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Crustal structure of the continent–ocean transition zone along two deep seismic transects in north-western Spitsbergen

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ABSTRACT: Deep seismic sounding measurements were performed in the continent-ocean transition zone of north-western Spitsbergen, during the expedition ARKTIS XV/2 of the RV *Polarstern* and the Polish ship *Eltanin* in 1999. Profile AWI-99200 is 430 km long and runs from the Molloy Deep in the Northern Atlantic to Nordaustlandet in north-eastern Svalbard. Profile AWI-99400 is 360 km long and runs from the Hovgård Ridge to Billefjorden. Seismic energy (airgun and TNT shots) was recorded by land (onshore) seismic stations (REF) and ocean bottom seismometers (OBS) and hydrophone systems (OBH). Good quality refracted and reflected P waves were recorded along the two profiles providing an excellent data base for a detailed seismic modelling along the profile tracks. Clear seismic records from airgun shots were obtained up to distances of 200 km at land stations and 50 km at OBSs. TNT explosions were recorded even up to distances of 300 km. A minimum depth of about 6 km of the Moho discontinuity was found east of the Molloy Deep. Here, the upper mantle exhibits P-wave velocity of about 7.9 km/s, and the crustal thickness does not exceed 4 km. The continent–ocean transition zone to the east is characterised by a complex seismic structure. The zone is covered by deep sedimentary basins. The Moho interface dips down to 28 km beneath the continental part of the 99200 profile, and down to 32 km beneath the 99400 profile. The P-wave velocity below the Moho increases up to 8.15 km/s. The continental crust consists of two or three crystalline layers. There is a lowermost crustal continental layer, in the 99400 profile's model, with the P-wave velocity in order of 7 km/s, which does not exist in the continental crust along the 99200 profile. Additionally, along the 99200 profile, we have found two reflectors in the lower lithosphere at depths of 14–42 and 40–50 km dipping eastward, with P-wave velocity contrasts of about 0.2 km/s. The characteristics of the region bears a shear-rift tectonic set-

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ting. The continent–ocean transition zone along the 99200 profile is mostly dominated by extension, so the last stage of the development of the margin can be classified as rifting. The uplifted Moho boundary close to the Molloy Deep can be interpreted as a south-western end of the Molloy Ridge. The margin in the 99400 profile area is of transform character.

Key words: Arctic, Spitsbergen, continent-ocean transition, seismic crustal structure.

Introduction

Spitsbergen is the main island of the Svalbard Archipelago located at the north-western corner of the Barents Sea continental platform and bordered to the west and north by passive continental margins (Fig. 1, insert map). The development of these margins is strongly connected to the history of rifting and subsequent sea-floor spreading in the North Atlantic Ocean (Jackson *et al.* 1990; Lyberis and Manby 1993a; Lyberis and Manby 1993b; Ohta 1994). The Svalbard continental margin has been studied by geophysical surveys over the last 30 years, mainly based on multichannel seismic reflection, sonobuoy refraction, gravity and magnetic measurements, as well as deep seismic soundings. The old investigations provided only limited information about the crystalline basement and deep crustal structure of this area (Guterch *et al.* 1978; Sellevoll 1982; Davydova *et al.* 1985; Faleide *et al.* 1991; Sellevoll *et al.* 1991; Czuba *et al.* 1999).

Based on a dense system of refraction and wide-angle reflection seismic vertical component data (Fig. 1), this paper presents a study of the transition zone between the continental and oceanic crust in the region of north-western Spitsbergen, where the North-Atlantic Ocean is opening. The profiles cross a transition zone between oceanic and continental crust with sedimentary basins, fracture zones and, presumably, a rifting zone. The experiment was carried out in the end of August 1999 by the expedition ARKTIS-XV/2 of RV *Polarstern* in co-operation with the Polish ship *Eltanin* (Jokat *et al.* 2000). It was a co-operation of the Alfred Wegener Institute for Polar and Marine Research, the Institute of Geophysics, Polish Academy of Sciences, the Hokkaido University and the University of Bergen. A total of 30 TNT shots and 3817 airgun shots were performed in the sea along the profiles (Fig. 1). The seismic energy was recorded in-line (and partially semi in-line along the 99200 profile) by land stations (REF), ocean bottom hydrophones (OBH) and ocean bottom seismometers (OBS). A detailed list of coordinates for stations and TNT shots is published by Jokat *et al.* (2000). This is the first experiment of a such quality in this region (Ritzmann *et al.* 2004; Czuba *et al.* submitted).

Tectonic setting

Spitsbergen is composed of different kinds of rocks ranging in age from Precambrian to Cenozoic (*e.g.* Birkenmajer 1993; Ohta 1994; Harland 1997; Dall-

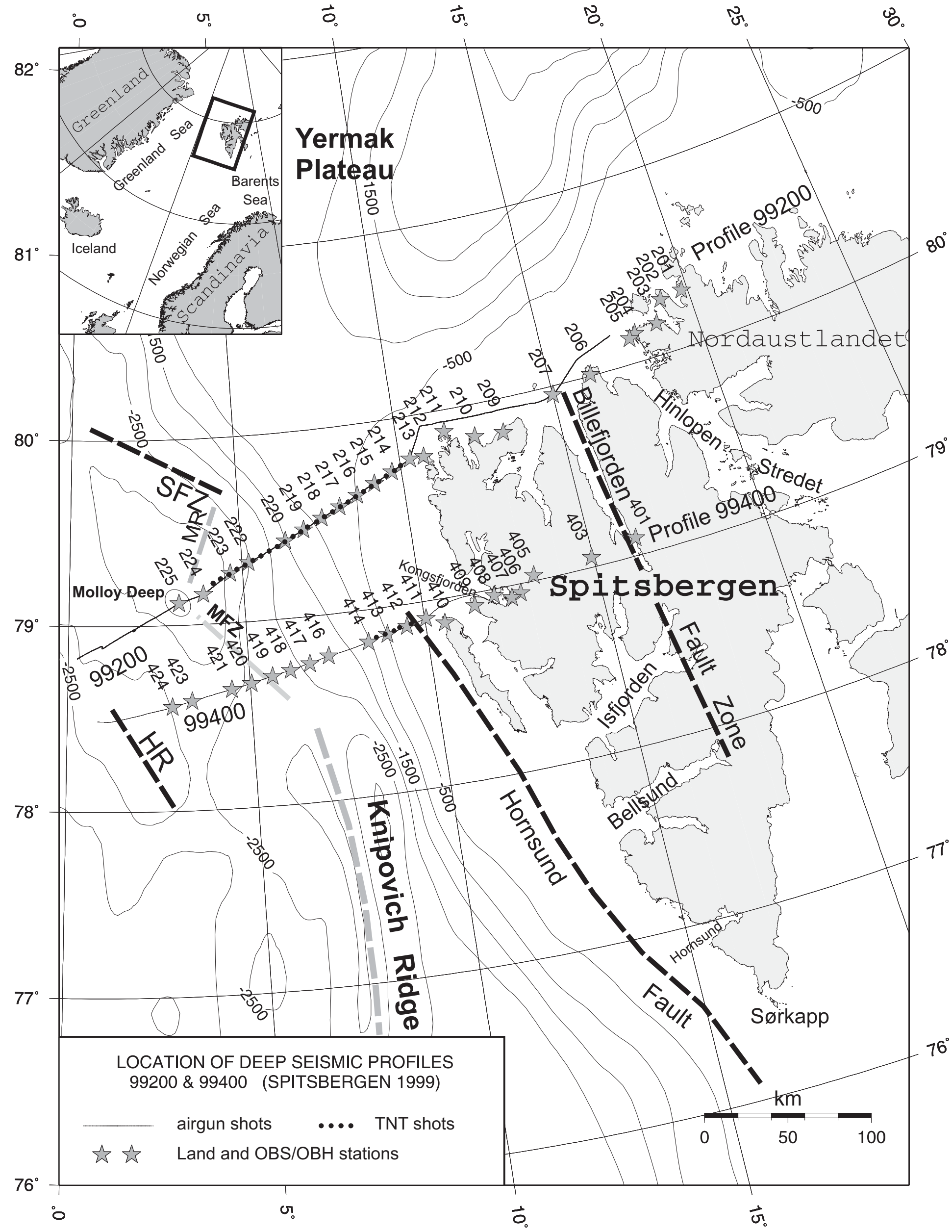

Fig. 1. Location map of the seismic transects 99200 and 99400 in the north-western Spitsbergen region. Stars are receivers, thin lines and dots are airgun and chemical shots, respectively. HR – Hovgård Ridge, MFZ – Molloy Fault Zone, MR – Molloy Ridge, SFZ – Spitsbergen Fault Zone.

mann 1999). The structure of the Svalbard Archipelago is a result of complex geological history reflecting the relative movements of the Eurasian and the North

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American plates (Eldholm *et al.* 1987). It is described in detail by Harland (1997). The tectonic development of the region can be simplified into terms of three main geological events (Sellevoll *et al.* 1991), however others divide it into more stages (Harland 1997). The first main tectonic phase is related to the Caledonian Orogeny (Birkenmajer 1981) whose effects are particularly well recognized in the eastern Svalbard (Sellevoll *et al.* 1991).

The next major tectonic phase is called the Late Devonian Svalbardian event. During this event the present-day eastern Spitsbergen and Nordaustlandet moved northward from eastern Greenland along the Billefjorden Fault Zone to a location north of Greenland (Sellevoll *et al.* 1991). This is a disputable hypothesis (Harland 1997). As a result, the eastern part of the Svalbard Archipelago attached to western Spitsbergen. Western Spitsbergen was supposed to be located in the north before these movements or slightly moved northward from a shorter distance (Harland and Cutbill 1974; Sellevoll *et al.* 1991). The N-S trending lineaments which controlled the Devonian sedimentation and tectonism were reactivated in Mid-Carboniferous during a new period of crustal extension (Steel and Worsley 1984). The last major deformation took place in the Tertiary time. Dextral transcurrent movements occured along a N-S trending fault west of the Palaeozoic fault lines (predominantly along the Hornsund Fracture Zone; Fig. 1). The early stages (Paleocene) of the movements along the west coast of Spitsbergen were probably transtensional. The regime changed to transpressional later when the present margins of the Greenland and Eurasian plates were deformed. This process led to the West Spitsbergen Orogeny (Harland and Cutbill 1974; Steel *et al.* 1985; Harland 1997). During the orogeny a narrow thrust and fold belt developed along the west coast of Spitsbergen. Today the fold belt crops out south of Kongsfjorden. Extensive erosion led to increased sedimentary load along the western margin of Svalbard and turned the epicontinental, littoral basin of central Spitsbergen into a rapidly subsiding foreland basin (Eiken and Austegard 1987; Sellevoll *et al.* 1991).

The subsequent tectonic history of Svalbard can be considered in terms of a postorogenic relaxation of tectonic stresses. The main depocentre of Neogene sedimentation shifted to the west, where thick clastic wedges have been accumulated offshore (Eiken and Austegard 1987). Cenozoic tectonic processes in the Svalbard region were closely related to the structural history of the western Barents Sea margin. The relative motion between Svalbard and Greenland took place along the NNW-SSE trending Hornsund Fault Zone with no accompanying crustal extension in the Greenland Sea. This regional fault zone acted as an incipient plate boundary between the Barents Sea shelf and the emerging Arctic Ocean. The initial opening of the southern Greenland Sea apparently began in Early Eocene (Faleide *et al.* 1988). No significant separation between Svalbard and Greenland occured until about 36 Ma ago. The seafloor spreading in the Norwegian Sea and the Arctic Ocean began approximately 57–58 Ma ago (Vogt and Avery 1974; Labrecque *et al.* 1977;

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Talwani and Eldholm 1977). The spreading axis in the Greenland Sea is today represented by the Knipovich Ridge (Fig. 1). Its northward extension to the Fram Strait is represented by the Spitsbergen Transform Fault Zone (SFZ), the Molloy Ridge (MR) and the Molloy Transform Fault Zone (MFZ) (Fig. 1).

The Hornsund Fault, the prominent tectonic structure which parallels the Knipovich Ridge to the east, can be traced from just south of Bjørnøya at ca. 75°N, to about 79°N (Sundvor and Eldholm 1979; Sundvor and Eldholm 1980). Another main tectonic structure, the Yermak Plateau, is located north of western Spitsbergen (Fig.1). It was considered a volcanic body probably connected with early spreading along the Gakkel Ridge (Feden *et al.* 1979; Harland 1997). Recent results indicate that at least its south-western part is a fragment of the Svalbard continental crust with no indications of extensive volcanic activity in the middle and lower crust (Ritzmann and Jokat 2003).

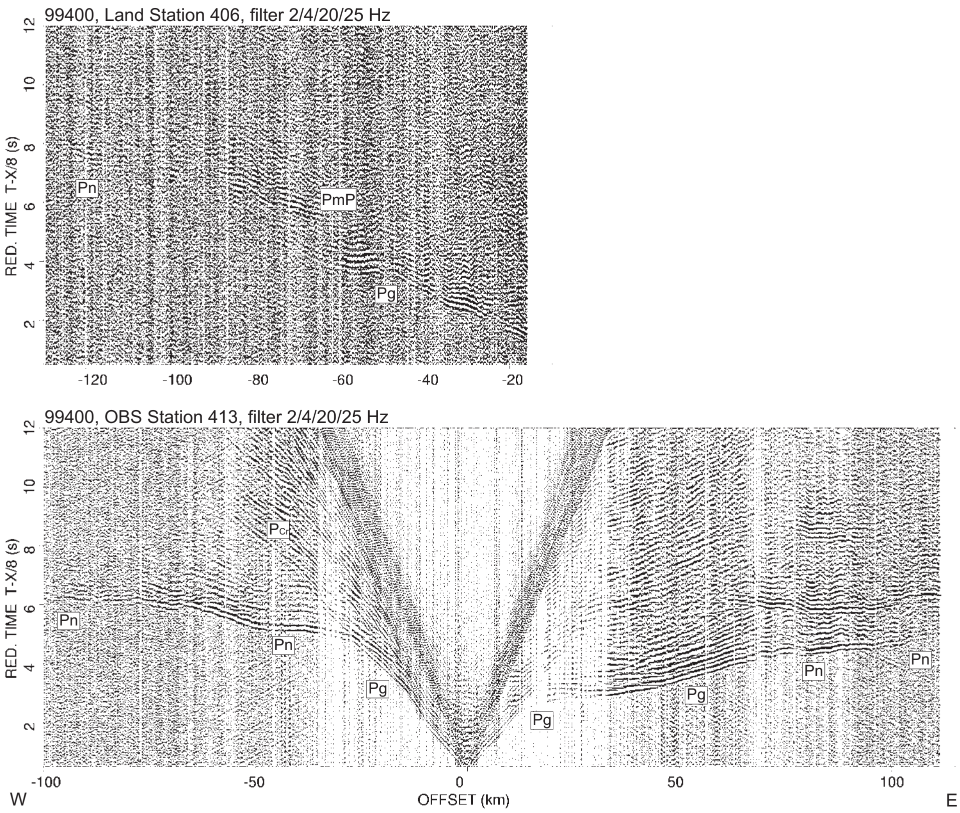

Fig. 2. Examples of amplitude-normalized airgun seismic record sections along the 99400 profile, stations 406, Reftek land station (top) and 413, OBS (bottom). Note high apparent Pg velocity at the Reftek land station (upper section), low apparent Pn velocity, clear P_{Cr} phases and later Pn arrivals to the west at the OBS (lower section). Pg – first arrivals of crustal P-waves; P_{Cr} – later crustal refracted P-waves; Pn – refracted P-waves beneath the Moho; PmP – Moho P-wave reflections.

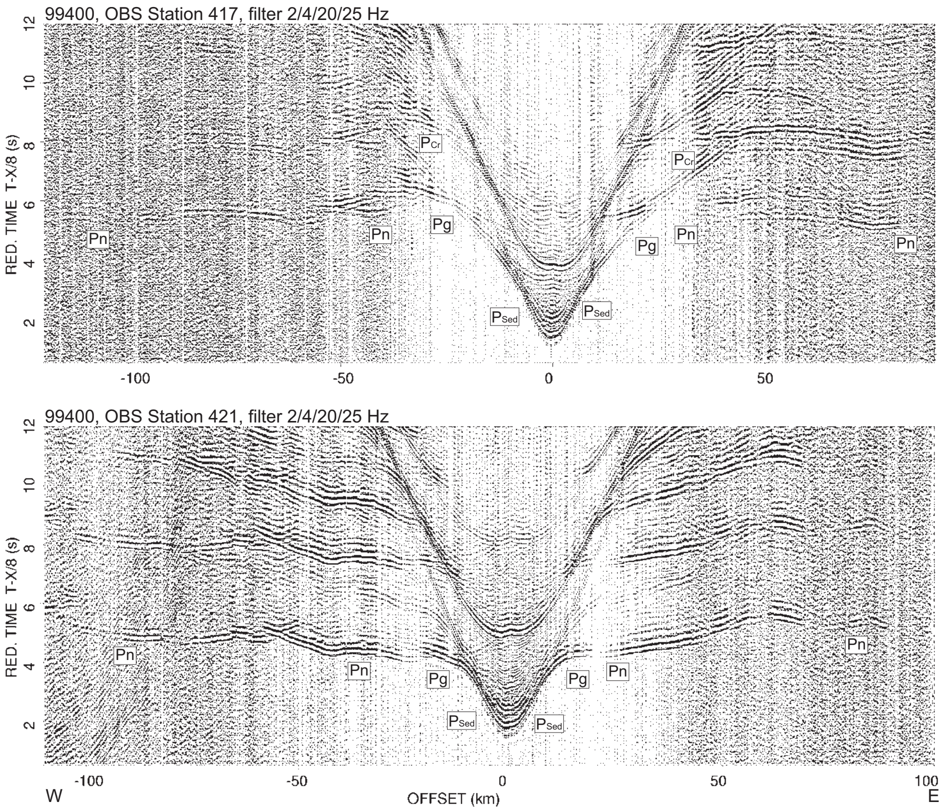

Fig. 3. Examples of amplitude-normalized airgun seismic record sections along the 99400 profile, stations 417, OBS (top) and 421, OBS (bottom). Note early Pn arrivals at about 5 s of reduced time and almost no Pg phases (both sections), and quite long P_{Sed} branches (lower section). P_{Sed} – refracted P-waves in sediments. Other descriptions as in the Fig. 2.

Seismic processing, wave field and method of interpretation

In general, good quality recordings allowed us to perform a detailed study of the seismic wave field and crustal structure. The seismic processing procedure is described in Ritzmann *et al.* (2004) and Czuba *et al.* (submitted).

Examples of seismic record sections from the profiles studied are presented in Figs 2–6. Figs 7 and 8 show examples of seismic record sections together with synthetic seismogram sections and ray diagrams calculated using the SEIS83 software package (Červený *et al.* 1977; Červený and Pšenčík 1983) and the graphical interface MODEL (Komminaho 1993) for comparison of the observed and calculated data. All airgun record sections from onshore stations and ocean bottom seismometers show distinct P-wave first arrivals and most of them Moho reflections. The seismic energy is clearly visible from airgun shots over distances of about 200 km and 300 km from TNT shots at the onshore stations, and up to 50 km and 100 km

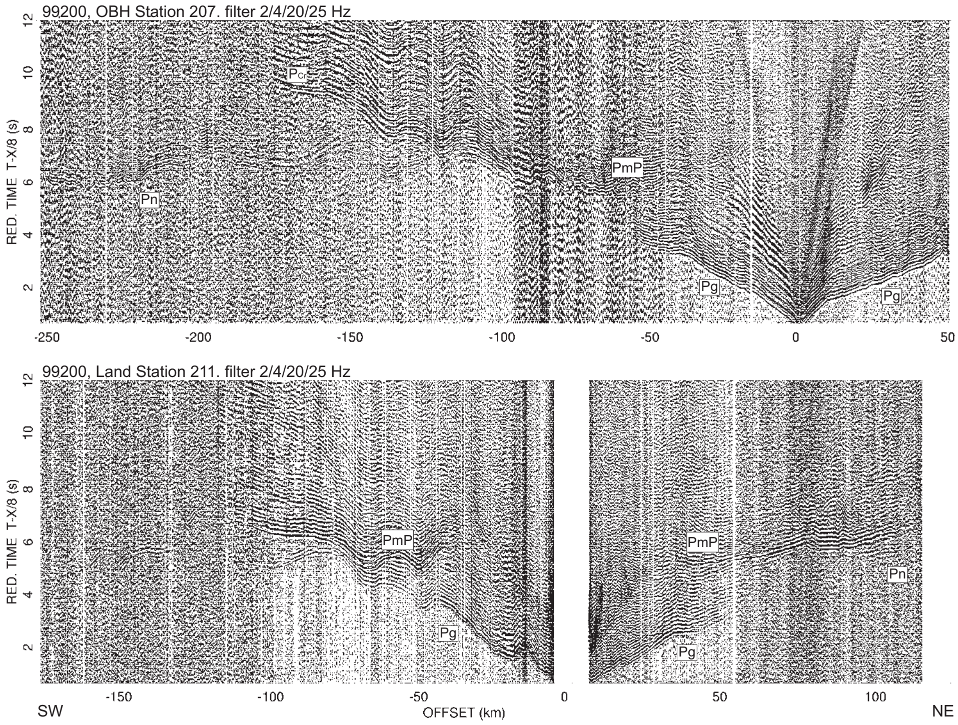

Fig. 4. Examples of amplitude-normalized airgun seismic record sections along the 99200 profile, stations 207, OBH (top) and 211, Reftek land station (bottom). Note extremely good energy at the OBH (upper section), complicated shape of arrivals envelops left of -100 km (upper section) and -50 km (lower section). Other descriptions as in Fig. 2.

from airgun and TNT shots at the OBSs, respectively. There is very good signal to noise ratio in the TNT record sections enabling very precise identification of the first arrivals even for the P_n or lower lithospheric phases. The seismic signal range recorded by the OBHs is very short. The only exception is the OBH station 207 which exhibits an extremely good signal range (Fig. 4).

First arrivals with apparent seismic velocities lower than about 2 km/s (P_{Sed}) generally associated with upper sedimentary layers have been recorded at offsets up to 10 km, *e.g.* at stations 417, 421 (Fig. 3) and 215 (Fig. 5). Long distance P_g waves in the eastern part of the 99200 profile with apparent velocity of 5–6 km/s show a refractive interface present in this part of the crust. There are almost no P_g waves on the western side of the profiles (station 225, Fig. 5), or they appear at distances no greater than 30 km (stations 413, 417, 421, 217 and 223, *e.g.* Figs 2, 3 and 9). There are branches with apparent velocity in the range of 8 km/s close to the stations. This indicates the presence of a very shallow Moho interface in that area. Clear and distinct Moho reflections with large relative amplitudes in the eastern part of the 99200 profile with reduced travel time of more than 6 s reflect a high

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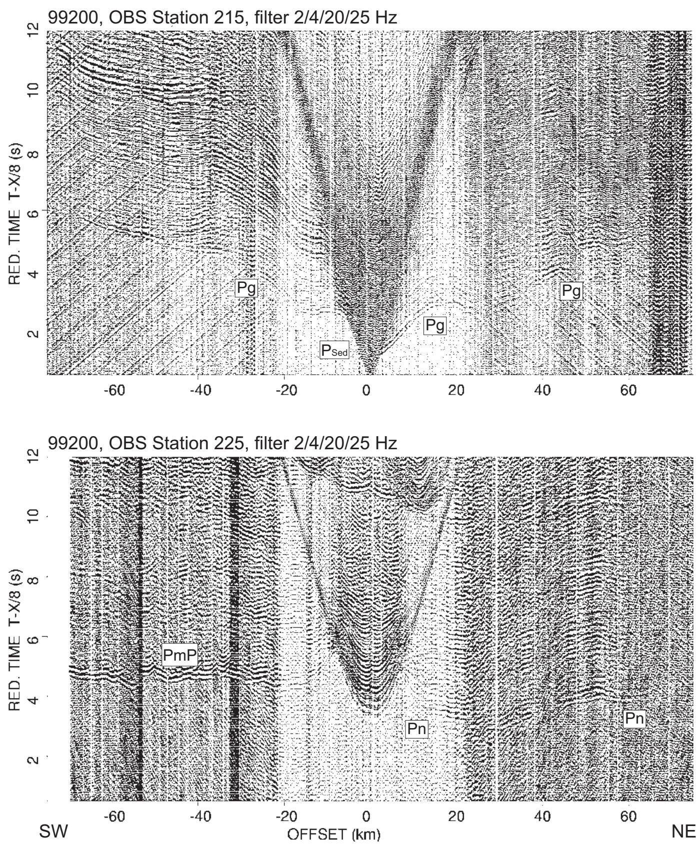

Fig. 5. Examples of amplitude-normalized airgun seismic record sections along the 99200 profile, OBS stations 215 (top) and 225 (bottom). Note high Pn apparent velocity and early PmP arrivals on the bottom section. Other descriptions as in Fig. 2.

velocity contrast at the interface (Fig. 7). Earlier than in the eastern part, the arrivals from Moho reflections and refractions in the western part of the profiles clearly show the Moho dipping from the west to the east along the profiles (station 413 – Fig. 2; compare stations 406 and 421 – Figs 2 and 3; compare stations 202 with 225 and 207 – Figs 4, 5 and 7). It is worth noting a "ringing" character of Moho reflections along the 99200 profile. Weak, difficult to distinguish, Moho reflections in the middle part of the 99200 profile indicate low velocity contrast at the Moho interface (Fig. 4, station 211; Fig. 5, station 215 – no Moho visible). Complicated, undulating shape of the branches is a result of the sea bottom morphology and of the shape of the underlying sedimentary layers.

The TNT explosion records allow to get information from longer distances and from deeper boundaries. It is possible to find deep reflections from the lower lithosphere. They are clearly visible, for example at stations 203, 205, 207, 210, 212 (*e.g.* Figs 6 and 8, P1 branches).

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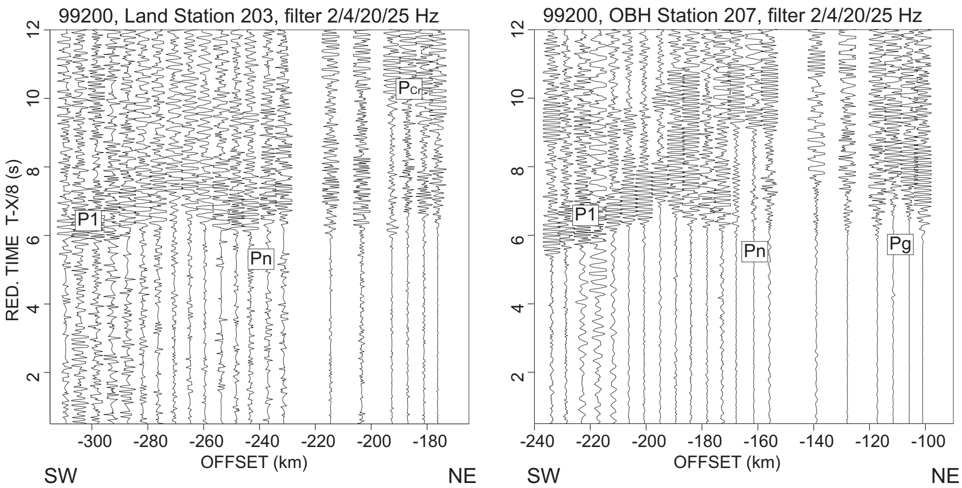

Fig. 6. Examples of amplitude-normalized TNT seismic record sections along the 99200 profile, stations: Reftek 203 (left) and OBH 207 (right). P1 – lower lihospheric reflections. Note clear first arrivals on all the traces. Other descriptions as in Fig. 2.

Seismic models

Profile 99200.—The 2-D seismic velocity model shown in Fig. 9a was constructed using the SEIS83 software package (Červený and Pšenčík 1983). It focuses on the crust which can be horizontally divided into 3 parts: oceanic, transitional and continental (Czuba *et al.* submitted).

The oceanic part extends from the western termination of the model (km 0) to the location of station 225 (km 70, Molloy Deep). The water depth ranges from approximately 2.5 km down to 5.5 km in the Molloy Deep. The seismic structure is quite simple and consists of two layers only: (1) the upper, sedimentary 1–2 km thick layer with P-wave velocity of about 1.95 km/s, and (2) the lower layer, crystalline, 5–10 km thick. The P-wave velocity in the crust is 6.60–6.75 km/s. The Moho boundary is located at a depth of about 13 km. The thickness of the crystalline crust below the Molloy Deep is determined to be about 4.5 km only. The crustal model west of the point km 70 is inaccurate because the station 225 is the most western station along the profile, so the transect is unreversed west of it.

The transitional part, extending from about km 70 of the model to about km 240 (stations 212 and 213), is more complex. It consists of two sedimentary layers forming a large, complex cover reaching a depth of about 8 km at about km 140 of the model. The upper sedimentary layer, characterised by P-wave velocity of about 1.9 km/s, is about 1–2 km thick and shows four depressions. The lower sedimentary layer exhibits P-wave velocity of 3.60–3.85 km/s and is 1–4 km thick. The basement is determined as a rather complex block structure. The section of km

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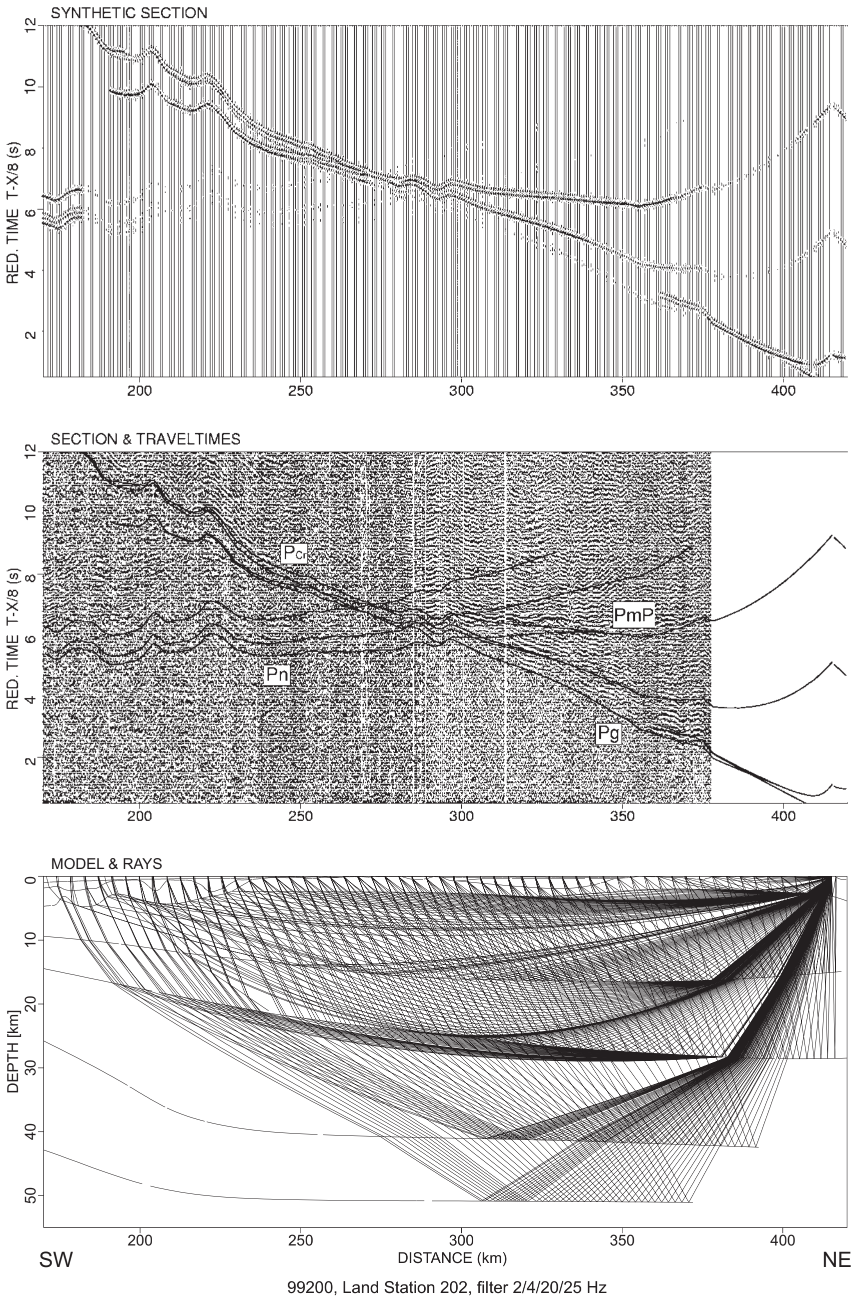

Fig. 7. Example of the amplitude-normalized airgun seismic record section along the 99200 profile for the 202 station (Reftek) and theoretical travel times calculated for the model (in the middle, VA plot). The upper diagram shows the synthetic seismograms (WT/VA plot), which should be compared with the data in terms of the relative amplitudes within a trace. The bottom portion of the figure is a diagram with refracted and reflected rays in the model calculated from 100 shots only for better presentation. Other descriptions as in Fig. 2.

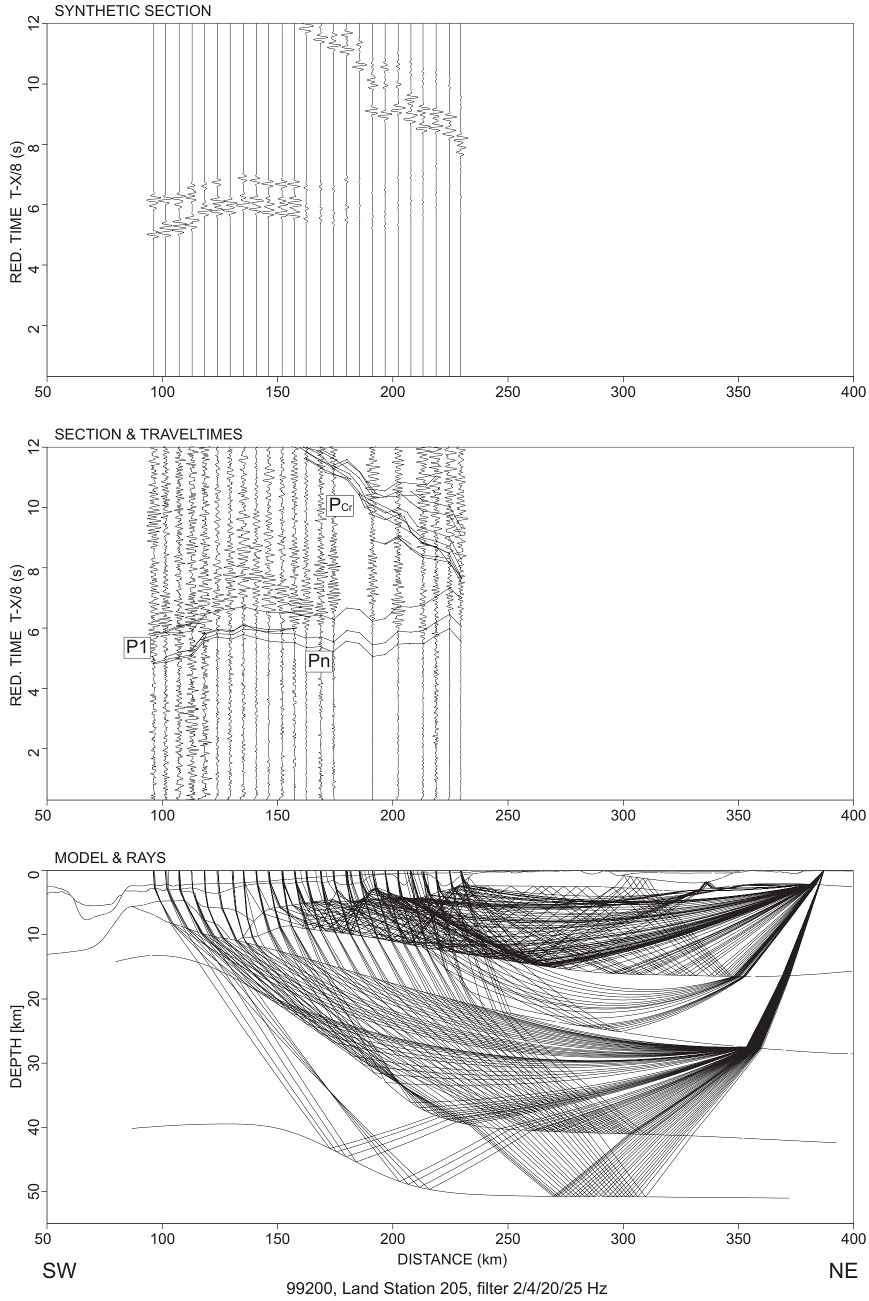

Fig. 8. Example of the amplitude-normalized TNT seismic record section along the 99200 profile for the 205 station (Reftek) and theoretical travel times calculated for the model (in the middle). P1 – lower lihospheric reflections. The upper diagram shows the synthetic seismograms, which should be compared with the data in terms of the relative amplitudes within a trace. The bottom portion of the figure is a diagram with refracted and reflected rays in the model. Note clear P1 arrivals and good kinematical and dynamical fit of later crustal arrivals. Other descriptions as in Fig. 2.

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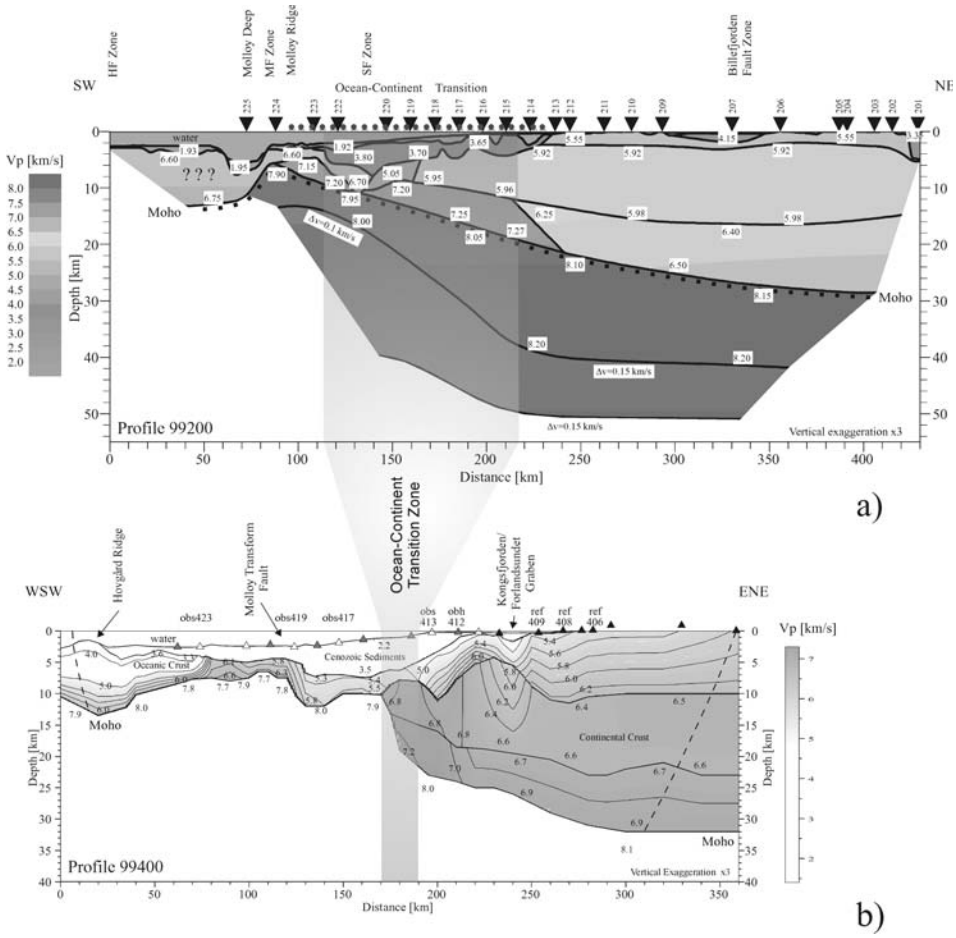

Fig. 9. Two-dimensional seismic P-wave velocity models along the 99200 (a) and 99400 (b) profiles developed by raytracing techniques. Black lines represent seismic discontinuities (boundaries); solid/dotted line marks the Moho interface (top); shades represent the distribution of the P-wave velocity limited to the area of rays coverage; dashed lines limit the area of rays coverage (bottom); numbers in the models – P-wave velocities in km/s. Triangles show location of the stations, dots show location of the TNT shots. Vertical exaggeration is 3:1. Question marks point out uncertainty in the model.

85–110 of this part is composed of eastward dipping layers: the upper layer with P-wave velocity of 6.6–6.7 km/s, and the lower block with P-wave velocity of 7.15–7.20 km/s. The shallowest Moho boundary occurs beneath this block dipping from 6 km eastward, with P-wave velocity below the shallowest location less than 7.9 km/s (km 90). To the east, a small block exhibits P-wave velocity of about 5.05 km/s. It is followed, from the point of km 160, by a layer characterised by P-wave velocity of 5.9–6.0 km/s which is continued in the third, continental, part of the model. The layer and block are underlain by a body extending from 9 to 21 km

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deep and from about the point of km 135 to the point of 240 horizontally. The P-wave velocity is determined there in the range of 7.2–7.3 km/s. The lower boundary of this body is the Moho interface, descending eastward. The P-wave velocity below the Moho increases from 7.9 to 8.1 km/s eastward.

The continental part extends from the point of km 240 to the eastern termination of the profile. It consists of a sedimentary cover overlying a thin (3 km thick) layer with the P-wave velocity in order of 5.5 km/s and two layers representing the crystalline crust. The upper layer of the crystalline crust is characterised by P-wave velocity in the range of 5.9–6.0 km/s. Beneath this layer, a lower crustal layer with P-wave velocity of 6.25–6.60 km/s has been determined. There is no layer characterised by the P-wave velocities around 7 km/s in the continental part of the profile. The Moho interface is determined at a depth of 21–29 km, and it is characterised by P-wave velocity shift to about 8.1 km/s (east of km 240).

Additionally, based on TNT shots only, we have revealed two reflectors in the mantle lithosphere (see in Fig. 9, P1 wave at distance km 100–130) (Czuba *et al.* submitted).

Profile 99400.—The 2-D seismic velocity model shown in Fig. 9b was constructed using the raytracing software RAYINVR (Zelt and Smith 1992). It focuses on the crust which can be horizontally divided into 2 parts: oceanic and continental (Ritzmann *et al.* 2004).

The thickness of the sedimentary cover, which partly overlies the oceanic and continental crust, varies from 1 km close to the Hovgård Ridge to 6 km east of the Molloy Fault Zone (Molloy Transform Fault). P-wave velocities are in the range of 2.3–3.8 km/s.

The continental (eastern) part of the profile consists of three layers. The uppermost layer is characterised by P-wave velocities in the range 5.4–6.2 km/s. Its thickness varies from 5 km beneath the Forlandsundet Graben to 10 km in the east. The middle layer shows a complex structure. It is characterised between points of km 250 and 360 by P-wave velocities of 6.4–6.6 km/s at depths of 10–23 km. In the western part of the layer, beneath the Forlandsundet Graben, the P-wave velocities decrease to 5.8 km/s in the upper part of the layer, at a depth of about 5 km. This vertical feature of the velocity isolines is 20–30 km wide along the profile. The lower layer is characterised by P-wave velocities of 6.7–6.9 km/s, and it thins westward from 9 to 5 km. The Moho interface depth varies from 32 km at the eastern end of the profile to 23 km at ca. point of km 190 of the model, steeply ascending to 10 km at km 170, which marks a continent–ocean boundary. At the continent–ocean transition, P-wave velocities in the continental crust increase to 6.8–7.2 km/s. The mantle P-wave velocity beneath the continental part of the profile increases from 7.9 km/s in the transition zone to 8.1 km/s at the eastern end of the profile.

The crust in the oceanic (western) part of the profile is thin, varying between 2 and 5 km, and being of about 12 km thick in the Hovgård Ridge area. The P-wave velocities increase westward from 5.3–5.8 km/s to 6.3–6.6 km/s. In the vicinity of the Hovgård Ridge (km 0–80), the P-wave velocities decrease to about 4–6 km/s. The mantle P-wave velocity varies between 7.7 and 8.0 km/s. The lowest velocity is observed west of the Molloy Transform Fault.

Because the station 424 is the westernmost station along the profile, the transect is unreversed west of the point of km 60. Therefore, the crustal model along this part is quite unreliable. We can only conclude that the crust there is probably oceanic (Ritzmann *et al.* 2004).

The accuracy of the models is discussed in Ritzmann *et al.* (2004) and Czuba *et al.* (submitted).

Discussion

The basin located in the continent-ocean transition zone in both crustal models is a well known west Spitsbergen foreland basin (Eiken and Austegard 1987), which contains probably the Cenozoic sequence. The P-wave velocity is there 1.9 km/s only (Fig. 9), what indicates high water saturation of the rock body.

The uplifted Moho boundary close to the Molloy Deep can be associated with the Molloy Transform Fault Zone or with the Molloy Ridge connecting the Molloy Transform Fault Zone with the Spitsbergen Transform Fault Zone. We argue for an existence of a transform margin in the near past (Czuba *et al.* submitted) in the ocean-continent transition region of the profile. The thinned continental crust along only ca. 100 km long transition zone indicates quite young or weak extensional regime. The thinnest crust (~3 km thick), in the 99200 model, at about km 90, with the P-wave velocity of 7.9 km/s in the upper mantle marks probably the location of the rift axis. The well developed "ringing" wave form of the Moho reflections in the eastern part of the profile suggests a transitional character of the Moho discontinuity (Czuba *et al.* submitted).

Ritzmann *et al.* (2004) suggest the existence of a transform margin in the study region which is confirmed by the short (less than 20 km) continent–ocean transition zone along the 99400 profile. We probably observe the oceanic crust west of the km 170 in the 99400 model.

There has been revealed a lower continental crustal layer along the 99400 profile characterised by P-wave velocities reaching 7 km/s. A similar layer was not found in the continental part of the 99200 profile. A similar feature (disappearing of the P-wave velocity 7 km/s layer in the northern Spitsbergen) was observed in the results of previous geophysical study in the NW Spitsbergen (along K2 and C3/C1 transects; Czuba *et al.* 1999).

The results of this study are in good agreement with other experiments in this region. The continental structure is similar for all the models, except of high P-wave velocities below the Moho boundary in the continental part of the K1

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profile (Czuba *et al.* 1999). As described above, the character of the west Spitsbergen continent–ocean margin changes from the south to the north in the NW Spitsbergen region. The margin crossed by the 99400 profile can be classified as a transform margin (Ritzmann *et al.* 2004). The margin crossed by the 99200 profile has more complex history. Although at the present the margin seems to be a rifted margin, we have found features indicating the transform history of the margin in the near past (Czuba *et al.* submitted).

Conclusions

A dense system of airgun shots allows to model very accurately the seismic crustal structure along two parallel profiles in the NW Spitsbergen region. It provides an important base for geological interpretation and modelling deeper layers in the region.

We have obtained detailed seismic P-wave velocity models down to 50 km along the 99200 transect and down to 35 km along the 99400 profile. We have found deep sedimentary basins (probably of Cenozoic sequence) with a low seismic velocity and the thinned continental crust in the continent-ocean transition zone. The continent-ocean transition zone is ca. 100 km wide in the 99200 model, while only 20 km wide in the 99400 model. We interpret it as a result of an extensional regime, acting here from the anomaly 13 time (Eiken and Austegard 1987; Faleide *et al.* 1991; Harland 1997), which probably has masked previous shear structure of the margin crossed by the 99200 profile (Czuba *et al.* submitted). The 99400 profile demonstrates rather standard transform margin structure.

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