Dynamics of the Late Weichselian Svalbard-Barents Sea Ice Sheet and its deglaciation based on high-resolution bathymetric mapping and raised beach records

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Jarðvísindadeild
Háskóli Íslands
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10 eininga ritgerð sem er hluti af
Baccalaureus Scientiarum gráðu í jarðfræði

Leiðbeinandi
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Jarðvísindadeild
Verkfræði- og náttúruvísindasvið
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ABSTRACT

This thesis aims at giving a literature-based overview of the dynamics of the Late Weichselian Svalbard-Barents Sea Ice Sheet and its deglaciation, based on high-resolution bathymetric mapping and raised beach records. Extensive patterns of large- and medium scale submarine landforms formed by differences in ice-flow regimes are shown by high-resolution bathymetric mapping from the fjords and their adjacent cross-shelf troughs around Svalbard. Mega-scale glacial lineations, lateral moraines and transverse ridges occur superimposed on the fjord, shelf and cross-shelf trough morphology of the margin. From these landforms the ice-flow and dynamics of the last ice sheet over Svalbard can be inferred. The major fjords and their adjacent cross-shelf troughs are interpreted to be the main routes for fast-flowing ice streams draining the ice sheet. Along the west coast of Svalbard, the Bellsund, Isfjorden and Kongsfjorden troughs were major drainage pathways of the ice sheet. On northern Svalbard the main drainage routes were through Woodfjorden and Wijdefjorden-Hinlopen Strait. The pattern of postglacial emergence on Svalbard is important in correlating the timing of the last deglaciation and ice extent and thickness in time and space, along with the submarine landforms. The maximum area of glacier loading is observed on Kongsøya where postglacial raised beaches occur >100 m above sea level. Deglacial unloading began on western and northern Spitsbergen c. 13–12 ^{14}C ka ago, and by c. 10.5 ^{14}C ka on eastern Svalbard.

ÚTDRÁTTUR

Ritgerðin dregur saman gögn sem varpa ljósi á virkni jökulhvelsins er huldi Svalbarða og Barentshaf á hámarki síðasta jökulskeiðs, einkum gögn frá hafsbotni umhverfis Svalbarða og strandlínum frá tímanum eftir jöklaleysinguna. Mynstur stórra og meðalstórra neðansjávar landforma sem myndast hafa við mismunandi aðstæður við flæði skriðjökla og ísstrauma frá jökulhvelinu koma fram við neðansjávarkortlagningu fjarða og landgrunns Svalbarða. Stór línuleg landform, jaðargarðar og þverhryggir eru algeng set- og landform fjarðarbotna og landgrunns Svalbarða. Hægt er með hliðsjón af þessum landformum að draga ályktanir um ísflæði og virkni skriðjökla síðasta jökulhvels yfir Svalbarða. Stærstu firðir og fjarðartrog eru talin hafa verið farvegir skriðjökla og ísstrauma sem leystu jökulhvelið. Meðfram vesturströnd Svalbarða voru það fjarðarkerfi Bellsund, Isfjorden og Kongsfjorden sem veittu stærstum hluta ísstreymisins. Á norður Svalbarða voru aðal rennslisleiðir íss um Woodfjorden og Wijdefjorden-Hinlopen sundið. Mynstur landlyftingar og aldur strandlina á Svalbarða eru, ásamt mynstri og gerð neðansjávar landforma, mjög mikilvæg til þess að álykta um tímatal síðustu jökulhörfunar og breytingar í útbreiðslu og þykkt síðasta jökulhvels. Hámarksferging af völdum jökulhvelsins var á svæðinu við Kongsøya, í austurhluta eyjaklasans, þar sem hæstu strandlínur eftir jöklaleysingu eru >100 m yfir núverandi sjávarmáli. Jökulhörfum hófst á vestur og norðurhluta Svalbarða fyrir um 13–12 þúsund ^{14}C-árum síðan og fyrir um 10.5 þúsund ^{14}C árum síðan á austur Svalbarða.

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1 INTRODUCTION

For several decades the extent of the Late Weichselian Ice Sheet over Svalbard and the Barents Sea has been debated on. Scientist have suggested that the ice sheet reached the continental slope covering Svalbard and the Barents Sea with at least 3000 m thick ice in the central part of the Barents Sea Shelf and > 800 m thick ice along the western coast of Svalbard (Lambeck, 1995, 1996) while other scientist have suggested that the Late Weichselian Ice sheet did not reach Svalbard and that the ice covering Svalbard was characterized by local ice caps over central Spitsbergen with large outlet glaciers and fast flowing ice streams localized in major valleys, fjords and troughs (Lambeck, 1995; Lambeck 1996; Mangerud *et al*., 1998; Landvik *et al.*, 1998; Andersson *et al.*, 1999)

In fjords and on the continental shelf around Svalbard extensive data of high-resolution bathymetric mapping have been collected (e.g. Ottesen *et al*., 2005; Ottesen and Dowdeswell, 2007; Ottesen *et al*., 2007; Dowdeswell *et al*., 2007; Ottesen *et al*., 2008; Ottesen and Dowdeswell, 2009) showing extensive patterns of submarine landforms such as mega-scale glacial lineations (MSGL), lateral moraines and transverse ridges that were formed by differences in ice-flow velocities and dynamics. By looking at these landforms the flow and dynamics of the last ice sheet covering Svalbard, the Late Weichselian Ice Sheet, can be interpreted (Ottesen *et al*., 2007). Landvik *et al*. (2005) combined new marine evidence and land records from north-western Svalbard and suggested that ice streams draining the Late Weichselian Ice Sheet were drained through the major fjords and cross-shelf troughs of north-western Svalbard, Woodfjorden and Wijdefjorden-Hinlopen fjord system. On the west coast of Svalbard Bellsund, Isfjorden and Kongsfjorden were major drainages pathways as well for the Late Weichselian Ice Sheet (Landvik *et al.*, 2005; Ottesen *et al*., 2007). Their results indicated that the Kongsfjorden cross shelf trough was filled by a fast flowing ice stream constrained by a slower moving ice on the adjacent shelves and strand flats favoring the preservation of older geological records adjacent to the main pathway of the Kongsfjorden glacial system. They suggested that the same model might apply to the other major drainage pathways for the Late Weichselian Ice Sheet (Landvik *et al*. 2005).

Ice streams are fast flowing curvilinear features that are constrained within slower moving ice. They drain very large interior basins of our modern ice sheets, Antarctica and Greenland (10^6 to 10^7 km^2) and are one of the main mechanisms of mass loss to the oceans. They are typical features for modern ice sheet dynamics and are commonly tens of kilometers in width, up to 2 km in thickness and up to hundreds of kilometers in length (e.g. Bentley, 1987; Bamber *et al*., 2000; Fahnestock and Bamber, 2001; Whillans *et al*., 2001).

The altitude and age of raised-beach deposits are a critical field observation in determining the magnitude and distribution of past-glacier loads and the timing of the last deglaciation (Forman *et al.* 2004). On eastern Svalbard and islands (Storøya, Kongsøya, Barentsøya, Edgeøya and Hopen) well preserved and extensive Late Weichselian and Holocene raised beach deposits occur. They are from 60 to 130 m above present mean high tide mark (m aht) and reflect the area of maximum ice sheet loading (Salvigsen, 1981; Forman, 1990; Landvik *et al.*, 1998; Forman *et al.*, 2004). On the other hand postglacial emergence on northern and western Svalbard is at 65 m aht or lower (Forman, 1990), indicating relatively less loading at the margin of the ice sheet. Kurt Lambeck proposed a model of post-glacial crustal rebound and estimated that the ice sheet had to be at least 3000 m thick in the central part of the Barents Sea Shelf (Lambeck, 1996) and > 800 m thick along the western coast of Svalbard (Lambeck, 1995). When the ice front started to retreat it retreated rapidly and the shelf west of Svalbard was deglaciated by ca 15 ka BP (Landvik *et al.*, 1998) and middle of Isfjorden was ice-free ca 10,5 ka BP (Elverhøi *et al.*, 1995). The raised beach record in Svalbard reflects principally two competing processes; the postglacial rise in sea level and isostatic uplift of the lithosphere with disintegration of the Late Weichselian Ice Sheet (Forman *et al.* 2004). Holocene patterns of uplift (Landvik *et al.*, 1998) and marine sediments of widely different ages, that are found at approximately same altitudes, indicate that the uplift is due to repeated vertical glacio-isostatic movements, that are caused by the loading and unloading of major ice sheets (Mangerud *et al.*, 1998).

In this paper glacial landforms characteristic for ice sheet activity will be described with examples from major fjords and from the continental shelf around Svalbard. These landforms will then be used to reconstruct the dynamics of the former ice sheet covering Svalbard. By using the raised beach deposits we may confirm the dynamics of the last ice sheet and reconstruct the deglaciation of it.

1.1 Physiographic setting

Svalbard is a high-arctic archipelago located near the western edge of the Barents Sea Shelf between 74 and 81 degrees north and 10 and 35 degrees east. Svalbard consist of many islands but the main islands are Spitsbergen, Nordaustlandet, Barentsøya, Edgeøya, Kong Karls Land, Prins Karls Forland, and Bjørnøya. The archipelago covers ca. 62.000 km^2 and about 60% of the area is presently covered with glaciers, both icecaps, valley, tidewater and cirque glacier. Svalbard is characterized by an Alpine mountain landscape, glacially eroded fjord systems and coastal strandflats. Raised beaches and post glacial marine terraces commonly occur along the coast (Harland, 1997; Hjelle, 1993).

2 METHODS

2.1 Multibeam echo sounder bathymetry

The Norwegian Hydrographic Service (NHS) collected multibeam bathymetry data (EM1002, depth range 20-1000 m) during the years 1999-2004 in the fjords and on the adjacent shelf areas around Svalbard. These data were gridded with a cell size of 10 or 50 m, depending on the sampling density. The University of Tromsø (EM300) and the RV Jan Mayen (1500 km^2) have collected multibeam bathymetry data in several areas. In October 2004 one line of bathymetric data gridded with a cell size of 10 m was collected north of Svalbard (Ottesen *et al*. 2005; Ottesen *et al.* 2007).

All multibeam bathymetry data presented in this paper have been published before (e.g. Ottesen *et al*., 2005; Ottesen and Dowdeswell, 2007; Ottesen *et al*., 2007; Dowdeswell *et al*., 2007; Ottesen *et al*., 2008; Ottesen and Dowdeswell, 2009; Hogan *et al.*, 2010).

2.2 Raised beach record

Nordenskiøld (1866), a pioneering Scandinavian geologist and De Geer (1919) were first to scientifically recognize raised marine landforms on Svalbard, and in the 1950s the first strandlines were dated by radiocarbon in Billefjorden, central Spitsbergen, and in Lady Franklin Fjord, Nordaustlandet, by Freyling-Hanssen and Olsson (1959) and by Blake (1961) (Forman *et al.* 2004). There are 28 separate assessments of post-glacial emergence for Svalbard (Fig. 2.1) (Forman *et al*., 2004).

The raised beach data record presented in this paper have been published before (e.g. Salvigsen, 1981; Forman *et al.*, 1990; Landvik *et al.*, 1998; Forman *et al*., 2004).

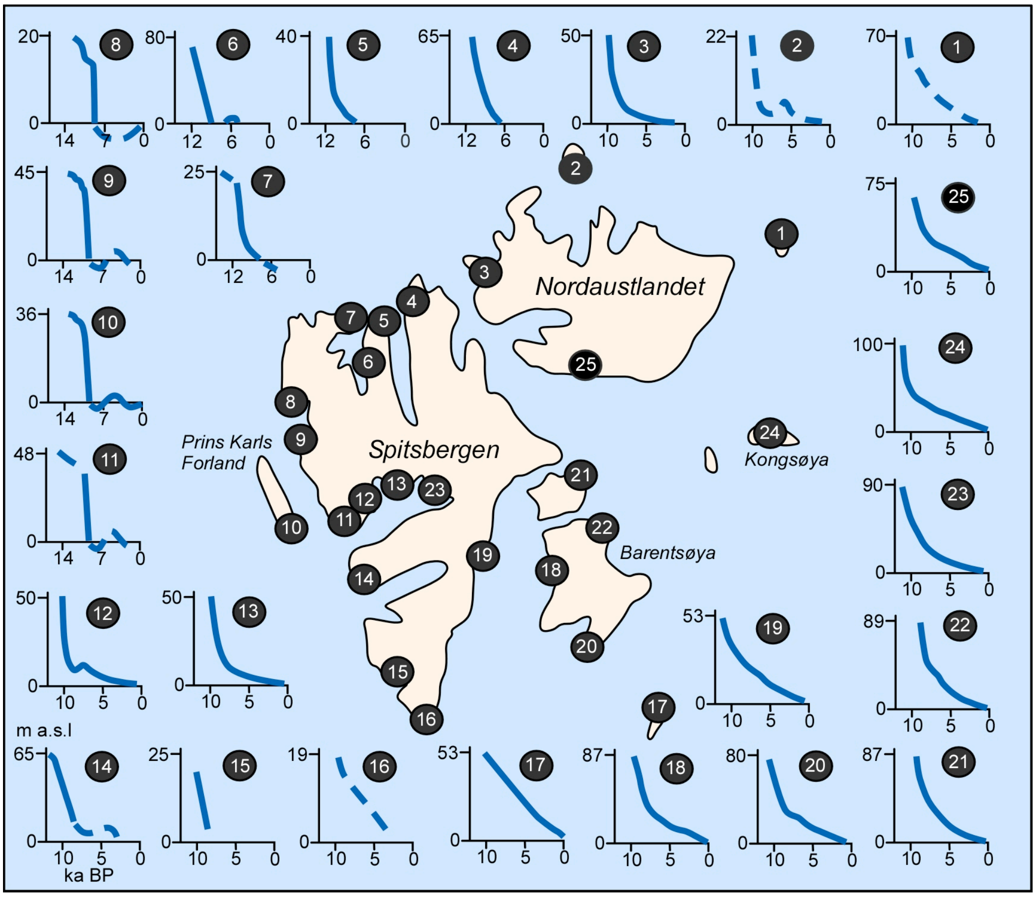

Fig. 2.1 Relative sea level curves from Svalbard. (Modified by Ólafur Ingólfsson after Forman et al. 2004).

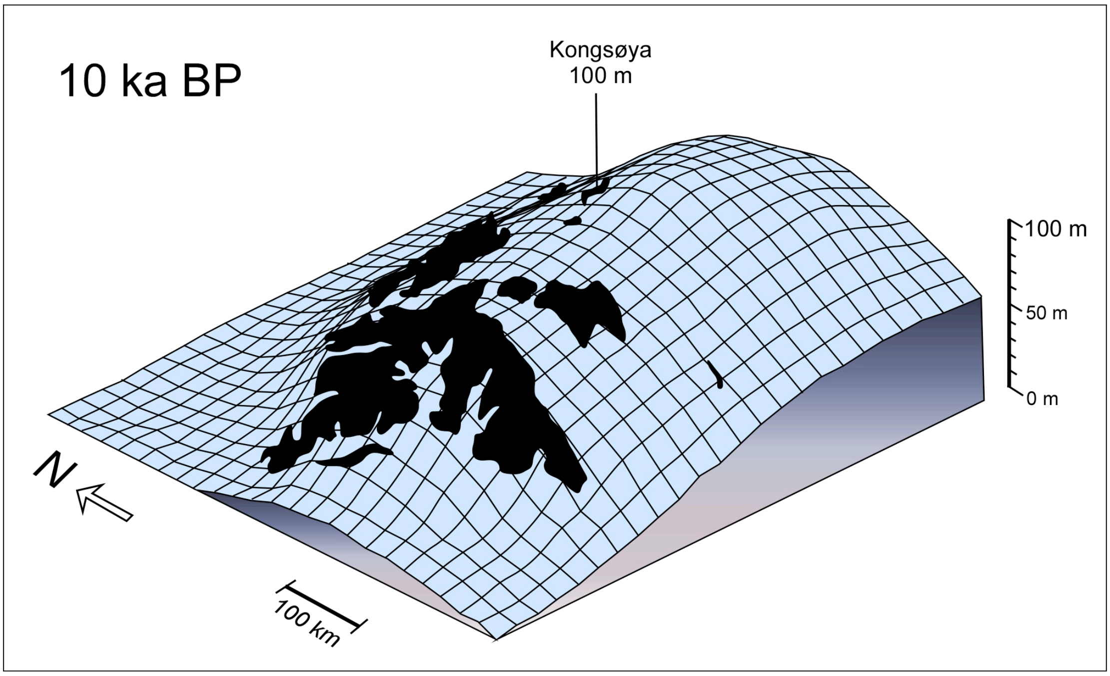

Fig. 2.2 The pattern of postglacial raised beaches, combined with well-dated relative sea-level curves, fingerprints the isostatic depression caused by the Svalbard-Barents Sea ice sheet. (Modified by Ólafur Ingólfsson from Bondevik 1996).

3 SUBMARINE LANDFORMS

In major fjords and on the continental shelf around Svalbard detailed marine geophysical observations of submarine morphology have been systematically examined for the presence of submarine landforms linked to the presence of the Late Weichselian Ice Sheet (Fig. 3.1). The postglacial sediments on the shelf and in most outer fjords are limited to only a few tens of cm (Elverhøi and Solheim, 1983; Elverhøi, 1984) and therefore the glacial morphology is well shown and can easily be mapped (Ottesen *et al*. 2007). In some of the inner fjords the postglacial sediments can be much thicker ranging from few meters up to more than 10 m , making it harder to recognize the glacial landforms (Ottesen *et al. 2007)*. The morphology and the distribution of a few submarine landforms characteristic for the presence of grounded glacier ice and, in some cases, fast flowing ice will now be described.

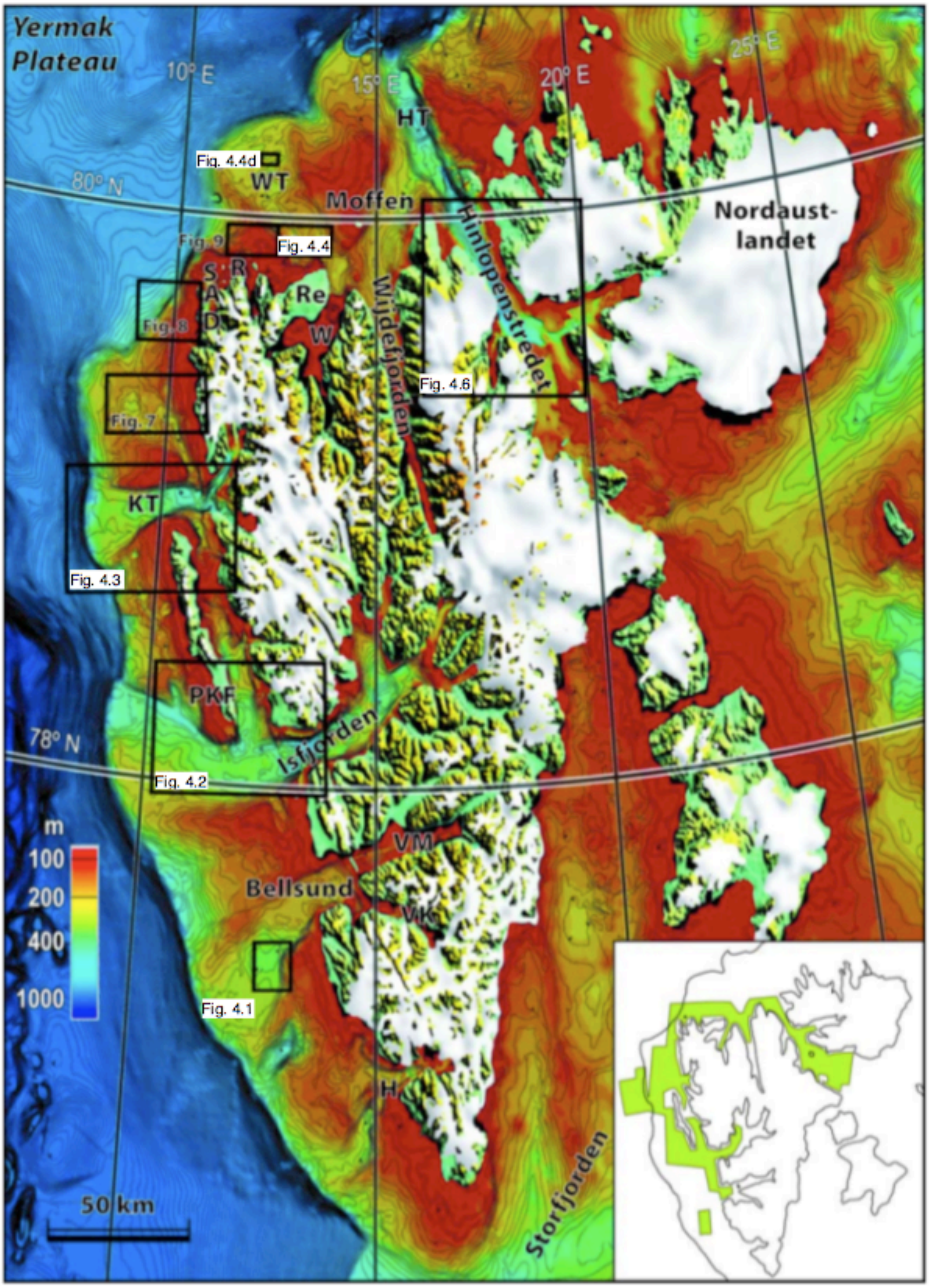

Fig. 3.1 Overview map with regional bathymetry of the fjord and shelf areas around Svalbard. 20 m depth contours. Data from single beam echo-sounder from the Norwegian Hydrographic Service. All data gridded with a cell size of 1 km. A - Amsterdamøya, D -Danskøya, H - Hornsund, HT - Hinlopen Trough, KT - Kongsfjorden Trough, PKF - Prins Karls Forland, R - Raudfjorden, Re - Reinsdyrflya, S - Smeerenburgfjorden, vK- van Keulen fjorden, vM - van Mijenfjorden, W - Woodfjorden, WT - Woodfjorden Trough. Green colour on inset map hows extent of swath bathymetric data coverage and the solid line west of Svalbard is the shelf edge. (Ottesen et al. 2007).

3.1 Megascale lineations

3.1.1 Description

Streamlined submarine features, with individual lineations ranging from a few hundreds of meters up to more than ten kilometers in length, are observed in several major fjords and cross-shelf troughs on the Svalbard margin (Figs. 4.1, 4.2, 4.3, 4.4, 6.2). The bedforms are aligned parallel to one another and are oriented along the main axes of the depressions. The wavelength of the lineations range from 0.1-2 km and amplitudes are up to 15 m. Where fjords and troughs join, some sets of lineations show signs of convergence while in one case the divergence of lineations have been observed. On the shelf beyond the mouth of Woodfjorden in northern Spitsbergen a single set of lineations diverges into two sets of lineations around the island Moffen, the eastern set converges with the Wijdefjorden set while the western set continuous to the shelf edge (Fig 6.2, 3.2) (Ottesen *et al.*, 2005, 2007).

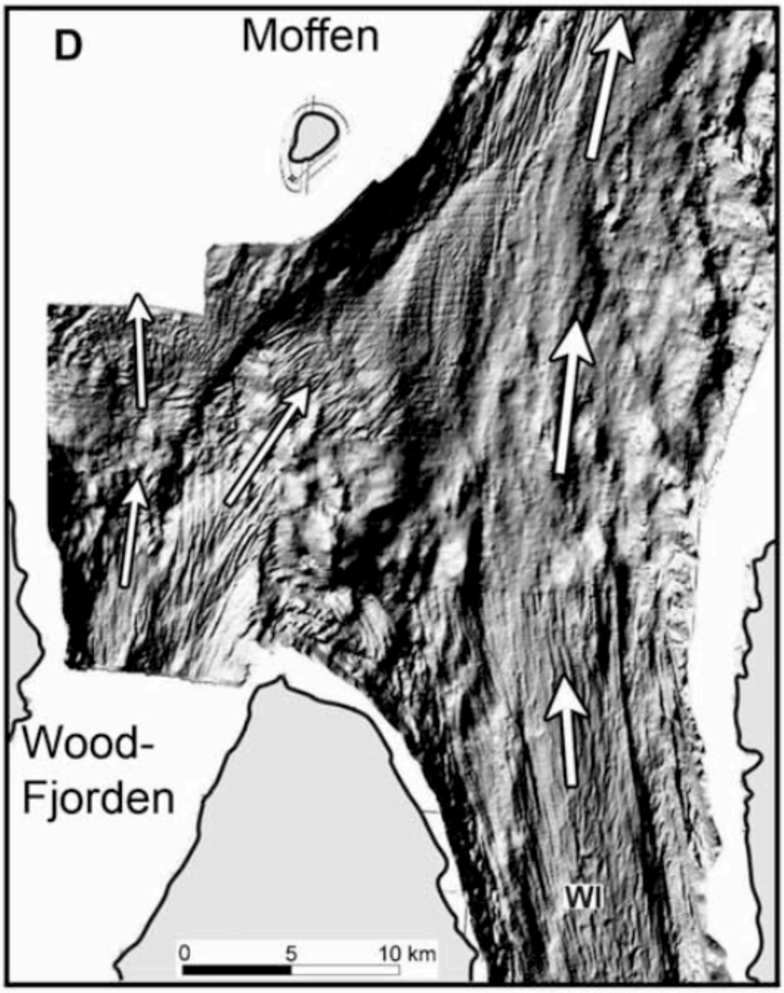

Fig. 3.2 EM1002 shaded-relief image of the seafloor of outer Wijdefjorden (Wi) and Woodfjorden on the northern side of Svalbard. MSGL (arrows) follow the long axis of Wijdefjorden, whereas the lineations in Woodfjorden diverge around the island of Moffen. Maximum water depth is 200 m. (Ottesen et al. 2005).

3.1.2 Interpretation

The streamlined lineations are similar in morphology to features describes as megascale glacial lineations (MSGL) (Clark, 1993; Stokes and Clark 1999, 2001; Ottesen *et al.* 2005). Similar features have been observed on the continental shelves of, for example, Norway and Antarctica (e.g. Canals *et al.*, 2000, 2002; Wellner *et al.,* 2001; Ó Cofaigh *et al.,* 2002; Dowdeswell *et al.*, 2004; Ottesen *et al.,* 2005) as well as on satellite imagery of parts of northern Canada (Clark, 1993). These landforms are thought to be formed due to soft sediment deformation at the base of a fast-flowing ice stream that is draining a large ice sheet (Tulaczyk *et al.,* 2001; Dowdeswell *et al.*, 2004; Ó Cofaigh *et al.,* 2005; Ottesen *et al.*, 2005).

The distribution of the lineation sets is compatible with formation beneath grounded glacier ice (Ottesen *et al.,* 2005, 2007). The fact that extensive lineation sets are found in the fjords and the cross-shelf troughs, links them to the former presence of relatively thick ice, presumably at the pressure melting point (Dowdeswell and Siegert, 1999; Ottesen *et al.*, 2005, 2007). Therefore basal melting and, later on, sediment deformation would be expected (Ottesen *et al.,* 2005, 2007). This interpretation is consistent with the calculations of ice-flow, thermal structure and sediment delivery from Svalbard fjords and troughs, e.g Isfjorden, where observations and models suggest fast ice-flow and high sediment delivery rates from ice-sheet drainage basins to the continental margin (Hooke and Elverhøi, 1996; Elverhøi *et al.*, 1998).

3.2 Ridges parallel to the former ice-flow direction

3.2.1 Description

Along the lateral margins of cross-shelf troughs in Svalbard individual ridges have been observed. The ridges are of tens of kilometers in length and between 40-60 m in hight. The ridges occur as single ridges on either side of a trough or as pairs on both sides. These ridges are suggested to be of course diamictic sediments due to little acoustic penetration by sub-bottom profilers. Examples are observed at the mouth of Isfjorden, Kongsfjorden troughs and Wijdefjorden, where they approach the shelf edge west and northwest of Svalbard (Fig. 4.3, 6.1) (Ottesen *et al.*, 2005, 2007).

3.2.2 Interpretation

The extensive ridges parallel to fjords and troughs around Svalbard are interpreted as glacier-derived moraine systems that determine the lateral limits of fast flowing former ice streams. They occur with MSGL as described above (Ottesen *et al.*, 2005, 2007). MSGL were not found beyond the moraine ridges and are therefore thought to mark the boundaries of fast flowing ice streams (Ottesen *et al.*, 2005, 2007). From terrestrial settings in the Canadian Arctic similar features have been observed and described. The association between the ridges and MSGL are thought to be former ice streams draining parts of the Late Wisconsin North American Ice Sheet

(Boulton and Clark, 1990; Stokes and Clark, 2001). Similar features have been described as well from Trænedjupet, a trough on the Norwegian margin (Ottesen *et al.*, 2005, 2007).

The formation of the extensive lateral ridges are thought to be the result of the shear zone and high stress gradient between fast and slow moving ice at ice stream lateral margins (e.g. Bentley 1987).

3.3 Transverse ridges and ice marginal grounding-zone wedges

3.3.1 Description

Large ridges transverse to the former ice-flow direction are found both at the shelf edge and in troughs and fjords of Svalbard. Theses ridges are several kilometers in width, up to tens of meters in hight and tens of kilometers in length (Fig. 4.2b, 4.3). With acoustic stratigraphic records, where penetration was achieved, it is possible to see that the ridges form sedimentary wedges lying above strong basal reflectors (Fig. 4.3c). Sometimes it may be quite difficult to detect the sedimentary wedges on the sea floor, but they are clearly visible on acoustic records. The Ridges may consist of a few cubic kilometers of sediments (Ottesen *et al.*, 2005, 2007).

3.3.2 Interpretation

The transverse ridges located at or near the shelf edge between large submarine fans are interpreted as terminal moraines recording the farthest advance of slow moving grounded glacier ice across the continental shelf (Elverhøi *et al.*, 1998; Dowdeswell and Elverhøi, 2002; Ottesen *et al.*, 2005).

The transverse ridges and underlying sedimentary wedges found in the fjords and troughs are interpreted to mark major still stands of the ice margin during the last deglaciation (e.g. Landvik *et al.*, 2005; McMullen *et al.*, 2006; Ottesen *et al. 2005)*. During a major still stand it allows the sediments to build up and form grounding zone wedges often tens of meters in thickness. These wedges are found either as single features or as sets of large ridges. They are sometimes associated with constricted and/or shallower areas of the fjords and troughs, which would have formed natural pinning points during retreat (Ottesen *et al.*, 2005, 2007).

3.4 Small transverse ridges

3.4.1 Description

Clusters of small transverse ridges, each ridge spaced a few hundred meters apart, are observed on the shelfs and in troughs and fjords around Svalbard. A single ridge can reach 15 m in height and have an average width of about 50-150 m. Outside the mouth of Woodfjorden there is an example of a cluster of small transverse ridges located in water depths of up to 150 m (Fig. 4.4) (Ottesen *et al.*, 2007).

3.4.2 Interpretation

The ridges described above are interpreted to be moraines that record short still stands or winter-summer ice front oscillations during general retreat of the ice margin during deglaciation (Ottesen and Dowdeswell 2006; Ottesen *et al.*, 2007). These ridges are generally recorded short-lived events during deglaciation due to the significant smaller sediment volumes in them compared to the large terminal moraines or grounding zone wedges. These ridges are thought to be formed during minor winter readvances of a tidewater glacier, as push moraines, when the calving is largely suppressed by the presence of sea ice (Boulton, 1986; Ottesen and Dowdeswell 2006; Ottesen *et al.*, 2007).

4 FJORDS AND TROUGHS ON THE SVALBARD MARGIN

Series of deep fjord and trough systems separated by intervening shallow banks between one and other are characteristic for the sea floor morphology of the Svalbard margin west and north of the archipelago. These systems are a result, mainly, of the advance of ice sheets, intercalated ice streams to the shelf edge on a number of occasions and other glacial activity during the Pleistocene (e.g. Solheim *et al.*, 1996; Mangerud *et al.*, 1998). The fjords and adjacent cross shelf troughs system and their characteristic submarine landforms will now be described and interpreted.

4.1 Western Spitsbergen

4.1.1 Bellsund trough and fjord system

The Bellsund system consists of van Mijenfjorden and van Keulenfjorden, ~60 km long and 10 km wide, and a cross-shelf trough that ends west of the outer coast at the shelf edge and is up to 260 m deep and 80 km long (Fig. 3.1). At the fjord mouth the trough is 15 km wide, but widens to the shelf edge up to about 40 km. The trough has steep walls on both sides in the inner part of it, for the trough is cut into sedimentary rock, but in the outer part the side-walls are more ridge like sedimentary features. The bathymetry data for the Bellsund area only covers a part of the Bellsund Trough and not the inner fjords (Ottesen *et al.*, 2007).
At the floor of the Bellsund Trough MSGL occur with a predominantly south-westward direction (Fig. 4.1). They are up to 15 km in length, less than 5 m in height and with an average spacing between the ridges of about 200 m. At the outer part of the trough, ridges parallel to the lineations are observed. The ridges are 5 km wide, about 35 km in length and up to 80 m in height. The presence and dimensions of a fast flowing ice stream can be inferred from the MSGL and the parallel ridges on either side of them. The ice stream was about 40 km wide close or at the shelf edge. Small transverse ridges are superimposed on the MSGL (Fig. 4.1). They are up to 5 m high and an average spacing of about 300 m. These ridges are interpreted as push moraines formed due to small readvances during a deglaciational retreat after the deposition of the MSGL. The ridges indicate that the ice margin was grounded during the retreat, at least in parts of the trough (Ottesen *et al.*, 2007).

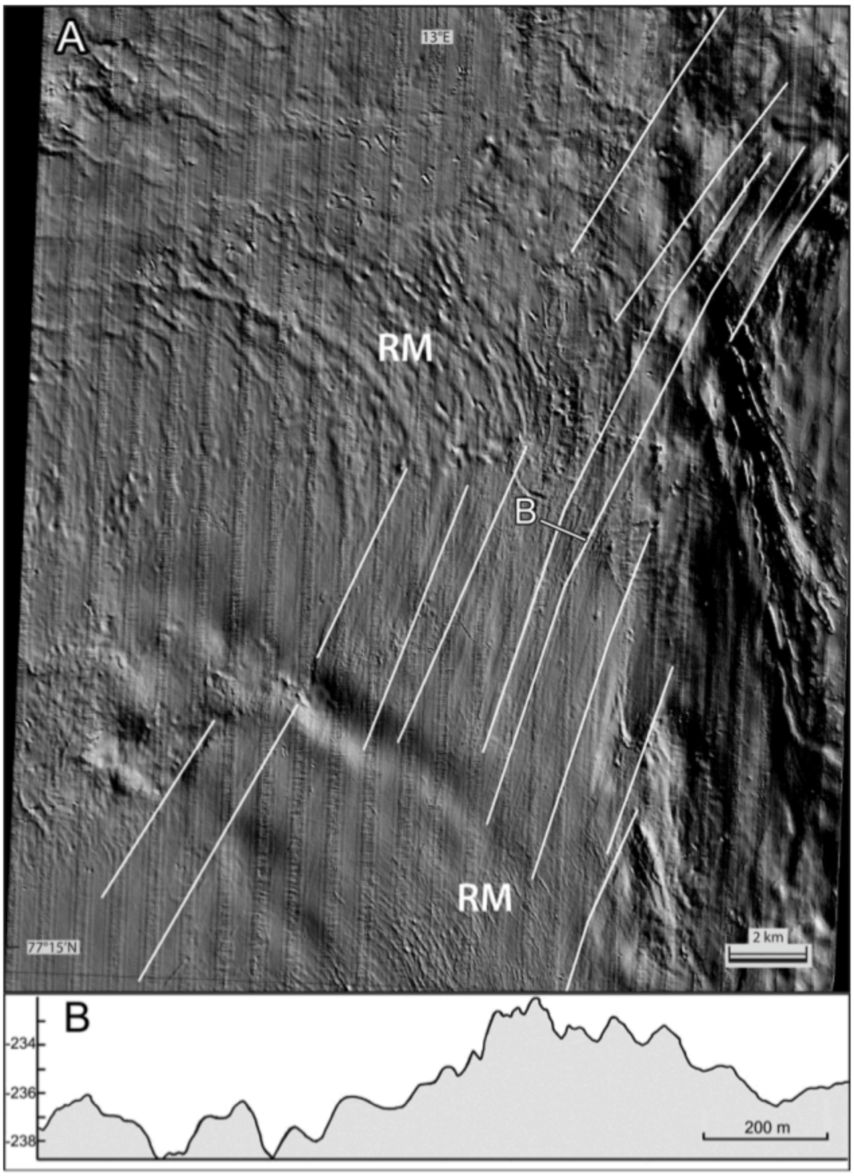

Fig. 4.1 A. Detailed swath bathymetry of the Bellsund Trough. The image shows glacial lineations (white lines) in the eastern and lower part with a NNE-SSW direction. A series of recessional moraines is marked with RM. The image is from the deepest part of the Bellsund Trough with water depth between 150 and 260 m. B. Cross-section of the amplitude and wavelength of glacial lineations. (Ottesen et al. 2007).

4.1.2 Isfjorden trough and fjord system

The Isfjorden glacial system is about 200 km long and extends from the fjord heads in the east to the shelf break west if Svalbard. Isfjorden is from 10 to 25 km in width and comprises of several basins which are up to 400 m deep and are separated by bedrock sills or moraines. Beyond the outer coast the trough widens to about 50 km close to the shelf edge, and reaches its greatest depth, ~350 m, on its northern side (Fig. 3.1) (Ottesen *et al.*, 2007). The inner fjord systems and fjord heads are not presented with bathymetry data in this paper.

MSGL are observed in the trough and the outer fjord, with an average spacing of 200-250 m. Some of the lineations are more than 10 km long and they are a few meters high. Ridges transverse to the main fjord axis along with some glacially sculpted bedrock features, most likely rock drumlins, are observed in the trough and outer fjord (Ottesen *et al.*, 2007). Svendsen *et al.*, (1992,1996) showed that that during the Last Glacial Maximum (LGM) the Late Weichselian Ice Sheet extended to the mouth of the cross-shelf trough and that during the deglaciation a major readvance occurred in the trough. A curved wide ridge extends to the shelf edge on the southern side of the trough, representing the troughs southern limit (Fig. 3.1). The ridge has a curved form and is thought to have been formed at a lateral ice stream margin, deposited during several glaciations with the Late Weichselian being the last one to modify it.

A large curved ridge transverse to the former ice-flow direction is located in the Isfjorden trough, about 50 km from the coastline and 15 km southwest of Prins Karls Forland (Fig. 4.2). The ridge crosses most of the trough and is about 10 km in width and up to 40 m in height and is in water depths of 150 to 250 m. The ridge is interpreted to be a grounding zone wedge. The ridge is both superimposed on MSGL and has similar features on its surface (Ottesen *et al.*, 2007). The ridge is similar in features to grounding zone wedges reported from Antarctica and the north Norwegian margin (McMullen *et al.*, 2006; Ottesen *et al.*, 2007). Assuming that the grounding zone wedge represents a still stand of the ice margin during retreat, it can be argued that the ice was grounded but continued to flow rapidly during and after the deposition of the wedge. Therefore during deglaciation there was an ice stream. Similar ridges are not found between this ridge and inner Isfjorden (Ottesen *et al.*, 2007).

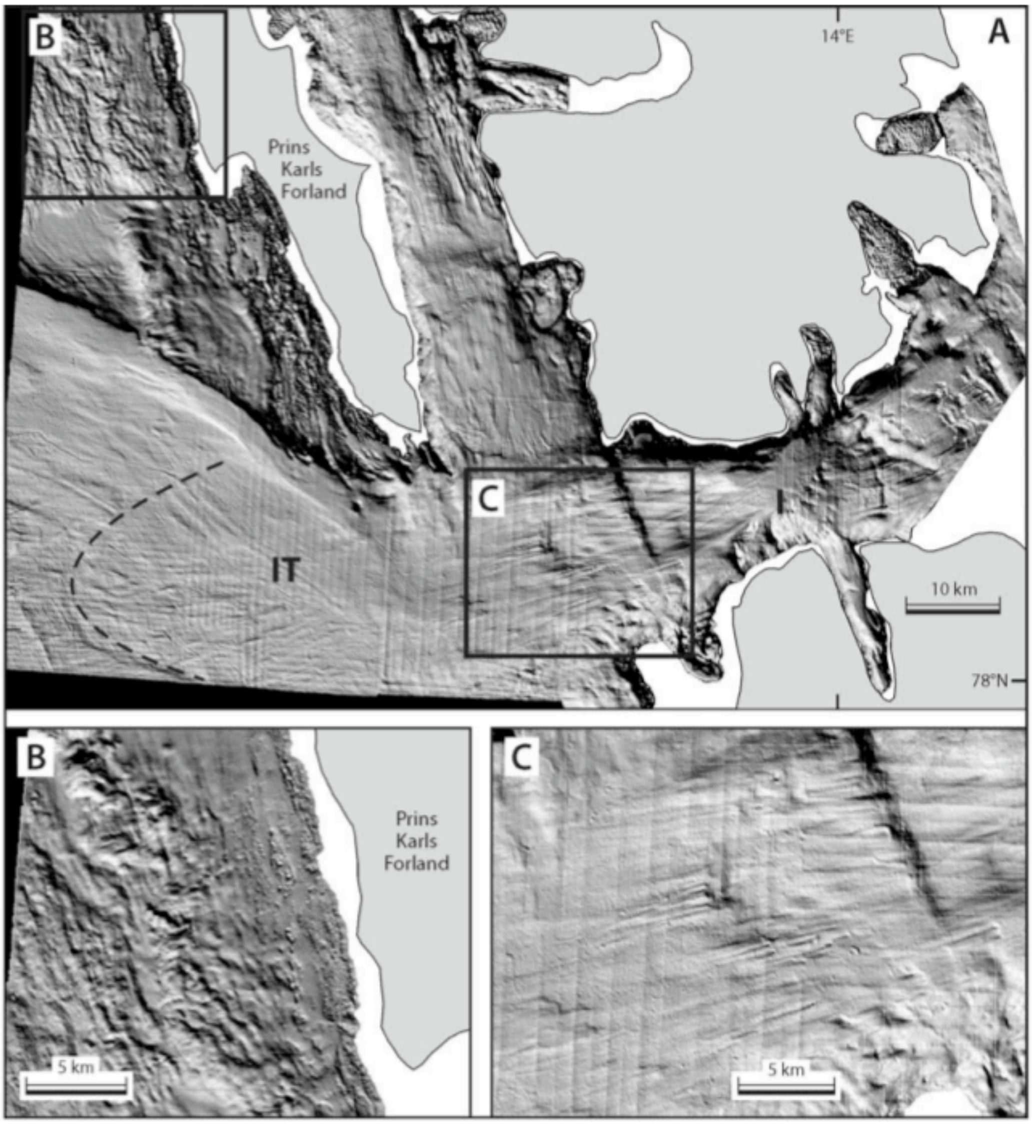

Fig. 4.2 A. Swath bathymetry of Isfjorden (I) and Isfjorden Trough (IT). The dashed line indicates a grounding-zone wedge, which is also described in Svendsen et al. (1992, 1996). B. A series of ridges on the shallower banks north of Isfjorden Trough. C. Mega-scale glacial lineations at the mouth of Isfjorden. (Ottesen et al. 2007).

4.2 Northwestern Spitsbergen

4.2.1 Kongsfjorden trough and fjord system

The Kongsfjorden trough and fjord system comprises of Kongsfjorden, 5-10 km wide and about 30 km long with a maximum water depth of 400 m, Krossfjorden, similar in length but narrower with water depth of 370 m (Sexton *et al.,* 1992), and the Kongsfjorden Trough which is 10 km wide at its inner most part. The inner most part of the trough reaches depths of 350 m but shallows to 200 m and widens to 30 km across toward the shelf edge (Fig. 3.1, 4.3) (Ottesen *et al.*, 2007).

MSGL are observed in most of the Kongsfjorden Trough (Fig. 4.3). In the middle of the trough a large ridge transverse to the former ice-flow directions is observed. The ridge is about 5 km wide and up to 30 m in hight and superimposed on the MSGL with similar features on top of it (Ottesen *et al.*, 2007). During the LGM the lineations in the outer part of the trough were most likely formed while the lineations in the fjord were probably formed during deglaciation (Landvik *et al.*, 2005).

A ridge parallel with the trough axis is observed on the southern side of the trough. It is about 5 km wide and up to 50 m high and is interpreted as a lateral ice stream margin, developed in the shear zone between the fast flowing ice stream in the Kongsfjorden Trough and the slower moving ice covering the shallow shelf outside Prins Karls Forland (Fig. 4.3). A large transverse ridge marking the maximum extent of the ice stream is located at the shelf break in the southern part of the trough (Fig. 4.3) (Ottesen *et al.*, 2007).

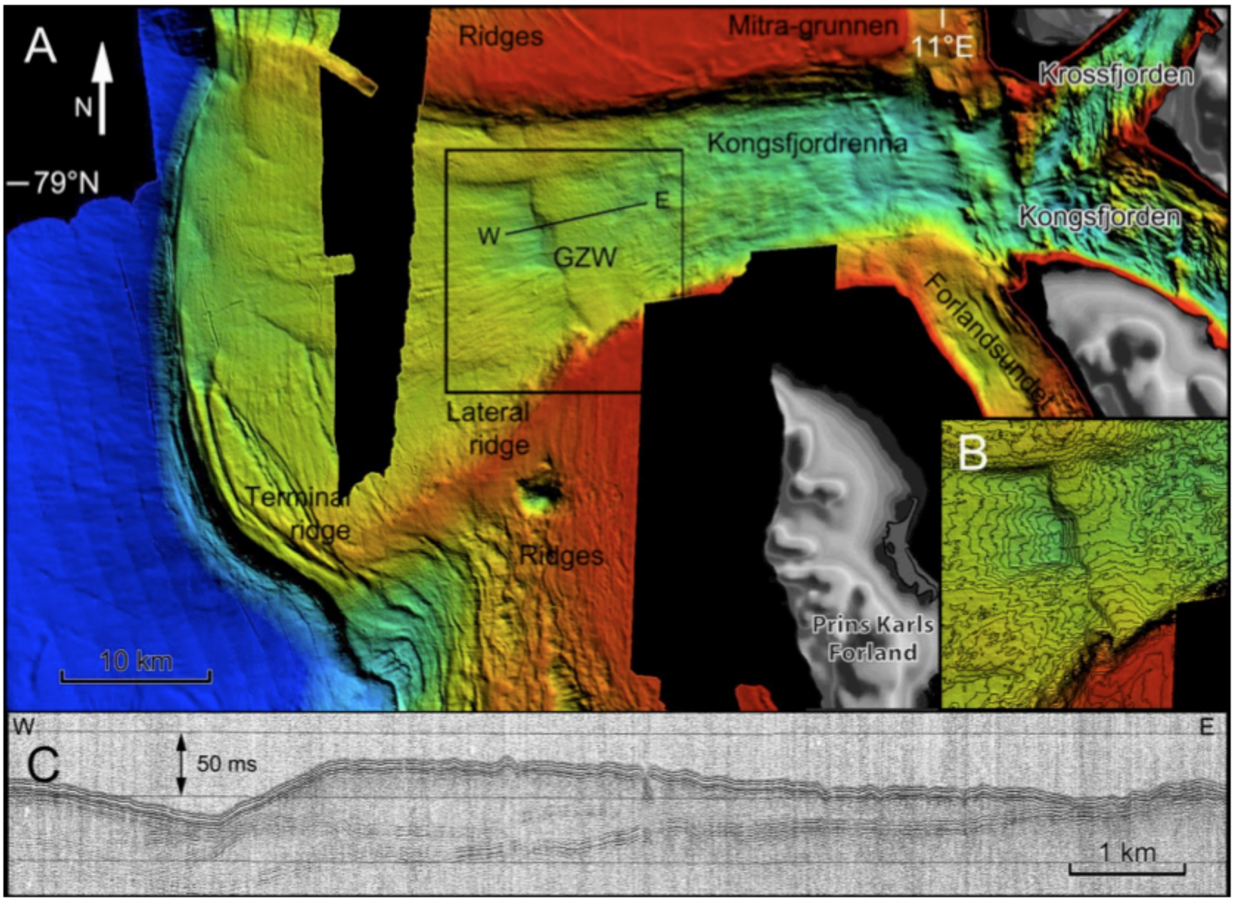

Fig. 4.3 A. Swath bathymetry of the Kongsfjorden – Kongsfjorden Trough system. Extensive glacial lineations are found on the sea floor, generally orientated parallel with trough axes. Outside the curved shelf edge, a series of downslope-orientated features, which may represent glacigenic debris flows related to maximum ice front position, are imaged. B. In the middle of Kongsfjorden Trough a grounding-zone wedge is found (GZW). 5 m depth contour interval. C. Sparker profile NP05-11-10 across the grounding zone wedge. (Ottesen et al. 2007).

4.2.2 Woodfjorden trough and fjord system

The Woodfjorden glacial system is in total about 150 km long, extending from the fjord head in the south to the cross shelf break north of Svalbard (Fig. 3.1). In the inner part of Woodfjorden the trough is about 5 km in width widening to 10 km at the mouth and further to a maximum of 35 km at the shelf edge. Water depths in the cross-shelf trough are in general about 200 m (Ottesen *et al.*, 2007).

MSGL parallel to the trough axis are observed in the Woodfjorden and Woodfjorden cross-shelf trough (Fig. 4.4). On the shelf beyond the mouth of Woodfjorden the set of lineations diverges into two sets of lineations around the island Moffen, the eastern set converges with the Wijdefjorden set while the western set turns to the NW continuing parallel with the trough axis to the shelf edge (Fig 3.2, 4.4a, 6.2) (Ottesen *et al.*, 2005, 2007). The lineations have an average spacing of 300 m. They are several km in length and are in average 3 m in height. Figure 4.4d shows a single swath-bathymetry line with MSGL parallel with the trough axis in the outer part of the cross-shelf trough. Transverse ridges, 10 km long and up to 10 m in hight, are found superimposed on the lineations in several areas of the trough (Fig. 4.4c). Large lateral ridges are found on both sides of the outer part of the trough (Fig. 3.1). These large lateral ridges are at least 30 km long, up to 50 km in height and 5 km wide. They are interpreted to be ice stream shear marginal moraines, and together with the lineations, they indicate substantial flow through the trough system during the LGM. The transverse ridges indicate a stepwise retreat during the deglaciation (Ottesen *et al.*, 2007).

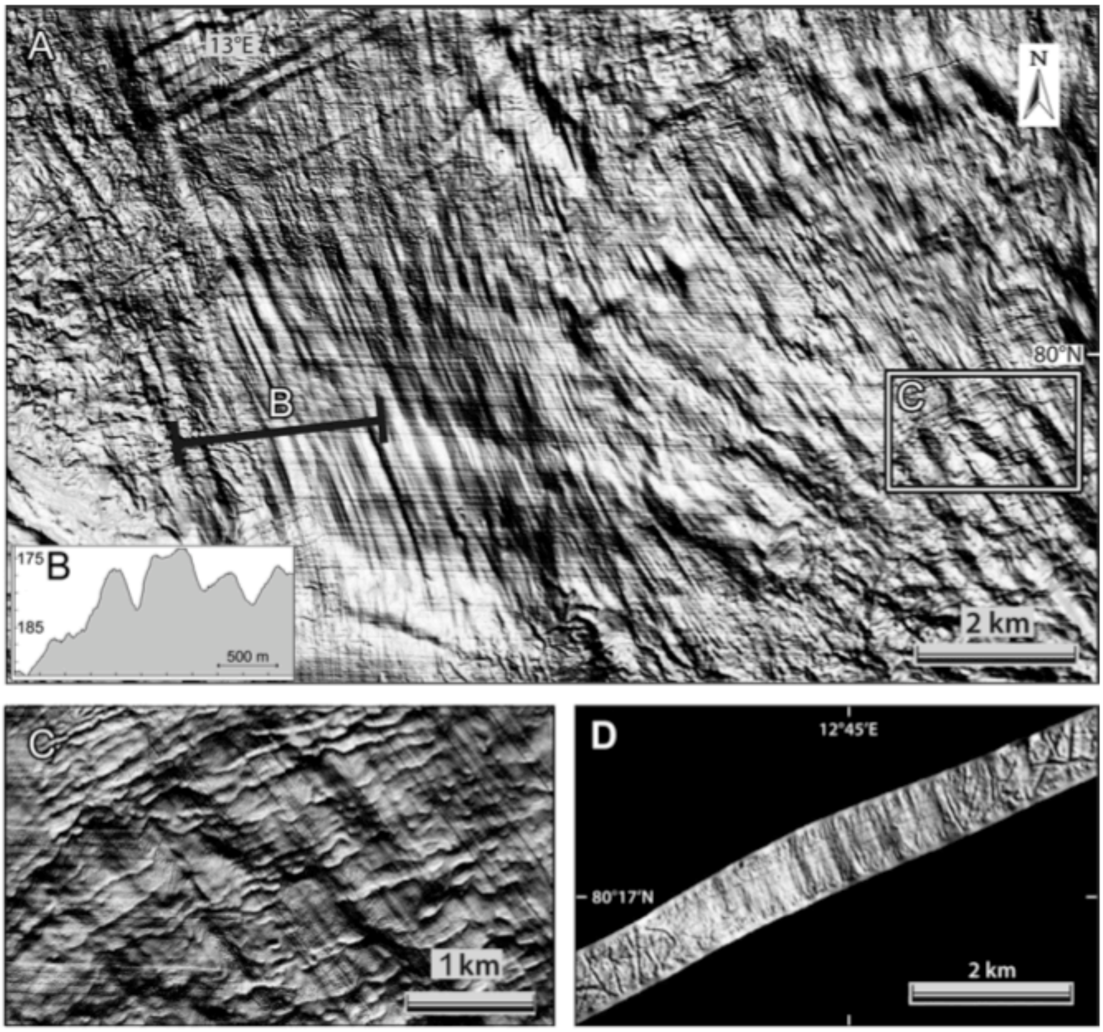

Fig. 4.4 A. Shaded-relief image of the sea floor of the Woodfjorden cross-shelf trough on the northern side of Svalbard. Glacial lineations follow the long-axis of the trough. B. Vertical profile across a series of glacial lineations in the Woodfjorden trough. C. A series of recessional moraines on a surface with glacial lineations. D. A single multibeam bathymetric line from the outer part of the Woodfjorden cross-shelf trough. Glacial lineations in the deepest part of the trough generally follow the trough axis. Maximum water depth is 200 m. (Ottesen et al. 2007).

4.2.3 Wijdefjorden and Hinlopen trough and fjord system

The Wijdefjorden trough is 110 km long extending from the head of the fjord in the south, across the shelf until it emerges with the Hinlopen trough northeast of the island Moffen (Fig. 3.1). Wijdefjorden is 5 km wide in the inner part, 20 km wide at the mouth and slightly wider crossing the shelf. Water depths in the fjord are usually less than 200 m. Hinlopenstredet is a 5-10 km wide and 110 km long trough between Spitsbergen and Nordaustlandet that continues north to the shelf edge as a 400 m deep cross-shelf trough, Hinlopen Trough, (Fig. 4.5). A shallow bedrock threshold is located west of Wahlenbergfjord in Nordaustlandet. North of this threshold the rather narrow trough is ~400 m deep towards another threshold near the shelf edge (Ottesen *et al.*, 2007).

In Wijdefjorden an extensive pattern of lineations parallel to the troughs axis are observed (Ottesen *et al.*, 2005). The lineations are up to 20 km in length, between 100-400 m in width and up to 20 m in height. The distance between each ridge range from 500-100 m. MSGL parallel to the troughs axis are observed in Hinlopenstredet (Fig. 4.5). The pattern of the MSGL in Wijdefjorden shows that the ice that flowed through the trough was confluent with an ice stream in the Hinlopen Trough ending at the shelf break 70 km further north (Fig. 6.2) (Ottesen *et al.*, 2007).

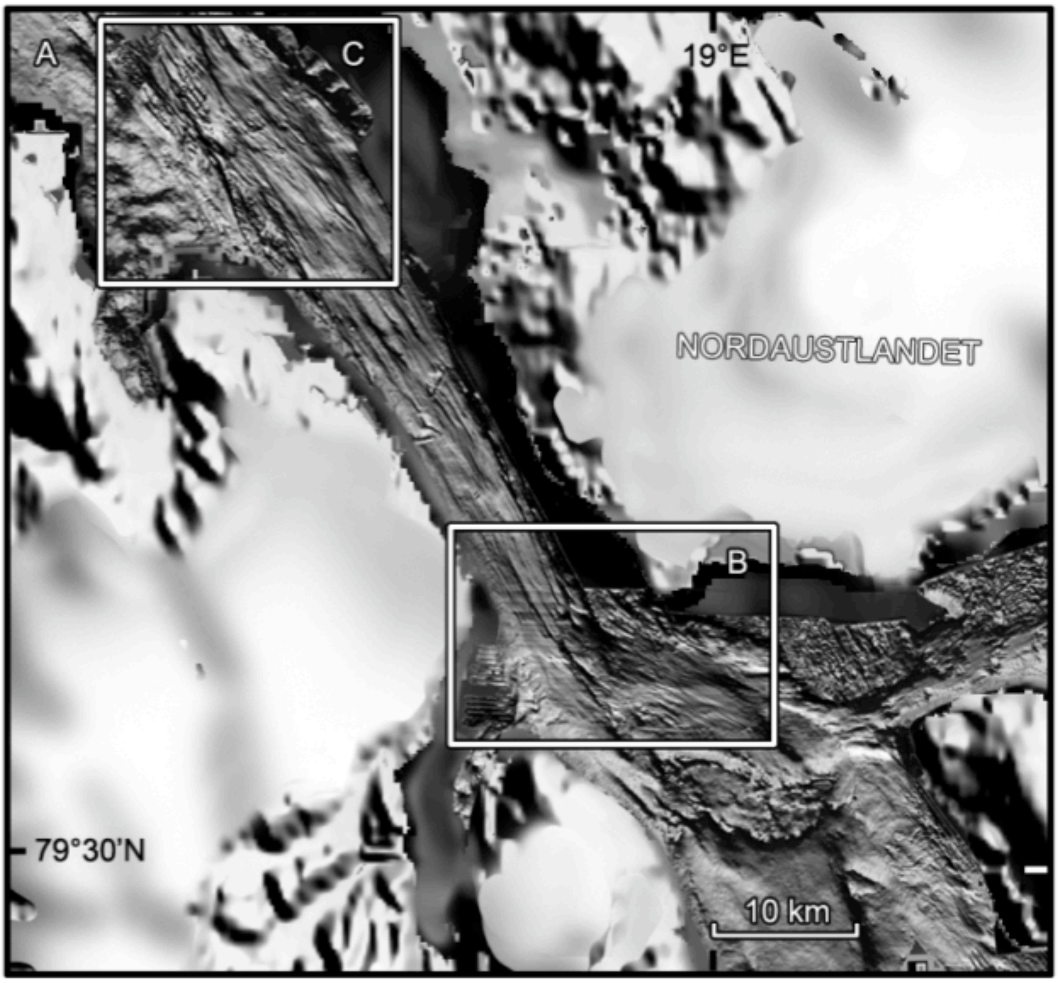

Fig. 4.5 A. Overview map of the detailed bathymetry of Hinlopenstredet area. (Ottesen et al. 2007).

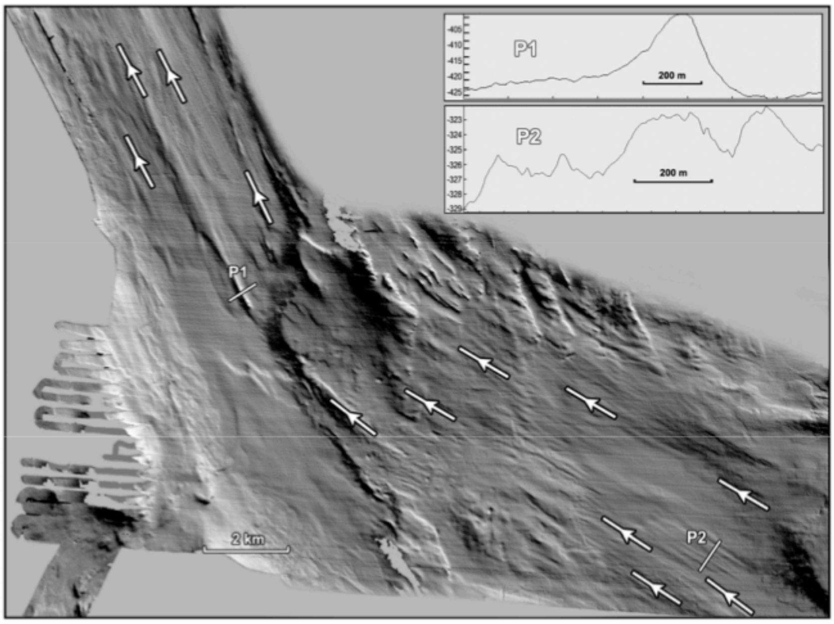

Fig. 4.5 B. Shaded relief image of the sea-floor morphology of the southern part of the Hinlopenstredet with two surface profiles. Extensive glacial lineations (white arrows) show ice movement from ESE into the narrow part of the Hinlopenstredet where it turns parallel to the trough axis towards NW. (Ottesen et al. 2007).

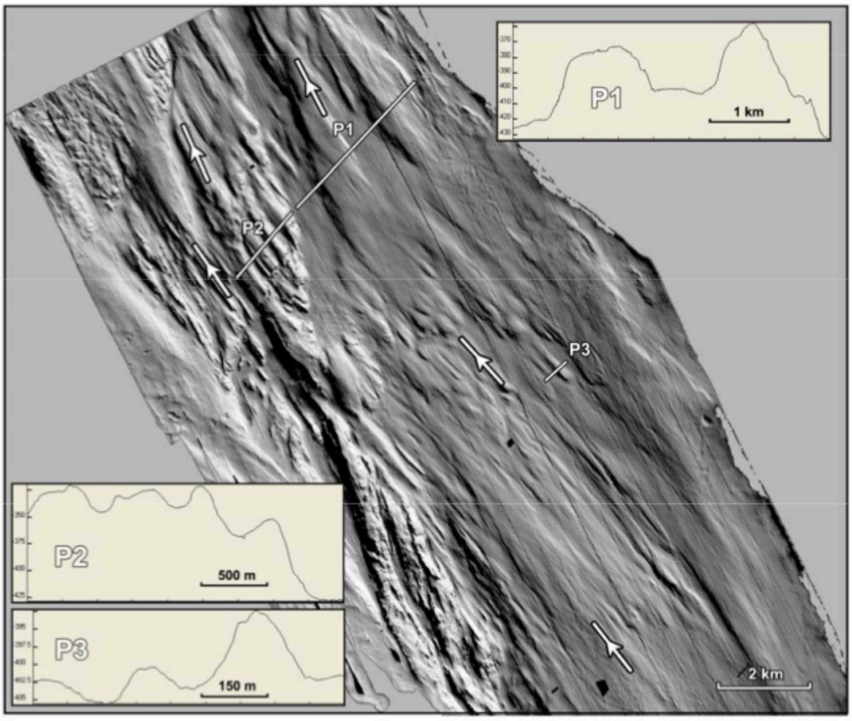

Fig. 4.5 C. Shaded relief image of the sea floor morphology of the northern part of the Hinlopenstredet with three surface profiles. Extensive glacial lineations (white arrows) parallel to the main axis of the Hinlopenstredet show ice-flow towards northwest. (Ottesen et al. 2007).

5 DEGLACIATION

5.1 Western Spitsbergen

On Brøggerhalvøya on western Spitsbergen there occurs a sequence of raised beaches (Forman *et al.*, 1987; Landvik *et al.*, 1987; Forman, 1990; Forman *et al.,* 2004) that has been divided into three distinct age groups (Fig. 5.1) on the basis of the degree of terrace dissection, preservation of individual shorelines and the extent of pedogenesis (Forman and Miller, 1984; Forman *et al.,* 2004).

Between 80 and 55 m aht the oldest terrace sequence is observed, c. >140 ka. It is highly dissected with only 20-40 m long remnants of the original surface preserved. These deposits can be traced intermittently and lack distinctive shoreline morphologies. They contain a silt-rich B-horizon that exceed 80 cm in thickness. Between 44 and 55 m aht there is a sequence of an intermediate age, c. 60-80 ka, with the thickness of 70-50 cm (Forman and Miller, 1984; Forman, 1990; Forman *et al.,* 2004).

At 45 m aht there is a distinct change in geomorphic expression (Fig. 5.1); at and below this level the terrace exhibits exceptionally well preserved beach morphologies with B-horizons that are less than 30 cm thick. This prominent geomorphic boundary has been dated at c. 13-11 ka on Brøggerhalvøya as well in other areas of western and northern Svalbard (Forman *et al.*, 1987; Landvik *et al.*, 1987; Forman, 1990; Forman *et al.,* 2004).

By looking at the lowest and youngest raised beach sequence from the last deglacial hemicycle we can see changes in the beach ridge morphology with altitude that can be related to the rate and direction of the relative sea level change (Fig. 5.1). Between 20 and 45 m aht there are three large beach ridges, they have broad crests (100-200 m wide) and a relief up to 5 m. Below 20 m aht there are numerous low (< 2 m) and narrow (5-10 m) strandlines that reach down to the present shore (Forman *et al.,* 2004).

The sequence form below 45 m aht show that as the relative sea level fell slowly it was interrupted by at least three short periods of still stands or possibly transgressions that caused the formation of the three large beach ridges between 20 and 45 m aht (Fig. 5.1). After the formation of the lowest ridge, at c. 30 m aht, sea level fell rapidly leaving only small numerous strandlines down to the present shore where a large barrier beach ridge is actively forming in response to the ongoing transgression (Forman *et al.*, 1987; Forman *et al.,* 2004). On Kapp Guissez and Mitrahalvøya (Forman, 1990), to the north, we find morphologically similar raised beach sequences with massive beach ridges near the marine limit (found at different altitudes due to the regional variations in isostatic depression) followed by minor ridges reaching the present shore correlates to the reconstruction of relative sea level dynamics (Forman *et al.* ,2004).

At 45 m aht there is another striking feature, as there is a breach in the shoreline accentuated by curved terrence remnants, oblique to the lower beaches (Fig. 5.1, shown by arrows). This breach may represent the building of the spit when the LWML was established, and subsequently eroded. The formation of spits is usually in shallow coastal waters where there is abundant sediment supply and long-shore drift predominates, rather than storm generated fetch. On Brøggerhalvøya an ice-covered sea would have dampened severely the dominant westerly fetch and favored longshore drift in the near shore leading to the built up of the spit. Later a switch to the modern wave conditions and intensity, caused by ice-free conditions, that resulted in the truncation of the spit (Forman *et al.* 2004). Remnants of spits have been found at the marine limit on other forelands of western Spitsbergen, including Mitrahalvøya, Sarsøya, Daudmansøya and southern Prins Karls Forland (Forman, 1990; Andersson *et al.*, 1999; Forman *et al.,* 2004).

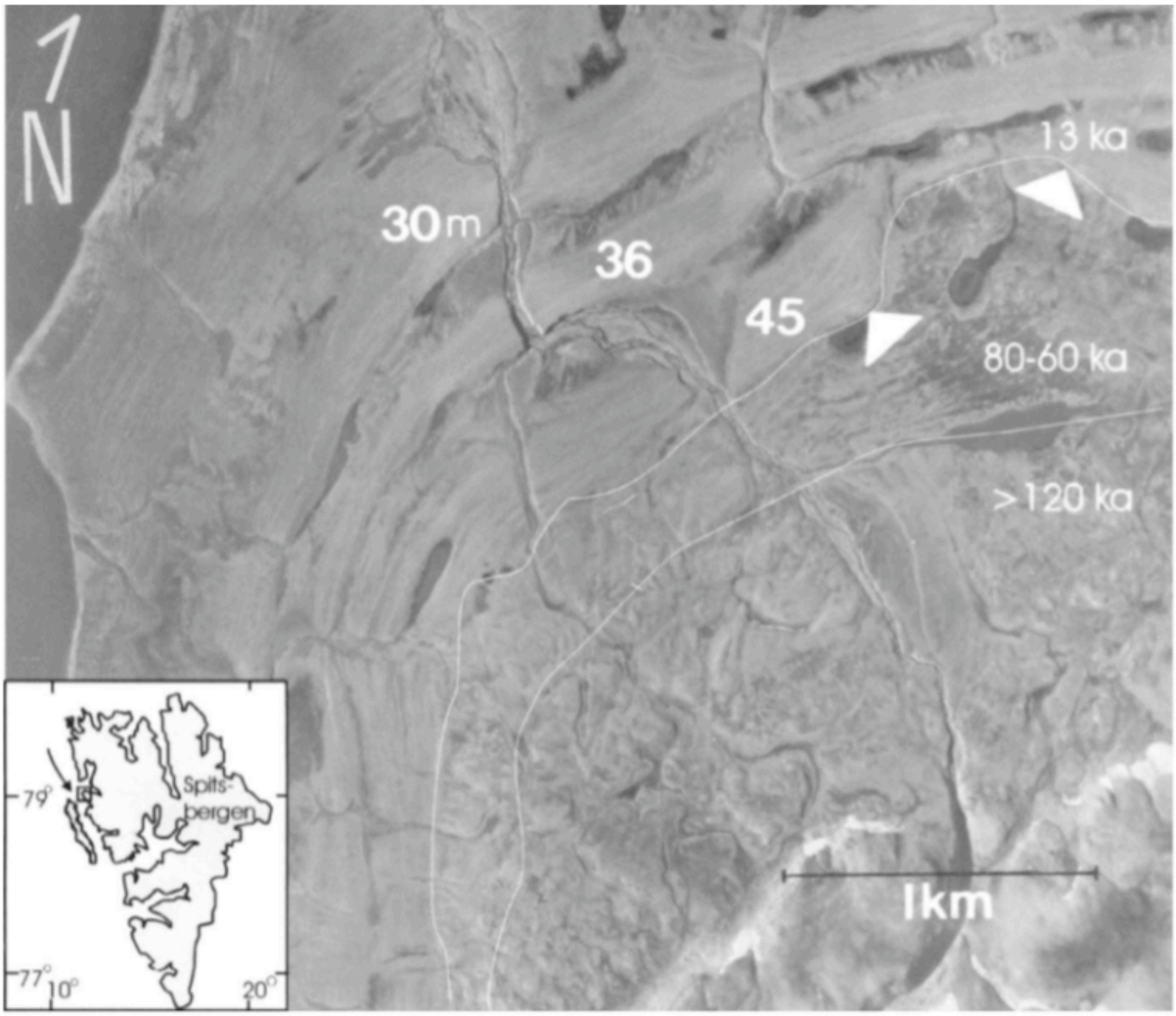

Fig. 5.1 Aerial photograph (S70-4231, copyright Norsk Polarinstitutt, Oslo) of Brøggerhalvøya, western Spitsbergen. Showing the raised-beach sequence the oldest extending up to >80 m aht and dated to >120 ka (note that this date is not the same as in the text, correct date is >140 ka (Forman et al. 2004)). The youngest sequence is demarcated by Late Weichselian marine limit at 45 m aht and is expressed as a truncated spit-cusp, shown by arrows. Three large raised barrier beaches occur at 45, 37, and 29 m aht. Below 29 m aht, numerous discrete strandlines occur down to the modern shore. (Forman et al., 1984; Forman et al. 2004).

On Brøggerhalvøya two whalebones that were collected above the LWML yielded infinite ^{14}C ages (>36 ka) on the collagen fraction, supporting pedologic and geomorphic interpretations that on Brøggerhalvøya there are raised beaches from the middle Weichselian or earlier (Forman and Miller, 1984). From the LWML at 20 m aht at Mitrahalvøya whalebones were dated to 12,960+/-190 yr. BP (B-10986) (Forman *et al.*, 1987). At these two sites the marine limit was established basically at the same time because Kapp Mitra and Brøggerhalvøya were either glaciated or unglaciated early at similar times (ca >13 ka) (Lehman and Forman, 1992). From a collection of bones on the 37 m beach ridge on Brøggerhalvøya a whale rib was dated to 11,760+/-430 yr BP (GX-9990) (Forman *et al.* 1987). This is one of the oldest ages associated with raised beach deposits on Svalbard so for verification another half of the rib was dated in a second laboratory. The second age, 11,800+/-180 yr BP (I-13793) (Forman *et al.* 1987), correlates with the original age, giving a good confidence to the dating (Forman *et al.*, 2004). Also associated with LWML landforms are deposits with the radiocarbon ages of ca 12.5-11 ka near Bellsund (Landvik *et al.*, 1987) and southern Prins Karls Forland (Forman, 1990; Andersson *et al*. 1999) and to the north in Woodfjorden (Brückner *et al.*, 2002). From 30 m aht to the present shore on Brøggerhalvøya whalebones range in age between 10,275+/-90 yr BP (DIC-3122) (Forman *et al.*, 1987), and 9230+/-340 yr. BP (GX-9908) (Forman *et al.*, 1987) with most ages overlapping at two standard deviations (Fig. 5.2). For other sites on western Spitsbergen including Daudmannsøyra, Southern Prins Karls Forland (Forman, 1990), Bellsund (Landvik *et al*., 1987), and Erdmannflya and Bohemanflya (Salvigsen *et al*., 1990), similar apparent rapid rates (2-3 m/100 yr.) of emergence have been documented. The rapid emergence probably reflects an elastic crustal response to ice unloading ca 10,000 yr. BP (Forman *et al*., 2004).

Over broad areas of western and northern Svalbard there are abundant evidences for a sea level oscillation in the middle Holocene. At 5 m aht near a modern storm beach at Tønsneset, on the north shore of Kongsfjorden, a whalebone is yielded at an age of 5900+/-210 yr. BP (GX9899) (Forman *et al*., 1987), indicating that middle Holocene sea level was similar or slightly higher than today (Fig. 5.2). This high sea level event between 6 and 4 ka ago is also found at other sites along the western Spitsbergen (Landvik et el., 1987; Forman, 1990), northern Spitsbergen (Brückner et al., 2002) and on Phippsøya (Forman and Ingólfsson, 2000). Between Isfjorden and Bellsund, on western Spitsbergen (Landvik *et al*., 1987) there is prominent constructional beach and erosion and truncation of older raised beaches on Phippsøya (Forman and Ingólfsson, 2000) that are associated with this high stand (Forman *et al*., 2004). This dated sea level event is often associated with the first pumice level which across Svalbard (Blake, 1961; Salvigsen, 1978, 1984) and is widely recognized, but on Erdmannflya and Bohemanflya, on the northern shore of Isfjorden (Salvigsen *et al.*, 1990), it may have occurred up to 2 ka earlier (Forman *et al*., 2004).

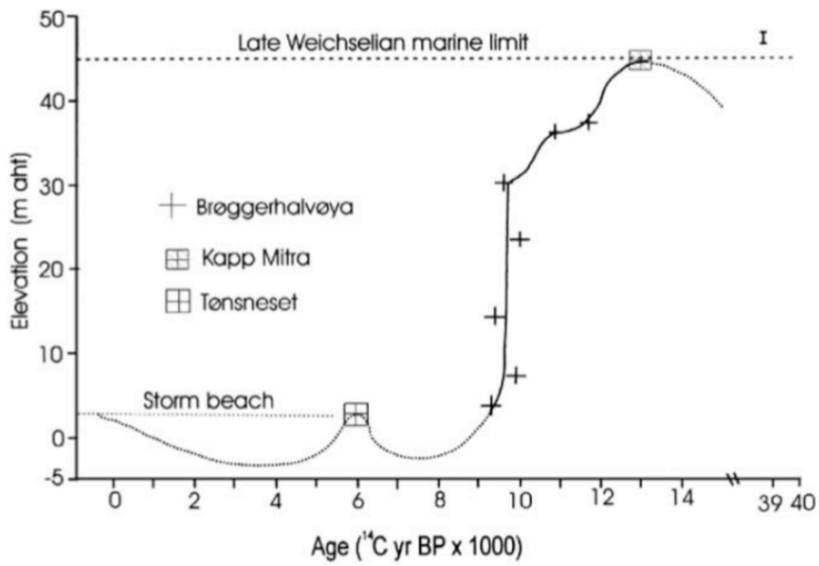

Fig. 5.2 Late Weichselian and Holocene relative sea level for Brøggerhalvøya, Spitsbergen. Note rapid fall in relative sea level ca 9.5 ka and inferred transgressive and regressive event ca 6 ka. (Forman et al., 1987; Forman et al., 2004).

5.2. Eastern Svalbard

Single generation of raised beaches exceeding 50 m aht are found on eastern Spitsbergen (Bondevik *et al*., 1995) and islands in the Barents Sea, like Storøya (Jonsson, 1993) and Hopen (Hoppe *et al*., 1969), that indicate full coverage and erosion by the Barents Sea ice sheet (Landvik *et al.,* 1998). For Nordaustlandet the postglacial emergence is not well constrained. Blake (1961) and Salvigsen (1978) only found two emergence records for this island, although it is one potential source for the ice sheet that covered the Barents Sea during the Late Weichselian (Forman *et al*., 2004).

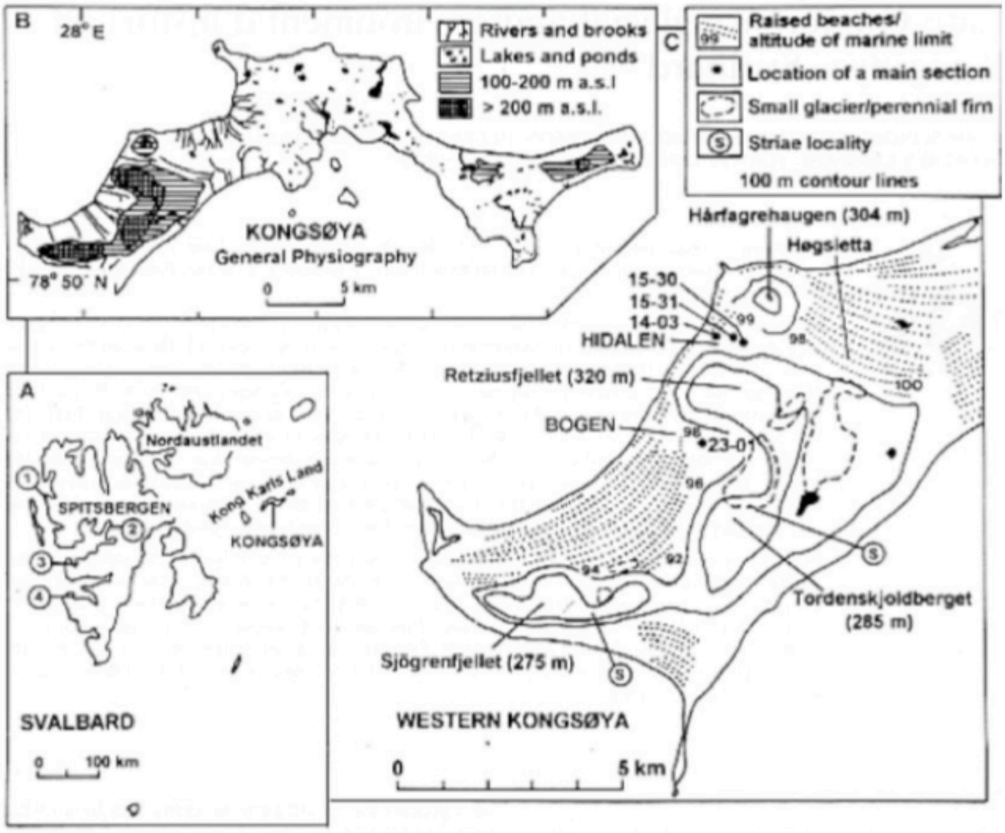

Fig. 5.3 Svalbard location map. B: The general physiography of Kongsøya. C: Western Kongsøya. The location of raised beaches is shown as an approximation of areas where beach ridges occur rather than location of individual ridges. (Ingólfsson et al., 1995).

On Kong Karls Land the highest deglacial strandlines are recognized (Salvigsen, 1981; Ingólfsson *et al*., 1995). On western Kongsøya, which is the largest of a number of islands comprising Kong Karls Land, the marine limit is at 94-98 m aht (Fig. 5.3c). On Høgsletta (Fig. 5.3c) the highest beach ridges are leveled to ca 100 m aht (Fig. 2.2) (Ingólfsson *et al*., 1995) although many of them are obscured by solifluction (Forman et al., 2004) (Fig. 5.4). Between ca 40 m and 60-70 m aht on Høgsletta the beach ridges are dated to have been formed between ca 9000 BP and 7000 BP (Salvigsen, 1981) and are composed of well-rounded cobbles and boulders, with little gravel or sand, while the finer material is more characteristic for the beach ridges above and below. Ingólfsson et al. (1995) interpreted that this could indicate that during the formation of the intermediate altitude beach ridges there was higher energy in the coastal environment and less annual sea-ice cover. Lack of driftwood on the beach above 35-40 m aht supported this (Ingólfsson *et al*., 1995). The 100 m aht raised beach on Kongsøya is securely constrained by a ^{14}C age on *Larix* sp. log of 9850+/-40 yr BP (GSC-3039), indicating full deglaciation by at least ca 10 ^{14}C ka BP (Salvigsen, 1981; Forman *et al*., 2004). Indication of appreciable loading are found at marine limits at 90 m aht in Billefjorden, Spitsbergen, and at 85-90 m aht on Barentsøya and Edgeøya (Bondevik et al., 1995) with deglaciation of the latter dated to ca 10-10.4 ^{14}C ka (Landvik et al., 1998; Forman et al., 2004). On Hopen emergence shows unusual near linear emergence since ca 9.4 ^{14}C ka (Fig. 2.1), which may reflect initial emergence under a thinning ice sheet, with a subsequent declining rate of emergence post deglaciation (Forman *et al*., 2004). From southern Nordaustlandet (Salvigsen, 1978), Barentsøya and Edgeøya (Bondevik *et al*., 1995) emergence records show a fluctuation emergence rate at ca 6 ka, which may reflect the middle Holocene transgression documented at other localities where total emergence is < 70 m (Forman *et al*., 1987; Landvik *et al*., 1987; Forman, 1990; Forman and Ingólfsson, 2000; Brückner *et al*., 2002; Forman *et al*., 2004) (Fig. 2.1).

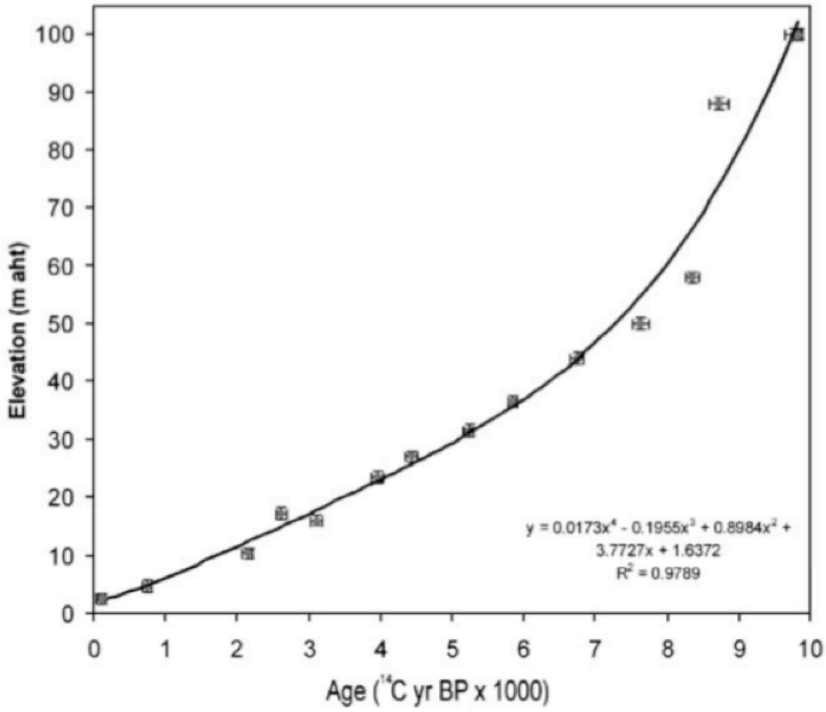

Fig. 5.4 Holocene relative sea level for Kongsöya, Svalbard, which exhibits the greatest post-glacial emergence in the Barents Sea. (Salvigsen, 1981; Forman et al., 2004).

5.3. Post-glacial emergence on Svalbard

Most chronologic evidence places the retreat of the western and northern margins of the Barents Sea ice sheet on to Svalbard by ca 13,000-12,000 ^{14}Cyr ago (Forman *et al.*, 1987; Svendsen *et al*., 1992; Elverhøi *et al*., 1995; Lubinski *et al*., 1996; Landvik *et al*., 1998; Brückner *et al*., 2002; Forman *et al*., 2004). Between ca 13,000 and 10,500 ^{14}Cyr BP the LWML occurred and many forelands on western and northern Spitsbergen were characterized by construction of spits, with longshore drift predominating over storm-generated fetch. The dominance of long shore drift and the paucity of driftage associated with the LWML may reflect extensive sea-ice coverage of coastal regions of the western Spitsbergen, blocking the passage of whales and driftwood laden sea ice and weakening the prevailing westerly waves. The presence of whalebones, although rare, indicate that the sea ice periodically scattered in the Norwegian Sea to allow the migration of whales to Svalbard (Forman *et al*., 2004). The Last Glacial Maximum is characterized by the variability in sea surface conditions when there were submillennial-scale oscillations in the dominance of North Atlantic and Arctic water masses off of Svalbard (Hebbeln *et al*., 1994; Forman *et al*., 2004). the emergence rate during ca 13,000 and 10,500 ^{14}Cyr was relatively slow (1.5-5 m/ka) reflecting the rate of isostasy just exceeding eustatic rise of sea level for approximately 15 m/ka for this interval (Fairbanks *et al*., 1989; Forman *et al*., 2004).

For many sites on western and northern Spitsbergen relative sea level fell rapidly (15-30 m/ka) between 10,500 and 9000 ^{14}Cyr BP (Fig. 5.2 of Brøggerhalvøya) depositing beach ridges parallel to the present shoreline, indicating that wave directions were similar to that of today. For areas on eastern Spitsbergen (Salvigsen and Mangerud, 1991), Barentsøya and Edgeøya (Bondevik *et al*., 1995) and Kong Karls Land (Salvigsen, 1981) emergence started after 10,500-10,000 ^{14}Cyr BP (Forman *et al*., 2004).

On many localities on western and northern Spitsbergen, mid-Holocene (6-4 ^{14}C ka) transgressive-regressive cycles are recognized (Forman *et al*., 1987; Landvik *et al.*, 1987; Forman, 1990; Forman and Ingólfsson, 2000; Brückner *et al.*, 2002). The transgression did not reach higher than 7 m aht and is limited by a constructional terrace that cuts a early Holocene regressional strandlines. Marine subfossils associated with this transgressive feature were radiocarbon dated and the dates indicated that the sea occupied this level between 6000 and 4000 ^{14}Cyr BP. Also in areas on eastern Svalbard, where the total emergence is >70 m, there are a noticeable fluctuation in emergence rates centered at 6000 ^{14}Cyr BP, which may reflect the same sea level oscillation (Salvigsen, 1981; Salvigsen and Mangerud, 1991; Bondevik *et al.*, 1995).

6 DISCUSSION

The view of the beds beneath former ice sheets given by marine-geophysical records provides important data for the reconstruction of former ice sheet flow directions and dynamics. During the last 20 years a number of geomorphical and stratigraphical studies, along with glaciological modeling, have established that an ice sheet covered Svalbard, its continental margin and the Barents Sea during the Late Weichselian glacial maximum (Svendsen et al. 1996; Landvik et al. 1998; Siegert & Dowdeswell 2001; Hogan *et al.* 2010). From the glaciological and isostatical modeling the thickens of this ice sheet has been identified to have been up to 1500 m thick over the Barents Sea, with its center just east of Kong Karls Land in the northern Barents Sea (e.g. Siegert & Dowdeswell, 2001; Hogan *et al.,* 2010).
Ottesen *et al.* (2007) used the distribution of the submarine landforms shown in figure 6.1 to reconstruct the former ice-flow pattern in the fjords and adjacent shelves of western and northern Svalbard. Mega-scale glacial lineations are interpreted to indicate the presence of fast flowing ice streams within former ice sheets and lateral moraines often define their margins (Clark 1993; Ottesen *et al.* 2005, 2007). Major landforms observed in the fjords and troughs in western and northern Svalbard are mapped in figure 6.2. The distribution of the MSGL show that ice streams extended across the shelf and beyond each of the major fjords of western and northern Svalbard (Ottesen *et al.* 2007)

Submarine morphological evidences demonstrate that a large ice sheet reached the shelf of western and northern Svalbard at the LGM (Fig. 6.1) (e.g. Andersen *et al.*, 1996; Elverhøi *et al.* 1995; Landvik *et al*. 1998; Mangerud *et al.* 1998; Ottesen *et al.* 2005; Ottesen *et al*. 2007). It is therefore unlikely that a restricted ice sheet was over Svalbard (e.g. Boulton 1979; Troitsky *et al.* 1979), with some areas on the west coast ice free at the LGM (e.g. Miller 1982).

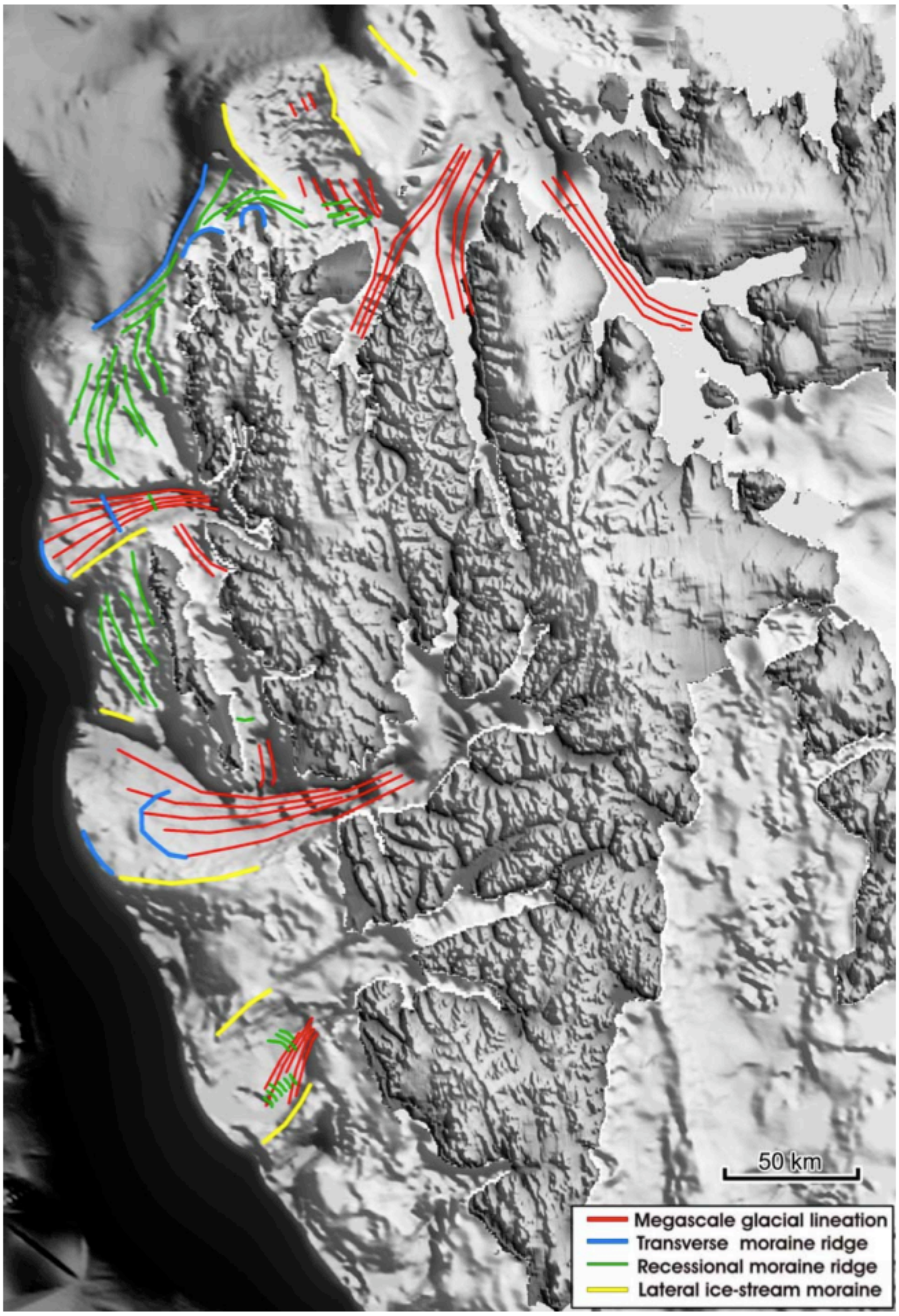

Fig. 6.1 Geomorphic map of the submarine landforms on the Svalbard margin derived from sea-floor imagery. (Ottesen et al. 2007).

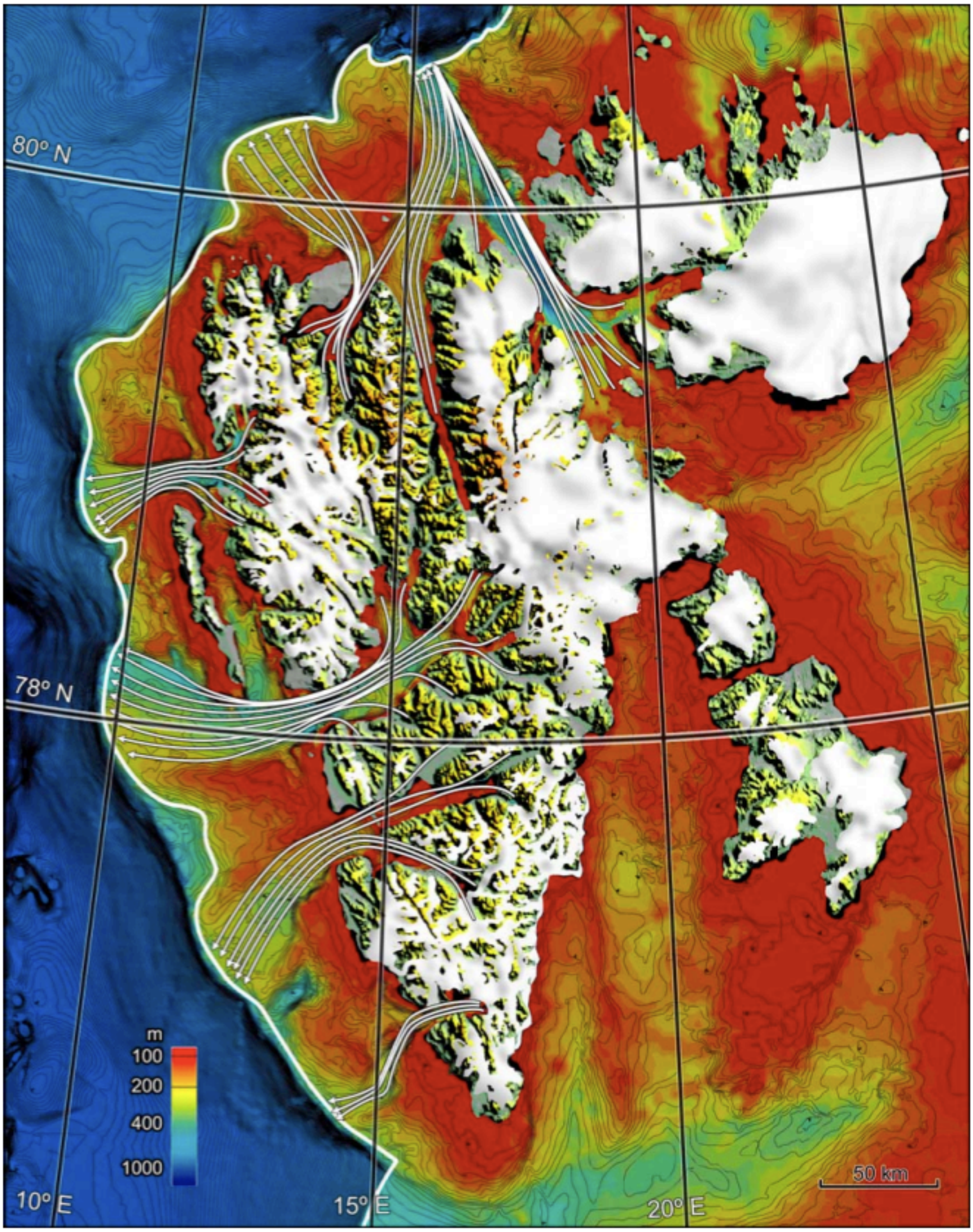

Fig. 6.2 Reconstruction of ice-sheet flow regime, including ice streams, on the western and northern margin of the Late Weichselian Barents/Svalbard ice sheets. An ice stream is assumed to be present in the fjord of Hornsund and the trough beyond, on the basis of analogy with the other systems of western and northern Svalbard. (Ottesen et al. 2007).

Western and northern Spitsbergen emergence curves provide the oldest (ca 13,000 ^{14}Cyr BP) but discontinuous post-glacial record of relative sea level for Svalbard (Fig. 2.1). This pattern of emergence correlates well to predictions for sites that were at or near the margin of a large ice sheet (Transition zone I/II of Clark *et al.*, 1978). On the other hand, shoreline displacement on eastern Svalbard and islands in the Barents Sea started ca 10,000 ^{14}Cyr BP and is continuing at present (Forman *et al.*, 1997). These records are similar to relative sea level predictions for areas that were beneath a substantial (>1 km) ice sheet load (Clark et al., 1978; Lambeck, 1995, 1996). For Svalbard the recognized spatial variations in emergence indicate that western and northern Spitsbergen was near the reactive margin of the ice sheet. Studies of the continental shelf and slope north of the Barents Sea place the maximum extend of the Late Weichselian ice sheet to its northern limit by ca 21-23 ^{14}Cyr BP and that the northern areas of Svalbard and Franz Josef Land were deglaciated by ca 15 ka (Landvik *et al.*, 1998; Kleiber *et al.*, 2000). This indicates that the marginal areas on northern Svalbard may have sustained relatively brief ice coverage (3000-8000 yr) under a variable ice sheet or ice stream flow and were probably not isostatic equilibrium, with glacio-isostatic unload/load half life of ca 2000 yr (Forman and Ingólfsson, 2000). On the other hand central and eastern Svalbard deglaciated by at least 10,500 ^{14}Cyr BP (Landvik *et al.*, 1998) and were covered by a substantial ice load and probably sustained this load for at least 10,000 yr and achieved isostatic equilibrium (Forman *et al.*, 2004). Earth rheology-based ice sheet models are predicated on isostatic equilibrium. This ice sheet model often model well the former ice sheet centers, but deviate from field observations for non equilibrium areas at the ice margin (Forman and Ingólfsson, 2000).

In southern most Spitsbergen the decline in elevation (19 m aht) of the marine limit and associated isobases (Forman, 1990; Forman et al., 1997; Landvik et al., 1998) as well as the lack of emergence on Bjørnøya (Salvigsen and Slettemark, 1995) indicate a minimal glacier loading and/or early deglaciation (before 10 ka); the former is supported from marine geologic records from the adjacent Bear Island Trough (Faleide et al., 1996).

The post-glacial pattern of emergence since 9000 and 5000 yr ago is assessed for Svalbard, Franz Josef Land and for the latter period for Novaya Zemlya (Fig. 6.3). The justification for using combinations of emergence data from Franz Josef Land, Svalbard and Novaya Zemlya is that the glacio-isostatic compensation reflects past glacial loads over 100s of kilometers, with half life response of approximately 2000 yr. Forman *et al.* (2004) hand contoured these isobases from ^{14}C-dated relative sea level records for individual raised strandplain sequences (Fig 2.1; Appendices A and B in Forman *et al.*, 2004). By looking at the 9000 ^{14}Cyr BP isobase we can define a broad zone of maximum emergence through the east and center of the Svalbard archipelago, with islands in the Barents Sea and eastern Svalbard registering the greatest emergence. Westward of isobases into Isfjorden and Van Mijenfjorden there is a noticeable deflection indicating areas of substantial ice sheet loading. The 5000 ^{14}Cyr BP isobase, though it shows at least 50% less emergence than the 9000 ^{14}Cyr BP isobase, portrays a similar pattern to the older isobases, there for an effective measure of past glacier loading (Forman *et al.* 2004).

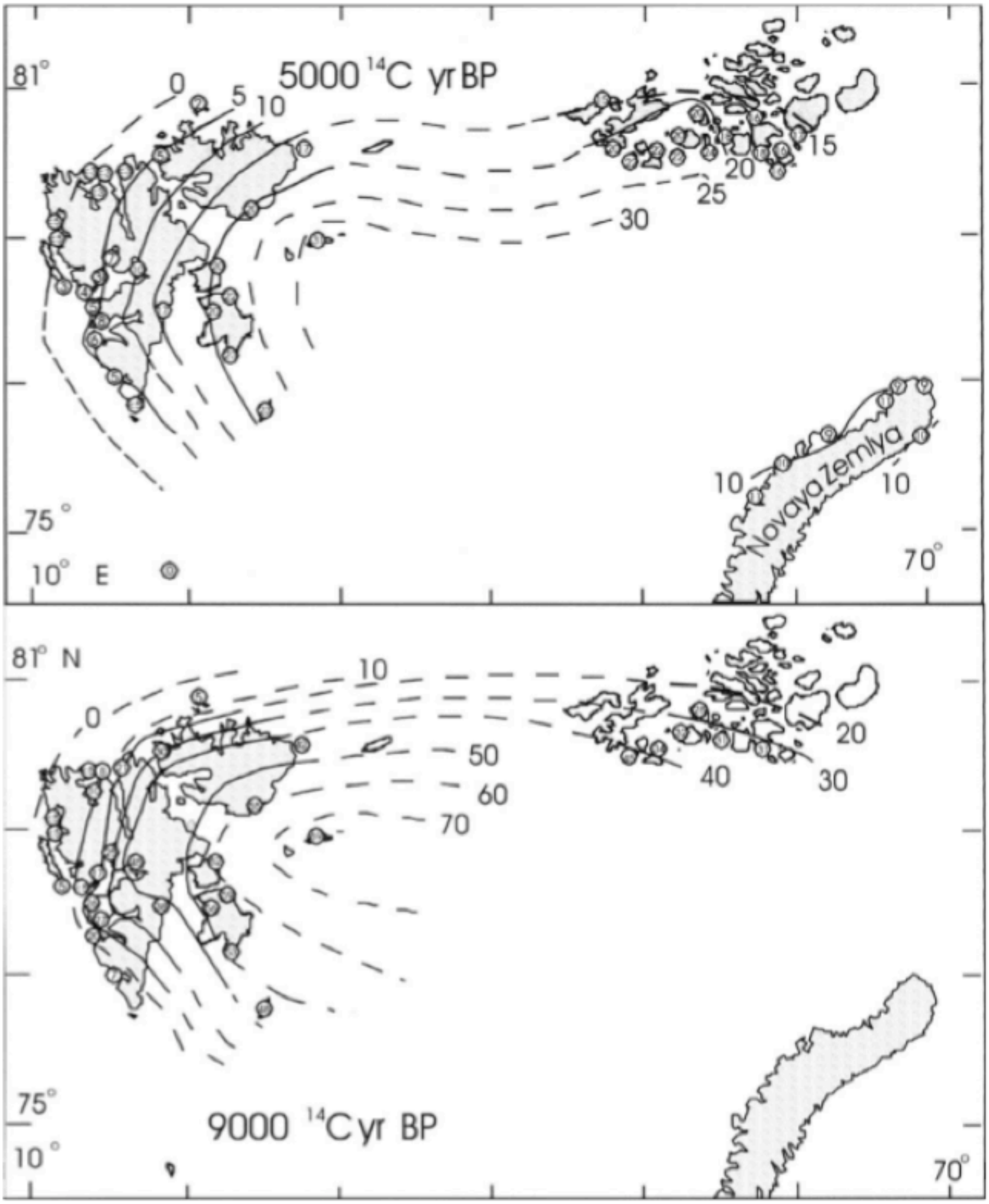

Fig. 6.3 Estimated emergence isobases for the Barents Sea area since 5000 and 9000 14Cyr BP. Circles indicate placement and value of emergence data point, for Svalbard see Fig. 2.1. (Forman *et al*., 2004).

7 CONCLUSIONS

The view of the beds beneath former ice sheets given by marine-geophysical records provides important data for the reconstruction of former ice sheet flow directions and dynamics. Geomorphic seafloor features have been described in fjords and troughs that show the dynamics of the Late Weichselian ice sheet that covered Svalbard and its retreat in the troughs and fjord systems.

Submarine landforms produced by the Late Weichselian ice sheet are characteristic for the continental shelves and major fjords west and north of Svalbard. MSGL are the result of soft sediment deformation at the base of fast flowing ice streams. The presence of fast flowing ice streams within the former ice sheet, suggest that the Late Weichselian ice sheet was divided into fast flowing ice stream separated by slower moving ice.

Submarine morphological evidences demonstrate that a large ice sheet reached the shelf of western and northern Svalbard at the LGM (Fig. 6.2). It is therefore unlikely that a restricted ice sheet was over Svalbard with areas on the west coast ice free at the LGM.

The behavior of the ice margin during retreat can be inferred from submarine landforms. Well preserved MSGL that are not overlain by small transverse ridges or grounding zone wedges indicate that the ice streams underwent rapid thinning and retreat through cross-shelf troughs and deep fjords. Grounding zone wedges indicate an episodic retreat. While small transverse ridges indicate a slow quasi-continuous ice front retreat, more common on shallower banks between troughs (Dowdeswell *et al*. 2008).

For Spitsbergen, Edgøya and Barentsøya the pattern of postglacial emergence is very well constrained, but for other islands of the archipelago, like Nordaustlandet, they have scattered data coverage (Forman *et al.* 2004).
For northwestern Spitsbergen there is clear evidence for an early deglaciation by c. 13 ka ago, resulting with variable relative sea level response with transgressive and regressive events, compared to deglaciation at c. 10.5-10 ka the uplift was essentially exponential for eastern Svalbard. On Kongsøya the maximum isostatic compensation of >100 m aht is registered (Fig. 2.2). Kongsøya and adjacent eastern Svalbard and the adjacent Barents Sea are inferred to have sustained the maximum ice sheet load (Lambeck, 1995).

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