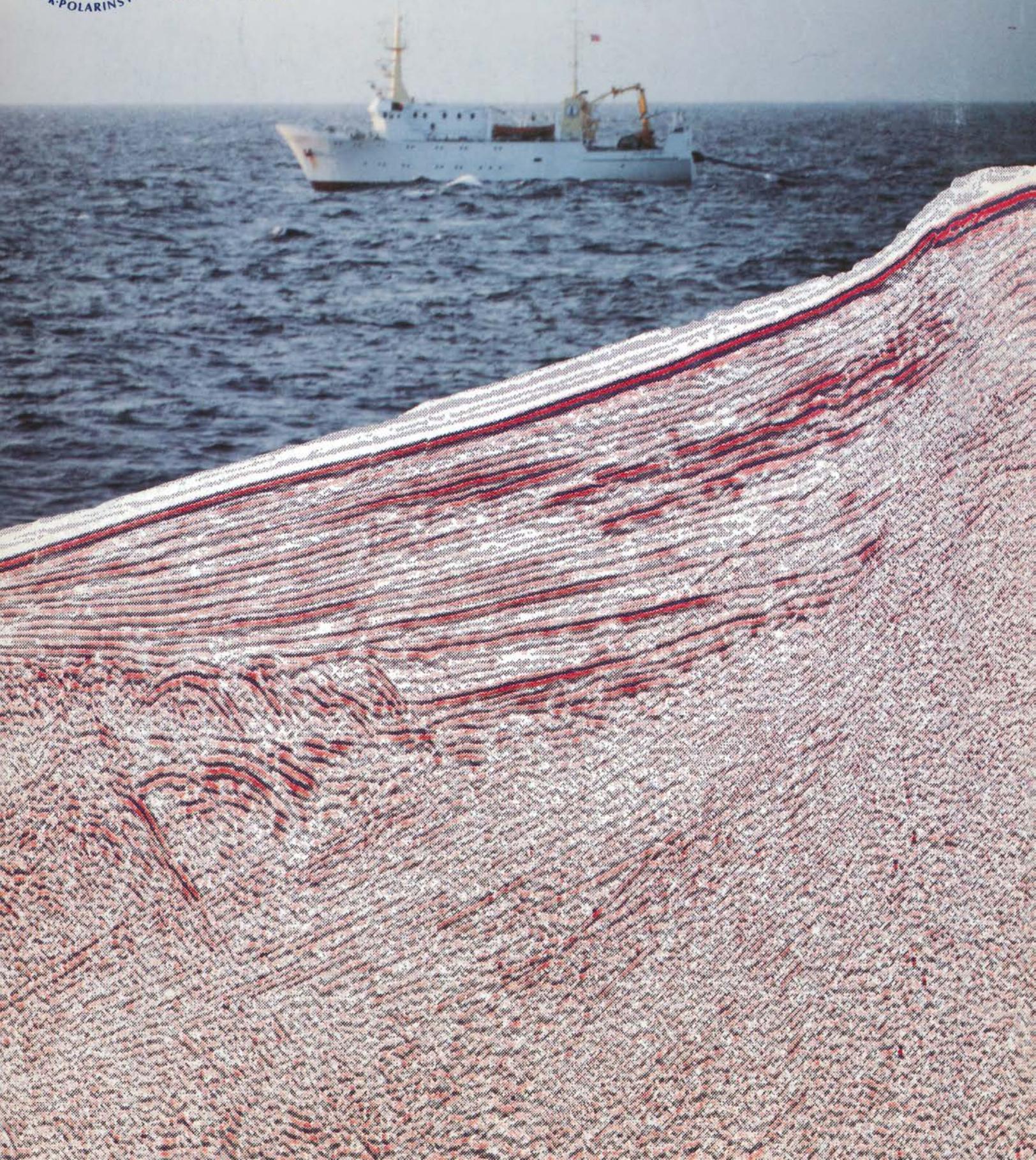


OLA EIKEN (EDITOR)

Seismic Atlas of Western Svalbard



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SEISMIC ATLAS OF WESTERN SVALBARD

A selection of regional seismic transects

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Preface

After fifteen years of geophysical work in the Svalbard area, a large set of seismic reflection profiles were at hand at the Institute of Solid Earth Physics, University of Bergen. Other important subsets of seismic data have been acquired by a number of institutions and oil companies, and there is clearly a need to make the seismic database more easily available for the geoscientific community.

We sincerely appreciate the positive attitude of all institutions we have approached for release of data. We thank Bundesanstalt für Geowissenschaften und Rohstoffe in Hannover, Norwegian Petroleum Directorate in Stavanger, Norwegian Polar Research Institute in Oslo, Laboratory of Regional Geodynamics in Moscow, IKU Petroleum Research in Trondheim, Statoil in Harstad, Norsk Hydro in Bergen and Harstad, British Petroleum in Stavanger, Mobil Exploration in Stavanger, Barentz Petroleum in Tromsø and Nopec at Nærnes for their cooperation and contributions. Data acquisition by University of Bergen during 1976–1981 was carried out under the direction of Eirik Sundvor. Compilation of this atlas was made possible by financial support from Statoil and Norsk Hydro, and the Norwegian Polar Institute carried through the publication process.

I thank the authors of the different chapters for their contributions. The manuscript was reviewed by Prof. Arild Andresen, Dr. Jan Inge Faleide, Prof. Yngve Kristoffersen and Prof. Ron Steel. Yngve Kristoffersen provided the encouragement to undertake the effort and gave numerous helpful suggestions during the work. The staff at Institute of Solid Earth Physics, University of Bergen gave unfailing practical support, and colleagues at Statoil's Kristiansund office have shown patience during the last phase of my work. I hope the atlas will stimulate the use of seismic data for geologists working in the Svalbard area.

Bergen, February 1994

Ola Eiken

Chapter 1

Introduction

1.1 History and Geography

The Svalbard archipelago is situated between 74° and 81° North and between 10° and 35° East. When Willem Barents first sighted land on the west coast of the main island, 17 June 1596, he named it Spitzbergen, inspired by the alpine landscape that met him. The highest peaks reach 1717 m above sea level¹ Fjords cut deeply into the main island, where they continue onshore as broad U-shaped valleys with seasonal rivers. Mountains in the eastern islands are commonly flat-topped.

The name Svalbard has been found in Icelandic annals back to the year 1194. The Norsemen mentioned land "four days of sailing north of Iceland". After Willem Barents re-discovered Svalbard, a short and intense period of a few decades of seasonal whaling by mainly Dutch and British, but also Danish, French and Spanish vessels followed. In the eighteenth and early nineteenth centuries Russian trappers spent all the year on the islands, while Norwegians increased their hunting and fishing activities from the late 1800's. Regular coal mining and settlements started early in this century and has continued up to now with small Norwegian and Russian communities in Longyearbyen, Barentsburg and Pyrami-

den. The treaty awarding Norway sovereignty of the archipelago was agreed on in 1925.

The shelf widths west and north of Spitsbergen vary from 20 to 100 km, as shown in Fig. 1². The submarine continuations of the fjords extend to the shelf edge as major troughs with water depths of 200–400 meter. Water depths on the shallow bank areas between the troughs are less than 100 meter. The continental slopes are between 1.5° and 5° steep. Prominent bathymetric features in the deep sea north and west of Svalbard are: Yermak Plateau, Knipovich Ridge, Hovgård Ridge and Molloy Ridge (Fig. 1).

About 60 % of the land area is covered by glaciers. Average air temperatures in Isfjorden are $+5^\circ C$ in July and $-12^\circ C$ in January, and precipitation is low. This relative mild climate (compared with the high altitude) is due to the warm West Spitsbergen Current (e.g. Johannessen 1986). This current also leads to favourable ice-conditions (indicated in Fig. 1) with open water along the western coast of Spitsbergen most of the year.

¹Topographic maps at scale 1:100,000 or larger covering the islands have been published by the Norwegian Polar Institute.

²Bathymetric maps have been published by Perry et al. (1980, 1986), Kristoffersen et al. (1988) and Cherkis et al. (1990).

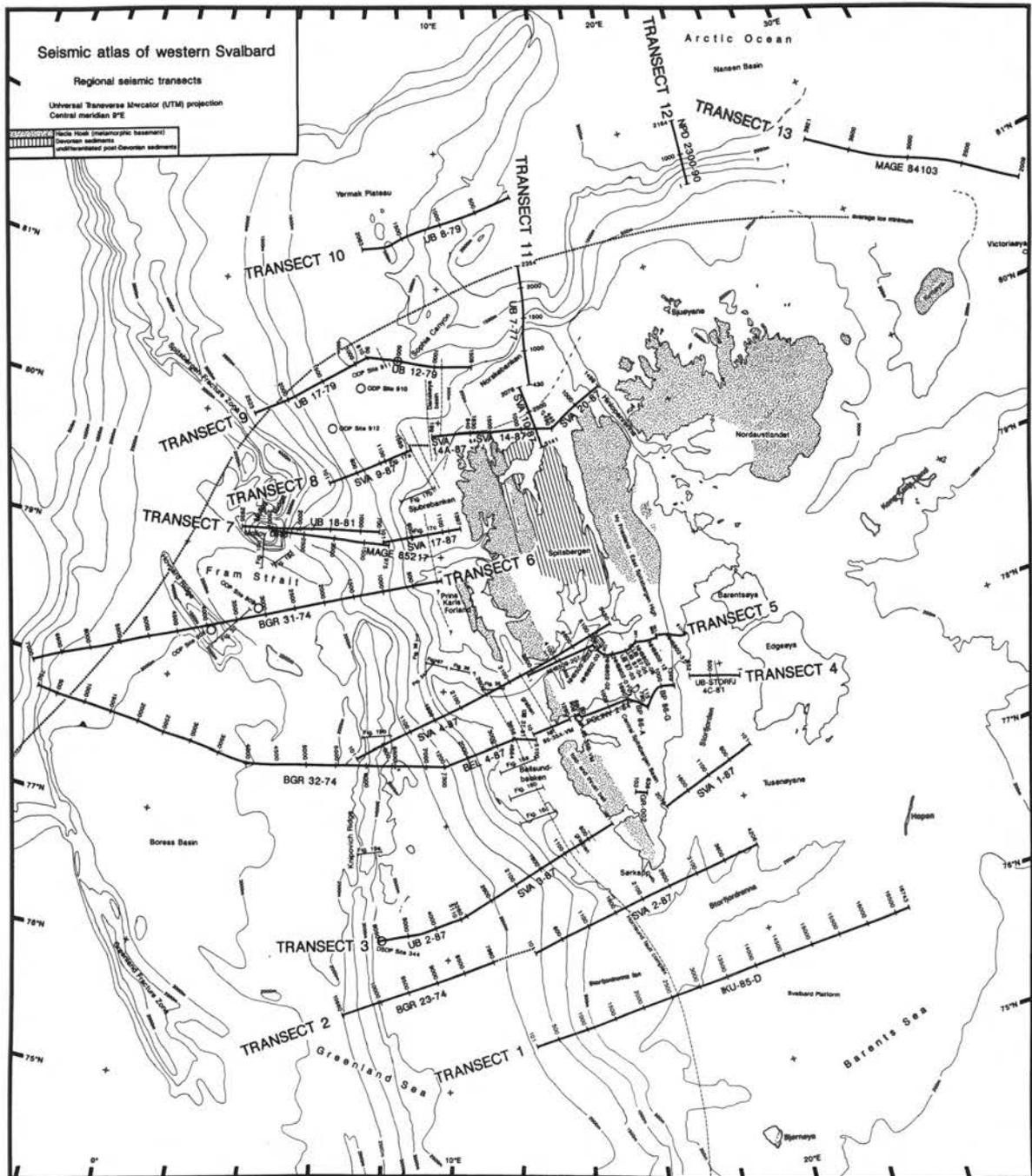


Figure 1: Simplified map of Svalbard with location of seismic transects and close-ups shown in the atlas. Bathymetry after Sundvor et al. (1982b), Perry et al. (1986), Kristoffersen et al. (1988) and Cherkis et al. (1990). Surface geology after Winsnes (1988) and average minimum ice-limit after Vinje (1985). An enlarged version of the map, at scale 1:2,000,000, is enclosed.

1.2 Geology

Investigations started with the expedition of the Norwegian geologist B.M. Keilhaug in 1827. Later in the 1800's Swedish expeditions headed by among others G. Torell and A.E. Nordenskiöld did pioneering work. From Norway, G. Isachsen started systematic research in 1906. This work was followed up by A. Hoel, and together they founded Norges Svalbard- og Ishavs-undersøkelser, which later became the Norwegian Polar Institute. In the first decades after second world war, a British group headed by W.B. Harland did extensive field work and published numerous papers (see list of references). In recent years, international academic studies have increased both in number of people and nations, and include Norwegian, Swedish, Danish, Russian, British, German, French, Polish and American geologists. A major effort in sedimentology by the universities in Bergen and Oslo from 1975 to 1985 resulted in some 35 student theses from the area, summarized by Steel and Worsley (1984). Oil and gas exploration since the 1960's has not resulted in any discoveries, but has contributed to our common knowledge of the geology. At present, economic exploitation is limited to coal mining of Paleocene and Carboniferous seams around Isfjorden.

Precambrian through early Tertiary rocks and Quaternary glacial sediments are exposed on the islands³ (Fig. 1). Caledonian and Precambrian metamorphic and crystalline rocks dominate the alpine mountains along the western and northern coasts. Devonian sediments are present in N-S oriented grabens in northern Spitsbergen, with small remnants of Tertiary plateau lavas

³Four geological maps at scale 1:500,000 covering the archipelago have been published by the Norwegian Polar Research Institute (Flood, Nagy and Winsnes 1971, Winsnes and Worsley 1981, Hjelle and Lauritzen 1982 and Lauritzen and Ohta 1984). This institute has also published geological maps of selected areas at scale 1:100,000.

and Quaternary volcanoes. Late Paleozoic, Mesozoic and Tertiary sedimentary rocks are exposed to the south and east and form the Central Spitsbergen Basin. Reviews of Svalbard's geology include Orvin (1940), Hjelle (1993), Ohta et al. (1989) on Caledonian terranes, Harland (1979) on the major fault zones and Steel and Worsley (1984) on the post-Caledonian rock succession.

Hecla Hoek basement shows several phases of Caledonian deformation, with a N-S to NNW-SSE structural grain (e.g. Harland 1985, Ohta et al. 1989, Ohta 1992). After Devonian extension and compression had formed grabens and deformed its sediment infill (Ziegler 1988, Manby and Lyberis 1992), tectonic movements were modest during late Paleozoic and Mesozoic times. Early Tertiary compression or transpression formed the fold- and thrust-belt along the western coast of Spitsbergen. When seafloor spreading moved Lomonosov Ridge and Greenland away in early-middle Tertiary times, the northern passive and western sheared-passive continental margin formed (e.g. Eldholm et al. 1990b, Kristoffersen 1990b).

15 oil exploration wells have been drilled onshore to date, but only results from Grumant 1 in Colesbukta south of Isfjorden have been published (Skola et al. 1980). Sediments offshore have been sampled in a number of shallow cores and dredge hauls, one scientific well has been drilled on the lower slope west of Svalbard: Deep Sea Drilling Project Site 344 (Talwani, Udintsev et al. 1976, Fig. 1), and five scientific wells (Ocean Drilling Project) were drilled late 1993 in the Fram Strait (Fig. 1). Only preliminary descriptions of these latest well data were available when this atlas went into press (Thiede, Myhre et al. 1993).

The stratigraphic division for Svalbard is shown in Table 1. Also shown are various seismic stratigraphic divisions and ages proposed for sediments within the different basins along the Svalbard margin. The

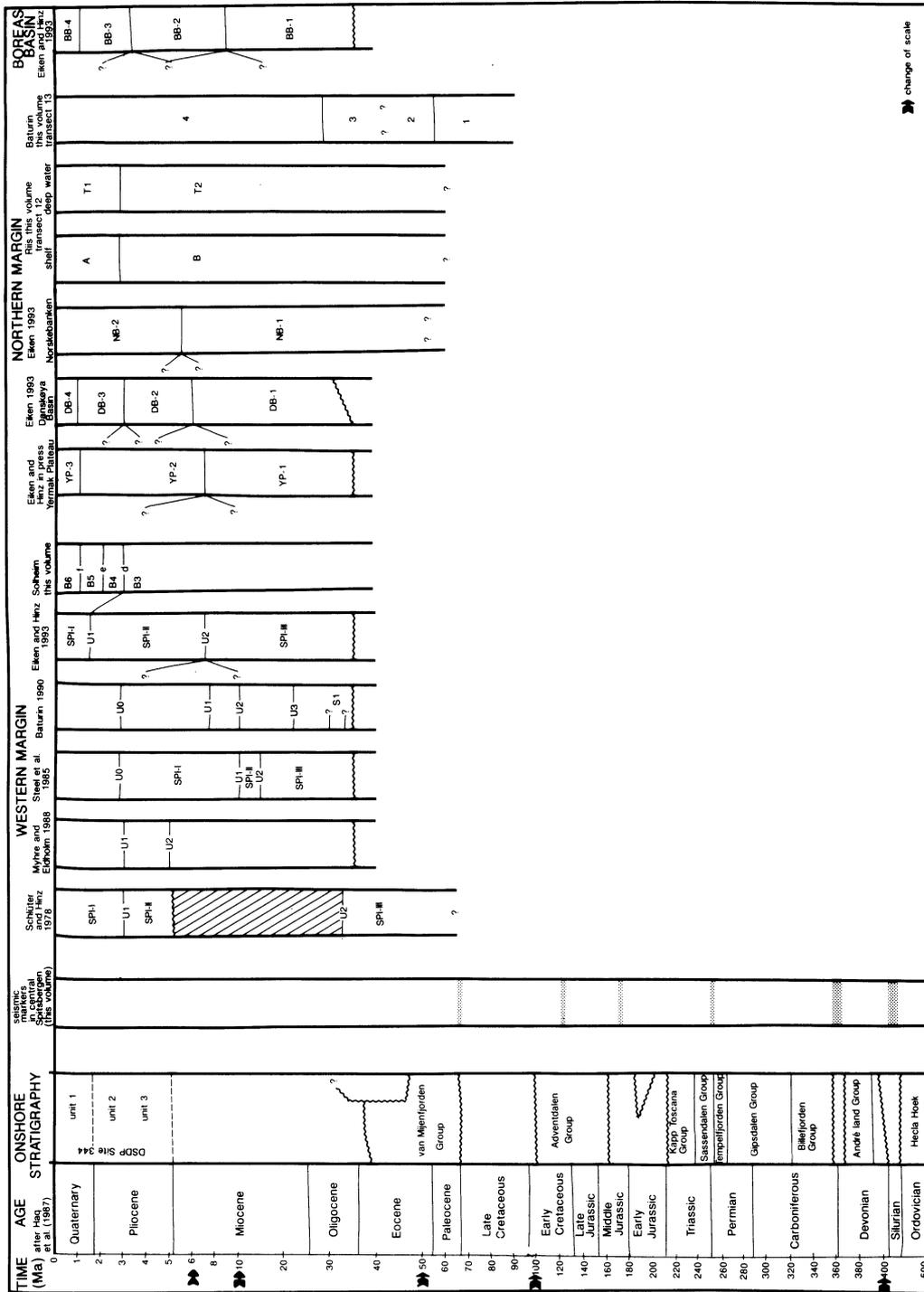


Table 1: The stratigraphic division of exposed Svalbard rocks and their assigned ages (to the left), and various seismic stratigraphic divisions for the sedimentary basins along the Svalbard margin.

recent Ocean Drilling Project wells, not included in this Table, generally support Pliocene and Pleistocene ages for most of the sediments within the seismic sequences YP-2/YP-3 and BB-3/BB-4.

1.3 Geophysical exploration

All seismic surveys in the Svalbard area known to us have been listed in Table 2. The first surveys were shot by Soviet crews onshore central Spitsbergen, as well as in the west-facing fjords and off the western coast in 1962–63, and little has been known about these surveys in the western world. Marine reconnaissance surveys with long, widely spaced lines were shot in the late 1960's and the 1970's by western academic institutions. The first single-channel analog data were recorded in the Greenland Sea and on the western margin with R/V "Vema", as part of Lamont-Doherty Geological Observatory's famous world-wide efforts to decipher the plate-tectonic history. Karl Hinz at Bundesanstalt für Geowissenschaften und Rohstoffe conducted the first large multichannel survey in 1974. University of Bergen has carried out extensive marine reflection surveys during 1976–

1981 on the western and northern continental margins, under the direction of Eirik Sundvor. From 1982 to 1988 Soviet offshore surveys were run by Marine Arctic Prospecting Geophysical Expedition. All these programs focused mainly on the structural framework of the continental margin and its relation to the plate-tectonic evolution of the area (e.g. Schlüter and Hinz 1978, Sundvor and Eldholm 1979, Baturin et al. 1990). The Norwegian Petroleum Directorate has acquired several regional seismic lines north, east and south of Spitsbergen during the last fifteen years as part of their large Barents Shelf mapping program. The offshore areas have not yet been opened for commercial exploration by Norwegian authorities, but onshore central Spitsbergen and in the fjords a number of oil companies have since 1984 acquired seismic lines through geophysical contractors. In the fjords, Nordisk Polarinvest, Norsk Hydro and Statoil have been most active, and onshore British Petroleum and Norsk Hydro have shot large surveys, the latter company using the snowstreamer technique. High-resolution test profiles for mapping of coal seams, initiated by Store Norske Spitsbergen Kullkompani, have been shot near Longyearbyen.

Year	Institution	Area	Remarks	Line-km	Reference
1962-64	Russian	central Spitsbergen			Potzeev 1965
1962-64	Russian	offshore west			Kogan and Milashin 1970
1966	LDGO	offshore west	analog		
1969	LDGO	offshore west	analog		
1970	CNEXO	offshore west	analog	ca. 1400	Malod and Mascle 1975
1972-74	Fina	offshore near coast			
1974	BGR	offshore west	normal	2100	Schlüter & Hinz 1978
1974	NPD	Storfjorden	normal	ca. 250	
1976	PAN	west of Isfjorden	(poor quality)	160	
1976	UB	offshore west	(weak source)	589	Sundvor et al. 1977
1977	UB	offshore north	(weak source)	795	Sundvor et al. 1978
1977	NPD	offshore southwest	normal	ca. 600	
1979	UB	offshore north	(weak source)	1397	Sundvor et al. 1982a,b
1979	UB	Agardhdalen, Sabine Land	test	1	Eiken 1981
1980	SNSK/UB	Platåfjellet, Nordenskiöld Land	high resolution	2	Bruland et al. 1982
1980	NPD	Storfjorden	normal	ca. 350	
1981	UB	offshore	normal	5000	
1981	NPD	offshore south	normal	ca. 200	
1982	MAGE	offshore southeast	normal	350	Baturin 1986
1982	NPD	Storfjorden	normal	ca. 350	
1984	SNSK/UB	Adventdalen	high resolution	5	Iversen 1986
1984	NPD	offshore north	normal	ca. 600	
1984	MAGE	offshore north	normal	450	Baturin 1987
1984	Statoil	Isfjorden	normal	284	
1984	NP	Van Mijenfjorden	normal	180	
1985	BP	Heer Land	(glaciers)	13	
1985	MAGE	offshore west and north	normal	1000	Baturin and Netchaev 1989
1985	NP	fjords and near coast	normal	1500	
1985	NH	fjords and near coast	normal	935	
1985	Statoil	fjords, near northern coast	normal	1120	
1985	NH	Adventdalen	test	4	
1986	BP	Heer Land	(glaciers)	325	
1986	Statoil	Grimfjellet	(glaciers)	55	Orheim et al. 1988
1986	Geoteam	Berzeliusdalen	normal	46	
1986	MAGE	offshore west	normal	1000	Baturin 1990
1987	UB/NH	Adventdalen	normal	40	Eiken et al. 1987
1987	MAGE	offshore west	normal	1000	
1987	UB/NPI	offshore	deep-penetration	2027	Austegard et al. 1987
1987	NPI/UB	off Bellsund	high resolution	733	Austegard et al. 1987
1987	UB	western slope	normal	1073	Austegard et al. 1987
1987	NH	Isfjorden	normal	300	Nøttvedt and Rasmussen 1988
1988	NH/UB	Sassendalen and surroundings	(snowstreamer)	300	
1988	BGR	Greenland Sea	normal	400	Hinz 1988
1988	MAGE	offshore west	normal	1000	Baturin et al. 1990
1988	GSN	offshore northwest	high resolution	2000	
1988	Statoil	Isfjorden	normal	742	
1989	NH	Reindalen and surroundings	(snowstreamer)	100	
1989	Petroarctic	Kvalvågen	(snowstreamer)	30	
1989	NPD	offshore northwest	normal	573	
1990	NPI	off Isfjorden and Bellsund	high resolution	3000	Solheim et al. 1991
1990	NPD	offshore northwest	normal	1473	
1991	NH	Nordenskiöld Land	(snowstreamer)	157	
1991	SNSK/NH/UB	Gruvefonna	high resolution	7	
1991	AWI	offshore north	icebreaker	ca. 500	Jokat et al. 1992
1992	Statoil	Sabine Land	(snowstreamer)	183	
1992	SNSK/NH/UB	Gruvefonna and surroundings	high resolution	17	

Abbreviations of institutions:

AWI: Alfred-Wegener-Institute, Bremerhafen
 BGR: Bundesanstalt für Geowissenschaften und Rohstoffe, Hannover
 BP: British Petroleum, Stavanger
 CNEXO :Centre National de l'Explorations Oceaniques, Brest
 GSN: Geological Survey of the Netherlands
 LDGO: Lamont-Doherty Geological Observatory, New York
 MAGE: Marine Arctic Prospecting Geophysical Expedition, Murmansk
 NH: Norsk Hydro, Harstad and Bergen
 NP: Nordisk Polarinvest, now Barentz Petroleum, Tromsø
 NPD: Norwegian Petroleum Directorate, Stavanger
 NPI: Norwegian Polar Research Institute, Oslo
 PAN : Polish Academy of Sciences
 SNSK: Store Norske Spitsbergen Kullkompani, Longyearbyen
 UB: University of Bergen

Table 2: Seismic reflection surveys in western Svalbard known to us. Some of the offshore surveys (particularly NPD-surveys) extend farther to the south and east of the area covered by this atlas (which is limited by Storfjordrenna-Edgeøya-Kvitøya).

Chapter 2

Descriptions of seismic transects

The area covered by this atlas is limited to the northwest-northeast by ice, deep water and lack of seismic data (Fig. 1), while Storfjordrenna–Edgeøya–Kvitøya define the southern and eastern limit towards the Barents Sea. Transect 1, south of this limit, offers a link to the Barents Shelf.

Thirteen seismic transects have been compiled and are displayed at horizontal scale 1:200,000 and time scale 1.5 cm/second. The seismic transects are enclosed as loose leaf displays. The plotting parameters are a compromise between high resolution and overall clarity of the section; variable area shading with 40 % bias, and amplitude normalization in about 1 second sliding windows. The transects are composites of lines from several surveys with different acquisition and processing parameters. Small mismatches are thus unavoidable. The data are mostly in the form of stacked profiles, but for some sections without penetration into igneous or metamorphic basement, migrated versions of the lines are used.

Line drawing interpretations and brief descriptions of the transects are presented in the following sections. Throughout the text we refer observed (two-way) reflection times as just times. Selected topics are treated more thoroughly in chapter 3, and some aspects of the seismic method relevant to the Svalbard environment are described in chapter 4.

2.1 The continental margin between Spitsbergen and Bjørnøya

by Steinar Thor Gudlaugsson¹ and Jan Inge Faleide¹

A first order division of the margin into three structural zones has been established on the basis of seismic reflection data (Riis et al. 1986, Gudlaugsson et al. 1987, Myhre and Eldholm 1988, Riis and Vollset 1988) and expanded spread profiles (ESP)² (Kitterød 1986): (1) The Svalbard Platform, where a basement of probable Caledonian origin is covered by an upper Paleozoic - Mesozoic platform sequence; (2) the Hornsund fault complex³, a downfaulted marginal terrace of continental rocks; and (3) oceanic crust in the Greenland Sea. The margin is covered by a thick late Cenozoic sedimentary wedge derived from the Barents Shelf (Schlüter and Hinz 1978, Eiken and Austegard 1987, Myhre and Eldholm

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²An outline of this two-ship multichannel marine acquisition technique for deep-crustal studies has been described by Stoffa and Buhl (1979).

³This zone has been termed Hornsund fault (Sundvor and Eldholm 1979), Hornsund fault zone (Myhre et al. 1982), Hornsund lineament (Eiken and Austegard 1987) and Hornsund fault complex (between 73°30'N and 77°N, Gabrielsen et al. 1990). Here we use the term Hornsund fault complex (see also chapter 3.1).

1988).

Fig. 2 shows major crustal units inferred from the transect superimposed on the ESP-derived velocity field. Eight units are recognized.

In the eastern part of the transect, highly reflective crust (unit 7) characterized by hyperbolic reflections overlies a reflection-free mantle (unit 8). Strong reflections coinciding with a sharp decrease in reflectivity define a stepped Moho rising from 10.5 seconds (32 km) at the eastern end of the transect to 9 seconds (27 km) at shotpoint 3100. Wide angle reflections from the Moho are also observed on ESP-profiles 5 and 134, but refracted arrivals are lacking. Thus, the identification of the base of the crust is based entirely on reflected arrivals. The hyperbolic reflection signature indicates complex small-scale structuring, which we interpret as pervasive folding and faulting of the crystalline crust. On a larger scale, we recognize two subunits separated by a westward dipping structural discontinuity. The western subunit (7A) appears to have been thrust over the eastern subunit (7B), forming a large ramp anticline. The irregular form of the basement surface bounding the western subunit above indicates extensional movements on westward dipping normal faults postdating the inferred compressional event.

Unit 6 is characterized by indistinct, mildly disturbed layering, which is infilling small half-grabens at the crest of the ramp anticline and a wider and deeper basin to the east. It represents the first sediments deposited after the compressional event. Units 4 and 5 are platform sediments characterized by regular sub-horizontal layering. The boundary between these units is marked by a change from a thin eastward tapering wedge at the top of unit 5 to westward directed onlap in the basal part of unit 4.

The seismic reflection character, seismic velocities and regional stratigraphic consi-

derations lead us to interpret units 4 and 5 as being Triassic and Permo-Carboniferous in age, respectively. The boundary between them correlates approximately with the top Permian reflector on the Svalbard Platform (Riis et al. 1986). This correlation is supported by the seismic refraction velocities which increase from 4.4 km/s in the upper part of unit 4 to just above 6 km/s within unit 5. The Upper Carboniferous - Permian carbonate platform of the Barents Shelf and Svalbard is generally associated with seismic refraction velocities close to 6 km/s (Eiken 1985, S.Sanner pers. comm., see also chapter 4.2).

The Caledonian Orogeny was the most important compressional tectonic phase affecting Svalbard and Bjørnøya in Paleozoic times. Given the severity of the deformations observed within unit 7 in the seismic transect and its pre-Mesozoic origin, a Caledonian origin appears most likely. An alternative interpretation which relates the event to large-scale sinistral strike-slip movements postulated to have occurred in Late Devonian time (Harland 1969, Ziegler 1988) appears unlikely because of the large thickness of Permo-Carboniferous strata that this interpretation implies (the maximum thickness of units 5 and 6 combined is approximately 12 km). Consequently, we interpret unit 6 as Devonian molasse, possibly including Lower Carboniferous deposits at the top. Linear NNW-striking magnetic anomalies characterize the western Svalbard Platform from Bjørnøya to Spitsbergen (Skilbrei et al. 1990). This implies that the Caledonian basement terrain proposed here to underly the western Svalbard Platform has a NNW-oriented grain.

In the western part of the transect (west of sp. 3050), a downfaulted terrace of continental rocks (unit 3) abuts oceanic crust (unit 2) and is overlain by a wedge of Cenozoic sediments (unit 1). In the terrace region, which corresponds to the Hornsund fault complex, we can identify several nor-

mal faults and a single faulted horizon stepping down to the west. The poor data quality in this area does not allow for detailed definition of structure or stratigraphy. Reflections from the base of the crust are lost, but this part of the transect is assumed to cross thinned continental crust.

Oceanic crust extends from the Greenland Sea to approximately sp. 1900 just west of the continental terrace. From the western end of the transect to sp. 350, a strong reflector marks the top of oceanic crust. Oceanic basement between sp. 350 and 1400 is defined by a less reflective, more irregular surface. Due to deteriorating data quality east of this point, top basement is lost. Its eastward continuation in Fig. 2.1 to sp. 1900 is based on information from crossing reflection profiles. Refracted Moho-arrivals observed on ESP-profiles 1, 2 and 3 (Kitterød 1986, Myhre and Eldholm 1988) are consistent with this interpretation, and yield an oceanic crustal thickness of 6 to 8 km.

ESP-profiles 2 and 3 record anomalously high velocities of more than 7 km/s in the lower 3–5 km of the oceanic crust (Kitterød 1986, Myhre and Eldholm 1988). An elongate positive gravity anomaly with a maximum amplitude of 142 mgal is present along the margin (Faleide et al. 1984). Gravity models (Myhre and Eldholm 1988, Breivik

1991) constrained by the seismic transect and ESP-profiles described here require anomalously high densities in a 90 km wide strip of oceanic crust west of the Hornsund fault complex.

Unit 1 comprises post-opening sediments prograding into the Greenland Sea. The base of subunit 1A corresponds to reflector U_2 of Schlüter and Hinz (1978). This sequence boundary can be followed towards the Knipovich Ridge where it overlaps oceanic basement not older than 5–6 Ma (Myhre and Eldholm 1988). The sediments of subunit 1A are therefore probably of Pliocene - Pleistocene age. Sequence boundary U_2 has also been tied by regional seismic lines (B.O. Hjelstuen and A. Fiedler, pers. comm.) to two wells (7117/9-1 and 7117/9-2) in the southwestern Barents Sea where it forms the base of fan deposits of glacial origin deposited in Late Pliocene and Pleistocene times (Eidvin and Riis 1989, Eidvin et al. in press). The transect shows a total thickness of 3 seconds for subunit 1A, indicating that an up to 4 km thick sedimentary wedge was built over a short time span, mainly due to glacial erosion of the Barents Shelf region. Subunit 1B was deposited over a much longer time interval with lower average sedimentation rates.

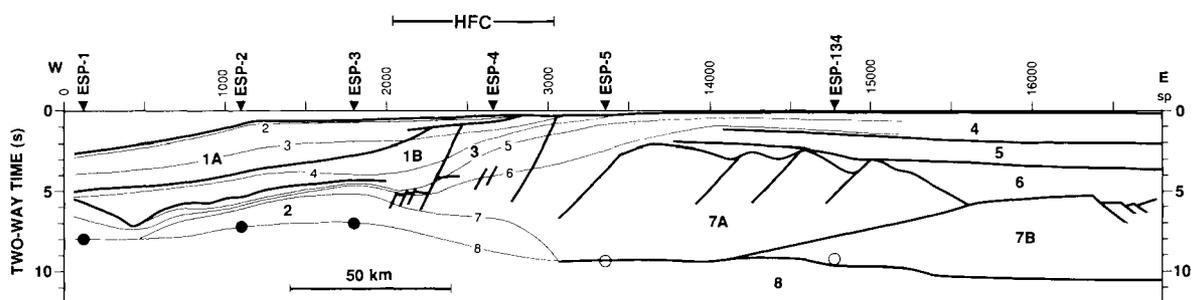


Figure 2: Line-drawing interpretation of transect 1. Major crustal units discussed in the text are numbered 1 - 8. Iso-velocity contours derived from ESPs are shown by thin lines and small numbers (contour interval 1 km/s). Filled and open circles show the position of Moho as interpreted from refracted and reflected ESP-arrivals, respectively. HFC: Hornsund fault complex.

2.2 Knipovich Ridge - Sørkapp - south of Tusenøyane

by Ola Eiken⁴

The transect runs from the Greenland Sea and across the continental margin just south of Spitsbergen with the southernmost exposures of the Tertiary fold- and thrust-belt. The Knipovich rift-valley and half-grabens on its western flank contain at least 1 second (0.8–1.0 km) of thick well-stratified sequences, while the eastern rift-valley flank consists of one major fault (Fig. 3). The thick deposits here have been attributed to heavy sediment supply from Svalbard and the Barents Shelf in the late Cenozoic (Vogt et al. 1978, Myhre and Eldholm 1988). Turbidity currents sourced by glacial sediments on the upper continental slope have probably reached the rift-valley, and sediment transport by bottom currents along the rift-valley has also been important (Eiken and Hinz in press). The sequences may consist of a mixture of these deposits, hemipelagic sediments and volcanic detritus.

Basement gets buried increasingly deeper eastwards at the continental rise. Top oceanic basement can be followed as a rough surface east to line SVA 2-87 sp. 600, where the sediment cover is more than 3 seconds (4–5 km) thick.

The sediments beneath the slope in this area and north to about 78°N can be divided into the three sequences SPI-I through SPI-III (Schlüter and Hinz 1978), separated by two unconformities U_1 and U_2 . This division has been adopted in a number of later papers (Steel et al. 1985, Eiken and Austegard 1987, Myhre and Eldholm 1988), where different ages and depositional environments have been suggested (Table

1, page 8). Sequence SPI-III has fairly parallel layering, fills in basement topography and is generally believed to consist of early post-rift sediments. Sequence SPI-II is characterized by abundant diffractions. Its thickness increases westwards from near zero beneath the upper slope on this transect as well as farther north (Fig. 31, page 50). Sequence SPI-II may have been formed by large-scale slumping (Schlüter and Hinz 1978, Myhre and Eldholm 1988) or bottom currents (Eiken and Hinz 1993). The uppermost sequence SPI-I is dominated by prograding subsequences and may consist of Plio-Pleistocene glacial marine sediments (Schlüter and Hinz 1978, Myhre and Eldholm 1988).

East of the shelf edge, total sediment thickness decreases rapidly, with erosional truncations near the seafloor, and with high-velocity (3–4 km/s) rocks present close to the seafloor east of sp. 1900. The volume of low-velocity sediments beneath the shelf is less here than at transect 1, due to the transect position on the flank of Storfjordrenna fan, a major depocenter of late Cenozoic sediments (Myhre and Eldholm 1988). The particularly well developed unconformity beneath sequence SPI-II is attributed to the line orientation oblique to Storfjordrenna fan. Beneath sp. 1950–2350, westward-dipping events are observed at 1–3 seconds. Hornsund fault complex is situated in this area, but it lacks the distinct signature seen farther north (transects 3 and 4). The fault complex makes an eastward bend south of Sørkapp (Fig. 1), and Riis and Vollset (1988) interpreted a transition from the fold- and thrust-belt in the north to the mainly extensional Knølegga fault farther south in this area.

Farther east the line crosses the Svalbard Platform, with high-amplitude seafloor multiples. Steeply dipping energy down to 4 seconds at sp. 2600 and a broad anticline within the uppermost 2–3 seconds at sp. 3300–3500 are noise; seafloor diffrac-

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tions and multiple reflections. Near horizontal reflections at 1.5–2 seconds in Storfjordrenna are probably from Permian and Carboniferous platform sediments. Highly reflective middle and lower crustal levels may correspond to the western part of unit 7 as described in transect 1 (and there interpreted as results of Caledonian deformation). Bright reflections and diffractions at 10–10 1/2 seconds form the base of the reflective crust and are most likely associated

with Moho. A slight westward dip of this reflective level (on the time-section) may indicate crustal thickening associated with the root zone of the fold- and thrust-belt. Gravity modelling supports this (Austegard unpublished results), and similar patterns are seen beneath the fold- and thrust-belt at transect 5. Thus, the fold- and thrust-belt seems to extend southward at least some distance south of Sørkapp.

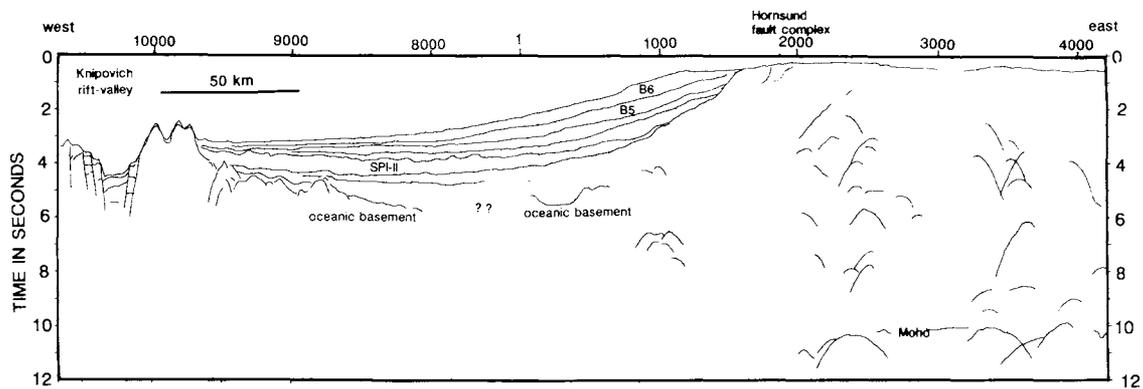


Figure 3: Line-drawing interpretation of transect 2.

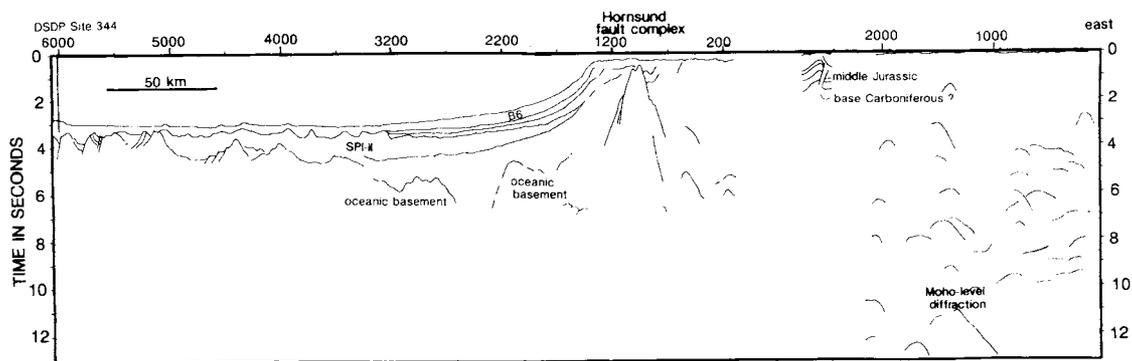


Figure 4: Line-drawing interpretation of transect 3.

2.3 Knipovich Ridge - Hornsund - Hornbreen - Storfjorden

by *Ola Eiken*

The transect has a well-tie in deep water, and is the southernmost of three transects crossing Spitsbergen. Oceanic basement shows a pattern of structures which resembles rotated fault blocks and suggests extensional tectonics within the crust beneath the continental slope and rise. Seismic sequence SPI-II (Fig. 4) exhibit particularly well developed diffraction patterns. Near the western end (sp. 6000 on line SVAL 2-87), the transect crosses Deep Sea Drilling Project site 344, which penetrated 377 m of glacial marine sediments and 10 m into a sill (Talwani, Udintsev et al. 1976). Eiken and Hinz (1993) correlated the rocks encountered in the well with regional seismic data and concluded that the lowermost sediments in the borehole, probably of ages around the Pliocene/Miocene boundary (Talwani, Udintsev et al. 1976, Warnke and Hansen 1977), correspond to the upper part of sequence SPI-II.

The sediment thickness increases towards the shelf edge, and top basement reflections disappear east of sp. 2600 on line SVA 3-87 below about 3 seconds (about 4 km) of sediments, but is locally seen also farther east, beneath the slope possibly at around 6.5 seconds (10–12 km beneath the seafloor), indicating very thick sediment accumulations in this area. Sundvor and Austegard (1990) interpreted these reflectors as representing the top of oceanic basement. Nearly 10 km thick sediments have been suggested here from gravity modelling (Austegard and Sundvor 1991), and Aalerud (1986) explained this by a decoupling of the oceanic and continental lithosphere during sediment loading. Hornsund fault complex is seen as a steep fault bounding a basement high

(Fig. 4). Farther east is a graben containing Tertiary (Eiken and Austegard 1987, 1989) or Mesozoic (Townsend and Mann 1989) sediments.

Austegard and Sundvor (1991) suggested on basis of gravity modelling an abrupt eastward thickening of the crust (from about 18 km to about 33 km) on the landward side of Hornsund fault complex. Aalerud (1986) modelled a profile just north of transect 3 (NPD 7700-77) with major crustal thickening at Hornsund fault complex, and with a dense peridotitic body landward of the fault complex.

There is a data gap across the fold- and thrust-belt in Hornsund, where exposed Mesozoic and older rocks are strongly deformed (e.g. Dallmann et al. 1992, Dallmann 1992). The land-seismic profile in the Grimfjellet area, where fairly flat-lying Cretaceous strata outcrops, shows an asymmetric anticline within the Mesozoic and upper Paleozoic strata. A reverse fault zone on the eastern flank of the anticline terminates within the Cretaceous sediments (Fig. 4). The anticline parallels the trend of the fold- and thrust-belt and has probably been formed during Tertiary compressional or transpressional events (Orheim et al. 1988).

The deep seismic profile across Storfjorden has poor signal-to-noise ratios in the upper 3–4 seconds and reflections from the Mesozoic and upper Paleozoic strata cannot be discerned here. Seismic lines with a shallower focus (Eiken 1985) have revealed flat-laying reflectors at 1.0–1.5 seconds (2–3 km depth) of probable Permian-Carboniferous age in this area. Several hyperbolic events are observed from 3–4 seconds and down to 10–11 seconds. Most of the hyperbolas at mid-crust levels are too broad to be in-line point diffractions, and they may arise from undulating reflector surfaces or from diffractors striking oblique to the line. The high amplitudes (relative to the noise) and abundant number of hyperbolas suggest that the

crust is significantly heterogeneous. Events are barely seen beneath 10–11 seconds, and we correlate this level to the Moho transition. A slight Moho dip to the southwest ($1^\circ - 1,5^\circ$) is then suggested.

Around sp. 1300 distinct diffraction tails dipping both to the east and the west suggest an anomaly at Moho level, perhaps a vertical offset. This may correlate with similar diffractions at Moho-level on transect 2 (sp. 3600) and represent a NNW-SSE trending crustal scale lineament, parallel to and in the extension of the large fault zones on Spitsbergen.

2.4 Boreas Basin - Bellsund - Van Mijenfjorden - Heer Land - Stor-fjorden

by Arvid Nøttvedt⁵, Ola Eiken and Peter Steven Midbøe⁶

The 600 km long transect (the longest in this atlas) crosses the entire Greenland Sea, Svalbard's western continental margin and the Central Spitsbergen Basin, and images a fairly complete succession of Svalbard rocks.

Sediment thicknesses at the Greenland lower slope are about 1.0–1.2 seconds (about 1 km), considerably less than on the western Svalbard margin. The upper strata are truncated at the seafloor, which suggest recent submarine erosion. Eiken and Hinz (1993) explained this by slope-parallel bottom currents. Top basement is irregular and highly reflective and basement may be of volcanic nature (see also transect 6).

Eiken and Hinz (1993) divided the 3 seconds (3–5 km) thick sediments in Boreas Basin into four seismic sequences, labelled BB-1 through BB-4 (Table 1 and Fig. 5). Basement becomes shallower and appears

faulted towards Knipovich Ridge. West of the rift-valley (line BGR 32-74 sp. 5000–5500), rotated fault blocks with thick sediments above (Fig. 5) are evidence of a late phase of rifting in this area. The flanks of Knipovich rift-valley show down-faulted basement blocks containing up to 0.5 seconds of sediments which are faulted as well. Within the rift-valley, the transect crosses an axial mountain (volcano), and abundant seafloor diffractions are observed. The sediment thickness increases rapidly towards the Svalbard continental margin and top basement is not well defined beneath the slope. The uppermost sequence SPI-I (Schlüter and Hinz 1978) is thicker here than farther south, due to the proximity to a depocenter west of Bellsund. Eiken and Austegard (1987) observed an unconformity U_o within sequence SPI-I beneath the shelf, and proposed that it represents the onset of glacial erosion. In addition, two shallower unconformities can be observed on high-resolution profiles in the Bellsund-banken area, and allow a division of the sediments into six seismic sequences B1–B6 (Table 1 and chapter 3.4).

The westward dip of the strata beneath U_o increases east to line BEL 4-87 sp. 3600. Farther east, structures within the Hornsund fault complex were interpreted by Eiken and Austegard (1987) as two east-verging faults bounding a rotated fault block, with strata dipping as steep as $40^\circ - 60^\circ$ to the west (line UB 37-81 sp. 4700–4900). An alternative interpretation of the steep dips may be offline diffractions from a shallow (0.3–0.4 seconds) basement horst. The top of the feature is masked by water bottom multiples, but is tentatively proposed to dip gently westwards down to about 1 second (Fig. 5).

East of this complicated area (east of line UB 37-81 sp. 4900), we observe a graben, about 28 km across and about 2 seconds (about 4 km) deep on the profile. This graben is likely to be of Ter-

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tiary age (Eiken and Austegard 1987), similar to the Forlandsundet Graben (Steel et al. 1985, Gabrielsen et al. 1992, Kleinspehn and Teyssier 1992). The graben appears to be nearly symmetrical with a dome in the middle. Within the lowermost second of the graben fill, internal reflectors parallel the basal reflector, which suggest initial subsidence along near vertical boundary faults. Higher up in the section, divergent reflectors indicate increased subsidence and rotation against the marginal faults. Alternatively, but less likely, the dome could be explained by an early phase of compression. Gently undulating reflectors in the upper part of the fill may represent small-scale folding caused by late (? post fill) compression. Farther east, at the location of the fold- and thrust-belt in Bellsund (Craddock et al. 1985, Hjelle et al. 1986, Maher et al. 1986, Dallmann 1988, Dallmann and Maher 1989, Maher et al. 1989, Dallmann et al. 1990), the seismic data are noisy in the uppermost second and scattered reflections at 1–2 seconds are difficult to relate to geological structures.

Line POLINV 2-84 in Van Mijenfjorden and lines BP 86-A and BP 86-G on Heer Land show the marked asymmetry of the wide Central Spitsbergen Basin. The basin is a little more than 2 seconds (5–6 km) deep, down to the top Hecla Hoek basement reflector, at line POLINV 2-84, sp. 400–500. In our interpretation of the lines we have followed age assignments of the different reflection events as given by Norsk Hydro (Fig. 5). There is a clear westward divergent geometry of seismic reflectors towards the fold- and thrust-belt. This is interpreted to be a result of thrust-tip wedging of imbricated thrust planes into the basin succession, similar to what Dallmann (1988) has described from the Berzeliustinden area. A strong and slightly undulating reflector at 1.5 seconds (sp. 500) can be followed for some 25 km eastwards. It probably represents a doleritic sill intruding Lower Triassic

strata (Eiken 1985). Beneath sp. 800–1200, the near base Upper Jurassic - intra Lower Cretaceous package can be seen to thicken, probably as a result of tectonic thickening within incompetent (Janusfjellet Subgroup, Dypvik et al. 1991) shales (Faleide et al. 1988, Nøttvedt and Rasmussen 1988, Nøttvedt et al. 1993a). Similar deformation, in the form of thrust duplex structuring, has been described from surface mapping farther east (Haremo et al. 1990, Andresen et al. 1992). Below sp. 1100 several thrusts and a possible backthrust originating from the same decollement are interpreted to merge and ramp to the surface, causing tectonic thickening also of the intra Lower Cretaceous - base Tertiary package and the formation of a pronounced anticline at the base Tertiary reflector. This structure was drilled by the Amoseas Group in 1965 at Blåhuken (well Ishøgda 1, see also Nøttvedt et al. 1993a).

Farther east, the land-seismic part of the transect makes a sharp bend above the inferred southwards continuation of the Billefjorden Fault Zone⁷ (at the intersection of lines BP 86-A and BP 86-G). The bend may be the reason for the somewhat unclear reflection patterns observed here, as other seismic data to the north define the fault zone well. The fault zone is interpreted as a single normal fault, bounding a low-relief half-graben to the east. The apparent graben width is about 20 km, but the seismic line is highly oblique to the graben axis and true width, therefore, is more like 12–13 km. The fault is not observed to penetrate Permian strata, and the graben fill is supposed to be mainly of early-middle newpage Carboniferous age, similar to the Billefjorden Graben⁸ farther north. Unconformably

⁷The lineament has been termed Inland Fault by Orvin (1940) and Billefjorden Fault Zone by Harland et al. (1974) and in most later papers.

⁸This feature has been termed Billefjorden Trough by Cutbill and Challinor (1965), Gjelberg and Steel (1981) and Johannessen and Steel (1992) and Billefjorden Graben by Nøttvedt et al. (1993a).

below the graben fill, a set of strong, but partly masked seismic reflectors are seen. They define a large syncline, about 25 km wide and 3 seconds (8–10 km) deep, and may represent Devonian and/or Early Paleozoic deposits. The Cretaceous sequence can be correlated to outcrops, as mapped by Salvigsen et al. (1989).

In Storfjorden, the transect crosses a NNW-SSE striking structure at depths between 1 and 3 seconds (2–8 km). Eiken (1985) interpreted the about 15 km wide

feature as a system of rotated fault blocks, called it Storfjorden fault zone and related it to late Devonian - late Carboniferous extension analog to grabens exposed in Billefjorden and Hornsund (Steel and Worsley 1984, Johannessen and Steel 1992). Alternatively the feature is of Devonian or early Paleozoic age. The size and similarity in seismic character to the syncline farther west in Heer Land may favour the latter interpretation.

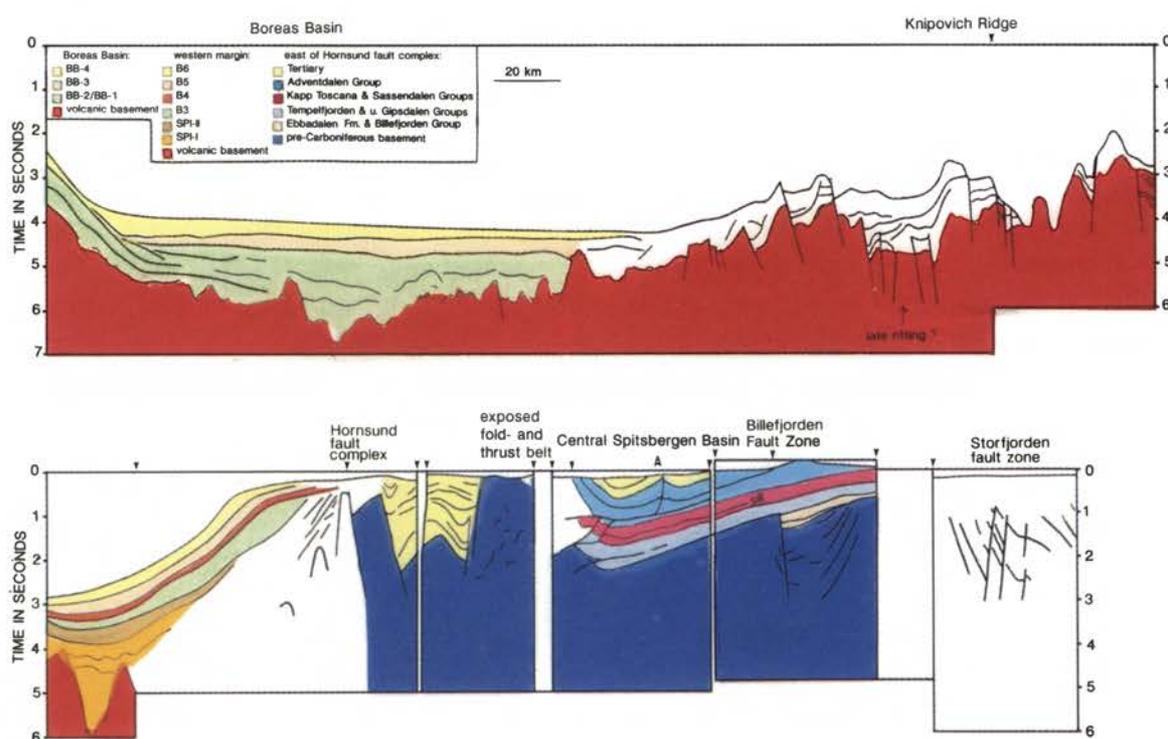


Figure 5: Line-drawing interpretation of transect 4.

2.5 Knipovich Ridge - Isfjorden - Adventdalen - Sassendalen - Sabine Land

by Arvid Nøttvedt, Ola Eiken and Peter Steven Midbøe

The transect crosses, within a distance of 300 km, an ocean spreading ridge, a passive continental margin, a fold- and thrust-belt and a continental platform area. Beneath the Knipovich rift-valley, only diffractions

are observed, in contrast to the horizontal layering seen in profiles farther south. Top basement is not clearly observed east of the rift-valley and diffractions (Fig. 6a) may here arise from basement or deformed sediments. The sequences SPI-I through SPI-III are less clearly defined in this area than farther south. Sequence SPI-I thins rapidly seaward at the lower slope, and its upper part (sequences B4–B6) wedges out. Kristoffersen (1990a) and Eiken and Hinz (1993) attributed this to bottom currents flowing along the lower slope during the late

Neogene and Quaternary.

A basement high at 4 seconds beneath sp. 1100–1150 has been crossed by a slope-parallel expanding spread profile (ESP 3-87, Austegard and Sundvor 1991), which reveals velocities above 8 km/s at about 8 km depth (Fig. 6a). These high velocities may represent anomalous mantle material within the crust. The eastern flank of this basement high is steeply dipping, down to 6.5 seconds at sp. 1400 (Fig. 6a), and can be traced farther down-dip to about 8 seconds (about 10–12 km) in a two-ship wide-aperture stacked⁹ seismic section (shown in the enclosed transect). One possible interpretation of this feature is a crustal scale shear zone.

On the shelf west of Hornsund fault complex, sediments dipping 6° – 8° are truncated beneath the U_o -unconformity and well-stratified sequences wedge out around sp. 2630. Farther east is an area with few primary reflections, which suggest that pre-Carboniferous basement is present near the seafloor. At sp. 3000–3350, a graben, about 15 km wide and about 2 seconds (3–4 km) deep, is probably part of a Tertiary graben system along the inner shelf (Eiken and Austegard 1987, also transects 3,4 and 7). A high-resolution refraction velocity profile with 3 km penetration (Fig. 6c) show the low-velocity sediments within the graben and the strong lateral velocity variations across Hornsund fault complex.

Two expanded spread profiles along the shelf (ESP 1-87 and ESP 2-87), together with gravity data, suggest that the main zone of crustal thinning occurs beneath the inner shelf (Fig. 6, Austegard et al. 1988, Sundvor and Austegard 1990, Austegard and Sundvor 1991, Faleide et al. 1991), and that a 15–20 km thick crust with a "continental" velocity distribution extends some distance west of Hornsund fault complex.

At the mouth of Isfjorden, on strike with

⁹This technique has been described by Buhl et al. (1982).

the fold- and thrust-belt north and south of the fjord (Ohta et al. 1992), reflections and diffractions extend from 5–6 seconds and down to 12–13 seconds. In Isfjorden a clear and fairly smooth reflection at about 11 seconds marks the base of the reflective crust and can be traced from about sp. 3800 and eastward to the end of line SVA 4-87. The reflector may be associated with the crust - mantle transition (Austegard et al. 1988). Chan and Mitchell (1982) modelled a 28 km or 35 km thick crust beneath central Spitsbergen and, for the shallow alternative, a low-velocity layer in the upper mantle. Based on all this, crustal thicknesses of about 35 km seems likely in Isfjorden. The reflector geometry, as well as gravity modelling (Sundvor and Austegard 1990), suggest crustal thickening beneath the fold- and thrust-belt (Fig. 6b). This is in accordance with Moho depth interpretations from refraction profiles in the fjords and off the Spitsbergen coast (Sellevoll et al. 1991). This paper proposed a block model with vertical displacements of Moho in Isfjorden. The reflection data do not support this.

Beneath the inner shelf and in Isfjorden, the upper 3–4 seconds of profile SVA 4-87 (which is shot with a deep-crustal focus) is obscured by high-amplitude water-bottom multiples. In Isfjorden we have duplicated the deep-seismic profile with line NH8509-207, acquired and processed with parameters tuned at shallower targets. Like in Van Mijenfjorden, the asymmetric Central Spitsbergen Basin is clearly defined. The basin is slightly less than 2 seconds (5–6 km) deep. We have used the same subdivision of seismic units as in van Mijenfjorden (Fig. 6c). The western flank of the basin is interpreted to be composed of several imbricated thrust nappes which involves basement, Upper Paleozoic and Mesozoic strata. This resembles the deformation in the frontal part of the fold- and thrust-belt that is exposed immediately north of Isfjor-

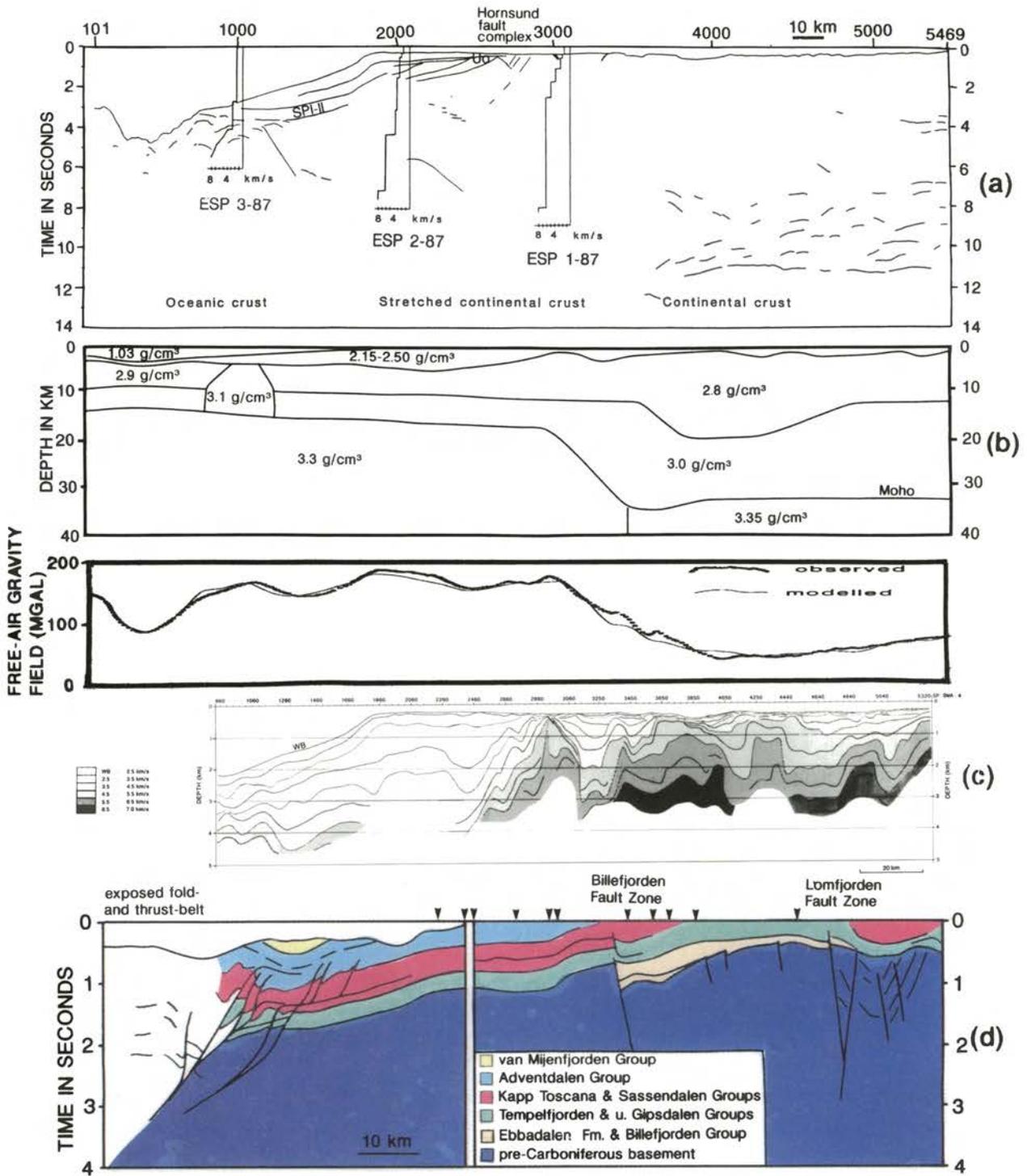


Figure 6: Line-drawing interpretation of transect 5: (a) the deep-seismic part (line SVA 4-87), with velocity-depth functions from expanding spread profiles included, after Austegard and Sundvor (1991). (b) Density model and free-air gravity anomalies, after Sundvor and Austegard (1990). (c) Iso-velocity section along line SVA 4-87, from Evertsen (1988), calculated from refractions observed on two-ship common midpoint gathers with shot-receiver offsets up to 12 km. Contour interval is 0.5 km/s. (d) Line-drawing interpretation of the land-seismic part.

den (Bergh et al. 1988, Bergh and Andresen 1990). Farther east, below sp. 1500–1900, the base Triassic - intra Lower Cretaceous package is thickened, most likely as a result of small-scale imbrication above a decollement surface in the Triassic Sassendalen Group. High-amplitude water bottom multiples disturb the upper part of the seismic section, but intra Cretaceous - Tertiary horizons appear to be relatively undisturbed by tectonic deformation. The imbrication causes, however, the formation of a weak monocline at the intra Lower Cretaceous and base Tertiary horizons. This monocline comprises the northwards, offshore prolongation of the Colesbukta anticline in Nordenskiöld land (Dalland 1979).

On the land-data, line NH 8802-03 shows a strong and slightly undulating reflector at 0.7 seconds (1.5 km). It can be followed eastwards to line UB 87-03, sp. 350 and probably represents a doleritic sill intruding Lower-Middle Triassic strata, similar to what is observed in Van Mijenfjorden (transect 4). Farther east (line UB 87-03, sp. 375), the transect crosses a distinct fault zone, the southerly striking Billefjorden Fault Zone (Major et al. 1992). The half-graben structure to the east is thought to be continuous with the half-graben observed in Heer Land (transect 4) and with the Billefjorden Graben north of Isfjorden (Gjelberg and Steel 1981, Lauritzen et al. 1989, Johannessen and Steel 1992, Nøttvedt et al. 1993a). The graben is between 15 km and 20 km in width and is bounded by a single, listric shaped normal fault. The fault does not appear to penetrate above middle Carboniferous, but a slight flexural bending and possible thickening of Upper Carboniferous strata across the fault is observed. High amplitude reflections characterize the graben fill, which appears to be little deformed. Several, less important normal faults are observed cutting the basement reflector on line NH 8802-12, sp. 350–450 and sp. 630, but they do not seem to involve the

sedimentary strata above. A backthrust is interpreted immediately above the western graben boundary fault, but is apparently detached from it. It cuts through Permian strata and is believed to sole out in Upper Carboniferous strata, similar to what has been described in outcrops farther north at Gipshuken (Ringset and Andresen 1988).

A gentle, east-facing monocline or flexural bending of the sediments on line NH 8802-31, sp. 200–300, is believed to represent the northward continuation of the Eistraryggen anticline in Agardhdalen (Andresen et al. 1992, Haremo and Andresen 1992), which forms part of the southerly striking Lomfjorden Fault Zone¹⁰. The line crosses the glacier Normannsfonna with a total surface relief of close to 500 m, however, and as no static corrections have been applied on the transect display, the relief of the monocline is suppressed. The base of the glacier is defined by a strong, irregular reflector at 0.05–0.1 seconds. An east-dipping reverse fault is interpreted to cut the top basement and intra middle Carboniferous horizons below sp. 200. It is interpreted to represent an antithetic fault to a larger west-dipping reverse fault causing the formation of the larger east-facing monocline. The sedimentary succession appears to thin eastwards, primarily as a result of onlap of Carboniferous strata against the pre-Carboniferous basement reflector. Below the basement reflector, between sp. 200–550, some fairly strong reflectors define a large syncline, some 20 km wide and 3 seconds (8–10 km) deep. These rocks are most likely Devonian or early Paleozoic in age, and possibly part of the same syncline as observed in Heer Land (transect 4).

¹⁰The lineament has been termed Lomfjorden fault belt by Cutbill (1968), Lomfjorden fault by Harland (1969), Lomfjorden Fault Zone by Harland (1979) and Steel and Worsley (1984) and Lomfjorden Fault Complex by Andresen et al. (1992).

2.6 Greenland slope - Hovgård Ridge - Molloy transform fault - off Prins Karls Forland

by Ola Eiken

The transect crosses the Fram Strait, the only deep-water gateway to the Arctic Ocean, and enters the lower continental slope east of Greenland. The total sediment thickness is at the western end about 1.2–1.5 seconds (1.1–1.6 km). Basement has, as in transect 4, an irregular short-wavelength and high-amplitude character, which suggest it is of volcanic origin. Possibly large volumes of volcanics were extruded at the margin when the continents split apart during the Eocene. Volcanics are inferred at the conjugate margin, Vestbakken volcanic province west of Bjørnøya (Eldholm et al. 1987, Faleide et al. 1988).

In Boreas Basin, top basement is relatively smooth east of an escarpment at sp. 5700 (Fig. 7). Eiken and Hinz (1993) subdivided the 3 seconds (3–5 km) thick section of sediments in the basin into four sequences, BB-1 through BB-4, of poorly known composition and ages. The uppermost two sequences thin and possibly wedge out towards the Greenland slope and they onlap Hovgård Ridge¹¹. The ridge is fault-bounded to the northeast (Fig. 7), while on the southwestern side, sequences BB-2 and BB-1 may continue at shallow levels within the ridge. The ridge may have been uplifted (upthrust) after the opening of Greenland Sea. Beneath the crestal parts a 0.1–0.2 seconds thick sequence represents Plio-Pleistocene mud and clay, confirmed by ODP wellsite 908. Beneath

¹¹The ridge was first recognized by Johnson and Eckhoff (1966) and called Hovgård Fracture Zone. Myhre et al. (1982) and later Johansen (1985) suggested it is a continental fragment split off from the Svalbard margin. Due to its uncertain origin I prefer the non-genetic name Hovgård Ridge.

this sequence, sediments of upper Oligocene age at the wellsite showed no indications of glacial influence.

Top basement is not well defined north-east of Hovgård Ridge (east of sp. 3700). Well-stratified sediments down to 1.0–1.5 seconds (1.1–2.0 km) below the seafloor are cut by normal faults with 10–50 m throw (the extent of this area is hatched in Fig. 20, p. 39) and suggest moderate extension after the igneous crust was formed. Rotated fault-blocks are interpreted deeper in the section, each 10–15 km across and tilted to the southwest as the case is for Hovgård Ridge. ODP wellsite 909 penetrated mostly clays and silty clays down to 1061 m below seafloor, with an early Miocene to late Oligocene age of the lowermost sediments (Thiede, Myhre et al. 1993). Deeper in the seismic section, intra-basement reflections are observed at 6.5 seconds (sp. 3100–3400), and Baturin (1990) showed similar intra basement reflections west of the ridge. He suggested the latter was related to a tectonic zone between two crustal blocks. Johansen (1985) interpreted the area east of Hovgård ridge as part of a same continental block as the ridge itself. An alternative interpretation may be extended oceanic crust. As a third alternative, Baturin et al. (1990) interpreted the complex seismic patterns beneath the well-stratified sediments as a pile of slices of oceanic crust related to overthrusting and transpressional faulting.

The Molloy transform fault¹² (sp. 1750–1950) is imaged as a zone of incoherent reflections beneath a bathymetric depression. The seafloor expression of the fault trace appears as two parallel fault scarps about 2 km apart in SeaMarc side-scan data (Sundvor et al. 1990, Crane and Solheim in press).

Top basement is not seen below the Svalbard continental slope and more than 2.5

¹²The feature is also called Molloy Fracture Zone (Vogt 1986b, Eldholm et al. 1990a). Since the topographic signature is subdued, I prefer to use the term *transform fault* (Bates and Jackson 1987).

seconds (3 km) thick sediments are present. The sequences are truncated at the seafloor at the steep lower slope, interpreted as a result of bottom current erosion (Eiken and Hinz 1993). A bottom simulating reflection 0.2–0.3 seconds beneath the seafloor probably represents the base of a zone of frozen gas hydrates (Fig. 7). Wavy patterns may be attributed to sediment waves, to slumps or to normal faults. A prominent depocenter for the uppermost sediments, sequence

YP-3 of Eiken and Hinz (1993), is present along the shelf edge. Reflectors downlap at the upper slope and the sequence thins rapidly. Strong seafloor multiples dominate the seismic section beneath the shelf. Hornsund fault complex is not imaged clearly, but a possible fault block at sp. 450 at 2.6 seconds (about 3 km) and steeply westward dipping events at shallower levels farther east may be parts of it.

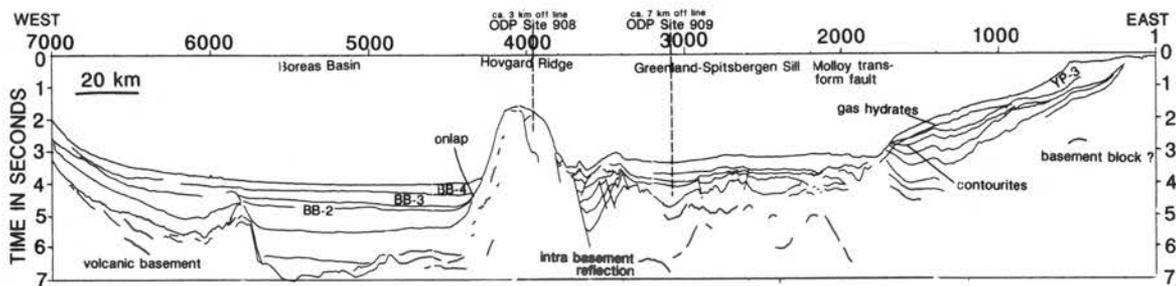


Figure 7: Line-drawing interpretation of transect 6.

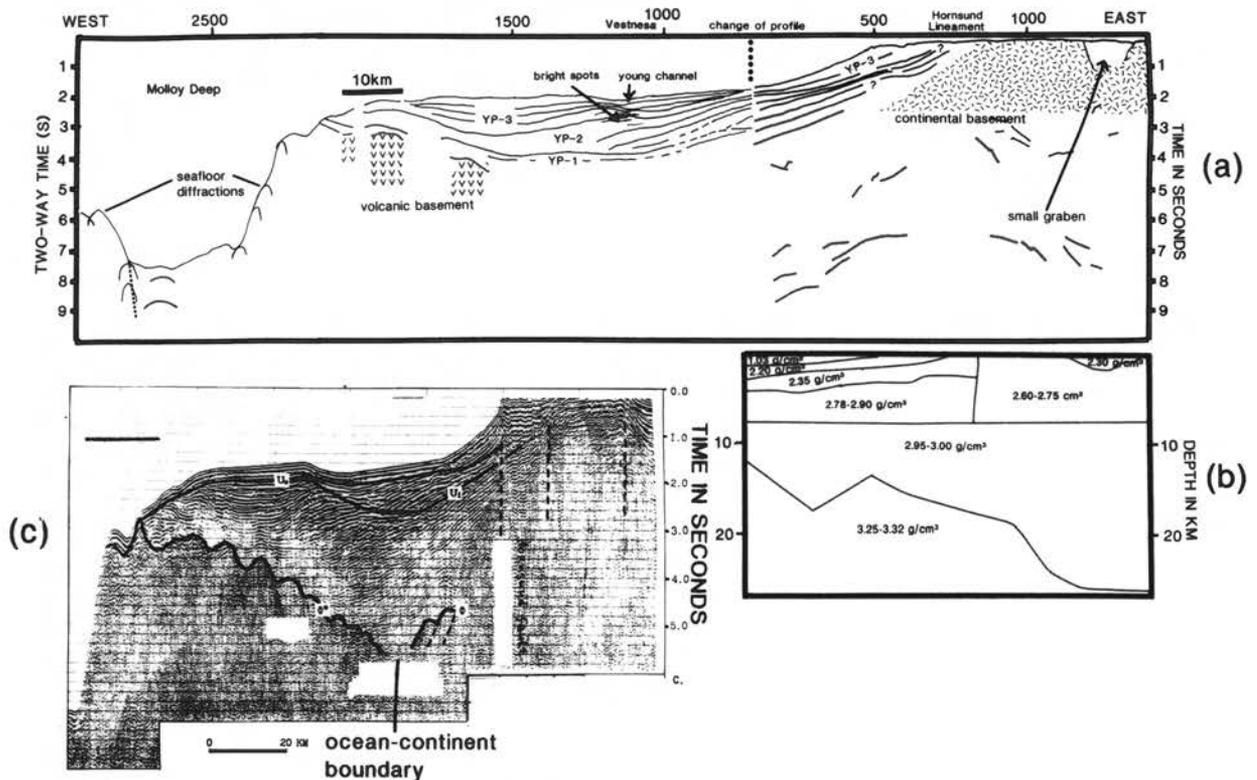


Figure 8: (a) Line-drawing interpretation of transect 7 after Eiken (1993). (b) Density model from Austegard (unpublished results). (c) Additional profile PR 85217 across Vestnesa with top basement interpretation, from Baturin et al. (1990).

2.7 Molloy Deep - Vestnesa - Sjubrebanken

by Ola Eiken and Dmitri Baturin¹³

The Molloy Deep is a nodal basin at the intersection of Molloy transform fault and Molloy Ridge (Figs. 1 and 20) and is more than 5600 m deep (Thiede et al. 1990a). A few weak diffractions are observed beneath a bright and relatively smooth seafloor. East of Molloy Deep, the 1500 m depth contour (Fig. 1) define a submarine ridge on the lower continental slope, called Spitsbergen Plateau by Matishov (1978) and Vestnesa¹⁴ by Eiken and Hinz (1993). Its westernmost part is cored by a basement high, while the eastern part of Vestnesa is formed by a more than 3 km thick pile of sediments (Fig. 8a). The uppermost sequences (YP-2 and YP-3 of Eiken and Hinz 1993) have well-defined depocenters on the ridge itself (Fig. 33, page 51). Kristoffersen (1990a) and Eiken and Hinz (1993) interpreted these sequences as contourites, and Eiken and Hinz (1993) suggested ages between 4 and 10 Ma for the oldest contourite deposits. Baturin and Nechkhaev (1989) suggested that a basal unconformity (U_1 of Fig. 8c) represents deep current erosion associated with opening of the deep-water passage between the Norwegian-Greenland Sea and the Arctic Ocean, and they suggested an age around 6.6 Ma for this.

Baturin and Nechkhaev (1989) further interpreted basement beneath Vestnesa to be downfaulted to the east in a series of step faults with individual throws up to 1.5 km (Fig. 8c). They also reported Moho refractions (with P-wave velocities around 8.1 km/s) at only 13 km depth beneath the central part of the plateau, and at 24 km depth

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¹⁴We use Vestnesa here because of its resemblance to a ridge rather than a plateau as seen on recent bathymetric maps.

beneath the shelf.

The sequences YP-1 through YP-3 are truncated near the seafloor between sp. 600–800 on line SVA 17-87, which is the approximate location of Hornsund fault complex (Eiken and Austegard 1987, Eiken 1993). An about 10 km wide graben close to the coast is probably filled with early Tertiary sediments similar to Forlandsundet Graben and other grabens on the inner shelf farther south (transects 3, 4 and 5).

The deep-crustal part of the transect shows bands of high-amplitude reflections at 6–7 seconds beneath the shelf, dipping westwards (on the time-section) to 7–8 seconds beneath the slope. These dips are difficult to relate to expected oceanward crustal thinning. Gravity modelling (Fig. 8b) suggests locally thicker crust beneath the upper slope. A different crustal transition here compared to farther south and north may be related to Spitsbergen Fracture Zone, which projects onto this area (Fig. 1). Max and Ohta (1988) suggested that fractures in the continental crust control the orientation of the plate boundary west of Svalbard.

2.8 Southern Yermak Plateau - Northern coast of Spitsbergen

by Ola Eiken

The transect crosses the continental slope off the northwestern tip of Spitsbergen, and this area may be classified as a part of Yermak Plateau. Farther east the transect passes close to the northern coast of Spitsbergen, paralleling the coast line as well as the shelf edge, and probably crossing offshore extensions of the N-S trending faults exposed in northern Spitsbergen.

The southwestern Yermak Plateau shows up to 3.5 seconds (about 5 km) thick sediments above an irregular and bright basement reflector, with a basement high be-

neath the western slope (Fig. 9). The upper sediments, sequences YP-2 and YP-3 of Eiken and Hinz (1993), thicken westwards. Sequence YP-2 downlaps to the west and can be mapped as a mound striking parallel to the continental slope (Fig. 33, page 51). Eiken and Hinz (1993) interpreted it as a contourite mound. A horst complex composed of a main high and a tilted western block at the eastern end of line SVA 9-87 separates sequence YP-1 on western Yermak Plateau from the deeper sediments in *Danskøya basin*¹⁵. The horst complex is located at the northward extension of the Hornsund fault complex, and it is probably continuous with the easternmost of two blocks on Sjubrebanken (Fig. 17, page 35). Eiken (1993) traced the lineament north to 80°30' N, farther north there is no data.

Danskøya basin is up to 3 seconds (nearly 5 km) deep, about 20 km wide and about 100 km long in the N-S direction (Fig. 1). On basis of three low-resolution (deep-seismic) profiles within the basin, the sediments can be divided into four sequences, labelled DB-1 through DB-4 (Table 1). The sequences DB-2 through DB-4 are truncated at the seafloor towards land, suggesting a hinge line for late Cenozoic uplift of northwestern Svalbard seaward of the basin rim. Rapid subsidence in central parts of the basin is suggested by large scale onlap

patterns and recent subsidence is suggested by the present narrow shelf and water depths of 400–500 m only 20–30 km away from land. Farther east the transect passes close to the coast, most likely with Devonian and Hecla Hoek rocks at the seafloor. No primary reflections can be discerned in the upper part of the section in this area—only seafloor multiples.

The crust-mantle transition is probably seen beneath the southwestern Yermak Plateau (line SVA 9-87 sp. 400–900) as sub-horizontal reflections at 6–7 seconds (12–15 km). Towards the horst complex, these reflections originate somewhat deeper (Fig. 9). Combined with gravity modelling (Austegard and Sundvor 1991) this suggest an eastward thickening of the crust. Eiken (1993) interpreted the horst complex as the boundary between volcanic or oceanic crust on the southwestern Yermak Plateau and stretched continental crust on southeastern Yermak Plateau.

North of Spitsbergen we observe an abundance of eastward dipping middle and lower crustal reflections and diffractions along the transect. This dip preference suggests crustal thickening towards east. Eiken (1993) speculated if thin continental crust beneath the mountaneous northwestern tip of Spitsbergen was due to simple shear and asymmetric margin development.

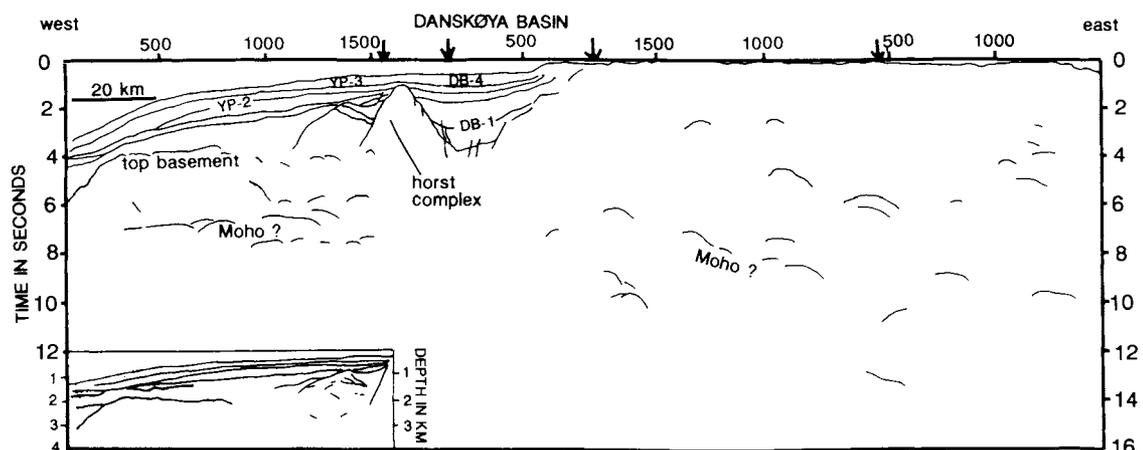


Figure 9: Line-drawing interpretation of transect 8, with a depth-converted version of the western part in the lower left corner.

¹⁵The basin was described and termed Danskøya basin by Eiken (1993).

2.9 Southern Yermak Plateau

by Ola Eiken

The transect crosses the Yermak Plateau near the crestal parts of Nansen Bank. At the western slope, long wavelength irregular reflection patterns make it difficult to divide the sediments into sequences (Fig. 10). Slumping and bottom current controlled deposition may have caused these patterns. Top basement reflections are not identified, maybe because the data have been acquired with a weak source. Beneath the upper slope and Nansen Bank, sub-parallel reflections are observed down to the first seafloor multiple, which is the penetration limit here. These sediments belong to sequences YP-2 and YP-3 of Eiken and Hinz (1993). The transect crosses the recent ODP wellsite 911, which penetrated 505 m of Pliocene and Pleistocene muds within the seismic sequences YP-2 and YP-3. High sedimentation rates, above 100 m/m.y., were found in the well. Beneath southeastern Yermak Plateau, east of sp. 200 on line UB 12-79, bright top basement reflections are observed through the multiples. Basement has a rough surface, prob-

ably it represents the top of faulted blocks. This may be the top of stretched continental crust (Jackson et al. 1984, Baturin et al. 1990, Sundvor and Austegard 1990, Eiken 1993). These works, as well as Feden et al. (1979), suggested the southern Yermak Plateau has a continental type of crust, while Crane et al. (1982) and Eiken (1993) suggested that only the southeastern plateau is of continental type, and the southwestern plateau of oceanic or volcanic nature.

Beneath sp. 550-850 on line UB 12-79, the strong and smooth reflections at the base of the well-stratified sediments may arise from sills (Eiken 1993) or lava flows. In this area, Eiken and Hinz (1993) interpreted wavy patterns as buried moats caused by intensified bottom currents around ancient topographic highs (Fig. 10).

High heatflow has been measured on the southern and western parts of Yermak Plateau (Crane et al. 1982, Vetaas 1990). Particularly high heatflow around Sophia Canyon may be related to the inferred sills or lava flows. Late Cenozoic volcanism is found in northern Spitsbergen (Hoel 1914, Prestvik 1978), and several papers have speculated as to whether the volcanism is related to the tectonic evolution of the pla-

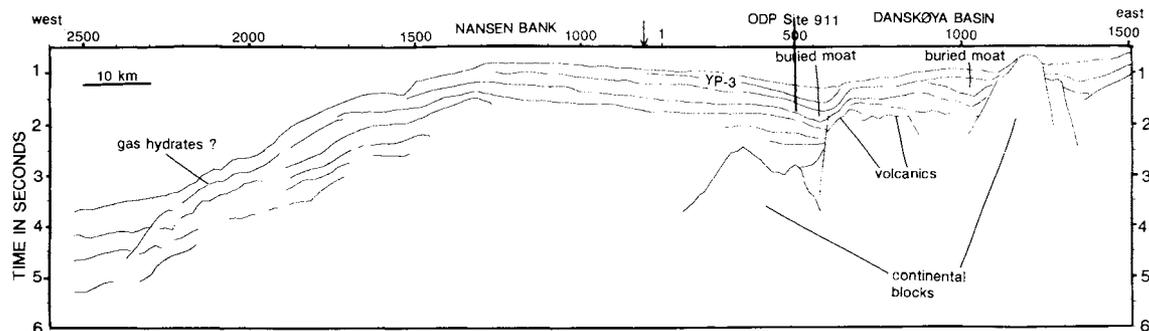


Figure 10: Line-drawing interpretation of transect 9.

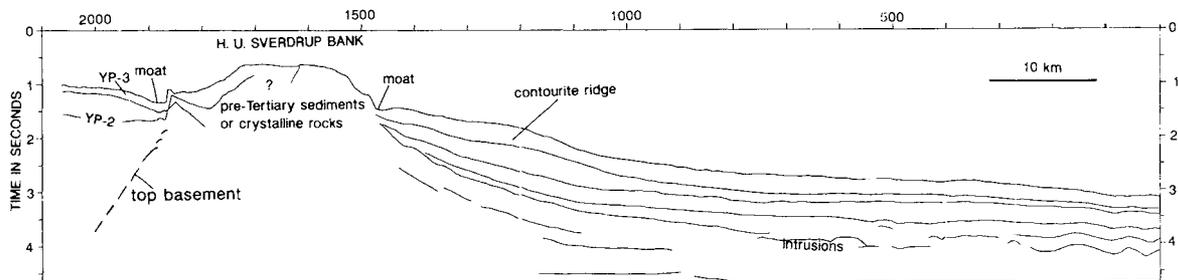


Figure 11: Line-drawing interpretation of transect 10, after Eiken (1993).

teau or to a thin mantle lithosphere, which also may be responsible for the late Cenozoic uplift of Svalbard (Jackson et al. 1984, Amundsen et al. 1987, Skjelkvåle et al. 1989, Vågnes and Amundsen in press).

2.10 Yermak Plateau - Nansen Basin

by *Ola Eiken*

This is the northernmost transect across Yermak Plateau, and it extends eastwards across the eastern slope and into Nansen Basin. Farther north on the plateau, where Feden et al. (1979) inferred volcanic (oceanic) rocks on basis of high-amplitude magnetic anomalies, no reflection seismic coverage exists south of 82°40' N (Kristoffersen and Husebye 1985).

H.U. Sverdrup Bank is flat-topped at the seafloor and is bounded by well-defined slopes and moats, probably caused by bottom current erosion (Sundvor et al. 1982b, Kristoffersen 1990a, Eiken 1993). It is not clear from seismic profiles whether the bank is fault-bounded. A sonobuoy refraction profile across H.U. Sverdrup Bank (Austegard 1982) shows laterally varying seismic velocities, generally above 5 km/s, and this indicates lithological variations within the bedrocks making up the bank. A Precambrian high grade gneiss sample has been dredged from the bank (Sundvor et al. 1982b, Jackson et al. 1984), but no fresh surfaces were observed and the sample may be glacial erratic. Nevertheless, a continental origin of the bank (and all the southeastern Yermak Plateau) seems most likely, as its aeromagnetic signature is very smooth (Feden et al. 1979) and the basement surface is block faulted. Possibly large blocks were transported along the shear zone between Svalbard and Greenland during early-middle Tertiary.

West of the bank, about 1 second (1.0–1.2

km) of sediments can be discerned above the seafloor multiple. Top basement dips steeply westwards, may be as a series of tilted blocks (Fig. 11).

In deeper water east of H.U. Sverdrup Bank up to 2 seconds of well-layered sediments have been penetrated, without clear definition of basement beneath. A low relief contourite ridge (Fig. 11) is attributed to deposition by bottom currents. Farther east the seismic character becomes more transparent, and probably represents hemipelagic sediments and turbidites. High-amplitude and laterally smooth reflections around 4 seconds at sp. 200–800 have been interpreted as sills (Eiken 1993).

2.11 Gråhuken - Norskebanken - northern continental slope

by *Ola Eiken*

This north-south transect crosses the westernmost part of the northern Svalbard margin. The upper part of the seismic section is dominated by seafloor multiples beneath the inner shelf, just offshore Devonian exposures at Gråhuken. A fault beneath sp. 2450–2500, called *Moffen fault* by Eiken (1993), bounds the low-velocity sediments at Norskebanken towards land. Based on the present seismic line coverage, it seems like Moffen fault north of Wijdefjorden curves and changes from an E-W striking fault in the west to a hinge line trending NE-SW farther east (Fig. 1). The fault was probably activated during opening of Eurasia Basin (Eiken 1993). More than 2 seconds of sediments are observed on the outer shelf (Fig. 12). Top basement is not imaged, but sonobuoy measurements suggest total sediment thicknesses of 3–4 km (Sundvor et al. 1978). Eiken (1993) divided the sediments into an upper prograding sequence, NB-2, with a 1.4 sec-

ond (about 1.5 km) thick depocenter near the shelf edge, and a lower (mega)sequence NB-1. He tentatively correlated sequence NB-2 with Miocene and younger strata, and related it to uplift of Svalbard and glacial erosion. Sequence NB-1 may be related to Cretaceous and Tertiary syn-rift and post-rift deposits.

At the lower slope, sequence NB-2 thins rapidly seaward (Fig. 12), possibly because of rapid deposition of coarse material. Beneath this sequence is a transparent and discontinuous reflection pattern, called *fuzzy* reflector by Sundvor et al. (1978). Deeper in the section we observe smooth, parallel reflections between 4 and 5 seconds. These horizons are probably shallower than crystalline basement.

A band of laterally smooth lower crustal reflections at 7–8 seconds just north of Mofen fault increases in depth to 9–10 seconds towards land. Sundvor and Austegard (1990) suggested these arise from the crust/mantle transition, and that major crustal thinning occurs here. They supported this by gravity modelling (Fig. 12). A more gradual continent/ocean transition here compared to the western margin may be due to extensional movements without significant shear components. Austegard and Sundvor (1991) modelled a positive gravity anomaly of about 100 mgal around the shelf edge with a dense and shallow upper basement and a shallow Moho (Fig. 12). Farther seaward oceanic crust is probably present.

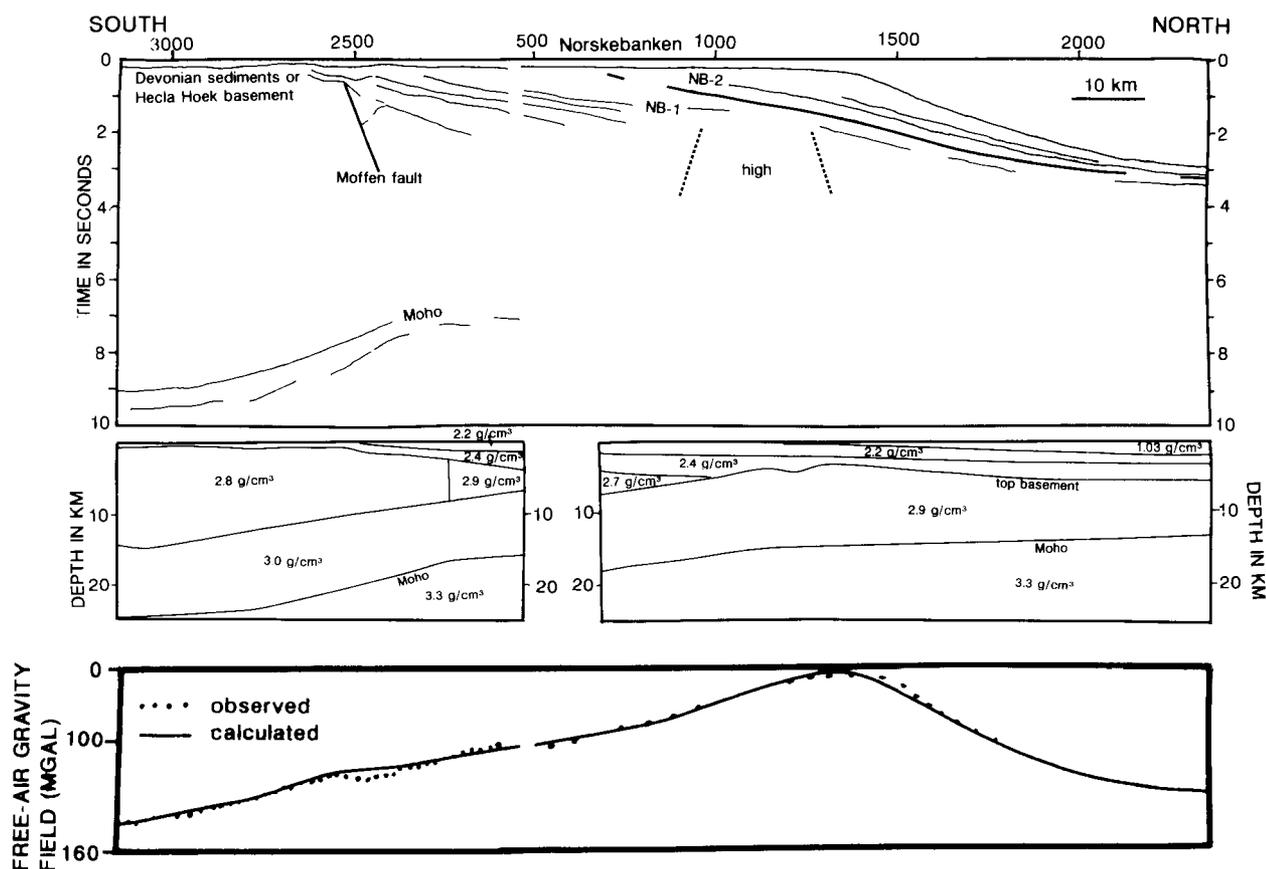


Figure 12: Line-drawing interpretation of transect 11, after Eiken (1993). Density models after Sundvor and Austegard (1990) on the inner part and Austegard and Sundvor (1991) on the outer part.

2.12 North of Nordaustlandet

by Fridtjof Riis¹⁶

The southern, continental, and the northern, oceanic, parts of transect 12 are separated by a large fault referred to as the boundary fault (Fig. 13), and the geology across the fault cannot be easily correlated.

Southern part

The southern part of the transect is located on the continental platform north of Nordaustlandet (Fig. 1). The data indicate that the surface of the crystalline basement slopes gently towards the Nansen Basin, with a fairly thin cover of sediment. A few graben structures have been observed to break the basement surface, which might be analogous to the structures presented in transects 11 and 13.

In transect 12, the basement is overlain by an approximately 1 km thick sequence of sediments characterized by intensive slumping (Fig. 13). Because of the slumping, a subdivision into seismic sequences is difficult. Interpretation of scattered seismic lines farther south on the platform indicates that a sequence boundary can be defined at the base of the major slumped section A (Fig. 13). The boundary is characterized by downlap on the underlying sequence B, and onlaps onto the low relief structural highs. Onlap is interpreted at the outermost high in the transect (sp. 660).

The age of the sediments cannot be established by seismic correlation to dated sections. It is suggested that sequence A correlates with the thick wedges of Upper Pliocene to Pleistocene sediments which cover the shelf margins west and south of Spitsbergen. It is probable that sequence B also belongs to the Tertiary. In fact,

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the sloping basement surface observed both on this line and along the southern part of transect 11 might correlate with the peneplain observed in northern Spitsbergen south of Gråhuken. Here, Tertiary lavas flowed over a well defined peneplain which is today elevated to 1000 m to 1200 m above sea level, and which dips to the north (Hjelle and Lauritzen 1982). Skjelkvåle et al. (1989) indicate an age range of 2.5 Ma to 11.5 Ma for the volcanics. This indicates a Tertiary and possibly Neogene age for the peneplain. This young age supports the interpreted young age for the sedimentary cover offshore.

Northern part

The northern part of the transect extends across the deep floor of the Nansen Basin. The water depth increases abruptly from 1000 m to 3500 m where the line intersects the boundary fault. Unfortunately, the line was only recorded to 7 seconds, and the top of the oceanic crust can be seen only between sp. 1400 and 1650. The oceanic basement is covered by approximately 2 seconds of sediments with interval velocities in the range 2000 m/s to 3000 m/s. The cover can be divided into two major seismic sequences, each of which exhibiting a thickness of approximately 1 second.

The lower seismic sequence (T1) comprises a series of mainly parallel reflectors with good continuity. The sequence both onlaps and is flexured upwards towards the boundary fault, and internal onlaps can be identified where the oceanic basement forms a structural high. Normal faults with small throws penetrate the sequence.

The upper sequence (T2) can be divided into several subsequences. Most of the subsequence boundaries in addition to the base of T2, are characterized by downlap. The ocean floor itself has a smooth surface with scattered irregularities apparently caused by slumping, and the internal reflectors in

T2 are less continuous than in T1. Sequence T2 thickens towards the boundary fault, and very little drag is observed.

A possible interpretation is that sequence T1 was formed as a distal deep water deposit during the opening of the Eurasia Basin when the boundary fault was active. Sequence T2 has a more proximal source, and slumping could be an important mechanism of sediment transport, which apparently took place when there was limited vertical movement along the boundary fault. Sequence T2 is therefore tentatively correlated with sequence A and the Upper Pliocene/Pleistocene wedges, while T1 might span a large part of the Tertiary.

at sp. 700. The fault is interpreted to define the boundary between continental and oceanic crust. According to sparse bathymetric data (Cherkis et al. 1990), the orientation of the fault is WNW-ESE. Thus the southwestern corner of the Nansen Basin has an exceptionally narrow transition zone between continental and oceanic crust. The geometry might suggest a translational component of motion along the fault, although this does not seem to be compatible with current ideas on the opening history of the Eurasia Basin. A correct interpretation of the fault will give an important additional constraint for the tectonic development of the region.

The boundary fault

The boundary fault seems to be steeply dipping with a throw of at least 2000 m (Fig. 13). It is associated with a basement dome

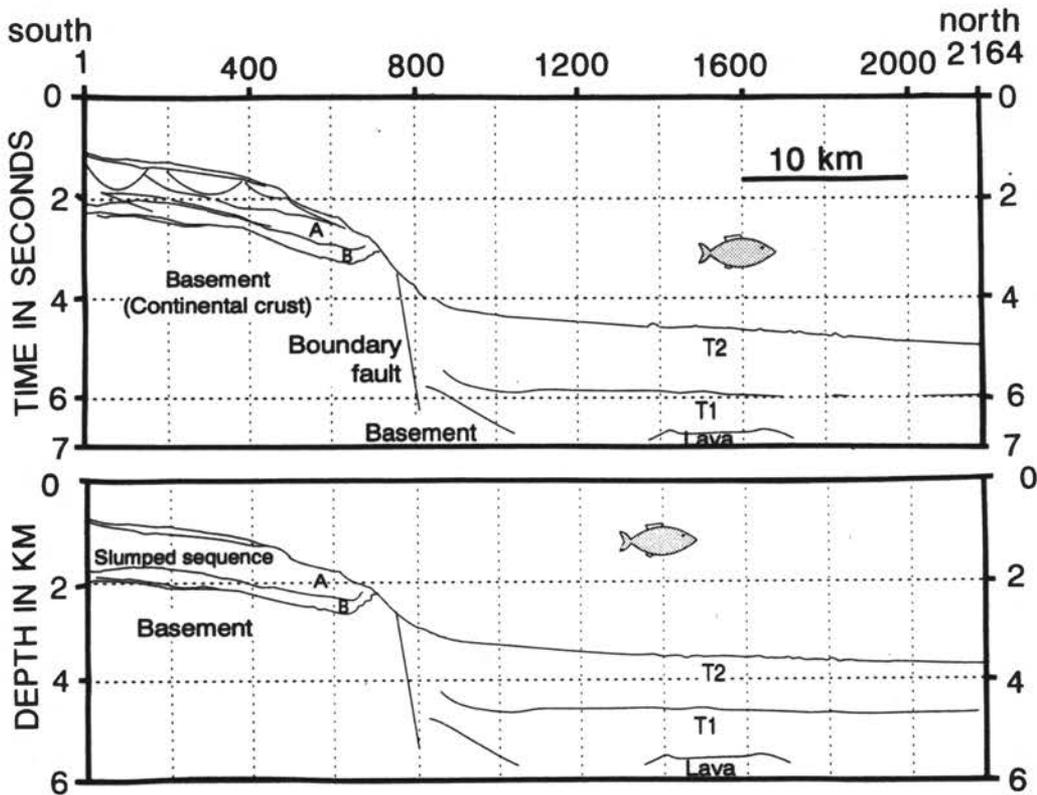


Figure 13: Line-drawing interpretation of transect 12. The depth conversion was made by assuming water velocity 1480 m/s and sediment velocity 2000 m/s.

2.13 Between Nordaustlandet and Franz Josef Land

by Dmitri Baturin

This northeasternmost transect is situated in a poorly explored area, and there is a discrepancy in water depths seen on the seismic line and on the map of Cherkis et al. (1990).

The total thickness of sediments on the continental slope along this transect does not exceed 3.5 km. A major structural boundary is present beneath the shelf edge. The continental substratum dips seaward in a system of listric faults with southward tilt of individual block surfaces (Baturin 1987).

The lower seismic sequence constitute the inter-block graben fill. The sequence is expressed in the seismic wave field as a chaotic area with no continuous reflectors. The lower part of the sequence is acoustically transparent. The sequence is interpreted as synrift deposits. The age of the sediments is probably Cretaceous if their deposition is

related to the beginning of extension and intracontinental rifting within the area of the future Eurasia Basin (Baturin 1987). The extent of the second seismic sequence is limited to the foot of the continental slope. The reflectors of the sequence form a well-bedded uninterrupted succession which onlaps the basement flanks. Basal horizons of the sequence are probably of upper Paleocene age, deposited after seafloor spreading within Eurasia Basin had started, and the continental margin began to evolve.

The third seismic sequence is defined within the outer shelf and the upper part of the continental slope. It represents a large sedimentary body, characterized by propagational internal structures with clinoforms.

The depocenter of a fourth seismic sequence is situated on the lower part of the slope. The unconformity between the third and the fourth sequences has been interpreted to result from the most significant sea level drop during the Cenozoic (29 Ma), which also resulted in the depocenter of strata moving from the upper continental slope into the basin (Baturin 1987).

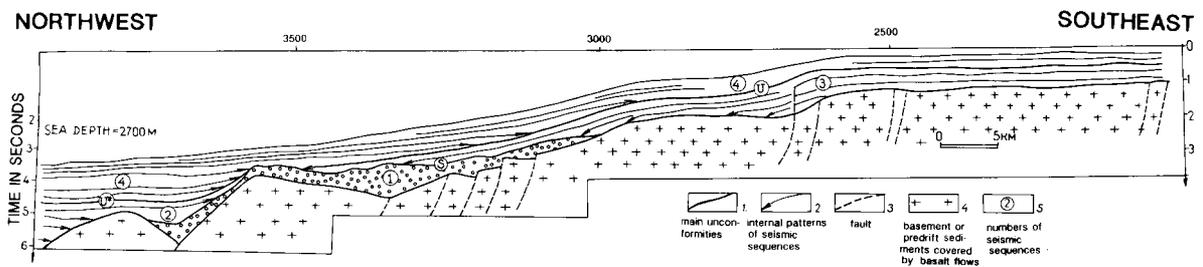


Figure 14: Line-drawing interpretation of transect 13, after Baturin (1987).

Chapter 3

Seismic expressions of the main geological features

3.1 Svalbard: The lower crust

by Ola Eiken and Atle Austegard¹

Continental crust

Few data exist on crustal thickness and structure in Svalbard: expanded spread profiles across transects 1 and 5, deep-seismic reflection profiles as shown in the transects (and 4–5 additional short lines), geobarometry of the mineral assemblages in xenolithes of the Quaternary volcanic centers in the Bockfjorden area in northern Spitsbergen, suggesting a crustal thickness of about 27 km (Amundsen et al. 1987, 1988), and maps of free-air gravity anomalies (Faleide et al. 1984, Austegard and Sundvor 1991). Austegard and Sundvor (1991) compiled a map of crustal thicknesses, shown in Fig. 15, based on deep-seismic and gravity data.

The continental crust can be divided into three geological provinces, each with different seismic reflection patterns and crustal thicknesses:

i) *The continental platform.* Isfjorden, Storfjorden and Spitsbergenbanken show a highly reflective lower crust, with abundan-

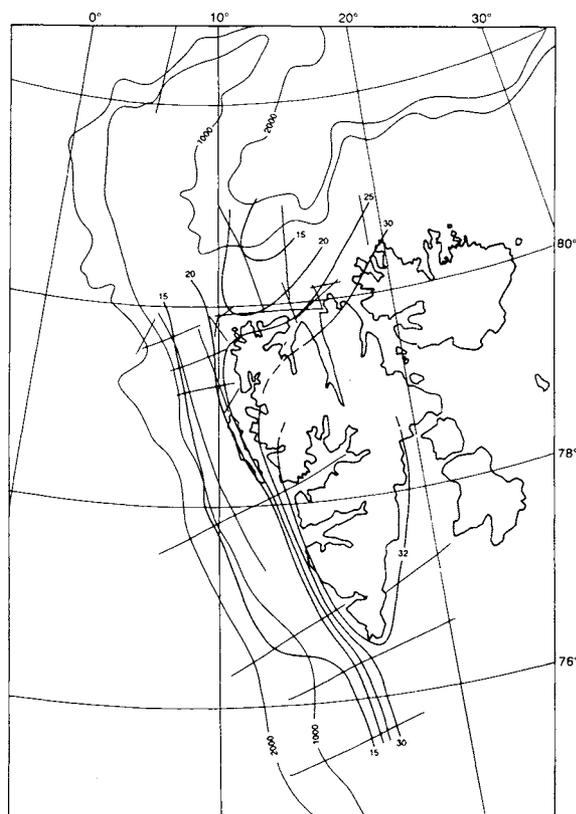


Figure 15: Map showing contours of depth to Moho (km), estimated from deep seismic reflection and refraction measurements and gravity modelling (from Austegard and Sundvor 1991). Shown is also bathymetry (in meters) and location of deep-seismic reflection lines.

ce of diffraction hyperbolas between 4 and 11 seconds. Crustal thicknesses are about 30–35 km. This area has probably been

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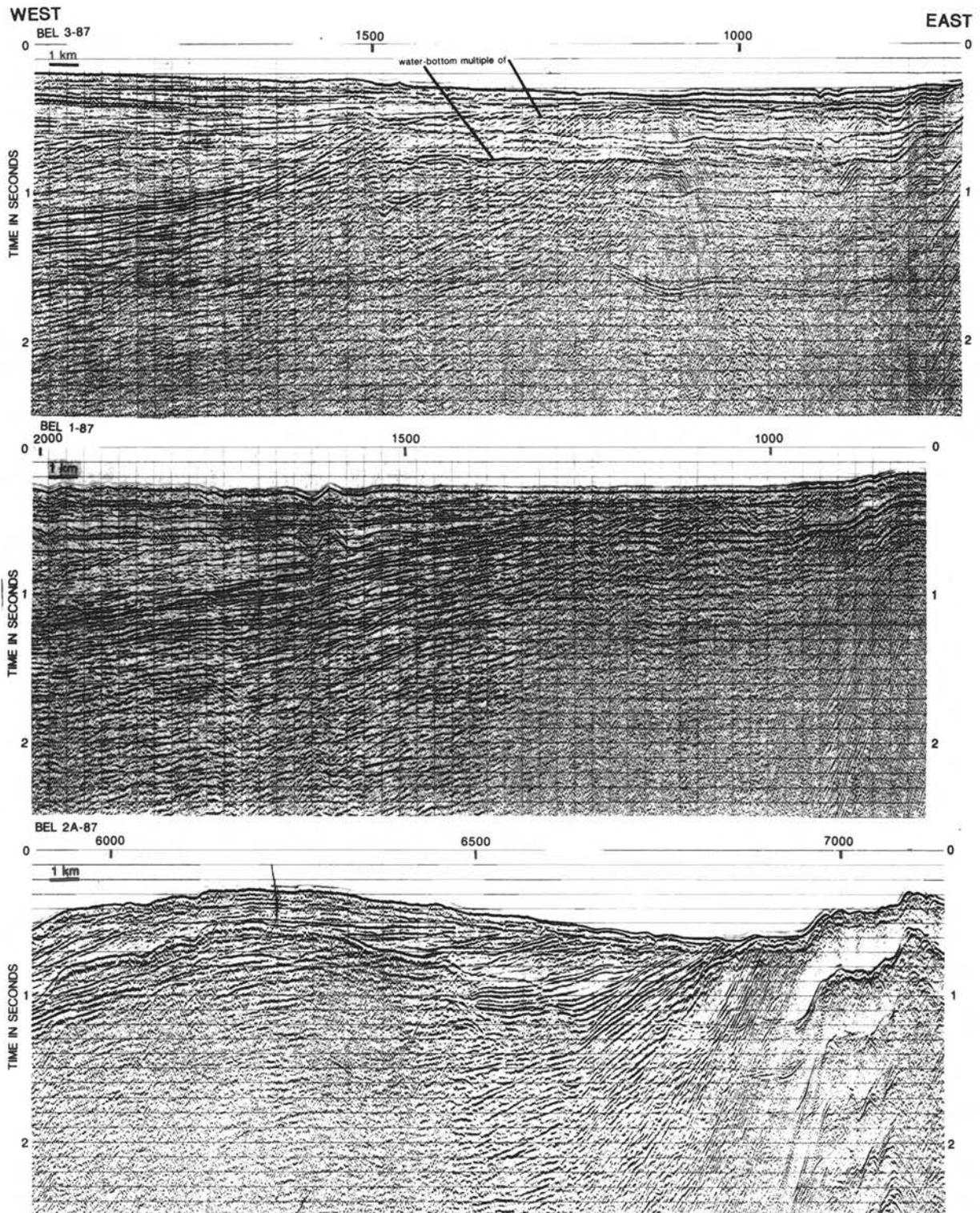


Figure 16: Aligned profiles across Hornsund fault complex west of Hornsund - Bellsund. Locations are shown in Fig. 1. (above) line BEL 3-87 (middle) line BEL 1-87 (below) line BEL 2-87.

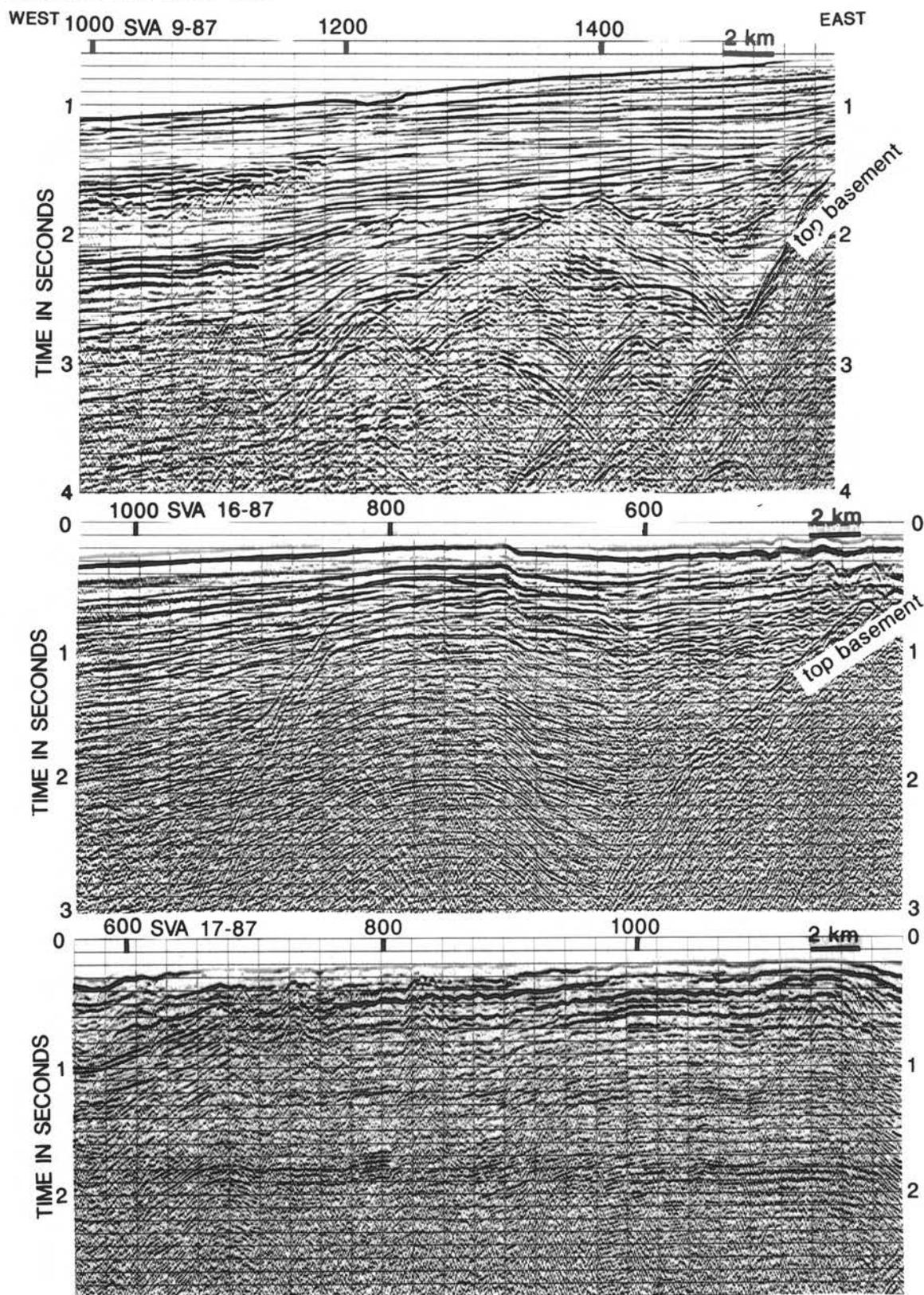


Figure 17: Aligned close-ups of profiles across Hornsund fault complex, north of Prins Karls Forland. Locations are shown in Fig. 1. (above) line SVA 9-87 (middle) line SVA 16-87 (below) line SVA 17-87.

tectonically stable since late Paleozoic time.

Two zones of intraplate earthquakes have been observed within the platform: in Heer Land and Nordaustlandet. Bungum et al. (1982) and Chan and Mitchell (1985) attributed these to the interaction of a regional stress field of plate tectonic origin with local zones of weakness.

The cause of about 3 km of late Cenozoic uplift in central Spitsbergen (Manum and Throndsen 1978, Eiken and Austegard 1987) is poorly understood, as is the uplift along western Fennoscandian shield and western Barents Shelf during the same time frame (Riis and Fjeldskaar 1992, Nyland et al. 1992). Vågnes et al. (1992) and Vågnes and Amundsen (in press) suggested the uplift has been caused by an anomalously thin mantle lithosphere in Spitsbergen, giving high heat-flow both onshore and offshore.

ii) *The Tertiary fold- and thrust-belt.* The base of the reflective lower crust extends down to 12–13 seconds (35–40 km) beneath the Tertiary fold- and thrust-belt (transects 2 and 5). Crustal thickening of about 5 km is also supported by gravity modelling along these transects. This probably represents the root zone of the exposed tectonic belt.

iii) *The continental margins.* A relatively narrow zone of crustal thinning is observed along the western margin (transects 1–7). Only few deep reflections are observed beneath the western shelf south of Prins Karls Forland, while distinct reflections are shallowing to the north beneath Sjubrebanken.

The upper crust of the western margin can be divided into an inner part with horsts and grabens and an outer part with a thick wedge of well-stratified sediments. In between is Hornsund fault complex, a zone of distinct faults in the Hornsund - Sørkapp area and steeply dipping strata truncated near the seafloor at Isfjordbanken and Sjubrebanken, with rapid east-west changes in seismic velocities. The variation in seismic expressions along the strike

of Hornsund fault complex is shown in Figs. 16 and 17.

Several papers (Sundvor and Eldholm 1976, Myhre et al. 1982, Eldholm et al. 1987, Myhre and Eldholm 1988, Faleide et al. 1991) have proposed that the Hornsund fault complex played a major role in early Tertiary shear movements between Svalbard and Greenland, and that the ocean-continent boundary is located near the fault complex. Austegard et al. (1988) interpreted thin continental crust in places west of Hornsund fault complex as a result of later extension overprinting shear movements.

Baturin et al. (1990) interpreted the ocean/continent transition at the junction of the Knipovich Ridge / Molloy transform fault to have a rifted origin, and they traced the transition in a NE direction parallel to the strike of Molloy Ridge. Furthermore, they interpreted the ocean/continent transition north of 79.5° N as a stepped zone seaward of Hornsund fault complex, originating from strike-slip and rifted segments. Jackson et al. (1984) and Eiken (1993) suggested that continental blocks on the southern Yermak Plateau were formed during early Tertiary shearing at the triple junction north of Spitsbergen.

A laterally more variable zone of crustal thinning seems likely north of Spitsbergen (transects 11–13). Beneath Norskebanken strong and continuous lower crustal reflectors dip towards the continent (to the south-southeast).

Oceanic crust

Eurasia Basin opened at about 58 Ma, between magnetic anomaly 25 and 24B (Kristoffersen 1990b, Lawver et al. 1990) with separation of Lomonosov Ridge from the northern Svalbard/Barents Shelf margin. A detailed reconstruction of the seafloor spreading history of the Greenland Sea and Fram Strait has not yet been in-

ferred, due to lack of well-defined magnetic anomalies (e.g. Vogt 1986b). Constraints from magnetic anomalies in the Norwegian-Greenland Sea and Eurasia Basin suggest opening of the northern part after 35 Ma (magnetic anomaly 13 time), when a major plate reorganization took place (Talwani and Eldholm 1977, Vogt et al. 1978, Myhre et al. 1982, Reksnes and Vågnes 1985, Vogt 1986b), as outlined in Fig. 18.

The present plate boundary follows the Knipovich rift-valley, the Molloy transform fault and the Molloy Ridge. The location of the boundary is inferred on the basis of bathymetry (Perry et al. 1980, Eldholm et al. 1990b), seismicity (Husebye et al. 1975, Savostin and Karasik 1981) and heat-flow (Crane et al. 1982, 1988).

Knipovich rift-valley is generally 8–10 km wide, about 1 km deeper than the surrounding seafloor, and it has a distinct gravity low (Grønlie and Talwani 1982). Ohta (1982) argued from detailed bathymetric data that it consists of a large number of small spreading and transform segments, each only a few kilometres long, but the dominating feature seen in long-range side-scan data is one linear, broad rift-valley (Sundvor et al. 1990, Crane and Solheim in press). Neumann and Schilling (1984) dredged pillow basalt and

sediments in the rift-valley north to $77^{\circ}40'$ N, farther north they found only sediments. The valley floor south of $77^{\circ}50'$ N is highly reflective with weak, sub-parallel reflections beneath (Fig. 19a and 19b). The inner walls of the rift-valley consist in places of a series of basement blocks with sediment cover, each downfaulted some hundred meters along fault planes dipping about 45° (Fig. 19b). In other places basement and sediment cover form a smooth slope towards the valley (Fig. 19a). Just south of 77° N a 6–8 km wide graben branches off to the east from the main rift valley (Fig. 20), with throws of more than 1 km. The amount of extension within this graben can be calculated to a few kilometres. Sediments observed within the rift-valley have been attributed to heavy sediment supply from Svalbard and the Barents Sea in the late Cenozoic (Vogt et al. 1978, Myhre and Eldholm 1988, Eiken and Hinz 1993, chapter 2.2).

The northwestern part of the aseismic Molloy transform fault is well defined in seismic profiles (Fig. 19c), while profiles crossing the southeastern part show a more diffuse zone, with small normal faults cutting through the sediments in a wide area (transect 6). The character of the seismic

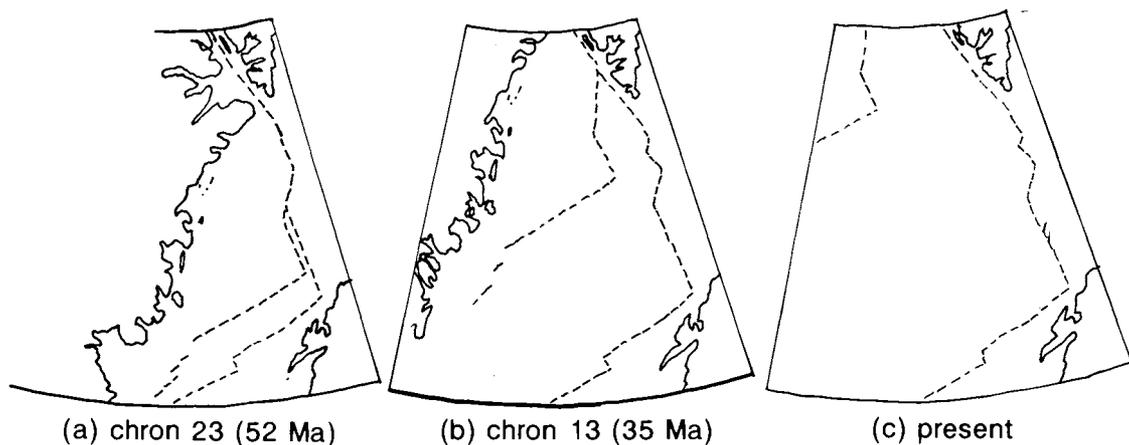


Figure 18: Schematic plate-tectonic reconstructions of the Greenland Sea, after Reksnes and Vågnes (1985).

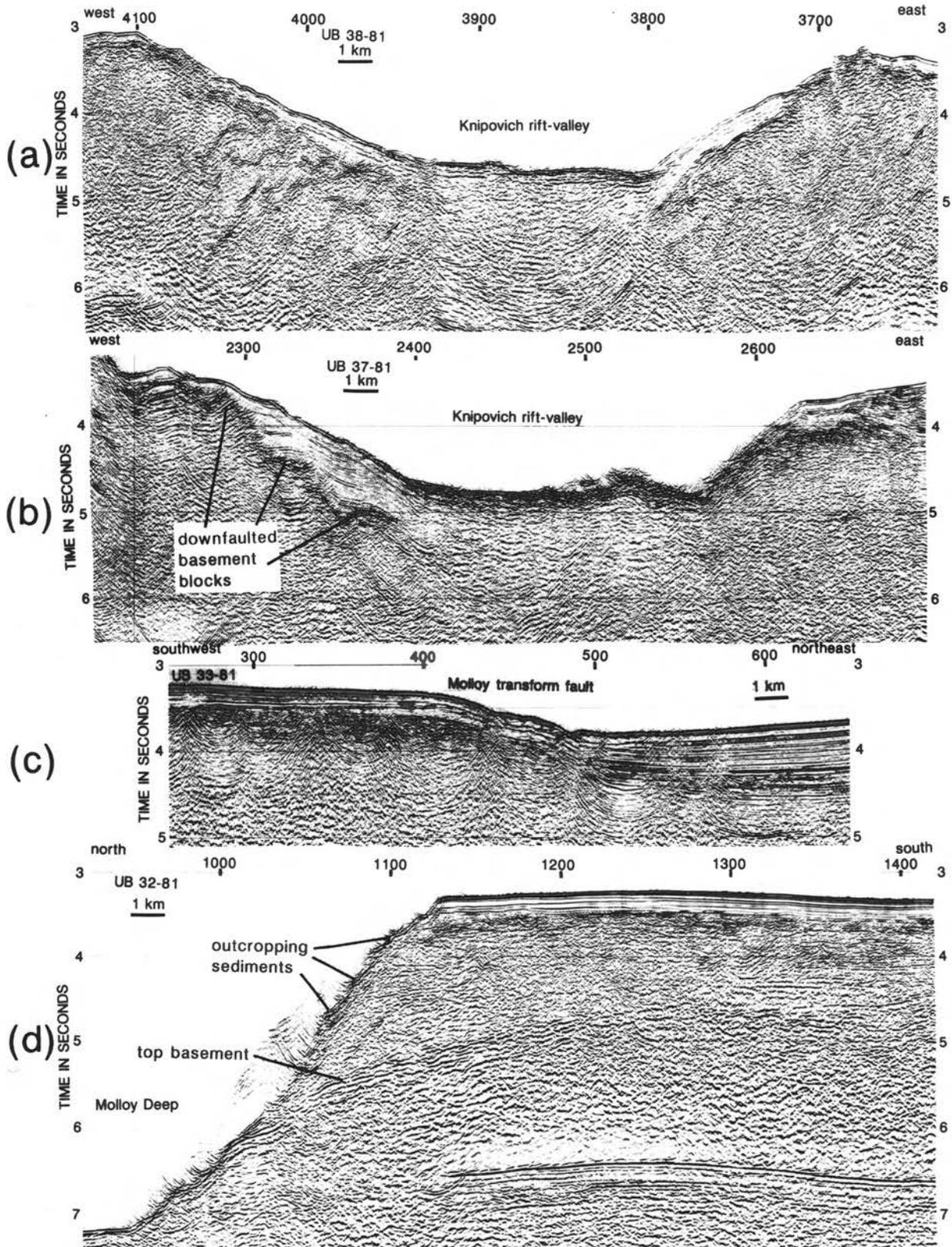


Figure 19: Seismic profiles across the plate boundary. Locations are shown in Fig. 1. a. across Knipovich Ridge (line UB 38-81). b. across Knipovich Ridge (line UB 37-81). c. across Molloy transform Fault (line UB 33-81). d. at the flank of Molloy Deep (line UB 32-81).

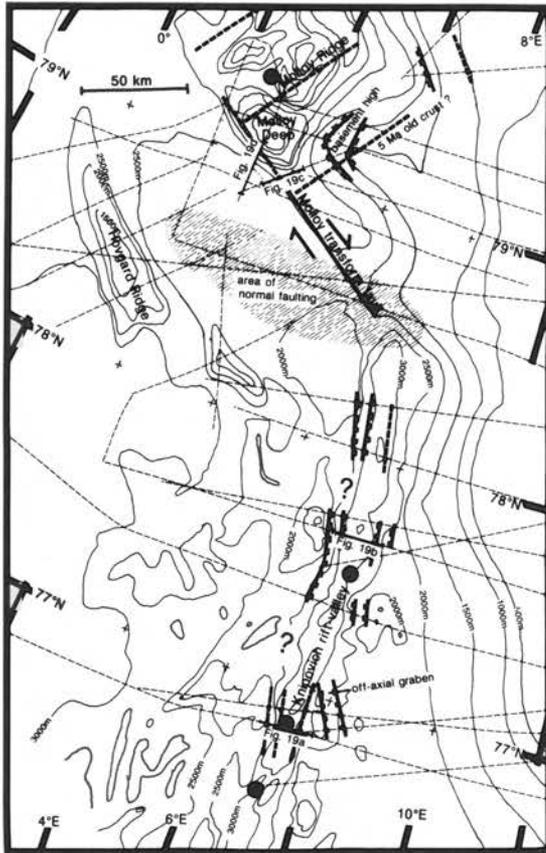


Figure 20: Schematic tectonic map of the plate boundary. Multichannel seismic line coverage (except Soviet data) is marked with hatched lines and dredged basalt probes from Neumann and Schilling (1984) are marked with dots.

sequences changes across the fault (Fig. 19c), and basement is not clearly imaged.

Molloy transform fault projects along the southwestern side of Molloy Deep (Fig. 20). As seen in Fig. 19d, basement is here dipping gently towards the deep, whereas the whole sedimentary section is truncated at the seafloor. The southern wall of Molloy Deep may be a good area for surface sediment sampling of older strata.

The basement high 30 km east of Molloy Deep (Fig. 20 and transect 7) may be a "high inside corner", common at ridge-transform intersections at slow spreading ridges (Severinghaus and MacDonald 1988).

Baturin et al. (1990) suggested from gravity modelling only 2–4 km crustal thicknesses beneath the Molloy Ridge. Seismic reflection data give a chaotic pattern of seafloor diffractions in this region of rough seafloor topography. Farther north there is no seismic data coverage, due to the perennial sea ice cover.

Top basement is frequently faulted and tilted in the crestal mountains of Knipovich Ridge (transect 4). Gradual westward deepening of basement into Boreas Basin can be explained by subsidence of cooling oceanic crust (Fig. 21). If we correct basement depths for the sediment load and compare them with thermal subsidence curves from Parson and Sclater (1977), they are consistent with the empirical depth of basement having an age of 30–60 Ma.

Only few measurements of refraction velocities within oceanic basement exist. Myhre and Eldholm (1981) found from sonobuoy recordings low average velocity, 2.83 km/s, along the Knipovich Ridge and attributed this to mixed sequences of basalt and rubble.

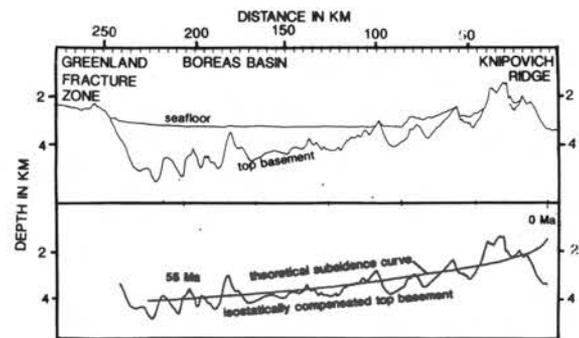


Figure 21: Above: Sediment thickness and depth to basement along line UB 38-81 across Boreas Basin at 76°50'N. Below: Depth to basement, isostatically compensated for weight of the sediments, and compared to a subsidence curve from Parson and Sclater (1977) assuming a crustal age of 55 Ma 200 km west of Knipovich Ridge.

3.2 Post-Caledonian sediments on Spitsbergen

by Arvid Nøttvedt

The Central Spitsbergen Basin is bounded to the west by the exposed west Spitsbergen fold- and thrust-belt and to the east by the Ny-Friesland - East Spitsbergen High (Harland et al. 1974, Nøttvedt et al. 1988a,

Fig. 1). The latter is bounded between the Billefjorden and Lomfjorden Fault Zones. The basin boundaries parallel a dominant N-S to NNW-SSE structural grain on Spitsbergen, inherited through several episodes of similarly aligned tectonic deformation. Caledonian faulting, folding, thrusting and metamorphism of pre-Cambrian to Silurian sediments and igneous complexes formed the Hecla Hoek basement of Svalbard (Ohta

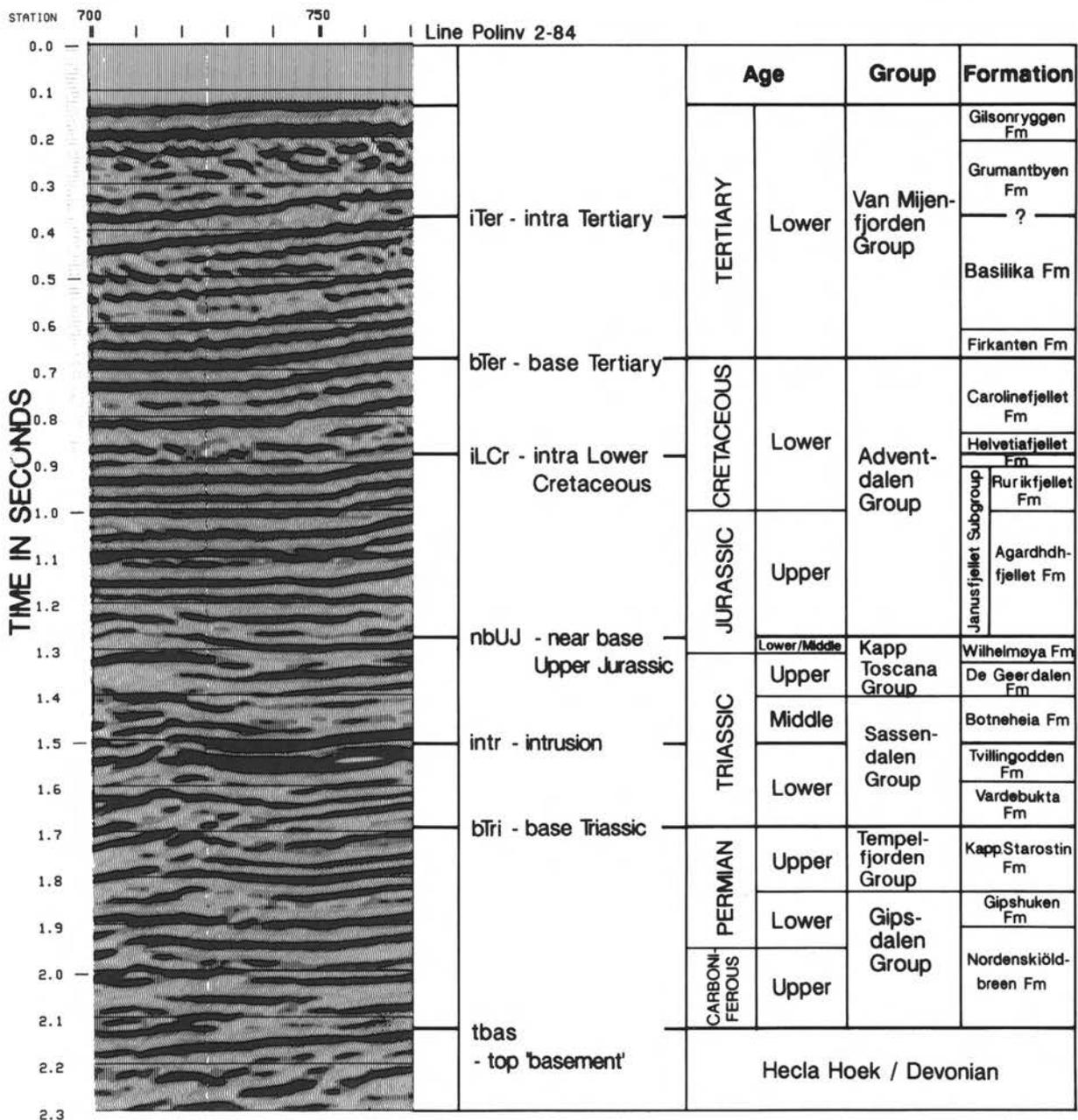


Figure 22: Seismic stratigraphic subdivision of the Mesozoicum in van Mijenfjorden (from line POLINV 2-84).

1992). Hecla Hoek has a stratigraphic thickness in the order of 15–20 km and varying degree of metamorphism and structural complexity. The seismic response from the Hecla Hoek is generally poor with few and scattered reflections visible. A possible exception exists on central-east Spitsbergen and in north Storfjorden (transects 4 and 5) where some strong, dipping reflections down to more than 3 seconds may represent lower Paleozoic sediments. Alternatively, they may be Devonian in age (see below).

A seismic stratigraphic subdivision of the post-Caledonian section is shown in Figs. 22 and Fig. 23, in accordance with datings from Norsk Hydro.

Devonian basins

Following the Caledonian orogeny, Devonian extension and possible shear caused the formation of major north-south trending grabens (Harland et al. 1974, Lamar et al. 1986). Up to 8 km (stratigraphic thickness) of mostly continental redbeds have been reported from two such grabens in

northern Spitsbergen (Friend and Moody Stuart 1972, Tab. 1). These grabens continue southwards below Isfjorden, but transect 5 does not allow any clear definition of their subsurface extent. From other seismic data in eastern Isfjorden (Fig. 24), a steeply dipping reflector down to about 4 seconds is interpreted as the sediment-basement boundary of a Devonian half-graben. Internal reflections are scattered and diffuse, but some sub-horizontal, undulating reflectors can be observed to lap onto this boundary (see also Johannessen and Steel 1992, Nøttvedt et al. 1993a). The reflections down to 3 seconds on central-east Spitsbergen and in northern Storfjorden resemble those in the Devonian in Isfjorden. Upper Carboniferous - Permian strata rest unconformably on Hecla Hoek basement rocks immediately north of these seismic profiles, however, favouring an early Paleozoic origin here.

The Devonian on Spitsbergen underwent late Devonian east-west compression (Lamar et al. 1985) and an angular unconformity marks the boundary to the overlying Carboniferous strata.

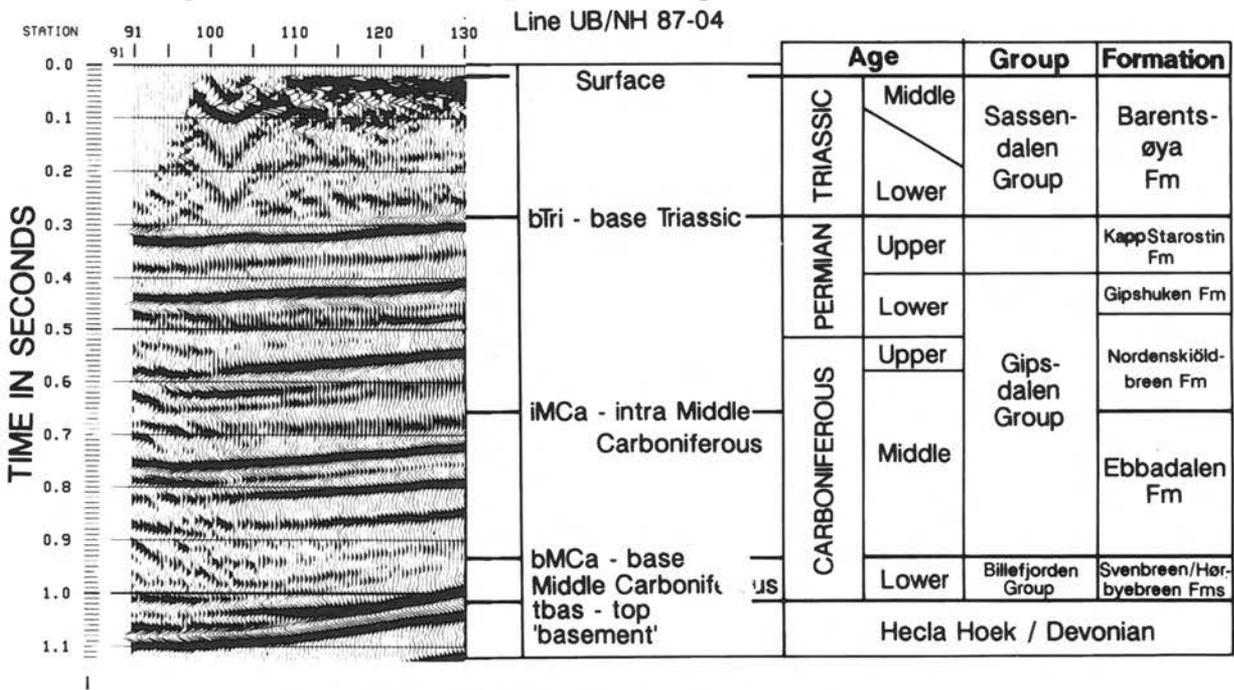


Figure 23: Seismic stratigraphic subdivision of the upper Paleozoic (from line UB 87-04).

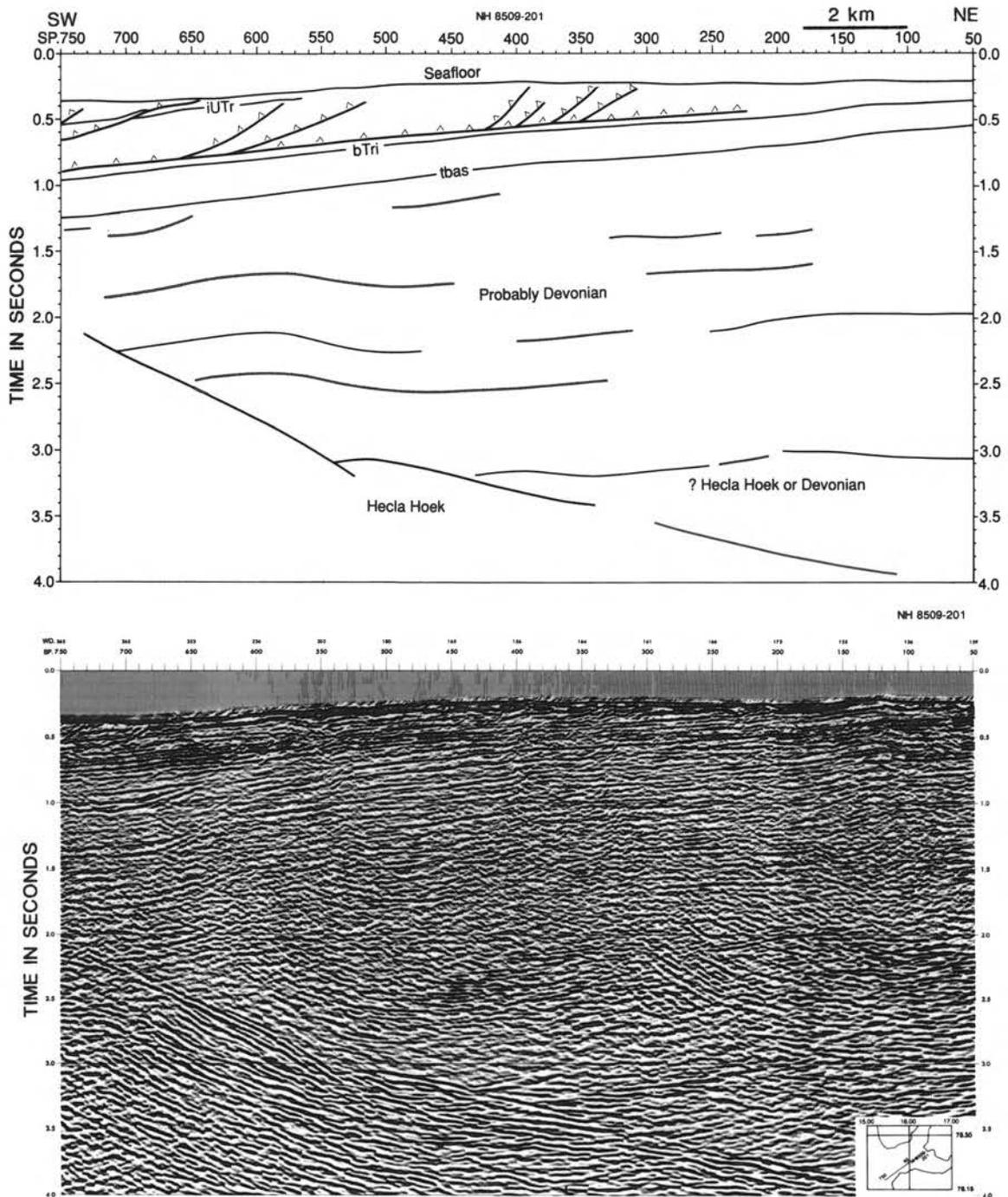


Figure 24: Close up of seismic profile in inner Isfjorden (line NH 8509-201).

Carboniferous - Tertiary basins

The post-Devonian history of Svalbard has been studied in great detail for the past 100 years. Various summary papers include Orvin (1940), Birkenmajer (1981) and Steel and Worsley (1984) on stratigraphy, Harland (1969) and Dallmann et al. (1988, 1992) on post-Caledonian deformation and Nøttvedt et al. (1993a) on the petroleum potential.

The Carboniferous - Tertiary succession on Spitsbergen rests unconformably on Hecla Hoek and on Devonian strata. There are no well log data describing this boundary, but it is thought to be represented by

a fairly strong and continuous reflector separating continuous, parallel reflectors in the Carboniferous - Tertiary from more diffuse reflectors in the Devonian and Hecla Hoek. Five stages of basin evolution are recognisable here, as well as across large parts of the Barents Shelf (Steel and Worsley 1984, Faleide et al. 1984, Worsley et al. 1986, Harland and Dowdeswell 1988, Nøttvedt et al. 1993b):

- 1) Late Devonian - middle Carboniferous rifting.
- 2) Late Carboniferous - Permian carbonate platform establishment.

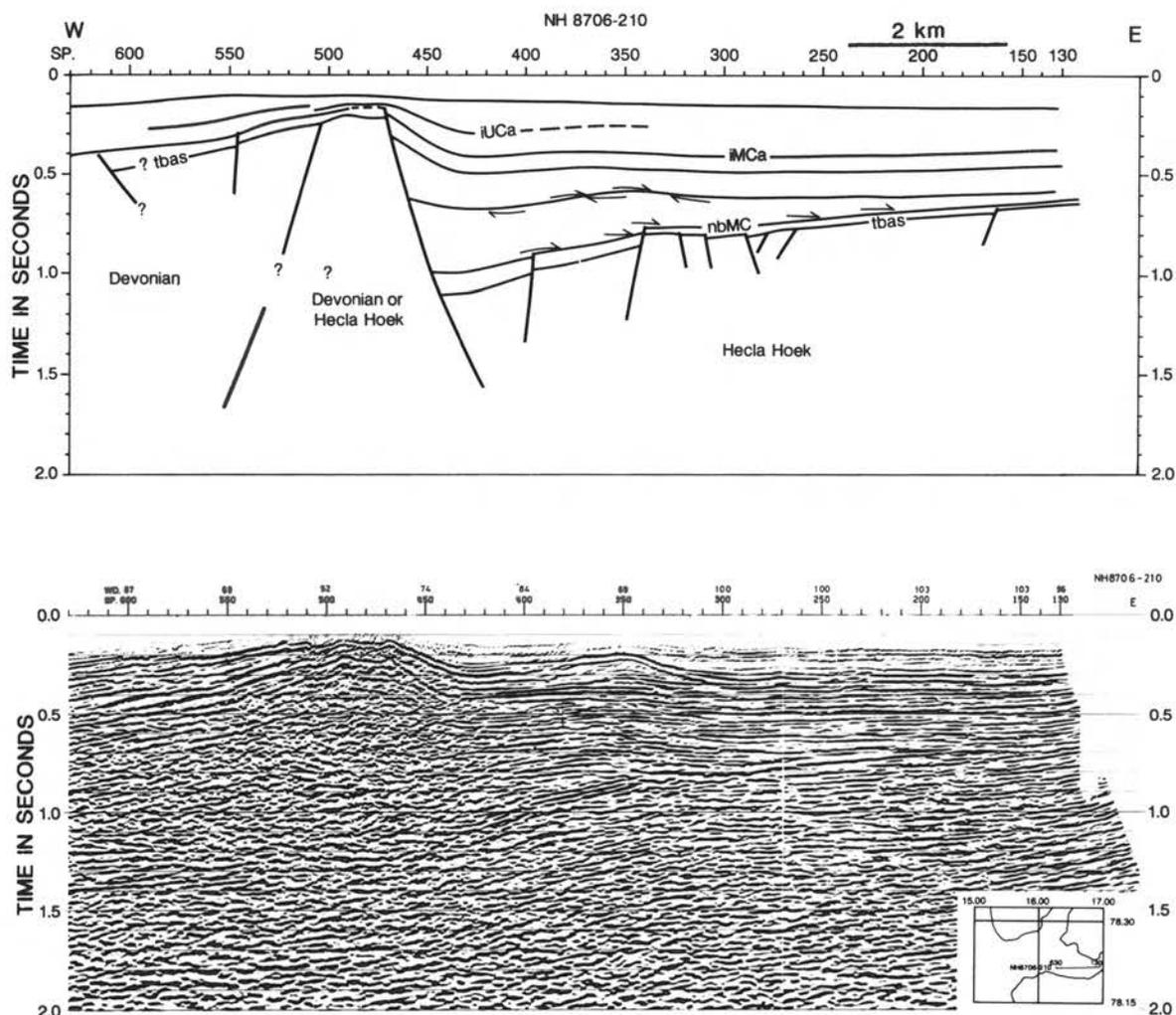


Figure 25: Close-up of seismic profile in Tempelfjorden (line NH 8706-210).

- 3) Triassic - Cretaceous siliciclastic shelf development.
- 4) Early Tertiary crustal breakup.
- 5) Late Cenozoic passive margin development.

Early - middle Carboniferous extension succeeded late Devonian compression. Fault bounded grabens and half-grabens with throws up to 2 km developed—i.e. St. Johnsorden Trough, Inner Hornsund Trough and Billefjorden Graben (Cutbill and Challinor 1965, Gjelberg and Steel 1981, Johannessen and Steel 1992). Lower Carboniferous sediments include interbedded fluvial conglomerates, sandstones, shales and coals, whereas middle Carboniferous is characterized by alluvial redbeds, carbonates and evaporites. Only the Billefjorden Graben has been crossed with seismic lines. It appears as a continuous half-graben east of the Billefjorden Fault Zone from inner Isfjorden to Heer Land, but with diminishing relief. The base of the graben floor drops from 1.0 seconds (about 2.5 km) on Nordenskiöld Land (transect 5) to 1.5 seconds (about 3.5–4.0 km) on Heer Land (transect 4). The bounding Billefjorden Fault Zone is described in more detail below. Lower Carboniferous deposits are commonly less than 0.1–0.2 seconds (250–500 m) thick, and can be identified within the half-graben structure where they thin gently against the basement reflector in an eastward direction (Fig. 25). In Billefjorden, Lower Carboniferous deposits also crop out to the west of the half-graben. In Fig. 25 is shown a tentative interpretation of some downfaulted Lower Carboniferous strata to the west of the Billefjorden Fault Zone in inner Isfjorden is shown. Middle Carboniferous strata reaches up to more than 0.5 seconds (1.2 km) in thickness within Billefjorden Graben in inner Isfjorden (Fig. 25) and has a characteristic wedge shaped geometry. Internal reflectors lap down onto Lower Carboniferous strata (Fig. 25). This downlap surface is thought

to represent the basal progradational surface for alluvial fan redbeds, and is taken therefore as the base middle Carboniferous. From correlation against outcrops, a strong and continuous reflector around 0.5 seconds on line UB 87-05 is believed to be intra middle Carboniferous in age, possibly near the top of Ebbadalen Formation. The boundary to Upper Carboniferous is lithologically indistinct and difficult to define in the seismic data. The internal reflection pattern changes from weak, discontinuous-chaotic near the boundary fault to strong, sub-parallel, continuous up the half-graben flank. This is interpreted to reflect the lateral facies change from alluvial redbeds to carbonates and evaporites.

In late Carboniferous - Permian time, most of the Svalbard archipelago became drowned and a series of carbonates and evaporites overlain by cherts accumulated (Steel and Worsley 1984). Sonic log velocity data from the Grumant 1 and Ishøgda 1 wells show a marked velocity increase across the Triassic - Permian boundary, reflecting the passage from Triassic shales to Permian cherts. The corresponding seismic reflector is generally easily recognizable. No precise datings exist for the reflectors below. Upper Carboniferous - Permian strata commonly varies between 0.2 and 0.4 seconds (0.5–1.1 km) in total thickness and seismic reflectors are generally sub-parallel and continuous. They can be mapped sub-regionally and dated against outcrops. A near base upper Carboniferous reflector has been tentatively interpreted in seismic data from inner Isfjorden (Fig. 25). The unit is little deformed except along the western flank of the Central Spitsbergen Basin where it is repeated on several imbricated thrust plates below the Mesozoic cover.

Stable conditions prevailed during Triassic - Cretaceous times, but sedimentation was now characterized by siliciclastic shelf aggradation. Repeated deltaic-marine coarsening upward sequences from shales to

siltstones and sandstones in the Triassic and early Jurassic pass into more homogeneous offshore marine shales and back into deltaic - shallow marine sandstones and siltstones with coals in late Jurassic - early Cretaceous (Dypvik 1980, Mørk et al. 1982, Steel and Worsley 1984, Dypvik et al. 1991). Sonic log velocity data from the Grumant 1 and Ishøgda 1 wells show a significant downwards velocity increase across the Upper-Middle Jurassic boundary, from Adventdalen Group shales to Kapp Toscana Group sandstones. A weak corresponding reflector is present in the seismic data, but this reflector is sometimes difficult to follow, especially in tectonized areas. A somewhat smaller downwards velocity increase occurs in intra Barremian or near top Festningen Member, giving rise to a moderately strong reflector that can be followed on the seismic data. Finally, there is a slight downwards velocity decrease across the base Tertiary boundary, reflecting the passage from Tertiary sandstones to Lower Cretaceous siltstones and sandstones. A moderately strong corresponding reflector is present in the seismic data from central Isfjorden and central van Mijenfjorden. The Triassic - Cretaceous unit is commonly between 0.6 and 0.8 seconds (1.4–1.8 km) thick, but may reach up to 1 second (2.2–2.4 km) in areas of tectonic thickening. The unit shows abundant internal deformation. It served as the main decollement unit for the eastward translation of compression from the exposed fold- and thrust-belt into and across the Central Spitsbergen Basin. The deformation structures and kinematics are described in detail below.

As a consequence of opening of the Arctic and North Atlantic Oceans, Svalbard sheared against Greenland from Paleocene through early Oligocene. The shearing was accompanied by initial warping (Paleocene) and subsequent compression (Eocene) along the plate boundary (Harland 1969, Lowell 1972, Kellogg 1975, Steel et al. 1981, 1985,

Vågnes et al. 1988, Müller and Spielhagen 1990). Tertiary deposits include thick, interbedded shale and sandstone units of deltaic and shallow marine origin (Kellogg 1975, Steel et al. 1981). The Tertiary is a little more than 0.5 seconds (1.1–1.2 km) thick in van Mijenfjorden (transect 4) and appears to be relatively little deformed.

Billefjorden and Lomfjorden Fault Zones

The Billefjorden and Lomfjorden Fault Zones have a complex tectonic history of both extension and compression, probably from Precambrian through Tertiary (Orvin 1940, Parker 1966, Harland et al. 1974, Kellogg 1975, Haremo and Andresen 1988, 1992, Haremo et al. 1990, Andresen et al. 1992). The surface expression of the Billefjorden Fault Zone changes along strike from a set of large, basement involved normal faults in the middle Carboniferous section in Billefjorden, to a large backthrust in the Upper Carboniferous - Permian strata at Gipshuken and Flowerdalen, to a pair of smaller thrust faults and ramp anticlines in the Jurassic - Tertiary section between Adventdalen and Kjellstrømdalen (Gjelberg and Steel 1981, Ringset and Andresen 1988, Haremo et al. 1990). From structural field data and seismic data these structures are clearly uncoupled vertically (Haremo and Andresen 1988, Haremo et al. 1990, Nøttvedt et al. 1988b). The normal fault, or fault zone, in Billefjorden (Gjelberg and Steel 1981) involves lower - middle and partly upper Carboniferous strata. It cannot be traced in surface outcrops south of Pyramiden, but is clearly present in seismic data from inner Isfjorden (Fig. 25, line NH 8706-210). The backthrust at Gipshuken probably sole out in middle - upper Carboniferous evaporites eastwards in the Billefjorden Graben and is likely to have been forced by a facies change from evaporites to conglomerates close to the graben mar-

gin (Ringset and Andresen 1988). This backthrust in turn forced ramping from a decollement surface in the Jurassic Janusfjellet Subgroup to create the partly oversteepened anticlines between Adventdalen and Kjellstrømdalen. The relationship between normal faulting and backthrusting is demonstrated on line UB 87-03 sp. 350–400 (Fig. 26). Moreover, the surface anticlines that can be observed in the mountains next to the line do not seem to involve the Triassic and Permian strata, as seen on the line. This suggests that they are rooted in a decollement level basically in the Jurassic.

The Lomfjorden Fault Zone is developed as a basement rooted reverse fault in northern Spitsbergen (Nøttvedt et al. 1988b) passing into a large monocline with superimposed small-scale thrust faulting from a Mesozoic decollement in Agardhdalen (Andresen et al. 1992). The small-scale thrusting predates the folding which is probably coupled in the deep to the reverse fault seen further north. Seismic line NH 8802-12 crosses the fault zone at sp. 200–300, but it is difficult to interpret (Fig. 6). A gentle, east-facing monocline is interpreted as a low amplitude, asymmetrical fold above a

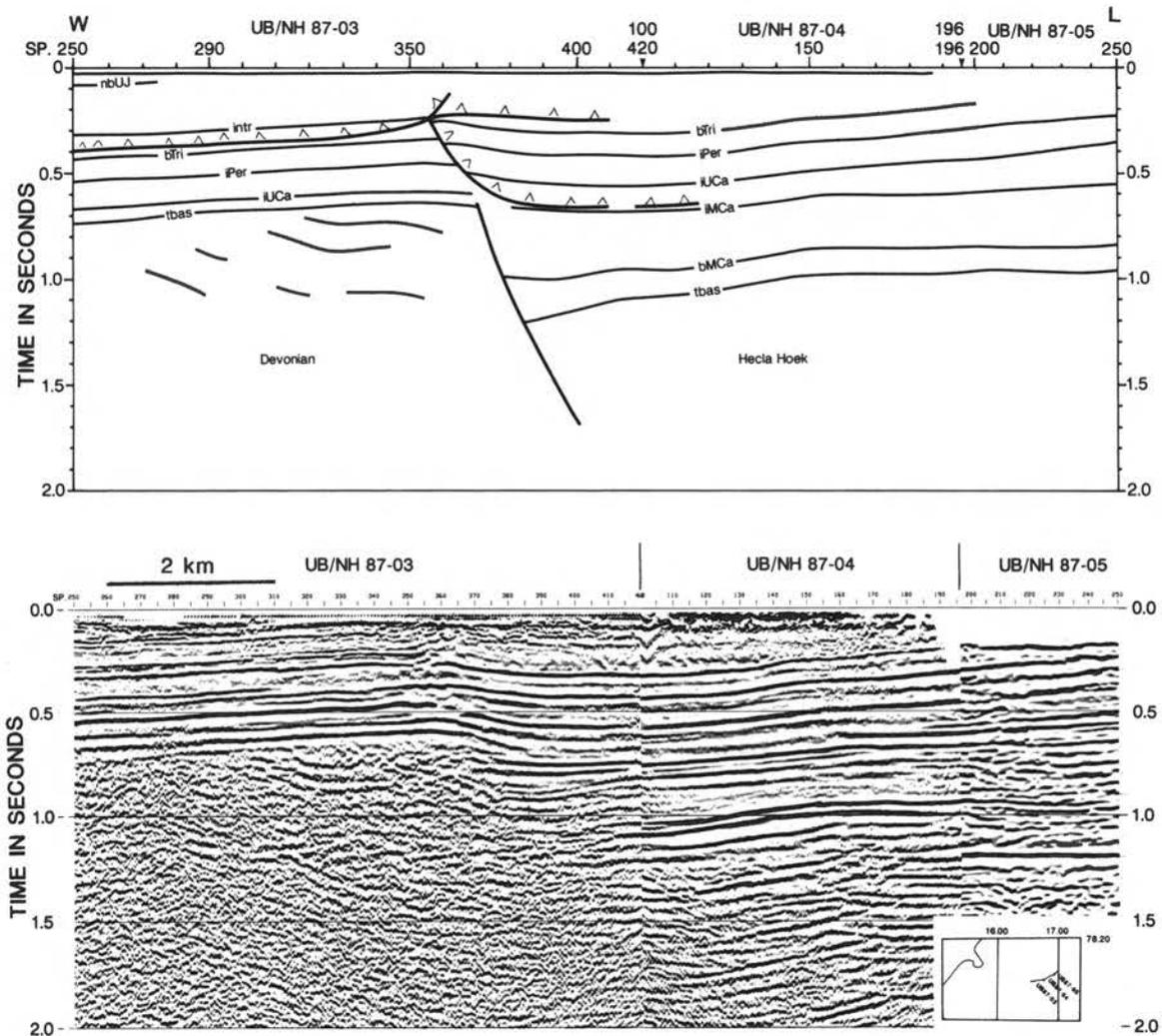


Figure 26: Close-up of seismic profile in Adventdalen (line UB 87-03/04/05).

deeper basement fault. An east-dipping reverse fault is observed to the west of the monocline (sp. 200). In innermost Sassendalen, a west-facing, open, monoclinial fold (200–300 m) is present in hill exposures immediately south of the trace of the seismic line (Nøttvedt et al. 1988b). The monocline involves Permo-Triassic beds and is interpreted as a blind fold above the east-dipping reverse fault. This fault is interpreted to represent an antithetic fault to a main west-dipping reverse fault strand controlling the formation of the gentle east-facing monocline.

Tertiary compressional structures

Tertiary compressional structures include a complex suite of thrust wedges and larger thrust nappes in west Spitsbergen to small-

scale thrust ramps and duplex structures from regional decollement levels in central-east Spitsbergen. The style of structuring is controlled both by distance from the fold- and thrust-belt and by the stratigraphy involved in the deformation. The larger thrust nappes occur in brittle basement and upper Paleozoic carbonate lithologies and represent the transitional thick-skinned to thin-skinned part of the fold- and thrust-belt proper. Fig. 27 shows a segment of line NH 8509-204 from outer Isfjorden, demonstrating some details of the deformation. Several low-angle thrust faults have been interpreted, imbricating basement and Upper Carboniferous - Permian stratigraphy (see also Nøttvedt et al. 1993b). The section has not been balanced, but several kilometers of shortening may be envisaged. More duc-

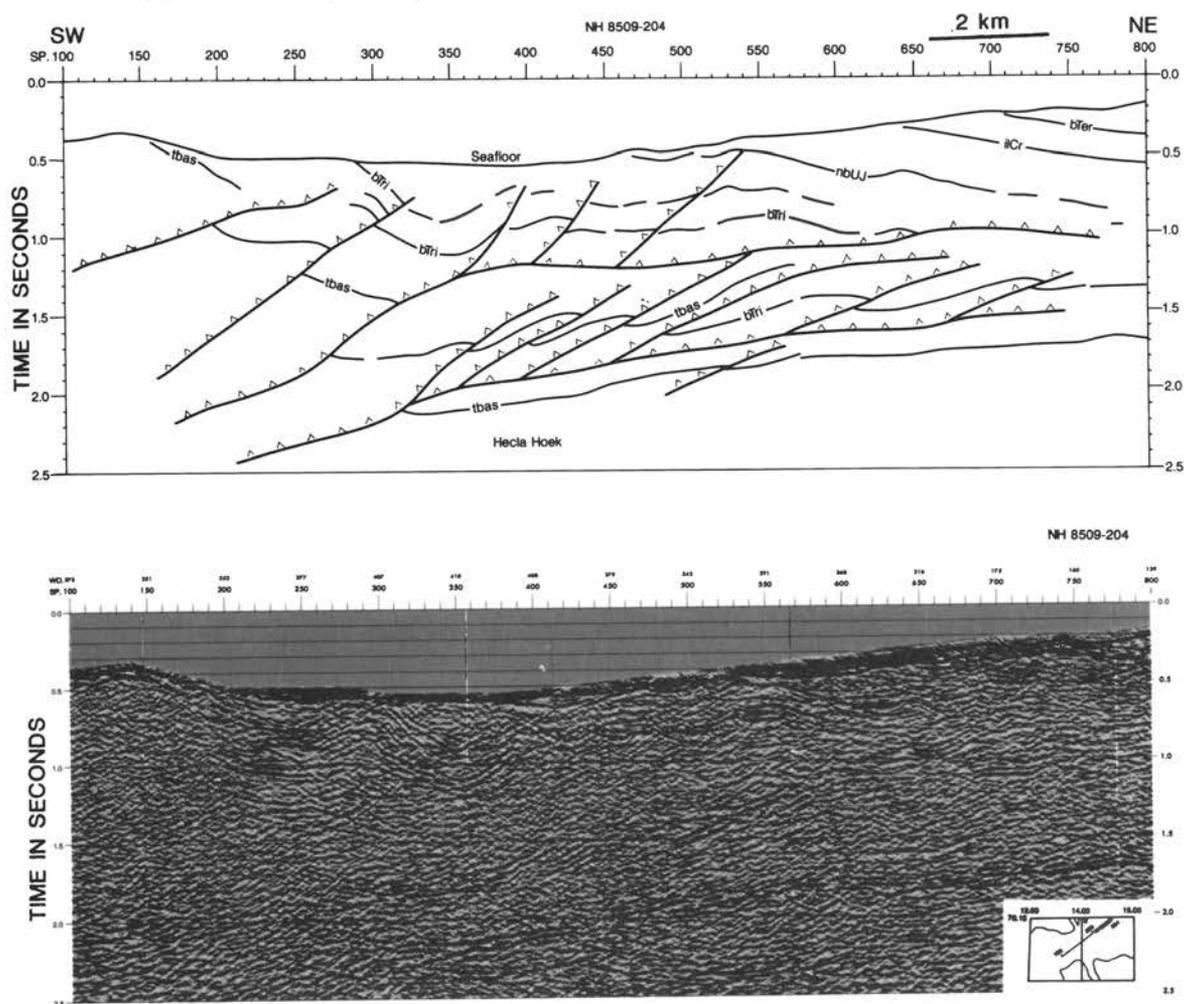


Figure 27: Close-up of seismic profile in western Isfjorden (line NH 8509-204).

tile lithologies in the Mesozoic favoured the transformation of compressional strain along extensive decollement surfaces. Extensive fracturing at several levels in the black shales of the Janusfjellet Subgroup and Sassendalen Group in both Grumant 1 and Ishøgda 1 wells confirm the presence of such decollement surfaces (Nøttvedt and Rasmussen 1988). Generally speaking, the main decollement appears to climb southwards, from within the Sassendalen Group in Isfjorden to within the Janusfjellet Subgroup in van Mijenfjorden (transects 4 and 5). This is suggested to be in response to the pre-existing late Cretaceous tilting of Spitsbergen. These decollement levels have been further mapped in outcrops to the east

(Haremo et al. 1990, Andresen et al. 1992), suggesting that they are continuous across most of central Spitsbergen. An example of this deformation is shown in Fig. 28. The detailed line segment is from line NH 8509-202 in central Isfjorden and shows small-scale thrust faults and ramp anticlines originating from a decollement in the Sassendalen Group. Duplex structures within the Mesozoic succession are sometimes present and may cause local thickening of the strata. Recent work suggest that much of the thickness variations described from outcrops have a similar origin and that previously inferred Mesozoic faulting is negligible (Haremo and Andresen 1988, Andresen et al. 1992).

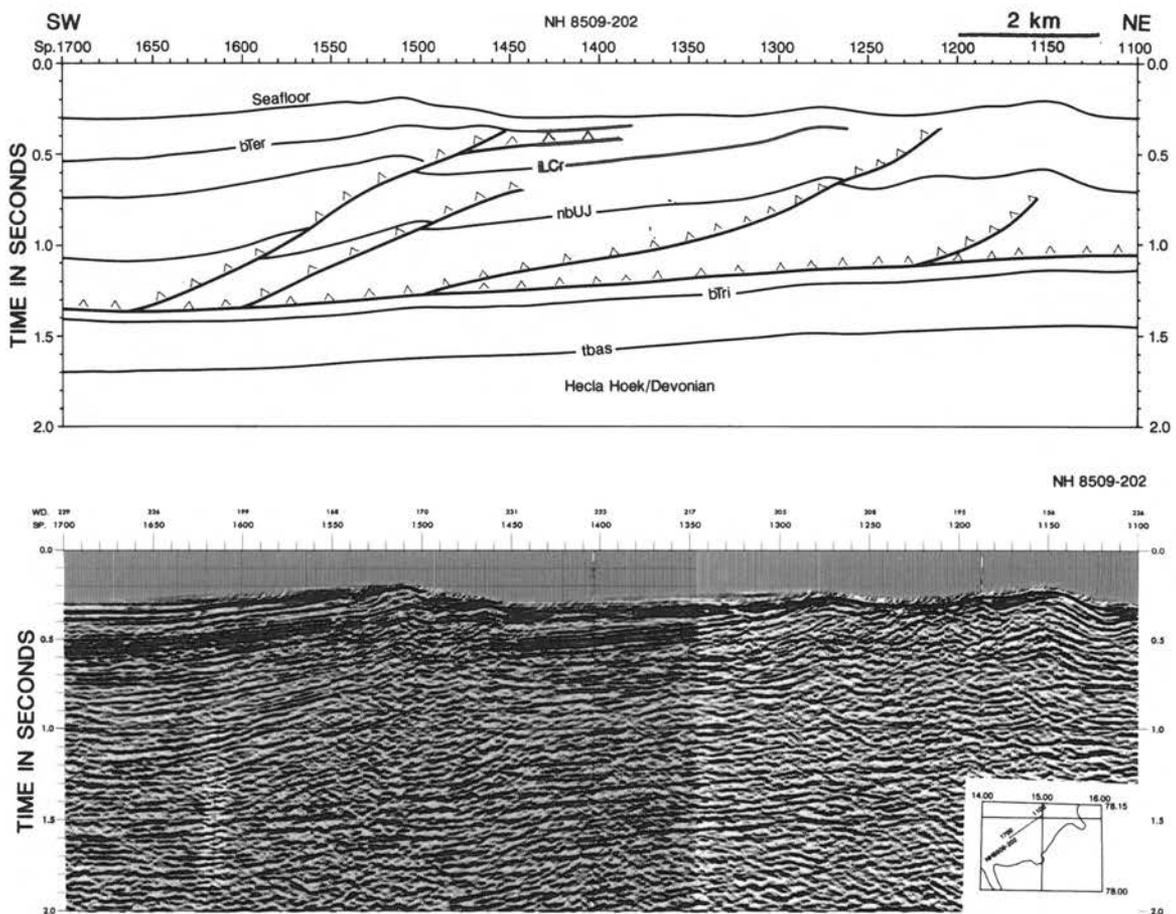


Figure 28: Close-up of seismic profile in central Isfjorden (line HN 8509-202).

3.3 Sediments west and north of Svalbard

by Ola Eiken

Western margin

East of Hornsund fault complex several grabens and half-grabens are present (Eiken and Austegard 1987, Fig. 1, transects 3,4,5 and 7), up to 20–30 km wide and 4 km deep, and offset with respect to each other. Only a coarse outline of the structures can be obtained from the present data coverage. The graben fill is weakly deformed and believed to be of Tertiary age (Birkenmajer 1972, Kellogg 1975, Eiken and Austegard 1987), similar to Forlandsundet Graben (Atkinson 1962, Steel et al. 1985, Kleinspehn and Teyssier 1992, Gabrielsen et al. 1992) and the Tertiary outcrops at Renarodden and Øyrlandet. The presence of older sediments, may be of Carboniferous age, in the lower part of the grabens cannot be excluded (Townsend and Mann 1989). The relationship between graben formation and the Tertiary fold-and thrust-belt is debated, and models include post-orogenic collapse, a strike-slip releasing bend, and extension during continental break-up (Steel et al. 1985, Kleinspehn and Teyssier 1992, Gabrielsen et al. 1992).

West of Hornsund fault complex, a section of up to 4–6 km thick sediments of presumably Oligocene and younger ages is present below the outer shelf and slope (Fig. 29). A close-up showing characteristic reflection patterns within sequences SPI-I, SPI-II and SPI-III is shown in Fig. 30. The

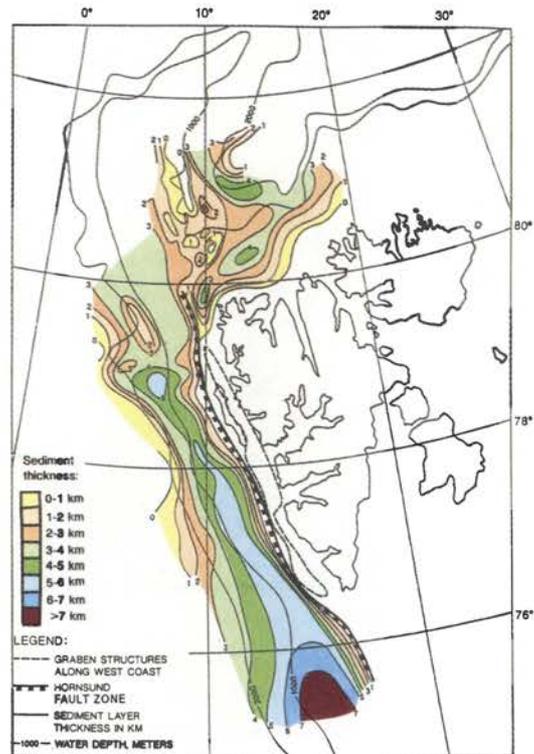


Figure 29: Sediment thickness map in km along the western and northern Svalbard margin, from Austegard and Sundvor (1991).

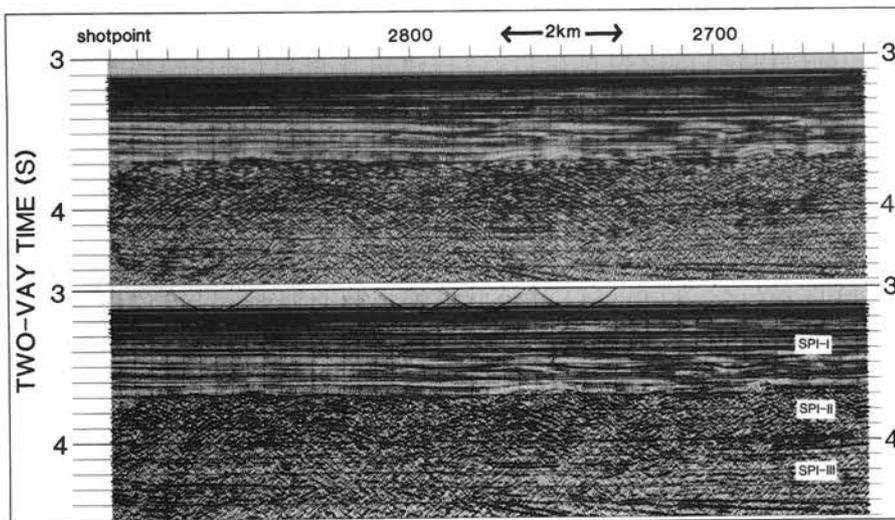


Figure 30: Unmigrated and migrated portion of profile UB 38-81. Sequence SPI-I shows migrating sediment waves with amplitudes 10–20 m and in-line wavelengths of 3–4 km, while sequence SPI-II shows abundant diffractions.

thickness and extent of the slump-related (Schlüter and Hinz 1978), shale diapirism-related (Eiken and Austegard 1987) or current-formed (Eiken and Hinz 1993) sequence SPI-II north of 76° N is shown in Fig. 31.

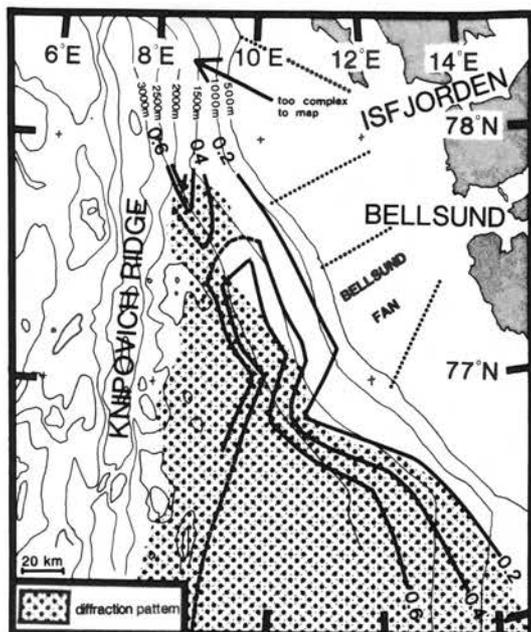


Figure 31: Isopach map of sequence SPI-II west of Spitsbergen in seconds reflection time. The areal distribution of the diffraction pattern is marked with dots. From Eiken and Hinz (1993).

The deep-water sediments are generally 1–3 km thick, with greatest thickness near the continental margins (Vogt 1986a). Hemipelagites and turbidites are probably volumetrically dominating. Bottom current influence on sediment deposition was reported by Sundvor et al. (1982b) on the Yermak Plateau and by Kristoffersen (1990a)

also around Molloy transform fault. In addition to the transects, an example of this is the section across the eastern slope of Hovgård Ridge shown in Fig. 32, where layers within a sediment drift are truncated at the seafloor in a characteristic pattern and a moat is present at the base of the slope. A thick contourite sequence has been deposited in lee of the ridge. At the current shaped Vestnesa (transect 7), Eiken and Hinz (1993) attributed a change from a N-S elongated depocenter within their sequence YP-2 to a circular depocenter within sequence YP-3 (Fig. 33) to a change from oppositely directed N-S trending currents to a persistent eddy above the depocenter. Strong currents in Fram Strait are related to exchange of warm and cold water masses through Fram Strait (Aagard et al. 1985, Thiede et al. 1990b), and this exchange probably did not become effective until late Miocene (Eiken and Hinz 1993).

Northern margin

The Yermak Plateau, roughly defined by the 800 m depth contour, is linked to the Svalbard margin through the 4–5 km deep Danskøya basin (transect 8). Most or all of the sediments in the basin are of late Cenozoic age. The sediment cover is generally between 1 km and 3 km on the surveyed part of the Yermak Plateau (Fig. 29).

The northern shelf is about 50–60 km wide west of Sjuøyane and widens eastward. The about 400 m deep Hinlopenrenna cuts

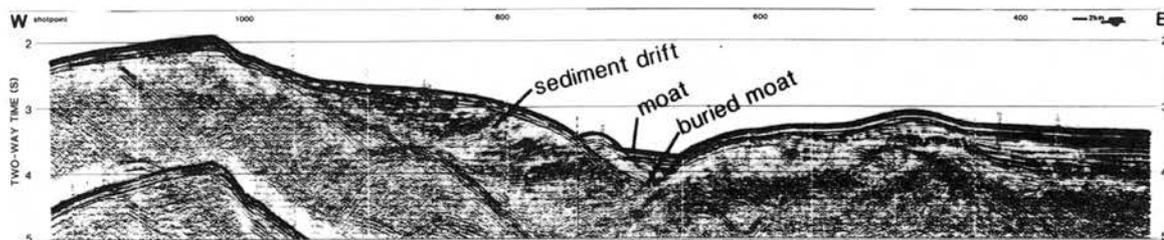


Figure 32: Seismic line across the eastern slope of Hovgård Ridge (line UB 26-81) showing bottom current influence. Location is shown in Fig. 1.

through the 100–200 m deep bank areas. Only few seismic lines exist, and the geology is less known than at the western margin. Also, the dissimilarities between transects 11, 12 and 13 makes it difficult to establish geological models. The margin has prob-

ably a longer history of subsidence and deposition than the western margin, due to earlier continental separation, but a large part of the sediments may be of late Cenozoic age (Table 1).

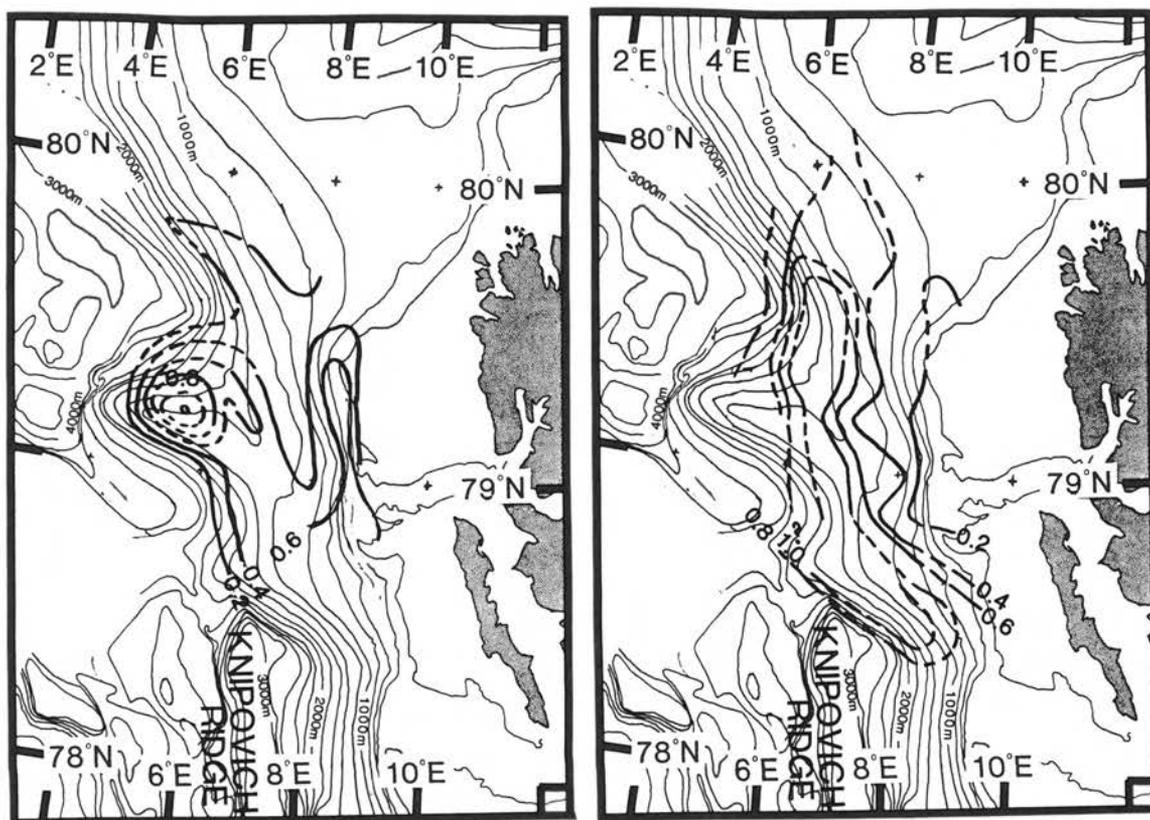


Figure 33: Isopach map (thick lines) of sequence YP-3 (left) and the upper part of sequence YP-2 (right) on Vestnesa in seconds reflection time. From Eiken and Hinz (1993). Thin lines show water depth in meters.

3.4 Glacial deposits on the western Svalbard margin

by *Anders Solheim*²

Glacial history

The glacial history of Svalbard and the adjacent shelf areas has been a matter of discussions for a number of years. Until recently, most interpretations were based on land work, and therefore, most of the discussion has been on the last glacial period, the Weichselian (the last 120 000 years), and in particular on the Late Weichselian glaciation which had its maximum around 18–20ka. Based on raised shorelines Schytt et al. (1968) proposed the existence of an extensive Late Weichselian ice sheet which covered the entire area, including the Barents Sea, and which extended all the way to the shelf edge. Boulton (1979a,b) claimed that the Late Weichselian glaciation was concentrated in the central and eastern parts of Svalbard, and that the ice cover did not reach significantly beyond the present-day coasts of the archipelago. Based on several lines of evidence from data collected through the last decade, a major, grounded ice sheet seems to have covered most (if not all) of the Barents Sea continental shelf (Elverhøi and Solheim 1983, Solheim and Kristoffersen 1984, Vorren and Kristoffersen 1986, Vorren et al. 1988, Elverhøi et al. 1990, Solheim et al. 1990).

From terrestrial data, Mangerud and Svendsen (1992) have concluded that the west coast of Spitsbergen (Fig. 1) has been glaciated three times during the last 120 000 years. Although some field studies have indicated that parts of the west coast remained ice free during the Late Weichselian (Salvigsen and Nydal 1981, Miller

et al. 1989), other studies have concluded that the ice extended at least to the present-day coast line (Mangerud et al. 1987). Recent data from the continental shelf west of Spitsbergen, however, indicate that the Late Weichselian ice sheet may have covered the entire shelf and extended to the shelf edge, where also moraine ridges have been identified (Fig. 34) (Solheim et al. 1991). The ice sheet over Svalbard was contiguous with the Barents Sea ice sheet, but the situation along the north coast of the archipelago remains unknown due to lack of data.

As the preservation potential for older (pre-Weichselian) deposits is low on land, information about earlier glaciations must be sought in the offshore regions, where an archive of glacial - interglacial variations is stored in the sediments. Both in the deep sea and on the outer parts of the continental shelf, we infer from seismic data that thick glacial deposits are present. A main problem in these regions, however, is the lack of geological samples to provide stratigraphic tie points for the interpretation of the seismic data. Only one drillhole exists off the Svalbard margin (Deep Sea Drilling Project Site 344, transect 3). The results from this drilling indicate glacial conditions at least since the Late Miocene or Early Pliocene (i.e. 5–7 Ma) (Talwani, Udintsev et al. 1976). These results are supported by the more recent Ocean Drilling Program Leg 104 at the Vøring Plateau off Norway. Input of ice-rafted detritus at this location appears to have commenced at about 5.5 Ma (Jansen and Sjøholm 1991). However, the exact source area for the ice-rafted detritus around the Norwegian-Greenland Sea is uncertain. Hence, only the very latest part of the glacial history has been interpreted and discussed on the Svalbard margin, but the seismic geometry on the outer parts of the margin strongly indicates the repeated action of waxing and waning ice sheets.

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The distribution and character of the glacial deposits

Approximately 60% of Svalbard is covered by glaciers and large amounts of glacial sediments are being produced and delivered through meltwater streams to the marine environment. Most of these sediments are deposited in the fjord basins where relatively thick layers of fine grained deposits accumulate. Depositional rates of 10 cm/year have been measured in the inner parts of Kongsfjorden (Elverhøi et al. 1983), but the rate decreases rapidly to values significantly lower than 1 mm/year farther along the fjord. In general, the thickness of postglacial sediments in the fjords varies from a few meters to a few tens of meters, depending largely on the areal cover-

age of glaciers in the drainage area and the dynamic behaviour of the glaciers. There are clear indications that the glacial coverage and also the output of glacial erosional products have varied significantly through the Holocene. The maximum glacial extension and the highest output of glacial sediments to the fjords occurred during the Little Ice Age (Sexton et al. 1992). Although this period is poorly defined in Svalbard, it probably lasted approximately 4-500 years, up to the end of the last century. Although the fjord basins also contain thin layers of till, representing subglacial conditions, they mainly act as temporary storage for the sediments. During a glacial advance, most of the fjord sediments are probably reworked and redeposited farther out on the continental shelf and slope.

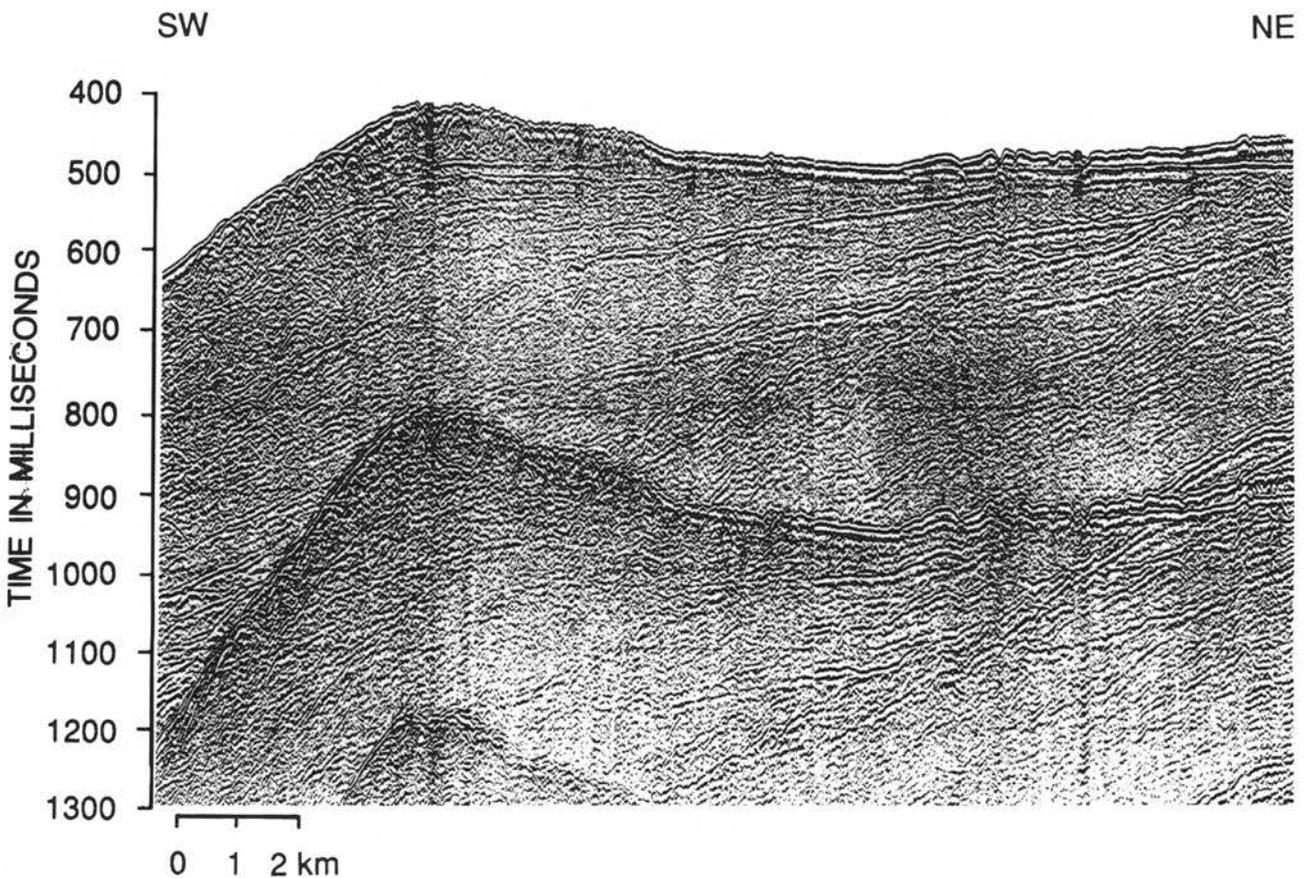


Figure 34: Part of single channel seismic line NP 90-119 showing moraine ridge at the shelf edge. For location, see Fig. 1. (From Solheim et al. 1991).

Most of the seismic data acquired on the Svalbard margin until the last few years, have been conventional multichannel data, with vertical resolution generally limited to approximately 20 m. The distribution of sediments within the upper few hundred meters below the sea floor have therefore been only coarsely mapped on the shelf. Particularly on the inner shelf, east of the Hornsund fault complex, where the thickness of unlithified sediments is the thinnest, strong sea-bottom multiples have been another problem which hampers the quality of the seismic data. During the last 3–4 years, high resolution seismic data have been acquired (Table 2). The Dutch Geological Survey and the University of Edinburgh acquired both single- and multi-channel data in 1988, mainly in the area from Isfjorden and northwards, while the Norwegian Polar Institute carried out single channel seismic profiling and piston-coring off Isfjorden and Bellsund in 1990. In addition to this, multichannel seismic data of relatively high resolution was acquired jointly by the Norwegian Polar Institute and the University

of Bergen, using a commercial seismic vessel, M/V "Mobil Search", outside Bellsund in 1987. The following is based on data from the shelf and slope off the Isfjorden - Bellsund area, but it is considered representative for the Svalbard margin in general.

The inner shelf is characterized by a thin (< 20m) cover of unlithified sediments overlying the bedrock. The boundary is easily defined as a regional angular unconformity (Fig. 35). On the middle and outer shelf there is a steady increase in the thickness of unlithified sediments which form a thick wedge under the outer shelf and the upper continental slope (Figs. 36 and 37). The thickness of glacial sediments may exceed 2 seconds (about 2.5 km, using velocities within the range shown in Fig. 41) as inferred from deep seismic reflection and refraction surveys off Isfjordrenna (Myhre and Eldholm 1988, sequence SPI-I). These authors suggest a glacial thickness reaching 1.8 seconds (about 2.2 km) below the upper slope off Isfjorden.

The thickness of glacial sediments be-

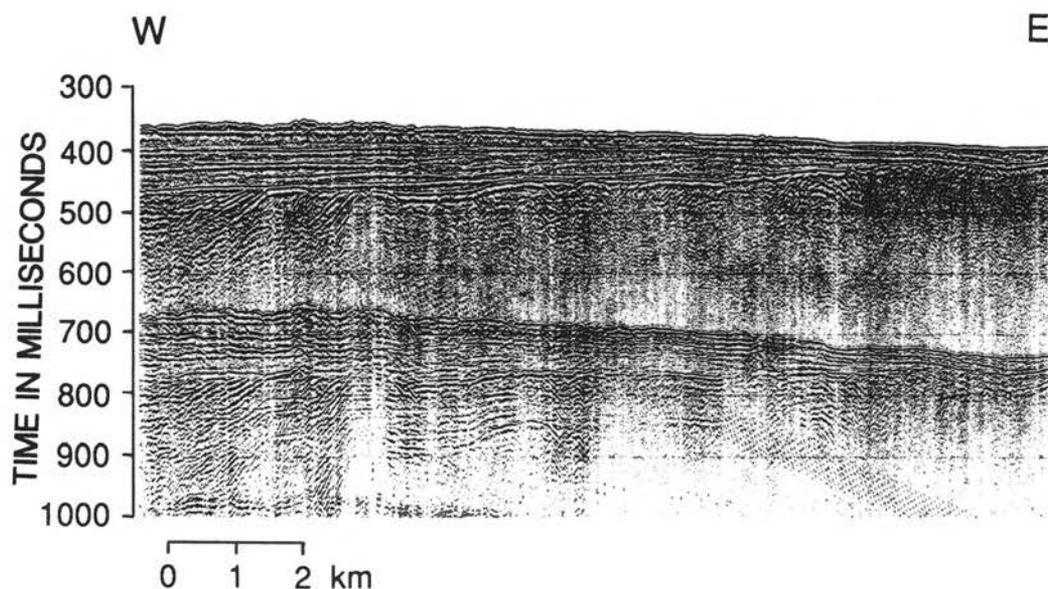


Figure 35: Part of single channel seismic line NP 90-114 showing the increase in thickness of glacial sediments (above the angular unconformity) from the inner to the outer shelf. For location, see Fig. 1. (From Solheim et al. 1991).

low the outer shelf and slope have not been confirmed by drilling, and the identification of the preglacial/glacial boundary in the seismic records is not straightforward. The boundary does no longer form a distinct angular unconformity, as in the inner parts of the shelf. Estimates of glacial thicknesses are based on unconformities and changes in the acoustic character of the seismic sequences, interpreted to reflect changes in the depositional environment. Schlüter and Hinz' (1978) interpretation of the youngest sequence, SPI-I, as glacially influenced deposits was based on a marked change in character across unconformity U_1 , from a chaotic reflection pattern in the seawards thickening sequence SPI-II to a more regular character with subparallel to divergent internal reflections in the seaward thinning sequence SPI-I. Later, Eiken and Austegard (1987) defined a shallower unconformity on the shelf (U_0), and proposed this to mark the base of glacial erosion.

High resolution data from the "M/V Mo-

bil Search" investigations in 1987 allow us to define six seismic sequences, B1–B6, in the area off Bellsund (transect 4). Sequence B3 overlies the chaotic sequence B2 (equivalent to SPI-II), but thins rapidly westwards. The three upper sequences, B4–B6, have more continuous internal strata and thin more gradually westwards than B3. This change in depositional pattern, across an unconformity, d , which coincides with U_0 of Eiken and Austegard (1987) on the shelf, is interpreted to mark the onset of glacially influenced sedimentation. Based on this stratigraphic interpretation, the thickness of glacial sediments off Bellsund exceeds 2 km.

In general, the upper, strongly prograding sequences have been interpreted to be glacially influenced. These sequences have also been tentatively correlated with B4–B6 further to the north of the Isfjorden-Bellsund region. In the northern regions, however, the assumed glacial sequences are generally less than 1.0 seconds in total

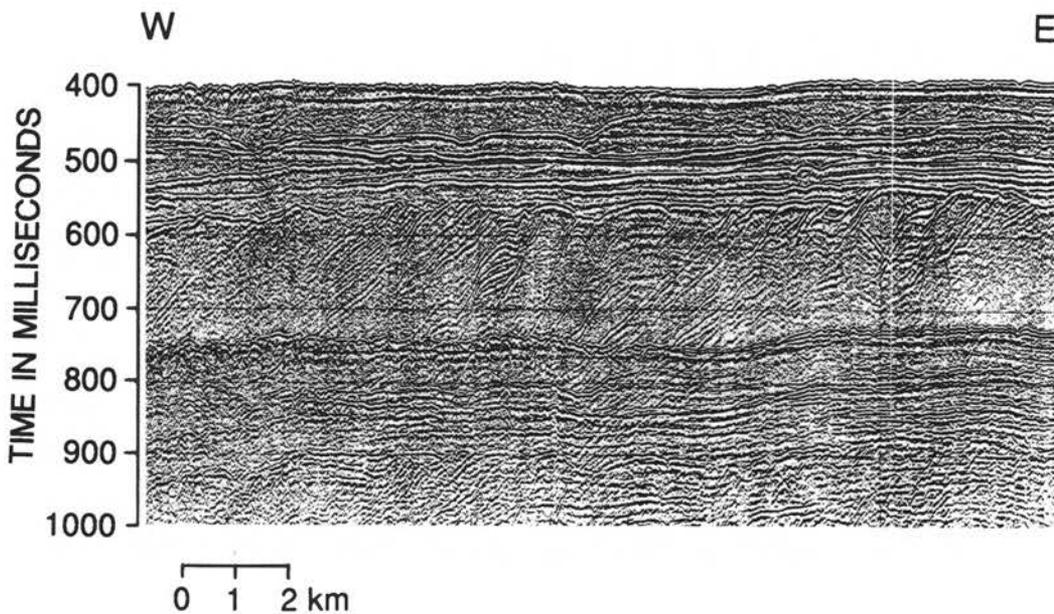


Figure 36: Part of single channel seismic line NP 90-131 showing increased thickness of glacial sediments (above the distinct angular unconformity at 500-600 mseconds). Local erosional surfaces are seen in the glacial sequence. For location, see Fig. 1. (From Solheim et al. 1991).

thickness. The exact onset of glacial sedimentation and hence the thickness of glacially influenced sediments can, however, only be determined by future drilling.

Large transverse bathymetric troughs form the submarine continuation of the major fjord systems (Fig. 1). These troughs most likely contained ice streams which drained large parts of the ice sheet during the glacial periods, and at least the last part of their evolution is caused by glacial erosion. The bulk of the glacial erosional products was probably transported out by these ice streams and, during glacial maxima, deposited directly at the shelf break. In the region off Isfjorden and Bellsund we observe a change from one depo-center for seismic sequence B3, to two for sequence B6. This demonstrates the increased influence of the fjords and transverse shelf trough systems for drainage of glacial erosion products through the Plio-Pleistocene. The enhanced supply of material by the ice streams is seen in the large fan complexes built out in front of the troughs. In the fan

complex off Isfjorden (Fig. 38), the main down-slope transport mechanism seems to be small scale slumping where individual slump bodies are 1-2 km across (Solheim et al. 1991). This observation is in accordance with a high deposition rate at the shelf edge.

Towards the lower continental slope and the deep sea, the slump complexes seem to grade into turbidites, interbedded with ice-rafted glacial marine sediments, which also formed the major part of the sediment drilled in Deep Sea Drilling Project Site 344 (Talwani, Udintsev et al. 1976). In some regions, particularly to the north, where the present spreading axis approaches the Svalbard margin, the glacial sediments overflow directly into the Knipovich spreading axis valley.

The interpretation of the seismic stratigraphy from the western Svalbard margin indicates that the onset of a glacial regime marked a large increase in the sedimentation rates and that a significant amount of the sediments on the margin are glacially influenced. This is in accordance with re-

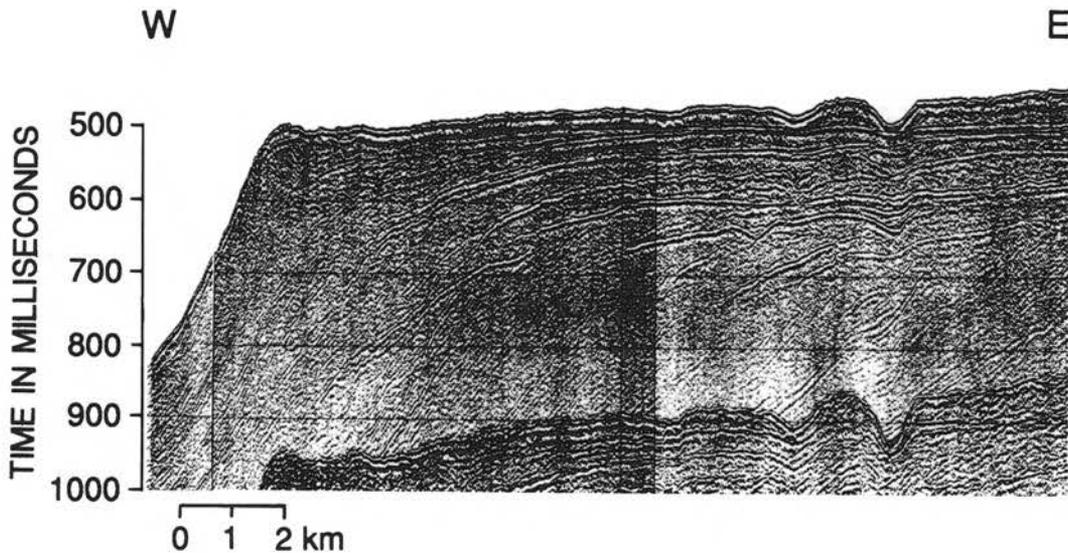


Figure 37: Part of single channel seismic line NP 90-131 from the outer shelf. The base of the glacial sequence can no longer be readily defined. Note erosional features at the sea floor. For location, see Fig. 1. (From Solheim et al. 1991).

sults from the major sedimentary fan off the southwestern Barents Shelf, the *Bjørnøya Fan*, where a Late Pliocene age have been found at approximately 1100 m subbottom depth in two boreholes (Eidvin and Riis 1989). This implies that a considerable part of the Tertiary erosion of the Barents Shelf is glacial (> 1000m). The same may apply to Svalbard, where Manum and Throndsen (1978) suggested a Tertiary erosion of at least 1.7 km based on vitrinite reflectance, and Eiken and Austegard (1987) proposed nearly 3 km of average erosion from volu-

metric estimates of the offshore sediments. A rapid uplift must have occurred to maintain the relief necessary for such high rates of glacial erosion. With regards to the Barents Shelf, there are indications that the last (Late Weichselian) glaciation, which covered a relief like the present, may have had only limited erosional power (Nyland Berg 1991). Hence, during the main part of the glacial period, the Barents Shelf may have had a topographic relief, caused by uplift, which was significantly different from the present.

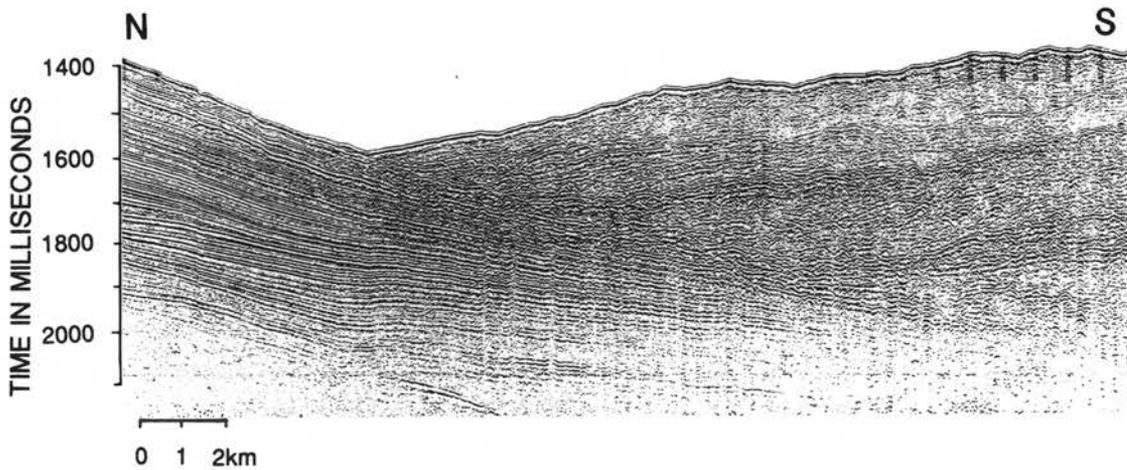


Figure 38: Part of single channel seismic line NP 90-303 crossing the northern boundary of the fan off Isfjordrenna. The line is roughly a strike-line following mid-slope depths. Note the marked fan build-up, with a more chaotic reflection pattern relative to the layered sequence dipping under the fan from the north. For location, see Fig. 1. (From Solheim et al. 1991).

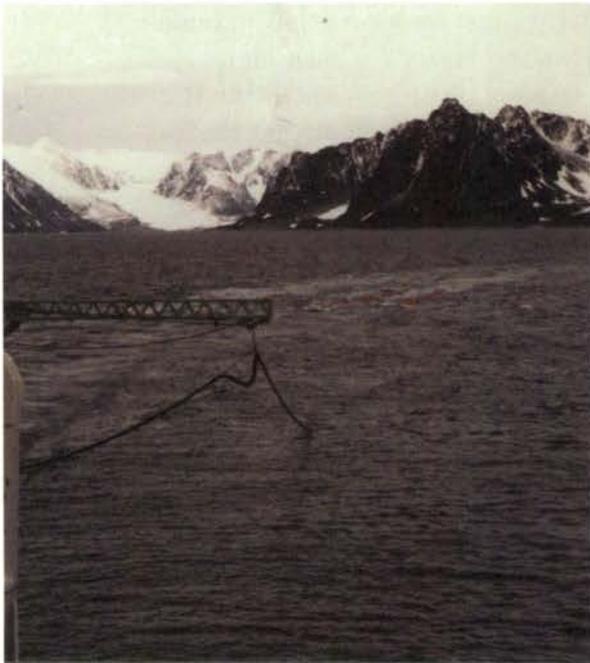


Figure 39: Seismic field work onshore and offshore Svalbard.

Chapter 4

Aspects of the seismic method

by Ola Eiken

4.1 Field operations in Svalbard

The quality of the seismic sections have improved much over the last two decades. Technological developments include better and more repeatable sources, a larger number of receivers and recording channels and better processing software. Several surveys offshore Svalbard have been shot by state-of-the-art instrumentation in the 1980's. Lines have been acquired with up to 3000 m long hydrophone cables in narrow fjords such as Van Mijenfjorden, Van Keulenfjorden and Tempelfjorden. Deep-crustal reflection profiles have been obtained with a 4500 m long cable and the powerful source array of M/V "Mobil Search", and the data quality compares favourably with other continental offshore areas. Fair seismic coverage has been obtained with normal marine operations north to the limit of minimum sea ice cover, which is about 82° N north of Spitsbergen in good summers. In the perennially ice-covered areas farther north, only small amounts of data have been acquired (Kristoffersen and Husebye 1985). Future acquisition techniques may here be ice-breakers and mobile units on the ice (Chalmers 1990, Jokat et al. 1992, Kristoffersen et al. 1992) and technological improvements are expected in years to come.

Onshore, commercial surveys on glaciers were conducted by British Petroleum/-Horizon in 1985 and 1986 and by Statoil/Geco in 1986, with good-quality data. Ice proved to be a good wave-propagating medium, and the ice-rock interface did not set up too much noise. However, the costs of these surveys were large. Transportation to and within Svalbard, and the severe climate makes seismic operations expensive. A breakthrough in onshore data acquisition came with the snowstreamer technique developed by University of Münster/University of Bergen/Norsk Hydro in 1986–1988 (Eiken et al. 1989). The snowstreamer consists of gimbaled geophones attached to a main cable and towed on the snow behind a tracked vehicle. Norsk Hydro was since 1988 able to shoot lines at a fraction of the cost of previous surveys, using detonating cord at the surface as source, and advancing up to 3 km per hour !

Data from the transition zone between land and water have been obtained by operating on the fjord ice during winter (British Petroleum and Norsk Hydro). Vibrators have been used on Svalbard only in small-scale experiments by Norsk Hydro/University of Bergen.

4.2 Seismic velocities

Sediment velocities are much higher beneath the inner shelf and within the Sval-

bard archipelago than what is normal for sediments at such depths. Compressional wave velocities are mostly above 4 km/s even close to the surface, and velocities above 6 km/s are often observed within Permian - Carboniferous carbonates (Fig. 40). The latter velocities are close to the highest values measured in sedimentary rocks (Gardner et al. 1974) and reflect their high densities (Kurinin 1965), low porosities and their degree of diagenetic alteration (Grønlie 1978, Elverhøi and Grønlie 1981). These rock properties are results of earlier greater depth of burial. The outer shelf and deepwater sediments have more normal velocity distributions, and on the basis of velocity differences between Tertiary sediments at the outer shelf and in central Spitsbergen, Eiken and Austegard (1987) estimated a Cenozoic uplift of 1-2 km at the inner shelf grabens and 3 km in central Spitsbergen. This is supported by similar estimates made from vitrinite reflectance (Manum and Throndsen 1978).

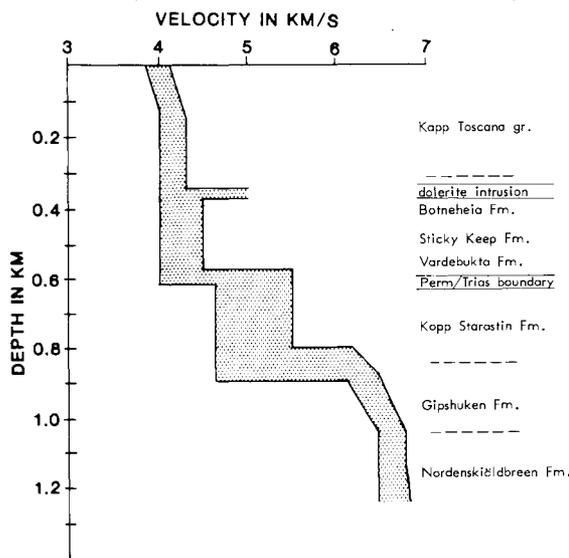


Figure 40: Range of possible velocity-depth distributions obtained from both refractions and reflections in Agardhdalen, eastern Spitsbergen, and their geological correlation (from Eiken 1985).

Myhre and Eldholm (1981) found velocity-depth gradients on the outer western

shelf which were higher than elsewhere on the shelves around the Norwegian - Greenland Sea, and they attributed this to coarse clastic material and high degree of lithification caused by high thermal gradients. Eiken and Austegard (1987), on the other hand, found lower velocity gradients in the same region (Fig. 41), more in correspondence with observations from other areas.

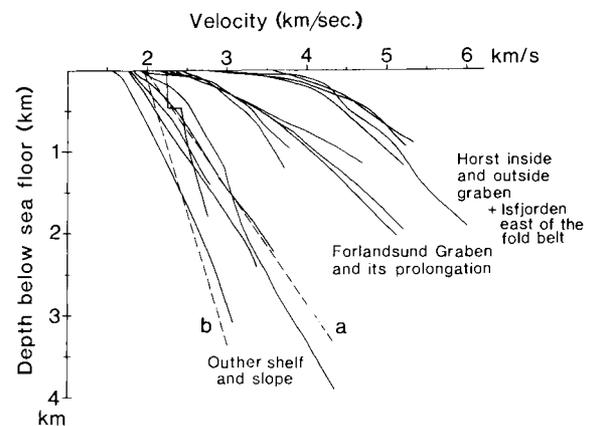


Figure 41: Velocity-depth functions derived from sonobuoys at the western Svalbard continental margin, grouped into three areas: The outer shelf and slope, inner shelf and Isfjorden (from Eiken and Austegard 1987). Also Myhre and Eldholm's (1981) gradient for the shelf 74.5° - 80° N (a) and for other shelves around the Norwegian-Greenland Sea (b) are shown.

Seismic responses to different velocity distributions are illustrated in three marine common midpoint gathers in Fig. 42. On the continental slope, first-arrival refractions have low velocities, and several reflections can be followed across the gather to large offsets. In the fjords (Fig. 42c), first arrival refractions have velocities starting at about 4 km/s and increasing to 6 km/s. Multiple refractions and reflections dominate the record. These areal differences require different acquisition and processing parameters for optimal imaging (shotpoint interval, maximum offset, mute, velocity analysis, etc.).

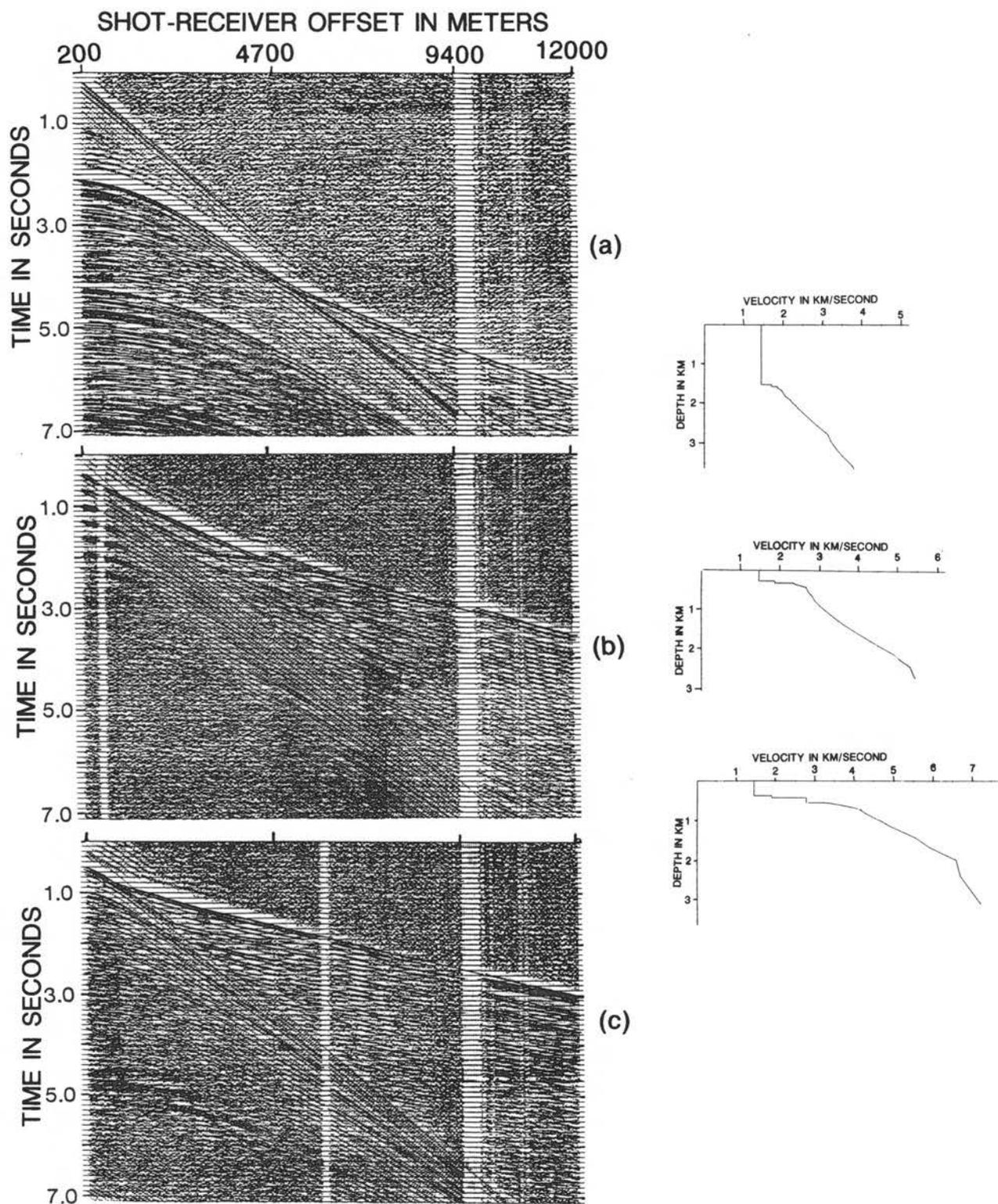


Figure 42: Three marine common midpoint gathers with shot-receiver offsets up to 12 km, recorded during two-ship wide-aperture profiling. Amplitudes have been scaled in sliding windows. These examples are from transect 5, line SVA 4-87 at sp. 1260 (a), 2740 (b) and 4010 (c). Velocity-depth functions, shown to the right, have been derived from first arrival refractions (from Evertsen 1988).

4.3 Near-surface sources of coherent noise

Onshore: Unconsolidated sediments, permafrost and surface waves

An example of a land-seismic shot profile is shown in Fig. 43. Note the first-arrival P-wave refractions with velocities increasing from around 3.5 km/s at near offsets to 5.5 km/s at far offsets, S-wave refractions

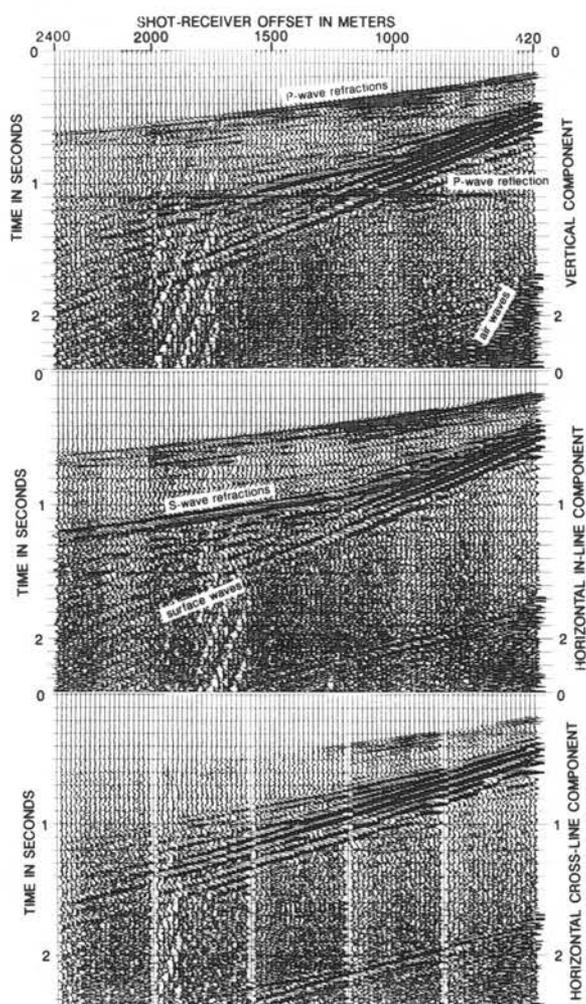


Figure 43: Example of a land seismic shot-profile from Adventdalen, Spitsbergen, recorded with single 3-component geophones (from Bru-land et al. 1988). Surface dynamite point source was used, and amplitudes have been scaled in sliding windows.

(on the horizontal in-line component) with velocities around 4.5 km/s at large offsets, surface waves with velocities around 1.3 km/s and a strong P-wave reflection at 1.1 seconds (on the vertical component).

Unconsolidated Quaternary sediments are present in the broad valleys above the high-velocity early Tertiary and older rocks. The permafrost extends down to 100–400 m beneath the surface (Liestøl 1976), with increasing thickness away from the coast. Most of the unconsolidated sediments are thus usually frozen, and P-wave velocities are then around 4 km/s, close to those of the consolidated rocks beneath. These sediments cause therefore little distortion of the wavefields passing through. Hence, onshore Spitsbergen there is little need for complicated static corrections.

In many nearshore and central parts of the valleys one observe inverse dispersion of the surface waves, and I explain lower velocities of the lower frequencies with a deeper, unfrozen zone of the Quaternary (Fig. 44).

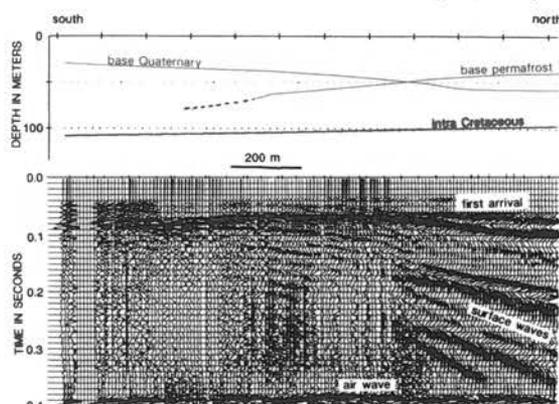


Figure 44: A profile with 130 m common shot-receiver offset, from southern into central Adventdalen (modified after Iversen 1986). Note the later arrival-times for surface waves on the right side of the profile, interpreted as an effect of unconsolidated sediments which are unfrozen and have low velocities.

Snow and ice

Seismic data acquisition during late winter allows easy transportation of equipment along the lines, and is both environmentally and economically a sound approach. Furthermore, the data quality does not seem to suffer if we record on top of the snow: A thin snowlayer attenuate upgoing seismic waves very little, as shown in Fig. 45. The velocity of wind-packed snow has been measured at 1.2–1.4 km/s (Eiken et al. 1989), while the velocity structure of the several meters thick and several years old snow on the glaciers is poorly known.

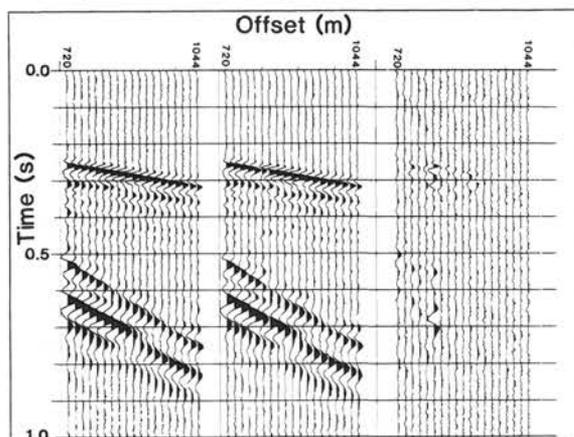


Figure 45: Comparison of recordings with geophones placed above (left) and below (middle) a snow-layer of 0.5 m thickness (from Eiken et al. 1989). The difference is shown to the right.

Waterlayer multiples

High-amplitude waterlayer multiples cause poor signal-to-noise ratios in marine reflection surveys within the Svalbard archipelago, especially for target depths between 0.5 and 5 km. Seafloor reflection coefficients do in places exceed 0.5 and multiples dominate the recorded wave field completely (Fig. 46). These are particularly difficult to suppress where water depths are more than 150 m and predictive deconvolution is less effective (e.g. Isfjorden and Storfjordrenna). Acceptable signal-to-noise ratios have in some cases been obtained with long

streamers (3000 m) and short hydrophone group intervals (12.5–25 m), and with including in the stack wide angle reflections which are obscured by multiple refractions and reflections (Fig. 42c), instead of muting this part of the records, as usually done in other shelf areas.

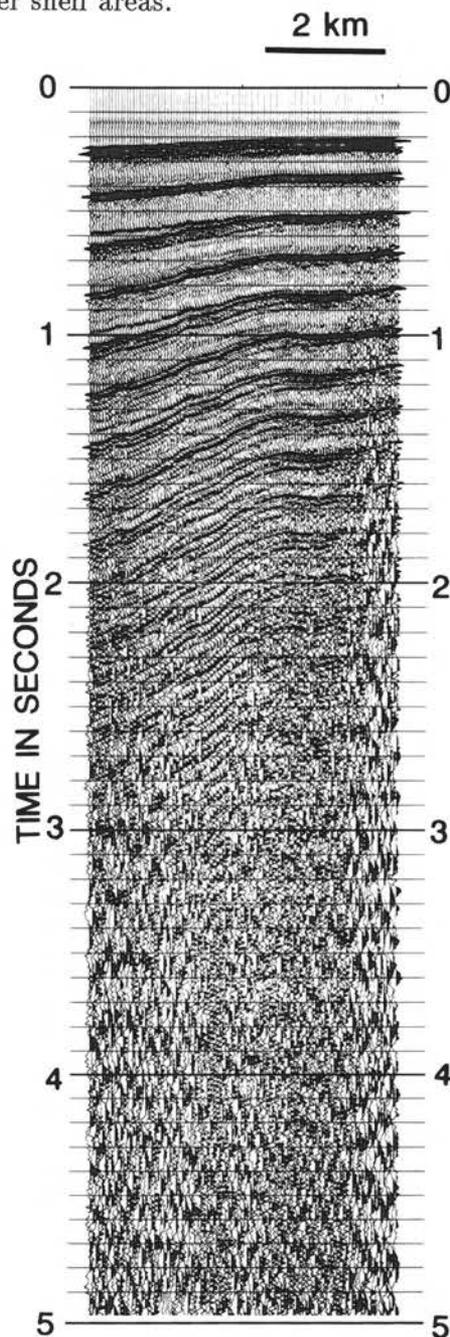


Figure 46: Example of a neartrace section from Storfjorden, showing the seafloor reflection and at least 10 waterlayer multiple reflections.

Chapter 5

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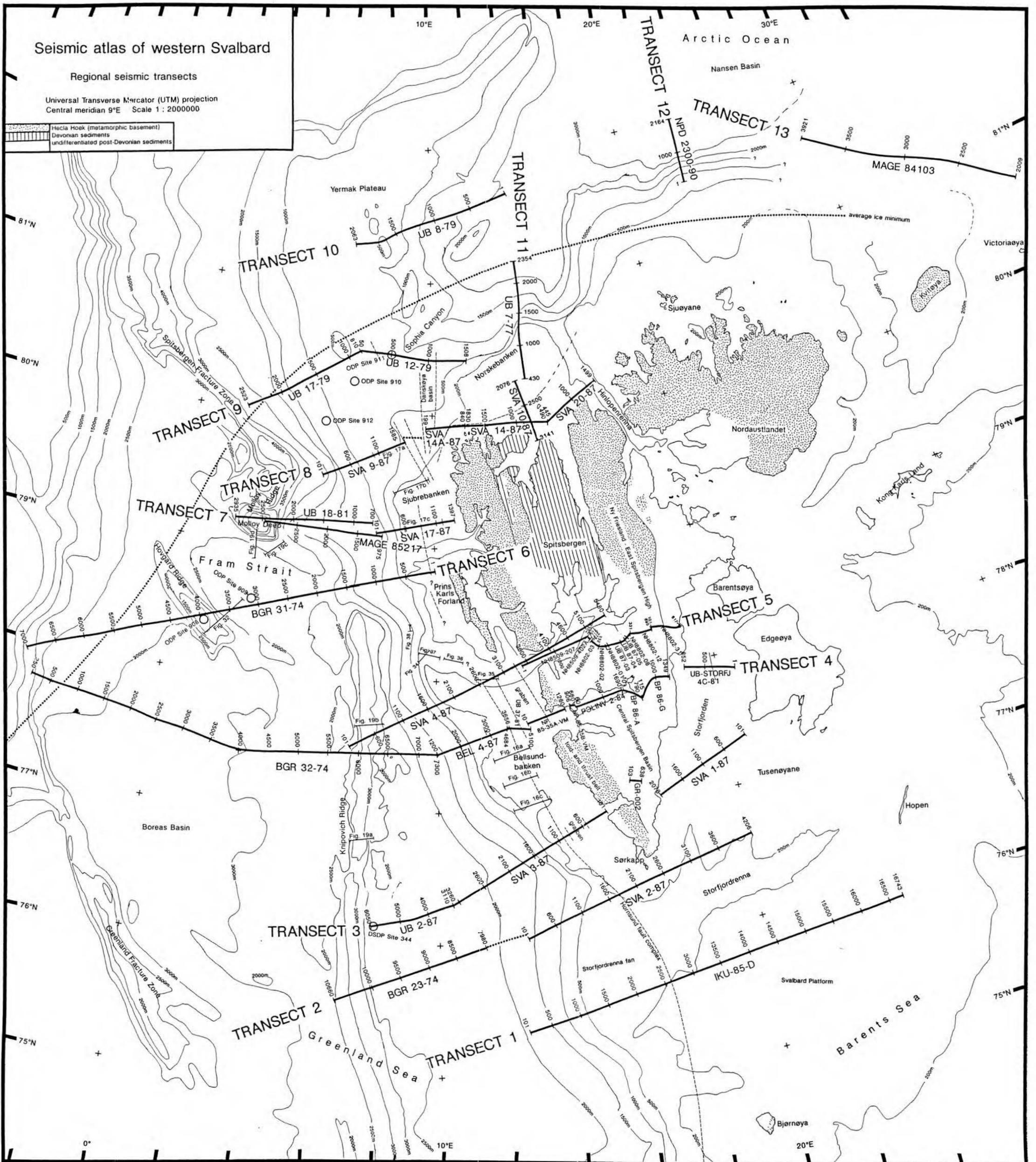
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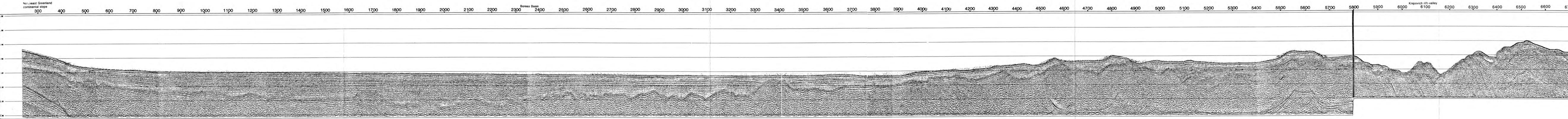
Seismic atlas of western Svalbard

Regional seismic transects

Universal Transverse Mercator (UTM) projection
Central meridian 9°E Scale 1 : 2000000

Hecla Hoek (metamorphic basement)
Devonian sediments
undifferentiated post-Devonian sediments





continue
below left.

Processing

SECT 101	SECT 102	SECT 103	SECT 104
Processing: 101	Processing: 102	Processing: 103	Processing: 104
Processing: 105	Processing: 106	Processing: 107	Processing: 108
Processing: 109	Processing: 110	Processing: 111	Processing: 112
Processing: 113	Processing: 114	Processing: 115	Processing: 116
Processing: 117	Processing: 118	Processing: 119	Processing: 120
Processing: 121	Processing: 122	Processing: 123	Processing: 124
Processing: 125	Processing: 126	Processing: 127	Processing: 128
Processing: 129	Processing: 130	Processing: 131	Processing: 132
Processing: 133	Processing: 134	Processing: 135	Processing: 136
Processing: 137	Processing: 138	Processing: 139	Processing: 140
Processing: 141	Processing: 142	Processing: 143	Processing: 144
Processing: 145	Processing: 146	Processing: 147	Processing: 148
Processing: 149	Processing: 150	Processing: 151	Processing: 152
Processing: 153	Processing: 154	Processing: 155	Processing: 156
Processing: 157	Processing: 158	Processing: 159	Processing: 160
Processing: 161	Processing: 162	Processing: 163	Processing: 164
Processing: 165	Processing: 166	Processing: 167	Processing: 168
Processing: 169	Processing: 170	Processing: 171	Processing: 172
Processing: 173	Processing: 174	Processing: 175	Processing: 176
Processing: 177	Processing: 178	Processing: 179	Processing: 180
Processing: 181	Processing: 182	Processing: 183	Processing: 184
Processing: 185	Processing: 186	Processing: 187	Processing: 188
Processing: 189	Processing: 190	Processing: 191	Processing: 192
Processing: 193	Processing: 194	Processing: 195	Processing: 196
Processing: 197	Processing: 198	Processing: 199	Processing: 200

Display

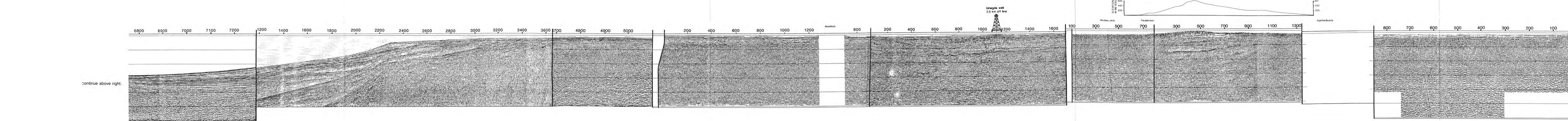
Display of traces with 50 m interval
Variable area shading, 40% bias
Horizontal scale 1: 20000
Vertical scale 1.5 cm, second

SEISMIC ATLAS OF
WESTERN SVALBARD
Transect 4: Boreas Basin -
Bellsund - Van Mijenfjorden
- Heer Land - Storfjorden

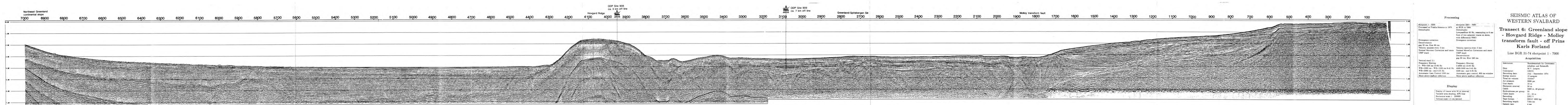
Lines BGR 32-74 shotpoint 250 - 7300,
BEL 4-87 shotpoint 1200 - 3656,
UB 37-81 shotpoint 4684 - 5100,
NP 85-35A-VM shotpoint 10 - 1280
NP 85-35B-VM shotpoint 680 - 906
POLINV 2-84 shotpoint 60 - 1691,
BP 86-A shotpoint 101 - 790,
BP 86-G shotpoint 115 - 1349
and UB-STORFJ-4C-81 shotpoint 1 - 852

Acquisition

Line	Shotpoint	Receiver	Group	Station	Depth
BGR 32-74	250	100	100	100	100
BGR 32-74	7300	100	100	100	100
BEL 4-87	1200	100	100	100	100
BEL 4-87	3656	100	100	100	100
UB 37-81	4684	100	100	100	100
UB 37-81	5100	100	100	100	100
NP 85-35A-VM	10	100	100	100	100
NP 85-35A-VM	1280	100	100	100	100
NP 85-35B-VM	680	100	100	100	100
NP 85-35B-VM	906	100	100	100	100
POLINV 2-84	60	100	100	100	100
POLINV 2-84	1691	100	100	100	100
BP 86-A	101	100	100	100	100
BP 86-A	790	100	100	100	100
BP 86-G	115	100	100	100	100
BP 86-G	1349	100	100	100	100
UB-STORFJ-4C-81	1	100	100	100	100
UB-STORFJ-4C-81	852	100	100	100	100



continue above right.



Processing

shotpoint 1 - 3200
Processed at Prins-Karls in 1978
Demultiplex

shotpoint 3201 - 7000
at BGR in 1989
Demultiplex
Lowpassfilter 60 Hz, resampling to 6 ms
Sum of two adjacent traces on shots, with differential NMO
Divergence correction

Divergence correction
Deconvolution:
gap 20 ms, filter 80 ms
Velocity analysis every 3 km
Normal MoveOut Correction and mute
CMP stack

shotpoint 3201 - 7000
at BGR in 1989
Demultiplex
Lowpassfilter 60 Hz, resampling to 6 ms
Sum of two adjacent traces on shots, with differential NMO
Divergence correction
Velocity spectra every 5 km
Normal MoveOut Correction and mute
CMP stack
Deconvolution:
gap 30 ms, filter 240 ms

Vertical stack 2:1
Frequency filtering:
0 - WB=500 ms 10-60 Hz
WB=1000 ms - WB=1500 ms 8-45 Hz
WB=2000 ms - end 6-35 Hz
Automatic Gain Control 1000 ms
Mute above seafloor reflection

Frequency filtering:
0-4000 ms 10-60 Hz
4000-5000 ms 8-45 Hz
5500 ms - end 6-35 Hz
Automatic gain control, 800 ms window
Mute above seafloor reflection

Display

Display of traces with 50 m interval
Variable area shading, 40% bias
Horizontal scale 1 : 200000
Vertical scale 1.5 cm/second

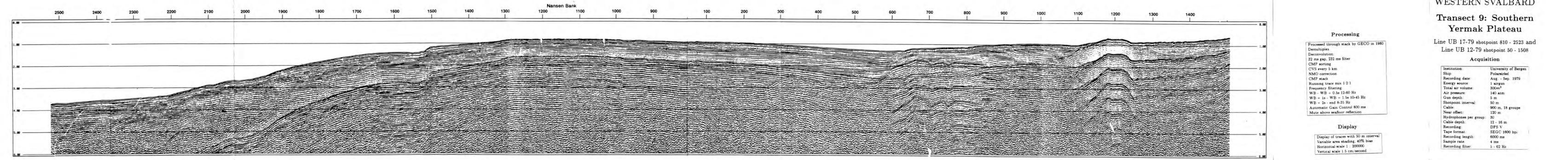
SEISMIC ATLAS OF WESTERN SVALBARD

Transect 6: Greenland slope - Hovgard Ridge - Molloy transform fault - off Prins Karls Forland

Line BGR 31-74 shotpoint 1 - 7000

Acquisition

Institution: Bundesanstalt für Geowissenschaften und Rohstoffe
Ship: M. V. Longva
Contractor: GECO
Recording date: July - September 1974
Energy source: 10 airguns
Total air volume: 1510m³
Air pressure: 2000 psi
Gun depth: 10 m
Shotpoint interval: 50 m
Cable: 2400 m, 48 groups
Hydrophones per group: 30
Cable depth: 12 - 16 m
Recording: DFS V
Tape format: SEGOC 1600 bps
Recording length: 7164 ms
Sample rate: 4 ms
Recording filter: 8 - 62 Hz



Processing

Processed through stack by GECO in 1980
Demultiplex
Deconvolution:
32 ms gap, 212 ms filter
CMP sorting
CIVS every 3 km
NMO correction
CMP stack
Running trace max 1:2:1
Frequency filtering:
WB = WB = 0.5s 12-60 Hz
WB = 1s - WB = 1.5s 10-45 Hz
WB = 2s - end 8-35 Hz
Automatic Gain Control 800 ms
Mute above seafloor reflection

Display

Display of traces with 50 m interval
Variable area shading, 40% bias
Horizontal scale 1 : 200000
Vertical scale 1.5 cm/second

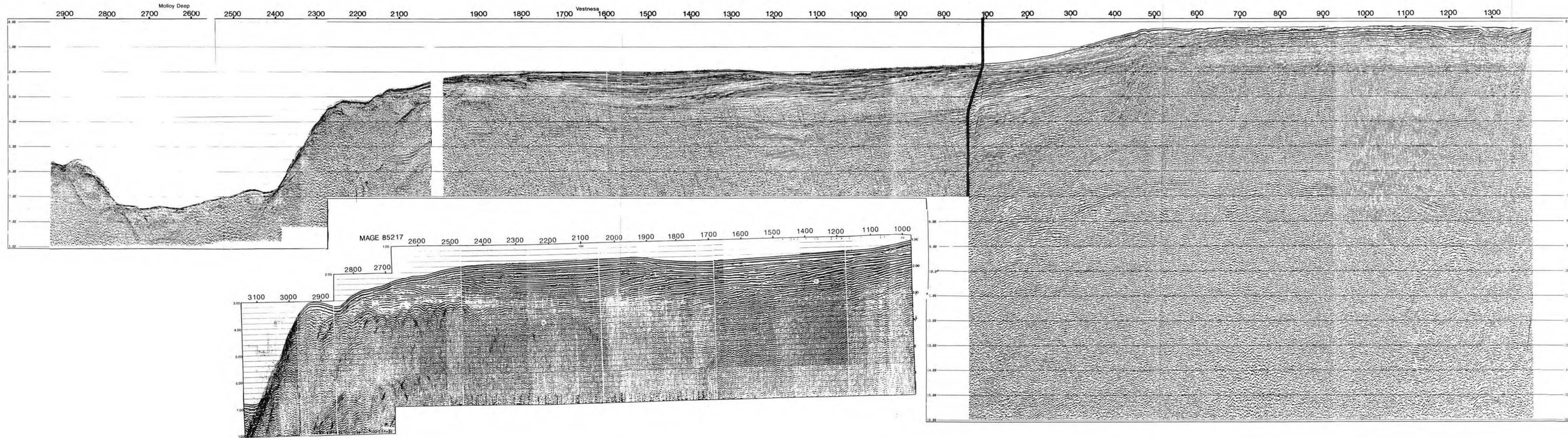
SEISMIC ATLAS OF WESTERN SVALBARD

Transect 9: Southern Yermak Plateau

Line UB 17-79 shotpoint 810 - 2523 and
Line UB 12-79 shotpoint 50 - 1508

Acquisition

Institution: University of Bergen
Ship: Polarstern
Recording date: Aug - Sep 1976
Energy source: 1 airgun
Total air volume: 300m³
Air pressure: 140 atm
Gun depth: 3 m
Shotpoint interval: 50 m
Cable: 900 m, 18 groups
Near offset: 120 m
Hydrophones per group: 30
Cable depth: 12 - 16 m
Recording: DFS V
Tape format: SEGOC 1600 bps
Recording length: 6000 ms
Sample rate: 4 ms
Recording filter: 5 - 62 Hz



SEISMIC ATLAS OF
WESTERN SVALBARD
Transect 7: Molloy Deep -
Vestnesa - Sjubrebanken

Lines UB 18-81 shotpoint 700 - 2935
and SVA 17-87 shotpoint 101 - 1397

Acquisition

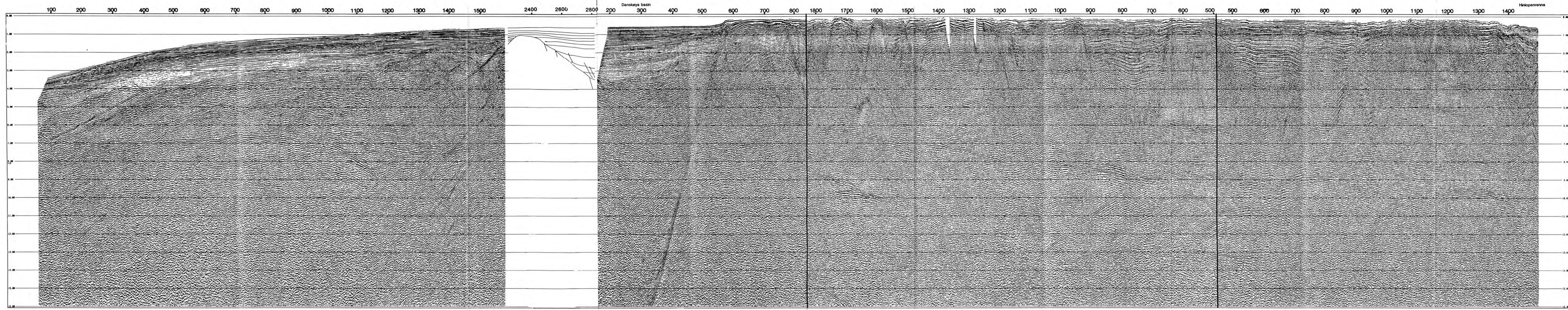
Line	UB 18-81	SVA 17-87
Institution:	University of Bergen	University of Bergen
Ship:	Håkon Mosby	Mobil Search
Contractor:		Mobil Norway
Recording date:	August - September 1981	August 1987
Energy source:	3 airguns	48 airguns
Total air volume:	1380m ³	7300m ³
Air pressure:	2000 psi	1850 psi
Gun depth:	7 m	12 m
Shotpoint interval:	50 m	50 m
Cable:	1200 m, 24 groups	4500 m, 90 groups
Near offset:	180 m	260 m
Hydrophones per group:		64
Cable depth:	10 m	15 m
Recording:	DFS V	DFS V
Tape format:	SEGC 1600 bpi	SEGY 6250 bpi
Recording length:	7000 - 9000 ms	16000 ms
Sample rate:	4 ms	8 ms (AQ1)
Recording filter:	out - 64 Hz	5-45 Hz

Processing

at Geo Harstad 1985 - 1986	Onboard Mobil Search in 1987
Demultiplex	18 ms static shift Instrument dephase filter Divergence correction
Divergence correction	Divergence correction
CMP sorting	CMP sorting Trace balancing in one window Mute
Velocity analysis every 3 km	Running trace mix 5:1 Deconvolution: 0.2-6 sec 32 ms gap, 248 ms filter 6-13 sec 64 ms gap, 352 ms filter Constant Velocity Stack every 3 km
Normal MoveOut Correction and Mute	Normal MoveOut Correction and Mute
CMP stack, offset weights: (near-mid-far) 0-1-1	CMP stack, offset weights: (near-mid-far) 1-3-1
Vertical stack 2:1	Vertical stack 2:1
Frequency filtering: 0 - WB+500 ms 12-60 Hz WB+500 - WB+1000 ms 10-50 Hz WB+2000 - end 8-40 Hz	Frequency filtering: 4000 - 6000 ms 8-25 Hz 8000 - 10000 ms 6-30 Hz 12000 - 16000 ms 5-25 Hz
Automatic Gain Control, 1000 ms	Automatic Gain Control 2000 ms
Mute above seafloor reflection	Mute above seafloor reflection

Display

Display of traces with 50 m interval
Variable area shading, 40% bias
Horizontal scale 1 : 200000
Vertical scale 1.5 cm/second



SEISMIC ATLAS OF
WESTERN SVALBARD
Transect 8: Southern
Yermak Plateau - Northern
coast of Spitsbergen

Lines SVA 9-87 shotpoint 101-1585,
SVA 14A-87 shotpoint 99-540,
SVA 14A-87 shotpoint 2340-2810 (line-drawing),
NPD 7950-80 shotpoint 490-1830 (line-drawing),
SVA 14-87 shotpoint 490-1830
and SVA 20-87 shotpoint 445-1499

Acquisition

Institution	University of Bergen
Ship	Mohit Search
Contractor	Mohit Search
Recording date	August 1987
Energy source	48 charges
Total air volume	7200m ³
Air pressure	1800 psi
Gun depth	15 m
Shotpoint interval	50 m
Cable	4000 m, 90 groups
Group offset	200 m
Hydrophones per group	64
Cable depth	15 m
Recording	5000 V
Trace format	SEG Y 8200 kps
Recording length	10000 ms
Sample rate	8 ms (nom)
Recording filter	5 - 45 Hz

Processing

Processed onboard Mohit Search in 1987

18 ms static shift

Deconvolution

Dispersive correction

CMP sorting

Trace balancing in one window

Gain

Residual time max 5:1

Decomposition

0.5-6 sec 22 ms gap, 245 ms filter

6-13 sec 44 ms gap, 352 ms filter

Constant Velocity Stack every 3 km

3000 correction

CMP stack, offset weights

0.5

Vertical stack 2:1

Frequency filtering

0.2 - 10 Hz filter

4.4 - 8.5 Hz filter

6-10 - 6-30 Hz

15-16 - 1-25 Hz

Automatic gain control 2000 ms

Mean above and below reflection

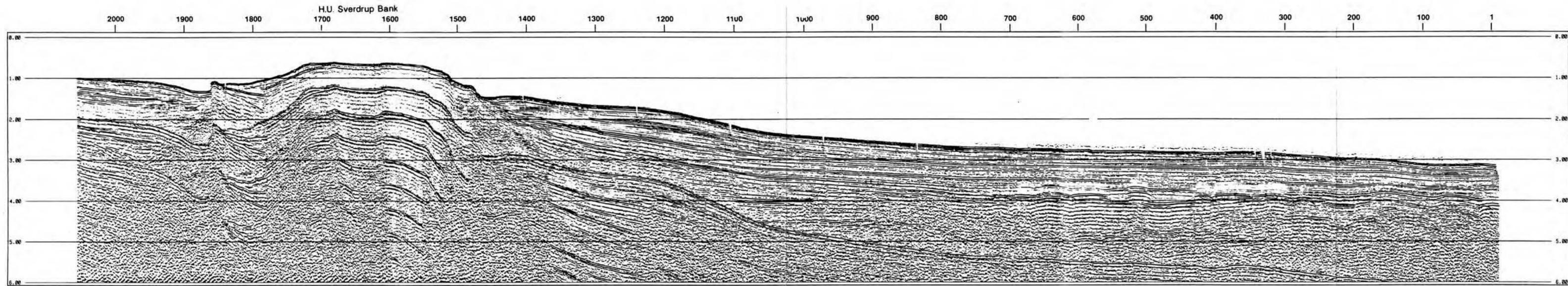
Display

Display of traces with 50 m interval

Variable area shading, 40% base

Horizontal scale 1: 200000

Vertical scale 1: 5 cm/second



SEISMIC ATLAS OF
WESTERN SVALBARD

**Transect 10: Yermak
Plateau - Nansen Basin**

Line UB 8-79 shotpoint 1-2063

Acquisition

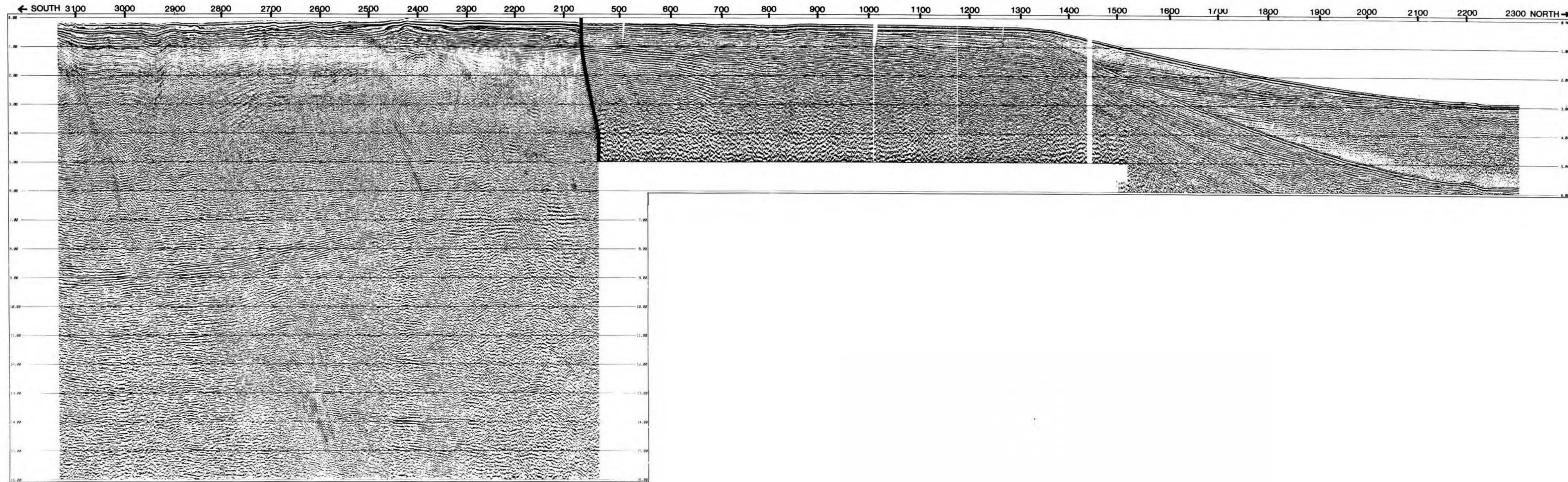
Institution:	University of Bergen
Ship:	Polarisraket
Recording date:	Aug - Sep. 1979
Energy source:	1 airgun
Total air volume:	300m ³
Air pressure:	140 atm
Gun depth:	5 m
Shotpoint interval:	50 m
Cable:	900 m, 18 groups
Near offset:	120 m
Hydrophones per group:	30
Cable depth:	12 - 16 m
Recording:	DPS V
Tape format:	SEGC 1600 bps
Recording length:	6000 ms
Sample rate:	4 ms
Recording filter:	5 - 62 Hz

Processing

Processed through stack by GECO in 1980
Demultiplex
Deconvolution
32 ms gap, 232 ms filter
CMP sorting
CWS every 5 km
NMO correction
CMP stack
Running trace mix 1:2:1
Frequency filtering:
WB - WB + 0.2s 12-40 Hz
WB + 1s - WB + 1.5s 10-45 Hz
WB + 2s - end 4-35 Hz
Automatic Gain Control 800 ms
Mute above seafloor reflection

Display

Display of traces with 50 m interval
Variable area shading, 40% bias
Horizontal scale 1 : 200000
Vertical scale 1.5 cm/second



SEISMIC ATLAS OF
WESTERN SVALBARD

**Transect 11: Gråhukken -
Norskebanken - northern
continental slope**

Lines SVA 10-87 shotpoint 2076-3141
and UB 7-77 shotpoint 430-2354

Acquisition		
Line	SVA 10-87	UB 7-77
Institution	University of Bergen	University of Bergen
Ship	Mobil Search	H. T. Sverdrup
Contractor	Mobil Search	H. T. Sverdrup
Recording date	August 1987	Aug. - Sep. 1977
Energy source	48 airguns	1 airgun
Total air volume	73000 m ³	2000 m ³
Air pressure	1850 psi	140 atm
Gun depth	12 m	6 m
Shotpoint interval	50 m	50 m
Cable	4500 m, 90 groups	900 m, 18 groups
Shot offset	280 m	180 m
Hydrophone per group	64	30
Cable depth	15 m	12-16 m
Recording	DSS V	DPS V
Tape format	SEG Y #250 lps	SEG Y 1600 lps
Recording length	18000 ms	5000 - 8000 ms
Sample rate	8 ms (AG1)	4 ms
Recording filter	5 - 45 Hz	5 - 62 Hz

Processing

SVA 10-87	UB 7-77
Processed on Mobil Search in 1987	On University of Bergen in 1987
18 ms static	Resampling to 8 ms
Instrument delay filter	Edit: scale of trace = 15
Divergence correction	Divergence correction
	On the shelf
	Fl. filter with passband above
	- 3000 m/s with preliminary NMO
CMP sorting	CMP sorting
Trace balancing in new window	
Mute	Mute on the shelf
	up to 100 ms at 230 m offset
	up to 800 ms at 430 m offset
	up to 800 ms at 730 m offset
	up to 1100 ms at 1030 m offset
Running trace max 5.1	
Demultiplexion	Demultiplexion
0.24 sec 32 ms gap, 248 ms filter	32 ms gap, 240 ms operator
6.13 sec 64 ms gap, 352 ms filter	
Constant Velocity Stack every 3 km	Velocity spectra and CVS
NMO correction	NMO correction
CMP stack, offset weights	CMP stack, offset weights
1:3:1	25: 5: 75: 1: 1: 1: 1: 5
Vertical stack 2:1	Vertical stack 2:1
Frequency filtering	Frequency filtering
0.2 x 10-40 Hz	WB - WB - 1a 10-40 Hz
4.0 x 6-35 Hz	WB - 1 b - WB - 2 1a 6-45 Hz
6.0 x 6-30 Hz	WB - 2a - and 6-35 Hz
12.18 x 6-25 Hz	
Automatic gain control 2000 ms	Automatic gain control 1000 ms

Display

Display of trace with 50 m interval
Variable area shading, 40% time
Horizontal scale 1: 200000
Vertical scale 1: 5 m second

SEISMIC ATLAS OF WESTERN SVALBARD

Transect 12: North of Nordaustlandet

Line NPD 2300-90 shotpoint 1 - 2164

Acquisition

Ship	Norwegian Petroleum Directorate
Recording date	October 1990
Energy source	airgun
Trawl air volume	8220 m ³
Air pressure	2000 psi
Cable depth	6 m
Shotpoint interval	25 m
Cable	3000, 120 groups
Hydrophones per group	40
Head offset	40 m
Cable depth	10 m
Recording	DPS V, GDR 1000
Tape format	SEGG
Recording length	7000 ms
Sample rate	2 ms
Recording filter	2.5 - 180 Hz

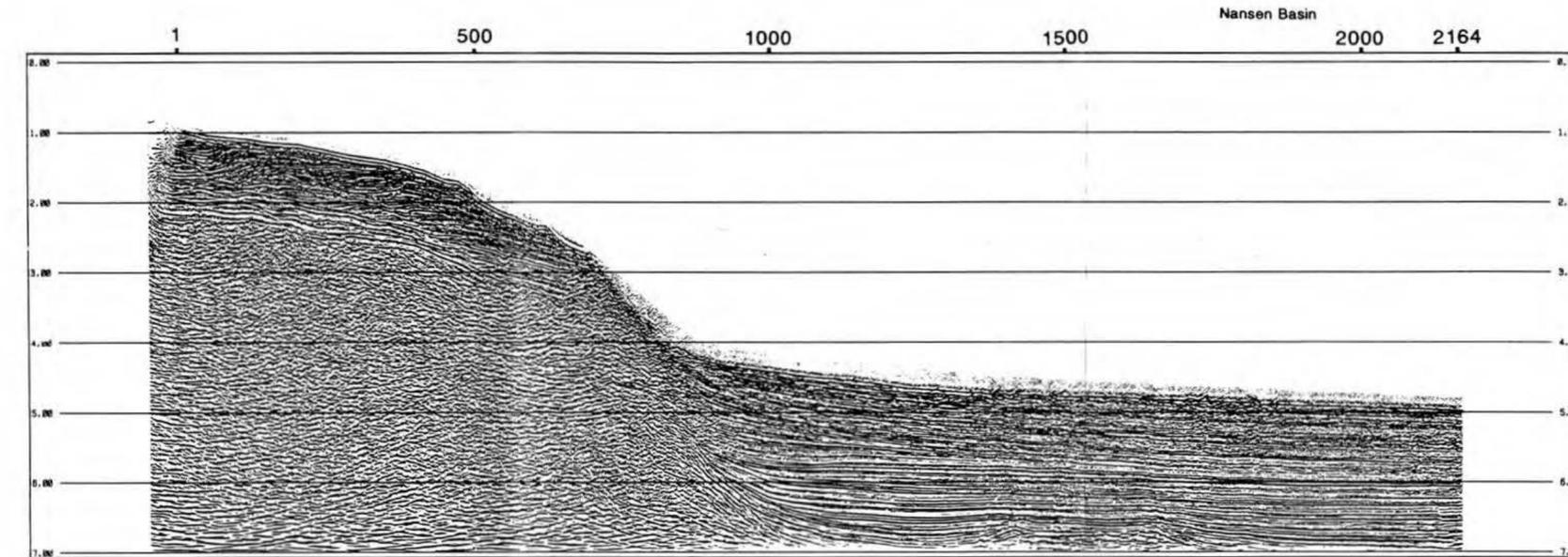
Processing

Processed by Compagnie Generale de Geophysique in 1992

- Anti-alias filter, resample to 4 ms
- Amplitude recovery
- CMP sorting
- 1st velocity analysis, every 4 km
- Wave equation multiple attenuation (eq. 1-16)
- Partial stack, reduction to 30 fold
- Signature denoising, conversion to maximum phase
- Dip correction
- Dispersion
- 2d velocity analysis, every 2 km
- NMO correction and stack
- Trace equalization
- CMP stack
- Basic correction 11 ms
- Phase-Response migration
- Vertical stack 4.1
- Frequency Shaping 12-60 Hz
- Automatic Gain Control 3000 ms
- Mean above seabed reflection

Display

Display of trace with 50 m interval
Variable area shading, 40% level
Horizontal scale 1: 20000
Vertical scale 1: 5 m/sec



SEISMIC ATLAS OF WESTERN SVALBARD

Transect 13: Between Nordaustlandet and Franz Josef Land

Line MAGE 84103 shotpoint 2009 - 3921

Acquisition

Ship	Marine Arctic Prospecting
Recording date	Geophysical Expedition, March 1988
Energy source	airgun
Trawl air volume	10 litres
Air pressure	120 atm
Cable depth	15 m
Shotpoint interval	50 m
Cable	3000, 48 groups
Head offset	45 m
Hydrophones per group	41
Cable depth	17 - 19 m
Recording	"Program" recording system
Tape format	SEGB
Recording length	8000 ms
Sample rate	4 ms
Recording filter	10 - 83 Hz

Processing

Processed by Marine Arctic Prospecting Geophysical Expedition in 1988

- Demultiplex
- Stack
- CMP sorting
- Frequency Shaping 10 - 80 Hz
- CYS every 10 km
- NMO correction
- CMP stack
- Frequency Shaping 10 - 80 Hz
- Scaling

Display

Display of trace with 50 m interval
Variable area shading
Horizontal scale 1: 20000
Vertical scale 1: 45 m/sec

