Project Tor: Deep lithospheric variation across the Sorgenfrei-Tornquist Zone, Southern Scandinavia

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The Tor project makes use of teleseismic tomography across the Sorgenfrei-Tornquist Zone and has now revealed significant variations in the deep lithosphere under northern Germany, Denmark and southern Sweden. Here we present the first interpretations of P-wave traveltime anomalies from the Tor project. The project utilised 120 seismographs placed in a rectangular array, the largest seismic antenna so far used in Europe, for half a year in the period 1996–1997.

The present investigation establishes a 3D crustal/upper mantle model of the P-wave velocity based on existing data. A picture of the crustal influence on the seismic P-wave rays is established by ray tracing through the model. When this is subtracted from that observed by the Tor array, a picture of the influence of the lower lithosphere/asthenosphere system emerges. For several earthquakes it is shown that the observed P-wave traveltime anomalies of nearly 2 seconds can be divided almost equally between known crustal effects and lower lithosphere/asthenosphere differences. The transition appears gradual from most directions but for rays coming from the north-east direction the transition appears sharper. This means that the broad scale deep lithosphere transition is gradual with the sharpest discontinuity plane dipping down steeply in a north-easterly direction from the Sorgenfrei-Tornquist Zone. Based on existing knowledge of the area we conclude that the transition from thin to thick lithosphere occurs within a short distance, and that the lithosphere/asthenosphere boundary dips steeply down from the surface expression of the Sorgenfrei-Tornquist Zone.

Key words: Tor project, deep lithospheric variation, lithosphere/asthenosphere boundary, Sorgenfrei-Tornquist Zone, 3D crustal model.

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The Tornquist Zone which trends NW-SE from Denmark to Romania has traditionally been regarded as the suture between (1) the Precambrian shield of Fennoscandia and Eastern Europe, and (2) the Paleozoic regions of Central Europe. Age determination of bedrock material from deep boreholes (Frost et al., 1981) and crustal studies starting with EUGENO-S (EUGENO-S Working Group,1988), and continuing with BABEL (BABEL Working Group, 1993) and MONA LISA (MONA LISA Working Group, 1997) have pointed to other important lineaments in the crustal structure. It is now customary to distinguish the northern part of the Tornquist Zone in Sweden and Denmark as the Sorgenfrei-Tornquist Zone and the southern part between Poland and Romania as the Teisseyre Tornquist Zone. Within the large geoscientific program Europrobe (Gee and Zeyen, 1996) the Tornquist Zone is still considered to be an important component of what is now called the Trans-European Suture Zone. Large seismic velocity contrasts have been recognized in the area of the Tornquist Zone (e.g. Guggisberg & Berthelsen, 1987; Blundell et al., 1992; Spakman et al., 1993 ; Zielhuis & Nolet, 1994). It was expected that detailed geophysical investigations could delineate this important lithosphere boundary at depth more precisely. A teleseismic tomography project across the Tornquist Zone, the Tor project, was set up to do this with a resolution of 20–30 km as a

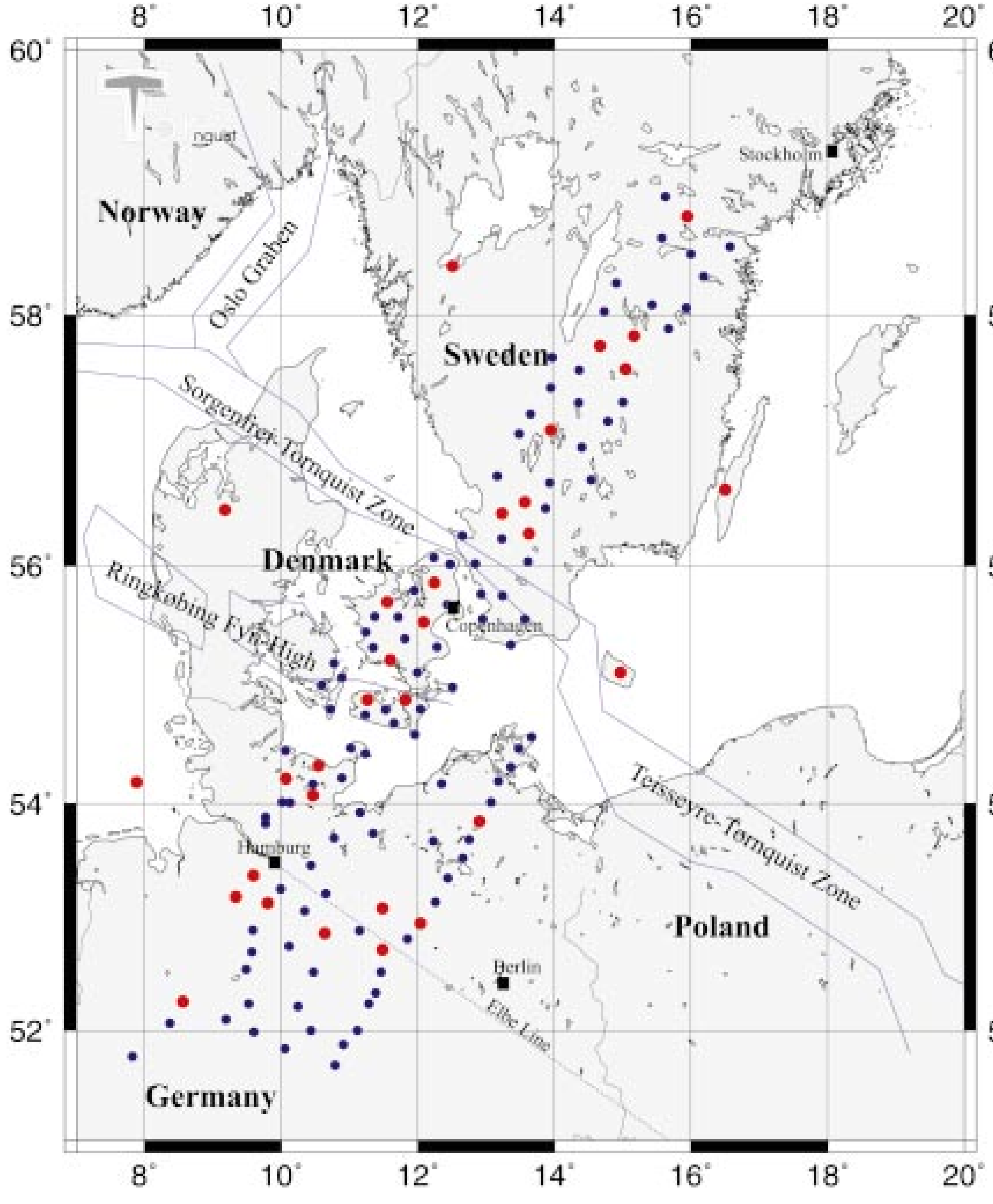

Fig. 1. The location of seismometers during the fieldwork of the Tor project 1996-1997. The blue dots indicate short-period seismometers and the red dots indicate broadband seismometers. The geologically important structural lines are also shown (EUGENO-S Working Group, 1988).

result of dense seismograph spacing (Gregersen, 1995, 1997; Gregersen et al., 1997; Voss, 1998). This is an improvement compared to previous studies with a resolution of more than 100 km (Spakman et al., 1993).

The area around the NW-SE trending Tornquist Zone has been tectonically active for the last 440 million years. The zone has been tectonically reactivated many times through complicated motions including shearing and extension (e.g. Berthelsen, 1998). Tectonic inversion occurred in Cretaceous-Tertiary time, when the present areal extent of the Tornquist Zone was established (e.g. EUGENO-S Working Group, 1988). In Sweden, Denmark and Germany the exact location and form of the lithospheric transition is unclear. Several lineaments besides the Tornquist Zone are candidates (Fig. 1). The transition could, on the other hand, be gradual, with no preferred sharp boundary. The aim of the Tor project is to delineate the plate boundary in the deep part of the lithosphere. This paper summarizes the field work and presents the pattern of the P-wave arrival times, taking into account the shallow geological structure determined in previous studies. Finally, the P-wave traveltime anomaly pattern and the consequences for the structure of the deep lithosphere are discussed.

Data Acquisition

Two belts perpendicular to the NW-SE trending Tornquist Zone have been chosen for investigation. The studies are part of the large-scale project on the Trans-European Suture Zone (TESZ) within the geoscientific program Europrobe of the International Lithosphere Program. The northern belt, Tor-1, across the Sorgenfrei-Tornquist Zone, is the one currently under study. A similar belt across the Teisseyre-Tornquist Zone, Tor-2, will be investigated in the future. A pilot study in the northern belt was carried out, in the winter of 1994–1995 to evaluate if the signal to noise ratio was sufficiently large for a full-scale investigation to be worth while (Kind et al., 1997). The pilot project also resulted in a supplementary indication of the sharpness of the crustal contrast across the Sorgenfrei-Tornquist Zone near the west coast of Sweden, in agreement with previous studies (e.g. EUGENO-S Working Group, 1988).

The Tor-1 seismic antenna was placed along a very well studied area of the European Geotraverse, where the sedimentary and crustal structures are well known (EUGENO-S Working Group, 1988; Thybo, 1990). A belt of seismographs, 100 km wide and 900 km long, was placed around profile 1 of the EUGENO-S study (Fig. 1). Approximately 100 short period seismographs were in the field for 6 months (October 1996 – April 1997) to be used for teleseismic traveltime tomography. By a stroke of luck a local earthquake was also recorded (Schmidt, 1998). A total of 31 broad band seismographs were placed mainly in the belt of the short period seismographs to supplement the geographical coverage and also to provide data for studies of receiver functions (Gossler et al., in press), anisotropy (Wylegala et al., in press) and eventually surface waves. These broad band seismographs were in their Tor positions for approximately a year (July–August 1996 to July–August 1997). The field locations were typically in outhouses or basements at the end of small roads, far away from traffic and other potential man-made disturbances. The geographical distribution of the seismographs was such that the broadband instruments were placed in triangles with side lengths of 40-60 km, a distance chosen to be comparable to wavelengths of the surface waves of interest. Each triangle was placed within one geological unit (Fig. 1) in order to be able to distinguish geological differences directly, one geological unit being an area with approximately the same crustal structure. The short period seismographs supplemented the coverage to give a horizontal resolution of 20–30 km.

Table1. The a priori models used to construct the 3D crust model

Profile	Name	References
E1	EUGENO-S profile 1	(Thybo, 1990)
E2	EUGENO-S profile 2	(Thybo & Schönharting, 1991)
E3	EUGENO-S profile 3	(EUGENO-S Working Group, 1988)
E4	EUGENO-S profile 4	(EUGENO-S Working Group, 1988)
E5	EUGENO-S profile 5	(Thybo et al., 1989)
E6	EUGENO-S profile 6	(Green et al., 1988)
FENNOLORA	FENNOLORA profile	(Stangl, 1990)
BABEL A	BABEL line A	(BABEL Working Group, 1993)
BABEL B	BABEL line B	(BABEL Working Group, 1993)
EUGEMI	EUGEMI	(Aichroth et al., 1992)
ZIPE	ZIPE	(Rabbel et al., 1995)
LT7	LT7	(Thybo & Zelt, in press)

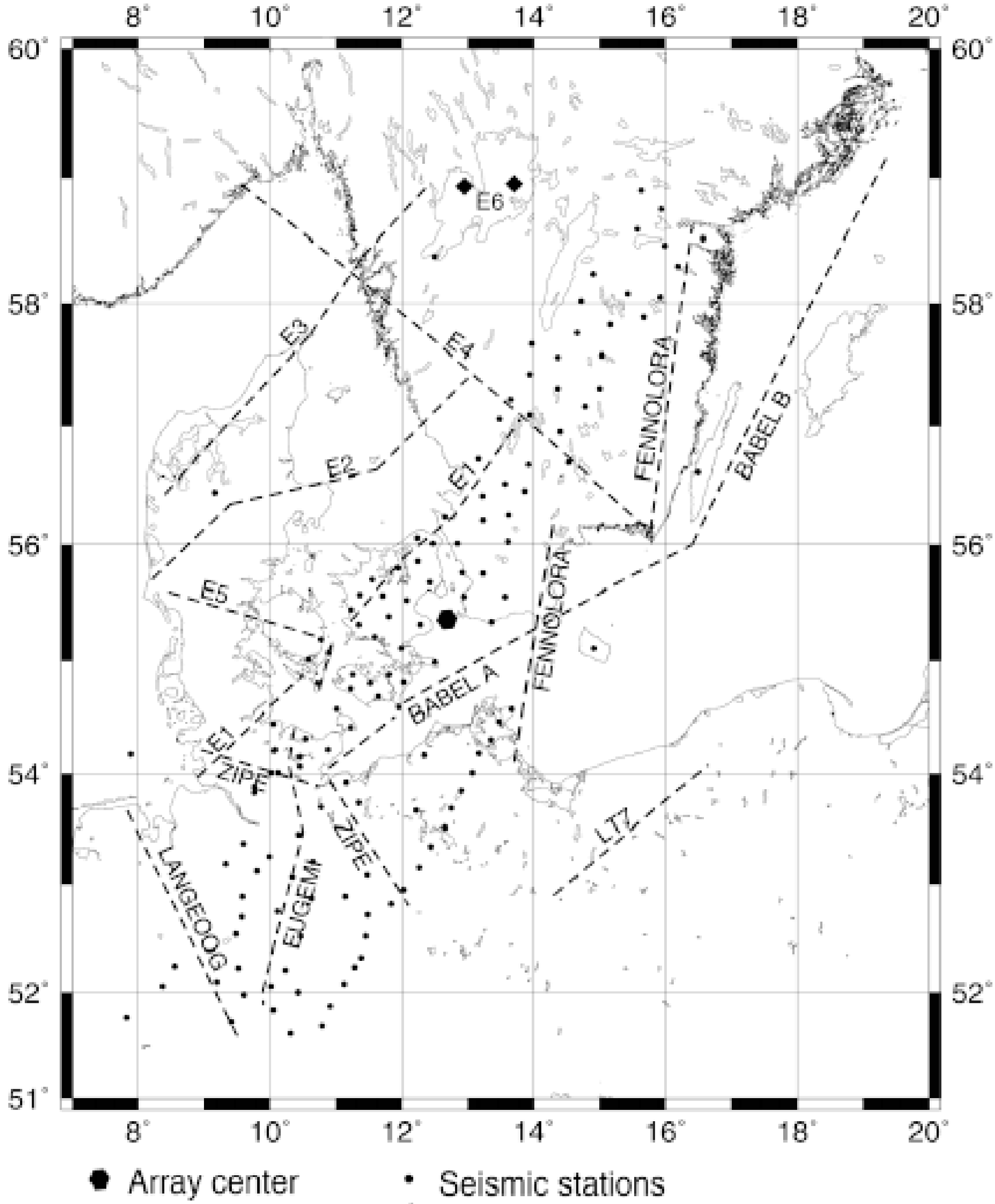

Fig. 2. The location of the crustal profiles, which the model is based on. See also Table 1.

The 3D crustal model

In the planning of the Tor project it was realised that waves from teleseismic earthquakes i.e. angular distance larger than 35°, would traverse the shallow parts of the lithosphere, the sediments and crystalline crust, almost vertically. This gives very poor vertical resolution. An important issue in the first phase of the project was therefore to establish the best possible model for the crustal part of the lithosphere based on other geological and geophysical data.

Along the well-studied crustal profile 1 of the EUGENO-S project two versions of the 3D crustal model have emerged, with different arguments for the choice of interpolation method and use of a priori information. The one presented here was derived by Pedersen (1999) based on the geophysical and geological interpretations by individual scientists presenting their 2D models, and carefully considering the choice of interpolation between 2D P-wave velocity models.

The Pedersen 3D crustal model, presented here, has been constructed to cover the volume 600 km (EW) × 860 km (NS) around a centre at 55.4°N 12.6°E to a depth 50 km. The uppermost mantle is therefore included in the model. Several wideangle reflection/refraction investigations have been carried out in southern Sweden, northern Germany and Denmark (Table 1 & Fig. 2). Data from these projects have been interpreted by individual geoscientists, giving 2D and 1D P-wave velocity models. All these models have been used to construct the 3D crustal model (Table 1). The lines in Figure 2 indicate the locations of the 2D models and the two diamonds indicate the locations of the two 1D models. These models are referred to as *a priori models* containing a priori information about the P-wave velocity.

The a priori models were digitised by reading the velocity in points spaced 5 km vertically and 20–30 km horizontally. Smaller horizontal spacings were chosen, when there was a step in the velocity. In this case the velocity is read at both sides of the step. In order to represent the sedimentary structures, the near-surface information has been supplemented by an extra level of digitised points at 2.5 km depth. These sets of points form the basis for the 3D crustal model. The 3D model has been parametrised by a 3 dimensional grid of nodes. The nodes are equally spaced horizontally (50 km × 50 km). The vertical grid space is 5 km plus an extra horizontal grid plane at 2.5 km depth. These depths of the horizontal grid planes match the depths where the a priori information is read. Each grid plane has been treated separately: The scattered a priori information of a plane was first grouped in triangles by the use of Delaunay triangulation (Fortune, S.J. 1987). The web page by Chew (1997) offers an excellent, and interactive visualization of the concepts behind Delaunay triangulation. The corners of a triangle are three digitised points. Interpolation and extrapolation were then used to calculate the velocity in the nodes. Various interpolation and extrapolation schemes have been tested (Pedersen, 1999) and the preferred methods were linear interpolation and the "nearest neighbour" scheme for extrapolation. This chosen procedure involves direct acceptance of the 2D a priori models and simple interpolation and extrapolation where no a priori information was available.

- Linear interpolation: A plane surface is fitted to the three points of a triangle. By drawing a line from each point to a node within the triangle, three subtriangles are constructed. The node gets a value, which is the weighted sum of the point values. The weighting of each point is proportional to the area of subtriangle opposite the point (Watson & Philip, 1984).
- "Nearest neighbour" is a very simple scheme: A node is given the same value as the nearest point. This scheme can also be used for interpolation, but here it is only used for extrapolation.

The 3D crustal model is presented as planes in Figure 3 and shows how the velocity increases with depth. The high velocities at depth 30 km below the northern part of Germany shows that it is below Moho. Beneath Denmark Moho is reached between 30 km and 35 km, and beneath Sweden between 40 km and 45 km depth. The first three planes, the surface-plane 0 km and the planes at 2.5 km and 5 km are shown in more detail. The border between the low velocity sediments and bedrock in Sweden is very clear at the surface-plane (Fig. 3b) . The Ringkøbing-Fyn high is seen as the two yellow high velocity zones in the blue low velocity area, which represents the surrounding basins in the plot of the plane at depth 2.5 km (Fig. 3c). The Ringkøbing-Fyn high gets wider at depth 5 km (Fig. 3d).

The other version of a 3D model was presented by Arlitt et al. (in press), following ideas of Waldhauser et al. (1998). It includes a 3D model of the Moho-interface based on well determined Moho-reflections, and interpolation that replaces the interpreted Moho depths made by geophysicists/geologists. This is in contrast to the method used for the model presented above, that accepts the geophysical and geological interpretations. The sediments are included in the model by Arlitt et al. (in press) as homogeneous polygons, which results in sharp velocity discontinuities along the edges of the polygons. These discontinuities are visible in the display of residuals, calculated using wavefront calculations through the model (Arlitt et al., in press; Gregersen et al., 1999). This shows the choice of sediment representation as very unfavourable.

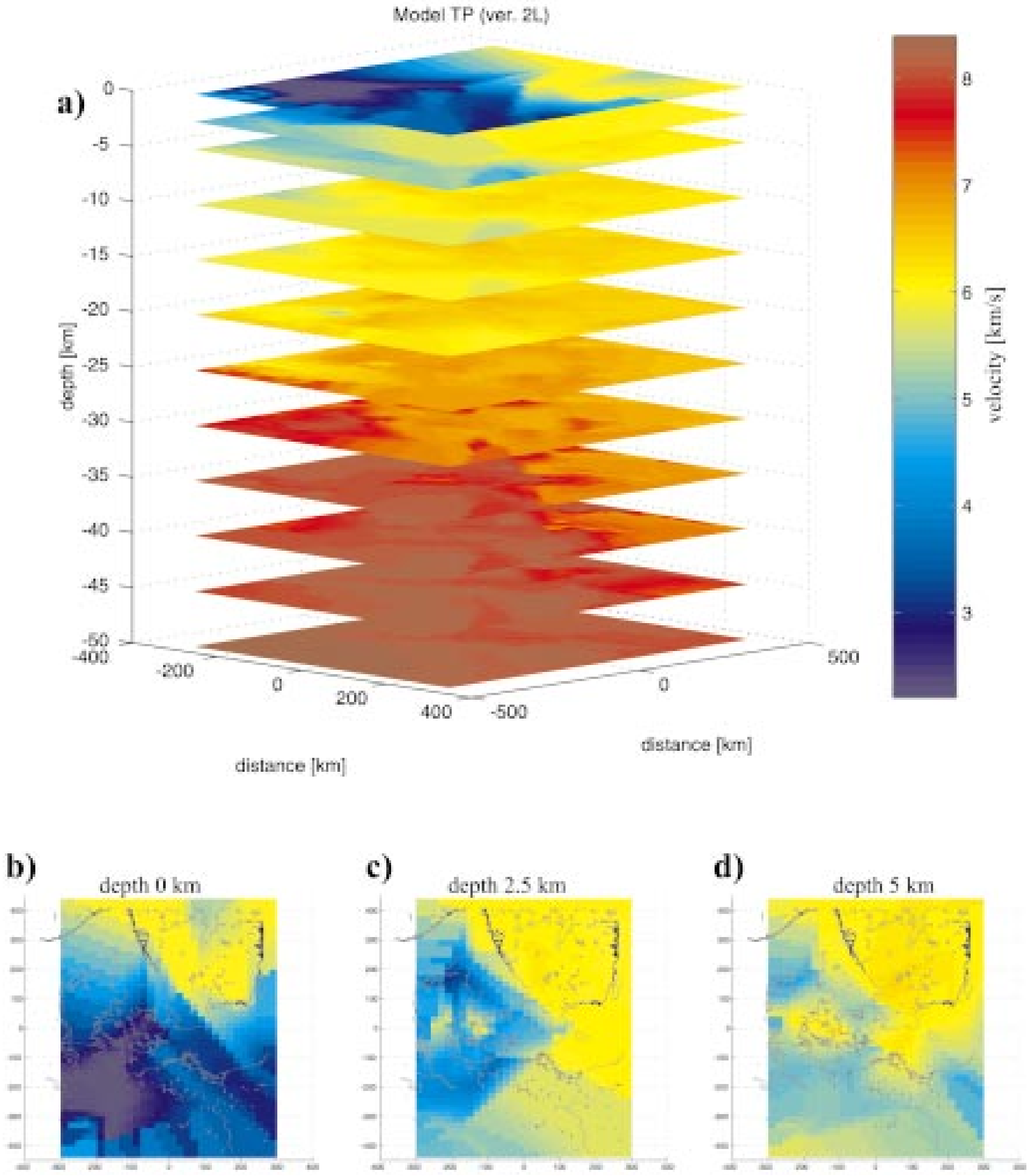

Fig. 3 a. Model of P-wave velocity in the top 50 km (TP-2L). Each plane is made by linear interpolation and nearest neighbour extrapolation. b, c & d are separate displays of the upper 3 planes. All axes are in [km] relative to the array centre at sea level.

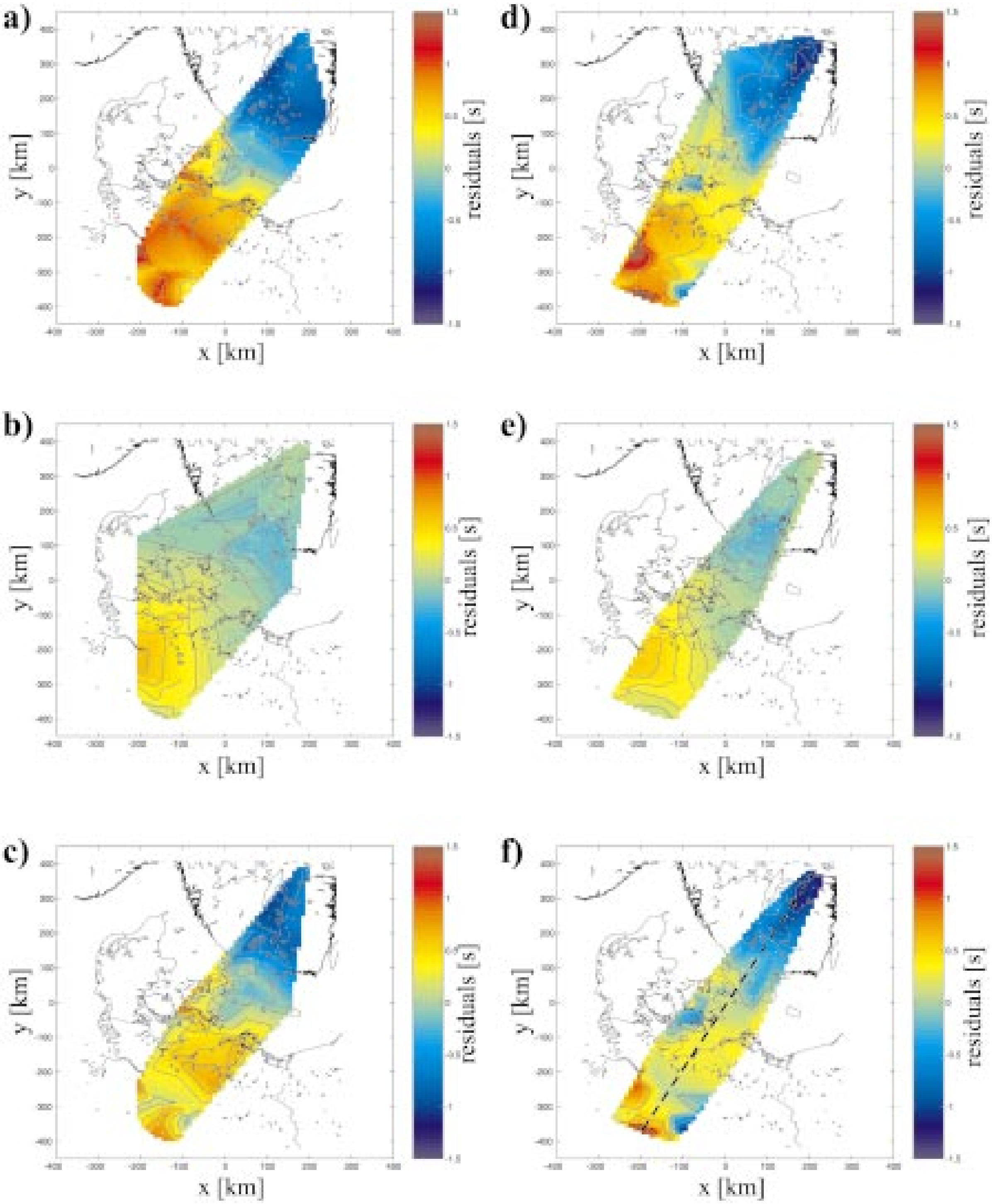

Fig. 4. Residuals from two earthquakes. a, b & c. Residuals of an event arriving along the Tor array from north-east (event No. 2 in Table 2). d, e & f. Residuals of an event arriving perpendicular to the Tor array from north-west (event No. 4 in Table 2). a & d. Observed residuals. b & e. Model residuals for the upper 50 km. c & f. Stripped residuals (Observed minus model residuals) caused by structure deeper than 50 km.

Table 2. List of picked earthquakes, where AZ is the azimuth angle from the central point of the array. m_b is the magnitude determined from P-wave amplitudes.

No.	origin	Latitude	longitude	depth	m_b	AZ
1	1996–10–18 10:50:22.79	30.594° N	131.034E	10.8 km	6.0	50°
2	1996–10–19 14:44:43.27	31.909° N	131.516E	28.9 km	6.3	49°
3	1996–12–10 08:36:19.42	0.958° N	29.945W	9.2 km	6.0	228°
4	1997–01–11 20:28:29.67	18.262° N	102.700W	49.3 km	6.5	300°
5	1997–02–27 21:30:37.13	29.953° N	67.969E	24.5 km	6.0	99°
6	1997–02–28 11:32:21.08	43.993° N	147.806E	6.1 km	6.1	32°

Traveltime residuals

To explain the results of our work it is necessary to introduce three expressions for different kinds of traveltime residuals: The *observed traveltime* residuals were produced by picking the P-wave arrival times at all the stations and subtracting the iasp91 traveltime. The iasp91 traveltime tables are the standard traveltime tables used for global earthquake location (Kennett & Engdahl, 1991). These observed residuals are a sum of residuals caused by inhomogeneities in the crust and upper mantle. The crustal effect is computed by ray tracing through the 3D model presented in the previous section (Pedersen, 1999). The ray tracer calculates the traveltimes of the rays inside the crustal model, from the bottom of the model to the receiver (Steck & Prothero, 1991). The traveltime was calculated for each earthquake, and each station. We refer to the computed residuals as *model residuals*. By subtracting the model residuals from the observed residuals, we find the residuals caused by the structure beneath the model. These resulting residuals are stripped for the crustal effect and are referred to as *stripped residuals*.

These three different residuals were calculated for six earthquakes with waves travelling across the Tor array from different azimuths (Table 2). Figure 4 illustrates the three types of residuals for two of the six events. The wavefront of the event in the left column travels from azimuth 49°, i.e. along the array of seismometers from NE. The wavefront of the other event travels from azimuth 300°, i.e. perpendicular to the array from NW. The white dots indicate where residuals are respectively observed and calculated. For visualisation the residuals are interpolated to a regular grid using linear interpolation. Therefore the areal coverage varies according to the availability of data.

Observed residuals (Fig. 4a&d) show that the P-wave arrives early in Sweden and late in Germany for both events with a difference of approximately 2 seconds between the overall minimum and maximum. Small local variations on the plots can be attributed to local inhomogenities in the crust. On both plots there is evidence of a high velocity zone near (x,y)=(–100, –370), which is the Harzen Block (Berthelsen, 1992). Evidence of two low velocity zones in Germany is also observed on both plots. Only the zone near (x,y)=(–200,–250) can be related to a known low-velocity zone. It is a low-velocity zone located in the middle crust, which has been found in many seismic refraction profiles as a typical property of the continental crust in central and southern Europe (Aichroth et al., 1992).

Model residuals for the two events show a similar pattern of early arrivals in Sweden and late ones in Germany (Fig 4b&e). Here a delay of 0.8 second is observed between the overall minimum and maximum.

Observed residuals minus model residuals – the stripped residuals (Fig. 4c&f) are caused by differences in the lower lithosphere and asthenosphere. Here we disregard local variations, which were not included in the model, due to lack of a priori information. The average residual variation along the array of seismometers is about 1.2 seconds.

From previous investigations it is expected that the lithosphere/asthenosphere boundary dips to the NE below the shield (Guggisberg & Berthelsen, 1987, Blundell et al., 1992; Ansorge et al., 1992; Calcagnile et al., 1990; Spakman et al., 1993; Pedersen et al., 1995). Also the Tor-measurements of surface waves (H.A.Pedersen, personal communication) show large difference in lithosphere thickness across the Trans-European Suture Zone, i.e. across the Danish area of the Tor belt. The major questions within the Tor project are, where does the transition occur, and how steep is it?

To get a better overview of the variation of the stripped residuals along the Tor array, all the data have been projected onto one profile representative of the array belt (the dashed line in Fig. 4f). The resulting plots for the six investigated earthquakes are shown in Figure 5. Here the vertical axis is the stripped residual at each station, caused by the deep structure. The difference between the level of the residuals in the shield and in the basin areas is obvious in all of the six plots. There is a difference in level of slightly more than 1 second. And the change from one level to the other does not occur at one single line, the change is somewhat gradual. The sharpest part of the transition is concentrated below the Sorgenfrei-Tornquist Zone or in close connection to it in several of the plots.

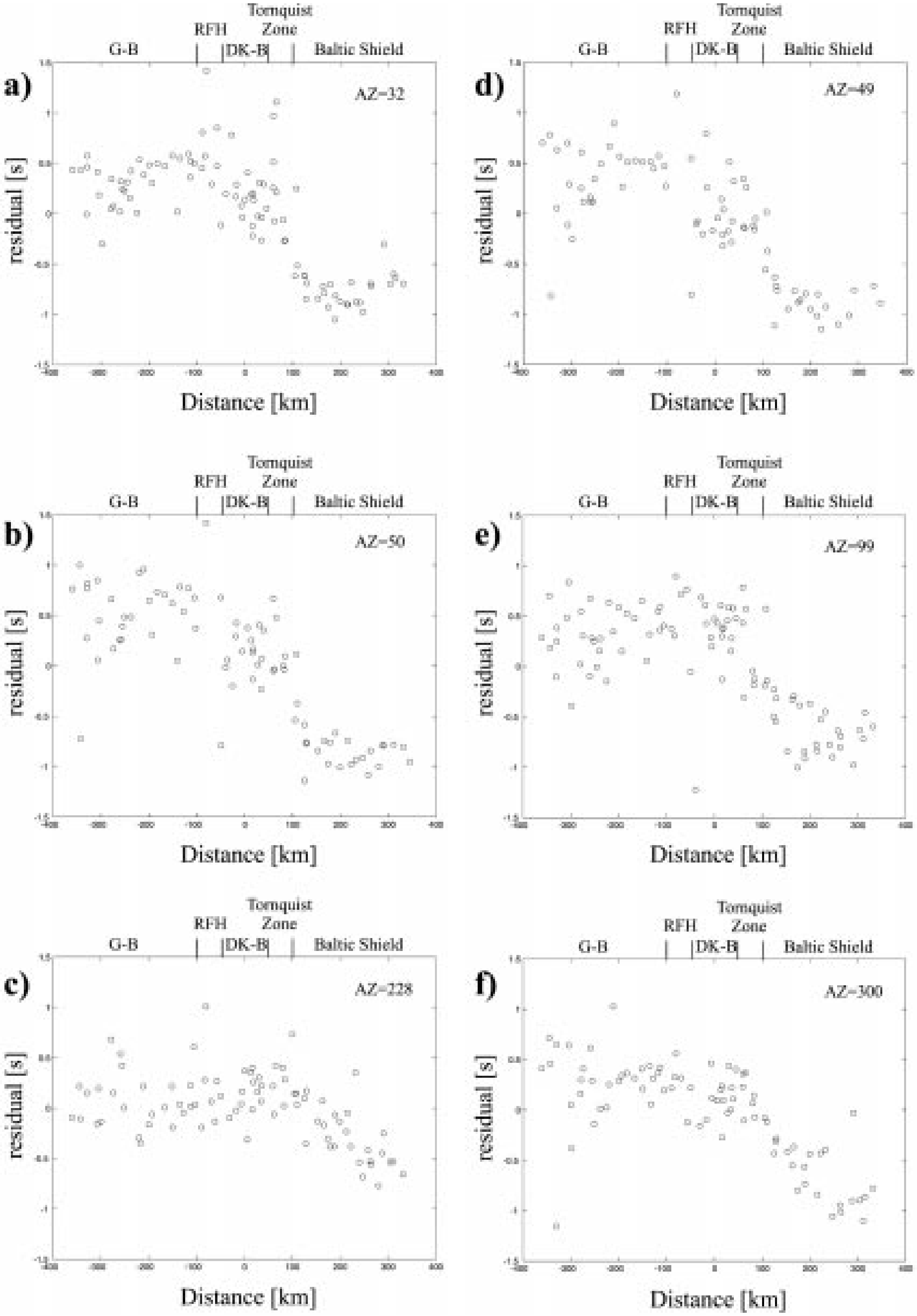

Fig. 5. Stripped residuals for the 6 events (listed in Table 2) projected onto a line perpendicular to the Tornquist Zone (Fig. 4f). G-B North-German Basin, RFH Ringkøbing-Fyn High, DK-B Norwegian-Danish Basin. AZ is the azimuth from which the P-waves arrive. Zero on the horizontal axis refers to the array centre (se Fig. 2).

Discussion

We are careful not to overinterpret these plots. However, we do see that the sharpest transition is near the Sorgenfrei-Tornquist Zone, for the earthquakes with waves crossing the array from azimuths 32°, 49° and 50°. These are waves coming along the array from earthquakes near Japan, which means that the P-rays arrive rather steeply from the north east, 15°–20°from the vertical (Pho & Behe, 1972). From previous geophysical investigations, summarized in the book on the European Geotraverse (Blundell et al., 1992) we have inferred that the dominating lithospheric difference is the thickness change across this area. We interpret that the rays from Japan travel almost along the deep separation zone between the low velocity asthenosphere and the thick lithosphere under the shield. This means that the transition is very steep (15°–20° from vertical).

The geographical location of the transition is steeply down from the Sorgenfrei-Tornquist Zone as indicated by the arguments on the sharp transition for the earthquakes from NE. Additionally, it is also in agreement with the plots for waves travelling perpendicular to the Tor array, azimuths 99° and 300° (Fig. 5), where the level of the shield residuals close to the Sorgenfrei-Tornquist Zone is delayed by some low velocity material below the edge of the shield. Similarly, it is in agreement with the very smooth image of the residuals from azimuth 228°, i.e. waves travelling along the Tor belt from the SW end.

The Tor project is an investigation of lithospheric variations from the Precambrian Shield to Paleozoic Europe. In this teleseismic tomography investigation of the lithosphere, we have mentioned that the geometry of the seismic rays in the crust is such that the rays are almost parallel, namely almost vertical. This gives poorer resolution in the crust than in the deeper parts of the lithosphere, so poor that it has been decided that the best possible 3D crustal model can not be determined from the teleseismic P-wave arrival times, but rather from previous 2D explosion studies. In the present paper we have derived and discussed this 3D crustal model. By ray tracing through this model, we have derived the residual pattern of the deep lithosphere/asthenosphere system by stripping away the influence of the upper 50 km.

Conclusion

Tomographic imaging of the deep lithospheric structure will eventually be done for close to 100 earthquakes. The pattern observed in a limited part of the data set has been reported in this investigation. We have determined the above-mentioned deep anomaly pattern for the P-wave arrivals of six earthquakes in different, representative azimuths. Hereby we are able to point out some typical features which serve as a preliminary result of the Tor project and a conclusion on sharpness of the transition, which will serve as a test for future tomographic inversions of the Tor data set.

The pattern of change from one end of the Tor array to the other has been observed as rather gradual. This implies that the physical change of P-velocities across the area from the Ringkøbing-Fyn High to the Sorgenfrei-Tornquist Zone is gradual in the deep part of the lithosphere. However, the change does have the sharpest gradient just under the Sorgenfrei-Tornquist Zone, indicating a deep extension of what is known to be a sharp transition zone in the sediments and in the Moho. In the first three plots of Figure 5 (Fig. 5a, b & d) a jump of the order of 1 second can be seen below the Tornquist Zone. These three plots illustrate the residuals for three earthquakes in the Pacific ocean, close to Japan. The rays come up rather steeply, i.e. with angles of incidence 15°–20°. Comparing with the three other plots (Figure 5c, e & f) for P-waves coming in from two directions perpendicular to the Tor belt (99° and 300°), and for P-waves coming in from the SW end of the Tor belt, it is argued that the rays coming up from Japan experience the sharpest inhomogeneity by following the sloping bottom of the lithosphere in the transition between thick lithosphere below the Baltic Shield and thin lithosphere below Denmark. The low velocity asthenosphere dips NE at a steep angle about 15°–20° from the vertical, and this low velocity material under the shield edge results in the gradual transition from one residual level to the other, seen in the three last plots of Figure 5 (c, e & f). The change in lithosphere thickness across the Sorgenfrei-Tornquist Zone is calculated to be more than 100 km, with a reasonable assumption on the P-wave velocity contrast of 5–7%. Whether or not there is a distinct lower boundary of the shield lithosphere can not be distinguished through these preliminary Tor investigations.

We have tried to eliminate the possibility that this effect could be caused by large source effects below Japan by choosing several earthquakes with different azimuth and distance; the same pattern emerges for the different sources .

In the first three plots of Figure 5 (a, b & d) a smaller step of approximately one third of a second is recognized below the Ringkøbing-Fyn basement High, so there is also a change in lithosphere/asthenosphere thicknesses here. This step is so small that its consequences are not distinguished in Figure 5c, e & f. If this obeservation turns out to be significant this smaller lithosphere change close to the Ringkøbing-Fyn High is more like 40 km, and steep like the larger change at the Tornquist Zone. Of the many possibilities in the Tor area, we find that the geological surface zone, which has the most prominent deep structure is the Tornquist Zone, while the Ringkøbing-Fyn High is of less significance. Further and more accurate conclusions are expected to emerge from the Tor project in the coming years.

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