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Seismic crustal structure along the deep transect Horsted'05, Svalbard

Wojciech CZUBA[1*], Marek GRAD [2,1], Aleksander GUTERCH [1], Mariusz MAJDAŃSKI [1], Michał MALINOWSKI [1], Rolf MJELDE [3], Mateusz MOSKALIK[1], Piotr ŚRODA[1], Monika WILDE-PIÓRKO [2] and Yuichi NISHIMURA[4]

[1] *Instytut Geofizyki, Polska Akademia Nauk, Księcia Janusza 64, 01-452 Warszawa, Poland <wojt@igf.edu.pl> <mgrad@mimuw.edu.pl> <aguterch@igf.edu.pl> <mmajd@igf.fuw.edu.pl> <michalm@igf.edu.pl> <mmosk@igf.edu.pl>*

[2] *Instytut Geofizyki, Uniwersytet Warszawski, Pasteura 7, 02-093 Warszawa, Poland <mgrad@mimuw.edu.pl> <mwilde@igf.fuw.edu.pl>*

[3] *Institute of Solid Earth Physics, University of Bergen, Post Box 7800, N-5020 Bergen, Norway <Rolf.Mjelde@geo.uib.no>*

[4] *Institute of Seismology and Volcanology, Hokkaido University, N10S8 Kita-ku Sapporo 060-0810, Japan <nishi@eos.hokudai.ac.jp>*

Abstract: During the Polish-Norwegian-Japanese Polar Expedition in August 2005 a deep seismic sounding of the Earth's crust was performed in the southern Svalbard area. It was an introduction to the international Cluster Program 77: *Plate Tectonics and Polar Gateways in Earth History*, closely related to the International Polar Year. During the expedition onboard r/v *Horyzont II*, a seismic transect Horsted'05 along Hornsund and crossing Storfjorden and Edgeøya was performed. Seismic energy (provided by an airgun and TNT shots) was recorded by land (onshore) seismic stations and ocean bottom seismometers (OBS) deployed along the transect, resulting in high quality refracted and reflected P-waves record sections. Clear seismic records from airgun shots were obtained up to distances of 150 km at land stations and OBSs. TNT explosions were recorded along the whole transect length even up to distances of 300 km. The obtained 2-D seismic model is complicated in the western part which is located ~100 km east of Knipovich Ridge. The Moho shallows in this area up to 14 km, while beneath the Hornsund it dips down to 33 km. The eastern part of the model is characterized by simpler structure, with two layers in the upper crust Vp = 5.5–6 km/s and in the lower crust Vp = 6.1–6.2 km/s. There is no P-wave velocity higher than 6.5 km/s in the crust of the Storfjorden and Edgeøya area. The Moho depth varies between 26 and 31 km.

Key words: Arctic, Svalbard, Spitsbergen, seismic crustal structure.

* *Corresponding author*

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. 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 region has been studied by geophysical surveys over the last 40 years, mainly based on multichannel reflection seismic, sonobuoy refraction surveys, gravity and magnetic measurements, as well as deep seismic soundings. The old investigations provided only limited information about the crystalline basement and the deep crustal structure of this area (Sellevoll 1982; Davydova *et al.* 1985; Faleide *et al.* 1991; Sellevoll *et al.* 1991; Czuba *et al.* 1999; Ljones e*t al.* 2004; Czuba *et al.* 2005).

This paper presents a study of the crustal structure in the southern Spitsbergen region along a deep seismic sounding transect based on a dense system of refraction and wide-angle reflection seismic data (Fig. 1). The Horsted'05 experiment was carried out in August 2005 as an introduction to the international Cluster Program 77: *Plate Tectonics and Polar Gateways in Earth History* (PLATES and GATES), closely connected with the 4th International Polar Year. It was a co-operation of the Institute of Geophysics, Polish Academy of Sciences, the University of Bergen and the Hokkaido University. 370-km long Horsted'05 profile starts in the Atlantic Ocean, crosses Spitsbergen along Hornsund, Storfjorden, then Edgeøya and ends in the Barents Sea.

A total of 26 TNT shots and 235 airgun shots were performed in the sea along the transect (Fig. 1). Seismic energy was recorded in-line by 11 land stations and 10 ocean bottom seismometers (OBS). This was the first experiment of the deep seismic sounding using TNT charges in this area of Svalbard.

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; Dallmann 1999). The structure of the Svalbard Archipelago results from the complex geological history reflecting the relative movements of the Eurasian and the North American plates (Eldholm *et al.* 1987). It is described in detail by Harland (1997). The tectonic development of the region can be simplified into three main geological events (Sellevoll *et al.* 1991), however others divide it into more stages (Harland 1997).

The first tectonic phase is related to the Caledonian Orogeny (Birkenmajer 1981) whose effects are particularly well recognized in the eastern Svalbard

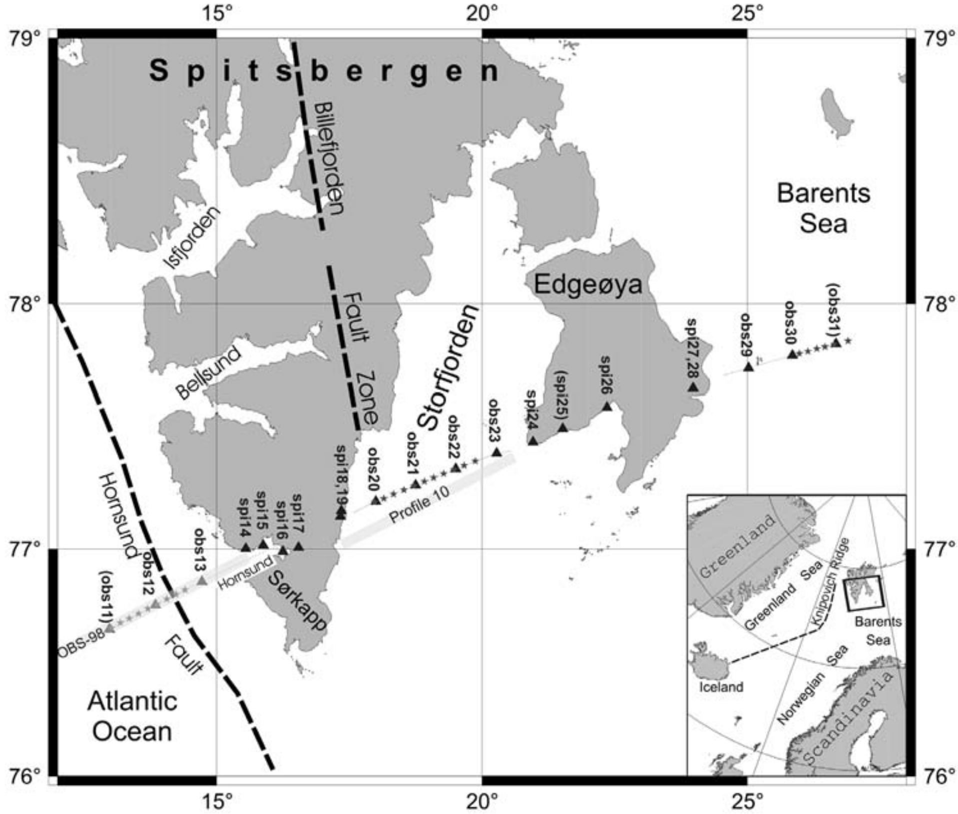

Fig. 1. Location map of the seismic transect Horsted'05. Stars are TNT shots, dots are airgun shots, triangles are seismic stations. In brackets – corrupted stations with no data recorded. Gray-shaded bars – previous seismic profiles in the area: OBS-98 (Ljones *et al.* 2004) and Profile 10 (Breivik *et al.* 2005).

(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). As a result, the eastern part of the Svalbard Archipelago attached to the 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 last major deformation took place in the Tertiary time. During the West Spitsbergen Orogeny (Harland and Cutbill 1974; Steel *et al.* 1985; Harland 1997) a narrow thrust and fold belt developed along the west coast of Spitsbergen. 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 the tectonic stresses. 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 (Fig. 1) 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). 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; Talwani and Eldholm 1977). The spreading axis in the Greenland Sea is represented today by the Knipovich Ridge (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).

Experimental setup and seismic wave field

Seismic investigations along the Horsted'05 transect were performed in August 2005 during the expedition of the Polish ship r/v *Horyzont II*. The experiment was carried out in cooperation of the Institute of Geophysics, Polish Academy of Sciences, the University of Bergen and the Hokkaido University. Two seismic sources were used during the experiment: (i) The entire sea part of the transect was covered by airgun shots. Seismic energy was generated by 3 airguns with a total volume of 60 l. 235 airgun shots were performed in total with an average spacing of about 800 m; (ii) The ship fired also 26 shots of 25–100 kg TNT charge (Fig. 1). Firing depth of the chemical explosions was approximately 60 m. The shot spacing of the TNT shots was approximately 5 km. The seismic energy was recorded in-line by 11 land stations deployed onshore and 10 ocean bottom seismometers (OBS). There were no data from two outlying OBSs (obs11, obs31 see Fig. 1). Some land station were corrupted by polar bears but we have got most of the data. Land station 25 was corrupted probably by tourists and there is no data at all (Fig. 1).

In general, good quality recordings allowed us to perform a detailed study of the seismic wave field and crustal structure along the transect. We have used vertical component data only in this study. Examples of seismic record sections are presented in Figs 2–4. All the sections show calculated travel times. Figs 2 and 4 show additionally ray diagram and synthetic section. There is a comparison of the airgun and TNT record sections in Fig. 3. All airgun record sections from onshore stations and ocean bottom seismometers show distinct P-wave first arrivals and most of them contain Moho reflections. Quite clear seismic records from airgun shots were

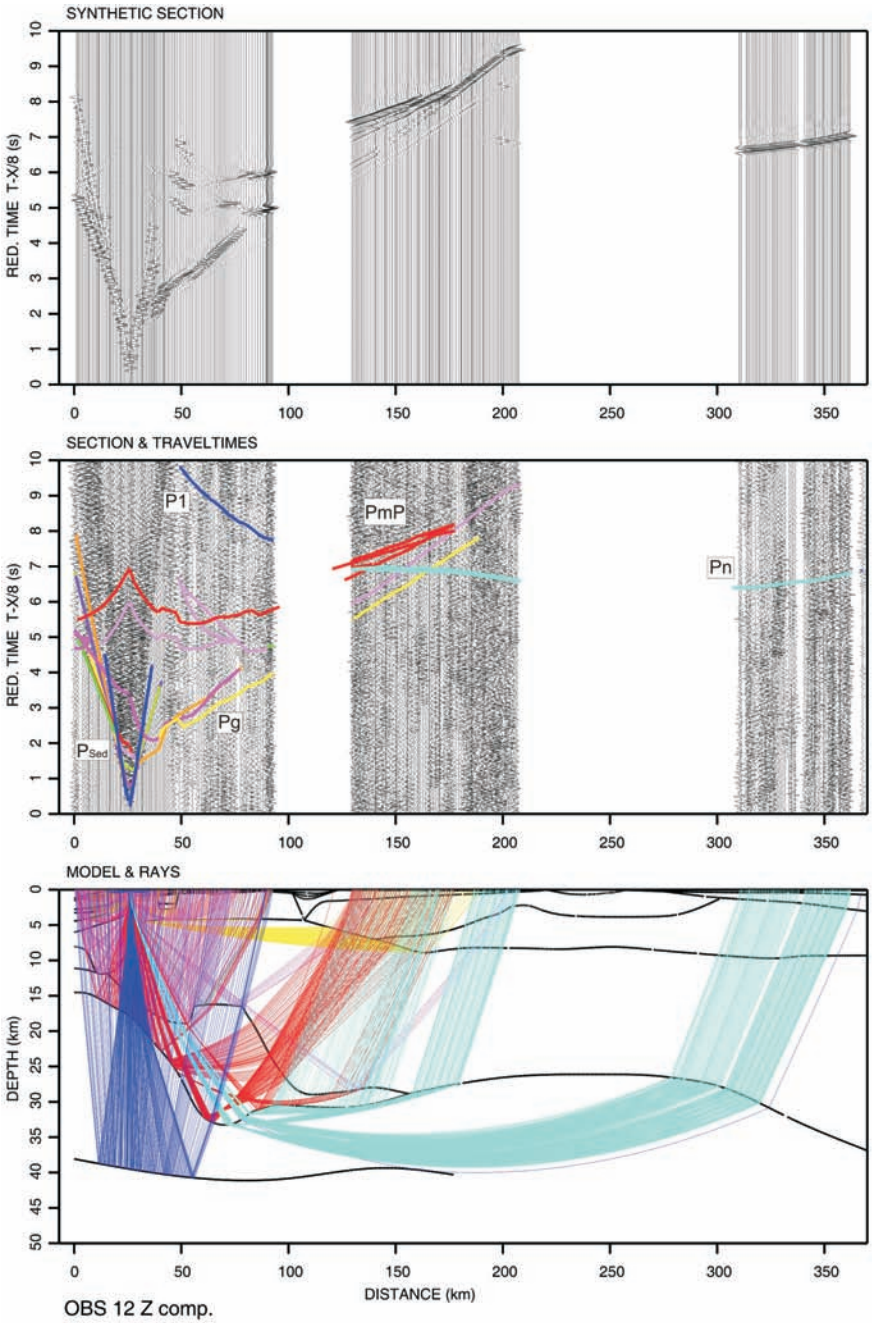

Fig. 2. Example of the 2-D seismic modelling diagram for the OBS 12. Normalized experimental record section in the middle with calculated travel times and ray diagram (below the record section), and synthetic record section (above the experimental record section). Filter 2–15 Hz. Note low sedimentary velocities. P_{Sed} – first arrivals of sedimentary P-waves; Pg – first arrivals of crustal P-waves; Pn – refracted P-waves beneath the Moho; PmP – Moho P-wave reflections; P1 – lower lihospheric reflections.

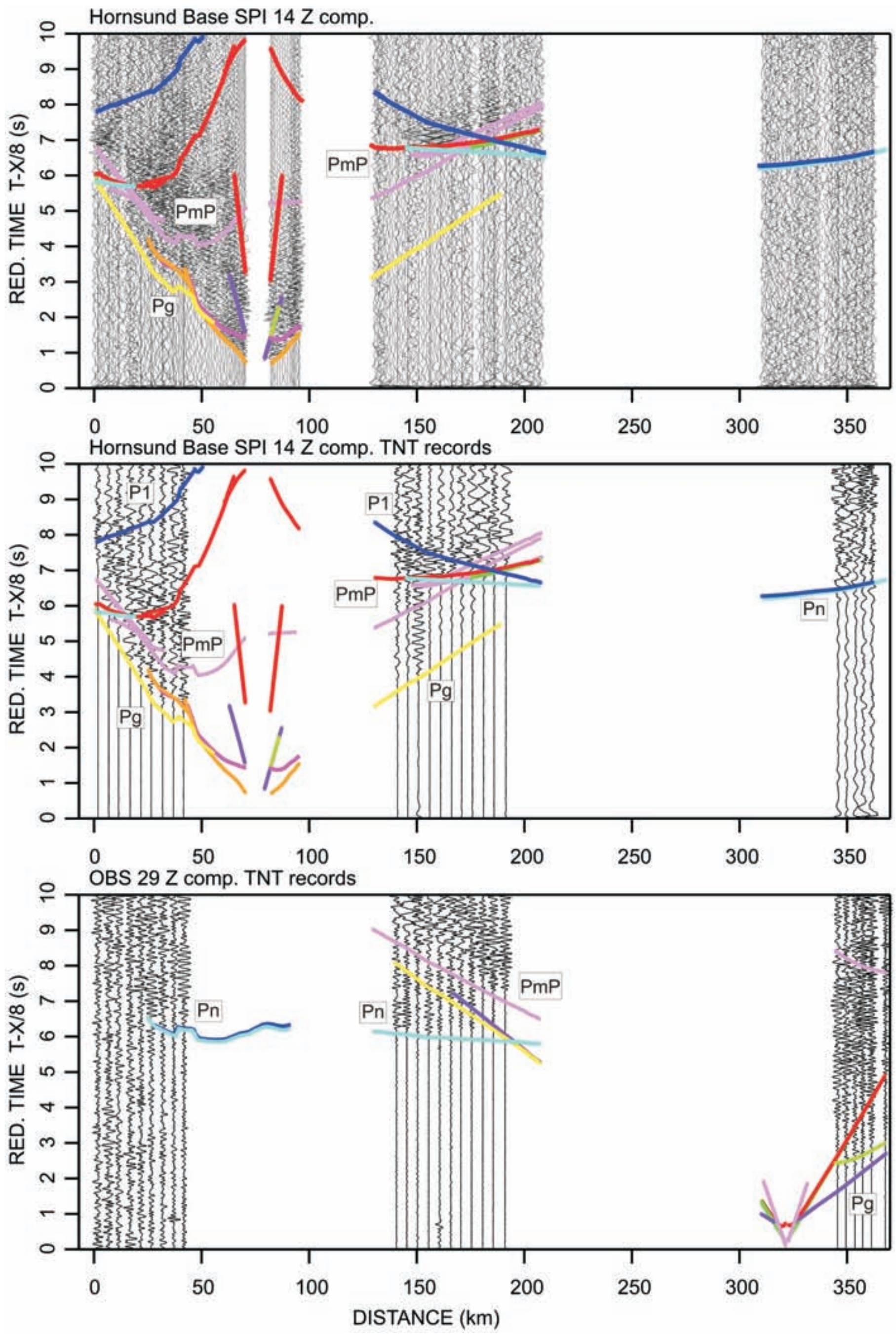

Fig. 3. Comparison of amplitude-normalized airgun seismic record section with calculated travel times from station 14 (top) and TNT seismic record sections with calculated travel times from station 14 (middle) and station 29 (bottom). Filter 2–15 Hz. Note distant Pg and Pn branches on the TNT record sections. Other descriptions as in Fig. 2.

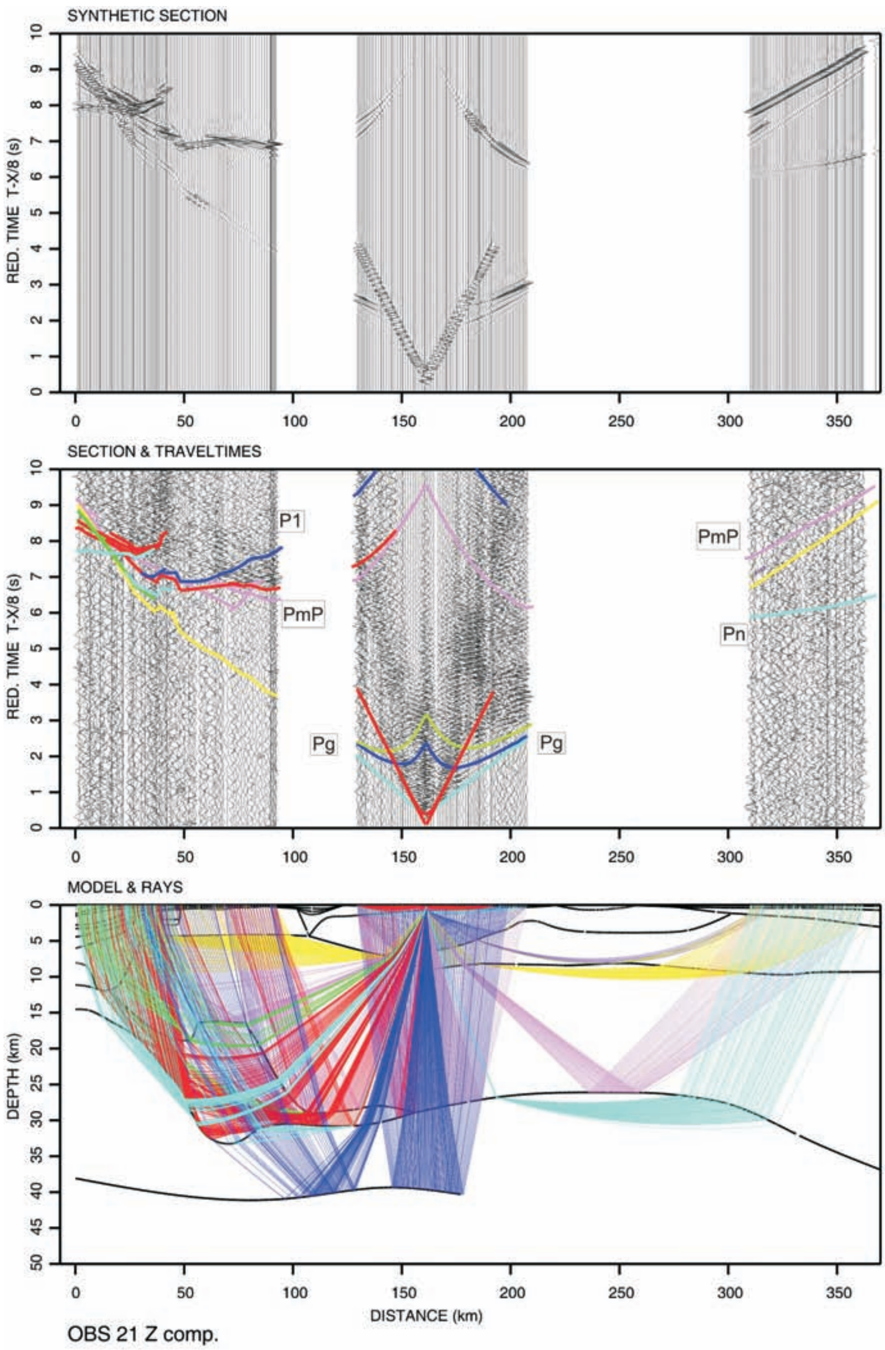

Fig. 4. Example of the 2-D seismic modelling diagram for the OBS 21. Normalized experimental record section in the middle with calculated travel times and ray diagram (below the record section), and synthetic record section (above the experimental record section). Filter 2–15 Hz. Other descriptions as in Fig. 3.

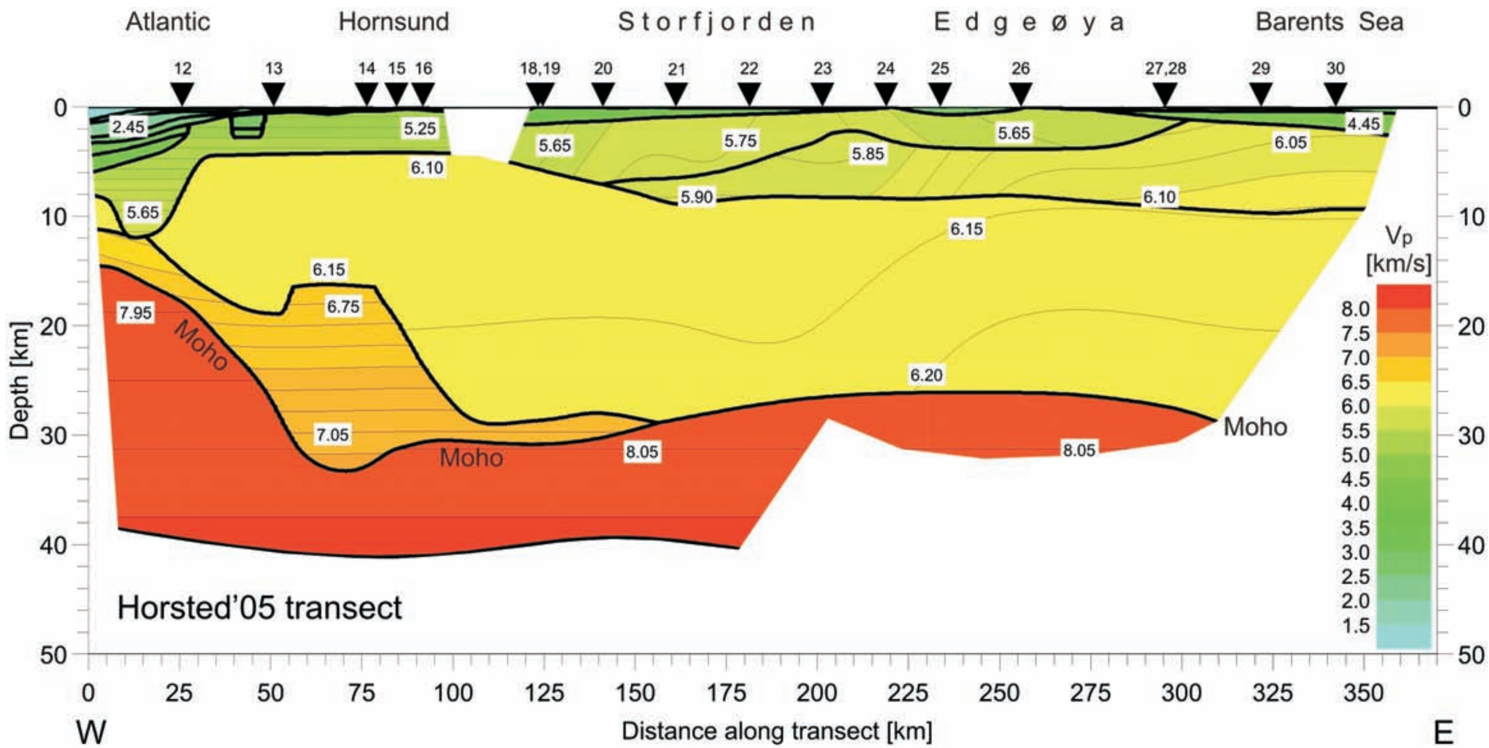

Fig. 5. Two-dimensional seismic P-wave velocity model along the Horsted'05 transect developed by ray-tracing techniques. Black lines represent seismic discontinuities (boundaries); shades represent the distribution of the P-wave velocity limited to the area of rays coverage; numbers in the model – P-wave velocities in km/s. Triangles show location of the stations. Station 14 is located in the Hornsund Polish Polar Base. Vertical exaggeration is 3:1.

obtained up to distances of 150 km at land stations and OBSs. TNT explosions were recorded even up to distances of 300 km (Fig. 3). There is very good signal to noise ratio in the TNT record section enabling very precise identification of the first arrivals at long distances (*e.g.* Fig. 3, 350 km, Pn).

First arrivals at close distances from the stations at the western part of the transect are characterized by the lower apparent seismic velocities than those from the central and eastern parts. There is the difference of almost 1 km/s. The western most part is covered by thick sediments, in comparison to other part of the transect, characterized by low apparent velocity (Fig. 2). The Moho reflections (PmP) are earlier and more complicated in the western part of the transect than in the central or eastern part (*e.g.* Figs 2–4). There are low velocities in the overcritical branches in the eastern and central part of the transect. It implies velocities lower than 6.5 km/s in the lower crust.

Seismic model

We have used SEIS83 ray tracing software package (Červený *et al.* 1977; Červený and Pšenčík 1983) with the graphical interface MODEL (Komminaho 1993) to model the 2-D crustal structure, and the ZPLOT software (Zelt 1994) for comparison of the observed and calculated data.

Results from the previous air-gun experiments from vicinity of the study area were used to build initial model. There are: OBS-98 Project (Ljones *et al.* 2004) at the western limit of transect studied and Profile 10 (Breivik *et al.* 2005) in the central part (Fig. 1). Such a model was then tested and modified according to our data.

The 2-D seismic model along the Horsted'05 transect (Fig. 5) is complicated in the western part, which is located east of the eastern escarpment of the Knipovich Ridge. Sedimentary layers with P-wave velocity in order of 2.5 km/s reach the depth of 5 km. The upper crustal layer with the velocities of 5.25–5.65 km/s dips down to 12 km to the lower crustal layer at 10 km of the model. Eastward, it shallows to 4 km. Below it there exists the middle crustal layer with velocities of 6.1–6.2 km/s which further to the east becomes the lower crustal layer in a two-layered crust. The high velocity layer in the western part (the P-wave velocity of 6.7–7.1 km/s) varies in thickness from 2–4 km on its ends to 17 km around the distance of 70 km along the transect. The Moho shallows up to 14 km in this area, while beneath the Hornsund it dips down to 33 km. The eastern part of the model is rather simple, with two layers in the upper crust (P-wave velocity in the order of 5.6–6.1 km/s) and the layer characterized by the P-wave velocity of 6.1–6.2 km/s in the lower crust. There is no P-wave velocity higher than 6.5 km/s in the continental crust in Storfjorden and Edgeøya area. The Moho depth varies between 26 and 31 km, shallowing beneath the central part of the continental crust. The

P-wave velocity below Moho interface is generally 8.05 km/s lowering to 7.95 km/s in the western part of the model.

Additionally, an upper mantle reflector was modelled in the western part of the transect at the depth of about 40 km.

Summary

A dense system of airgun shots with additional chemical explosions allows to model very accurately seismic crustal structure along the 370 km long transect. This paper focuses purely on the seismic modelling. We have obtained detailed seismic P-wave velocity model down to 40 km. It provides an important base for tectonic and geological interpretation. This interpretation will be the subject of the next study.

There are some differences between results of previous seismic investigation in this area. Profile OBS-98 generally coincides with the Horsted'05 transect but its middle crust is characterized by higher and more complicated P-wave velocity field. Profile 10, located a bit south of the transect Horsted'05, is characterized by an undulating basement and higher P-wave velocity gradient in the lower crust. The Moho boundary is of about 1–2 km deeper than in this study.

Complicated crustal structure of the western part of the transect must be connected with active rifting processes in the Knipovich Ridge. The high velocity layer in this part (P-wave velocity ~7 km/s) could be connected with the rifting, or it could be normal continental lower crustal layer deformed during tectonic history of western Spitsbergen. There is no explanation at this time for so low (6.2 km/s) P-wave velocity in the lower crustal layer in the eastern part of the transect.

There is rather difficult to find a prolongation of the Billefjorden Fault Zone but it could be localized in the vicinity of 115 km of the model in the upper crustal layer and deeper coinciding with the eastern limit of the high velocity lower crustal layer. There is no sign of the fault in the middle crustal layer. It is interesting that there are no upper mantle reflections in the eastern part of the transect.

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