a small crustal root of about 5 km displaced towards the cratonic area compared to the maximum topography, and low seismic velocities in the lithospheric mantle. The topography, together with the seismic structure and the gravity field, can be explained by buoyancy loads, in particular mantle buoyancy, deflecting an 80 km thick elastic plate broken at the edge of the craton (Lawrence et al., 2006). The model for the formation of the Transantarctic Mountains may however not apply to the southern Scandes, due to the smaller elastic thickness in Scandinavia and the anticorrelation between topography and Bouguer anomaly, which is not observed in Antarctica, suggesting a lesser degree of compensation in Antarctica. A recent receiver function study has shown that another intraplate mountain range in Antarctica, the Gamburtsev mountains, is underlain by a ca. 10 km thick crustal keel (Hansen et al., 2010), which may indicate that crustal isostasy maintains the high topography. The Colorado Plateau also has similarities with southern Norway: high elevation (1800 m) in intraplate setting of a Proterozoic lithosphere. The Plateau is also associated with contrasts in seismic velocities of up to 12%, but the low velocities are located at the edge of the high topography and not directly below, and are associated with Neogene magmatism at the surface. Edge convection that erodes the lithosphere of the Colorado Plateau is able to explain the observations well (van Wijk et al., 2010).

Although there are several intracontinental high elevation areas with some features similar to the ones of the southern Scandes, it turns out that each mountain chain has its own characteristics, making it unique and leading to different preferred mechanisms for their tectonic evolutions. This diversity probably reflects the diversity of the evolution of the continental lithosphere and the need to account for tectonic inheritance and a combination of different mechanisms to explain the presence of high topography in intracontinental settings.

7. Conclusion

Several high-resolution independent seismological models indicate the presence of low seismic wave velocities in the mantle below southern Norway. P- and S-wave body tomographies as well as a small depression of the 410 discontinuity suggest that the anomaly extends down to the transition zone. The anomaly is located approximately in-between the coast in the west and the Oslo Graben in the east, underlying the highest topography of the southern Scandes. A major issue in interpreting the low velocities is the relation between the lithospheric part of the velocity anomaly and the sublithospheric part. Establishing a connection between the sublithospheric anomaly and the Icelandic plume and comparing to the lithospheric and sublithospheric structure in the northern part of the Scandes could shed light on this issue. This would require extension of the tomographic and geophysical studies to a larger regional scale. New high-resolution crustal models suggest a crustal root that is not located right below the region of high topography. The seismic models combined with the gravity data show that crustal isostasy is not sufficient to explain the mountain chain but that it contributes, together with lateral variations in the lithospheric thermo-chemical structure, to a system that is likely to be close to isostatic equilibrium. As such, dynamic support from the mantle is not needed to explain the presence of the topography today. However, the presence of low velocities in the asthenosphere points to the possibility for a minor dynamic mantle contribution.

Independently of the process that led to the formation of the topography, it is clear that the geological history of the area has led to major lateral variations in crustal and mantle structures that have to be taken into account when searching for the mechanisms that created the topography.

Acknowledgments

This work has been done in the framework of the ESF EUROCORES TOPO-EUROPE Programme 07-TOPO-EUROPE-FP-014: The Scandinavian Mountain Chain — deep processes (TopoScandiaDeep; www.geo.uio.no/toposcanldia). We acknowledge financial support from the Research Council of Norway, the Danish National Science Research Council, Deutsche Forschungsgemeinschaft and Statoil. MAGNUS waveforms were recorded with the mobile KArlsruhe BroadBand Array (KABBA) of the Karlsruhe Institute of Technology, Germany as well as with permanent stations of the NORSAR array and the Norwegian National Seismological Network. Financial support for the MAGNUS experiment was provided by the Universities of Aarhus, Copenhagen, Karlsruhe and Oslo as well as NORSAR. Acquisition of the SCANLIPS data was supported by the Geological Survey of Norway and equipment loans from the UK Natural Environment Research Council’s Geophysical Equipment Facility. We thank two anonymous reviewers for detailed and very constructive reviews.

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